The Bedrock Geology of Massachusetts

A. The Pre-Silurian Geology of the Rowe-Hawley Zone

B. Stratigraphy of the Connecticut Valley Belt

C. Post-Taconian Structural Geology of the Rowe-Hawley Zone and the Connecticut Valley Belt West of the Mesozoic Basins

D. The Whately Thrust: A Structural Solution to the Stratigraphic Dilemma of the Erving Formation

U.S. GEOLOGICAL SURVEY PROFESSIONAL PAPER 1366-A-D

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NORMAN L. HATCH, JR., *Editor*

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Editor's Preface to Chapters A through D

This Professional Paper was planned as a companion to the bedrock geologic map of Massachusetts (Zen and others, 1983; hereafter referred to as the State bedrock map). It is being published as lettered chapters of Professional Paper 1366, four of which are included in this volume. Compilation of the geology for the State bedrock map was completed in 1980. Some of the chapters in this Professional Paper reflect field or laboratory data that were gleaned as much as six years later. Each chapter was prepared, however, with the objective of explaining and further describing the geology as portrayed on the State bedrock map. In some instances, information and interpretations developed since 1980 have caused chapter authors to suggest revisions that they would make to the map if they were able to redraw it, but in each case these suggested revisions are discussed in the context of the map as it was published.

The previous State bedrock map (which also showed the geology of Rhode Island) was published in 1917 by Benjamin K. Emerson as U.S. Geological Survey Bulletin 597. (The publication date of Bulletin 597 is 1917. Some confusion arises from the fact that the bedrock map of the two States, which is included in the pocket of the Bulletin, bears the date of 1916.) All who were involved in the preparation of the new bedrock map, particularly those responsible for the parts of the State in which Emerson himself had done the original field work, feel a great deal of respect for Professor Emerson and his remarkably perceptive and thorough understanding and portrayal of the geology. Although the new map is very different from Emerson's in many aspects, particularly with regard to the interpretation of the geologic history, the basic distribution of map units is remarkably similar.

The State bedrock map and this report are direct outgrowths of a cooperative geologic mapping program between the U.S. Geological Survey and the Commonwealth of Massachusetts, which was begun in 1938. They also include the results of more than 25 years of mapping and topical studies by faculty and students at the University of Massachusetts at Amherst and at many other colleges and universities.

The subdivision of the material in this Professional Paper into the constituent chapters is based on the grouping of the 343 individual lithic units on the State bedrock map into the eight lithotectonic packages discussed by Hatch and others (1984). The temporal and geographic distribution of these eight packages are indicated on figures 1 and 2. Also indicated on the figures are the geographic and geologic coverage of the chapters included in this volume. In this packaging scheme, the older, primarily pre-Silurian, rocks are grouped into five "zones" whose exposed and buried parts completely cover the State. From west to east these zones are the Taconic-Berkshire, the Rowe-Hawley, the Bronson Hill, the Nashoba, and the Milford-Dedham. Their mutual boundaries are, or could reasonably be interpreted to be, faults. Overlying and overlapping the zones in the central part of the State are the Connecticut Valley and Merrimack "belts" of primarily Silurian and Devonian strata. Their mutual boundary is somewhat arbitrarily taken to be the east contact of the easternmost exposed Silurian Clough Quartzite. Finally, the Mesozoic "basins" unconformably overlie the Connecticut Valley belt.

For some packages, all aspects of the geology are treated in the same chapter. For others, aspects such as the structure, metamorphism, and tectonics are discussed separately from stratigraphy and lithology. These differences in treatment resulted from peculiarities of the geology and the preferences of the individual authors. Most of the plutonic rocks of the State are described and discussed in separate chapters. A particularly knotty problem concerning the stratigraphic and structural relations of some of the strata in the Connecticut Valley belt on both sides of the Mesozoic basins is treated in a separate chapter on the Whately thrust.

Many of the lithologic subdivisions of formal units on the State bedrock map have not been given formal names. In order to avoid potentially cumbersome discussions of such things as "the thick-bedded micaceous quartzite and mica schist unit of the XYZ Formation," many chapter authors have chosen to refer to such units simply by their map symbols. Thus the micaceous quartzite, quartz-mica-garnet schist, and calc-silicate unit of the Devonian Goshen Formation may be referred to simply by its map symbol "Dgq," but in a context where the reader will be easily guided to the correct unit.

The terms "granulite" and "granofels" have been used rather arbitrarily and interchangeably throughout this Professional Paper, although on the State bedrock map the term "granofels" was used exclusively. Both terms are used to describe a metamorphic rock

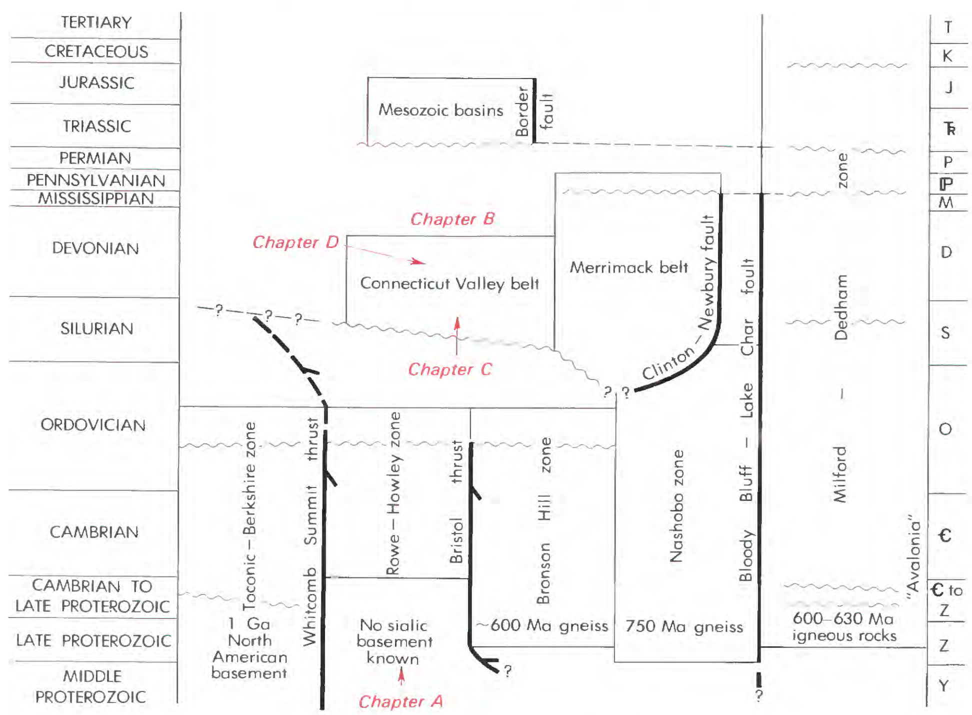

FIGURE 1.—Diagram simplified from the "Correlation of Map Units" on the State bedrock map showing the eight lithotectonic packages into which the rock units have been divided. Also indicated are the letter designation(s) of the chapter(s) in this volume covering various aspects of the geology. Modified from Hatch and others (1984, fig. 1).

composed predominantly of even-sized, interlocking granular minerals; no implication as to the grade of metamorphism is intended by either term. The choice of words merely reflects individual author preference, and we hope that no confusion to the reader will result from the unrestrained use of two words for the same kind of rock.

This volume contains four chapters. Chapter A on the pre-Silurian geology of the Rowe-Hawley zone describes the Cambrian and Ordovician stratified and plutonic rocks of the zone and also discusses the structures within these rocks that the authors attribute to the Taconian orogeny. Chapter B is devoted to description and discussion of the rocks in the Connecticut Valley belt. Chapter C treats the folds, faults, and metamorphism of the Rowe-Hawley zone and that part of the Connecticut Valley belt west of the Mesozoic basins that is believed to have formed during the Acadian orogeny. Finally, Chapter D is devoted to a discussion of the somewhat controversial Whately thrust fault, which the authors have proposed to explain the apparent incompatibility between the Silurian-Devonian stratigraphic sequences east and west of the Mesozoic basins.

We would herein like to acknowledge the invaluable contributions to this Professional Paper of two key people. Jewel Dickson did the cartographic work on the majority of the illustrations, the principal exceptions being those prepared by author Peter Robinson. Finally, Kathleen Krafft suffered bravely over the years with the editor and authors of this Professional Paper as its technical editor.

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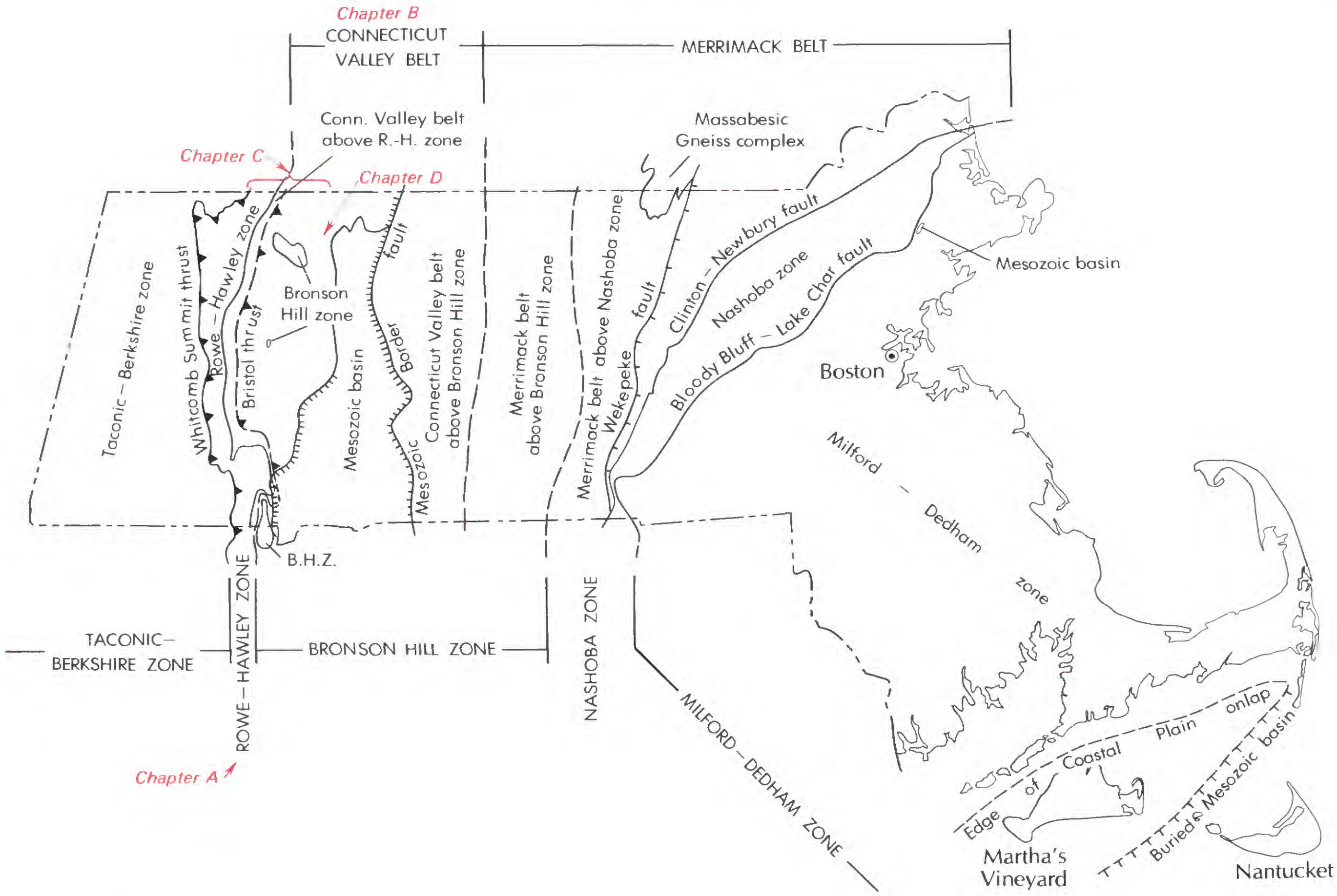

FIGURE 2.—Map of Massachusetts showing the geographic distribution of the eight lithotectonic packages into which the rock units of the State have been grouped, and the letter designation(s) of the chapter(s) in this volume in which aspects of the geology are discussed. Modified from Hatch and others (1984, fig. 2).

CONTENTS

[Letters designate the chapters]

The Pre-Silurian Geology of the Rowe-Hawley Zone

By ROLFE S. STANLEY, UNIVERSITY OF VERMONT, *and* NORMAN L. HATCH, Jr., U.S. GEOLOGICAL SURVEY

THE BEDROCK GEOLOGY OF MASSACHUSETTS

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CONTENTS

ILLUSTRATIONS

THE PRE-SILURIAN GEOLOGY OF THE ROWE-HAWLEY ZONE

By ROLFE S. STANLEY[1] and NORMAN L. HATCH, JR.

ABSTRACT

The Rowe-Hawley lithotectonic zone, western Massachusetts, consists of metasedimentary and metavolcanic rocks between the west contact of the Rowe Schist and the east contact of the Hawley and Cobble Mountain Formations of Late Precambrian to pre-late Middle Ordovician age. All rocks in this zone, with the exception of minor Devonian intrusions discussed elsewhere, predate the Taconian orogeny. The Rowe Schist consists of complexly intercalated lenticular masses of green schist, gray schist, amphibolite, and serpentinized ultramafic rock. Intercalation of these rocks may be largely tectonic. The Moretown Formation to the east consists of light-gray pinstriped granofels and mica schist. The granofels typically contains significant amounts of plagioclase. North of the Massachusetts Turnpike, the Moretown is bounded on the east by felsic and mafic metavolcanic rocks and sulfidic black slates of the Hawley Formation of supposed Middle Ordovician age. Near the turnpike, the black slates pass by facies change southward into the basal of four members of the silvery-gray mica schists and gneisses of the Cobble Mountain Formation. Plagioclase gneiss and amphibolite of the pre-Silurian Collinsville Formation core three of four domes that arch the Silurian-Devonian blanket of the Connecticut Valley synclinorium. The Cobble Mountain is thought to rest unconformably on the Collinsville Formation.

Lenses of ultramafic rock in the Rowe Schist and lower Moretown are thought to be slivers of oceanic crust mechanically emplaced into Rowe slope-rise sediments during an early stage of Taconian collision. Additional lenses in member C of the Cobble Mountain Formation are thought to represent olistostromal blocks shed eastward from an emerged part of the accretionary prism somewhat later in the Taconian orogeny.

Pre-Silurian intrusive rocks are represented only by a gneissic granite northwest of Plainfield, a few small sills of foliated granite of possible syn-thrust (Taconian) age at Borden Brook Reservoir 5 km south of Blandford, and a small body of diorite southwest of Rowe.

The exposed rock units of the Rowe-Hawley zone on the eastern limb of the Berkshire massif form a simple linear map pattern which contrasts with the more complex pattern of correlative rocks to the west in the Taconic allochthons and autochthonous platform. This apparently simple linear pattern is deceptive—detailed mapping in the 1970's has shown that major faults not only separate several of the formations but penetrate much of the Rowe Schist. The Middlefield thrust zone, which separates the Hoosac Formation from the Proterozoic Y rocks, and the Whitcomb Summit thrust, which separates the Rowe Schist from the Hoosac, are regionally extensive fault zones on which westward displacement during the Taconian orogeny was tens to hundreds of kilometers. These estimates are based on palinspastic restorations of the Taconic allochthons, which are rooted within the Hoosac Formation and beneath the Whitcomb Summit thrust. Displacement along thrusts in the serpentinite-bearing Rowe Schist is thought to be in the same order of magnitude, although there is no basis for palinspastically estimating the distance, without recourse to plate tectonic models. Thrusts may be present in the Moretown and Hawley Formations but have not yet been identified.

The feldspathic schists and gneisses of the Cobble Mountain Formation are thought to represent volcanogenic flysch eroded from the westward-advancing Bronson Hill volcanic arc and deposited in a forearc basin that contained black muds and cherts. Member C of the Cobble Mountain Formation is separated from the underlying member B by the Winchell Mountain thrust, which is thought to have displaced member C eastward as a near-surface backthrust during the latter part of the Taconian orogeny.

Plagioclase gneisses and amphibolites of the Collinsville Formation are tentatively correlated with the Ammonoosuc Volcanics and the Monson-Fourmile Gneisses of the Bronson Hill anticlinorium to the east. We suggest that the Hawley-Cobble Mountain-Partridge strata originally unconformably overlay the Collinsville-Bronson Hill gneisses. We further suggest that the present base of the Collinsville is a thrust contact with the underlying Moretown and that the inferred unconformity at the base of the Hawley-Cobble Mountain-Partridge cover truncates this thrust. Evidence in support of this model is found to the south in Connecticut in the Bristol and Waterbury domes, where the Taine Mountain (Moretown equivalent) Formation is exposed structurally below the Collinsville gneisses.

Tectonically, the Rowe-Hawley zone is an extensive belt of imbricated thrusts of Taconian age, which bound distinctive linear lithotectonic belts of pre-late Middle Ordovician rock. This configuration has been severely overprinted by Acadian deformation, which is considered in Hatch and Stanley (Ch. C, this volume).

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INTRODUCTION

The term "Rowe-Hawley zone," as used in the explanation for the Massachusetts State bedrock map (Zen and others, 1983; Hatch and others, 1984) and in this paper, is defined as the lithotectonic interval between the west margin of the Rowe Schist (the Whitcomb Summit thrust) on the west and the east edge of the principal outcrop belt of the Hawley and Cobble Mountain Formations, or the west boundary of the Bronson Hill zone (the postulated Bristol thrust), on the east. The zone consists of late Middle Ordovician(?) and older stratified rocks (fig. 1) thought to have been deposited on the eastern edge of pre-Taconian North America, on an island arc-microcontinent complex to the east, or on intervening oceanic crust. The collision of these two continental masses produced what is referred to as the Taconian orogeny.

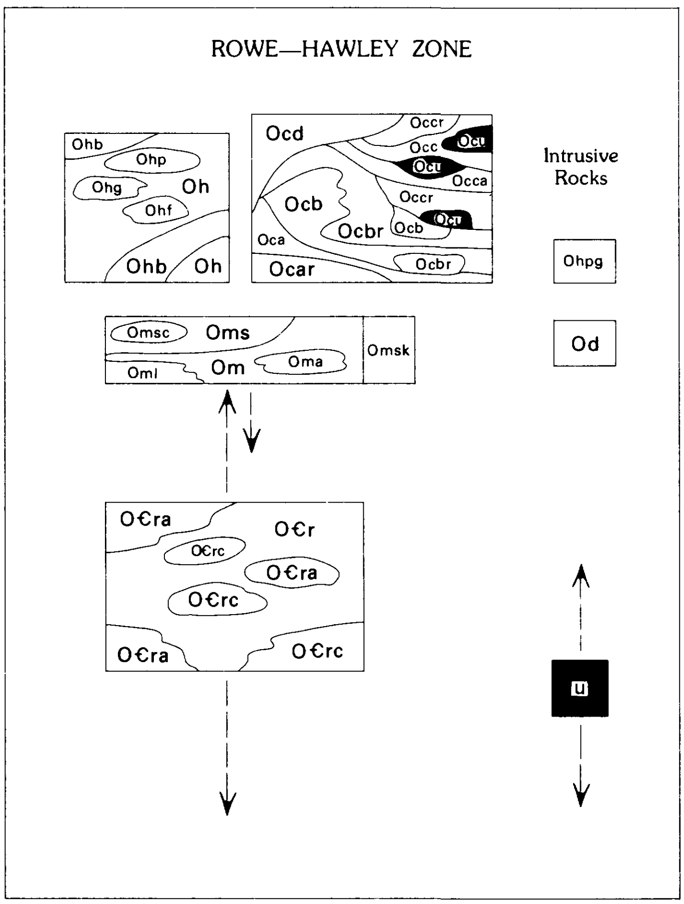


FIGURE 1.—Diagram showing Cambrian and Ordovician rocks of the Rowe-Hawley zone described in this chapter. From "Correlation of map units" of the State bedrock map (Zen and others, 1983). Symbols beginning with "Oh" represent members of the Hawley Formation. Symbols beginning with "Oc" represent members of the Cobble Mountain Formation. Symbols beginning with "Om" represent members of the Moretown Formation. Symbols beginning with "OꞒr" represent members of the Rowe Schist. Ohpg is gneiss at Hallockville Pond, Od is intrusive diorite, and u is ultramafic rock.

The Rowe-Hawley zone is a term that was coined and defined during preparation of the Massachusetts bedrock map. The regional extrapolation of the zone north and south of Massachusetts is discussed elsewhere (Stanley and Ratcliffe, 1983, 1985), but a brief summary is given here and is illustrated on figure 2. In Connecticut we would bound the zone on the west by Cameron's line, which separates Rowe- and Moretown-equivalent rocks on the east from Hoosac-equivalent rocks on the west. The east edge of the presently exposed zone is the narrow belt of Silurian-Devonian Straits Schist and Russell Mountain-equivalent rocks (Rodgers, 1985). In Vermont the west edge of the Rowe-Hawley zone is drawn along the west contact of the Ottauquechee Formation as shown on the Centennial Geologic Map of Vermont (Doll and others, 1961).

The pre-Silurian basement upon which the Silurian and Lower Devonian strata of the Connecticut Valley synclinorium (Hatch and others, Ch. B, this volume) were deposited is presently exposed in the cores of the Shelburne Falls, Goshen, Woronoco, and Granville domes and in the Whately anticline. These are the only areas of outcrop accessible to us of the Ordovician or older rock geographically between the Rowe-Hawley zone and the Bronson Hill zone to the east (Tectonic map of Massachusetts *in* Zen and others, 1983). Because of their resemblance to gneisses in the Bronson Hill domes, particularly the Monson and Fourmile Gneisses, the gneisses in the Shelburne Falls, Goshen, and Granville domes are thought to be continuous beneath the Paleozoic cover with rocks of the Bronson Hill zone and to represent the leading edge of the eastern microcontinent that overrode the ancient North American cratonic plate (Hatch and others, 1984, fig. 2). These gneisses are discussed in this chapter because of their present geography, their close relationship to Rowe-Hawley zone strata, and our familiarity with their lithology, sequence, and structure. Pre-Silurian black schists in the core of the Whately anticline could be assigned, on the basis of lithology and stratigraphic position, to either the Partridge or the Hawley Formation; they are here assigned to the Partridge because the lithic sequence within which they occur more closely resembles that of the Bronson Hill. The boundary between the Rowe-Hawley zone and the Bronson Hill zone to the east is concealed beneath the Silurian-Devonian strata of the Connecticut Valley belt somewhere west of the Shelburne Falls and associated domes and east of the present east contact of the Hawley Formation.

Until the late 1970's, the succession of metamorphic units along the east limb of the Berkshire massif was

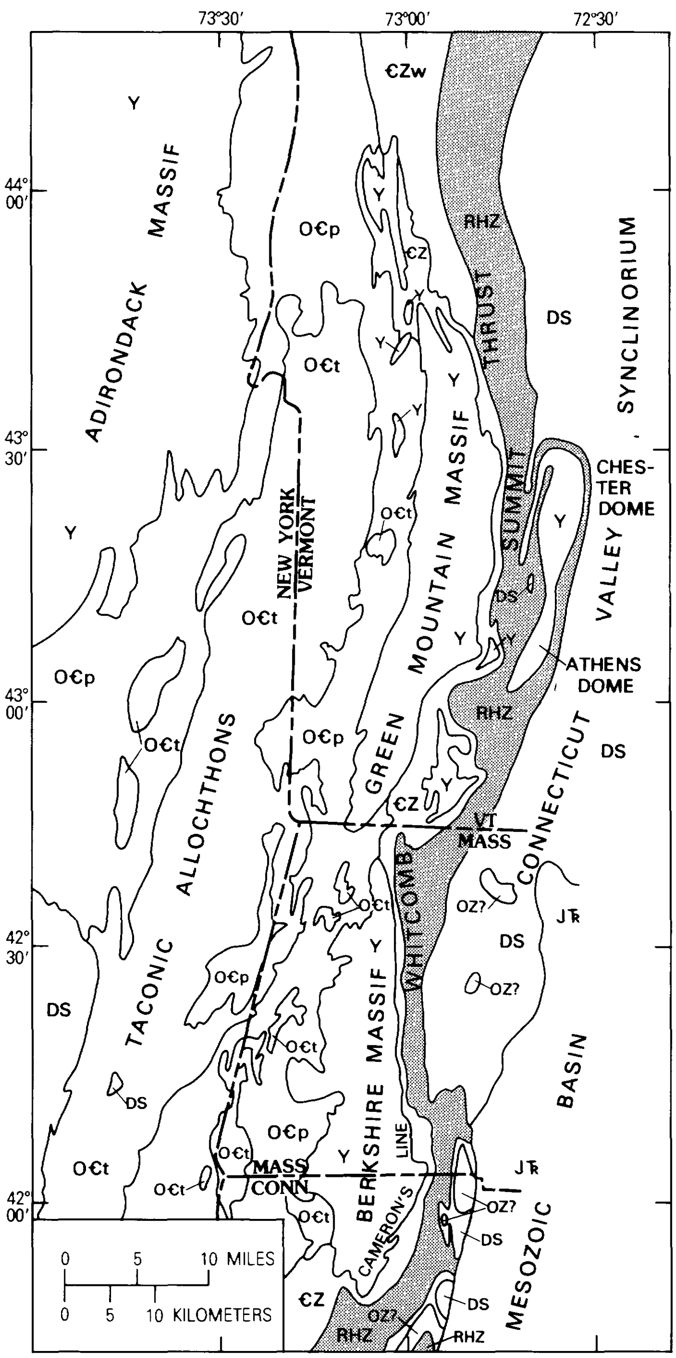


FIGURE 2.—Simplified lithotectonic map of western New England showing the Rowe-Hawley zone (shaded) and its regional setting. Letter symbols are as follows: JTR, Jurassic-Triassic rocks; DS, Devonian and Silurian strata; RHZ, Eugeoclinal Ordovician and Cambrian strata and ultramafic rocks of the Rowe-Hawley zone; ЄZw, Cambrian and Proterozoic Z strata of the western part of the eugeocline, which lack ultramafic rocks; OЄt, Ordovician and Cambrian strata of the Taconic allochthons; OЄp, Ordovician and Cambrian platform and thin western basal clastic rocks; OZ?, Ordovician to Proterozoic Z gneisses of the western Massachusetts and Connecticut domes; ЄZ, Cambrian to Proterozoic Z eastern clastic rocks; Y, Proterozoic Y rocks.

considered depositional (see, for example, Hatch and Stanley, 1973), although some faults were recognized in this interval to the north in Vermont (Chang and others, 1965; Thompson, 1972) and to the south in western Connecticut (Stanley, 1968, fig. 3; Hatch and Stanley, 1973, pl. 1). Norton (1971, 1975) was the first to suggest that a regionally extensive fault zone, the Middlefield thrust zone, separated the Proterozoic (Y) rocks of the Berkshire massif from the Hoosac Formation (allochthonous Hoosac of current usage; Zen and others, 1983). Berkshire gneisses intercalated with Hoosac rocks are found throughout this zone in which the thickness of the Hoosac changes drastically from Vermont to western Connecticut (Rodgers, 1985). Mylonitic fabrics of Taconian age are well preserved despite younger Acadian metamorphism.

During compilation of the State bedrock map, it became evident that the Hoosac-Rowe contact was also a significant thrust zone. Lithic units mapped by Norton (1967, 1974a,b), Hatch and others (1966), Hatch, Norton, and Clark (1970), Hatch and Hartshorn (1968), Hatch and Stanley (1976), and Ratcliffe (1979b) in both the Hoosac and the Rowe were truncated by their mutual formational contact. In 1978 Stanley described this zone and named it the Whitcomb Summit thrust for Whitcomb Summit about 6 km south of the Vermont State line (fig. 4). Earlier work along the Rowe belt (Chidester and others, 1967; Hatch and Hartshorn, 1968; Osberg and others, 1971; Hatch, 1969; Hatch, Norton, and Clark, 1970; Hatch and Stanley, 1976; Norton, 1967, 1974a,b) showed that the Rowe consisted of many lenses of three metamorphic lithologies plus numerous lenses of ultramafic rock. This fabric was originally interpreted by Hatch and others (1966) as a complex of sedimentary facies tongues and intrusive ultramafic pods within the stratigraphic interval represented by the Pinney Hollow, Ottauquechee, and Stowe Formations in Vermont. The lack of continuity of the sequence of these Vermont formations southward across Massachusetts resulted in Hatch and others' (1966) redefinition of the Rowe Schist. In 1968, Hatch and others (Hatch, Schnabel, and Norton, 1968, p. 179) did suggest the possibility that the complex interlayering might be in part tectonic. Zen (1972, p. 44) suggested that this lithic discontinuity was actually the "result of major thrusts that repeat as well as eliminate parts of the normal stratigraphic section." Analysis of the quadrangle mapping cited above and of detailed 1:13,000-scale mapping by Knapp (1977) in southernmost Massachusetts now suggests that the discontinuity of lithic units and the presence of the ultramafic rocks both result from imbricate thrusting. This hypothesis is supported by 1:10,000-scale mapping in equivalent rocks in the serpentinite belt in northern Vermont where Stanley and Roy (1982; Stanley and others, 1984) have mapped numerous faults along which are slivers of serpentinite. We thus conclude that both the remarkable linearity of formations and the internal lenticularity within these formations in the Rowe-Hawley zone, particularly the interval from the Hoosac-Rowe contact to the western part of the Moretown Formation, are largely tectonic in origin.

Although Acadian deformation and metamorphism profoundly influenced the observed structural and mineral fabric, particularly within the eastern part of the Rowe-Hawley zone, we believe that the present distribution of lithic units within the zone is largely due to a combination of lower Paleozoic depositional patterns and severe tectonism before and during the classical Taconian orogeny (Stanley and Ratcliffe, 1980, 1983, 1985). In this chapter we emphasize the Taconian and earlier heritage of the Rowe-Hawley zone; the Acadian deformational events are discussed elsewhere (Hatch and Stanley, Ch. C, this volume). We discuss briefly in this chapter the stratigraphy within the belt, the Ordovician and older intrusive rocks, the ultramafic and related rocks, the major tectonic surfaces and zones, and the relationship between the Rowe-Hawley zone and the rocks of the western edge of the Bronson Hill plate (Robinson and Hall, 1980) exposed in the Granville, Goshen, and Shelburne Falls domes. Finally, we show, by means of a sequence of successively retrodeformed cross sections (fig. 28), our interpretation of the tectonic evolution of this area. Although this interpretation is clearly speculative, we hope that figure 28 will enable and encourage future students of this area to see the basis for our thinking and our model and to correct it as new data and new ideas become available.

ACKNOWLEDGMENTS

Many of the ideas discussed in this chapter evolved directly from compiling the mapped geology of the Rowe-Hawley zone and developing a coherent synthesis that was compatible with the known geologic relations to the west in the Taconic allochthons and Berkshire massif. Stanley's research in equivalent rocks of the Rowe-Hawley zone in northern Vermont and his work in Taiwan in the western Pacific were particularly helpful in formulating a plate-tectonic interpretation of western Massachusetts. Much of the primary field mapping on which the compilation was based was done in the 1960's before the advent of plate-tectonic models for orogenic belts. Re-examination of critical areas within the zone, for compilation of the State map, not surprisingly resulted in some new interpretations of existing mapping, both of our own and of others.

We are indebted to a number of our colleagues for their contributions to the field mapping but more particularly for their contributions to the ideas presented here. Stephen A. Norton not only did much of the mapping along the westernmost margin of the Rowe-Hawley zone but also, by proposing and documenting the Middlefield thrust zone, introduced the concept of thrusting within the "eugeosynclinal sequence" east of the Berkshire massif, although he believed the age of thrusting to be Acadian. Ratcliffe and Mose (1978) later proved it to be Taconian in age. This chapter expands on that concept. Douglas A. Knapp did most of what little systematic detailed mapping was done in the zone after modern plate-tectonic ideas were developed. As a result, he was able to consider and test these concepts during his mapping. We have relied extensively on the results of his study (Knapp, 1977, 1978) and of Stanley and Roy's work in northern Vermont (Stanley and Roy, 1982; Stanley and others, 1984). The concept of the Bristol thrust evolved from a brief field trip and discussion with Philip H. Osberg in the fall of 1975, although he may or may not now wish to associate himself with the interpretation. Leo M. Hall's map of the Shelburne Falls dome provided the first detailed picture of both the stratigraphy and the structure of the core rocks of that dome, and his discussions with us have significantly influenced our thinking about those rocks and their probable source area. Discussions with David S. Harwood on his detailed mapping in the Berkshires and his ideas on the regional geology of southwestern New England have stimulated our thinking for many years.

Finally, our long-term association and collaboration with our neighboring compilers to the east and west have not only constrained our thinking about the Rowe-Hawley zone but have, we hope, made the ideas presented and summarized here compatible with a broader regional cross section. Peter Robinson, to our east, through his detailed knowledge of the Bronson Hill zone, has particularly influenced our interpretation of the pre-Silurian rocks between the true Rowe-Hawley zone and the main mass of the Bronson Hill zone east of the Mesozoic basin. We are most particularly indebted to Nicholas M. Ratcliffe, our neighbor to the west, for many days in the field together and for countless hours of discussion in and out of the field in an effort to understand better the evolution of the Taconian orogeny in western New England. The critical comments of Nicholas Ratcliffe and David Harwood substantially improved the manuscript.

STRATIGRAPHIC RELATIONSHIPS OF THE ROWE-HAWLEY ZONE

The four major formations within the Rowe-Hawley zone are the Rowe Schist, the Moretown Formation, the Hawley Formation, and the Cobble Mountain Formation. These units, and their mapped subunits, have been described in detail in the quadrangle maps in the belt (see references, State bedrock map) and those descriptions are not repeated here. A succinct stratigraphic summary was given by Hatch and Stanley (1973, p. 5-16). The present discussions emphasize critical contact relations and modifications that have resulted from compilation for the State bedrock map. Although the Collinsville Formation is shown as part of the Bronson Hill zone on the explanation to the State bedrock map, as noted in the Introduction, we include here a discussion of the Collinsville because it is germane to our consideration of the junction between the Rowe-Hawley zone and the Bronson Hill zone. The general stratigraphic and tectonic relations of these formations are shown diagrammatically in figure 3. Although we hope that the reader has the State bedrock map (Zen and others, 1983) at hand, figure 4 is provided as a simplified map of the Rowe-Hawley zone on which are shown the principal geographic features referred to in this chapter.

Other than the Chester Amphibolite of Emerson (1898), a term which we restrict to the large body of amphibolite immediately west of Chester village, none of the members or submembers of the formations have formal or informal stratigraphic names. Therefore, to avoid a proliferation of new names and to simplify discussion, members and submembers are referred to in the following discussion only by their letter symbol designation on the State bedrock map (see also fig. 1).

It should be noted here that although the stratified rocks of the Rowe-Hawley zone are all, by definition,

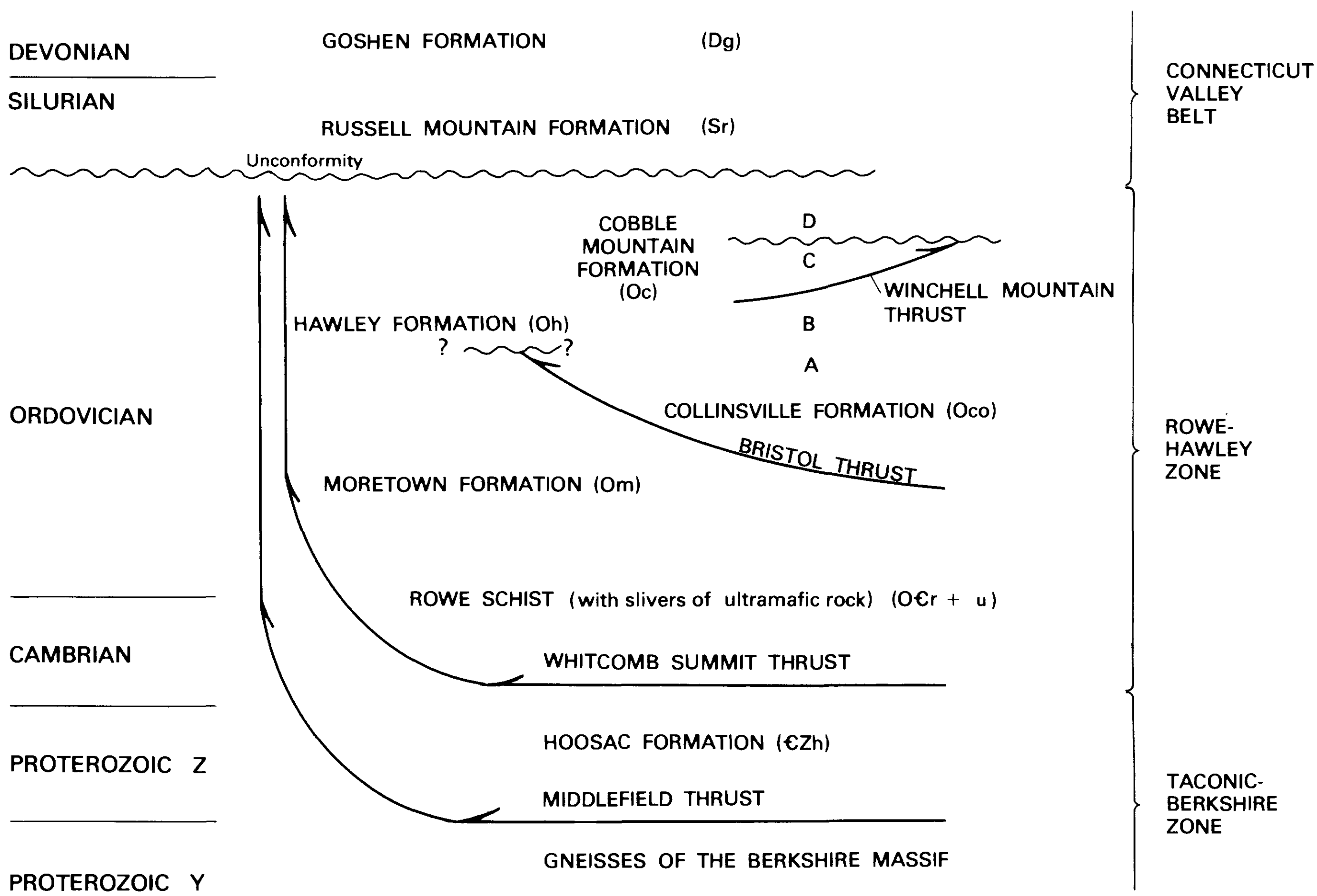


FIGURE 3.—Generalized stratigraphic and structural relations between the major formations in and immediately adjacent to the Rowe-Hawley zone, just prior to the deposition of Silurian and Devonian rocks.

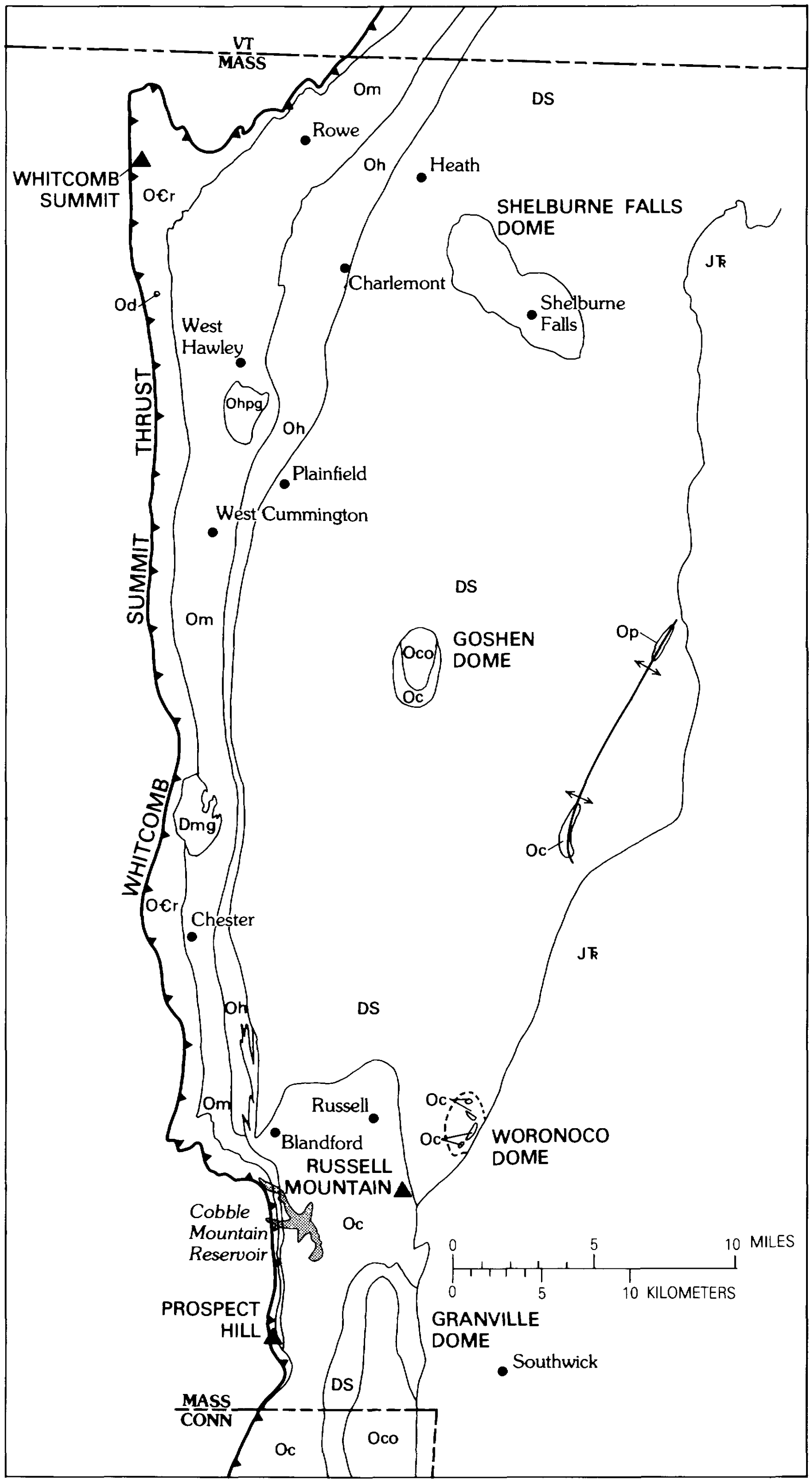


FIGURE 4.—Simplified geologic map of the Rowe-Hawley zone in Massachusetts showing principal geographic features referred to in the text. Letter symbols are as follows: O€r, Rowe Schist; Om, Moretown Formation; Oh, Hawley Formation; Oc, Cobble Mountain Formation; Oco, Collinsville Formation; DS, undifferentiated Devonian and Silurian rocks of the Connecticut Valley belt; Ohpg, gneiss at Hallockville Pond; Dmg, Middlefield Granite; Op, Partridge Formation; Od, intrusive diorite.

pre-Silurian in age, and the focus of this chapter is on the pre-Silurian geology of the zone, all of the rocks have also been subjected to the deformation and metamorphism of the later Acadian orogeny. Thus the rocks, as we see them today and as they are described herein, inevitably owe at least some of their present texture and mineralogy to post-Taconian events. (Also see the metamorphic map on the State bedrock map.)

ROWE SCHIST (O€r, O€rc, O€ra)

Hatch and others (1966) redefined Emerson's (1898) Rowe Schist for the following reasons: (1) A succession of rocks equivalent to the stratigraphic sequence Pinney Hollow, Chester, Ottauquechee, Stowe mapped in southern Vermont (Skehan, 1961; Doll and others, 1961) was not present at the Massachusetts-Vermont State line and could not be recognized anywhere south thereof in Massachusetts. (2) The interval in western Massachusetts between the eastern contact of the Hoosac and the western contact of the Moretown is occupied by a variable thickness of a variable number of lenses of Pinney Hollow/Stowe-like green schist, Ottauquechee-like gray schist, amphibolite, and ultramafic rock. (3) The Chester Amphibolite of Emerson (1898), rather than being a single stratigraphic horizon across Massachusetts separating lithically distinct strata above and below it, is a series of lenses of amphibolite scattered over a wide outcrop belt that in turn consists of lenticular masses of green (Pinney Hollow/Stowe-like) and gray (Ottauquechee-like) schist. (4) The schist, which Emerson (1898) called Rowe, immediately *below* (west of) the type body of Chester Amphibolite at Chester, Mass., is identical with the schist immediately *above* (east of) the type Chester body that Emerson assigned to the basal part of his Savoy. North and south of the terminations of the type Chester body, where the two schists are in contact, they are indistinguishable. The redefinition proposed by Hatch and others (1966) applied Emerson's term "Rowe Schist" to all of the rocks between the Hoosac to the west and the Moretown to the east with the tacit assumption that all lithic subdivisions would be mapped as informal members of the formation. Subsequent field work along this belt has proven this system to be highly satisfactory within the redefined Rowe but has suggested modification of the original mapping of Chidester and others (1967) along the Rowe-Hoosac contact in the northern part of the State.

Of the three lithic members of the Rowe Schist, designated O€r, O€rc, and O€ra on the State map, the most abundant and most diagnostic member of the formation is O€r. In northern Massachusetts, O€r is a light-green, fine-grained muscovite-quartz-chlorite schist containing scattered 1-mm garnets and, commonly, 0.5-mm octahedra of magnetite. Bedding can be recognized only very locally. The rock is pervasively schistose, and lenses of vein quartz as much as 1 cm wide and 10 cm long are common within the schistosity (fig. 5*A*). South of approximately the latitude of Blandford, as metamorphic grade increases, the color of the rock changes to a very light green or silvery gray, the grain size increases markedly, and staurolite and kyanite become important constituents (fig. 5*B*). The rocks of O€r are *not* rusty weathering. They are lithically similar to schists of the Pinney Hollow and Stowe Formations of Vermont.

Member O€rc of the Rowe is marked by distinct gray, rather than green, color, due to the presence of carbonaceous material. The schists of O€rc in both the northern and southern parts of the State consist primarily of quartz, muscovite, biotite, and plagioclase, and accessory carbonaceous matter. Chlorite, garnet, and aluminum silicates are relatively rare. In contrast to the general absence of bedding in O€r, bedding is locally indicated by layers as much as 15 cm thick of white, gray, or black quartzite. South of Blandford, an increase in grain size and an apparent decrease in carbonaceous matter make it increasingly difficult to distinguish O€rc at the base of the formation from rocks of the Hoosac Formation in contact to the west (Zen and others, 1983). Outcrops of O€rc may be somewhat rusty weathered, but this effect is generally not strong.

Member O€ra consists of well-foliated, locally well-layered, black to dark-green hornblende-plagioclase amphibolite (fig. 6). Epidote is locally abundant, and garnet and sphene are common accessories. Opaque minerals are generally absent. Grain size varies from fine to very coarse, and wide extremes of grain size and equigranularity may be present in a given exposure and even in adjacent beds. Pillow structure was not observed in these amphibolites; the fine compositional lamination of many exposures suggests that they are mafic tuffs.

Near the boundary between the Hoosac Formation and the Rowe Schist in the northern part of the State are intervals of light green quartz-muscovite-chlorite schist containing abundant, conspicuous porphyroblasts of albite. Although the light-green color and general appearance are compatible with the green schist (O€r) of the Rowe, the albite porphyroblasts are not characteristic of the Rowe and instead are typical of the Hoosac. Remapping in the North Adams quadrangle has shown that large areas of albite-rich, light-green, aluminous schist (€Zhga on the State bedrock map) are complexly interlayered with other members of the

A

B

FIGURE 5.—Light-green aluminous schist of member C€r of the Rowe Schist. *A*, Near Whitcomb Summit thrust at Whitcomb Summit, Route 2. Note small lenses of isoclinally folded white quartz (near 25-cent piece) that characterize the unit. *B*, 3.5 km west of Blandford. Dark spots are staurolite porphyroblasts that cut S_2 schistosity. Pen is 14 cm long.

FIGURE 6.—Banded amphibolite of member O€ra of the Rowe Schist. Pen is 14 cm long.

Hoosac Formation (Ratcliffe, 1979b, fig. 5). This intercalation probably is stratigraphic, and the presence of green metapelitic rocks throughout much of the Hoosac and the widespread occurrence of albite porphyroblasts so characteristic of the rest of the Hoosac have persuaded us that these green albite-rich schists are stratigraphically related to the Hoosac rather than to the Rowe. Therefore, we herein modify the definition of the Rowe Schist to exclude these green albite-porphyroblastic rocks; the State bedrock map reflects this change from the earlier mapping of Chidester and others (1967).

Earlier evaluations of the age of the Rowe (Hatch and others, 1966) were based on correlation with strata continuous with the Rowe in southern Vermont (Pinney Hollow, Ottauquechee, and Stowe Formations of Doll and others, 1961) and on the assumption that the succession of formations eastward from the Precambrian rocks of the Green Mountain and Berkshire massifs was a true eastward-younging stratigraphic sequence. Although we are still comfortable with the general correlation with Vermont, our present views on the structural history of these rocks are such that the assumption about eastward younging may not be completely valid in detail (see discussion of structure later in this chapter). Despite these uncertainties, however, our preferred model for the evolution of the Rowe-Hawley zone, discussed later in this chapter, still suggests that the strata of the Rowe Schist are probably only slightly, if at all, older than the earliest Cambrian, and almost certainly are no younger than Early or possibly Middle Ordovician.

MORETOWN FORMATION (Oml, Om, Oms, Omsk)

The Moretown Formation is a regionally persistent and distinct group of rocks that are characterized by very light gray, nonrusty-weathering, sandy-textured, "pinstriped" granofels and interlayered schist. In northernmost Massachusetts, it is divided into three members: Oml, Om, and Oms. A fourth member, Omsk, is distinguished in the southern part of the State.

The lowest member, Oml, is recognized only in the northernmost few kilometers of the State. It consists primarily of light-gray to buff, medium- to coarse-grained, irregularly schistose, poorly bedded quartz-plagioclase-muscovite-chlorite-biotite-garnet schist characterized by irregularly shaped and oriented aggregates or clots of chlorite 0.5-1 cm in diameter. Pinstripe granulite is rare. Oml is distinguished from the overlying principal member of the formation, Om, by the chlorite clots, coarser grain size, irregular (nonplanar) schistosity, poorer bedding, and paucity of pinstripe granulite (Hatch and Hartshorn, 1968).

The predominant and most diagnostic member of the Moretown, Om, is primarily very-light-gray, light-gray-green, or light-buff, fine-grained quartz-plagioclase-mica granulite or granular schist. In its two most common varieties this rock either is characterized by pinstripe structure, wherein light-colored granular layers of quartz and plagioclase 1-3 mm thick are separated by paper-thin partings of mica, particularly biotite (fig. 7), or consists of alternating 2- to 5-mm-thick granulose and schistose layers that are characteristically complexly folded (fig. 8). Fine-grained, light-gray garnet schist and fine-grained hornblende-plagioclase amphibolite are also present.

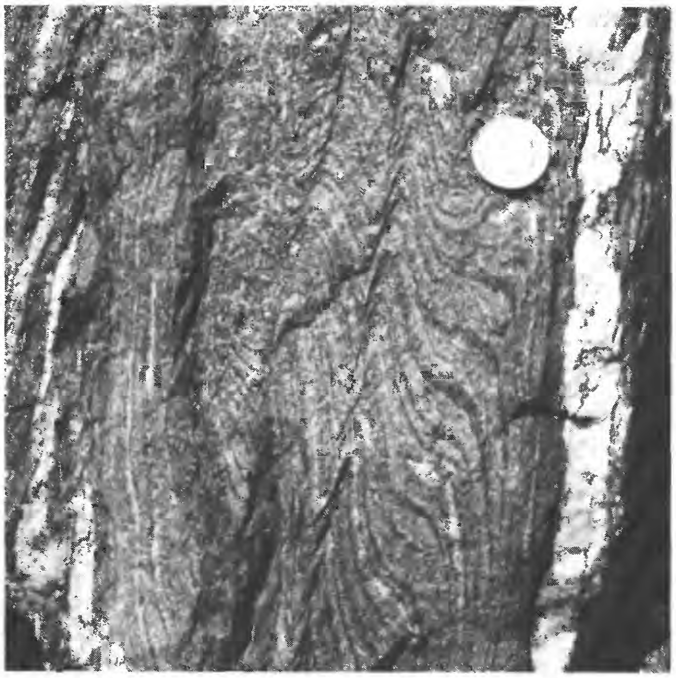
The second most widespread member of the Moretown is Oms, which is mapped separately at the top of the formation from the Vermont border south to a point 2 km south of Chester village. Oms is distinguished by the presence of beds, generally 15 cm to 1 m thick, of pale-brown to light-silvery-gray, fine- to medium-grained muscovite-quartz-plagioclase-biotite-garnet-chlorite schist characterized by round garnets 2-5 mm in diameter around which the schistosity is systematically deformed, leading to the field name of "nubbly garnet schist" (fig. 9). Intercalated with this schist in roughly equal proportions are pinstripe granulite, similar to that in Om, and dark-green to black, fine-grained hornblende-plagioclase amphibolite. Six thin lenses of carbonaceous garnet schist

FIGURE 7.—Well-bedded granulose rocks of the Moretown Formation (unit Om) at West Cummington. View looking north. Conspicuous folds assigned to F_4 fold both bedding and S_2 schistosity.

(Omsc) are mapped within the upper nubbly garnet schist member.

Near Blandford village, the principal Moretown granofels unit (Om) passes southward into more schistose rock (Hatch and Stanley, 1976) distinguished on the State bedrock map as Omsk. This facies (Omsk) is similar to the Ratlum Mountain Member of the Satans Kingdom Formation of Stanley (1964) in northwestern Connecticut (Orm on fig. 10). In southern Massachusetts, Omsk contains only minor pinstripe granulite and instead is characterized by light-gray, fine-grained muscovite-quartz-biotite schist containing conspicuous 2- to 3-mm porphyroblasts of staurolite, 2- to 3-cm porphyroblasts of kyanite, and abundant 1- to 3-mm layers (beds?) of pink, fine-grained quartz-garnet coticule granulite (Hatch and Stanley, 1976). In northwestern Connecticut, this schistose Moretown (Orm, fig. 10) occurs east of a more granulose Moretown facies (Taine Mountain Formation, Otm on fig. 10). This same granulose Moretown occurs in the Bristol dome to the east. Hatch and Stanley (1973, p. 28-29) originally interpreted these three belts to be east-west stratigraphic facies equivalents of each other. However, a tectonic repetition of Otm cannot be ruled out.

Whether the contacts of the Moretown with the Rowe and the Hawley are tectonic or depositional is critical to interpretation of the geologic history of the Rowe-Hawley zone. The following observations are relevant



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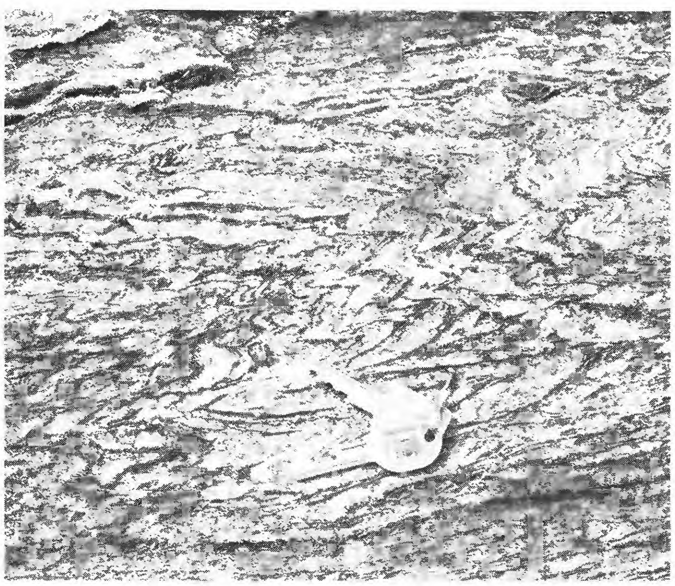


B

FIGURE 8.—Thin contorted granulose (light) and schistose (dark) laminae in unit Om of the Moretown Formation. *A*, Five km west of Plainfield. Coin is 1.75 cm across. *B*, At West Cummington. Keys are about 5 cm long. Minor folds in both photos are assigned to F_3.

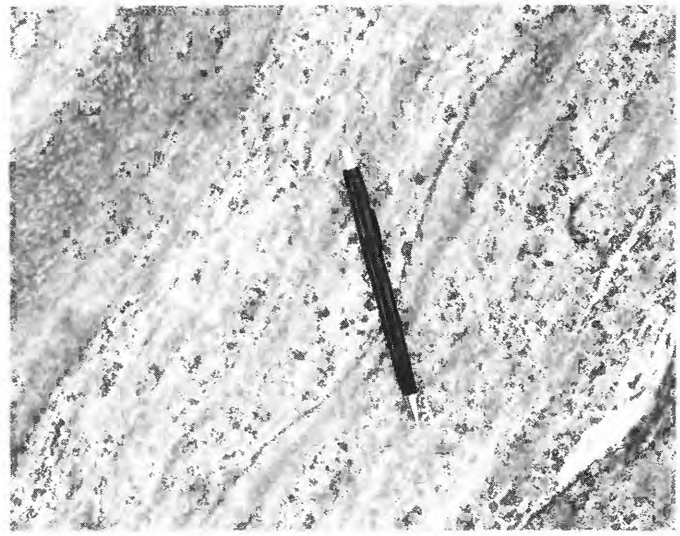


FIGURE 9.—Garnet schist characteristic of unit Oms of the Moretown Formation, 2.5 km northwest of Plainfield. Dark spots are garnets surrounded by rims of quartz-feldspar. Pencil is 14 cm long.

to this question: (1) The western Moretown contact is locally discordant with amphibolite (O€ra) and green schist (O€r) of the Rowe Schist in northernmost Massachusetts (Chidester and others, 1967; Hatch and Hartshorn, 1968). (2) Elsewhere, mappable units in the Moretown are conformable to the Rowe contact, which appears gradational by interlayering over an interval of 10-25 m. (3) Several small bodies of serpentinite are present in the western part of the Moretown near the Rowe contact. (4) Chidester and others (1967), Osberg and others (1971), and Hatch (1969) have mapped a thin amphibolite (Oma) within the Moretown for approximately 40 km along strike near the western edge of the Moretown Formation. (5) To the south, near Blandford, the surface separating schistose Omsk from the main, granulose, part of the Moretown (Om) to the north is shown arbitrarily on the State bedrock map as discordant to O€r of the Rowe. (6) Detailed mapping by Knapp (1977, fig. 4) and Brill (1980) at Prospect Hill (fig. 11) clearly shows an interlayered contact in which discontinuous lenses of staurolite-garnet schist of the Rowe are present throughout unit Omsk of the Moretown.

The available evidence thus does not clearly define whether the western Moretown contact is depositional or tectonic. The map relations in northern Massachusetts suggest the possibility of an unconformity followed by deposition of basal shales, minor quartz-feldspar wackes, and basalt (Oma), which spread across shales and volcanic rocks of the Rowe. The relations at Blandford and at Prospect Hill (fig. 11), however, could be interpreted as tectonic intercalation.

The nature of the eastern contact of the Moretown with the Hawley is uncertain. The State bedrock map and quadrangle maps show Hawley metavolcanic rocks (Oh) and Hawley black schist (Ohb) as discordant with the nubbly garnet schist (Oms) of the Moretown through much of the belt, but this discordance could equally well result from faulting at a very low angle to bedding, sedimentary facies changes in the Hawley, or disconformities. Hatch, Norton, and Clark (1970) described the contact east of Chester village as gradational by interlamination over a few tens of meters of Moretown and Hawley rocks. The map contact was placed east of the easternmost granofels and nubbly garnet schist. Moretown rocks were thus excluded by definition from the Hawley, although some Hawley-like metavolcanic rocks are present locally in the uppermost Moretown and may represent either sedimentary intercalation or premetamorphic tectonic imbrication. Martha M. Godchaux (oral commun., Oct. 1980) reported possible small-scale intercalation of Moretown and Hawley lithologies in the vicinity of the Deerfield River in northern Massachusetts that could be interpreted as either sedimentary or tectonic. Thus the nature of the contact is ambiguous; some data favor a tectonic origin, others support a depositional origin. If this contact is tectonic, the faulting must be premetamorphic, because mapped Acadian isograds are not interrupted by the contact. The relation to Taconian metamorphism, however, cannot be determined due to the severity of the Acadian overprint.

South of Chester, the eastern Moretown contact is tectonic (Hatch and Stanley, 1976; Knapp, 1977, 1978; Hatch and Stanley, Ch. C, this volume). Here the Prospect Hill thrust cuts out sections to both the east and the west, so that member B of the Cobble Mountain Formation is in contact with the Hoosac Formation at the Connecticut State line (fig. 10; Knapp, 1978). The thrust is expressed by mylonitized granites and truncated mesoscopic folds and is clearly synmetamorphic: the regional S_2 Acadian schistosity parallels the mylonitic foliation, which was recrystallized during this time (Knapp, 1977). The Prospect Hill thrust has been extended as far north as the latitude of Blandford village on the basis of discordant units in the eastern and inferred upper plate (Ocar in Oca; Oca-Oh relations, State map).

It is obvious from the above discussion that both the eastern and western contacts of the Moretown must be examined in greater detail than they have been to date if their character is to be determined. The critical factors are basically the mineralogical and textural gradation between the units and the distribution and continuity of small lithic units that probably can only be shown by mapping at a scale of 1:10,000 or larger. In northern Vermont, mapping at this scale has demonstrated that the western contact of the Moretown is tectonic (Stanley and Roy, 1982; Stanley and others, 1984).

The age of the Moretown has traditionally (see, for example, Doll and others, 1961; Hatch and Stanley, 1976; Hatch and Hartshorn, 1968) been considered to be Ordovician. This assignment derived largely from the assumption that the Moretown was conformably below the black slates assigned to the Cram Hill in Vermont that were in turn correlated with the black slates at Magog, Quebec, which contain Middle Ordovician fossils. Despite our present uncertainties about the nature of the Moretown-Hawley contact discussed above and the uncertainties about the southern continuation of the fossiliferous Magog rocks (Doolan and others, 1982), we still believe that an Ordovician age assignment for the Moretown is most reasonable, although perhaps for different reasons than previously used (see discussion of tectonic evolution at the end of this chapter).

HAWLEY FORMATION (Oh, Ohg, Ohp, Ohf, Ohb)

Emerson (1898) applied the name Hawley Schist to the sequence of green chloritic and hornblendic schists and feldspathic schists between his Savoy Schist and Goshen Schist. Hatch (1967), in a redefinition of the Hawley, pointed out that (1) sulfidic black schists are present in mappable intervals throughout the metavolcanic Hawley sequence, (2) these black schists, although grossly similar to the schists of the overlying Goshen, are distinguishable from them, and (3) Emerson (1898) had failed to recognize the distinction between these two types of black schist and had thus mapped them all as Goshen. Thus the redefined Hawley Formation became a unit of locally subdivided metavolcanic rocks, chiefly metabasalt, and mapped units characterized by sulfidic, carbonaceous schist and quartzite (Ohb) containing local quartz-garnet granofels (coticule).

Five unnamed members of the Hawley are distinguished on the State bedrock map, four consisting of metavolcanic rocks and one of black schists containing minor metavolcanic material. The most widespread member is Oh. In the northern part of the State, it consists primarily of green, medium-grained plagioclase-hornblende-epidote-chlorite greenstone and amphibolite and lesser amounts of feldspathic schist and granulite (fig. 12). Relict pillows were seen very locally (fig. 13) (Osberg and others, 1971; Hatch and others, 1967, p. 13, stop 5). Spectacular exposures of plagioclase-chlorite-hornblende-garnet schist, in which hornblende forms bundles of blades, or fascicules, as much as 10 cm long, and garnets are 3-20 mm in diameter, are common, particularly in the top half of the formation (fig. 14). The felsic rocks appear to form a decreasing percentage of the member south of Plainfield.

In the northern part of the State, from the Vermont line south to Heath, a distinctive unit of medium- to

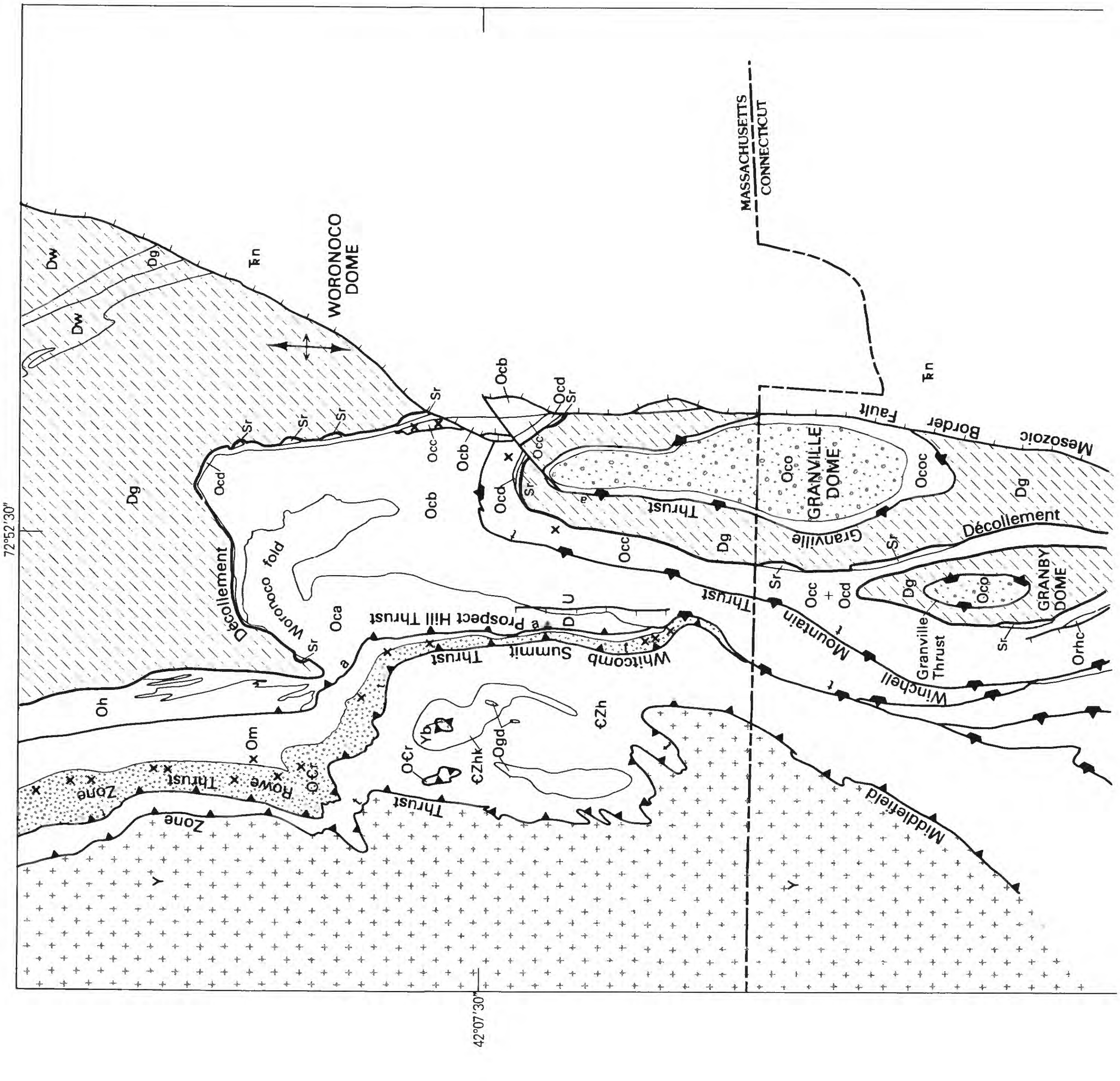

MASSACHUSETTS
CONNECTICUT
WORONOCO DOME
GRANVILLE DOME
GRANBY DOME
Mesozoic Border Fault
Granville Décollement
Granville Thrust
Winchell Mountain Thrust
Prospect Hill Thrust
Whitcomb Summit Thrust
Rowe Thrust Zone
Middlefield Thrust Zone
Woronoco fold
Décollement
72°52'30"
42°07'30"

FIGURE 10.—Geologic map of the Rowe-Hawley zone and infolded western Connecticut Valley belt rocks (diagonal pattern) in southern Massachusetts and adjacent western Connecticut, modified from Zen and others (1983), Stanley (1964, 1968), Hatch and Stanley (1973), Schnabel (1974), and unpublished detailed and reconnaissance mapping by Stanley. Lithic symbols corresponding to those on the Massachusetts State bedrock map are Dw, Waits River Formation; Dg, Goshen Formation; Sr, Russell Mountain Formation; Oca, Ocb, Occ, Ocd, Cobble Mountain Formation; Oh, Hawley Formation; Om, Moretown Formation; O€r, Rowe Schist; €Zh, Hoosac Formation; Oco, Ococ, Collinsville Formation; Ogd, diorite at Goff Ledges; Y, undifferentiated Proterozoic Y rocks. Additional symbols in western Connecticut are as follows: Otm, Taine Mountain Formation; Otmw, Otmsm, Otmwv are the Wildcat Gneiss, Scranton Mountain Gneiss, and Whigville Gneiss Members of the Taine Mountain Formation; Ocob, Ocoh, Ocos are the Bristol, hornblende gneiss, and Sweetheart Mountain Members of the Collinsville Formation (Ocob in the Collinsville dome may contain some Ocoh); Orhc, carbonaceous schist member of the Rattlesnake Hill Formation of Stanley (1964); Orm, Ratlum Mountain Member of the Satans Kingdom Formation of Stanley (1964); Yb, Proterozoic rocks of the Berkshire massif; TRn, Upper Triassic and Lower Jurassic rocks of the New Haven Arkose. Thrusts in Connecticut other than Cameron's line and the Ratlum Brook thrust are interpreted on the basis of relations in southern Massachusetts. Symbols "a", "t", and "m" on faults indicate Acadian, Taconian, or Mesozoic age, respectively. Age of faults is based on the age of their major movement. Although faults west of the Ratlum Brook thrust may have experienced some Acadian movement, the major movement on Cameron's line was Taconian. Small "x" indicates ultramafic rock. The inferred unconformity at the base of Ocoh in the Bristol dome corresponds to the inferred unconformity at the base of Ocos in the Collinsville dome. Ocoh in the north end of the Bristol dome is lithically similar to members A and B of the Cobble Mountain Formation in southern Massachusetts.

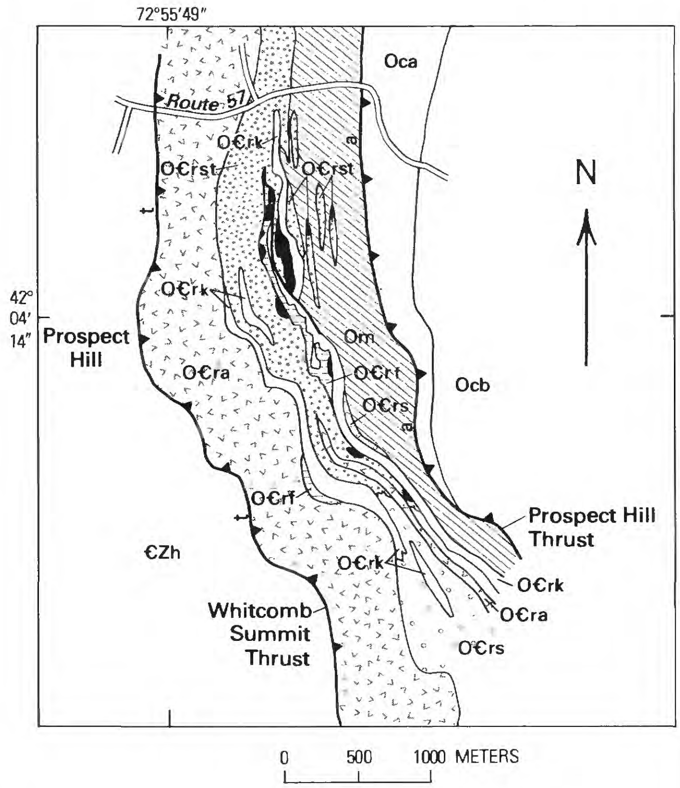


FIGURE 11.—Detailed geologic map of the Prospect Hill area, Massachusetts, in the north-central part of the West Granville 7.5-min quadrangle. Based on work by Knapp (1977, 1978). Lithic symbols corresponding to those on the State map are: ꞒZh, Hoosac Formation; OꞒra, amphibolite of the Rowe Schist; Om, Moretown Formation; Oca and Ocb, members A and B of the Cobble Mountain Formation. Additional units in the Rowe Schist are OꞒrst, staurolite-rich schist; OꞒrk, kyanite-rich schist; OꞒrs, sillimanite-rich schist; OꞒrf, feldspar-rich schist. The symbols "a" and "t" indicate Acadian and Taconian ages, respectively, for the designated thrust faults. Black lenses are serpentinite bodies.

FIGURE 12.—Outcrop of bedded volcanic rocks characteristic of member Oh of the Hawley Formation, near base of formation about 4 km west-northwest of Charlemont. View looking north. Key tag is 7 cm long.

coarse-grained plagioclase-quartz gneiss is distinguished as Ohg. The rock is light gray to greenish gray buff and may contain minor biotite, muscovite, epidote, chlorite, and garnet. Although Ohg resembles some of the felsic gneiss of the Collinsville Formation in the core of the Shelburne Falls dome, with which it was previously mapped (Hatch and Hartshorn, 1968), it is somewhat more schistose, more buff-green in color, and somewhat coarser grained. We now believe that although they may in part be contemporaneous, the Hawley gneiss (Ohg) and the Collinsville gneiss were originally formed in different stratigraphic belts (see discussion of the Bristol thrust at the end of this chapter).

Ohp forms two mapped lenses within the Hawley between Charlemont and Plainfield. The rock is a medium- to dark-gray plagioclase-hornblende-chlorite schist containing conspicuous 2- to 4-mm angular to rounded megacrysts of plagioclase (Osberg and others, 1971). Locally conspicuous angular fragments of light-buff feldspar granulite, light-green epidote-plagioclase granulite, and dark-gray-green amphibolite suggest a pyroclastic origin for the member.

Ohf is also only recognized in the area between Charlemont and Plainfield; it forms a lenticular body immediately west of the larger lens of Ohp. Ohf is primarily a very-light-green to light-buff plagioclase granulite containing minor chlorite and garnet. Interbedded with it are light-gray plagioclase-chlorite-calcite granulite and plagioclase-hornblende-garnet gneiss in which hornblende blades may be as much as 10 cm long and garnets may be as much as 2 cm across (Osberg and others, 1971). We suggest a felsic volcanic origin for these rocks.

The bodies mapped as Ohb are characterized by generally rusty-weathering, dark-gray, fine- to medium-grained, sulfidic quartz-muscovite-biotite schist containing conspicuous carbonaceous material (fig. 15*A*) that weathers to a characteristically splintery rubble (fig. 15*B*). North of Charlemont, Ohb contains as much as 50 percent admixed amphibolite and plagioclase-hornblende-chlorite-epidote gneiss. South of Charlemont, admixed metavolcanic material probably constitutes no more than 20 percent of the member. Very

A

B

FIGURE 13.—Pillow structure in mafic volcanic rocks of member Oh of the Hawley Formation, Chickley River, 4 km north of West Hawley. Hammer handle is 35 cm long.

FIGURE 14.—Long bundles of hornblende crystals forming fascicules, member Oh of the Hawley Formation, Route 2, 4 km west of Charlemont. Pen is 14 cm long.

finely laminated (1-2 mm), extremely fine grained, pink, gray, white, and black quartzites (fig. 16) are common in Ohb, particularly where carbonaceous schists are closely associated with metavolcanic rocks. The pink quartzite contains about 50 percent of tiny manganiferous garnets and is the coticule rock described by Emerson (1898, 1917). The black variety commonly contains magnetite. Although we have made no systematic study of these quartzites, their fine grain size, delicate laminations, and mineralogy all suggest a possible origin as volcanic cherts.

The mutual relations between the black schists and the metabasalt and minor meta-andesite (or metatonalite) of the Hawley are strikingly apparent on the State bedrock map, and, although we do not fully understand their significance, they are worthy of note. The Hawley Formation achieves its maximum outcrop width (and presumed maximum stratigraphic thickness) near the latitude of Charlemont; in this area of maximum thickness the formation is about 99 percent metabasalt and meta-andesite and only about 1 percent black schist. Both north and south from Charlemont not only does the Hawley thin but the ratio of black schist to metavolcanic material increases markedly.

These relations suggest a volcanic center of basalt in the Middle Ordovician section. Similar volcanic sections in the Ordovician are found to the north in southern Vermont (Barnard Volcanic Member of the Missisquoi Formation of Doll and others, 1961) and in southern Quebec (Bolton Lavas of Ambrose, 1942, 1957). The Barnard Volcanic Member consists of approximately equal amounts of metamorphosed quartz basalt (amphibolite) and metamorphosed dacite to rhyodacite (light-colored gneiss) as described by Chang and others (1965) whereas the Bolton Lavas are basalt having little or no felsic counterpart. The significance and plate-tectonic setting of these centers have not been studied on a regional basis, although local studies have been done or are in progress (Martha M. Godchaux, 1980, oral commun.). We suggest that at least two plate-tectonic settings are possible: (1) Volcanism in the forearc basin possibly where local converging boundaries were discordant and caused crustal elongation parallel to the junction, which resulted in volcanic

A

B

FIGURE 15.—Carbonaceous, sulfidic black schist of member Ohb of the Hawley Formation. *A*, The senior author at outcrop on Route 20, 3 km southeast of Chester. *B*, Splintery, acicular-weathering rubble from the same outcrop as *A*. Lens cap is 5.5 cm across. This weathering rubble characterizes the unit and is one means of distinguishing the rocks of Ohb from lithologically similar rocks of the overlying Goshen Formation (Hatch and others, Ch. B, this volume).

FIGURE 16.—Thinly laminated, fine-grained coticule quartzite of member Ohb of the Hawley Formation, 300 m south of Charlemont on road on south side of the Deerfield River. Magnet is 12 cm long.

centers along extensional fractures or (2) island-arc volcanism either associated with the Bronson Hill arc complex or associated with the smaller Ascot-Weedon complex, which disappears beneath the Silurian-Devonian section in northern Vermont. These suggestions are highly speculative and await further work.

Hatch and Stanley (1973, p. 7-8; 1976) have described and mapped in detail the well-documented facies change north of Blandford village between the sulfidic carbonaceous schist of the Hawley and the nonsulfidic noncarbonaceous schist and intercalated granofels of the Cobble Mountain Formation. Despite the number of premetamorphic thrust faults that have been introduced into the geology of the Rowe-Hawley zone in this chapter, we stand by our original interpretation of the Hawley-Cobble Mountain intercalation as sedimentary in origin. The critical evidence for this interpretation is the gradual change in composition between the metasedimentary units and the fact that some of the interlayered belts of the two formations can be mapped for 10-15 km along strike. These are shown on the State bedrock map, just north of Blandford, and on Hatch and Stanley (1976).

The western contact of the Hawley with the Moretown Formation has been discussed in the previous section of this paper. The eastern contact is interpreted to be a major regional unconformity and "surface of Acadian structural disharmony" (Zen and others, 1983). As such, it is discussed elsewhere by Hatch and Stanley (Ch. C, this volume).

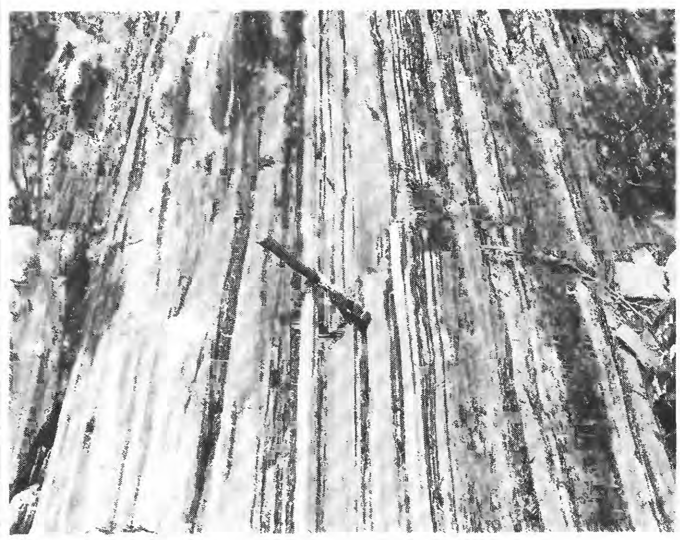
The age of the Hawley has generally been considered to be Middle Ordovician, although it has not been dated either isotopically or paleontologically. Rather, its age assignment derives from the tenuous lithic correlation of the black schists of the Hawley with the fossiliferous Middle Ordovician slates at Magog, Quebec (Berry, 1962), and northwest Maine (Harwood and Berry, 1967) and a lateral correlation with the Ammonoosuc Volcanics in the Bronson Hill anticlinorium to the east, which have been dated in New Hampshire by Naylor (1969) and Brookins (1968) as Ordovician.

COBBLE MOUNTAIN FORMATION (Oca, Ocar, Ocb, Ocbr, Occ, Occr, Occa, Ocd)

The name "Cobble Mountain Formation" was originally assigned to rocks around and near Cobble Mountain Reservoir (fig. 4) a few kilometers south of Blandford in southwestern Massachusetts (Hatch and Stanley, 1973, p. 9-16). The formation extends north almost to the latitude of Chester village, east to Russell Mountain, and south into Connecticut (fig. 10) where it was called the feldspathic schist member of the Rattlesnake Hill by Hatch and Stanley (1973, figs. 2, 3). It has subsequently been recognized in the core of the Goshen dome (Hatch and Warren, 1982) to the north and in the Woronoco dome (Stanley and others, 1982), and lithic equivalents occur in the Granville dome (Knapp, 1977, 1978), as well as in the Collinsville, Bristol, and Waterbury domes in Connecticut (Stanley, 1964; H.E. Simpson, 1974; Gates and Martin, 1967; Hatch and Stanley, 1973) (Sweetheart Mountain Member of the Collinsville Formation of Stanley (1964), for example).

The Cobble Mountain Formation is now divided into four members (Knapp, 1977, 1978; Stanley and others, 1980). These four members are designated A-D from bottom to top. Member A, which interfingers with the Hawley Formation north of Blandford, and member B occupy the largest area on the State bedrock map; members C and D are confined to the west and north sides of the Granville dome. The rocks of the Cobble Mountain Formation are very different from the carbonaceous schist and volcanic rocks of the Hawley; they are noncarbonaceous feldspar-mica-garnet-(± staurolite-kyanite-sillimanite)-rich rocks that are distinctly bedded (some are graded) in much of the section and generally are coarser grained than the Hawley. The bedding characteristics, the abundance of feldspar, and the presence of basaltic amphibolite and minor felsic volcanic rock indicate a turbidite sequence derived in part from a volcanic arc.

Member A consists of thin-bedded granofels and schist (fig. 17) and minor intercalated amphibolite (fig. 18) (Oca) and thin rusty-weathered schist at the base (Ocar). Graded beds at or near the upper contact indicate that B is younger than A. Member B consists of silvery-gray feldspathic schist and gneiss (Ocb) (fig. 19) and subordinate beds of amphibolite and rusty-weathering feldspathic schist and gneiss (Ocbr). Fault-bounded blocks of light-gray, fine- to medium-grained plagioclase-quartz-mica gneiss, directly northeast of the Granville dome, are correlated with member B to the west (Knapp, 1978). Member C contains nonrusty-weathering feldspar-rich schist (Occ), brown, rusty-weathering schist (Occr), and distinctive kyanite and sillimanite schist (Occa) containing magnetite and porphyroblasts of plagioclase. The contact of Occa with



A

B

FIGURE 17.—Well-bedded granulite and schist of member A of the Cobble Mountain Formation, Cobble Mountain Reservoir. Hammer handle is 35 cm long.

FIGURE 18.—Finely laminated amphibolite believed to represent metamorphosed mafic tuff in member A of the Cobble Mountain Formation, Cobble Mountain Reservoir. Coin is 2.5 cm across.

FIGURE 19.—Well-bedded feldspathic schist and gneiss of member B of the Cobble Mountain Formation, Cobble Mountain Reservoir. View looking to the northeast. F_3 folds with clockwise sense of rotation deform layering and F_2 schistosity. Hammer handle is 35 cm long.

adjacent rocks is locally sharp. Occ is lithically identical with the schist of member B. Included in the metasedimentary rocks of member C are lensoid bodies of serpentinite or steatitized ultramafic rocks and thinly laminated, fine-grained amphibolite containing epidote pods and laminae. The aluminous schist (Occa) and the finely laminated amphibolite are remarkably like the bluish-green, quartz-laminated schist and finely laminated amphibolite in the Rowe Schist (Hatch and Stanley, 1973, p. 12; Knapp, 1977, 1978). The serpentinites in member C mark the only occurrence of ultramafic rock above (east of) the lower part of the Moretown Formation in the Rowe-Hawley zone. The lower contact of member C is the Winchell Mountain thrust (Knapp, 1978; Stanley and others, 1982).

Member D (Ocd) is very thin in Massachusetts and occurs between the aforementioned members of the Cobble Mountain Formation and the Silurian and Devonian rocks of the Russell Mountain and Goshen Formations. It consists of dark-brown, fine- to medium-grained, thinly bedded (as much as 20 cm) schist and gneiss. It is locally associated with rusty-weathering, graphitic schist and thin beds (5-20 cm thick) of vitreous quartzite. The basal contact of member D is interpreted to be an unconformity because D rests discordantly on members C, B, and A as it is traced northward through the Woronoco and Blandford quadrangles (Hatch and Stanley, 1976; Stanley and others, 1980, 1982).

The age of the Cobble Mountain Formation is considered to be Middle Ordovician on the basis of the facies relations of member A with the Hawley Formation. This assignment is speculative, however, because of the uncertain age correlation of the Hawley with the fossiliferous Middle Ordovician black slates of the Magog Group along Castle Brook west of Magog, Quebec. The Hawley cannot be traced continuously from western Massachusetts to Magog. Stanley and others (1982) suggested that member D of the Cobble Mountain could be Silurian on the basis of their interpretation of an unconformity at the base of D, although a Middle Ordovician age is equally possible.

The north-south distribution of members and their lithic subdivisions are shown in figure 20 in which Acadian deformation has been removed. Note particularly the facies change between the Hawley and member A of the Cobble Mountain Formation (Hatch and Stanley, 1973).

COLLINSVILLE FORMATION (Ocoa$_1$, Ocog, Ococ, Ocoa, Oco)

The Collinsville Formation was originally defined in the Collinsville quadrangle in western Connecticut (Stanley, 1964); it has been redefined and extended into Massachusetts (Stanley, 1980) where it is exposed in the Granville and Shelburne Falls domes and, to a limited extent, in the Goshen dome (Hatch and Warren, 1982). It consists largely of various light-colored, plagioclase-rich gneisses and interlayered amphibolite and hornblende gneiss (fig. 21). The upper part of the Collinsville is commonly rich in amphibolite, whereas the lower part is predominantly feldspathic gneiss. In places, thin aluminous feldspathic schist, locally

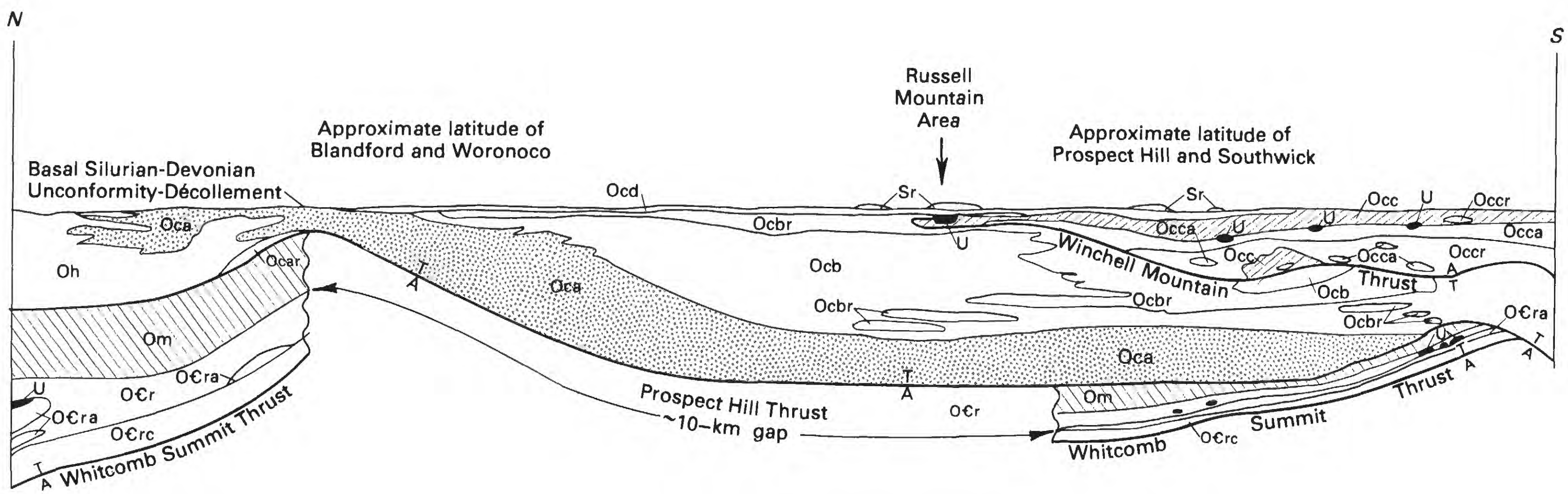


FIGURE 20.—Palinspastic reconstruction showing north-south distribution of the lithic units in the Cobble Mountain Formation. Only Acadian deformation has been removed. The future site and inferred movement sense of the Prospect Hill thrust are shown (A, away; T, toward). Lithic symbols corresponding to those on the State map are O€r, O€rc, O€ra, members of the Rowe Schist; Om, Moretown Formation; Oh, Hawley Formation; Oca, Ocar, Ocb, Ocbr, Occ, Occa, Ocd, members of the Cobble Mountain Formation; u, ultramafic rock; Sr, Russell Mountain Formation. Occu is undifferentiated Occ plus fragments of other lithologies. See also figure 27 for details of the Winchell Mountain thrust in the Russell Mountain area. The 10-km gap (marked by wiggly lines) in the Prospect Hill thrust occurs when Acadian folding of the basal Silurian-Devonian contact is unfolded. See State bedrock map or figure 4 for geographic locations.

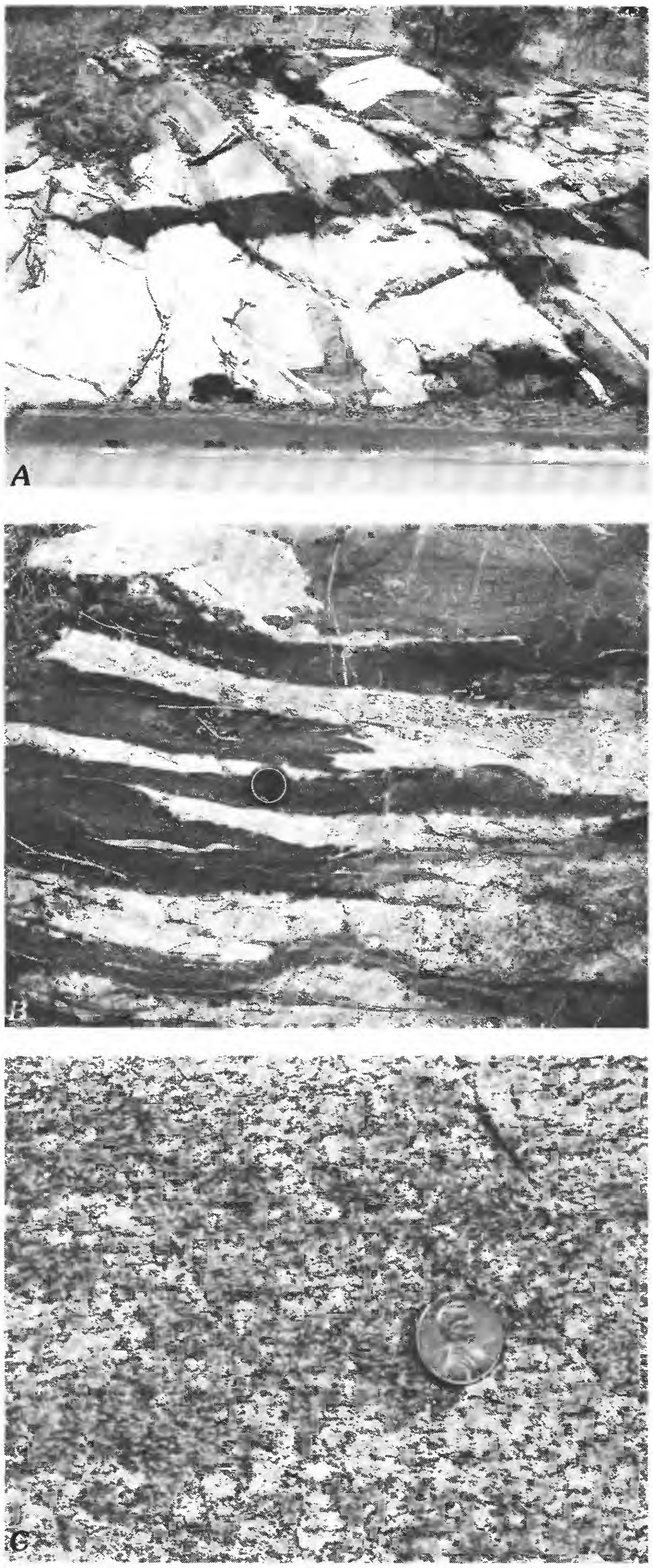


FIGURE 21.—Rocks of the Collinsville Formation in the Shelburne Falls dome (unit Ocog), Route 2 southeast of Shelburne Falls. *A* and *B* show deformed mafic dikes cutting felsic gneiss. Lens cap (*B*) is 5.5 cm across. Coin (*C*) is 1.75 cm across.

containing coticule and amphibolite, forms an upper member of the formation as, for example, Ococ in the Granville dome or the Sweetheart Mountain Member of Stanley (1964) in the Collinsville and Bristol domes in Connecticut (Ocos, fig. 10).

The Collinsville sequence in the Shelburne Falls dome, where Leo M. Hall (written commun., 1977) has recognized seven mappable units, is the most complete in Massachusetts. Unfortunately, however, its contact with the Moretown-Hawley section to the west is not exposed, and their stratigraphic or structural relations can only be inferred. To the south in the Bristol dome in Connecticut this contact is exposed, and we suggest later in this chapter that it is a thrust fault. Although the Collinsville Formation has been complexly folded, the section does not repeat and a large-scale recumbent fold is unlikely. Tectonic zones have not been recognized within the formation. Because this section is critical in establishing possible lithic correlation with the Ammonoosuc Volcanics and the Monson or Fourmile Gneiss in the Bronson Hill zone (Peter Robinson, oral commun., 1982; Leo M. Hall, written commun., 1977), it is described below. The letter symbols at the left are the map symbol designations of the units of the State bedrock map to which Hall's seven units have been assigned.

Stratigraphic section for core rocks of the Shelburne Falls dome (Leo M. Hall, written commun., 1977)

Combined with Ocoa on State map. Gray granulites that are tan-weathering locally and contain thin (2 mm to 3cm) coticule layers. Muscovitic quartzites with garnet and some amphibolite. The upper 15 cm-1 m contains segmented beds or rock fragments or both.

$Ocoa_1$ . . . . . Interbedded amphibolite and white felsic gneiss that contains biotite, hornblende, and commonly magnetite. Local calc-silicate beds contain abundant magnetite as do some of the amphibolites.

Ocof . . . . . . Felsic gneiss with scattered biotite ± magnetite, garnet, and hornblende. Orange-brown weathering is common and in some places occurs in patches. Contains local amphibolite layers typically less than 60 cm thick. Two to five meters of amphibolite is common near the top of the gneiss.

$Ocoa_2$ . . . . . Interlayered amphibolite and felsic gneiss, beds as much as 1.3 m thick but more commonly less than 25 cm thick. Coarse-grained hornblende gneiss with magnetite octahedra as much as 16 mm across is locally present.

Ocog .....Very homogeneous garnetiferous biotite gneiss that is commonly well lineated and poorly foliated. Chlorite is common in coarse clots and around garnet in places. Muscovite is present in places.

Ocor .....Rusty-weathering massive granulites that contain abundant anthophyllite(?) and tourmaline and that commonly have deep pits on weathered surfaces. Felsic gneiss associated with these rusty-sulfidic rocks also has rusty stained surfaces.

$Ocoa_3$.....Amphibolite with thick (1.2 m) and some thin felsic gneiss layers. Some of the amphibolite is strikingly garnetiferous, and garnets 16 mm across are present at one locality.

We have combined Hall's upper gray granulite unit with the highest unit of interlayered amphibolite and white felsic gneiss as $Ocoa_1$ because the gray granulite unit is too thin to show at the scale of the State bedrock map. It occurs along the northern part of the Shelburne Falls dome and is cut out by the base of the Goshen Formation. As shown on the explanation to the State bedrock map, the three units above the homogeneous gneiss (Ocog) are considered to be approximately equivalent to the Ammonoosuc Volcanics of the Bronson Hill anticlinorium. The homogeneous gneiss is lithically similar to the Monson and Fourmile Gneisses.

The exposed Collinsville sequence in the Granville dome to the south is not as complete, in part because of thick surficial cover in the center of the dome. Knapp (1977) divided the pre-Goshen rocks of the Granville dome into three units, an outer (upper) rusty-weathering quartz-plagioclase-muscovite-biotite schist and thin coticule (Ococ), a middle amphibolite-rich unit containing minor plagioclase gneiss and no rusty-weathering schist (Ocoa), and a lower plagioclase gneiss and minor amphibolite (Oco). Along the eastern side of the Granville dome, the lower part of the upper unit (Ococ) also contains abundant amphibolite intercalated with the rusty-weathering schist. Ococ is similar to the aluminous kyanite-sillimanite schist in the Cobble Mountain Formation (Occa) to the west and the Sweetheart Mountain Member (of Stanley, 1964) of the Collinsville Formation to the south in Connecticut. Although both Ococ and Occa are schistose and contain lenses of amphibolite, Ococ differs from Occa in that Ococ contains coticule, is characteristically rusty weathering, and lacks abundant kyanite and sillimanite. Ococ and its correlatives are absent in the Shelburne Falls dome and in the domes along the Bronson Hill zone. Of particular importance for our plate-tectonic interpretation, discussed in the final part of this chapter, is the presence of small pods of serpentinite in both the Sweetheart Mountain Member inside the domes and in member C of the Cobble Mountain Formation west of the domes.

The stratigraphic and (or) structural relations among the Collinsville Formation, members A, B, C, and D of the Cobble Mountain Formation, the Hawley Formation, and the Partridge Formation are shown schematically in figure 22. Member B is present in the Goshen dome, where it is mapped as overlying the gneiss and amphibolite of the Collinsville Formation (Hatch and Warren, 1982). Similar rocks are found in the northern part of the Bristol dome, Connecticut, on Nepaug Reservoir where Stanley (1964, pl. 1) mapped them as part of the Bristol Member of the Collinsville Formation because they could not be separated from the amphibolite and gneiss shown as Ocoh in figure 10. We correlate this unit with the Ammonoosuc Volcanics along the Bronson Hill anticlinorium. These relations suggest that Ocb interfingers with and, in part, overlies the Collinsville Formation (excluding Ocos). The silvery-gray feldspathic schist of Ocos is very similar to member C (Occ in fig. 10) in that both bear kyanite and contain pods of amphibolite and serpentinite. Ocos is found in all the domes of western Connecticut where it overlies the gneiss and amphibolite of the Bristol Member, thus supporting the relations shown in figure 22. Hawley-type black schists and volcanic rocks, which interfinger with Oca of the Cobble Mountain Formation north of Blandford, are not present in any of the domes west of the Mesozoic basins. Black schists, mapped as Partridge, are present, however, east of the domes, in the Whately anticline, which is closer to the Bronson Hill where the Partridge is extensive and rests on the Ammonoosuc or lower gneisses.

Although cross section *F-F′* on the State bedrock map shows Ococ and Oca as facies equivalents to the Partridge, other correlations are certainly possible because fossil control is lacking. The Partridge is lithically unlike Ococ and its equivalents and could be younger. An older age is unlikely because Partridge-like rocks are not found beneath Ococ or its equivalents and the gneisses and amphibolites of the Collinsville Formation in the domes. The Partridge is exposed in the Whately anticline but is absent from the Shelburne Falls, Goshen, and Granville domes in western Massachusetts as well as from the domes in western Connecticut. Its absence may be due to erosion before deposition of the Silurian-Devonian section and thus does not bear on whether the Partridge is equivalent to or younger than Ococ and its equivalents. The amphibolite-rich unit (Oca) in the Granville dome could also be equivalent to the Ammonoosuc Volcanics, although it too is shown as equivalent to the Partridge in section *F-F′*. The light-colored gneisses and amphibolite of the Collinsville, however, have many similarities to the Monson (Fourmile)-Ammonoosuc section of the Bronson Hill zone. Major unsolved problems are the position of the Ammonoosuc-Monson contact in the Collinsville Formation in each of the domes west of the Mesozoic basins in Massachusetts and western Connecticut and the age relationships among the various Middle Ordovician rocks of the Hawley-Cobble Mountain-Partridge cover.

Our correlation of the Ordovician and Silurian rocks across the Mesozoic basins, and our best guess as to the lithic and age relations among the different Middle Ordovician units and the core rocks of the Bronson Hill, as represented by the Collinsville Formation, are shown in figure 22. The localities where the critical "on-the-ground" evidence for this scheme was collected are marked by dashed lines and labelled at the top of the diagram. The unconformity at the base of the Cobble Mountain Formation that cuts the Bristol thrust (fig. 22) is speculative and is based on our interpretation of the Cobble Mountain Formation as having been shed off the Bronson Hill arc into the forearc basin. The erosional surfaces at the base of the Silurian-Devonian section and of member D of the Cobble Mountain Formation are shown by distinctive lines. In this scheme, the black schists and volcanic rocks of the Hawley are shown transgressing eastward over the Cobble Mountain Formation to the Bronson Hill domes where they become the Partridge Formation. Thus the base of the Hawley to the west would be older than the Partridge to the east.

Although we show the Collinsville Formation in the domes west of the Mesozoic basin as being continuous with the Monson-Fourmile core sequence of the Bronson Hill, future workers must consider the option that the Collinsville and equivalent rocks in the western domes could be part of an arc situated west of the Bronson Hill arc complex. This western arc could possibly be the southern continuation of the Ascot-Weedon arc in southern Quebec and northern Vermont (Doolan and others, 1982).

ULTRAMAFIC ROCKS

Fifty-three discrete bodies of ultramafic rock have been mapped in a relatively narrow belt within the Rowe-Hawley zone across Massachusetts. Of these, 42 are within the Rowe Schist and the basal part of the Moretown Formation, and 11 are in the Cobble Mountain Formation south of the Massachusetts Turnpike. (These 11 bodies are discussed at greater length in the section entitled "Structure of the Cobble Mountain Formation.") Two additional bodies are known from the Berkshire massif to the west. The Rowe-Hawley zone ultramafic rocks are very important in the tectonic reconstruction of the Rowe Schist because they are interpreted to represent fragments of ocean crust. Before the development of plate-tectonic theory, field workers had considered them to be intrusive bodies emplaced in a cold, semi-solid condition (Chidester, 1968, 1978). We take the plate-tectonic view that they were tectonically emplaced in light of evidence presented below and recent work by Stanley and Roy (1982; Stanley and others, 1984) in northern Vermont, which is summarized in part in the discussion of the Rowe thrust zone.

The Massachusetts ultramafic belt is part of a very extensive but narrow zone of ultramafic pods and lenses that extends north from Massachusetts through Vermont to Quebec and southwest across Connecticut. Although the Vermont ultramafic rocks are largely confined to formations that we interpret to be correlative with the Rowe and Moretown of Massachusetts, some are present in the Hazens Notch Formation of late Precambrian to Cambrian age west of the Ottauquechee Formation, a few are in the Cram Hill Formation of Middle Ordovician age east of the Moretown, one body is mapped in the Hoosac of Late Precambrian-Early Cambrian age, and several are reported in the Precambrian rocks of the Chester and the Athens domes (Doll and others, 1961). The ultramafic belt continues across the international border into Quebec where it has been traced out along the Gaspé Peninsula and thence to Newfoundland. Williams and St-Julien (1982) have applied the term "Baie Verte-Brompton line" to the zone of ultramafic bodies in Canada. South from Massachusetts, a narrow zone of similar ultramafic lenses has been traced across northwestern Connecticut, east of Cameron's line; they are in a belt of rocks that we correlate in part with the Rowe.

The ultramafic bodies in Massachusetts range in size from 0.5 m wide and 1 m long to 720 m wide and 4,000 m long (the large body near Chester). The great majority are 25-50 m wide and 100-300 m long. Exposures are rarely such that the exact size and shape of a body can be determined, but in the rare cases where mining or unusual natural exposures make this possible, the bodies are seen to be lensoid in plan and generally conformable with the dominant (S_2) Acadian foliation of the enclosing rocks. Little is known of the vertical or downdip extent of the bodies.

The mineralogy of the bodies is remarkably simple. Most are composed almost entirely of talc, serpentine, and minor amounts of magnesite, tremolite, magnetite, and chromite. The large body at Chester contains olivine, particularly in its central part, although even near the center of the body the olivine is extensively rimmed and veined by serpentine. Many of the smaller bodies either are steatitized throughout or have an inner core of serpentine and a rim of talc-carbonate rock. Killius (1974) gave a detailed description of two

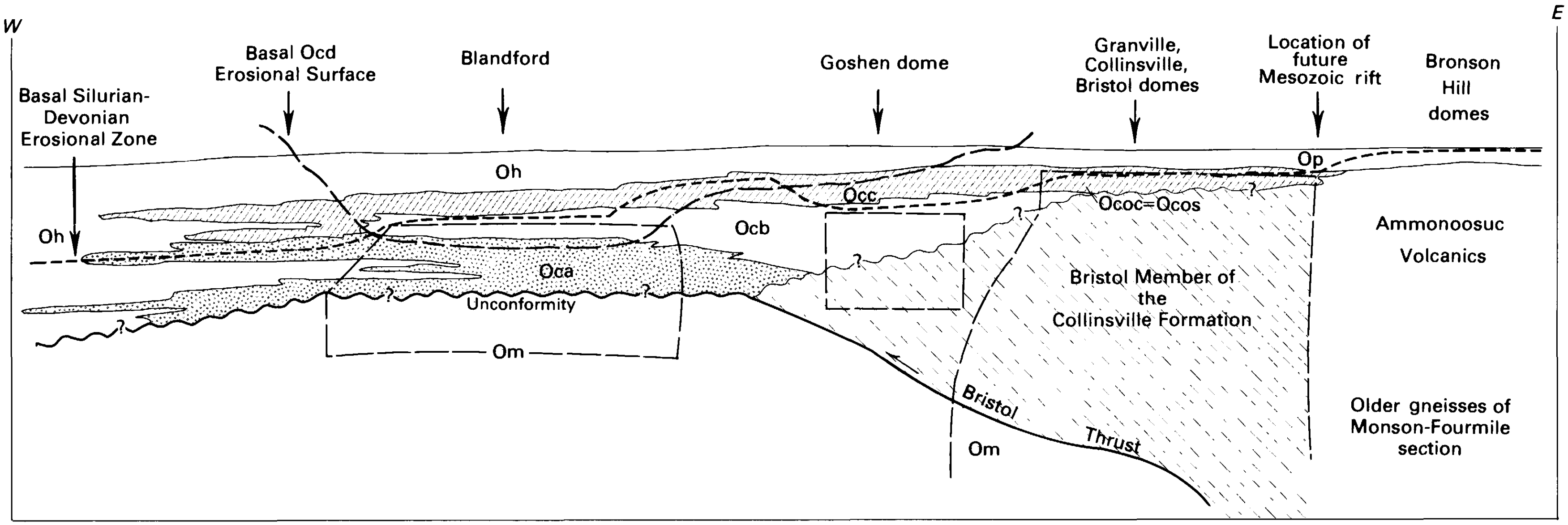


FIGURE 22.—Inferred east-west stratigraphic and structural relations among the Hawley (Oh), Cobble Mountain (Oca, Ocb, Occ), and Partridge (Op) Formations, the rocks of the Collinsville Formation (Ococ, Ocos), the Moretown Formation (Om), and generally correlative rocks in the Bronson Hill domes. Section is composite and is *approximately* 40 km wide. The Hawley-Cobble Mountain-Partridge Formations unconformably overlie the Bristol thrust, which places the Bristol Member of the Collinsville Formation over the Moretown Formation. The dashed lines outline areas in which the geologic relations are presently accessible to mapping. Relations outside dashed areas are thus much more inferential. Time represented by the youngest deposits on the diagram is late Middle Ordovician or somewhat later, before deposition of Ocd. See figure 10 for further description of lithic symbols.

small bodies on the north side of the northwest arm of Cobble Mountain Reservoir. Sanford (1982, p. 552, 556) described the mineralogy of the larger of these two bodies in more detail and discussed its origin petrologically.

Two small lenticular bodies of calcite-actinolite marble are present in member C of the Cobble Mountain Formation on and 1 km south of Russell Mountain (fig. 4) (Stanley and others, 1982). Both are within 100 m of, but are not in contact with, lenses of ultramafic rock. Although their origin is still uncertain, we suggest that they may have formed as a metasomatic reaction zone above the outer contact of the ultramafic rocks and then were isolated by later faulting.

None of the 53 ultramafic bodies in the Rowe-Hawley zone on the State map shows any evidence of having thermally altered the rocks that currently surround them. None shows any obvious signs of cataclasis along its contacts. Cryptic rhombohedral or phacoidal fragments flattened in the dominant Acadian schistosity (S_2) are separated by finer grained serpentinite. Acadian metamorphism and deformation in the Rowe-Hawley zone may have obscured the cataclastic fabric present in similar bodies in northern Vermont and southern Quebec that were not as affected by the Acadian events.

The origin of the ultramafic bodies, both in Massachusetts and in neighboring states, has been debated for many years. Emerson (1898) early noted the close association of many of the ultramafic bodies, particularly in the vicinity of Chester, with the Chester and other amphibolites. He suggested the possibility that they were "serpentinized amphibolite." Later Emerson suggested (1917, p. 156) that they may have been intrusive peridotites or norites (which presumably were subsequently serpentinized). Chidester (1968), in a summary discussion of the ultramafic rocks of Vermont and northern Massachusetts, suggested that the rocks had apparently been forcefully injected, perhaps as a crystal mush and perhaps already partly or highly serpentinized, sometime during the late Middle or Late Ordovician. This model was used by us and our colleagues in most of the quadrangle maps covering this zone in Massachusetts, as well as by the compilers of the Vermont State Map (Doll and others, 1961). The map pattern of the ultramafic bodies and the surrounding host rocks, recent detailed structural and petrographic data on and around individual bodies, and the advent of plate-tectonic theory have led us to modify considerably our interpretation of the origin of those bodies from that of earlier workers, including ourselves. We do, however, recognize that the small ultramafic bodies in the the Proterozoic Y rocks of the Berkshire massif (Ratcliffe, *in* Zen and others, 1983) and the Chester and Athens domes (Doll and others, 1961) may be intrusive in origin and, as such, distinct from those in the Rowe-Hawley zone. The following paragraphs summarize our observations and thoughts that influenced our interpretation of the Rowe-Moretown bodies in Massachusetts.

The 42 ultramafic bodies outside of the Cobble Mountain Formation in Massachusetts are not only restricted to the Rowe Schist and the Moretown Formation but are further restricted to the upper part of the Rowe (36 bodies) and the lower part of the Moretown (6 bodies). Of the 42, 18 are bounded by two or more host rock types, and 24 are mapped as being entirely within one host rock type: 18 are entirely within O€r, 5 are entirely within Om, and 1 is entirely within O€ra. None were found in O€rc. Of the 42 bodies, 16 are in contact with amphibolite of the Rowe or Moretown on at least one side (for example, 9 are on the eastern side), whereas 26 have no contact with amphibolite. Although all 42 bodies appear generally conformable with the schistosity and bedding in the host rocks, 4 are mapped as cutting across contacts between members of the host formations and thus cannot be considered to be truly conformable.

Evidence pertinent to the origin of the ultramafic bodies in the Rowe-basal Moretown section of Massachusetts can also be found in this same belt of rocks in northern Vermont where the effects of Acadian metamorphism and deformation are far less intense and consequently the older Taconian fabric can be seen. In the Troy-Jay area, near Vermont's northern border, Stanley and Roy (1982; Stanley and others, 1984) have shown that all the ultramafic bodies are located on faults or in fault zones. The faults are defined by mylonitic fabrics and by truncated depositional contacts in the footwall and hanging-wall blocks. Contact metamorphic aureoles are absent in the surrounding country rocks, and the serpentinites contain abundant interlacing slickensided slip surfaces, observed earlier by Cady and others (1963) and Chidester (1978). The relations are even more obvious to the north in Quebec, where the metamorphic grade is chlorite or lower.

The Belvidere Mountain Amphibolite is present as discontinuous fault slivers along the Hazens Notch-Ottauquechee boundary; it can be traced south to Belvidere Mountain, where Laird and Albee (1981a,b) reported medium-high-pressure sodium-rich amphibole. Gale (1980, 1986) has shown that the ultramafic complex and garnet amphibolite in this area form a series of southeastward-inclined thrust slices of adjacent metasedimentary rocks marked by mylonitic and cataclastic fault-zone features, suggesting repeated movement along this zone. Vestiges of epidote-amphibolite facies (garnet-barroisitic hornblende) metamorphism are preserved in the coarse- and fine-grained amphibolite where they have been only partially altered to greenschist-facies assemblages (Gale, 1980, 1986). Directly to the north at Tillotson Peak, a high-pressure low-temperature glaucophane-omphacite-garnet-phengite assemblage was reported by Laird and Albee (1981a) from the Belvidere Mountain Amphibolite. The whole complex is separated from the Hazens Notch Formation by a folded thrust. These field and petrographic data support a tectonic origin for the Belvidere Mountain Amphibolite rather than the intrusive origin suggested by Chidester (1978). We believe, therefore, that the ultramafic bodies in northern Vermont and Quebec are fault slivers of rocks that originally were intrusive-extrusive complexes, probably formed at an ancient ridge or transform fault in Iapetus. The Ordovician age commonly assigned to them on quadrangle maps is, therefore, only correct in the sense that they were emplaced into their present host rocks during the Ordovician Taconian orogeny. Their original age of formation depends on the age of the Iapetus. We further believe from these relations in northern Vermont that the Rowe-lower Moretown ultramafic bodies in Massachusetts were emplaced by similar tectonic processes.

INTRUSIVE ROCKS

Compared with many areas of middle- and high-grade metamorphism in New England, the Rowe-Hawley zone is remarkably lacking in intrusive igneous rocks, and most of those present appear to be Acadian or possibly younger. Only those intrusive rocks that are believed to be Ordovician or older are discussed here.

GNEISS AT HALLOCKVILLE POND

A 2- by 3-km body of granodiorite gneiss (fig. 23) was mapped by Osberg and others (1971) in the vicinity of Hallockville Pond, a few kilometers northwest of the village of Plainfield (fig. 4). They described it as "white to very light gray microcline-plagioclase-quartz-biotite gneiss. Microcline forms 1- to 2-cm insets." It is strongly foliated parallel to the regional (F_2) schistosity. Although they assigned this body to the Devonian(?), they pointed out that the only real constraints on its age were that it intruded the Moretown (of probable Ordovician age) and is cut by mafic dikes of questionable Middle Devonian age. No isotopic age has been determined on the body, and the rock has no obvious similarity to intrusive rocks elsewhere in central or southern New England. Although the diapiric rise of the Hallockville Pond body into its present position

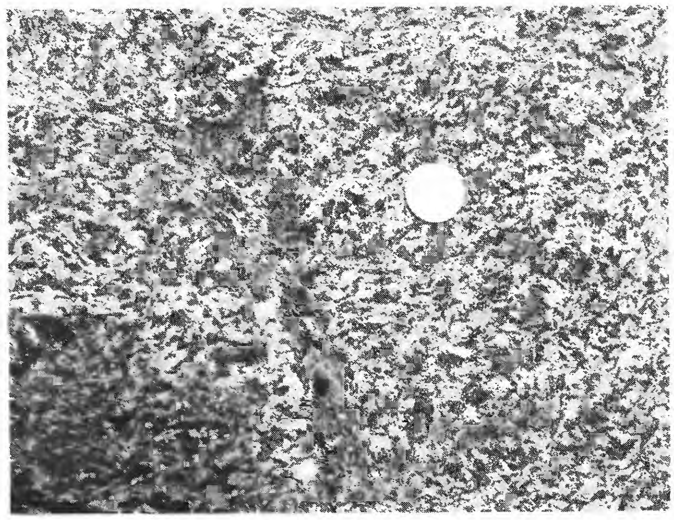


FIGURE 23.—Deformed F_2 foliation in gneiss at Hallockville Pond, 4 km northwest of Plainfield. Coin is 2.5 cm across.

must postdate the Acadian foliations which it has deformed into a cleavage arch, the original intrusion must have been considerably earlier because the gneiss itself contains the earlier Acadian foliations (Osberg and others, 1971). The Ordovician age assigned to the Hallockville on the State bedrock map is thus very tentative and based entirely on the fact that it is older (pre- rather than post-F_2 schistosity) than and texturally different from Devonian (Acadian) granitic rocks in the Rowe-Hawley zone and environs.

FOLIATED GRANITES

In the vicinity of Borden Brook Reservoir, about 6 km south of the village of Blandford, small bodies of foliated biotite-muscovite granite (Ogr) intrude the Hoosac Formation. Good exposures may be seen in the spillway of the reservoir. The rocks are well foliated parallel to the early Acadian (S_2) and (or) possibly Taconian foliation, which they therefore must predate. Furthermore, they closely resemble the granitic rocks described by Ratcliffe and Hatch (1979, p. 187, 205, 209) as commonly intruded along the Middlefield thrust zone and other faults of probable Ordovician age. The age of these rocks is uncertain, but on the basis of arguments relating the igneous rocks to metamorphic fabrics of known Taconian age they are interpreted by Ratcliffe and Hatch (1979) as Taconian. Although we have no age data on the Borden Brook rocks, we tentatively correlate them with the Middlefield fault zone granites exposed a few miles to the west. Finally, if this correlation is correct, and if the direct relationship between faulting and granite emplacement suggested by Ratcliffe is accepted, Taconian thrust faults may be more widespread than we recognized earlier

(Hatch and Stanley, 1976) in the vicinity of the Borden Brook Reservoir. However, we cannot rule out an early Acadian age for the Borden Brook Reservoir intrusive rocks.

A blastomylonitic quartz-microcline-muscovite-plagioclase-garnet gneiss with large (6-cm) megacrysts of microcline is interlayered with nonrusty-weathering, light-gray gneiss and schist of member B of the Cobble Mountain Formation (Knapp, 1977, 1978) in several horizons all within 100 m of the Prospect Hill thrust (Hatch and Stanley, Ch. C, this volume). The gneiss is a fine- to medium-textured rock with a strong anastomosing mica-rich foliation, which wraps around large microcline grains that form a distinctive spotted pattern on the weathered surface. Quartz forms long (as much as 5 cm) but very thin (less than 2 mm) stringers that parallel the strong foliation (Acadian, S_2, regional schistosity). Knapp (1977, p. 46-47) demonstrated conclusively that the blastomylonitic gneiss was intruded as granite before the regional S_2 Acadian schistosity. He further suggested that it may represent a late Taconian intrusion that was later caught up in the Acadian Prospect Hill thrust. We consider this rock to be generally similar to and approximately the same age as those described by Ratcliffe and Mose (1978) from the Middlefield thrust zone, although the composition is not strictly the same.

DIORITE

Norton (1967) mapped a small body of metamorphosed dioritic rock (Od on the State bedrock map and fig. 4) in the northeast corner of the Windsor quadrangle. This rock forms a body about 250 m across that cuts members O€r and O€ra of the Rowe Schist about 14 km south of the Vermont State line. Norton (1967, p. 41-42, 47, pl. 1) described the rock as a gray-green-weathering hornblende-quartz-clinozoisite-chlorite gneiss that is weakly foliated and has a grain size of 1-2 mm. Only this one body of the rock was recognized. It is tentatively assigned an Ordovician age because it predates the Acadian deformation and metamorphism and intrudes only the Rowe Schist. Although it is the only recognized body of mafic rock intrusive into the Rowe-Hawley zone in Massachusetts, larger bodies of mafic rock are common in the zone in southwestern Connecticut (for example, the Brookfield Gneiss (Stanley and Caldwell, 1976)).

PRE-SILURIAN DEFORMATION OF THE ROWE-HAWLEY ZONE

We discuss in this section the principal structures that were imposed on rocks of the Rowe-Hawley zone during the Taconian deformation. The discussion covers fold events, the Whitcomb Summit thrust that bounds the zone to the west along the Berkshire massif, the Rowe thrust zone and its ultramafic lenses, the Bristol thrust that bounds the zone to the east, and the complex structural history of the rocks of the Cobble Mountain Formation.

In addition to the Taconian deformation described here, all the stratified rocks of the Rowe-Hawley zone were also deformed during the subsequent Devonian Acadian orogeny. This Acadian deformation, which is described elsewhere in this volume (Hatch and Stanley, Ch. C), included at least three distinct, regionally extensive generations of folds and several episodes of thrust faults. The first of the three Acadian fold episodes, F_2, produced the strong regional schistosity in at least the eastern part of the Rowe-Hawley zone and overprinted an existing subparallel Taconian schistosity in at least the western part of the zone. The two later Acadian fold episodes, F_3 and F_4, refolded the earlier schistosities and locally developed crosscutting slip cleavage (Hatch and Stanley, Ch. C, this volume). East-over-west thrust faults, which can be chronologically related to the fold episodes, further complicate the structure of these rocks. Thus, an understanding of the Taconian deformation in the Rowe-Hawley zone can only be gained by palinspastically removing the recognized Acadian structures and by extrapolating documented Taconian structures eastward from the Taconic-Berkshire zone (Ratcliffe and Hatch, 1979).

FOLDS

Although the most readily recognized and apparently dominant structures of the Taconian orogeny in the Rowe-Hawley zone are the thrust faults discussed below, at least some Taconian folding is suggested by the field evidence. Hatch and others (1967), Osberg and others (1971), Stanley (1975), and Hatch (1975) all advocated an episode of pre-Acadian folding in the pre-Silurian rocks of the zone on the basis of differences in the attitudes of earliest Acadian fold axes in pre- and post-Taconian strata and of observation of a very few minor folds in the field. Most of the observed folds, designated F_1 by all the reports cited above, are in such thin-bedded parts of the section as member A of the Cobble Mountain Formation or mafic tuffs of the Rowe Schist (fig. 24). Here the pervasive S_2 schistosity of early Acadian age cuts across the axial surface of F_1. We have also suggested that the diversity in the amount and direction of plunge of F_2 folds in pre-Silurian strata, in contrast to the uniformly sub-horizontal plunge of F_2 axes in Devonian strata, results from the fact that at the onset of F_2 the pre-

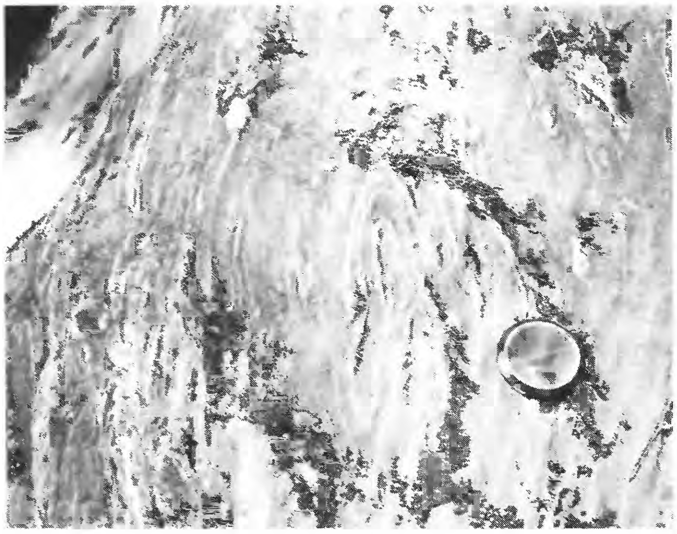


FIGURE 24.—Isoclinal fold interpreted as F_1 (Taconian) to left of lens cap, refolded by a second generation of isoclines interpreted to be F_2 (first Acadian fold stage; see Hatch and Stanley, Ch. C, this volume) in amphibolite of member O€ra of the Rowe Schist. Dominant schistosity is parallel to axial surface of the F_2 fold and cuts axial surface of F_1. Outcrop is 1.5 km west of Chester. Lens cap is 5.5 cm across.

Silurian strata were already folded whereas the Silurian and Devonian strata were not folded. If this interpretation is correct, F_1, Taconian, folding was widespread throughout the Rowe-Hawley zone. We attribute the paucity of observed F_1 folds to the facts that thrust faults dominate the fabric of the Cambrian-Ordovician section and that Acadian deformation is severe and abundant.

WHITCOMB SUMMIT THRUST

Compilation and local remapping along the Rowe-Hoosac contact in parts of the Rowe (Chidester and others, 1967), North Adams (Herz, 1961; Ratcliffe, 1979b), Windsor (Norton, 1967), and Peru (Norton, 1974a) quadrangles showed that mappable units in both the Hoosac and the Rowe were truncated along the contact. Therefore, an older, prethrust structure was truncated by this contact and was severely flattened and smeared out during subsequent deformation (fig. 25; also see Ratcliffe, 1979b, fig. 5). This surface was named the Whitcomb Summit thrust by Stanley (1978). Although truncated faint beds are visible in many waterwashed outcrops of the blue-green schist (O€r) of the Rowe, Acadian metamorphism and deformation have severely smeared out mylonitic fabrics along the Whitcomb Summit thrust zone so that they are easily overlooked. Very well developed mineral lineation and quartz rodding are oriented directly down the dip of the dominant schistosity and are similar to lineation described by Ratcliffe (1979b) as being associated with the Hoosac Summit (Middlefield) thrust on Hoosac Mountain and with the soles of imbricate fault zones in the Berkshire massif. Lenses of Rowe Schist (O€r) are present in the eastern part of the allochthonous Hoosac (€Zhga, fig. 25) and large lenses of Hoosac have been mapped locally in the Rowe 6-8 km west-southwest of Rowe village (Chidester and others, 1967). A continuous exposure through one of these Hoosac lenses shows that the different lithic types in the Hoosac are not symmetrically repeated across the lens, suggesting that the lenses are fault-bound slivers rather than isoclinal folds. Therefore we interpret these lenses in the Rowe and Hoosac to indicate that the Whitcomb Summit thrust is a zone of distributed thrust slices of varying width and length.

Further evidence for the Whitcomb Summit thrust is based on the drastic changes in thickness of the Rowe Schist from about 100 m to 2,000 m to 100 m as it is traced northward across western Massachusetts and on the discordant facies line separating the volcanic from the nonvolcanic parts of the allochthonous Hoosac (Ratcliffe and Hatch, 1979). As we described earlier in this chapter, the ultramafic bodies within the Rowe and basal Moretown are now believed to be fault-bounded fragments of oceanic crust rather than intrusions. Their restriction to the Rowe Schist and basal Moretown and their absence from the allochthonous Hoosac as well as from the late Precambrian to Cambrian sedimentary cover on the Grenville crust not only support the existence of the Whitcomb Summit thrust but indicate that considerable displacement has occurred across this boundary. This displacement postdated the faulting that interleaved the ultramafic bodies with the Rowe and Moretown metasedimentary rocks.

The Whitcomb Summit thrust continues southward into western Connecticut, where it joins Cameron's line (fig. 10). In northern Connecticut, the Moretown Formation equivalent, the Taine Mountain Formation (Otm, fig. 10), is in contact with the allochthonous Hoosac; the intervening Rowe is missing although it is present in an Acadian antiform directly to the east (Hatch and Stanley, 1973, pl. 1; Stanley, 1968, fig. 3). Moretown- and Hawley-equivalent rocks also are present directly east of Cameron's line farther to the south (Rodgers, 1985), further attesting to the magnitude of section lost along this surface. Cameron's line can be traced southward into New York (Hall, 1976, fig. 2; Hall, 1980, fig. 2), where it juxtaposes schists and metavolcanic rocks of the Hartland Formation (Rowe-Hawley zone of Massachusetts) against gneiss, marble, and schist of the Fordham-Lowerre-Inwood-Manhattan

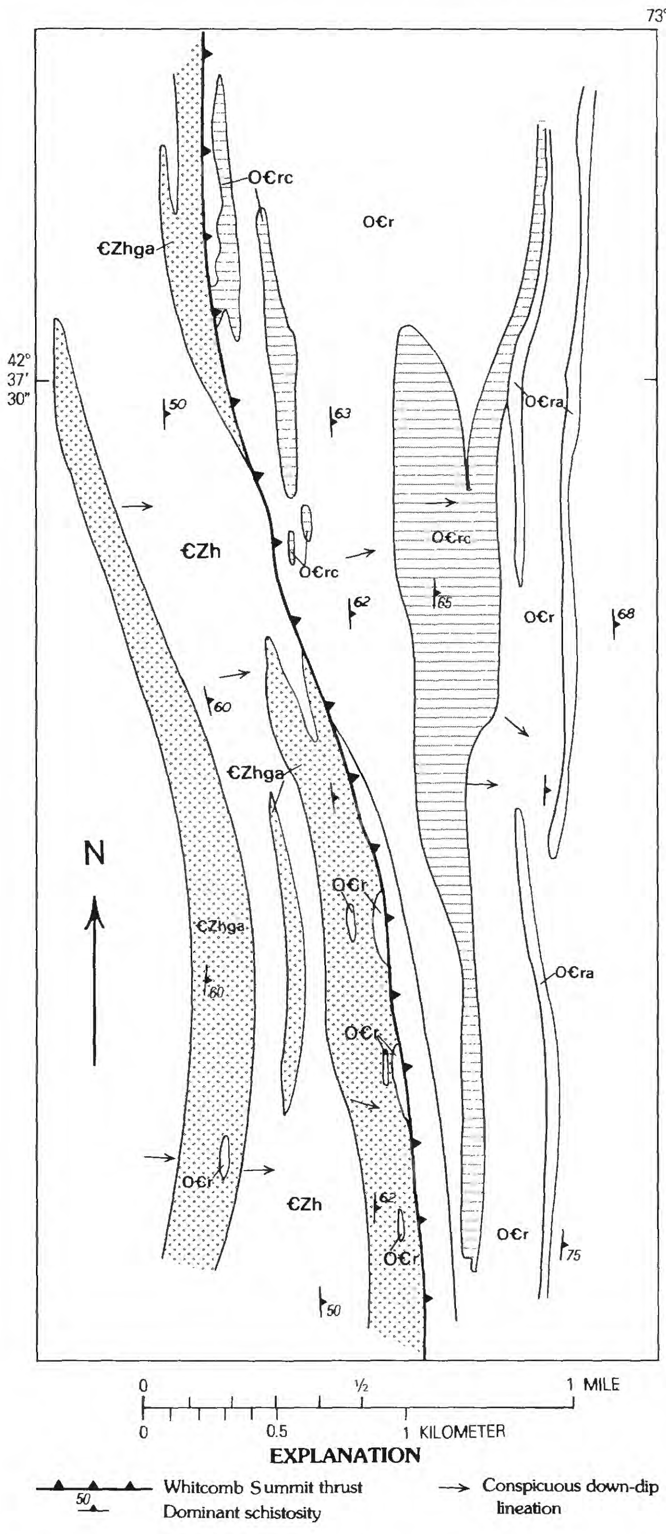


FIGURE 25.—Distribution of lithic units in the Hoosac Formation and the Rowe Schist in the vicinity of the Whitcomb Summit thrust near Borden Mountain. Geology modified from Norton (1967) and Chidester and others (1967). Note that the thrust truncates units in both formations. Lithic symbols corresponding to those on the State map are: ЄZh and ЄZhga, allochthonous Hoosac Formation; OЄr, OЄra, and OЄrc, Rowe Schist.

sequence (Taconic-Berkshire zone of Massachusetts). This line has long been recognized in Connecticut as a fundamentally important tectonic surface (see, for example, Rodgers, 1970, p. 94). Hall (1980, p. 304-305) showed it as the root zone for the Taconic allochthon, whereas Stanley and Ratcliffe (1980, 1983, 1985) described it as a thrust fault that has covered the root zone of the Taconian slices.

The Whitcomb Summit thrust has not been traced north into Vermont and its existence there may be questioned. However, if we consider the thrust to be an east-dipping surface separating the package of rocks on the east, which contain ultramafic slivers, from the package of rocks to the west, which lack ultramafic slivers, then its trace in Vermont could be drawn along the west contact of the Ottauquechee with the Pinney Hollow, at least as far north as the village of Moretown, Vt. (see Doll and others, 1961). At Moretown the contact between the Hazens Notch and the Pinney Hollow terminates against the western contact of the Ottauquechee. Thompson (*in* Chang and others, 1965, p. 20-21) described the Pinney Hollow-Ottauquechee contact in the Woodstock, Vt., quadrangle about 60 km south of Moretown, as "a transition zone a third of a mile wide" in which "at least three distinct bands of black phyllitic quartzite, each about 200 feet wide, are intercalated [with] pale green chlorite-sericite-quartz schist of the Pinney Hollow Formation [that] is here locally albitic and biotitic." He goes on to say that "similar conditions are found both farther north and south along the strike. The width of this transition zone, however, is not constant, and it may be wholly a zone of infolding of Pinney Hollow and Ottauquechee types." These relations are identical with those we have observed along and east of the Whitcomb Summit thrust in Massachusetts, and they lead us to propose that the Whitcomb Summit thrust continues northward at least to Moretown, Vt., approximately along the Pinney Hollow-Ottauquechee contact.

The problem of the location of the Whitcomb Summit thrust north of Moretown, Vt., is outside the realm of this paper; interested readers are referred to Stanley and Ratcliffe (1985) for further discussion.

ROWE THRUST ZONE

As noted above in the description of the lithostratigraphic units, the Rowe Schist consists of many lenses of green Pinney Hollow- and Stowe-like schist, gray Ottauquechee-like schist, amphibolite, and ultramafic rock. Although we long interpreted these lenses to be sedimentary and the ultramafic rocks to be solid intrusions, we now believe that the Rowe Schist is a complex tectonic zone in which many of the contacts between lenses are thrust faults and in which other thrusts pass unrecognized through lithic units.

Our views on the Rowe Schist are greatly influenced by Stanley's work in the Missisquoi River valley in northern Vermont, where ultramafic rocks abound in the Hazens Notch, Ottauquechee, Stowe, and Moretown Formations. Detailed 1:10,000-scale mapping by Stanley and Roy (1982; Stanley and others, 1984) has shown the following important relations: (1) The present configuration of map units is a result of tectonic juxtaposition of depositional sequences along faults of undetermined displacement. (2) Deformed metadiabasic dikes are restricted to fault slices distinguished by mafic volcanic rocks, fine-grained phyllites, and metasiltstones interpreted as having been deposited on the ocean floor. (3) Serpentinites and related rocks are only found as slivers along faults; ophiolitic olistostromes have not been recognized although they may have existed during the early tectonic history of the belt. (4) Four fold generations provide a relative time scale to which three fault generations can be related. Most of the faults cut F_1 and are coeval with F_2. These data suggest a complicated evolution involving early mafic intrusion into ocean-floor sediments, westward imbrication, and associated folding, followed by refolding and fault reactivation, rather than simple westward accretion above an east-dipping subduction zone. The structural fabric is similar to that described from modern exposed accretionary prisms (Moore and others, 1980; Stanley and others, 1982). On the basis of recent work by Laird and Albee (1981a,b), Sutter and others (1985), and Laird and others (1982), deformation and major metamorphism are considered to have occurred during the Taconian orogeny, and regional tilting and folding during the Acadian orogeny.

The Vermont ultramafic belt is traceable into the Rowe-Moretown interval in Massachusetts and its southward continuation in western Connecticut. It is apparent from our previous description that many of the structures and textures observed in Vermont are present, though far more cryptically, in and around the ultramafic bodies to the south. For example, detailed mapping by Knapp (1977; fig. 11) at Prospect Hill in the West Granville, Mass., quadrangle has shown that the whole belt of Rowe Schist from the Whitcomb Summit thrust into the western part of the Moretown Formation consists of discontinuous layers of mineralogically and texturally distinct aluminous schist bordered to the west by a fairly continuous Rowe amphibolite. Ten steatite and serpentinite lenses are scattered within and along contacts in the schist. Subsequent mapping by Brill (1980) at an even larger scale (1:6,000) confirmed Knapp's work. We, therefore, interpret the ultramafic rocks and their associated metagabbros and metamorphosed mafic volcanic rocks as fragments of ocean crust tectonically incorporated into continental rise-ocean floor sediments during underplating at an accretionary prism. Subsequent faulting, as suggested by the northern Vermont data, has reworked the original imbricated sequence to produce the relations observed today. The resulting composite fault zone of closely spaced, possibly anastomosing, thrust surfaces is here termed the Rowe thrust zone. Acadian movement on this zone seems unlikely because Acadian metamorphic isograds are undeflected across the Whitcomb Summit thrust (Zen and others, 1983; John Cheney, oral commun., 1982). Furthermore, early Acadian F_2 small-scale isoclinal folds fold these thrusts in the Rowe quadrangle.

The present dominant fabric of the metasedimentary rocks (O€r and O€rc) of the Rowe Schist is a very strong schistosity that is parallel to lithic boundaries in the formation. Bedding has been recognized only very locally. Previously we (Hatch, Osberg, and Norton, 1967; Hatch, 1975) had explained this schistosity as being axial planar to small isoclinal folds that we interpreted to be the same generation as the Acadian F_2 isoclines in the Lower Devonian Goshen Formation to the east. Strongly rodded lenses of vein quartz 1-2 cm thick and 10-20 cm long are particularly characteristic of the O€r units of the Rowe (fig. 26). The steep plunges of these rods, in contrast to the subhorizontal axes of F_2 folds in the post-Taconian Goshen Formation, had been explained by an episode of Taconian folding (F_1) that

FIGURE 26.—Reclined isoclinal folds outlined by deformed quartz veins in unit O€r of the Rowe Schist resulting from Taconian shearing along the Whitcomb Summit thrust. Outcrop is at Whitcomb Summit (fig. 4), 15 m above the thrust. Silver tip of pencil is about 2 cm long. View looking west.

left steeply dipping beds and foliation in the pre-Silurian strata, in contrast to the generally horizontal beds in the post-Taconian strata, at the time of the first Acadian (F_2) folding event (Hatch, Osberg, and Norton, 1967; Hatch, 1975). Although this model adequately explained the diverse attitudes of axes of F_2 folds in beds in the Moretown, Hawley, and Cobble Mountain Formations, it did not explain the remarkably consistent directly downdip attitude of quartz rods (if they were attributed to F_2 folding) in the Rowe.

The Middlefield thrust zone (Norton, 1971, 1975; Ratcliffe and Hatch, 1979), the Whitcomb Summit thrust (Stanley, 1978), and the Rowe thrust zone (this paper) and their associated fault fabrics appear to offer a better explanation for the absence of bedding, the strong schistosity fabric, and the consistently downdip orientation of the quartz rods (fig. 25). We now believe that the quartz rodding is a Taconian intersection or stretching lineation rather than an Acadian fold axis lineation. We further believe that the Rowe Schist fabric is roughly contemporaneous with the "fold-thrust fabric" of Ratcliffe and Harwood (1975) and Harwood (1975) in the Berkshire massif to the west and with the fault fabric in the Hoosac Formation adjacent to the Middlefield thrust (Ratcliffe, 1979b; Ratcliffe and Hatch, 1979). The absence of bedding in the Rowe could result either from its never having formed in the first place or, if once formed, from having been destroyed either at the time of emplacement of the ultramafic rocks into the Rowe or during subsequent Whitcomb Summit and Middlefield thrust time.

Ratcliffe (1979a) has argued that when the Hoosac Formation and Berkshire massif were displaced over the western autochthon in the Taconian orogeny, distinctive downdip lineations were produced by the axes of reclined folds and by mineral streaking. This lineation persists in the Taconian thrust fabric in the Hoosac and the Proterozoic Y rocks near the Whitcomb Summit thrust, in the Middlefield thrust zone (Hoosac Summit thrust), and in the sole thrust of the Berkshire massif westward all across the Berkshires, regardless of whether the present thrust fabric dips east or west. We had earlier been impressed, however, by the apparent continuity of the Rowe fabric eastward into a fabric that had to be Acadian because it affected the Lower Devonian strata of the Connecticut Valley synclinorium. The reason for this continuity appears to be that, in the area east of the Middlefield (Hoosac Summit) thrust zone where the Taconian faults are essentially vertical (see cross sections for the State bedrock map), the Taconian fault fabric and the Acadian F_2 isoclinal fold schistosity are closely parallel and superposed, and the schistosity in the rocks is a composite schistosity. East of the Rowe fault zone (the extreme western edge of the Moretown Formation), the strong downdip lineation dies out rapidly leaving only one apparent schistosity, parallel to Acadian F_2 schistosity that does indeed trace eastward into the Devonian strata. Although it is difficult to distinguish the two superposed foliations where they are parallel in the Rowe Schist and the eastern Hoosac, west of the Middlefield thrust zone the Taconian structures flatten out and a superposed vertical Acadian S_2 schistosity or crenulate cleavage is more conspicuous where present. Although a steep crenulation cleavage is locally superposed on older fabrics in the Hoosac schists beneath the Rowe and Whitcomb Summit thrust zones and in the Stockbridge valley to the west (Ratcliffe and Hatch, 1979; Ratcliffe, 1979b), the fact that a steep schistosity is not traceable west of the Rowe thrust zone indicates that the Acadian F_2 fold fabric has not imparted a schistosity to the rocks west of the Rowe and Whitcomb Summit thrust zones.

BRISTOL THRUST

The contact of the Rowe-Hawley zone with the Bronson Hill zone, or what Robinson and Hall (1980) called the Bronson Hill plate, is considered by us to be tectonic and is designated the Bristol thrust for its exposure in the Bristol dome, Connecticut (fig. 10). The evidence for the tectonic interpretation is the following: (1) The Taine Mountain Formation in the core of the Bristol dome is lithically similar to and apparently stratigraphically correlative with the Moretown Formation of probable Ordovician age (Hatch and Stanley, 1973). (2) The Collinsville Formation structurally overlies the Taine Mountain Formation in the Bristol and Waterbury domes. The contact is discordant with the three members of the Taine Mountain in the Bristol dome (fig. 10). (3) The interlayered gneisses and amphibolite (Ocoh, fig. 10) of the Collinsville Formation are lithically similar to the Ammonoosuc Volcanics of the Bronson Hill. The homogeneous plagioclase gneiss (Ocob) that is exposed in the Shelburne Falls, Goshen, Collinsville, Bristol, and Waterbury domes is lithically similar to the Monson and Fourmile Gneisses (OZmo, OZfm) of the Pelham dome. These gneisses may be as old as late Precambrian. These similarities constitute the basis of the correlation shown in figure 22 and the cross sections of the State bedrock map. Of particular importance is the fact that the Monson Gneiss is plutonic in origin, as indicated by its trondhjemitic composition (Robinson and Hall, 1980). Monson-like rocks are not present in the Taine Mountain Formation, which is largely metamorphosed turbidites. (4) As shown in figure 22 and the cross sections on the State bedrock map, the gneisses and amphibolites of the

Collinsville Formation (excluding Ococ and the Sweetheart Mountain Member of Stanley (1964)) form a westward-thinning wedge that is absent to the west in the Hawley and Moretown Formations. Support for the presence of a westward-thinning wedge is seen in the gravity data of R.W. Simpson (1974), who estimated the core gneiss in the Shelburne Falls dome to be 1,000-1,200 m thick, approximately the same thickness measured by L.M. Hall (written commun., 1977). Thus these rocks are substantially thinner than they are to the east in the Bronson Hill (Peter Robinson, oral commun., 1982).

On the basis of the foregoing evidence, we suggest that rocks of the Bronson Hill zone have been transported westward over the Moretown Formation of the Rowe-Hawley zone. We further suggest that major displacement on this thrust occurred before the deposition of the Hawley and equivalent rocks. Renewed movement may have occurred thereafter. The critical evidence is the fact that member C of the Cobble Mountain Formation and approximately equivalent black schists and volcanics of the Hawley and Partridge Formations rest on the core rocks of the domes both west and east of the Mesozoic basins as well as on the Moretown of the Rowe-Hawley zone (fig. 22).

STRUCTURE OF THE COBBLE MOUNTAIN FORMATION

The lithic complexity of the Cobble Mountain Formation and its relationship to the Collinsville and Hawley Formations have been discussed in previous paragraphs of this chapter. Our interpretation of these complexities, however, is critical to our model for the tectonic and stratigraphic evolution of western Massachusetts and thus is presented in some detail in the following paragraphs.

The base of member C of the Cobble Mountain was first recognized as a thrust by Knapp (1977, 1978) in the West Granville quadrangle and was named for Winchell Mountain in that area. The Winchell Mountain thrust extends south from the southeastern part of the Woronoco quadrangle through the Southwick and West Granville quadrangles into western Connecticut (fig. 10; State bedrock map). The thrust truncates lithic units in member B (Ocb, Ocbr) of the western plate (inferred footwall) and member C (Occ, Occr) of the eastern plate (inferred hanging-wall) (figs. 10, 20). The age of the Winchell Mountain thrust can be bracketed on Russell Mountain in the Woronoco quadrangle (figs. 10, 27; Stanley and others, 1982) where it is isoclinally folded with Ocbr and Occa and truncated by the inferred unconformity at the base of member D. Stanley and others (1982) suggested a Late Ordovician

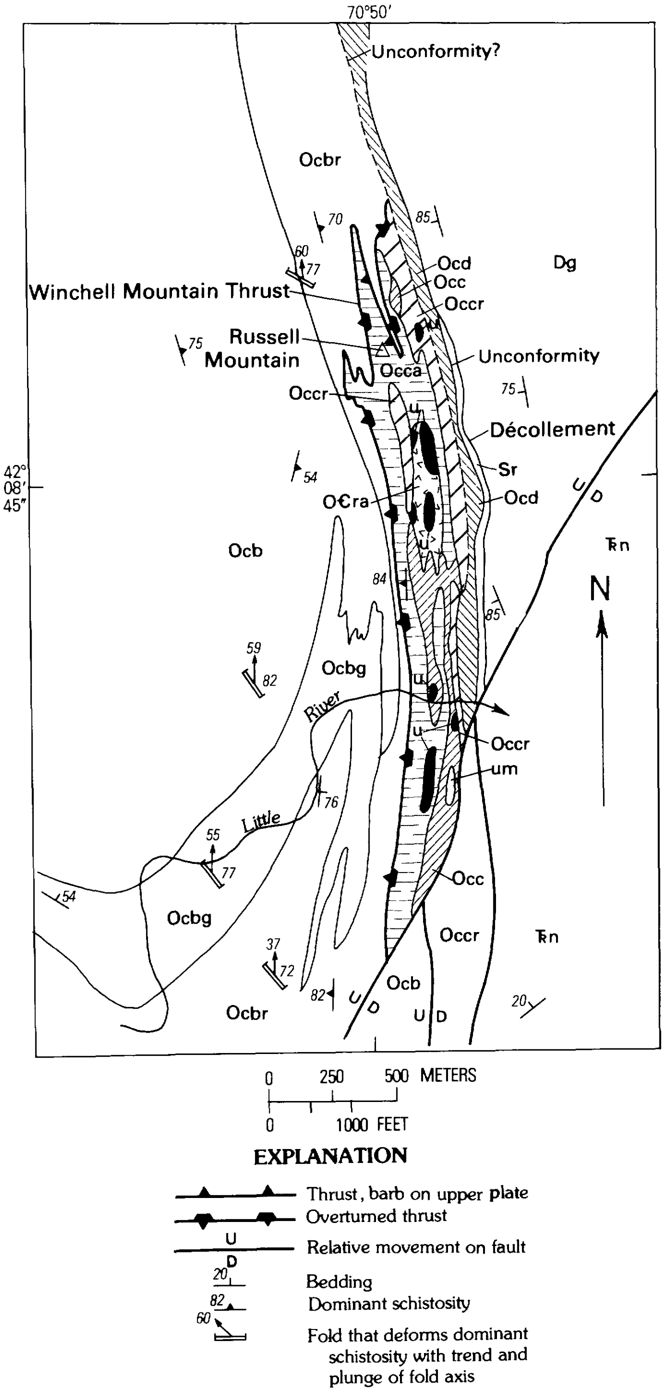


FIGURE 27.—Detailed geologic map of Russell Mountain showing distribution of lithologies within Cobble Mountain Formation member C, the complexly folded Winchell Mountain thrust, the unconformity(?) at the base of member D, and the unconformity-decollement at the base of the Silurian and Devonian section. Lithic symbols corresponding to those on the State bedrock map are O€ra, amphibolite in the Rowe Schist; Ocb, Ocbr, Occ, Occr, Occa, and Ocd, member B, rusty member B, member C, rusty member C, aluminous member C, and member D of the Cobble Mountain Formation; u, ultramafic rock; Dg, Goshen Formation; Ɨn, New Haven Arkose. Ocbg is a garnet-rich unit in member B. Um is a small lensoid body of actinolite marble. Geology enlarged from Stanley and others (1982).

to Early Silurian age for this unconformity because it appears to be cut by the regional Taconian unconformity at the base of the Silurian-Devonian section.

The upper plate of the Winchell Mountain thrust contains all of the 11 bodies of ultramafic rock in western Massachusetts that have been found in strata above the lower part of the Moretown Formation. As discussed in an earlier section of this chapter, the remaining ultramafic rocks are confined to the Rowe and the westernmost part of the Moretown. Ultramafic intrusions in the northern end of the Precambrian of the Berkshire massif are lithically and structurally distinct and are believed to have very different origins from those of either the Rowe and Moretown formations or the Cobble Mountain Formation.

The origin of the ultramafic bodies in the Cobble Mountain Formation is uncertain. Although there is little doubt of their original igneous parentage (Chidester, 1968), evidence bearing on their actual mode of emplacement into the surrounding metasedimentary rocks has been destroyed or obscured by Acadian deformation (Stanley, 1975; Knapp, 1977) and kyanite-sillimanite grade metamorphism. It is clear that they were transported with the upper plate of the Winchell Mountain thrust and that their emplacement must have predated that thrusting. We have argued earlier in this chapter that the ultramafic and associated rocks in the Rowe-lower Moretown Formations are tectonic slivers emplaced along major thrust slices in the Rowe thrust zone, which was then transported westward en masse on the Whitcomb Summit thrust. We believe, however, that the ultramafic rocks, as well as the Rowe-like aluminous schists and laminated amphibolites in member C of the Cobble Mountain Formation, have a sedimentary origin and represent exotic fragments in an olistostromal deposit. This interpretation is based on the following observations. (1) Six of the eleven ultramafic bodies in member C are along contacts with thinly laminated amphibolite, aluminous schist (Occa) or nonrusty-weathering silvery-gray schist (Occ). The silvery-gray schist forms the main part of the member. (2) The thinly laminated amphibolite and the aluminous schist are very similar to comparable rocks of the Rowe Schist. In fact, the aluminous schist is so like the Rowe that Emerson (1898) mapped it as Rowe. Hatch and Stanley (1973, p. 11-12) recognized this similarity but mapped the unit as Cobble Mountain Formation because it is complexly intermixed with rocks that are very different from the Rowe. Furthermore, the matrix aluminous schist (Occa) contains more biotite than do the schists of unit O€r of the Rowe and does not have the bluish cast or thin lenses and laminae of quartz so typical of the aluminous schist in the Rowe. (3) Although, on Russell Mountain, some of the contacts of the aluminous schist (Occa) are sharp, most are gradational there and elsewhere with Occ or Occr (fig. 10).

In order to understand the origin of the Cobble Mountain Formation within the otherwise black shale-volcanic terrain of Middle Ordovician age, it is useful to consider the possible plate-tectonic environment suggested by the rocks as they are presently known. Our model was derived by reversing, or retrodeforming, the present-day geological relations as shown on the State bedrock map and its cross sections and from the inferred lithic relations shown in figure 22. We also drew upon a present-day example in the eastern part of Taiwan, which Stanley has visited with Suppe, Liou, Lan, and Ernst, who have done the most recent work on the Lichi Melange (Page and Suppe, 1981). The schematic diagrams in figure 28 begin at the close of the Taconian orogeny before deposition of the Silurian and Devonian sequence and locally before deposition of member D of the Cobble Mountain Formation. The subsequent deformation in the Acadian orogeny is not included in this sequence of diagrams, because it is shown, for the area of cross section *F-F′*, in figure 17 of Hatch and Stanley (Ch. C, this volume). The diagrams in figure 28 are also at the latitude of cross-section *F-F′* where the Cobble Mountain Formation is exposed.

Figure 28*A* shows the inferred conditions just after movement on the Winchell Mountain thrust and before erosion of the older terrane. Subsequently, member D of the Cobble Mountain Formation was deposited across the eastern part of the forearc basin. The truncated map units in members B and C mapped by Knapp (1977, 1978) in the West Granville area and by Stanley and others (1982) in the Woronoco quadrangle clearly define the thrust zone and show that it was active before deposition of member D. On the basis of the interpretation that the rusty- and nonrusty-weathering schists of member C are a distal facies of member B, we show the Winchell Mountain thrust rooting to the west and climbing section to the east where it may have broken the submarine surface forming a ridge. Erosion along this eastern front may have produced some of the altered ultramafic rocks reported by Tracy and others (1984) in the Partridge Formation along the Bronson Hill anticlinorium. We further suggest that the Winchell Mountain thrust may have developed as an upper level backthrust possibly due to the resistance generated by the stacked Taconian slices to the west.

Returning the upper plate of the Winchell Mountain thrust to its western root zone results in figure 28*B*, which depicts the stratigraphic relations of the Middle Ordovician units after the transgression of the black shales and cherts eastward across the Bronson Hill suggested in figure 22. Volcanic rocks from the arc continued to mix with these black shales and cherts as they did with the older rocks of the Cobble Mountain and lower Hawley Formations. We believe the Middle Ordovician sequence of black shales, volcanic rocks, and volcanogenic flysch of the Hawley-Cobble Mountain-Partridge interval is a west-to-east, time-transgressive package that formed in a forearc basin west of the Bronson Hill arc complex receiving sediment from and, at times, covering the accretionary wedge. The wedge emerged periodically, forming nonvolcanic islands that tended to isolate the forearc region from the basin to the west along the margin of the continent (fig. 28*C*). With time, the sequence transgressed eastward to form the Partridge Formation, which unconformably overlies the Ammonoosuc and older rocks of the Bronson Hill arc complex. During times of reduced compression, however, the rocks of the forearc region were probably continuous with the Normanskill basin along the continental margin landward of the accretionary wedge. As the basin between the Bronson Hill arc complex and the North American plate continued to close, the slope-rise sequence was driven landward and formed the allochthonous terrain of the Taconics and isolated the still more landward basin (exogeosyncline) of the Walloomsac Formation (Stanley and Ratcliffe, 1985, pl. 2, sec. 7 and 8).

There is no evidence that the tectonic activity suggested by the volcanogenic flysch of the Cobble Mountain Formation extended eastward to the main part of the Bronson Hill or westward into the Walloomsac terrain. If it did extend to the east, it was eroded before the deposition of the Partridge Formation. The black shales and cherts of the Hawley do suggest either a period of subdued compressional activity in the overall collision between the Grenville and Bronson Hill plates or a considerable separation between the Bronson Hill plate and the North American continental edge at that time. The presence of black shales and cherts in what is believed to have been a forearc basin is anomalous; in similar basins in modern forearc environments, the sediments are coarse-grained clastics representing material from the volcanic arc and the emerged accretionary wedge (for example, the Takangkou, Chimei, and Lichi Formations in Taiwan (Chi and others, 1982); the Nias beds on Nias Island, Indonesian arc (Moore and others, 1980)). As we pointed out, however, in other respects the rocks of the Cobble Mountain Formation are quite similar to forearc deposits.

Figure 28*C* represents the conditions during the formation of member C of the Cobble Mountain Formation. As pointed out earlier, this unit contains mappable (at 1:24,000 scale) and smaller bodies of serpentinized ultramafic rock (u), Rowe-like amphibolite (O€ra), and aluminous Rowe-like schist (O€r). These bodies are enclosed in a matrix of silvery-gray, aluminous, nonrusty-weathering schist, and the whole unit is mapped as Occa within member C. As shown in figure 28*C*, we believe that the Rowe-like rocks and serpentinites were eroded from the steepened eastern flank of the accretionary wedge ("tectonized Rowe" on fig. 28*C*). The wedge consisted of imbricated ocean basin-continental rise sediments and represented an earlier stage in the development of the Rowe Schist. Fragments of the ocean crust had already been incorporated into the wedge as slices and slivers. Parts of the accretionary wedge that had emerged from below wave base and formed an outer, nonvolcanic arc were eroded, and olistostromes were deposited in the distal volcanogenic shales of member C. The aluminous schist (Occa) represents not only fragments of Rowe aluminous blue-green schist but finer Rowe-derived detritus mixed with feldspathic schistose wackes (Occ) from the island arc to the east. This mixture forms those parts of Occa that are aluminous but distinguishable from the Rowe green schist. Subsequent erosion and subdued compressional activity allowed black shales to transgress eastward forming the configuration of figure 28*B*. An excellent modern analog of our interpretation is the Pliocene Lichi Melange with its exotic blocks of ophiolitic material in the Coastal Range of eastern Taiwan (Hsu, 1956; Liou and others, 1977; Ernst, 1977; Page and Suppe, 1981; Suppe and others, 1981).

We interpret members A and B of the Cobble Mountain as the distal and proximal facies, respectively, of a volcanogenic flysch sequence largely derived from the westward-advancing Bronson Hill volcanic arc microcontinent complex. To the west, between the carbonate bank and the accretionary wedge, a stagnant basin received black muds. These deposits appear to have spread eastward through subaqueous depressions in the accretionary wedge, where they formed the Hawley Formation and interfingered with distal flysch (member A) eroded from a possible promontory of the Bronson Hill. Erosion from the emerged parts of the accretionary wedge undoubtedly contributed material both to the east and to the west. The only such deposit recognized south of Quebec is the Umbrella Hill Conglomerate in north-central Vermont (Badger, 1979). Volcanic material from the advancing arc spread westward and is represented today by the mafic and felsic volcanic rocks in the Hawley Formation and by the feldspathic wackes in the Cobble Mountain Formation. Amphibolites and felsic volcanic rocks are more abundant in members A and B than they are in C, which reinforces our interpretation that the lower two

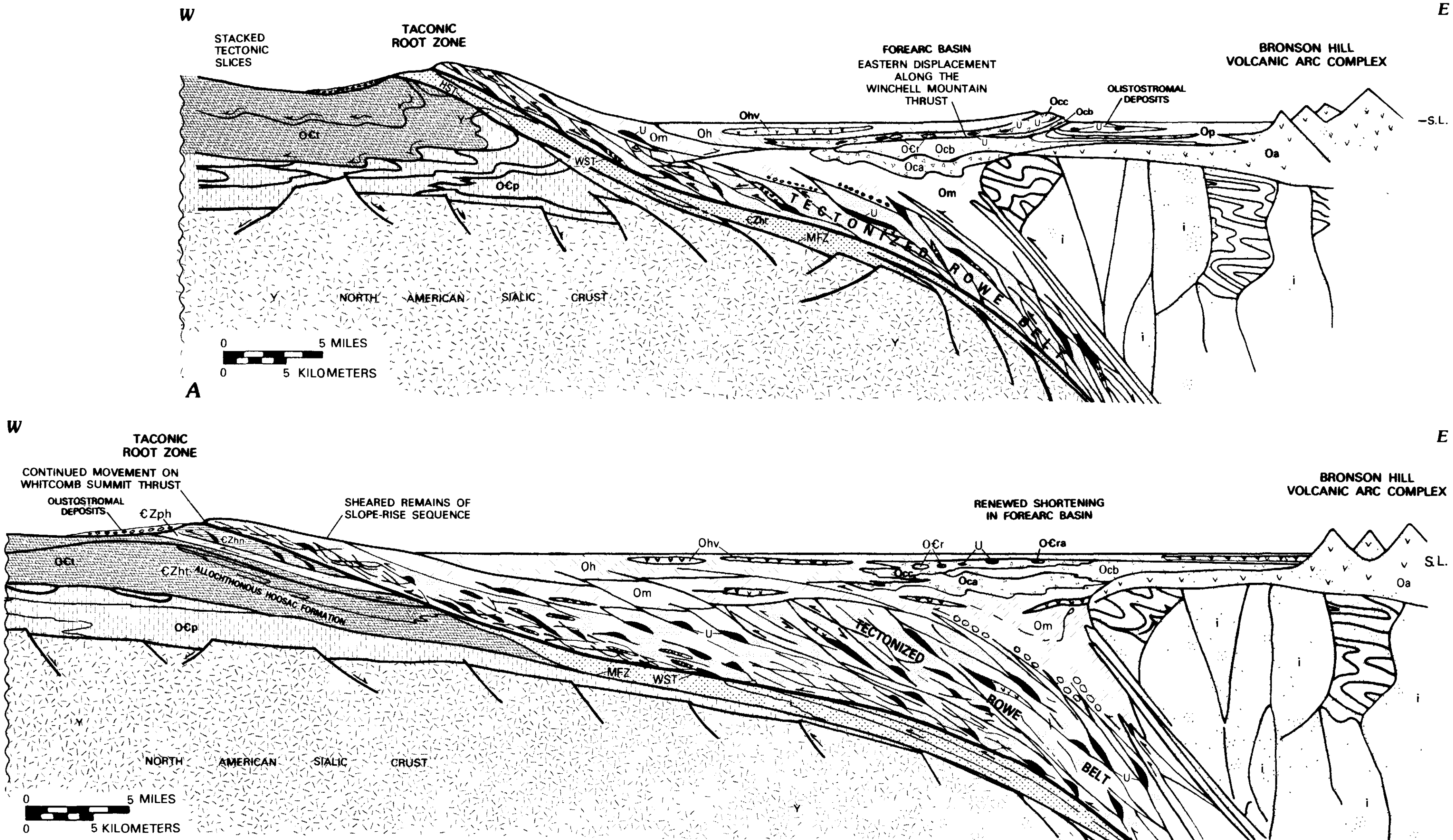
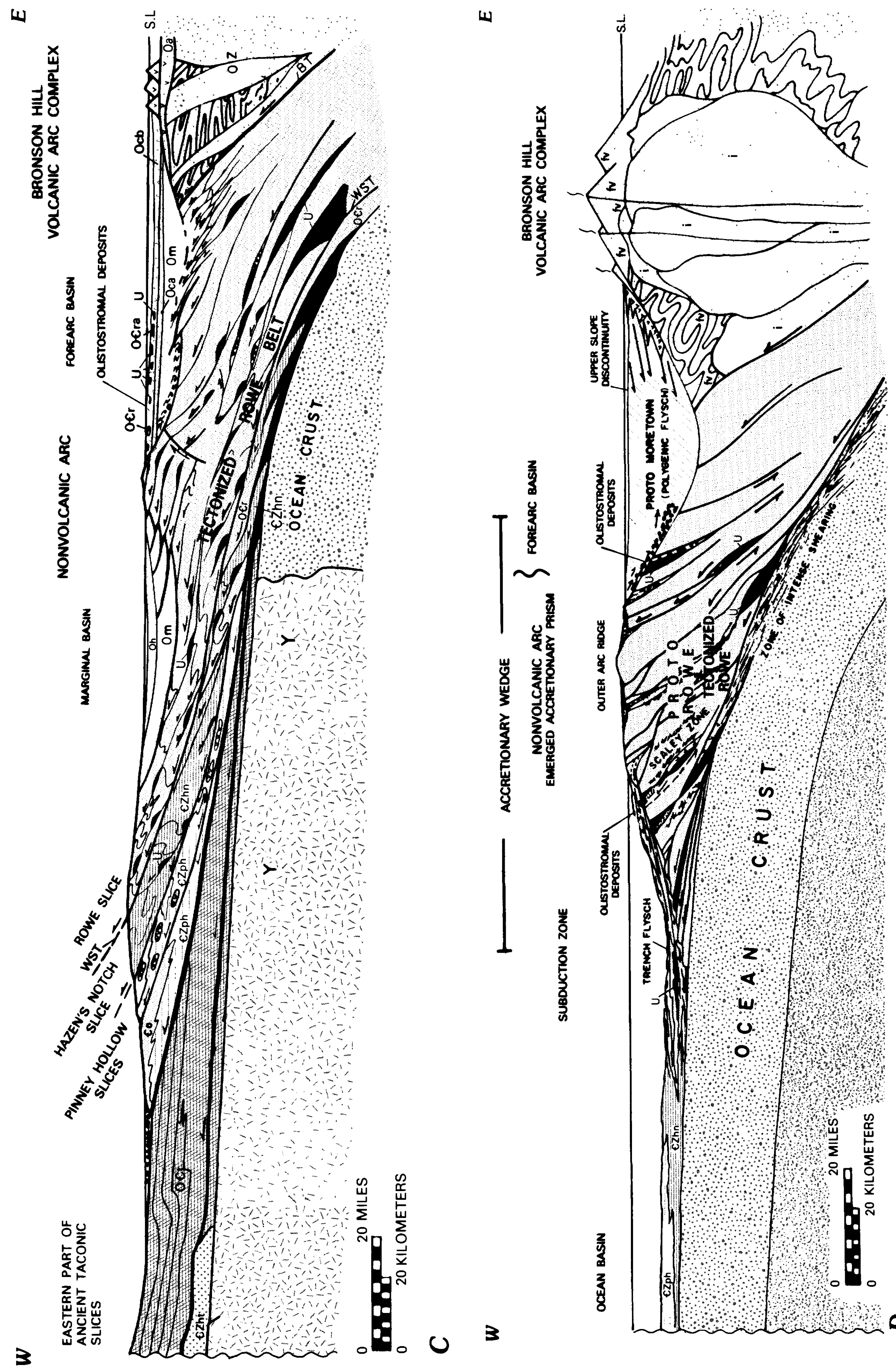


FIGURE 28.—Tectonic evolution of the Cobble Mountain Formation and the Winchell Mountain thrust as depicted in a sequence of west-east schematic diagrams at the latitude of State map cross section *F-F′*. Figure 28*A* shows the inferred configuration near the close of the Taconian orogeny immediately before deposition of Cobble Mountain member D and after eastward displacement on the Winchell Mountain thrust. Figures 28*B* and 28*C* successively remove deformation and rocks deposited through time to achieve the inferred configuration in pre-Cobble Mountain time depicted in figure 28*D* when the accretionary prism was an undetermined distance from the slope-rise sequence represented today by the rocks of the Taconic allochthons. The formation of member C of the Cobble Mountain Formation with its exotic fragments of Rowe Schist and ultramafic rocks is shown in figure 28*C*. These diagrams have been modified from sections 5, 6, 7, and 8 of plate 2 from Stanley and Ratcliffe (1985). S.L., sea level. Lithic symbols are described from west to east in each figure. *A*, Y, Proterozoic Y rocks of the North American sialic crust; O€p, rift-clastic rocks and carbonate-siliciclastic rocks of the North American platform; O€t, slope-rise rocks of the Taconic slices; €Zht, Hoosac Formation; HST, Hoosac Summit thrust; MFZ, Middlefield fault zone; WST, Whitcomb Summit thrust; Rowe Schist is represented by the tectonized Rowe belt; U, ultramafic rocks; lenses with "v" pattern represent mafic volcanic rocks; Om, Moretown Formation; Oh, Hawley Formation; Ohv, mafic volcanic rocks in the Hawley Formation; Oca, Ocb, Occ are members of the Cobble Mountain Formation; O€r, fragments of Rowe Schist in member Occ; Op, Partridge Formation; Oa, Ammonoosuc Volcanics; i, intrusive rocks. *B*, Additional symbols are €Zj and €Zhn, which refer to the Pinney Hollow and Hazens Notch Formations of Vermont. These rocks are considered thrust slices by Stanley and Ratcliffe (1985). O€ra, fragments of Rowe mafic volcanic rocks in member C of the Cobble Mountain Formation. *C*, €o, Ottauquechee Formation of Vermont; BT, Bristol thrust; OZ, Monson Gneiss and related rocks. *D*, fv, felsic volcanic rocks.


W
E
EASTERN PART OF ANCIENT TACONIC SLICES
MARGINAL BASIN
NONVOLCANIC ARC
FOREARC BASIN
BRONSON HILL VOLCANIC ARC COMPLEX
OLISTOSTROMAL DEPOSITS
PINNEY HOLLOW SLICES
HAZEN'S NOTCH SLICE
WST
ROWE SLICE
TECTONIZED ROWE BELT
OCEAN CRUST
S.L
0 20 MILES
0 20 KILOMETERS
C
W
E
OCEAN BASIN
SUBDUCTION ZONE
ACCRETIONARY WEDGE
NONVOLCANIC ARC
EMERGED ACCRETIONARY PRISM
FOREARC BASIN
OUTER ARC RIDGE
TRENCH FLYSCH
OLISTOSTROMAL DEPOSITS
UPPER SLOPE DISCONTINUITY
PROTO MORETOWN
(POLYGENIC FLYSCH)
PROTO ROWE
TECTONIZED ROWE
SCALEY ZONE
ZONE OF INTENSE SHEARING
OCEAN CRUST
S.L
0 20 MILES
0 20 KILOMETERS
D

members were deposited closer to the volcanic arc than was member C.

The geologic relations in pre-Middle Ordovician time are more difficult to define. Thus, the configuration we have shown in figure 28*D* is even more speculative than previous diagrams in the sequence. We have shown an accretionary wedge-volcanic arc complex with an intervening forearc basin. Oceanic crust attached to the North American plate forms an east-dipping slab beneath the accretionary wedge-arc complex to the east. We do not know when this subduction began, but it may have been in Late Cambrian to Early Ordovician time. Figure 28*D* represents the time before deposition of the Cobble Mountain Formation and after deposition of the Moretown Formation.

In summary, the diagrams in figure 28 are an attempt to show the plate-tectonic evolution of the Cobble Mountain Formation and the surrounding rocks in what we believe was Middle Ordovician time. These events are part of a total westward displacement on the order of 1,000 km. This figure is based on palinspastic restoration of the Taconic allochthons to their depositional sites and on estimated displacements along the Middlefield thrust, the Whitcomb Summit thrust, the Rowe thrust zone, and the Bristol thrust (see fig. 28). If our interpretation presented here is correct, the Cobble Mountain Formation provides an important clue to the plate-tectonic evolution of western New England. We believe that the events taking place today in such areas as eastern Taiwan lend support to our model of events that took place in the Middle Ordovician in the western part of Iapetus.

SUMMARY

From our discussion of the Rowe-Hawley zone, it is clear that major Taconian thrust surfaces (Whitcomb Summit thrust) and thrust zones (Rowe-western Moretown) are a principal factor contributing to the linearity and apparent simplicity of the formations on the east limb of the Berkshire massif. During the Taconian orogeny, westward displacement on the order of 1,000 km of an early-developing accretionary wedge (Stanley and Ratcliffe, 1980, 1983, 1985) tectonically interlayered slope-rise-ocean-floor sediments and fragments of the ocean crust and mantle and smeared them out parallel to thrusts such as the Whitcomb Summit thrust. Acadian metamorphism and deformation have severely overprinted fault-zone fabrics so that the principal surviving evidence of these major thrusts is the truncated lithic fabrics of the thrust-bounded packages.

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Stratigraphy of the Connecticut Valley Belt

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THE BEDROCK GEOLOGY OF MASSACHUSETTS

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ILLUSTRATIONS

STRATIGRAPHY OF THE CONNECTICUT VALLEY BELT

By NORMAN L. HATCH, JR., PETER ROBINSON,[1] and ROLFE S. STANLEY[2]

ABSTRACT

The Connecticut Valley belt of west-central Massachusetts includes metamorphosed Silurian and Lower Devonian strata between a major regional unconformity and "surface of structural disharmony" along the east side of the Berkshire Hills, on the west, and a north-south line on the east side of the first outcrop belt of Littleton Formation east of the main body of Monson Gneiss.

Silurian strata include the Russell Mountain Formation (unit Sr), the Clough Quartzite (unit Sc), and the Fitch Formation (unit Sf). The Russell Mountain Formation is a thin, discontinuous quartzite and calc-silicate granulite unit along the southern quarter of the western margin of the belt. The Clough Quartzite is a thin discontinuous unit of quartz-pebble conglomerate and quartzite that is largely restricted to synclines in and immediately east of the Bronson Hill anticlinorium. The Fitch Formation in Massachusetts consists of a few thin lenses of calc-silicate granulite and minor pelitic schist near Bernardston[3] and a short, narrow, north-south belt in the vicinity of Orange.

Lower Devonian strata of the belt are divided into five formations of primarily metasedimentary rocks and one localized thin metavolcanic unit. The Littleton Formation (unit Dl) has been mapped more or less continuously from its type area at Littleton, N.H. It consists predominantly of gray carbonaceous pelitic schist, present in eight areas within the Connecticut Valley belt. In areas 1 and 2, immediately north and west of the Mesozoic basins, the rocks are low- to medium-grade, dark-gray, graphitic, aluminous slate or phyllite with minor indistinct 1-mm- to 2-cm-thick graded beds of fine-grained, light-gray quartzite. East of the Mesozoic basins in areas 3 and 4, and the western part of area 5, in the Northfield, Wendell, and Great Hill synclines, the Littleton is an aluminous, gray staurolite schist with generally thin (1-2 cm), but locally thick (about 1 m), quartzite beds. These rocks reach sillimanite grade in the eastern part of area 5 and have in turn been pervasively retrograded to biotite and chlorite zone assemblages in the southeast part of area 5. The Littleton strata of area 6, in the Pelham-Shutesbury syncline, are coarse, gray, locally well bedded muscovite-biotite schists containing garnet, kyanite, and locally staurolite. Biotite-rich schist, feldspathic schist, mica-feldspar gneiss, and local lenses of garnet quartzite and magnetite iron formation characterize the Littleton of area 7, a complex belt generally along the east side of the Keene dome and the main body of Monson Gneiss, which forms an eastern facies of the Littleton.[4] The Littleton in the inliers of Paleozoic rocks around Amherst, area 8, resembles the Littleton in the higher grade parts of area 7.

Stratigraphically overlying the Littleton, the Erving Formation (unit De) is present in a narrow belt generally within 10 km east of the Mesozoic border fault in synclines along the Bronson Hill anticlinorium, and on the Whately anticline immediately west of the Mesozoic basin. The Erving is locally subdivided into light-gray, *non*carbonaceous quartz-plagioclase-biotite granulite and minor schist (unit Deg), amphibolite (unit Dea), and, near the base, interlayered amphibolite and gray to rusty schist (unit Dev).

The Goshen Formation, entirely west of the Mesozoic basin, has been subdivided into six informal members. The western part of the outcrop belt of the formation consists of three members (units Dg, Dgq, and Dgu). Units Dg and Dgu are characterized by thin beds (5-20 cm) graded from light-gray quartzite to dark-gray, aluminous, graphitic schist. Unit Dgq lies between units Dg and Dgu and is characterized by 15-cm- to 6-m-thick beds of light-gray massive quartzite and quartzose calc-silicate rock. Three other members, units Dgl, Dgc, and Dgp, are present in the northern, central-southern, and southern parts, respectively, of the east part of the Goshen outcrop belt. All consist largely of gray pelitic schist like that of units Dg and Dgu but also contain conspicuous calc-silicate rocks. They are mutually distinguishable by their apparent stratigraphic positions within the formation. East-west and north-south facies changes are believed to characterize the Goshen and to explain some of the geologic complexities.

The Waits River Formation, unit Dw, is characterized by poorly bedded, low-alumina, graphitic mica schist and punky brown-weathering siliceous marble. A member (unit Dwt), distinguished by

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[3]Recently discovered conodonts in the Fitch Formation marble at Bernardston, Mass. (Elbert and others, in press), indicate that it is earliest Devonian at this location.

[4]Recent stratigraphic interpretations in the Monadnock (Thompson, 1985) and Hinsdale (Elbert, 1986), N.H., areas, and remapping in the Mt. Grace, Mass., area (Robinson, 1987) have shown that most of the rocks of area 7 and all of those in area 8 should be assigned to the Lower Silurian Perry Mountain Formation.

relatively thicker siliceous marble beds, is locally mapped in northernmost Massachusetts. Veins of white quartz are widespread throughout the formation. Thin metavolcanic amphibolite present in the Waits River and Gile Mountain near the formation contact is mapped as Dwa or Dgma.

The Gile Mountain Formation (unit Dgm) is characterized by light-gray, micaceous quartzite in beds as much as a few meters thick interbedded with massive, muscovite-biotite schist and small amounts of punky brown-weathering siliceous marble. Member Dgmq is distinguished in the eastern part of the formation by more and thicker beds of quartzite. White quartz veins are common.

The Putney Volcanics form a very thin discontinuous unit of metamorphosed tuff of intermediate composition at the Gile Mountain-Littleton contact in northernmost Massachusetts.

INTRODUCTION

The term "Connecticut Valley belt," as used in this volume and on the State bedrock map (Zen and others, 1983; Hatch and others, 1984), includes the Silurian and Devonian strata in west-central Massachusetts bordered on the west by a major regional unconformity and on the east by an approximately north-south line east of the Bronson Hill anticlinorium (fig. 1). The unconformity along the west border of the distinctive gray schists of the Goshen Formation truncates various units of the underlying Rowe-Hawley zone (Stanley and Hatch, Ch. A, this volume) and is sharp and readily defined. The east border of the Connecticut Valley belt with the adjoining Merrimack belt is more arbitrarily defined as a line on the east side of the first outcrop belt of Littleton Formation east of the main body of Monson Gneiss (fig. 1). East of this line, in the Merrimack belt, the stratigraphic position of the Clough Quartzite is represented by a thicker, off-shelf, more turbidite-rich sequence of time-correlative units.[5] By this definition, the Connecticut Valley belt includes the post-Taconian strata of the Connecticut Valley-Gaspé synclinorium of Cady (1960), plus the post-Taconian strata of the Bronson Hill anticlinorium (Billings, 1956). Although the boundary between the Connecticut Valley and Merrimack belts is loosely defined on the basis of a facies change in the Silurian rocks, the overlying Lower Devonian strata of the Littleton Formation continue essentially unchanged across the boundary. The pre-Silurian strata beneath the Connecticut Valley belt are assigned to the Rowe-Hawley (Stanley and Hatch, Ch. A, this volume) and Bronson Hill zones (see Tectonic map on Zen and others, 1983, and figs. 1 and 2 of Hatch and others, 1984).

[5]Recent work (see footnote 4) shows that this Silurian facies change is telescoped by an early Acadian thrust, the Brennan Hill thrust, that carries Silurian strata of the Merrimack belt westward over Silurian-Devonian strata of the Connecticut Valley belt. This reinterpretation involves reassignment of most of the Littleton of areas 7 and 8 of this paper to the Lower Silurian Rangeley Formation and would place both areas in the Merrimack belt.

The Connecticut Valley belt thus includes the Russell Mountain, Goshen, Waits River, and Gile Mountain Formations and the Putney Volcanics of the classic Connecticut Valley-Gaspé synclinorium on the west, plus the Clough Quartzite and the Fitch, Littleton, and Erving Formations of the Bronson Hill sequence on the east (figs. 1, 2). Although the Mesozoic basins generally separate these two parts of the belt, areas of Littleton Formation are present west of the basins and a few patches of Gile Mountain and Waits River Formations are mapped east of the basins (see State bedrock map). The significance of these areas of Littleton and Gile Mountain-Waits River rocks and the stratigraphic and structural relations between them and the Littleton are thoroughly discussed in another chapter of this volume (Robinson and others, Ch. D). The purpose of the present chapter is to describe the various units that make up the Connecticut Valley belt in Massachusetts.

Because this paper is designed to supplement and clarify the State bedrock map (Zen and others, 1983), we assume in the following discussion that the reader has the map at hand or is at least generally familiar with it. Many of the units shown on the map are lithic members or submembers for which no formal or informal names have been proposed. These units are commonly referred to in this chapter by their letter symbol designations on the map (fig. 2).

Studies in the Connecticut Valley belt previous to those of the present authors and their colleagues and students are relatively few. The principal, most significant work upon which all subsequent studies are based is that of Benjamin K. Emerson. Emerson's numerous topical papers and maps and years of field work were summarized in his study of Old Hampshire County in 1898 and finally in his report and map of the whole state in 1917. Subsequent studies that preceded ours were chiefly those of Kenneth Segerstrom (1956a,b), Max Willard (1956), Robert Balk (1946, 1956), and Jarvis Hadley (1949).

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We are very much indebted to a large number of colleagues without whom the compilation of the Connecticut Valley belt would not have been possible. Stewart F. Clark, Jr., Douglas Knapp, Stephen A. Norton, Philip H. Osberg, Leo M. Hall, Walter Trzcienski, Richard A. Jackson, and Jeffrey Pferd contributed to the geologic mapping west of the Mesozoic basin. We acknowledge their contributions to the mapping and to our thoughts on these rocks over the years.

Contributions to the mapping and description of rock types in the vicinity of the Bronson Hill anticlinorium include work by John D. Peper, Newell J. Trask, Michael T. Field, David Halpin, G.W. Leo, David J. Hall, Richard A. Jasaitis, Stuart Michener, Thomas M. Pike, H. Scott Laird, J. Craig Huntington, R.J. Tracy, and Kurt T. Hollocher, plus a host of students in mapping classes at the University of Massachusetts. Critical advice and commentary in the field were provided by J.B. Thompson, Jr., Marland P. Billings, John D. Rosenfeld, and J. Christopher Hepburn. The work of Robinson during the 1970's was supported by grants from the U.S. Geological Survey and the National Science Foundation Geology and Geochemistry programs. Marie Litterer drafted some of the figures. To each of these institutions and persons we express our grateful acknowledgment.

Finally, this report benefited from thoughtful reviews by Leo M. Hall, Gerhard W. Leo, and Byron D. Stone.

SILURIAN STRATA

The Silurian of the Connecticut Valley belt is represented by three formations: the Russell Mountain Formation, the Clough Quartzite, and the Fitch Formation. Only the Clough Quartzite contains fossils in Massachusetts,[6] but all three formations are dated by Silurian fossils in the same or correlative formations to the north in Vermont or New Hampshire.

RUSSELL MOUNTAIN FORMATION (Sr)

Hatch, Stanley, and Clark (1970) first described a distinctive, thin, discontinuous unit of calc-silicate granofels and quartzite between the Ordovician Cobble Mountain Formation (see Stanley and Hatch, Ch. A,

[6]New information (see footnote 3) indicates that the Fitch at Bernardston, Mass., is earliest Devonian.

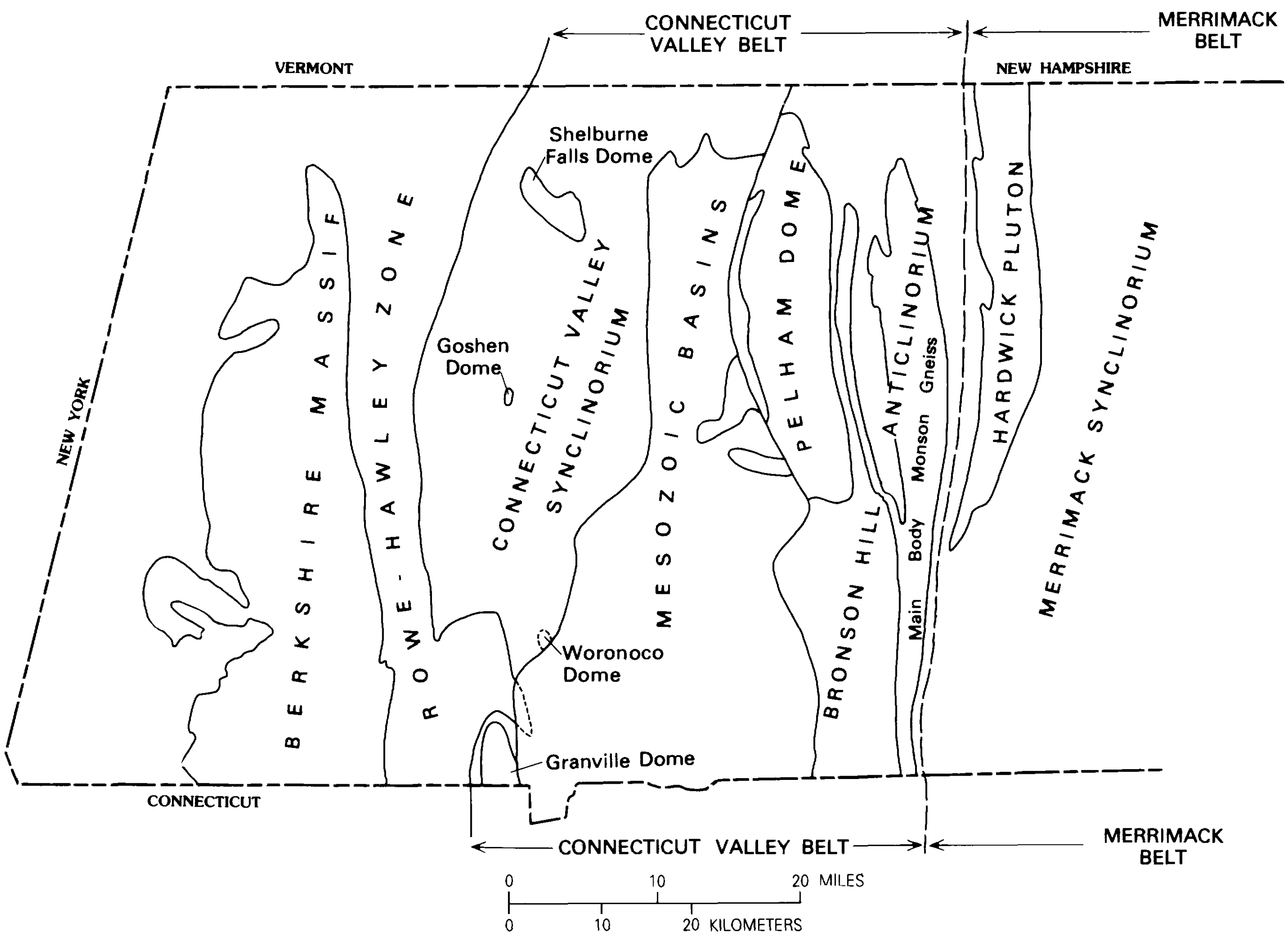

FIGURE 1.—Sketch map of central and western Massachusetts showing the boundaries of the Connecticut Valley belt in relation to major geologic features and the adjoining Merrimack belt.

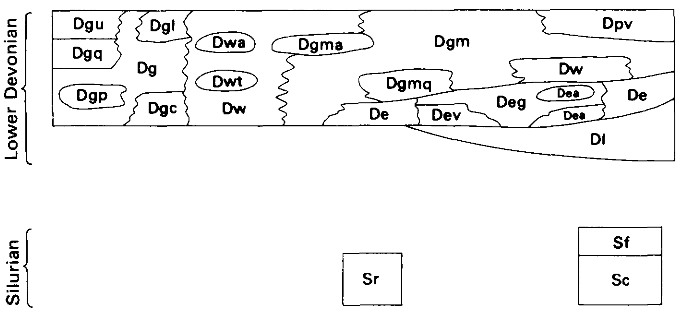

FIGURE 2.—Silurian and Devonian stratified rocks of the Connecticut Valley belt, from the Correlation of map units on the State map (Zen and others, 1983). Dg, Dgu, Dgq, Dgl, Dgp, and Dgc are members of the Goshen Formation. Dw, Dwa, and Dwt are members of the Waits River Formation. Dgm, Dgmq, and Dgma are members of the Gile Mountain Formation. Dpv is the Putney Volcanics. De, Deg, Dev, and Dea are members of the Erving Formation. Dl is the Littleton Formation, Sr is the Russell Mountain Formation, Sf is the Fitch Formation, and Sc is the Clough Quartzite.

this volume) and the Lower Devonian Goshen Formation in southern Massachusetts. Hatch and Stanley (1973, 1975) subsequently traced the Russell Mountain south across western Connecticut. Largely on the basis of a long-range lithologic correlation with the Shaw Mountain Formation in Vermont, this unit was assigned a Middle Silurian age. Because the rocks described here are not continuous with the Shaw Mountain in Vermont, and because of the structural complexities of the region, the name Russell Mountain Formation was given to these rocks in Massachusetts (Hatch, Stanley, and Clark, 1970).

In the original definition of the Russell Mountain (Hatch, Stanley, and Clark, 1970, fig. 1), the calcareous granofels on the Woronoco dome was included (figs. 1, 3, 4). Subsequent mapping (Stanley and others, 1982) has made it clear that these rocks are significantly different from the rest of the Russell Mountain rocks and that they are more logically included in the overlying Lower Devonian sequence (now mapped as unit Dgc of the Goshen Formation, fig. 3). Other than this modification and the finding and mapping of a few more lenses of Russell Mountain rocks just north of the Massachusetts Turnpike (Stanley and others, 1982) and around the northern part of the Granville dome (Knapp, 1977, 1978) (figs. 3, 4), the original definition of the Russell Mountain stands.

In Massachusetts, the Russell Mountain is highly discontinuous and, with the exception of the locality near the Shelburne Falls dome, discussed below (fig. 4), forms thin lenses that have not been found north of the exposures near Blandford and Woronoco (fig. 3) (Hatch and Stanley, 1976). It locally rests upon members A and D of the Cobble Mountain Formation (see Stanley and Hatch, Ch. A, this volume) and is everywhere overlain by rocks of the Goshen Formation.

Hatch (1981) described an "exposure of biotitic quartzite too small to be shown on this map [MF-855] at the contact between the amphibolites of the [pre-Silurian] dome sequence and the Goshen schists 160 m S. 35° E. of the summit of Goodnow Hill" on the southwest flank of the southeast lobe of the Shelburne Falls dome (figs. 3, 4), about 2 km north of Moonstone Hill (see State map). Available data are insufficient to determine whether this quartzite exposure is part of a feather edge or tiny remnant of the Silurian Russell Mountain Formation or Clough Quartzite or is simply a local and unique lens of very sandy material in the base of the overlying Lower Devonian Goshen Formation. No other exposures of comparable rocks have been reported in any of the previous studies of the Shelburne Falls dome (Balk, 1946; Segerstrom, 1956b; Hatch and Hartshorn, 1968; Leo M. Hall, written commun., 1977; Hatch, 1981). Furthermore, no comparable rocks have been reported from the margins of either the Goshen (Hatch and Warren, 1982) or Woronoco (Stanley, Clark, and Hatch, 1982) domes.

The very small area of exposure of the Russell Mountain Formation does not reflect its importance as a stratigraphic and structural marker bed. It is not known to exceed 35 m in thickness in Massachusetts, although its correlative in Connecticut, designated the basal member of The Straits Schist by Rodgers (1982, 1985), probably is locally at least twice that thick (Hatch and Stanley, 1973, 1975). Because of its distinctive lithology, however, it was mapped under a variety of names by many geologists in the eastern part of western Connecticut long before its presently interpreted stratigraphic position and correlation were proposed (see discussion in Hatch and Stanley, 1973, p. 17-57).

The Russell Mountain Formation is characterized by calc-silicate granulites and quartzites. The calc-silicate rocks are generally well layered, greenish gray, and medium grained and consist of varied proportions of quartz, epidote, feldspar, diopside, tremolite, sphene, calcite, and scapolite. Many exposures in Connecticut consist primarily of very dark-green amphibolite. The quartzites of the Russell Mountain are relatively clean and vitreous to micaceous, are locally conglomeratic (fig. 5), and have indistinct internal laminations. No consistent internal stratigraphic sequence within the Russell Mountain has been recognized that can be correlated with the sequence of lithologies in the Clough and Fitch to the east.

The Russell Mountain derives its age assignment from correlation with discontinuous lenses of similar

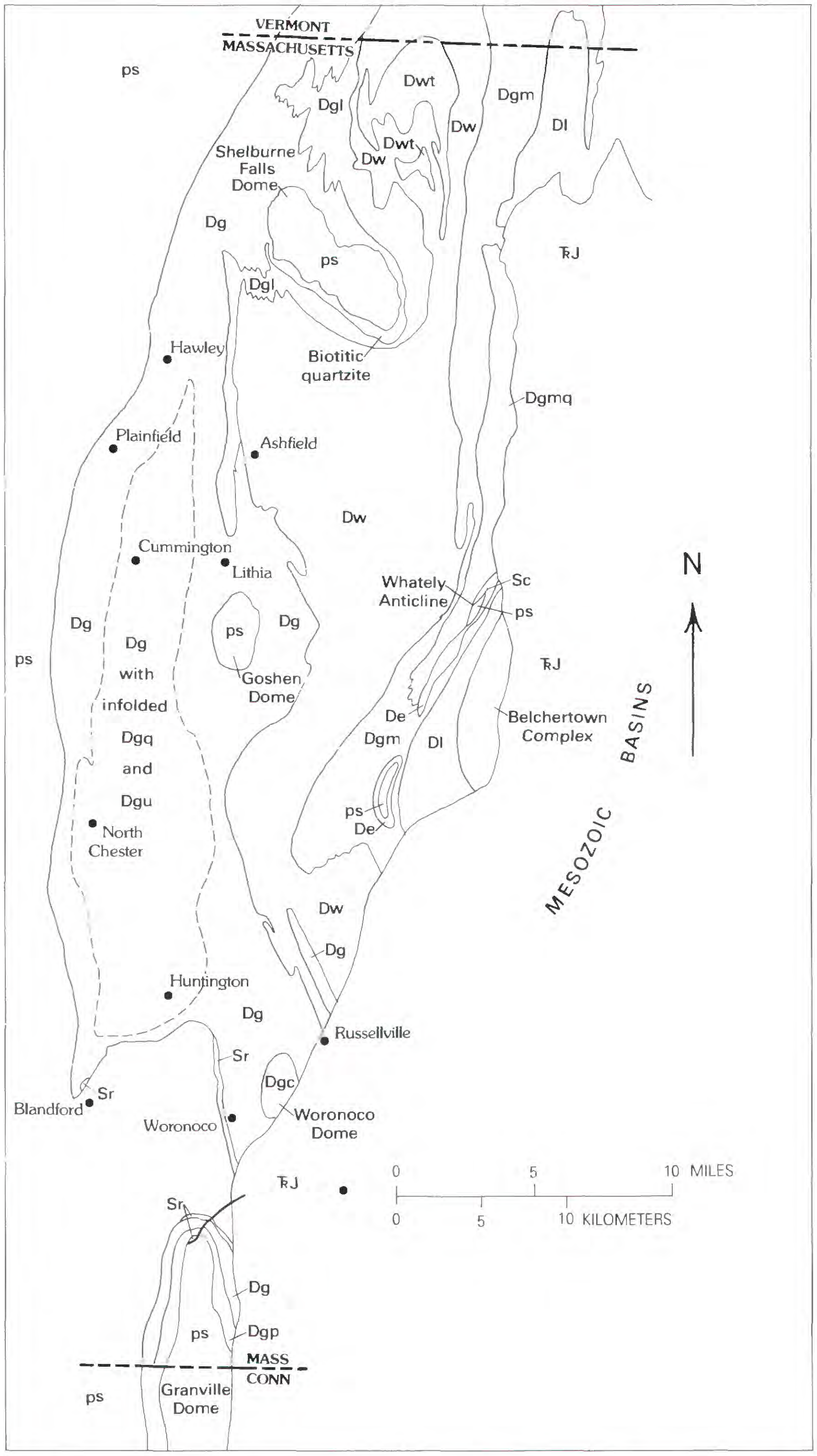

FIGURE 3.—Simplified map of the western part of the Connecticut Valley belt west of the Mesozoic basins showing approximate distribution of Silurian and Devonian strata. Outcrop areas of Russell Mountain Formation (Sr) exaggerated in order to show up on figure. Single outcrop area of Clough Quartzite (Sc) in east-central part of figure indicated only by leader. Other symbols are as follows: Dg, Dgl, Dgq, Dgu, Dgp, and Dgc are members of the Goshen Formation; Dw and Dwt are members of Waits River Formation; Dgm and Dgmq are members of the Gile Mountain Formation; De is Erving Formation; Dl is Littleton Formation; pS is all pre-Silurian strata undifferentiated; ꞦJ is Triassic-Jurassic strata. Intrusive rocks other than Belchertown Complex not shown.

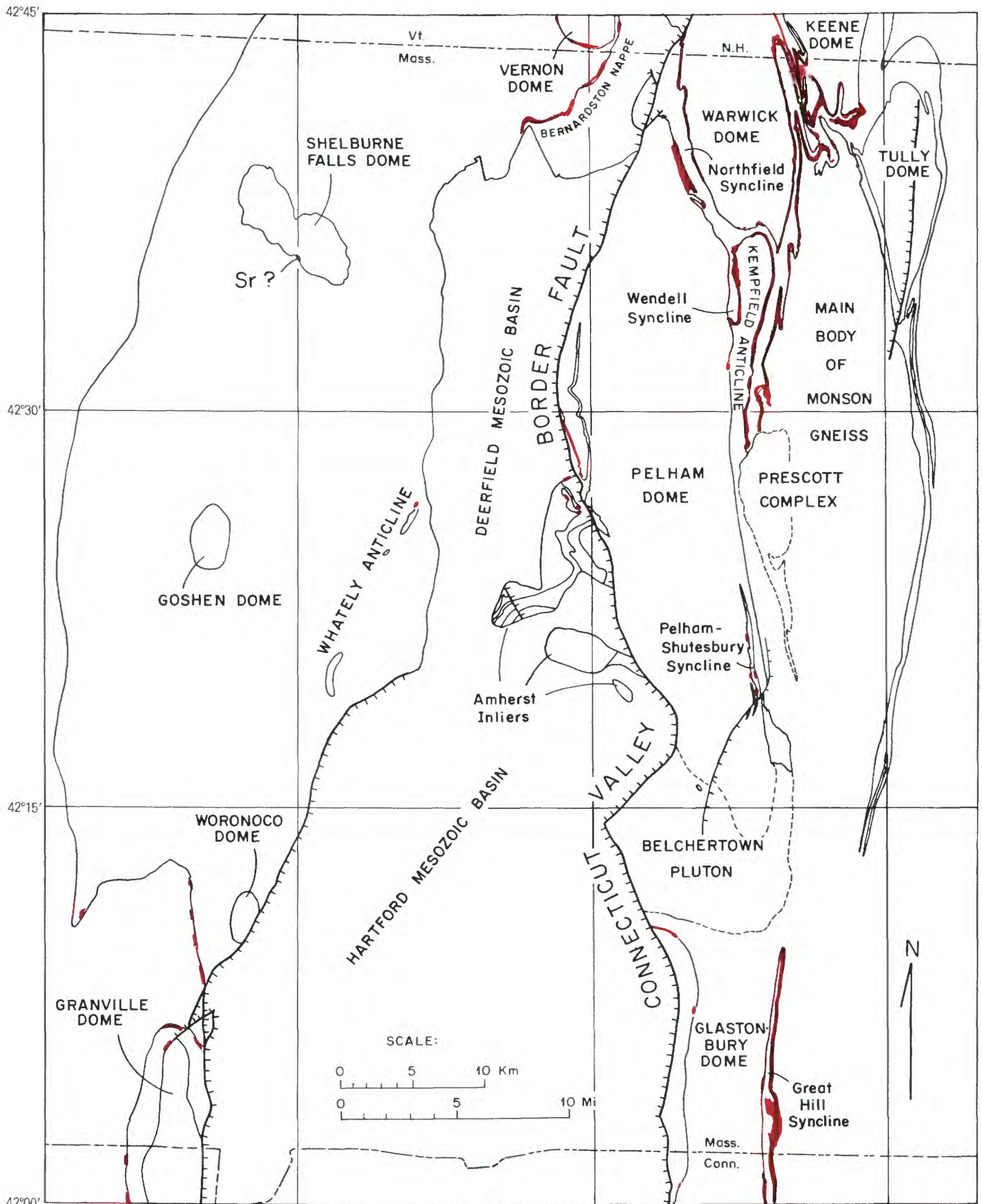

FIGURE 4.—Distribution of Silurian Russell Mountain Formation (Sr) and Clough Quartzite (Sc) in red in the Connecticut Valley belt. For distribution of Silurian Fitch Formation, see figure 6.

FIGURE 5.—Pebbly conglomerate in the Russell Mountain Formation, in the center of the fairground racetrack, Blandford. Exposed part of pocket knife is about 3 cm long.

rocks at the same stratigraphic position in Vermont assigned to the Shaw Mountain Formation. Recent field trips in southern Vermont with James B. Thompson, Jr., and others have raised questions in our minds about exactly how many of the rocks presently mapped as Shaw Mountain should, in fact, be considered part of that formation. However, a late Llandoverian to Gedinnian age assigned to the Shaw Mountain somewhat further north near Albany, Vt., on the basis of *Howellella* is firmly established (Boucot and Thompson, 1963, p. 1318; Konig, 1961, p. 32).

The lithology and the relative thinness of the Russell Mountain Formation suggest a continental-shelf environment of deposition. Although we generally correlate the Russell Mountain with the Clough Quartzite and Fitch Formation to the east, a continuous shelf at the east edge of post-Taconian North America between the present outcrop belts of the Russell Mountain and the Clough-Fitch can only be inferred. The near absence of recognized Russell Mountain-Clough-Fitch around the margins of the Shelburne Falls, Goshen, and Woronoco domes could equally well be explained by pre-Goshen erosion as by nondeposition.

CLOUGH QUARTZITE (Sc)

The Clough Quartzite (Billings, 1956) was first defined by Billings (1937) in the Littleton-Moosilauke area, New Hampshire, as the Clough Conglomerate. It was traced thence southward by Billings (1956) and students as far as the Massachusetts border (Moore, 1949) and was first extended into Massachusetts in the Mount Grace quadrangle east of the Warwick dome (fig. 4) by Hadley (1949). Robinson (1963, 1967a,b) and Peper (1966, 1967) traced the unit across Massachusetts and into Connecticut, where it was eventually adopted on the basis of mapping by Rosenfeld and Eaton (1958), Eaton and Rosenfeld (1960), and Snyder (1970). It was recognized in the Whately anticline west of the Mesozoic basin (fig. 4) by Walter Trzcienski (written commun., 1967).

The Clough Quartzite is the key stratigraphic unit in the Bronson Hill anticlinorium for three reasons (Thompson and others, 1968): (1) It is dominated by distinctive, readily recognized rock types. (2) Where present it is at the base of the Silurian-Devonian sequence, resting with detectable unconformity on older rocks. (3) It contains fossils assigned to the late Llandoverian at several localities in western New Hampshire and adjacent Vermont (Boucot and Thompson, 1958, 1963) and also at Bernardston, Mass. (Boucot and others, 1958).

The Clough Quartzite is irregularly distributed and shows wide variations in thickness and dominant rock type over relatively short distances (see, for example, Robinson, 1963). On the Massachusetts bedrock map the thickness of the Clough Quartzite is locally exaggerated because at many localities it is only a few meters or less thick and could not be shown at all if portrayed to scale at 1:250,000. The maximum measured thickness in northern Massachusetts is about 200 m on the west limb of the Northfield syncline (fig. 4), although this figure may be exaggerated by recumbent folding. Elsewhere along strike in the same area the Clough pinches out completely. In other areas, as for example east of the Warwick dome (figs. 4, 6), the Clough occurs as a continuous thin layer averaging about 10 m in thickness. In still other areas, the Clough occurs along the contact between underlying and overlying units as a series of lenses or boudins, some of which are no more than 1 m thick. Examples of extreme attenuation are in the Pelham-Shutesbury syncline (fig. 4) on an island in the western part of Quabbin Reservoir (Robinson, 1967b), where at one point the Clough, including three "mappable" (scale of 1:120) subunits, is 0.3 m thick and nearby is mappable at outcrop scale where only 5 cm thick!

The most common and most diagnostic rock type in the Clough is quartz-pebble conglomerate, in which the pebbles are typically deformed into the shapes of cigars, swords, or pancakes, making recognition as a conglomerate difficult for the uninitiated. More than

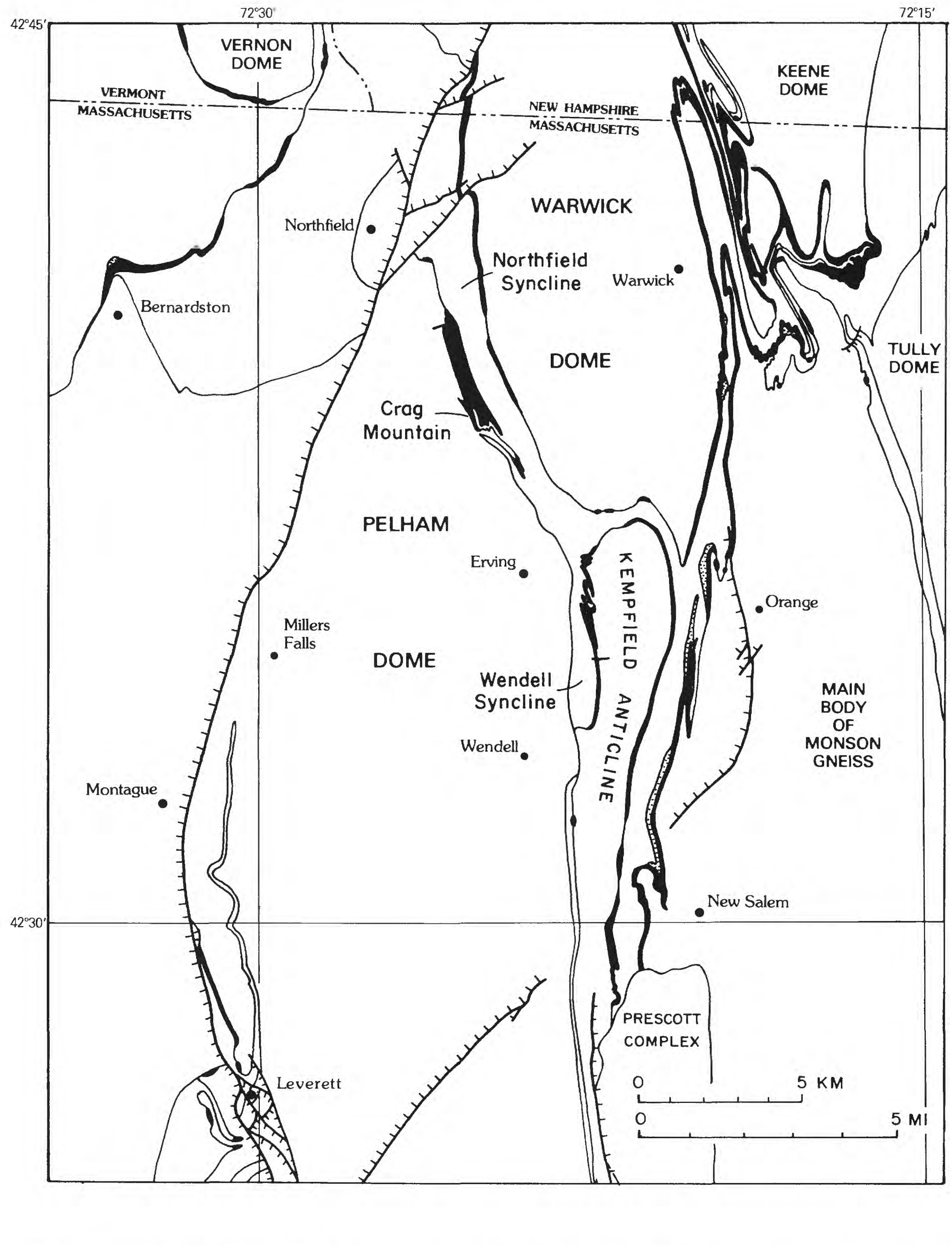

FIGURE 6.—Distribution of Silurian Fitch Formation (stippled) and adjacent Clough Quartzite (black) in the Connecticut Valley belt east of the Mesozoic basins.

95 percent of the pebbles are composed of coarse-grained metamorphic vein quartz. A few pebbles of black quartz-tourmaline vein material and fine-grained quartzite have also been identified (Robinson, 1979, p. 168), but conglomerate containing other rock types is extremely rare. The minerals that occur in the matrix of the conglomerate or in thin schist beds depend on metamorphic grade but include abundant quartz, muscovite, garnet, biotite, and sillimanite and less abundant staurolite, tourmaline, and magnetite, the last two possibly as detrital grains. In some locations, the conglomerate grades to coarse cobble or boulder conglomerate (fig. 7), in which the clasts are of the same materials, but the matrix is generally more micaceous. Such cobble conglomerates are particularly noticeable along the base of the Clough near the southwestern tip of the Keene gneiss dome. Elsewhere, the Clough ranges to quartz grit or well-bedded white to pink quartzite (fig. 8), commonly with very thin beds of mica schist. In a few areas, the Clough contains beds of brown- to gray-weathering mica schist several meters thick. The gray-weathering schists are hard to distinguish from schists of the Devonian Littleton Formation, especially in areas where the Littleton itself contains beds of quartzite. In a few areas, the top of the Clough is characterized by a thin zone of well-bedded calc-silicate rock, commonly dominated by hornblende with various combinations of epidote, clinozoisite, or zoisite (Robinson, 1963, p. 56; 1967b, Stop 1). This calc-silicate is distinguishable from those of the overlying Fitch Formation by being less feldspathic and may be equivalent to the fossiliferous horizon in the Clough described in New Hampshire (Robinson, Thompson, and Rosenfeld, 1979, p. 111-113).

FIGURE 7.—Deformed cobble conglomerate at base of the Clough Quartzite in contact with sulfidic schist of the Partridge Formation (left). Hammer handle lies on contact. Contact is along east limb of isoclinal syncline of Clough near the southwest extremity of the Keene dome in the woods a few hundred meters south of the powerline. Long axes of cobbles plunge about 50° south almost directly away from viewer.

The unconformable nature of the base of the Clough Quartzite (Billings, 1937) is well demonstrated in Massachusetts by regional contact relations and also locally by truncation of older rocks at individual outcrops, although the angle at which the base of the Clough cuts contacts between older units rarely exceeds 20° (Robinson, 1963, p. 61). In the Pelham dome (fig. 4; see also Robinson and others, Ch. D, this volume, fig. 3), the Clough rests directly on the Fourmile Gneiss (OZfm) over a distance of several miles. This contact was recognized as a probable unconformity by Emerson (1898b). The Clough also rests on Fourmile Gneiss in the Kempfield anticline (fig. 4), though here contact relations may be complicated by Mesozoic faulting. The Clough directly overlies the Ammonoosuc Volcanics (Oa) over most of the gneiss domes. Generally, it is in contact with the felsic upper member of the Ammonoosuc, but locally on the west side of the Warwick dome (fig. 4) (J.C. Schumacher, oral commun., 1979) it cuts down onto the mafic lower member. In many areas, for example the east side of the Warwick dome, the Partridge Formation (Ops) occurs discontinuously along the Clough-Ammonoosuc contact. Such lenses of Partridge are only a few tens of meters thick at most. These observations imply that although pre-Clough erosion was important regionally, the local amplitude of pre-Clough folding was very low. In the eastern part of the Bronson Hill anticlinorium, particularly around the Monson Gneiss, in the nappes at Bernardston and Amherst,[7] and on the west limb of the Pelham dome, the base of the Clough rests on rocks still higher in the

[7]The four mapped areas of Clough in the Amherst area west of the Connecticut Valley border fault have recently been reinterpreted as conglomerate lenses in the Lower Silurian Rangeley Formation (see footnote 4).

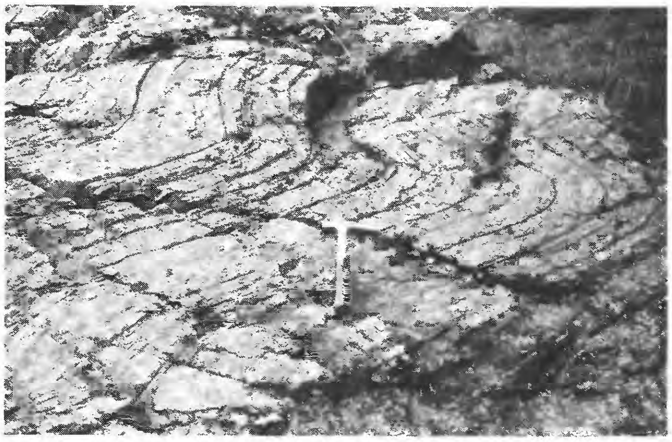

FIGURE 8.—Fine-grained, well-bedded Clough Quartzite. Outcrop surface dips about 45° south and prominent dome-stage folds plunge 30° north (away from viewer). Exposure on steep south face of hill north of Route 2 on the west side of the Kempfield anticline.

pre-Silurian stratigraphy. In these places it rests on the lower few hundred meters of the Partridge Formation characterized by abundant volcanics (Opsa), on the augen gneiss member (Opau), or on the volcanic-free mica schist facies of the Partridge (Ops). The Clough is extremely thin and attenuated near the north end of the main body of Monson Gneiss (fig. 6) and pinches out for the last time in Massachusetts (not recognized anywhere to the east) at a contact between the overlying Littleton Formation (Dl) and the augen gneiss member of the Partridge (Opau). The Clough Quartzite may connect with the Russell Mountain Formation as a discontinuous series of lenses, somewhere beneath younger cover to the west of the Whately anticline (fig. 4).

Several factors must be taken into account in interpreting the environment of deposition of the Clough Quartzite. A major rock type is pebble, cobble, and locally boulder conglomerate, the clasts of which are dominantly vein quartz set in a matrix rich in aluminous minerals. This conglomerate commonly rests unconformably on metamorphosed plutonic and volcanic rocks and shales in which vein quartz is present only in subordinate amounts, yet the coarse size of the conglomerate clasts suggests local derivation. One possibility is that the conglomerate was derived by wave or stream erosion on a surface that was undergoing such strong chemical weathering that only vein quartz remained intact. This model would account for both the monotony of the clasts and the aluminous character of the matrix. Less easily explained by this model, however, is the fact that the probably generally correlative quartzites of the Russell Mountain Formation to the west (the presumed source direction) only locally contain any conglomerates. Whether the basal part of the Clough was deposited in a terrestrial or marine environment is uncertain, but the uppermost calcareous rocks containing Early Silurian brachiopods reported by Boucot and others (1958) and Boucot and Thompson (1958, 1963) from southwestern New Hampshire, and probably also the finer sands, were deposited in a nearshore marine environment. When this proposed environment for the Clough is compared with the probable environments for the thick section of rocks of equivalent age to the east in the Merrimack synclinorium, it becomes clear that the environment was that of the then east coast of North America, as it existed after the Taconian and before the Acadian orogeny.

FITCH FORMATION (Sf)

In the Bronson Hill anticlinorium in Massachusetts, the Fitch Formation is represented only by a few lenses between the Clough Quartzite and the Littleton Formation (fig. 6). These are mainly on the east side of the Warwick dome and on the west side of the main body of Monson Gneiss, north of Quabbin Reservoir. Despite its limited occurrence, the Fitch Formation is very important in this region for three reasons: (1) It forms a unique member of a four-unit stratigraphic succession, Partridge-Clough-Fitch-Littleton, that has been extensively documented and mapped farther north and that is the key to many structural interpretations. (2) It is lithically and stratigraphically identical with Fitch strata in western New Hampshire that have yielded Late Silurian brachiopods (Billings and Cleaves, 1934; Berry and Boucot, 1970) and most recently latest Silurian conodonts and brachiopods (Harris and others, 1983).[8] (3) It is apparently the only unit in the Connecticut Valley belt deposited during the latter part of the Silurian, a time during which thick strata were being extensively deposited in the area of the Merrimack synclinorium to the east.

The most common rock type of the Fitch Formation in Massachusetts (Robinson, 1963) is gray, massive to weakly bedded, quartz-labradorite-biotite granulite containing a moderate amount of some combination of the calc-silicate minerals calcic amphibole, zoisite or clinozoisite, diopside, sphene, and, generally, microcline (fig. 9). These granulites are commonly interbedded with biotite-free granulites dominated by the same calc-silicate minerals plus variable amounts of calcite, grossular, and (or) calcic scapolite. One small exposure is nearly pure calcite marble with minor quartz (D.D. Ashenden, oral commun., 1978).

[8] See footnote 3.

FIGURE 9.—Fine-grained, poorly layered, gray-weathering feldspar calc-silicate granulite of the Fitch Formation. Behind gravel pit northeast of junction of Routes 2A and 78 west of Orange.

Additional rock types found in the larger lenses are rusty-weathering sulfidic-graphitic rocks including actinolite-biotite schist, labradorite-biotite schist, and pelitic biotite schist containing any or all of the minerals muscovite, sillimanite, staurolite, or garnet. These pelitic schists are lithically similar to schists of the Middle Ordovician Partridge Formation, but where best exposed, as, for example, on Nielson Road northwest of New Salem (Robinson, 1963) (about 1 km west of Morse Village on the State bedrock map), they occur between typical Fitch granulites and the gray mica schists of the overlying Littleton Formation and thus cannot be Partridge.

An isolated lens of Fitch Formation occurs stratigraphically above and structurally below the fossil locality in the Clough Quartzite north of Bernardston (fig. 6) in the garnet zone of regional metamorphism. The dominant rock of the Fitch here is tough calcite-garnet-biotite granulite that locally contains magnetite (Trask and Thompson, 1967).[9]

Another isolated lens of Fitch Formation, too small to show on the State bedrock map, occurs above the Clough Quartzite midway between the village of Warwick and the Tully dome (fig. 6). The rock in this lens is hard, very rusty-weathering, graphitic bytownite-hornblende-diopside-zoisite-quartz-biotite-sphene-pyrrhotite granulite that bears some resemblance to the Fitch Formation (Francestown type) in the western part of the Merrimack belt.

The best exposures of the Fitch Formation are in low hills west of the village of Orange (fig. 6), northeast of the junction of Routes 2A and 78, where the formation is 250 m thick. As presently mapped (State bedrock map and Robinson, 1979; but not Robinson, 1963, 1967a, 1977), the Fitch Formation everywhere overlies the Clough Quartzite and is never in contact with the Middle Ordovician Partridge Formation. This relationship suggests that the Fitch Formation may be the upper part of the same depositional sequence as the Clough Quartzite. It should be pointed out that the fossils dating the Fitch as Pridolian (Harris and others, 1983) are all from the Littleton, N.H., area, 200 km north of the Massachusetts border, whereas the fossils dating the Clough as late Llandoverian (Boucot and Thompson, 1958, 1963) are all from localities at least 100 km south-southwest of Littleton. Without additional fossil data, it is impossible to say whether or not either unit is time transgressive and thus whether or not they form a continuous depositional sequence in central Massachusetts. The discontinuous distribution of the Fitch Formation could, in part, be due to structural dislocations but is much more likely to be due to erosion before deposition of the Lower Devonian Littleton Formation. The Clough Quartzite is also cut out over extensive areas so that the Littleton Formation rests on the Partridge Formation or even older units, further confirming the postulated pre-Littleton erosion. A comparable uppermost Silurian-Lower Devonian unconformity has been described along strike in northern Maine (Roy and Mencher, 1976) and assigned to a "Salinic" disturbance.[10]

The Fitch Formation is interpreted as a metamorphosed sequence of calcareous dolomitic siltstone and shale containing local calcitic limestone and interbedded sulfidic, calcareous, and more aluminous shale. The environment of deposition was probably relatively near shore but away from sources of coarse clastic

[9]Also fossiliferous marble; see footnote 3.

[10]The new earliest Devonian fossils from the Fitch at Bernardston (see footnote 3) suggest that pre-Littleton erosion took place within the Early Devonian.

sediments and in regions of restricted circulation that produced at least local reducing environments.

DEVONIAN STRATA

Stratigraphically above the thin Silurian near-shore rocks of the Connecticut Valley belt is a variable thickness of predominantly gray metamorphosed flysch and very minor metamorphosed volcanic rocks. These rocks are all assigned with varying degrees of certainty to the Early Devonian. In the main belt of the Connecticut Valley synclinorium, between the outcrop belt of the Hawley and Cobble Mountain Formations (Stanley and Hatch, Ch. A, this volume) on the west and the Mesozoic basin on the east, these Devonian strata are divided into various informal members of the Erving, Goshen, Waits River, Gile Mountain, and Littleton Formations and the Putney Volcanics. East of the Mesozoic basin, in the area of the Bronson Hill anticlinorium, the Devonian strata are divided into members of the Littleton, Erving, and Waits River Formations. The Gile Mountain Formation occurs in small fault slices.

The ages of the rocks in the Connecticut Valley-Gaspé synclinorium of western New England have been variously interpreted. Emerson (1898b, 1917) considered these rocks in Massachusetts to be Silurian. Richardson (1916, 1919) reported numerous beds of slates from this sequence in Vermont to contain graptolites identified as Ordovician by Ruedemann. None of these reported occurrences are currently accepted as containing identifiable graptolites. Doll subsequently reported poorly preserved echinoderms (1943a) and a brachiopod (1943b) from this sequence in Vermont, to which an age of Middle Silurian to Early Devonian was ascribed. In 1950, Cady reported cup corals of probable Middle to Late Ordovician age from the Waits River Formation near Montpelier, Vt. This report apparently influenced Billings (1956) to assign an Ordovician age to the Meetinghouse-Gile Mountain-Waits River sequence in westernmost New Hampshire. However, in 1956 Cady, with no reference to his own 1950 coral paper, assigned a Silurian(?) to Devonian age to the Waits River and associated Northfield and Gile Mountain Formations in the Montpelier, Vt., area. He based this age assignment on the apparent stratigraphic position of these units above the (at that time not known to be fossiliferous) Shaw Mountain, which he correlated with the Peasley Pond Conglomerate, presumed to be Silurian, in southernmost Quebec (W.M. Cady, oral commun., 1979). Most subsequent reports and maps, including the Centennial Geologic Map of Vermont (Doll and others, 1961), have assigned these rocks to the Silurian and (or) Early Devonian, although many geologists have questioned the reliability or authenticity of the reported fossils. Restudy by William A. Oliver, Jr. (written commun., 12/3/79), of material collected by Hatch from Cady's (1950) locality 7 indicated that "none of the specimens at hand appear to be corals, and none have any markings or structure that suggest an organic origin, although this is a possible origin of the objects. Some specimens illustrated by Cady (1950, pl. 2) are more coral-like (as photographed) than the new specimens; others are very similar. I do not think it is possible to prove that the objects are not corals but I conclude that they are probably not, and that if they are, they are indeterminate and could be of any post-Early Ordovician age." Most recently, Doll (1984) has summarized a number of localities of possible fossils from the Waits River and Gile Mountain Formations in northern Vermont. Although he interprets them all to be Silurian or Devonian in age, a formal verdict on whether or not they are bona fide fossils and identifiable as to age is still pending. Approximately 20 samples collected by Hatch, Robinson, Leo M. Hall, and Anita Harris from carbonate beds in the lowest metamorphic grades throughout the synclinorium in Vermont and northernmost Massachusetts failed to produce conodonts or any other fossils.

Because few, if any, authenticated fossils are present in the units themselves, dating of this package of rocks of the Connecticut Valley synclinorium is based on the following: (1) their position east of and presumably stratigraphically above the fossiliferous (Boucot and Thompson, 1963) Silurian Shaw Mountain in Vermont and its unfossiliferous correlative, the Russell Mountain, in Massachusetts; (2) general correlation of this broad belt of rocks with its apparent continuation in southern Quebec where Devonian fossils have been found (see Boucot, 1968, p. 92, for summary); and (3) their general lithic similarity to rocks of the Littleton and Seboomook Formations of New Hampshire (Boucot and Arndt, 1960; Billings and Cleaves, 1934) and Maine (Boucot, 1969), respectively, both of which contain Early Devonian fossils. Recognizing the tenuous nature of the evidence, and the fact that the ages assigned to all of these units except the Littleton are dependent upon their stratigraphic relations to the fossiliferous Littleton (see Robinson and others, Ch. D, this volume, for more on this problem), we herein consider the Erving, Goshen, Waits River, and Gile Mountain Formations all to be Lower Devonian.

LITTLETON FORMATION (Dl)

The Littleton Formation has been traced southward from the type area near Littleton, N.H. (Ross, 1923; Billings, 1937), to the southern border of New

Hampshire (Billings, 1956; Moore, 1949). The name was first used in Massachusetts by Hadley (1949). The Littleton has been traced across Massachusetts along the Bronson Hill anticlinorium by Robinson (1963, 1967a, 1977), Trask (1964), and Peper (1966, 1967) and into central Connecticut (Rosenfeld and Eaton, 1958; Eaton and Rosenfeld, 1960; Snyder, 1970). The formation has also been extended eastward from the Bronson Hill anticlinorium into the Merrimack synclinorium in New Hampshire (Billings, 1956; Fowler-Billings, 1949; Greene, 1970) and, on somewhat different grounds, in Massachusetts (Field, 1975; Tucker, 1977).[11]

The Littleton Formation is critical to the geologic interpretation of west-central Massachusetts for four reasons: (1) It constitutes the most widespread unit of the Silurian-Devonian sequence in the eastern part of the Bronson Hill anticlinorium. (2) Over large areas it is the youngest stratified unit involved in intense deformation and regional metamorphism, thus setting a lower limit on the age of the Acadian orogeny. (3) It has been traced into Massachusetts from areas in New Hampshire containing Early Devonian fossils (Billings and Cleaves, 1934, 1935; Billings, 1937; Boucot and Arndt, 1960) that indicate correlation with the lower Onondaga or Schoharie of the New York State section and the Emsian of Europe. (4) It is interpreted to be a marine sandstone-shale unit that was the initial clastic deposit derived from tectonic land to the east (Hall and others, 1976), marking a sharp change in the style of sedimentation and the very beginning of the Acadian orogeny.

Although the map pattern is highly convoluted, the Littleton Formation of the Connecticut Valley belt can be described as being present in eight main areas (fig. 10) as follows: (1) west of the Connecticut Valley border fault and north and west of the Deerfield Mesozoic basin in northernmost Massachusetts; (2) near Whately, west of the Connecticut Valley Mesozoic basins and west of the Hatfield pluton; (3) as discontinuous lenses along the northeast flank of the Pelham dome and on the southeast side of the Wendell syncline; (4) in the Great Hill syncline in the southern part of the State just east of the Glastonbury dome; (5) in an extensive convoluted area that completely encircles the Warwick dome as well as the west and south flanks of the Keene dome as far as a crucial early fold hinge southeast of the south end of the Keene dome, and south to the north contact of the Prescott Complex; (6) within the deeply infolded Pelham-Shutesbury syncline completely surrounded by gneisses of the Pelham dome; (7) in a complex belt that enters the State from New Hampshire east of the Keene gneiss dome, encircles the Tully dome and the north end of the main body of Monson Gneiss, and extends down the east side of the Monson Gneiss to where it apparently hinges out southwest of Ware;[12] and (8) complexly interfolded with the Partridge Formation in the Amherst inliers.[13] Rock types and their contact relations within area 7 are crucial to eastward correlations; rocks mapped as Littleton Formation east of area 7 are included in the Merrimack belt.

Each of the areas of Littleton Formation requires some separate description because of local variations in rock types and because of significant differences in the grade of Acadian metamorphism. In brief, area 1 extends from the chlorite zone to the kyanite-staurolite zone; area 2 from the garnet zone through andalusite-staurolite and locally to sillimanite along the west contact of the Belchertown Complex; areas 3, 4, and 6 are in the kyanite-staurolite zone; area 5 is in the kyanite-staurolite and sillimanite-staurolite zones; area 7 is in the sillimanite-staurolite through sillimanite-K-feldspar zones; and area 8 is in the sillimanite-muscovite-K-feldspar zone.

The rocks in the western part of area 1, north of the Deerfield Mesozoic basin, are the least metamorphosed of the Littleton Formation in Massachusetts. They were originally mapped by Emerson (1917) as Leyden Argillite and were considered to be Silurian and immediately above the "Conway Schist" (primarily the Waits River and Gile Mountain of this report). Balk (1956) retained the name Leyden as well as the Silurian age and subdivided the formation into lower and upper members. He distinguished the members by a greater abundance of quartzite beds in the lower (western) member, a distinction that we were unable to recognize as mappable. It should be noted, however, that Moore (1949) mapped these same rocks as Littleton Formation immediately north of the State line in southeasternmost Vermont.

The Littleton rocks in the lowest grade part of area 1 are medium-gray slates and phyllites consisting of quartz, albite, white mica, chlorite, ilmenite, and graphite. The ilmenite occurs as conspicuous platelets that are commonly distorted and replaced by leucoxene. Although bedding is generally inconspicuous or invisible, light-gray silty quartzite locally forms beds 1 mm to a few centimeters thick, many of which preserve excellent grading. Despite the low metamorphic grade,

[11] Much of these strata are now reassigned to the Lower Silurian Rangeley Formation (see footnote 4).

[12] Most of the rocks of this area are now reassigned to the Lower Silurian Rangeley Formation or other units of the Monadnock, N.H., sequence (Thompson, 1985), but some true Littleton remains.

[13] These rocks and those shown as Partridge in the same area are now all reassigned to the Lower Silurian Rangeley Formation.

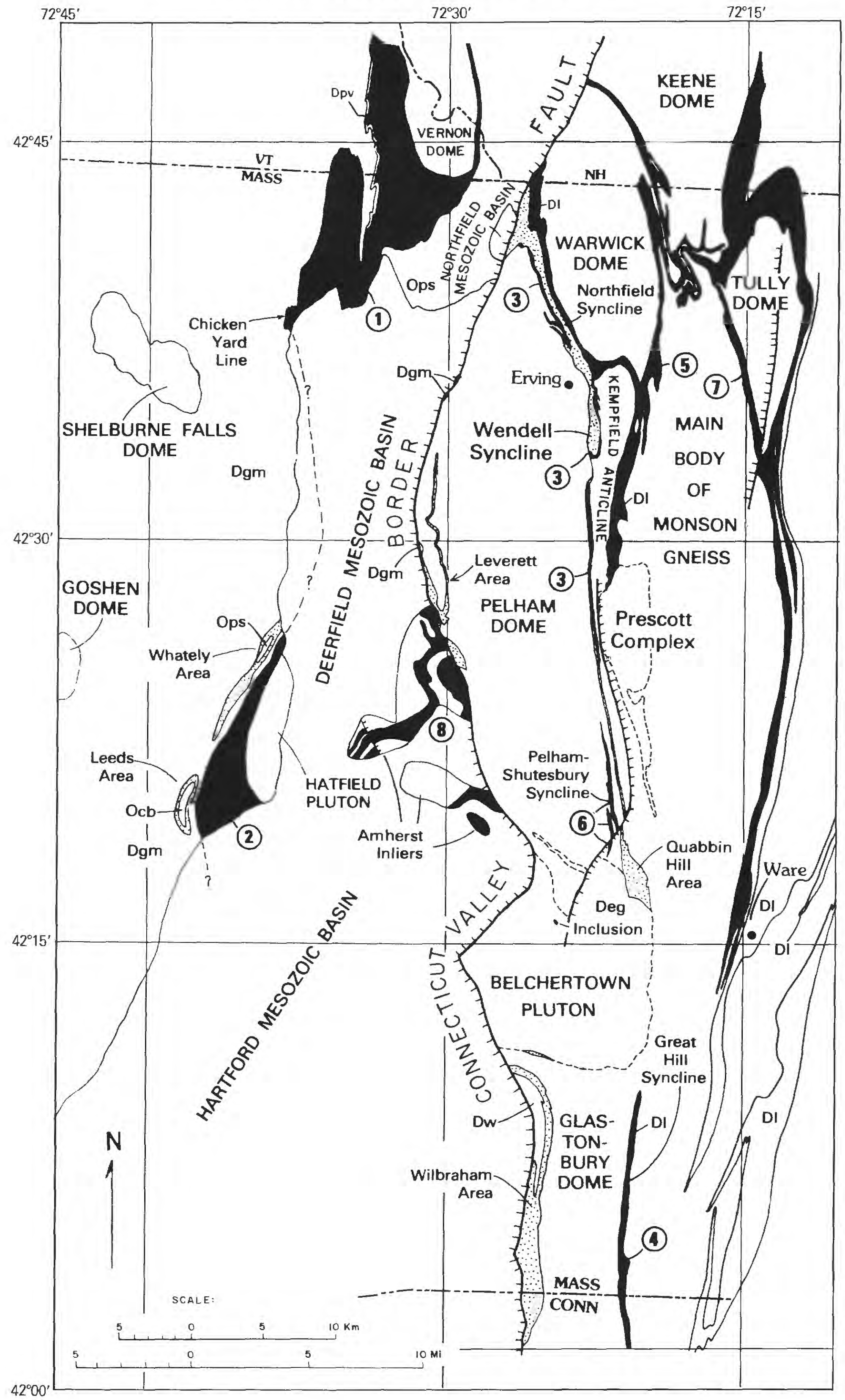

FIGURE 10.—Distribution of Littleton (black) and Erving (stippled) Formations in the Connecticut Valley belt of west-central Massachusetts. Numbers refer to eight areas of Littleton Formation described in text. For more details on distribution of the Erving Formation see figures in Robinson and others (Ch. D, this volume).

most exposures are characteristically complexly deformed by at least two sets of moderately tight folds. This low-grade Littleton is identical with the low-grade slates and phyllites, containing local thin graded quartzite beds, of the lower part of the Littleton Formation at Slate Ledge northwest of the Ammonoosuc fault in the Littleton, N.H., area (Billings, 1937), and with low-grade gray slate and phyllite and local thin graded quartzite beds of the Meetinghouse Slate at the east margin of the Connecticut Valley-Gaspé synclinorium in eastern Vermont (Eric and Dennis, 1958). As metamorphic grade increases eastward across area 1, randomly oriented porphyroblasts of biotite and garnet appear in the pelitic rocks. In the kyanite-staurolite zone in the eastern part of area 1, many outcrops of aluminous beds are studded with randomly oriented spongy staurolite porphyroblasts 3-5 cm long, commonly with 90° twins, set in a fine-grained weakly foliated matrix including biotite, muscovite, pinhead-sized garnets, and minor magnesian chlorite.

In the central and western parts of area 1, the Littleton is in contact with the Gile Mountain Formation. Where the Putney Volcanics are absent, the Littleton and Gile Mountain are distinguished on the basis of the following criteria: (1) The Gile Mountain has quartzite beds of metamorphosed sandstone that range in thickness from about 1 cm to about 1 m; quartzite beds in the Littleton are metamorphosed siltstone and generally range in thickness from 1 mm to a few centimeters (fig. 11). (2) Beds of punky brown-weathering siliceous carbonate are common in the Gile Mountain but virtually absent from the Littleton. (3) Lenses of vein quartz a few centimeters to about 10 cm thick and a few tens of centimeters long, generally conformable to the dominant schistosity, are very common in the Gile Mountain and very rare in the Littleton. Where the Putney Volcanics are present, they occur at the Gile Mountain-Littleton contact as independently defined by these three criteria.

The Littleton Formation in the Whately area (area 2) consists of thin-bedded carbonaceous quartz-mica schist similar to schist in area 1, generally containing tiny garnets. As is the case with the Littleton in area 1, Emerson (1917) assigned these rocks to his Silurian Leyden Argillite. Willard (1956) also applied the name Leyden to these rocks and assigned them to the Silurian. Stoeck (1971), in a report that synthesized mapping by herself, Walter Trzcienski, and P.C. Bazakas, called these rocks Devonian Whately Schist. We agree with the earlier interpretations that these rocks were continuous, when deposited, with the area 1 rocks, and thus assign them to the Lower Devonian Littleton Formation.

The Littleton of the Whately area is generally medium- to dark-gray, fine-grained phyllite intercalated with beds of light-gray, fine-grained quartzite a few millimeters to a few centimeters thick. The phyllite, which is locally finely crenulated, is composed primarily of quartz, muscovite, biotite, garnet, staurolite (at appropriate grade), and very finely disseminated graphite. The quartzite, which is commonly graded, shows a progressive increase upward from micaceous quartzite to increasing amounts of mica, garnet, and staurolite.

The complex mineralogical and structural modifications of the schist as one approaches the west contact of the Hatfield Pluton of the Belchertown Complex have been described in part by Stoeck (1971). These modifications include the appearance of andalusite and its subsequent complete or partial replacement by staurolite-muscovite, sillimanite-kyanite-muscovite, or coarse sillimanite depending on distance from the pluton.

At one locality 3-5 m from the western contact of the Littleton with the Gile Mountain Formation in the Whately area is a graded channel deposit, including quartz-pebble conglomerate. This locality is described and illustrated in detail by Robinson and others (Ch. D, this volume). Although this channel indicates an eastward younging across this contact (Littleton stratigraphically above Gile Mountain), complex regional

FIGURE 11.—Thinly laminated metamorphosed siltstone and shale of the Littleton Formation of area 1 about 5 km west-southwest of village of Bernardston. Lens cap is about 5. 5 cm across.

arguments discussed at length in Chapter D have led us to conclude that the Littleton is actually older than the Gile Mountain to the west and has been placed in its present position along the Whately thrust. This thrust, also known as the Chicken Yard line, forms the Gile Mountain-Littleton contact throughout areas 1 and 2 (fig. 10).

The Littleton Formation exposed in areas 3 and 4, and the western part of area 5 (that is, in the Northfield, Wendell, and Great Hill synclines), appears to have had a protolith similar to the Littleton in areas 1 and 2 but a different metamorphic history. The dominant rock type is homogeneous, gray-weathering, carbonaceous garnet-staurolite-mica schist in which bedding is generally very inconspicuous (Robinson, 1963). Where detectable, beds are generally thin and consist of 1- to 3-cm-thick quartzose laminae that rarely show grading. As a consequence of Mesozoic faulting, the level of exposure in areas 3, 4, and 5 is believed to be some 5 km deeper in the Acadian metamorphic pile than in area 1, and this appears to be reflected in the much coarser crystallization of micas and garnet. Typically, schist in these areas is strongly foliated and fissile, lead-gray to black on weathered outcrops and silvery-gray on fresh surfaces. The lead-gray or silvery appearance is due to a small amount of graphite finely disseminated in the coarse micas, which are matted together to form rippled silvery cleavage surfaces. The common mineral assemblages consist of quartz, muscovite, biotite, and garnet with or without staurolite. The garnets, as much as 1 cm in diameter, are light red and nearly spherical, and have well-developed dodecahedral and trapezohedral faces. They are dominantly almandine with subordinate pyrope and spessartine and minor grossular, and they show evidence of prograde growth zoning (Tracy and others, 1976). The staurolite occurs in relatively clear, dark, coffee-brown prisms as much as 7 cm long and 1 cm thick, some of which display 60° twins. Invariable accessories are brownish-green tourmaline and ilmenite, the former probably an indicator of former boron-bearing clay minerals deposited in a marine environment (Landergren, 1945; Volborth, 1955; Frederickson and Reynolds, 1960). Sodic plagioclase and retrograde chlorite are present in some specimens, but evidently no bulk compositions were appropriate for the formation of kyanite (Hall, 1970; Tracy and others, 1976); kyanite is present in other rock units in the vicinity. The only exception to this absence of kyanite is in the extremely narrow belt of Littleton that forms the east side of the southern extension of the Wendell syncline southeast of the village of Wendell. Here, typical Littleton-like graphitic garnet-staurolite-mica schist also contains minor kyanite. Garnet-quartz granulite (coticule) is present at one locality in this part of the Littleton, but the abundant occurrences of other rock types, such as the calc-silicate lenses found further east, are not documented.

The character of the schists of the Littleton Formation in area 5 changes rather abruptly near the sillimanite isograd, which passes through the hinge of the Ammonoosuc Volcanics of the extreme southeast corner of the Warwick dome (fig. 10). The presence of sillimanite (invariably as fibrolite) can be detected in outcrops by the milky-white appearance of white mica patches or quartz even where less than 1 percent is present. In this area, for reasons unknown, staurolite and garnet are finer grained than to the west; the garnet commonly is flattened and broken into irregular pods parallel to the foliation, and crystal faces are completely obliterated.

Northeast from the sillimanite isograd in area 5 along the east flank of the Warwick dome, fibrolitic sillimanite becomes increasingly abundant, first as a growth on muscovite, then also on biotite. In the northern and eastern parts of area 5, resistant quartz-sillimanite pods 1-2 cm thick are prominent in most outcrops, creating the distinctive "boot-grabber rock" (fig. 12). Primary bedding is commonly accentuated in outcrops by the distribution of these fibrous pods as well as by the gradual increase in abundance of quartzose beds and calcareous lenses. Typical assemblages in this region include quartz, muscovite, biotite, and garnet with sillimanite and (or) staurolite.

FIGURE 12.—Typical schist of Littleton Formation exposed in eastern part of area 5. White lenses and layers are quartz-sillimanite aggregates, which produce excellent traction resulting in the designation "boot-grabber rock."

Staurolite is less conspicuous and less abundant in the high-grade parts of area 5 but can always be found in large outcrops, usually together with sillimanite. In the sillimanite zone, graphite forms distinct crystal platelets and is not disseminated through the micas, so that the lead-gray to silvery sheen observed in the kyanite-staurolite zone is largely lost.

The southern branch of area 5 (fig. 10), east of the Kempfield anticline, contains the sillimanite-in reaction for Littleton bulk compositions. This branch extends into an area of intense late- to post-Acadian retrograde metamorphism and folding (Robinson, 1963; Hollocher, 1981, 1987) centered about 2 km north of the Prescott Complex. In this region, it has been possible to map successive retrograde isograds where sillimanite is replaced by muscovite, staurolite by muscovite and chlorite, garnet by chlorite, and finally biotite by chlorite (Hollocher, 1981). One sample contains retrograde chloritoid, and several have sphene replacing ilmenite, with Ca contributed from the breakdown of the grossular component of garnet. The result of this process is an area of outcrops of Littleton Formation that were once sillimanite-staurolite schists and are now gray chlorite phyllites, which superficially resemble the Littleton in area 1 and also the low-grade rocks near Worcester, Mass. (Emerson, 1917, p. 76). Within this retrograde zone, the additional folding gives many outcrops a knobby appearance typical of regions of fold interference. On weathered outcrops, the abundant beds of quartzose schist and quartzite stand out prominently from the pelitic beds, causing them to resemble adjacent exposures of the Clough Quartzite. Many large outcrops are covered by resistant white muscovite-chlorite pseudomorphs after staurolite (fig. 13), some as much as 8 cm long and 2 cm wide, and some showing typical 60° twins. Some outcrops show pits due to weathering of chlorite that replaced garnet or due to plucking of garnet that had retrograde chlorite shells.

The strata mapped as Littleton Formation in the Pelham-Shutesbury syncline (area 6) differ from all others in the region, and some would question their correlation, particularly since they are completely surrounded by older strata. Three separate patches of Littleton are shown on the State bedrock map in area 6. On the basis of detailed mapping by Michener (1983), we now believe that one of them, the gray-weathering kyanite-garnet-mica schist at the extreme northern tip of the syncline west of Wendell, is not Littleton but belongs to the extremely thin basal member of the Partridge Formation, which is here tightly pressed in the keel of the syncline. Michener has demonstrated that this distinctive gray schist occurs as a layer 2-10 m thick above the basal Partridge quartzite and below the much thicker rusty schists and amphibolites of the Partridge along the west limb of the syncline. On Michener's 1:24,000-scale map, the schist is included in the basal quartzite member of the Partridge, but neither is thick enough to show on the Massachusetts bedrock map except where they were mistakenly shown as Littleton and Clough southwest of Wendell.

The other two patches of Littleton Formation in area 6 lie in the southeastern portion of the Pelham-Shutesbury syncline where they are completely surrounded by Partridge Formation and locally extensive exposures of Clough Quartzite. The assignment of these rocks to the Littleton Formation is based on the fact that they are surrounded by a consistent descending sequence of units: Clough Quartzite with a recognized three-part internal stratigraphy, Partridge Formation, Ammonoosuc Volcanics, and Fourmile Gneiss. Because of very tight folding, the total exposed stratigraphic thickness of Littleton is probably less than 50 m, so it could be argued that the rocks are an unusual upper member of the Clough or an unusual lower member of the Littleton. For the present, at least, we prefer to assign them to the Littleton.

The rocks in these two patches of area 6 are light-gray, coarse-grained schists rich in quartz and muscovite and commonly containing conspicuous biotite and garnet. Kyanite is present in many samples and makes up as much as 30 percent of the rock in a few beds. It also occurs as bundles of blue prisms as much as 10 cm long along the walls of quartz segregations

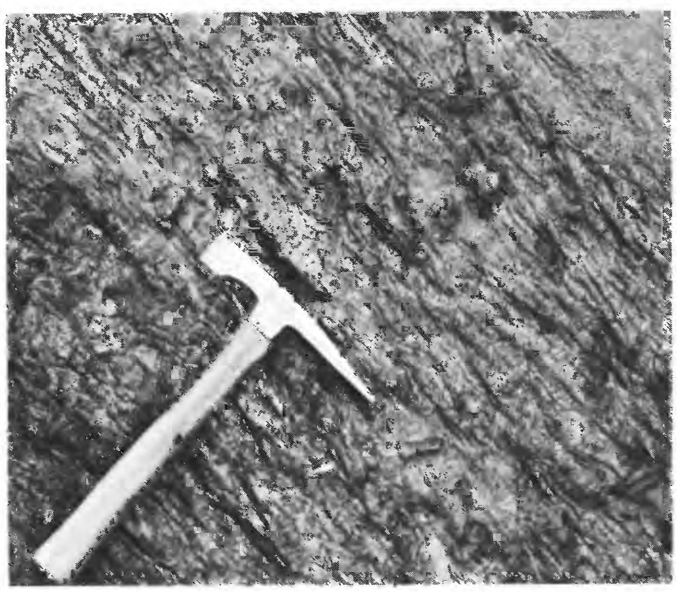

FIGURE 13.—Littleton Formation in zone of retrograde metamorphism near New Salem. Elongate masses are muscovite-chlorite pseudomorphs after staurolite.

and small pegmatites. In some beds, garnet is minor and crystals are about 0.5 cm in diameter; in others it is abundant and in crystals as much as 2 cm in diameter. Staurolite is commonly present but is everywhere inconspicuous. Graphite is not broadly disseminated in the micas as in other areas of Littleton in the kyanite-staurolite zone. Where best exposed on a tiny island in Quabbin Reservoir, three local members are distinguished (Robinson, 1967b): (1) gray, well-bedded muscovite schist and micaceous quartzite, (2) a big-garnet schist, and (3) poorly bedded, garnet-bearing, brown-weathering schist.

The Littleton Formation in area 7 (fig. 10) bears considerable resemblance to the sillimanite-grade Littleton in the northeastern part of area 5, with, however, some distinct differences. The rocks in area 7 were described in detail by Robinson (1963), who, because of structural complications, assigned them to a gray schist member of the Partridge Formation. The discovery of even greater structural complexity in 1966, including a new exposure of conglomerate of the Clough Quartzite, returned them to the Littleton Formation. As presently interpreted, these rocks represent a distinct eastern facies of the Littleton,[14] everywhere separated from the Littleton in area 5 by the root zone of an early nappe, which in some areas is a belt of older rock as narrow as 30 m.

The Littleton Formation of area 7 (fig. 10) is dominated by biotite-rich schist, feldspathic schist, and mica-feldspar gneiss, although some layers of sillimanite-staurolite-mica schist are identical to those in the eastern part of area 5. Typically, the schist in area 7 has a larger ratio of biotite to sillimanite plus muscovite, and the small amount of sillimanite typically occurs as isolated needles or clusters rather than in prominent quartz-sillimanite knots as in area 5. Normally, plagioclase is abundant. Many individual outcrops are strongly compositionally layered. Outcrops tend to be dark brown to black rather than dark gray as seen farther west. Mineral assemblages commonly consist of quartz, plagioclase, biotite, muscovite, garnet, and sillimanite in the sillimanite-staurolite and sillimanite-muscovite zones. Orthoclase is an additional phase in the sillimanite-muscovite-orthoclase zone, and it takes the place of muscovite in the sillimanite-orthoclase zone in the southern part of area 7, commonly in clear 1- to 3-cm porphyroblasts. Although staurolite is not typical, staurolite and staurolite-sillimanite schists are conspicuous in a curving belt that runs from northwest to northeast through Blissville on the northwest side of the Tully dome. Pegmatite and aplitic seams a few millimeters to 3-4 m in thickness are very common (fig. 14). They consist of quartz, plagioclase, biotite, and muscovite, or quartz, plagioclase, microcline, and biotite, and commonly are bordered by schists containing 30 percent to nearly 80 percent biotite. These pegmatites and aplites are interpreted to be the result of partial melting during metamorphism.

Accompanying this facies of generally more plagioclase-rich schists are inconspicuous lenses as much as 60 cm thick of gray calc-silicate granulite dominated by quartz and calcic plagioclase (An_{70-90}) containing various combinations of hornblende or actinolite, sphene, diopside, grossular, zoisite or clinozoisite, and calcite. On a tiny island in Sheomet Lake, northwest of Blissville, there is an apparently unique outcrop of well-foliated biotite marble.

FIGURE 14.—Typical schist of Littleton Formation exposed in northern part of area 7. White rock under hammer and in other lenses is anatectic pegmatite.

An unusual feature of area 7 is the scattered lenses, as much as 5 m thick, of "iron formation" that are apparently boudins of originally continuous beds.[15] They are mainly concentrated in the convoluted southwestern part of area 7, where they occur within a few meters of the base of the Littleton Formation. Early prospecting for magnetite in these lenses gave rise to the name "Iron Ridge" on early topographic maps. Robinson located many more occurrences in the 1960's, and they were the subject of a detailed mineralogical and petrological investigation by Huntington (1975),

[14] Most of these rocks are now reassigned to the Lower Silurian Rangeley Formation (see footnote 4).

[15] Now reassigned to the Silurian Perry Mountain Formation (see footnote 4).

who gave historical details and outcrop information. The dominant rock is well-bedded quartz-garnet-magnetite granulite with subordinate grunerite; the centers of the lenses contain more iron-rich rocks free of quartz and dominated by magnetite, Fe-Mn olivine, grunerite, garnet, and graphite. Other beds and cross-cutting veins contain unusual assemblages with pyroxmangite, Ca-Mn carbonate, Mn-rich grunerite, Fe-Mn orthopyroxene, and apatite. These well-bedded rocks appear to represent a local phase of chemical sedimentation under reducing conditions very early in the sequence of Littleton sedimentation. Unlike many other well-bedded garnet-rich rocks in the region, these appear to have had no relationship to volcanic activity.

The Littleton Formation of area 8 in the Amherst inliers (Jasaitis, 1983) resembles the Littleton in the higher grade parts of area 7 and in the western part of the Merrimack belt.[16] The rocks of the Amherst inliers are believed to be in the frontal regions of several high-level early Acadian nappes, brought down to their present position by Mesozoic movement on the Connecticut Valley border fault. For this reason, it is believed that in the middle and late Paleozoic before Mesozoic faulting there was a direct physical connection between these areas over the crest of the Bronson Hill anticlinorium. The outcrops in the Amherst inliers are very poor in most areas and nonexistent in other areas, where contacts have been drawn only on the basis of water-well data. Furthermore, most of the outcrops consist of pegmatite and granodiorite intrusions, and the schists themselves are strongly altered by late metamorphic and Mesozoic hydrothermal alteration and locally by pre-Triassic weathering. Nevertheless, the Littleton is recognized as relatively biotite-rich, feldspathic, generally poorly bedded schist, which shows evidence of having originally contained sillimanite-muscovite-orthoclase assemblages. Lenses of gray-weathering calc-silicate granulite occur widely and commonly contain abundant quartz and calcic plagioclase and minor diopside, grossular, and calcite.

A characteristic of the Littleton Formation in the Connecticut Valley belt, as compared with the underlying Partridge Formation, is the nearly total absence of metamorphosed volcanic rocks. Of the few lenses of amphibolite identified by Robinson (1963), all but two have since been reassigned to the Partridge Formation.

The thickness of the Littleton Formation can only be estimated where it is overlain, apparently unconformably, by the Erving Formation on the west side of the Bronson Hill anticlinorium. Where the Littleton is thickest along the west flank of the Warwick dome, it is about 700 m thick. Elsewhere in Massachusetts where the stratigraphic top is not exposed, the minimum thickness of the formation may be as much as 800 m.

In conclusion, the irregular distribution of the underlying Fitch Formation and Clough Quartzite suggests that the Littleton was deposited on an unconformity, at least near the Bronson Hill anticlinorium. The dominant rock type in the Littleton was marine black shale, deposited with abundant carbonaceous matter under reducing conditions but in an environment where sulfur- reducing bacteria were not active. Interbedded quartzose beds suggest deposition by turbidity currents. The eastward increase in quartzose beds and in plagioclase content is at least consistent with a source in that direction and with the model proposed for equivalent rocks in Maine (Hall and others, 1976) that the Littleton represents the distal portion of a marine deltaic complex originating in tectonic land to the east.

ERVING FORMATION (De, Deg, Dea, Dev)

The Erving was first assigned formational status by Thompson and others (1968), on the basis of work of Robinson (1963, 1967a,b) who originally described it as "the Erving Member of the Littleton Formation." Robinson (1963) defined the Erving to include the Erving Hornblende Schist of Emerson (1917) and quartz-plagioclase biotite granulites and mica schists associated with the hornblende schist near the type locality northeast of the village of Erving, Mass. (fig. 10). From this area, in the Northfield and Wendell synclines on the northeast flank of the Pelham dome, the Erving has been extended, mainly on the basis of lithic peculiarities and stratigraphic sequence, to the west flank of the Pelham dome (Robinson, 1967a; Laird, 1974), the Quabbin Hill area (Robinson, 1967b), the west and north flanks of the Glastonbury dome (Peper, 1967; Leo and others, 1977), and the Whately area west of the Connecticut Valley Mesozoic basins (Walter Trzcienski, written commun., 1968; Stewart Clark, written commun., 1977). Detailed description and distribution of rock types in these areas are given in Robinson and others (Ch. D, this volume).

The dominant rock type in the Erving Formation of the type area is gray, well-bedded, fine-grained, quartz-plagioclase-biotite granulite with very thin to thick beds of muscovite-biotite schist commonly containing dodecahedral garnet, ilmenite, staurolite, and (or) kyanite (fig. 15). These schists are distinct from schists of the adjacent Littleton Formation, which have abundant graphite, lack kyanite, have slightly more

[16] Now correlated with the Rangeley Formation for the same reasons.

FIGURE 15.—Typical well-bedded, fine-grained plagioclase-quartz-biotite granulite of Erving Formation. Contains beds of mica-kyanite schist to right and 1- to 3-cm layers of garnet-hornblende calc-silicate. From center of Northfield syncline in the back yard of house southeast of junction of Gulf and Orange Roads, Northfield.

iron-rich ferromagnesian minerals, and have trapezohedral garnets. The Erving granulites contain minor but nearly ubiquitous lenses and layers of fine-grained calc-silicate granulite containing prominent garnet and hornblende, along with fine-grained diopside, calcic plagioclase, clinozoisite, or zoisite. Calc-silicate rock is locally predominant in the lowest mapped Erving granulite lenses along the east limb of the Northfield syncline and in small outcrops nearest the Littleton contact on the east ridge of Round Mountain, Northfield, where it consists of dark hornblende calc-silicate with white centimeter-sized euhedral prisms of zoisite.

The granulites of the Quabbin Hill area (fig. 10) are, for the most part, entirely typical of the Erving, but the mica schists tend to be dominated by muscovite and biotite, generally lacking the abundant garnet, staurolite, and kyanite seen to the north. However, a granulite layer on Prescott Peninsula contains the only known graphitic schist layer in the Erving and also contains garnet, staurolite, and kyanite (Robinson, 1967b, p. 120-121). The basal granulite unit on Prescott Peninsula, about 10-40 m thick, is dominated by calc-silicate rocks, including coarse diopside-zoisite-microcline calc-silicate with matted diopside and zoisite prisms 5-25 cm long. Also present, particularly in the lowest 1 or 2 m, are beds of dark hornblende-rich calc-silicate with white 1- to 2-cm prisms of zoisite identical with the basal hornblende-zoisite rocks on Round Mountain in the Northfield syncline. As this basal calc-silicate unit is traced northwestward around the set of late folds on Prescott Peninsula, it grades into a finely laminated, brown-weathering, muscovite-bearing marble, with, however, the same hornblende-zoisite calc-silicate at the base.

The granulites of the Wilbraham area have a higher than normal abundance of mica schist beds. These rocks bear a close resemblance to the granulite and schist exposures near Quabbin Hill but appear to have reached slightly higher metamorphic grade as shown by local patches containing sillimanite (Leo and others, 1977). The granulite in the Whately area is predominantly a fine-grained, well-layered, quartz-biotite-oligoclase rock containing a few beds of gray mica schist and rare garnet schist. Small lens-shaped concordant quartz veins are characteristic.

The second most abundant rock type of the Erving Formation is hornblende-andesine-epidote-sphene amphibolite in sharply bounded mappable layers from 0.5 m to tens of meters thick. The predominant textural type is thinly laminated, with alternate hornblende-rich and epidote-plagioclase-rich layers, that is interpreted to have been laminated fine-grained basaltic tuff. Some light-colored layers also contain abundant diopside. Less commonly, the amphibolites are coarse grained and massive with prominent 1- to 2-cm-sized hornblende megacrysts suggesting porphyritic flows or coarse crystal tuffs. The megacrysts commonly contain inclusions of diopside and calcite. In one outcrop there are vague suggestions of pillows. The Erving amphibolite is relatively easily distinguished from adjacent amphibolite outcrops of the Partridge Formation and Ammonoosuc Volcanics by its mineralogical monotony and distinctive range of textures. In the type area, the amphibolite layers tend to be concentrated near the base of the formation. The hornblende megacrystic amphibolite is characteristic of the lowest amphibolite layers, although it does occur elsewhere.

The Erving Formation west of the Pelham dome is dominated by amphibolite, and an excellent chemical analysis of this rock from south Leverett was published by Emerson (1917). Most of the amphibolite of the Quabbin Hill area appears to belong to a single contorted layer, possibly a relict volcanic pile as thick as 400 m at the peak of the hill and thinning along strike to as little as 1 m. Amphibolite in the Wilbraham area occurs in relatively scarce isolated layers and within a separately mapped unit of amphibolite and rusty-weathering schist. The Erving of the Whately and Leeds areas is dominated by well-laminated epidote amphibolite, like that described above, but one exposure containing anthophyllite has been reported (Walter E. Trzcienski, oral commun., 1967).

Within the Erving granulites, commonly within 1 or 2 m of amphibolite contacts, are beds 0.25-3 cm thick or 0.5-cm irregular patches of pink coticule granulite containing ultra-fine-grained euhedral manganese-bearing garnet. The matrix surrounding the garnet is mainly a quartz-biotite granulite, typically with magnetite that is either fine grained or, more commonly, as discrete octahedra as much as 0.5 cm in diameter. The latter may also occur within the solid coticule beds, where they commonly are associated with feldspar pressure shadows. Unlike many coticule beds found in the Ordovician Ammonoosuc Volcanics and Partridge Formation (Robinson, 1963), the Hawley Formation (Hatch, 1969; Hatch, Norton, and Clark, 1970), and the Lower Devonian Littleton Formation (Huntington, 1975), which contain abundant grunerite, gedrite, or hornblende, the Erving coticules typically are dominated by green biotite with or without muscovite and potassic feldspar.

These coticules are widespread in the type area of the Erving Formation. They are also present in the Leverett area and are particularly spectacular near Quabbin Hill on the shores of Quabbin Reservoir. They are not reported in the Wilbraham area but are present within granulites in the Whately and Leeds areas. The origin of these coticules is still problematical, but they appear to have been derived from chemical precipitates that are somehow related to basaltic volcanic activity.

On the State bedrock map, the Erving Formation is shown by several different letter symbols. Unit De represents interbedded granulite and amphibolite that are too intimately interlayered and folded to be shown separately at 1:250,000 scale, as in the type area and at Whately. Elsewhere, areas of granulite, Deg, and amphibolite, Dea, are large enough or uncomplicated enough to be shown separately, as near Leverett, Quabbin Hill, and Wilbraham. Near the base of the formation in the Wilbraham area, a unit of interlayered amphibolite and gray to rusty schist, Dev, is mapped separately.

The Erving Formation rests on a variety of older units. The youngest of these is the Littleton Formation, which underlies the Erving in the type area only. Elsewhere, the Erving rests on Clough Quartzite, Partridge Formation, and Fourmile Gneiss. These relations suggest an unconformity at the base of the Erving, although no sedimentologic features are recognized in support of this interpretation.

In most areas, the Erving is the highest exposed unit, but in the Wilbraham and Whately areas the Erving is overlain by the Waits River and Gile Mountain Formations, respectively. Where the top of the Erving is exposed in the Wilbraham area, the formation is 500 m thick; in the Whately area it is 0-400 m thick. Where the top is not exposed, it is at least 900 m thick near Northfield, 200 m thick near Leverett, and 600 m thick near Quabbin Hill.

The sedimentary protoliths of the Erving granulites were probably calcareous siltstones and interbedded shales and beds and lenses of impure limestone and dolomite. Their provenance is uncertain. The amphibolites are metamorphosed basaltic tuffs and subordinate flows, locally pillowed, of surprisingly uniform character. In the granulites, invariably close to amphibolite contacts, are the coticule rocks dominated by fine-grained garnet. These are interpreted to be chemical sediments derived from hydrothermal solutions associated with submarine volcanic activity. The uniformity and continuity of bedding in all rock types except the massive amphibolites suggests very stable conditions of sedimentation relatively far from sediment sources but possibly close to some submarine volcanic vents.

GOSHEN FORMATION (Dg, Dgq, Dgu, Dgl, Dgc, Dgp)

DISTRIBUTION OF MAPPED MEMBERS

In order to clarify the ambiguities in Emerson's (1898a,b) mapping of the superficially similar gray phyllites of the Middle Ordovician Hawley and Lower Devonian Goshen Formations, Hatch (1967) redefined the Goshen Formation in western Massachusetts. Subsequent mapping and reconnaissance work by Hatch, Stanley, and Stewart F. Clark, Jr. (discussed below), require modifications to Hatch's 1967 redefinition.

The Goshen Formation underlies approximately the western half of that part of the Connecticut Valley belt west of the Mesozoic basin. At the Vermont State line, it is exactly continuous with the Northfield Formation of southern Vermont (Skehan, 1961; Doll and others, 1961). The name Goshen, rather than Northfield, is used for these rocks in Massachusetts because it has been applied to at least some of them in Massachusetts since 1898 (Emerson, 1898a,b), and because the relatively massive and only faintly bedded rocks of the Northfield of Vermont change character quite dramatically (as discussed below) a few miles south of the Massachusetts State line.

On the State bedrock map the Goshen Formation is subdivided into six informal members, units Dg, Dgq, Dgu, Dgl, Dgc, and Dgp. All six members are characterized by at least some gray, graphitic, generally nonrusty-weathering schist. Most are well bedded, and, although demonstrably isoclinally folded (Hatch, 1968, 1975; Hatch and Stanley, Ch. C, this volume), beds are remarkably planar even in large outcrops (fig. 16). Three of the members, Dg, Dgq, and Dgu, are relatively widespread and are found in a consistent stratigraphic sequence in a north-south belt across the State (fig. 3). The other three members, Dgl,

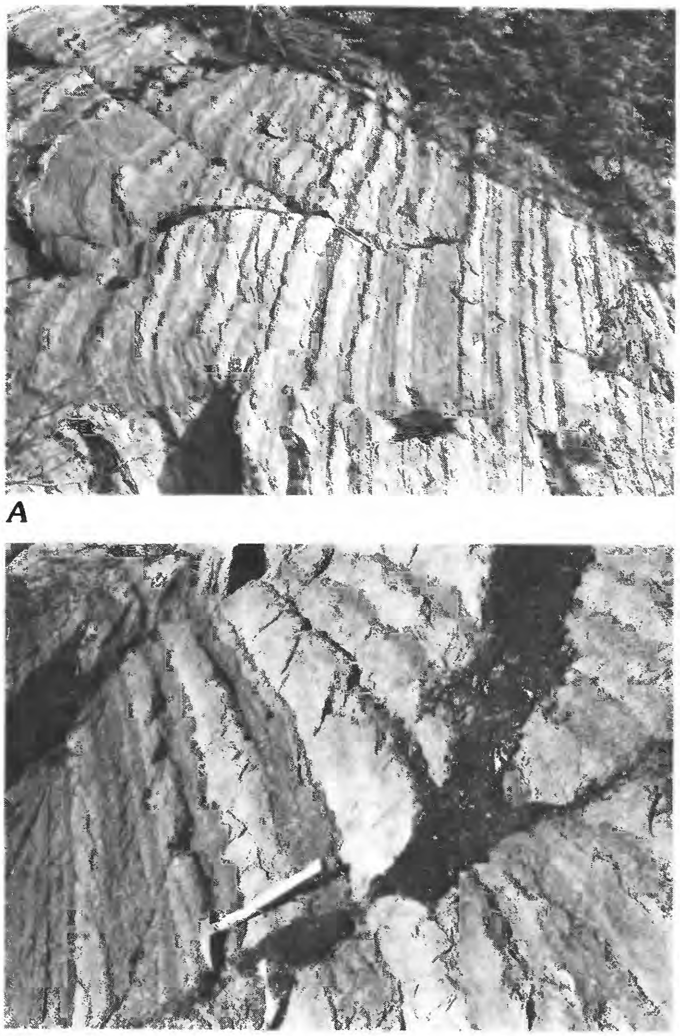

FIGURE 16.—Thinly bedded, graded-bedded aluminous schist and micaceous quartzite of member Dg of the Goshen Formation. Note straightness of bedding. Graded beds indicate tops are to the left in *A* and to the right in *B*. Route 9, 1.5 km west of Lithia.

Dgc, and Dgp, are localized around the vicinity of the Shelburne Falls, Woronoco, and Granville domes (Hatch and Stanley, Ch. C, this volume), respectively (fig. 3). They each can be locally related to the Dg-Dgq-Dgu sequence, but their physical isolation from each other precludes positive determination of their stratigraphic relations to each other.

Unit Dg is the most widespread of the six members of the Goshen and is the most diagnostic of the formation. It consists of very distinctly bedded, graded-bedded granulite and schist (fig. 16). Individual beds are generally 3-15 cm thick and grade upward from light-gray to light-tan quartz-mica granulite or micaceous quartzite through medium-gray quartz-muscovite-biotite-garnet schist to dark-gray graphitic muscovite-quartz-biotite-(staurolite)-(kyanite)-garnet schist or phyllite. Approximately the basal 50 m of the unit along the west border of the belt is somewhat less distinctly bedded, particularly in the northern half of the State. Despite well documented and very widespread isoclinal folding refolded by one or two episodes of refolding (Hatch and Stanley, Ch. C, this volume), even very large exposures of this member are characterized by remarkably planar bedding (fig. 16). The member is estimated to be about 300 m thick (Hatch, 1968), although complex folding makes this figure very approximate.

In the central part of the outcrop belt of the Goshen, roughly between the villages of Hawley and Blandford, consistent facing directions of graded beds indicate that member Dg is stratigraphically overlain by Dgq (figs. 3, 17). Beds in Dgq are characteristically 15 cm to 6 m thick (fig. 18). They are only locally graded but may be cross bedded. The member consists predominantly of light-gray to light-gray-brown, locally massively bedded ("elephant rock") micaceous quartzite and quartz-mica-garnet schist. The second, though less abundant, characteristic lithology of the member is calc-silicate granulite and granulose schist consisting of quartz, plagioclase, mica, garnet, tremolite, chlorite, calcite, zoisite, and sphene. Scattered beds of light-gray, dark-brown-punky-weathering quartzose marble consisting of ferroan calcite or dolomite plus quartz, biotite, garnet, and actinolite are 15 cm to 2 m thick. Dark-gray graphitic schist is minor and generally not as well graded as in Dg. Member Dgq is estimated to be about 300 m thick (Hatch, 1968).

In the south-central part of the Goshen outcrop belt, graded beds and cross beds consistently indicate that Dgq is stratigraphically overlain by a sequence of granulites and graphitic aluminous schists in 5- to 15-cm-thick graded beds. This member, designated Dgu (fig. 3), is distinguished from member Dg only by its stratigraphic position *above*, rather than below, Dgq, and by the presence in it of widely scattered beds about 15 cm thick of calc-silicate granulite or schist similar to the calc-silicate rocks in Dgq. As much as 150 m of Dgu are present in some of the deeper synclines near North Chester.

We now believe that the unit mapped as Waits River Formation (unit DSw of Hatch and Hartshorn, 1968) northeast of the Shelburne Falls dome (fig. 3) is more properly considered a member of the Goshen Formation. Although it contains scattered beds of punky brown-weathering carbonate rock as much as 6 m thick, "noncalcareous rocks, which constitute about 97 percent of the [Waits River] formation, are indistinguishable from the schists and phyllites of the underlying Goshen Formation" (Hatch and Hartshorn, 1968). Thus, that map unit, which is designated Dgl on the State bedrock map and figure 3, consists of well-

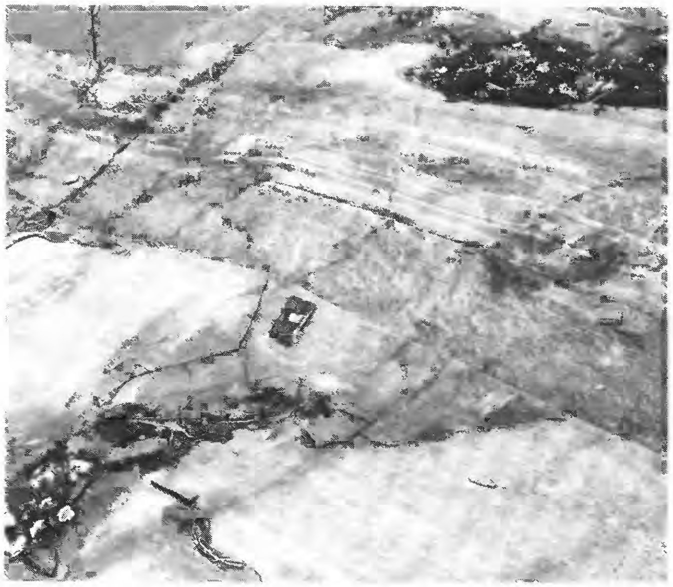

FIGURE 17.—Contact between thinner bedded, more micaceous strata of member Dg of the Goshen Formation to the right of key case and thicker bedded, more quartzose strata of member Dgq of the Goshen, State Route 9, Cummington. Graded beds indicate Dgq stratigraphically overlies Dg. Key case is 10 cm long.

FIGURE 18.—Thick-bedded micaceous quartzite of member Dgq of the Goshen Formation, 1.5 km north of Cummington. Key case is 10 cm long.

bedded, graded-bedded schist and granulite typical of Goshen member Dg plus as much as 3 percent calcareous rock. This unit has been traced around the Shelburne Falls dome and thence south to the latitude of Ashfield where it migrates westward, at the present ground surface, to a position within, rather than at the east contact of, the main body of Goshen, Dg (fig. 3). This relationship further supports the assignment of unit Dgl to the Goshen Formation rather than to the Waits River.

A corollary of this modification of the Goshen-Waits River boundary is the fact that a more fundamental lithologic change takes place east of member Dgl in the northern part of the State across a boundary that has been traced southward to the vicinity of Russellville. Eastward across this boundary, which is the western contact of unit Dw on figure 3, beds of brown-weathering impure carbonate rock ranging in thickness from a few tens of centimeters to 10 m increase abruptly in abundance, and the intercalated schist changes markedly in character. East of this boundary, the schist is characteristically inconspicuously bedded, contains innumerable quartz veins a few centimeters thick and tens of centimeters long, is highly contorted, and contains much less staurolite, garnet, and kyanite in appropriate metamorphic zones compared to the schist west of this boundary. In outcrop the schist east of the boundary appears messy and chaotically deformed, in marked contrast to the straight-bedded and planar schist of the Goshen Formation to the west (compare fig. 16 with fig. 21*A*).

For these reasons, we herein modify the definition of the Goshen-Waits River boundary to include as a member of the Goshen Formation (Dgl) the unit of planar-bedded schist and minor interbedded brown-weathering carbonate rock mapped as Waits River Formation by Hatch and Hartshorn (1968). We now define the formational boundary as the line along which, going eastward, the character of the schists changes from aluminous and planar-bedded, with only rare quartz veins, to relatively alumina-poor, contorted, and rich in quartz veins. This change may or may not coincide with the eastward appearance of significant beds of punky brown carbonate, depending upon the presence or absence of unit Dgl between the two redefined formations.

An elliptical area about 3 km east of Woronoco and 6 km northwest of Westfield is underlain by a unit designated on the State bedrock map as Dgc (fig. 3). Clark (1977) described these rocks as representing a transition between the well-bedded schist of the Goshen Formation and the more massive schist and massive calcareous rocks of the Waits River Formation. Although these rocks were previously assigned to the Russell Mountain Formation of Silurian age by Hatch, Stanley, and Clark (1970) as noted above, the combination of punky brown-weathering impure carbonate rock and gray pelite is more appropriately assigned to the Lower Devonian sequence. These rocks are lithologically similar to the rocks of unit Dgl described above but are separately designated on the State map because they appear to lie at the base of the Goshen Formation immediately upon three small areas of Ordovician Cobble Mountain Formation in the Woronoco dome rather than at or near the top (or east contact) of the Goshen, as does member Dgl.

Surrounding the Granville dome in southernmost Massachusetts and adjacent Connecticut is a tight syncline of Goshen Formation (fig. 3). This package of rocks was mapped by both Schnabel (1974) and Knapp (1978) as Straits Schist, and both subdivided it into two units. We agree with Schnabel and Knapp that these rocks are indeed part of The Straits Schist, but because The Straits of Connecticut is correlative with the Goshen of Massachusetts (Hatch and Stanley, 1973), we have used the name Goshen on the State bedrock map for continuity. The stratigraphically lower of the two members of the Goshen around the Granville dome is mapped as member Dg. The overlying member, Dgp, was described by Knapp (1978) as "brown to brownish-gray, medium- to coarse-grained quartz-plagioclase-muscovite-biotite-(garnet)-(sillimanite) schist." This member is distinguished from the adjacent Dg rocks by a greater abundance of "strongly boudinaged layers or lenses of dark-gray to greenish-gray, medium- to coarse-grained calc-silicate composed of varying amounts of quartz, plagioclase, hornblende, diopside, tremolite, garnet, zoisite and sphene" and a paucity of beds of "medium-grained, brown to brownish-gray quartz-plagioclase-mica gneiss" (Knapp, 1978).

Harris and others (1983) have modified the age of the Fitch Formation to latest Silurian. From stratigraphic position and lithology, as well as paleontology, the Fitch and underlying Clough are reasonably correlated with the Shaw Mountain of Vermont and the Russell Mountain of Massachusetts. These relations lead to the present interpretation that the gray strata of the Connecticut Valley synclinorium are not only all younger than latest Silurian but are also essentially correlative with or slightly younger than the fossiliferous Lower Devonian Littleton Formation and thus are Early Devonian in age. A Littleton or post-Littleton age for the entire Goshen-Waits River-Gile Mountain package is further supported by the discussion presented by Robinson and others in Chapter D of this volume. Thus, while we recognize the uncertainties of dating these polydeformed rocks from which *no* fossils of any sort have been reported in Massachusetts, the ages of the Goshen, Waits River, and Gile Mountain Formations are designated Early Devonian contrary to the previous designation of "Lower Devonian to Middle Silurian" (Hatch and Stanley, 1976).

The protolith of the Goshen Formation is uncertain. The well-bedded graded-bedded character of members Dg and Dgu (fig. 16) suggests turbidites deposited in at least moderately deep water. The abundant fine graphite suggests a closed reducing basin, although the relative paucity of sulfide minerals implies that sulfur-reducing bacteria were not present. The carbonaceous schists of the other members of the Goshen presumably imply the same closed-basin reducing conditions, but the thick beds of siliceous carbonate, calc-silicate rock, and micaceous quartzite all seem to suggest rather rapidly deposited detritus from not-too-distant carbonate-rich and quartz-rich sources. No firm evidence is presently available on the source direction for the Goshen sediment. Hall and others (1976), proposed an eastern source for the Lower Devonian Seboomook Formation of northern Maine, but their study area is too far removed both across and along strike to be applied with any confidence to western Massachusetts.

INTERNAL GOSHEN STRATIGRAPHIC RELATIONS

The three-part stratigraphic sequence Dg, Dgq, Dgu within the Goshen Formation has been traced from Plainfield south to the vicinity of Huntington in the terrane east of the Hawley-Goshen contact and west of the domes (fig. 3). Within that area, the character and thickness of the rocks in these three units are remarkably constant, both parallel to and across the structural trend. Graded bedding and data on minor structures indicate that the northern and southern terminations of members Dgq and Dgu are due to the gentle south and north plunge, respectively, of the synclines in which they lie. Regionally, however, the fact that the quartzite and calc-silicate rocks of member Dgq are not recognized anywhere to the north in Massachusetts or Vermont suggests that the unit also dies out through a facies change somewhere north of its structural termination. These relations further suggest that the original sedimentary source area for the Goshen may have been to the south or southwest and that the Goshen may represent deposits in a basin that filled by sediment transport parallel to the orogenic belt. We emphasize, however, that the abundant sedimentary features still preserved in these rocks have not been studied systematically for definitive information on source direction and depositional environment.

The basal unit (Dg) of the Goshen continues north without significant change to the vicinity of the Deerfield River. Here the more quartzose basal parts of the thin graded beds that characterize member Dg south of the river become thinner, more pelitic, and less distinguishable from their pelitic tops. Through the northern few kilometers of the unit in Massachusetts, bedding can be distinguished only locally, and member

Dg takes on the poorly bedded character typical of the Northfield Formation of Vermont with which it is continuous. This northward progression from well-bedded, graded-bedded lower (Dg) Goshen to poorly or indistinctly bedded Northfield seems most reasonably explained by a facies change (Hatch and Hartshorn, 1968).

Other north-south facies changes within the Goshen Formation are suggested by the State bedrock map. Member Dgl, which appears to occupy a stratigraphic interval around the Shelburne Falls dome either at the upper part of member Dg or possibly equivalent to the interval occupied by member Dgq to the south, clearly pinches out near the Vermont State line and south of Ashfield (fig. 3) (Hatch, 1981). Member Dgc, which forms the base of the Goshen around the Woronoco dome, is not present around the Goshen dome to the north, or at the base of the Goshen to the west, and thus must pinch out or undergo a facies change both north and west of the Woronoco dome.

Member Dgp occupies the same stratigraphic position around the Granville dome as does member Dgq to the north. Although both members Dgp and Dgq contain calc-silicate rocks, the thick beds of micaceous quartzite that characterize member Dgq are not present in member Dgp. Some southward facies change in the upper part of the Goshen section is clearly required between Huntington and the north end of the Granville dome.

East-west facies changes within the Goshen Formation are just as important as the north-south changes. Members Dgq and Dgu, which make up much of the formation west of the Goshen dome, are not present east of the dome. This absence east of the dome could theoretically be explained by a failure of their stratigraphic position to be structurally low enough to intersect the present ground surface. However, the failure of these units to appear *anywhere* east of the dome, in either the Goshen or other Lower Devonian units, would appear to support the concept of their having disappeared by changing facies eastward. Finally, as noted above, member Dgc fades out by facies change northward, and its absence at the base of the Goshen along the Hawley contact indicates that it also pinches out by facies change westward. Some of these facies changes could theoretically be explained by truncation of units along a major thrust fault on the surface at the base of the Goshen Formation along which we have mapped a "structural disharmony" or decollement (Zen and others, 1983). A structural explanation for the lithologic changes observed in the higher parts of the formation, however, would require a complex array of additional thrusts, for which there is no direct evidence. Although the thrust fault model is possible, we believe that the presently available evidence favors the facies model.

WAITS RIVER AND GILE MOUNTAIN FORMATIONS
(Dw, Dwt, Dwa, Dgm, Dgmq, Dgma)

Emerson (1898b, 1917) mapped a large area of rocks east of his Goshen Schist and west of the Mesozoic basin as Conway Schist. Within this area he locally mapped conspicuous beds of quartzite, now mapped as isoclinal synclines of member Dgq of the Goshen, and marble, now mostly included in the Waits River Formation. Segerstrom (1956a,b) and Willard (1956) subdivided Emerson's Conway Schist into a "marble conspicuous" member, a "quartzite conspicuous" member, a local "phyllite" member, an "amphibolite" member, and large areas of "undifferentiated."

Between the time of our work and that of Segerstrom and Willard, the Centennial Geologic Map of Vermont (Doll and others, 1961) was published. On this map the northern extension of the Conway of Massachusetts was subdivided into the Waits River and Gile Mountain Formations. Although controversy has existed and still does exist over their relative ages, our detailed and reconnaissance mapping has shown that the Gile Mountain and Waits River are readily distinguished in the field in Massachusetts and require no "undifferentiated" category. For these reasons, and to afford greater regional continuity across the State line, we chose to follow the Vermont subdivision and nomenclature for these rocks on the Massachusetts bedrock map. The configuration on the map results from extensive reconnaissance and local detailed remapping, particularly in the Colrain, Shelburne Falls, Williamsburg, and Easthampton quadrangles, by S.F. Clark, Jr., L.M. Hall, J.W. Pferd, and ourselves.

We have described the boundary between the Goshen and Waits River Formations in the previous section of this chapter. The fundamental difference, as mapped, between the Waits River and Gile Mountain Formations is the presence in the Gile Mountain of beds of generally noncalcareous, commonly micaceous, quartzite (fig. 19). *Both* formations contain conspicuous beds of punky brown-weathering impure marble or calcareous granulite (fig. 20), traditionally the hallmark of the Waits River, although they are less abundant in the Gile Mountain. The *predominant* lithology of both formations is typically contorted, gray, graphitic, locally very slightly sulfidic, moderately aluminous mica schist containing quartz veins (fig. 21). A gradational but definitely significant boundary can be mapped, however, delimiting areas in which beds of fine-grained, light-gray to light-gray-brown, variably micaceous quartzite 10 cm to a few meters thick are intercalated with the ubiquitous highly contorted

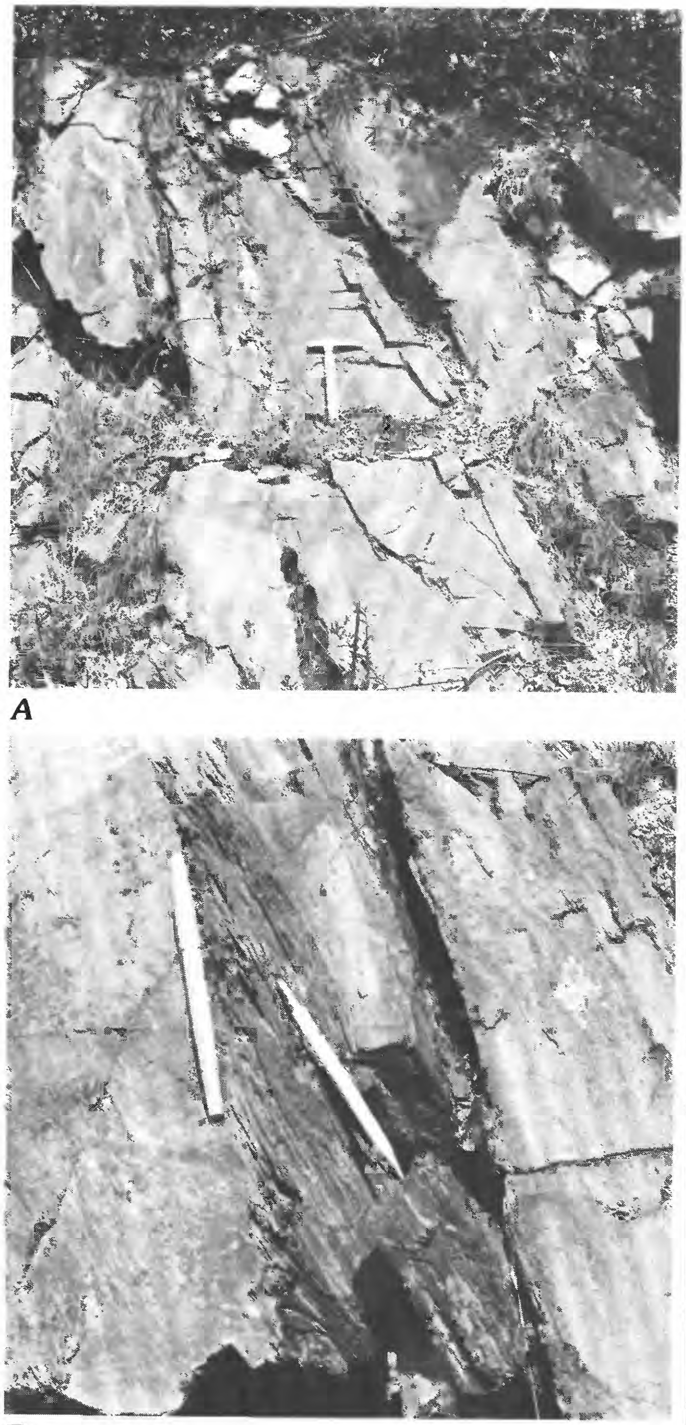

FIGURE 19.—Micaceous quartzite characteristic of the Gile Mountain Formation. About 3 km west of Greenfield. In *B* pen is parallel to bedding whereas pencil is parallel to pinstriping and schistosity.

schists that characterize these two formations. This is the boundary that is shown on the State bedrock map separating the Waits River from the quartzite-bearing Gile Mountain Formation.

Some of the unnamed but separately mapped members of the Waits River and Gile Mountain on the State

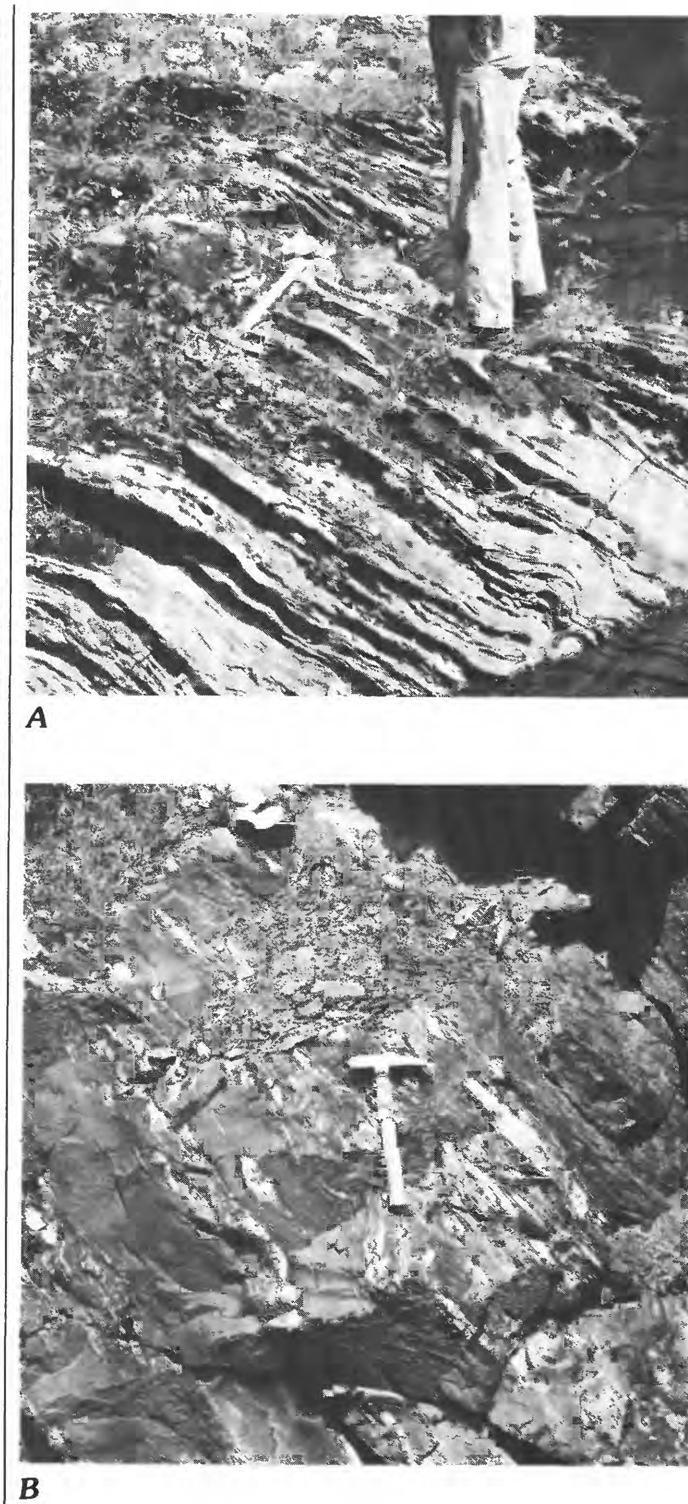

FIGURE 20.—Punky brown-weathering calcareous granulite common in the Waits River Formation but also present in lesser amounts in the Gile Mountain Formation. *A*, Calcareous granulite bed in the Waits River Formation, under hammer, Green River, 1.5 km north of Stewartsville (fig. 22). *B*, Calcareous granulite bed in the Waits River Formation, under hammer, 5 km north of Conway (fig. 22).

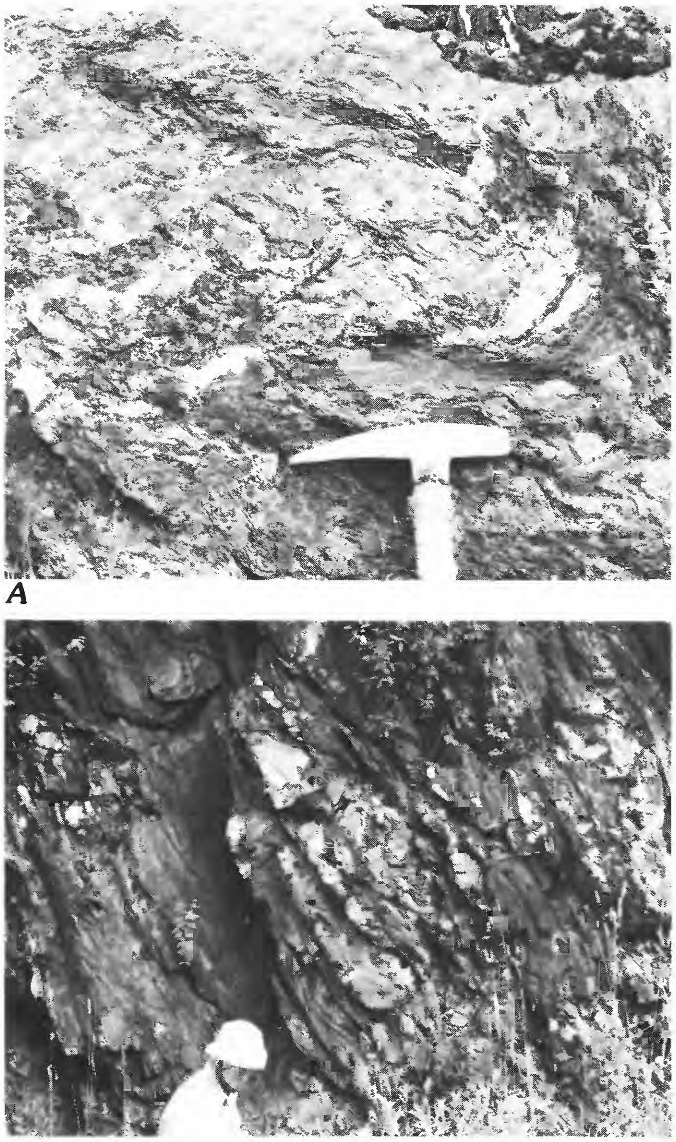

FIGURE 21.—Typically contorted schist characteristic of the Waits River and Gile Mountain Formations. Note abundant veins of white quartz. *A*, Waits River Formation southwest of Shingle Hill, southwest of Greenfield. *B*, Gile Mountain Formation, 6 km west-northwest of Bernardston.

bedrock map either have not been distinguished on previously published maps or have been used in a different sense. Both Emerson (1898b, 1917) and Segerstrom (1956a,b) mapped a belt of amphibolite south from the Vermont line to a point a few kilometers east of the village of Conway (State map; fig. 22). Both included it as a member of the Conway Schist. The rock is generally medium grained and dark green to black. It consists primarily of 2- to 3-mm crystals of hornblende plus plagioclase, quartz, and minor chlorite, biotite, and magnetite. We interpret the rock to be a metamorphosed mafic tuff. As a result of the definition of the boundary between the Waits River and Gile Mountain Formations presented above, some parts of this amphibolite are now entirely within the Waits River Formation whereas other parts are either at the formational boundary or within the Gile Mountain (fig. 22). Thus some parts are designated Dwa whereas others are designated Dgma. Because of the probable time-equivalent facies relation of the Gile Mountain and Waits River, however, all of these amphibolites may represent only one period of volcanism. Units Dwa and Dgma probably correlate with the Standing Pond Volcanics that occur at or near the Gile Mountain-Waits River contact in Vermont.

Unit Dwt, shown only in the northernmost part of the State, was recognized by Pferd (1978) in the Colrain 7.5-min quadrangle. It is distinguished from surrounding unit Dw by containing beds of calcareous granulite thicker than 1 m. We were not able to make a similar subdivision in the rest of the Waits River Formation further south where virtually all of the formation would have to be included in unit Dwt by Pferd's definition.

The unit Dgmq on the map is distinguished from the main body of Gile Mountain Formation by two criteria. First, beds of quartzite are more abundant and thicker in unit Dgmq than in the rest of the formation. Second, the schists of unit Dgmq are somewhat finer grained, more quartz rich, and less contorted.

The stratigraphic scheme followed on the State bedrock map thus represents a significant departure from the earlier systems of Emerson and of Segerstrom and Willard. Because Emerson did not recognize the isoclinal folding in the Goshen, he mapped his Goshen-Conway contact approximately at the west margin of the westernmost synclinal belt unit Dgq. Some of these synclines of unit Dgq appear on Emerson's earlier (1898b) map as mapped beds of quartzite in the western part of his Conway Formation. In the eastern part of his Conway, he mapped beds of limestone, most of which are now interpreted as being within the Waits River Formation. Thus Emerson's Goshen Schist was approximately equivalent to unit Dg of the presently defined Goshen west of the westernmost syncline of unit Dgq. Emerson's Conway Schist included most of the terrane in which unit Dgq is infolded with units Dg and Dgu, as well as the terrane now mapped as Waits River and Gile Mountain Formations (see fig. 3).

The stratigraphic scheme used on the State bedrock map also differs markedly from that used by Segerstrom (1956a,b) and Willard (1956), and the resulting map pattern is thus also significantly different from theirs. Although both systems are based largely upon presence and percentages of marble and quartzite and treat the phyllite pretty much as a uniform matrix for the granulose beds, Segerstrom's and Willard's extensive use of an "undifferentiated" unit appears to have greatly reduced the stratigraphic significance of their lithic subdivisions. Regardless of the time or facies relations between our Waits River and Gile Mountain Formations, we believe they are indeed valid lithostratigraphic units, unlike the subdivisions of the Conway used by Segerstrom (1956a,b) and Willard (1956).

East of the Mesozoic basins, three very small areas of Gile Mountain Formation and one larger area of Waits River Formation are shown on the State bedrock map (see also Robinson and others, Ch. D, this volume, figs. 2, 6, 14). These rocks contain beds of quartzite and of punky brown-weathering impure marble typical of the Gile Mountain and Waits River, respectively, and are distinctly different from the rocks of the Littleton, Erving, and other formations east of the Mesozoic basins. However, the exposures mapped as Gile Mountain Formation are too small to demonstrate convincingly that the proportions of constituent lithologies are correct for the Gile Mountain. Furthermore, because the three small patches are bounded on both sides by faults, little can be said about the stratigraphic position of the rocks involved.

Similarly, although the area mapped as Waits River Formation near Wilbraham on the east margin of the Hartford Mesozoic basin contains punky brown-weathering carbonate beds, these punky brown beds are no more than about a meter thick, whereas Waits River punky brown beds west of the basins are commonly 3-8 m thick. Secondly, Peper (1977) described the southern part of the Wilbraham "Waits River" as containing 15 percent "schist and granofels indistinguishable from schist and granofels of the Erving Formation." No such Erving-like rocks have been recognized in the main belt of the Waits River west of the Mesozoic basins. Thus, although the Wilbraham rocks are indeed like the Waits River in many ways, they are also significantly different from the Waits River in other ways. For this reason, and because we cannot document pre-Mesozoic continuity between the main belt of Waits River and the Wilbraham rocks, their formational assignment and their original stratigraphic position relative to the rocks west of the basins

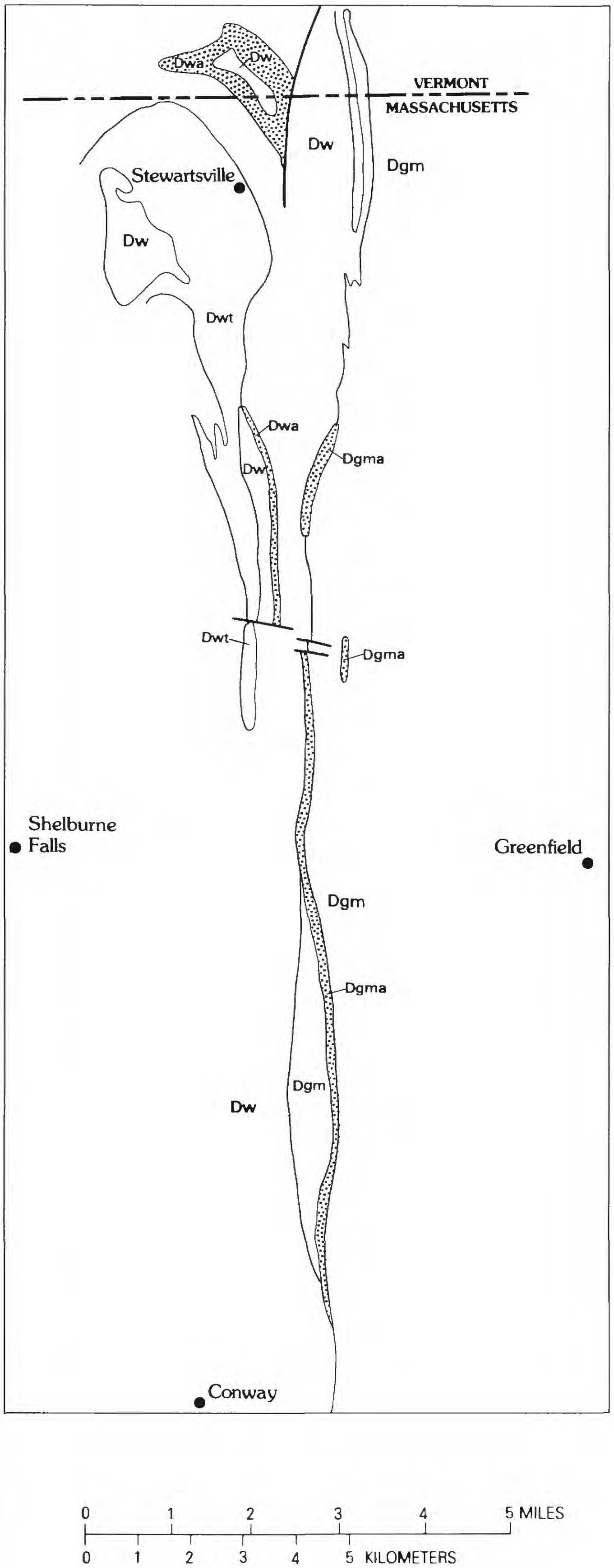

FIGURE 22.—Sketch map showing relations of metamorphosed volcanic rocks (patterned) to the Gile Mountain and Waits River Formations in northernmost Massachusetts. Symbols same as on State map: Dw and Dwt, metasedimentary members of the Waits River Formation; Dgm, metasedimentary rocks of the Gile Mountain Formation; Dwa, amphibolite apparently stratigraphically within and thus assigned to the Waits River Formation; Dgma, amphibolite either stratigraphically within the Gile Mountain or at the Gile Mountain-Waits River contact, both assigned to the Gile Mountain.

must be considered tentative. We do believe, however, that they are most reasonably assigned to the Waits River-Gile Mountain package.

PUTNEY VOLCANICS (Dpv)

The Putney Volcanics are exposed in Massachusetts in a small area at the north end of the Deerfield Mesozoic basin (fig. 10). These rocks were first mapped in Massachusetts by Balk (1956). He described them as beds and lenses of metatuff within his Leyden Argillite, which is approximately equivalent to the Littleton Formation of area 1 of this report. Doll and others (1961) in adjacent southern Vermont included these volcanic rocks in the Standing Pond Volcanic Member of the Waits River Formation, despite the fact that they showed them, as does the Massachusetts bedrock map, at the Gile Mountain-Littleton contact and totally isolated from the Waits River Formation. Trask (1964) also applied the name Standing Pond Volcanics to these rocks, but as a discrete formation, in northernmost Massachusetts and southernmost Vermont. Hepburn (1972a,b), however, in remapping southeasternmost Vermont, called these volcanic rocks between the Littleton and Gile Mountain Formations the Putney Volcanics in order to distinguish them from the Standing Pond (Hepburn, 1972b, p. 233). Trask (1980) then formally extended the use of the name Putney into Massachusetts. The following description is taken largely from Trask (1980) and Balk (1956).

The Putney consists of light-greenish-gray to white, very fine grained, poorly foliated phyllite to granulite in a discontinuous belt as much as 130 m wide. The rocks are composed of varied proportions of sodic plagioclase, quartz, sericite, clinozoisite, epidote, chlorite, and carbonate and are believed to represent mildly metamorphosed felsic tuff. These tuffs are locally interbedded with gray phyllite or slate.

As pointed out by Trask (1980), the Putney Volcanics are not physically traceable into the Standing Pond Volcanics of Vermont. Further, the amphibolites mapped as units Dgma and Dwa on the Massachusetts State bedrock map to the west, believed to be correlative with the Standing Pond, are much more mafic in composition. Hepburn (1972b) first noted that whereas the Standing Pond and correlatives, both in Vermont (Doll and others, 1961) and Massachusetts, are everywhere at or near the contact between the Waits River and Gile Mountain Formations, the Putney is everywhere at or near the Gile Mountain-Littleton contact. Until a much clearer understanding of the original stratigraphic relations among the Waits River, Gile Mountain, and Littleton rocks is achieved, it will not be possible to prove whether or not the Standing Pond (units Dwa and Dgma of Massachusetts) and the Putney were once continuous. In the interim, the Putney Volcanics and the amphibolites of the Gile Mountain and Waits River are maintained as two distinct stratigraphic units in Massachusetts.

Because it is sandwiched structurally or stratigraphically (see Robinson and others, Ch. D, this volume) between two Lower Devonian formations, an Early Devonian age is assigned to the Putney Volcanics.

STRATIGRAPHIC RELATIONS AMONG THE GOSHEN, WAITS RIVER, AND GILE MOUNTAIN FORMATIONS AND THE PUTNEY VOLCANICS

The Goshen, Waits River, and Gile Mountain Formations and the Putney Volcanics discussed above have all generally been considered to be younger than the Clough Quartzite and the Fitch and Shaw Mountain Formations, all of which have yielded Silurian fossils (Billings and Cleaves, 1934; Boucot and Thompson, 1958, 1963; Harris and others, 1983). Their *relative* ages and stratigraphic relations, however, have been the subject of continuing debate. This debate has centered in eastern Vermont where asymmetric belts of Waits River and Gile Mountain rocks are bounded on the west by the Northfield Formation and on the east by the generally similar Meetinghouse Slate. Although most workers have proposed that the Waits River and Gile Mountain are younger than the Northfield and Meetinghouse (see, for example, Doll and others, 1961), many combinations of facies changes, complex structures, and relative ages have been suggested to explain the observed regional distribution of the Waits River and Gile Mountain (see, for example, White and Jahns, 1950; Murthy, 1957, 1958, with discussion by White and Dennis, 1959; Doll and others, 1961; Eric and Dennis, 1958; Ern, 1963). The relations are still controversial in Vermont and thus do not offer any obvious resolution to the problem in Massachusetts. Fisher and Karabinos (1980) have shown from graded bedding that well-bedded Goshen-like rocks mapped as Gile Mountain Formation overlie Waits River beds at a group of exposures near Royalton, Vt., but more complex facies relations between the Gile Mountain and Waits River on a regional scale are certainly not excluded.

The Goshen Formation rests discordantly on the Hawley, Cobble Mountain, and Russell Mountain Formations along the west margin of the Connecticut Valley synclinorium in Massachusetts. These same relations continue northward into Vermont (Doll and others, 1961) where the basal Northfield (Goshen correlative) rests on the Cram Hill and Moretown Members of the Ordovician Missisquoi Formation (stratigraphic usage of Doll and others, 1961) and on lenses of the Silurian Shaw Mountain Formation. On the Massachusetts State bedrock map this unconformity is described as a "surface of Acadian structural disharmony" (see Hatch and Stanley, Ch. C, this volume) along which slippage with local displacement of as much as tens of kilometers may have taken place. However, we believe that in the general area of the presently exposed contact, the Goshen Formation was the basal Lower Devonian unit in Massachusetts. The minimum thickness of the Goshen in the area west of the Goshen dome (fig. 3) is estimated to be about 750 m (Hatch, 1968). As noted above, abundant graded beds in this area document an internal Goshen stratigraphy of, from bottom to top, Dg, Dgq, Dgu (fig. 23, column 2).

Other areas within the Connecticut Valley synclinorium in western Massachusetts in which Devonian stratigraphic sequences can be demonstrated are around the Shelburne Falls, Goshen, and Woronoco domes and on the flanks of the Whately anticline (fig. 3).

Even with no allowance for isoclinal folding, the thickness of the basal Dg unit of the Goshen overlying the core gneisses on the south side of the Shelburne Falls dome can be no more than about 170 m, and the upper two units (Dgq and Dgu) are both missing or at least not exposed. The progression from Dg to Dgl to Dw (Waits River) going north, east, or south off the dome strongly suggests a younging stratigraphic sequence in that order (fig. 23, column 3).

Although isoclinal folds are known to be present in the Goshen strata east of the Goshen dome, they cannot be mapped with sufficient accuracy to enable a confident estimate of thickness of the formation there. Unit Dg does, however, *appear* to be thicker here than around the Shelburne Falls dome. Because unit Dgl is not recognized east of the Goshen dome, the apparent stratigraphic sequence in that area is simply basal Goshen (Dg) up into Waits River (fig. 23, column 4).

These relations are complicated, however, by the available sedimentary tops data in the vicinity of the Goshen-Waits River contact in the Goshen (Hatch and Warren, 1982) and Westhampton (S.F. Clark, Jr., oral commun., 1979) quadrangles. Because of the gradational character of the contact between the two formations, locating stratigraphic tops data precisely at the contact has not been possible. In both quadrangles, however, most graded beds in the Goshen Formation

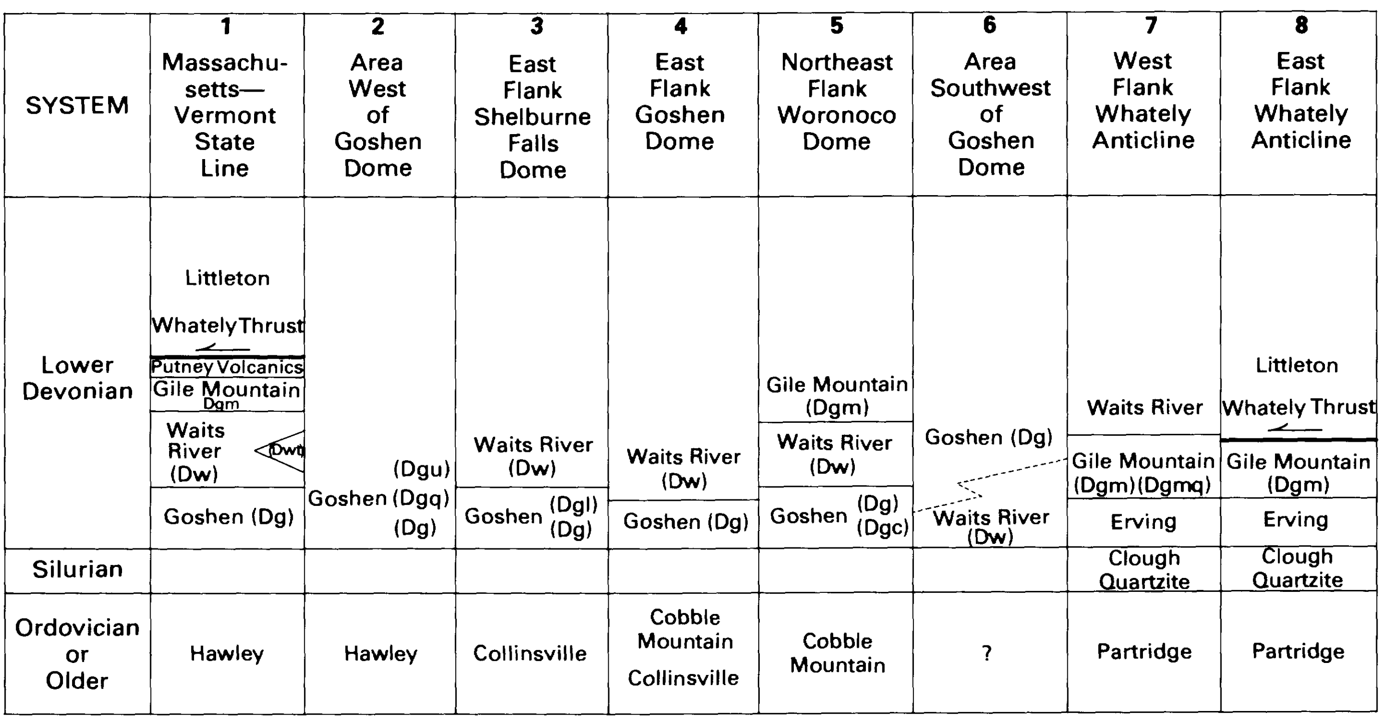

FIGURE 23.—Apparent stratigraphic columns above the pre-Silurian "basement" at various points within the western part of the Connecticut Valley belt. Named units are all of formation rank. Letter symbols refer to subdivisions of the formations. See text for discussion.

nearest to the contact indicate that the Waits River rocks are stratigraphically *below*, rather than above, the Goshen beds, at least in the vicinity of the presently exposed contact (fig. 23, column 6). We interpret these relations to result from major facies intertonguing of the two formations.

A somewhat different Devonian stratigraphic sequence appears to be present northeast from the patches of pre-Silurian rock exposed in the core of the Woronoco dome. Here the basal Devonian unit is Dgc. Although, as noted above, member Dgc appears to be a lithologic transition between the basal Goshen (Dg) and the Waits River Formation, Dg-type Goshen rocks crop out all around the exposed margin of member Dgc, and the quaquaversal dips and overall domal structure indicate that member Dg overlies member Dgc in this area. The Waits River borders unit Dg to the northeast. It is clear, therefore, that even though member Dgc is *lithologically* transitional between Dg and Dw, without otherwise unsupported structural complications it cannot lie stratigraphically between them. Northeast from the Woronoco dome, the apparent upward sequence is thus member Dgc, member Dg, Waits River, Gile Mountain (fig. 23, column 5).

In the Whately and Leeds areas, the Silurian to Lower Devonian sequence going westward from the Ordovician rocks in the center of the Whately anticline is Clough, Erving, members Dgmq and Dgm of the Gile Mountain Formation, and Waits River (fig. 23, column 7). Going eastward from the Ordovician rocks, the sequence is Clough, Erving, member Dgm of the Gile Mountain, and Littleton (fig. 24, column 8). A thrust fault, the Whately thrust, is shown on the State bedrock map along the Gile Mountain-Littleton contact. The nature of and rationale for this thrust are treated extensively in a separate chapter (Robinson and others, Ch. D, this volume).

It is apparent from the foregoing discussion and figure 23 that the sequence of lithofacies within the thick package of gray Lower Devonian metasedimentary rocks varies markedly from place to place around the synclinorium. Although these rocks are complexly and multiply deformed (see Hatch and Stanley, Ch. C, this volume), we believe that, with the exception of the Whately thrust (fig. 23, columns 1 and 8), the relations discussed above are largely the result of primary sedimentary facies changes.

In support of the facies model of the Goshen, Waits River, and Gile Mountain Formations, rather than a model based on thrust faults or other complex structures, we point out the following: (1) In contrast to the Littleton-Gile Mountain contact, which is relatively sharp, the Goshen-Waits River and Waits River-Gile Mountain contacts are gradational over intervals of tens of meters or more. We interpret these relations to suggest that the Goshen, Waits River, and Gile Mountain were probably deposited in contact with each other, whereas the Littleton west of the Mesozoic basins is believed to have been emplaced by later tectonism from a depositional site to the east (Robinson and others, Ch. D, this volume). (2) The Waits River and Gile Mountain Formations, in particular, are both characterized by the same rock types and are distinguished largely by the different proportions of those rock types. This lithologic similarity suggests to us that they started out as sedimentologically intercalated units, presumably in the same depositional basin. If they have subsequently been cut by thrusts, these thrusts have only compounded the complexities, which still derive primarily from sedimentary facies intertonguing. (3) If the Goshen-Waits River and Waits River-Gile Mountain contacts are thrusts, these thrusts continue all across Vermont into Quebec with remarkably similar gradational (minor) changes both along and across the slices and in the proportions of the constituent rock types in the slices. Again, these relations seem better explained as resulting primarily from original facies changes.

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Post-Taconian Structural Geology of the Rowe-Hawley Zone and the Connecticut Valley Belt West of the Mesozoic Basins

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THE BEDROCK GEOLOGY OF MASSACHUSETTS

U.S. GEOLOGICAL SURVEY PROFESSIONAL PAPER 1366-C

CONTENTS

ILLUSTRATIONS

POST-TACONIAN STRUCTURAL GEOLOGY OF THE ROWE-HAWLEY ZONE AND THE CONNECTICUT VALLEY BELT WEST OF THE MESOZOIC BASINS

By Norman L. Hatch, Jr., and Rolfe S. Stanley[1]

ABSTRACT

The Acadian structural history of the Rowe-Hawley zone and western Connecticut Valley belt is complex. Three distinct generations of folds are present. The first, F_2, formed large, high-amplitude isoclinal folds, whose axial surfaces parallel the regional schistosity in the Goshen Formation, and small isoclines with axial planar schistosity in the pre-Silurian rocks. The basal Devonian surface is described on the State bedrock map as a "surface of Acadian structural disharmony." F_3 and F_4 produced widespread minor crenulate or open folds with strong axial-surface cleavage but only very local map-scale folds. Pre-Silurian gneiss-cored domes, in a north-south chain, deform F_2 isoclines in the Lower Devonian Goshen Formation blanket. The domes are interpreted as having started to form after F_2 time by gravitational rise of the core gneiss and having subsequently been compressed and molded by both F_3 and F_4. Ramsay-type fold interference is considered an unlikely major mechanism, although it is responsible for some of the present configuration of the Shelburne Falls and Granville domes. The Granville and Prospect Hill thrusts both developed in the southern part of the area relatively early in the Acadian orogeny approximately synchronous with F_2 folding. Although the dominant structural transport direction was east over west, pronounced west-over-east "backfolding" in the southern part of the area is believed to have formed largely during F_3 time as a result of the incipient underthrusting of weaker eastern crust beneath more resistant tectonically thickened western crust.

INTRODUCTION

Soon after deposition of the Lower Devonian rocks described in an earlier chapter (Hatch and others, Ch. B, this volume), all of the rocks of the Connecticut Valley belt and of the Rowe-Hawley and Bronson Hill zones upon which they were deposited (Hatch and others, 1984) were multiply deformed in a series of events attributed to the Acadian orogeny. This chapter describes and discusses these Acadian events in the Rowe-Hawley zone and the Connecticut Valley belt west of the Mesozoic basins in Massachusetts. In the Connecticut Valley belt, the Acadian orogeny was the first to affect the rocks, and thus the deformational history for that area begins at this point. In the Rowe-Hawley zone, however, the Acadian events were imposed on rocks that had already been faulted, folded, and metamorphosed during the Taconian orogeny. Some of the following discussion thus builds upon and assumes a familiarity with the discussion of the pre-Silurian rocks and Taconian deformation in the chapter on the Rowe-Hawley zone (Stanley and Hatch, Ch. A, this volume). In the remainder of this chapter, references to the Connecticut Valley synclinorium (or belt) should be taken to mean that part of the belt west of the Mesozoic basins.

The recognized effects of the Acadian orogeny in the Rowe-Hawley zone and western Connecticut Valley belt of Massachusetts include at least three discrete widespread episodes of folding and associated cleavage development, thrust faulting, doming, a "surface of Acadian structural disharmony," regional metamorphism, and intrusion of the Middlefield Granite and the Williamsburg Granodiorite. Some aspects of the tectonic history of these areas have been discussed previously (Hatch, 1975; Stanley, 1975; Norton, 1975; Osberg, 1975). These earlier described aspects will only be summarized here; the emphasis of this chapter will be on more recent ideas, aspects of the structure not previously described, and an attempt to bring all of our current ideas together into one coherent synthesis.

Some of the members and submembers of formations mapped on the State bedrock map have not been formally named. To avoid a proliferation of unnecessary

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new names, these units may be referred to in this chapter only by their letter symbol designations on the State bedrock map. Figure 1 is simplified from the State bedrock map. It shows formations, but not members thereof, in the Rowe-Hawley zone and the western Connecticut Valley belt and geographic features referred to in the following text. It is designed as a supplement to, but not a substitute for, the State bedrock map while reading this chapter.

ACKNOWLEDGMENTS

Our interpretation of the structure of the Rowe-Hawley zone and western Connecticut Valley belt has been influenced by and has benefited from our close association with many colleagues. Although we believe that most of them agree with most of what we say here, many disagree with us on one or more points. These disagreements have stirred us, over the years, to re-look, re-think, and revise, and we thank these colleagues for their efforts, their persistence, and their good spirit. Chief among them are Stewart F. Clark, Jr., Leo M. Hall, and Philip H. Osberg. Many hours of field excursions and discussion with Nicholas M. Ratcliffe have helped us sort out the Taconian from the Acadian fabrics along the western part of the area, and comparable cooperative efforts with Peter Robinson have significantly improved our understanding of the Acadian orogeny across the Connecticut Valley belt. Thoughtful and thorough reviews by Leo Hall and Philip Osberg improved earlier drafts of this paper.

ACADIAN FOLD GENERATIONS

Three regionally persistent Acadian fold generations have been recognized throughout western Massachusetts between the Mesozoic basins and the east flank of the Berkshire massif (Hatch and others, 1967; Hatch, 1975; Stanley, 1975). All of these reports recognized an earlier, Taconian generation of folds designated F_1 and thus applied the terms F_2, F_3, and F_4 to the Acadian generations. For consistency and compatibility with the earlier papers, we will here follow that nomenclature. Osberg (1975) proposed an additional, pre-F_2 Acadian fold generation as a mechanism to explain inverted F_2 folds in the area of the Shelburne Falls dome. Although we disagree with Osberg on the magnitude of this structure, we do agree that smaller pre-F_2 folds are present locally, as, for example, in the basal contact of the Goshen Formation along the north side of the Shelburne Falls dome (Leo M. Hall, written commun., 1977; Zen and others, 1983). However, we interpret these as having formed locally during the early stages of east-over-west movement of the Devonian cover, which culminated, in this area, with the formation of the major F_2 isoclines.

The chronology and correlation of the Acadian fold generations are based on relations of superposed minor folds and associated cleavages that have been recognized and correlated by incremental analysis throughout western Massachusetts and Connecticut (Stanley, 1975). These generations and the methodology of analysis have been described in the papers just cited in U.S. Geological Professional Paper 888 as well as in many of the quadrangle maps referred to in Zen and others (1983) (hereafter referred to as the State bedrock map) and, therefore, are only briefly described here. We wish to point out, however, that the outcrop-scale folds for each fold generation have been correlated to major map-scale folds, which thus provides a relative chronology to which to relate major events such as faults, igneous intrusions, and metamorphism. More recently Ratcliffe and Hatch (1979, fig. 5) have extended this system of fold generations westward through the Berkshire massif and the Taconic allochthons. This chronology of events provides a basis for reconstructing the evolution of Acadian deformation.

Figures 2, 3, and 4 are summaries of the orientation of the axial surfaces of F_2, F_3, and F_4 minor and major folds. These data are interpolated into trend surfaces. The line length and spacing reflect the data population.

F_2 folds are conspicuous and widespread throughout the Connecticut Valley belt and Rowe-Hawley zone. They are tight to isoclinal, and the dominant regional schistosity is everywhere parallel to their axial surfaces. Throughout the two regions, other than on the flanks of the domes, F_2 axial surfaces trend north and dip very steeply. Because the folds are predominantly isoclinal, their axial-plane schistosity parallels beds in most outcrops. F_2 folds are identified by the fact that the dominant regional schistosity cuts their hinges, parallels their axial surfaces, and is not deformed by them. In the pre-Silurian rocks of the Rowe-Hawley zone, F_2 folds are generally small and have amplitudes measurable in centimeters to a few meters (figs. 5, 6). In the Rowe Schist and the adjacent Hoosac Formation, their axial surface schistosity is parallel or subparallel to schistosity of Taconian age traced eastward from the Berkshire zone (Ratcliffe and Hatch, 1979; Stanley and Hatch, Ch. A, this volume). The westward limit of true Acadian S_2 schistosity is thus difficult to ascertain, but we interpret it to extend at least to the western edge of the Rowe-Hawley zone. In the Connecticut Valley strata, F_2 isoclines have amplitudes measurable in tens of meters to kilometers (fig. 7) (Hatch, 1968, 1975; State bedrock map cross section *D-D'*). S_2 schistosity is pervasive and ubiquitous in these rocks.

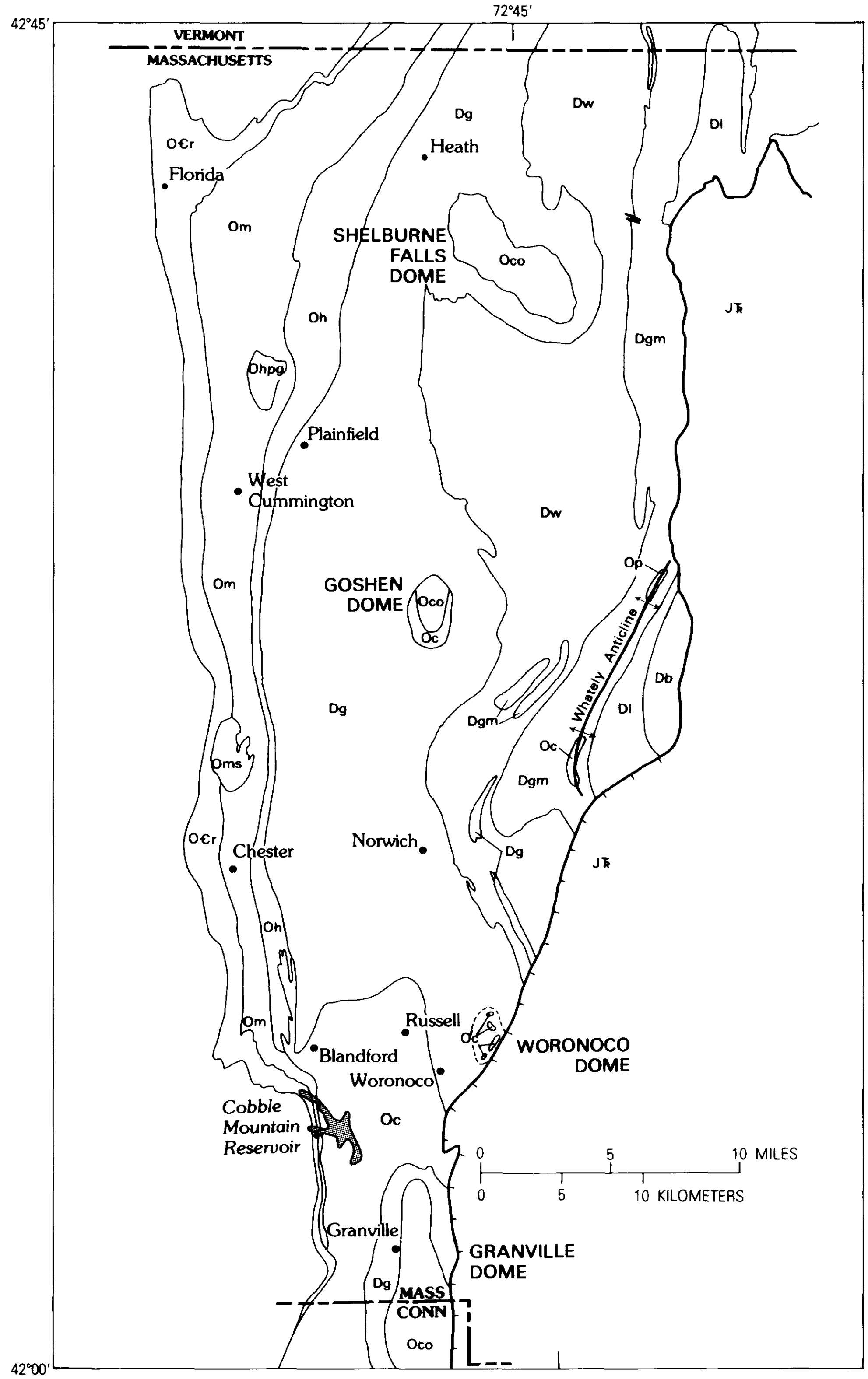
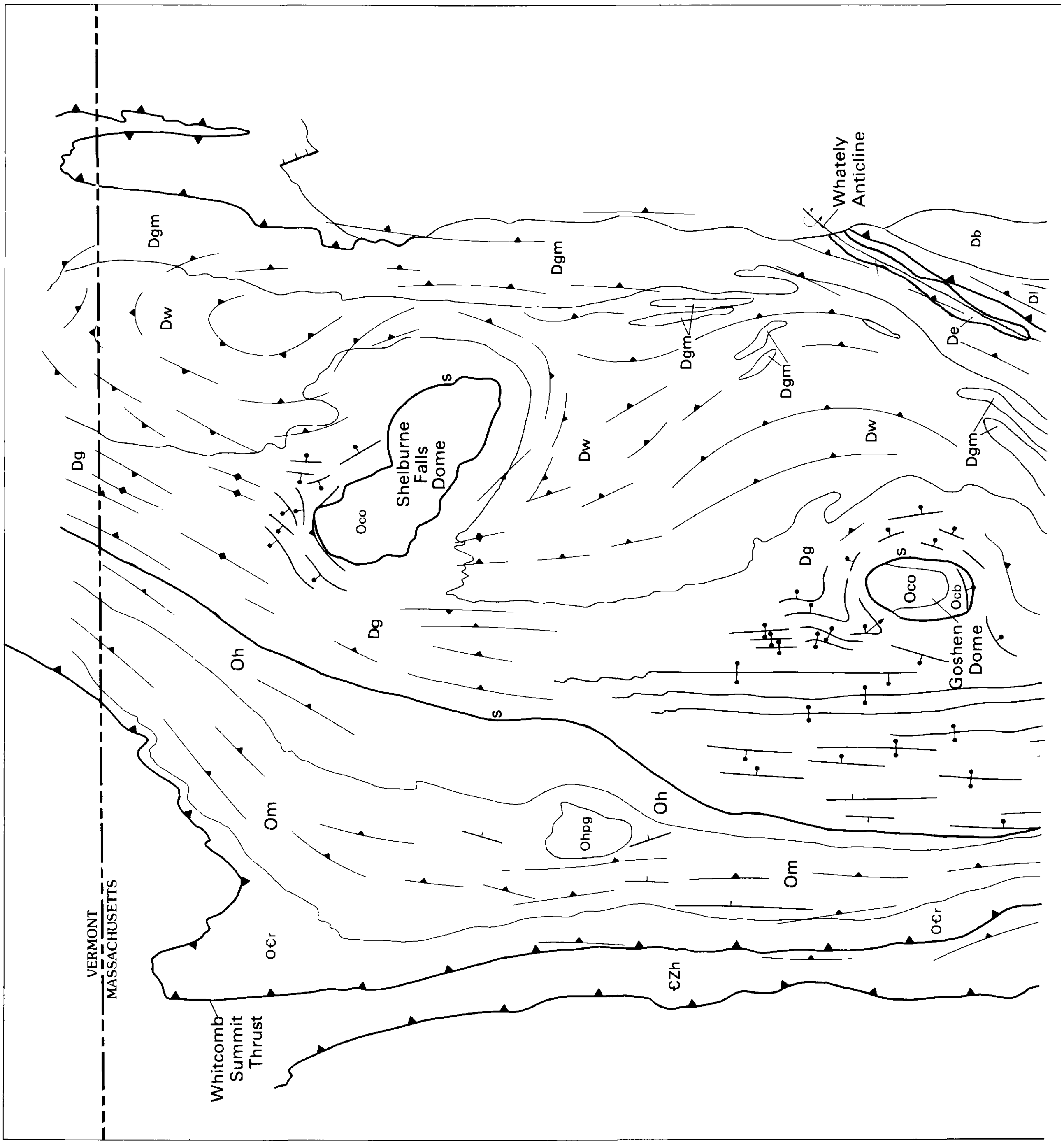


FIGURE 1.—Geologic map of the Rowe-Hawley zone and the western part of the Connecticut Valley belt, simplified from the State bedrock map (Zen and others, 1983). Explanation of letter symbols: JꞦ, undivided Jurassic and Triassic rocks; Db, Belchertown Complex; Dmg, Middlefield Granite; Dl, Littleton Formation; Dgm, Gile Mountain Formation; Dw, Waits River Formation; Dg, Goshen Formation; Ohpg, gneiss at Hallockville Pond; Oc, Cobble Mountain Formation; Oh, Hawley Formation; Oco, Collinsville Formation; Om, Moretown Formation; O€r, Rowe Schist.


Whately Anticline
Shelburne Falls Dome
Goshen Dome
Whitcomb Summit Thrust
VERMONT
MASSACHUSETTS
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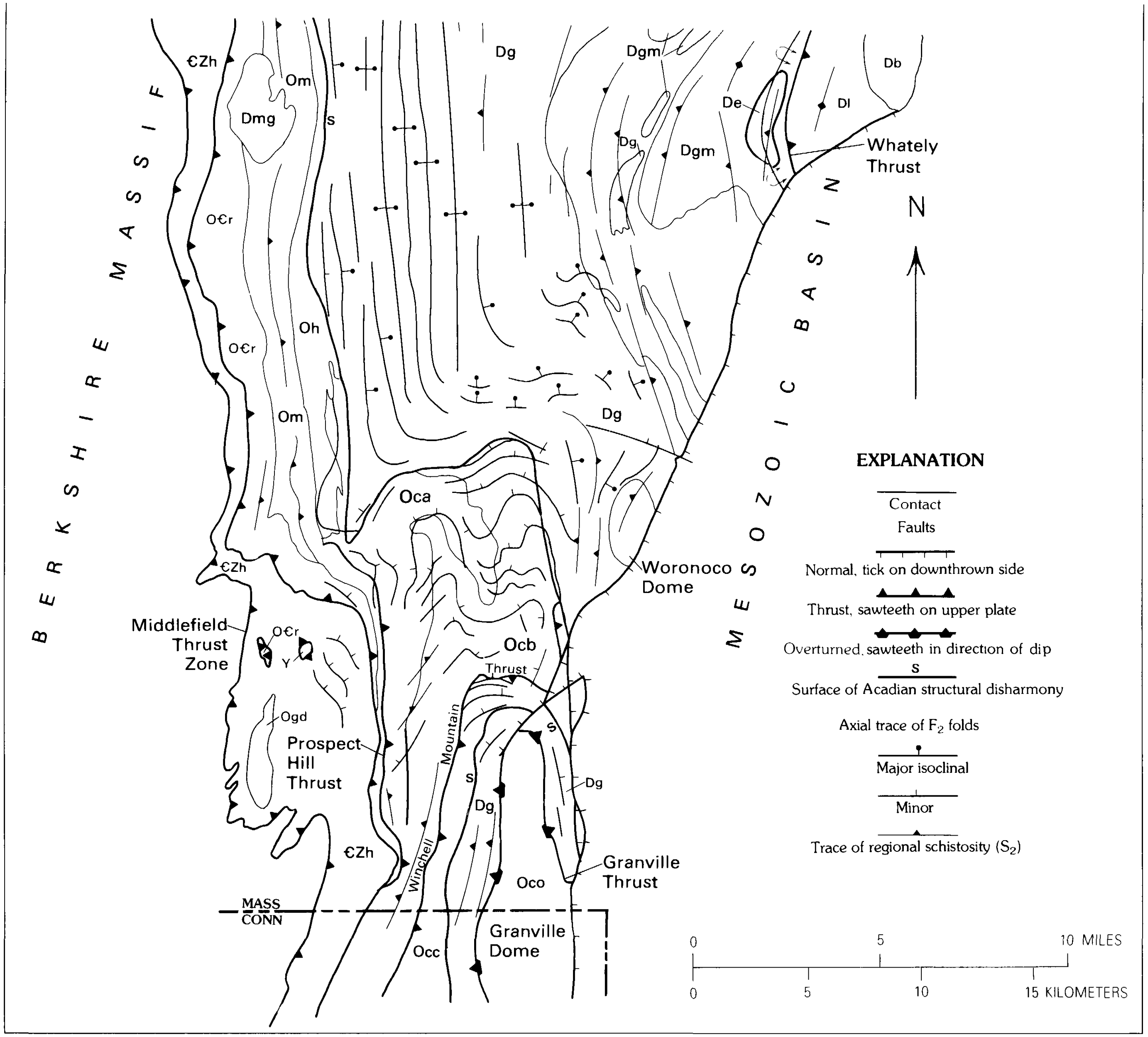
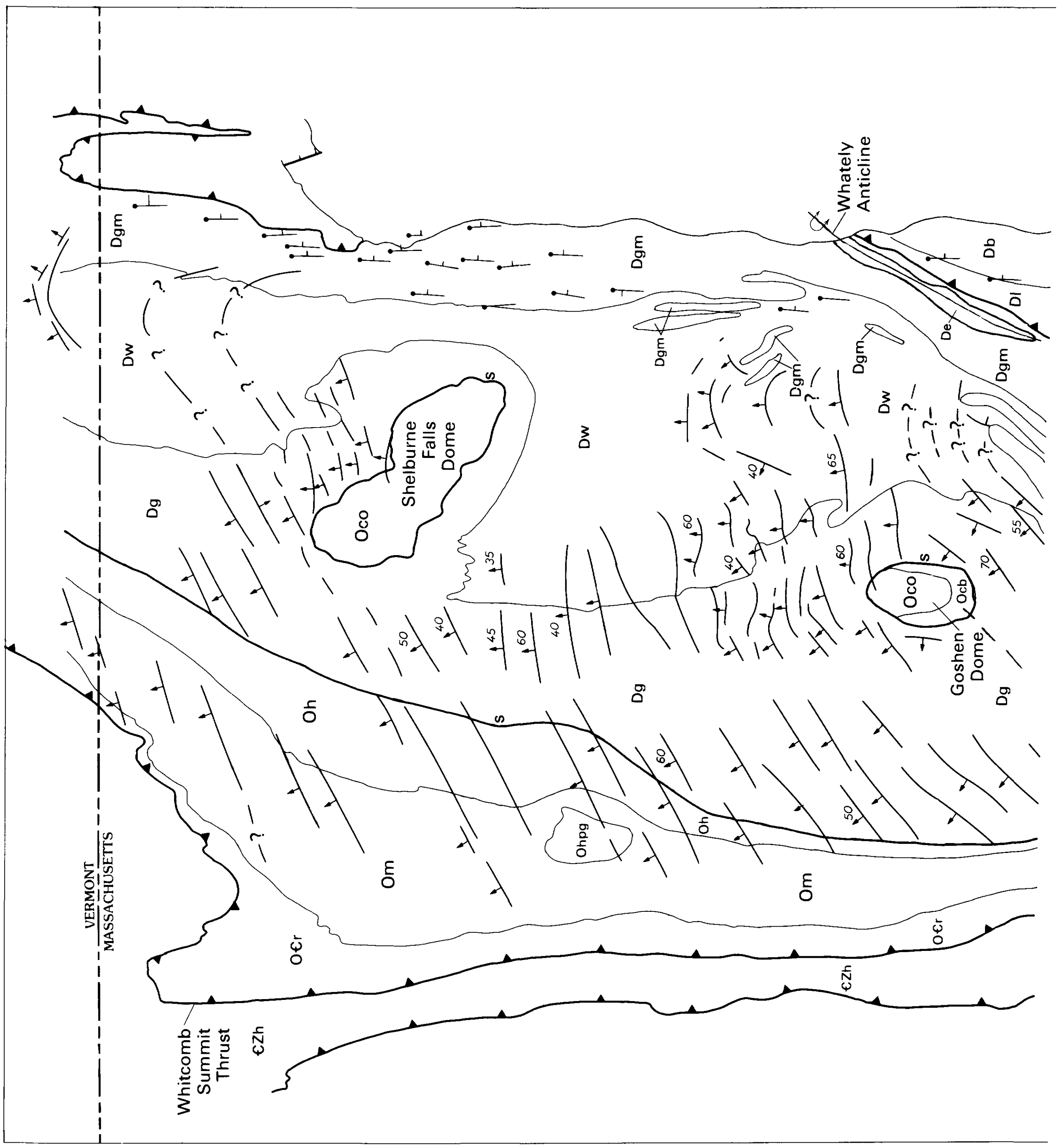


FIGURE 2.—Trend surface map showing the distribution and orientation of the axial surfaces of F_2 minor and major (isoclinal) folds and associated axial-surface schistosity. Explanation of letter symbols: Dmg, Middlefield Granite; Db, Belchertown Complex; De, Erving Formation; Dl, Littleton Formation; Dgm, Gile Mountain Formation; Dw, Waits River Formation; Dg, Goshen Formation; Ohpg, gneiss at Hallockville Pond; Ogd, diorite at Goff Ledges; Oca, Ocb, Occ, members of the Cobble Mountain Formation; Oco, Collinsville Formation; Oh, Hawley Formation; Om, Moretown Formation; O€r, Rowe Schist; €Zh, Hoosac Formation; Y, Proterozoic Y rocks.



VERMONT
MASSACHUSETTS
Whitcomb Summit Thrust
Whately Anticline
Shelburne Falls Dome
Goshen Dome
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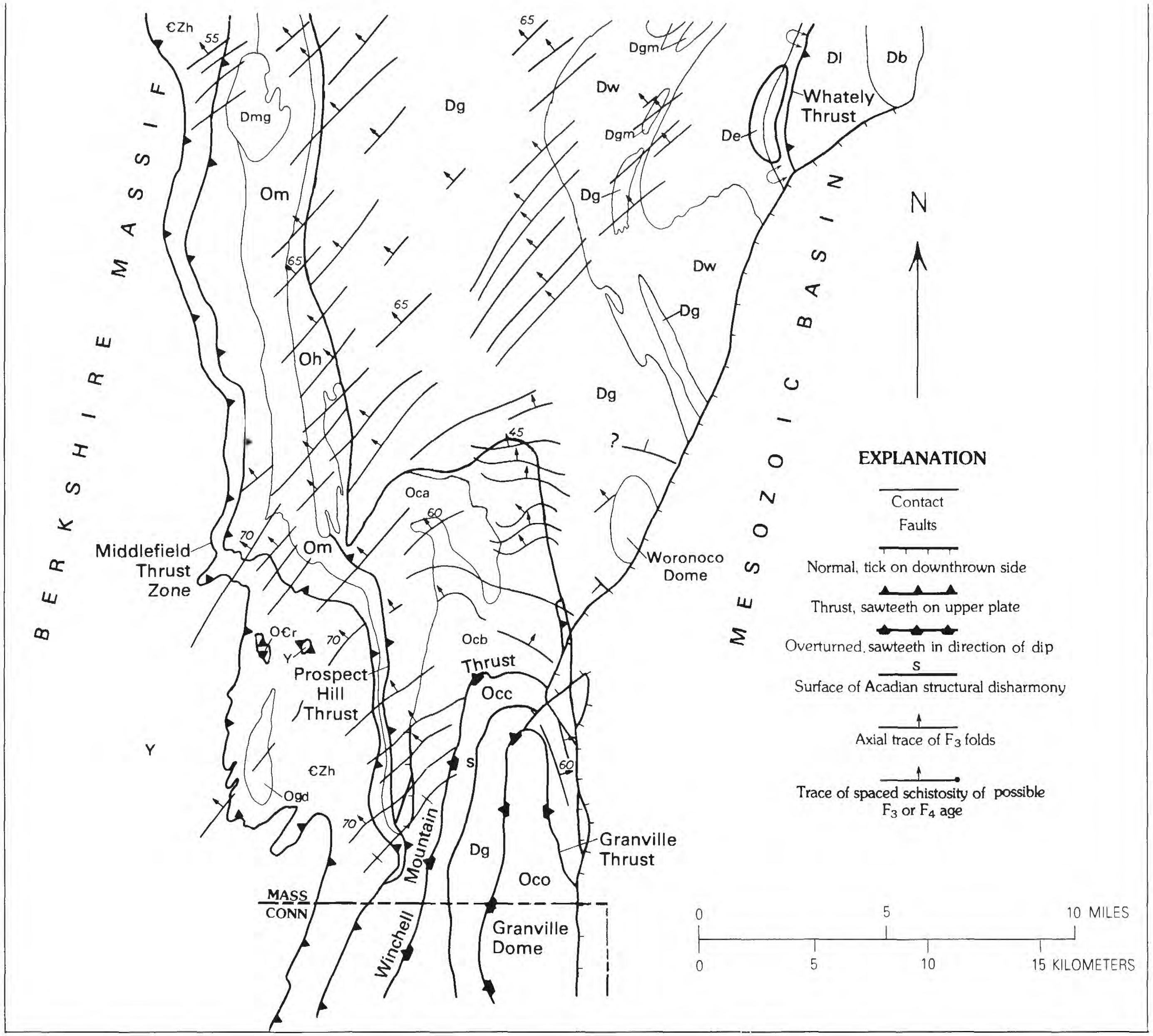
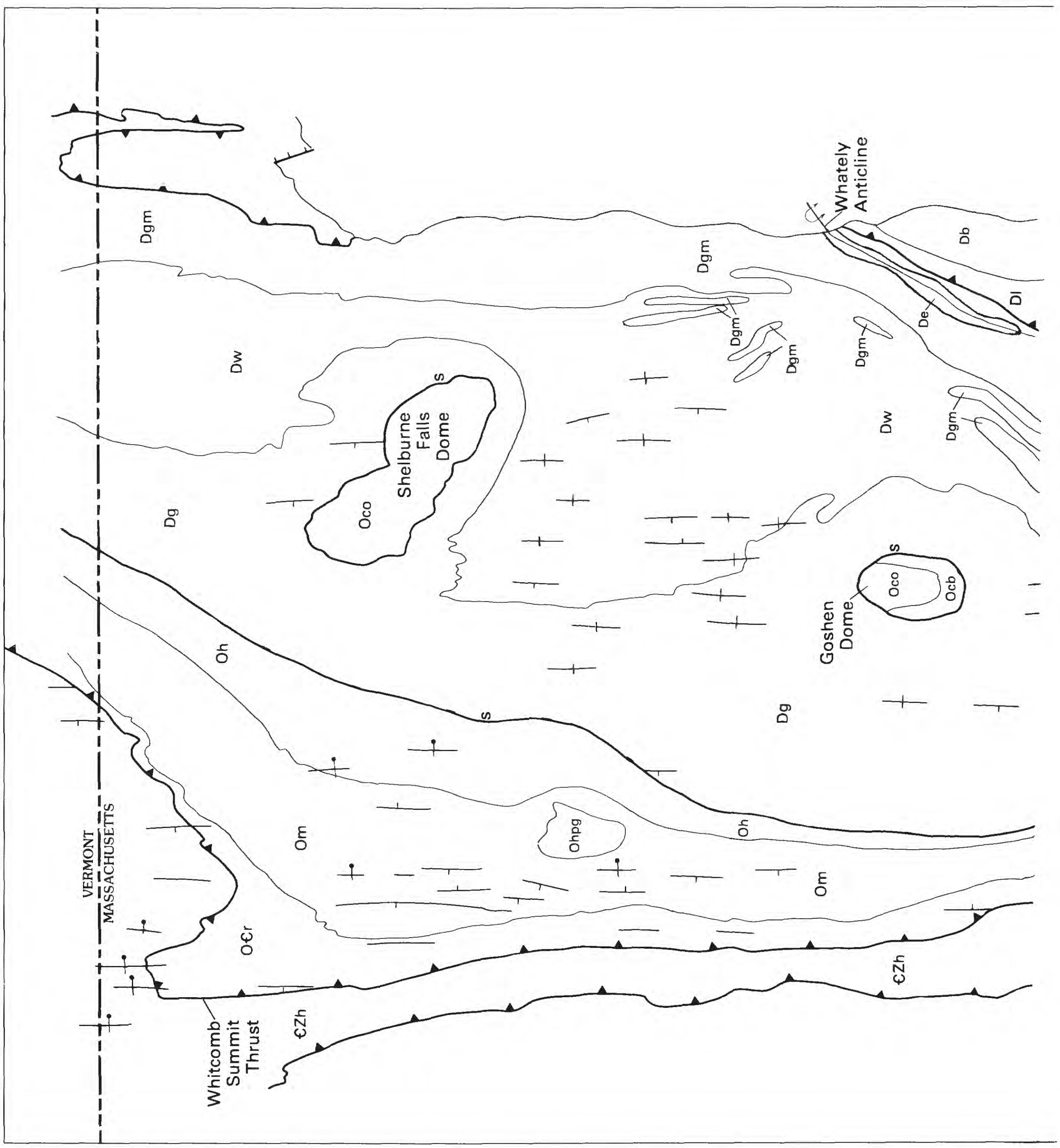


FIGURE 3.—Trend surface map showing the distribution and orientation of the axial surfaces of F_3 minor and major folds and associated axial-surface cleavage. Explanation of letter symbols: Dmg, Middlefield Granite; Db, Belchertown Complex; De, Erving Formation; Dl, Littleton Formation; Dgm, Gile Mountain Formation; Dw, Waits River Formation; Dg, Goshen Formation; Ohpg, gneiss at Hallockville Pond; Ogd, diorite at Goff Ledges; Oca, Ocb, Occ, members of the Cobble Mountain Formation; Oco, Collinsville Formation; Oh, Hawley Formation; Om, Moretown Formation; O€r, Rowe Schist; €Zh, Hoosac Formation; Y, Proterozoic Y rocks.



VERMONT
MASSACHUSETTS
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Shelburne Falls Dome
Goshen Dome
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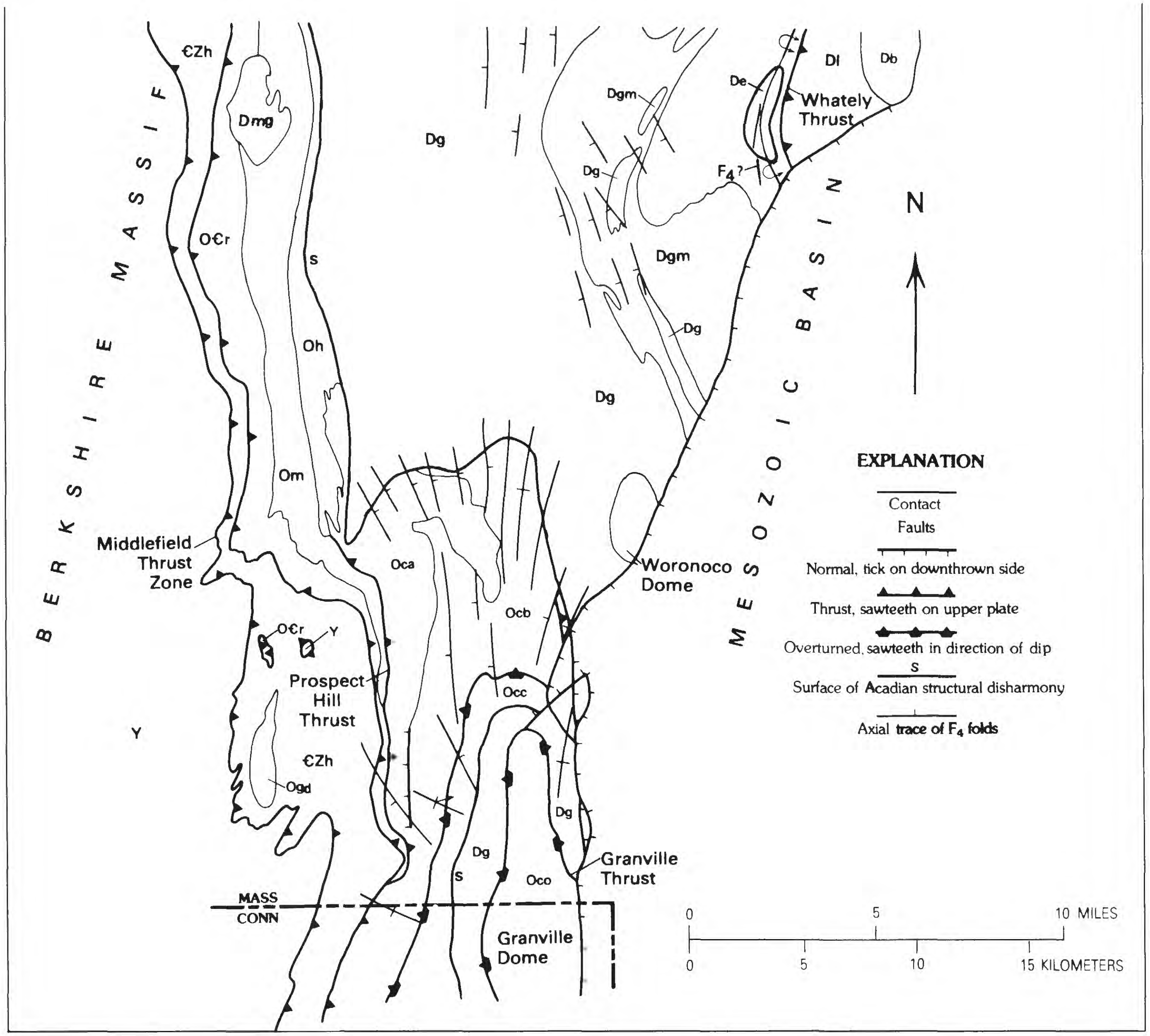
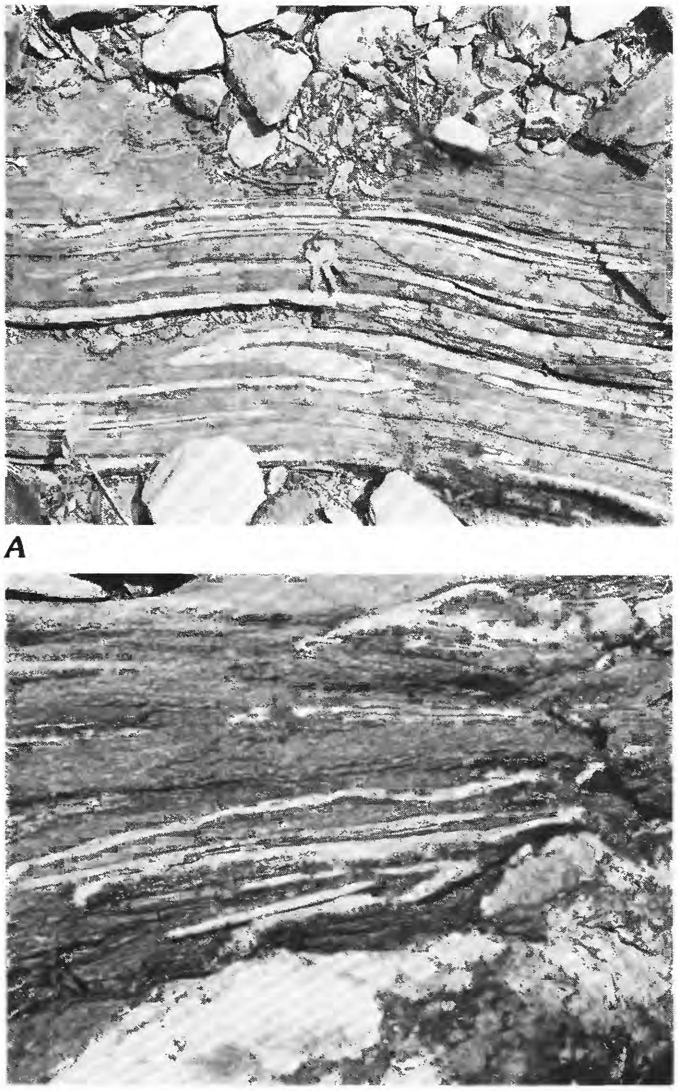


FIGURE 4.—Trend surface map showing the distribution and orientation of the axial surfaces of F_4 minor and major folds and associated axial-surface cleavage. Explanation of letter symbols: Dmg, Middlefield Granite; Db, Belchertown Complex; De, Erving Formation; Dl, Littleton Formation; Dgm, Gile Mountain Formation; Dw, Waits River Formation; Dg, Goshen Formation; Ohpg, gneiss at Hallockville Pond; Oca, Ocb, Occ, members of the Cobble Mountain Formation; Oco, Collinsville Formation; Oh, Hawley Formation; Om, Moretown Formation; O€r, Rowe Schist; €Zh, Hoosac Formation; Y, Proterozoic Y rocks.

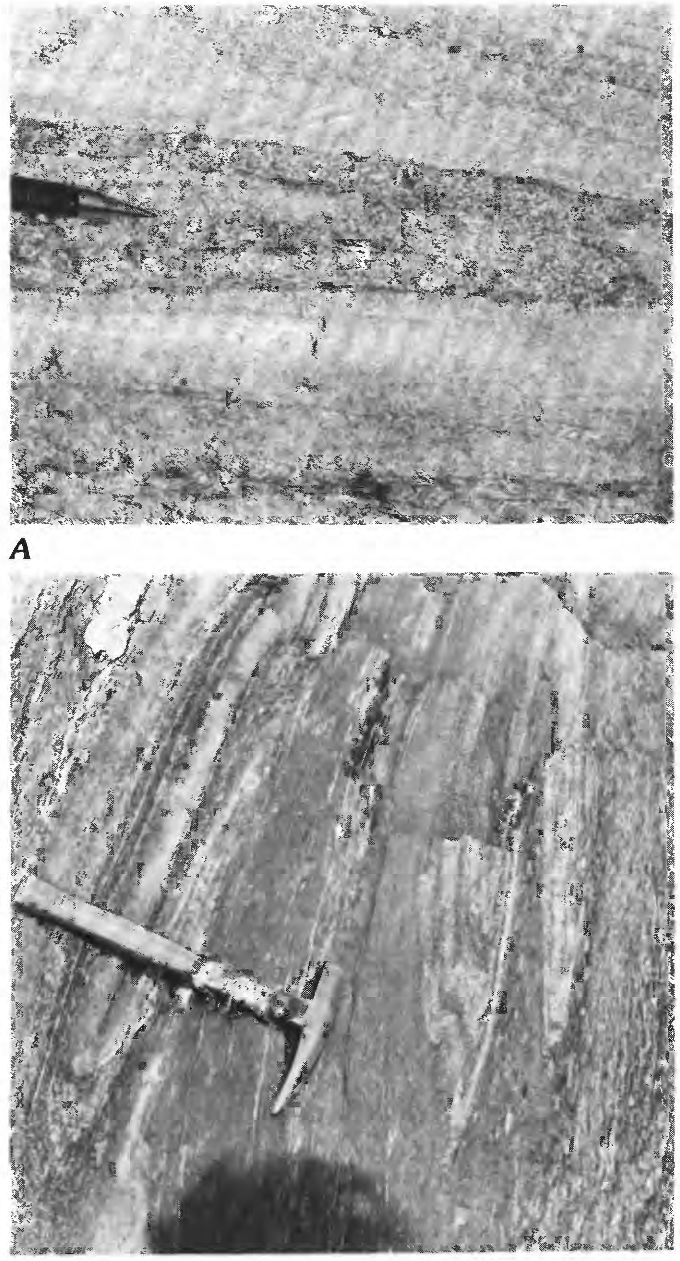


A

B

FIGURE 5.—F_2 minor isoclinal folds with penetrative S_2 schistosity in thinly laminated amphibolites of member A of the Cobble Mountain Formation, Cobble Mountain reservoir. Keys are about 5 cm long. Pencil is about 15 cm long.

The axial surface schistosity (S_2) of F_2 folds is parallel to the Prospect Hill thrust, the Granville thrust, the basal Devonian contact, and the axial surface of the Woronoco fold, all of which are discussed below. S_2 parallels the pre-existing fabric of the older Taconian thrusts (Middlefield thrust zone, Whitcomb Summit thrust, Rowe thrust zone, and Winchell Mountain thrust) (Stanley and Hatch, Ch. A, this volume; figs. 1, 2). F_2 axial surfaces and S_2 schistosity are clearly deformed by the Shelburne Falls, Goshen, Woronoco, and Granville domes as well as by F_3 and F_4 (fig. 2). The Williamsburg Granodiorite commonly forms sills parallel to S_2 and is generally unfoliated, suggesting a post-F_2 intrusion age.



A

B

FIGURE 6.—F_2 minor isoclinal folds in well-bedded schists and granulites of member A of the Cobble Mountain Formation, Cobble Mountain reservoir. Note that the penetrative schistosity is parallel to the axial surfaces of the folds and is not deformed by the folds. Steel point of pencil is about 2 cm long. Hammer handle is about 35 cm long.

F_3 folds and their associated axial plane cleavage are also well developed throughout both the Rowe-Hawley zone and the western Connecticut Valley belt (Hatch, 1975; Stanley, 1975) as well as in at least the eastern part of the Taconic-Berkshire zone to the west (Norton, 1975; Ratcliffe and Harwood, 1975). Axial surfaces of large and small folds of this generation have a consis-

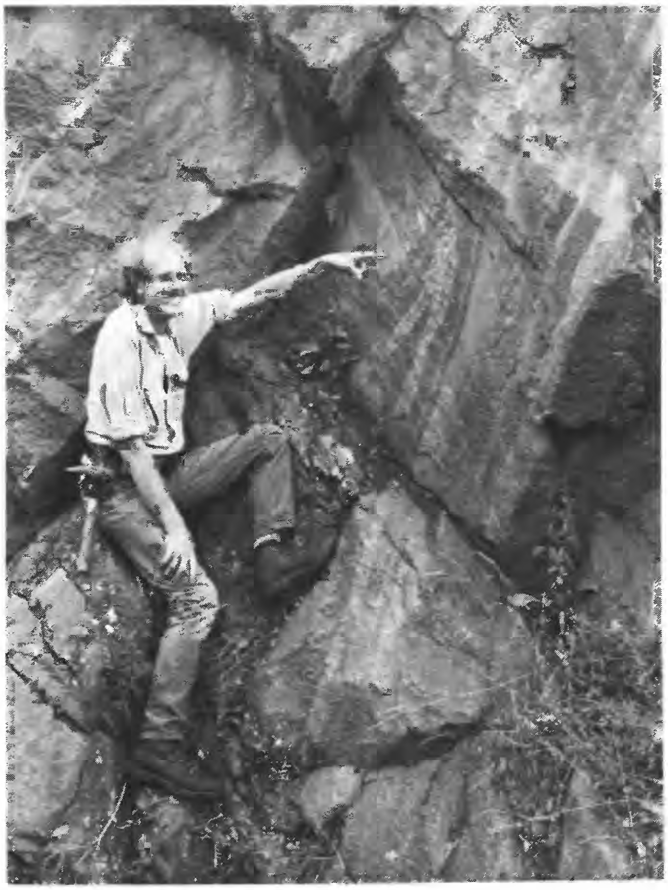


FIGURE 7.—Senior author points to the hinge of a major F_2 isocline in member Dg of the Goshen Formation, Route 9, 500 m west of Cummington. Graded beds indicate that this anticline is right side up. This fold has been traced more than 30 km along strike and is manifested on the State bedrock map as a narrow belt of Goshen unit Dg flanked on both sides by belts of Dgq.

tent northeast trend and northwest dip (fig. 3). The folds are typically crenulate in form (fig. 8). Their axial plane cleavage is generally a slip, crenulate, or spaced cleavage, and only locally is it well enough developed to be called a schistosity. The north-plunging fold axes consistently have a counterclockwise or east-over-west sense.

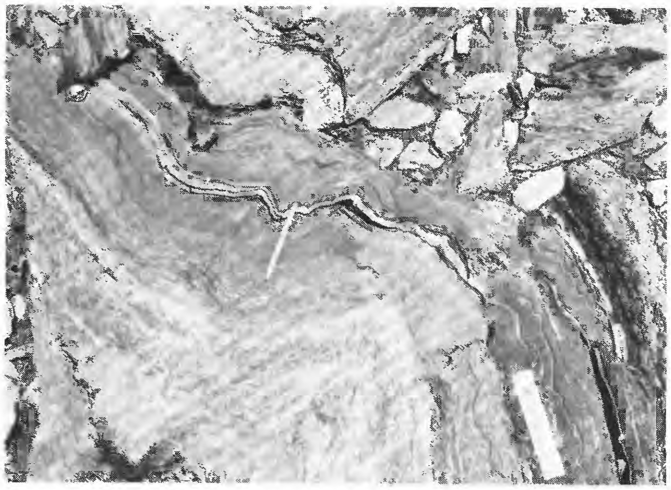
In the eastern part of the Connecticut Valley belt, in the Gile Mountain Formation, a very well developed, vertical, north-south, spaced schistosity strongly overprints an older, parallel schistosity (S_2). Although the data are far from complete, the gradual change of S_3 into an east-west orientation in the east-central part of the area of figure 3 suggests that S_3 may have been further rotated by a younger fold to produce the present north-south spaced schistosity in the Gile Mountain Formation. An alternative and to us more attractive interpretation is that this north-south spaced schistosity is not the same age as S_3 to the west. A comparison of the trend surface maps of F_2 and F_3 (figs. 2, 3) supports this interpretation because S_2 is not folded into a north-plunging anticline as it should be if S_3 is the same age as the north-south spaced schistosity of the Gile Mountain Formation. Instead, S_2 strikes northward with little or no deflection. Available field data therefore indicate that the Gile Mountain spaced schistosity is younger than S_3.

F_4 folds are far less abundant and pervasive than are folds of the older generations (fig. 4). They are well developed and easily recognized where S_3 is strongly deflected (compare figs. 3 and 4), particularly in the Connecticut Valley belt. In the Rowe-Hawley zone, S_3 is somewhat less widespread and S_4 is locally superposed directly on S_2. Axial surfaces of S_4 folds commonly trend northward and are vertical, and the folds have horizontal or north-plunging hinges.

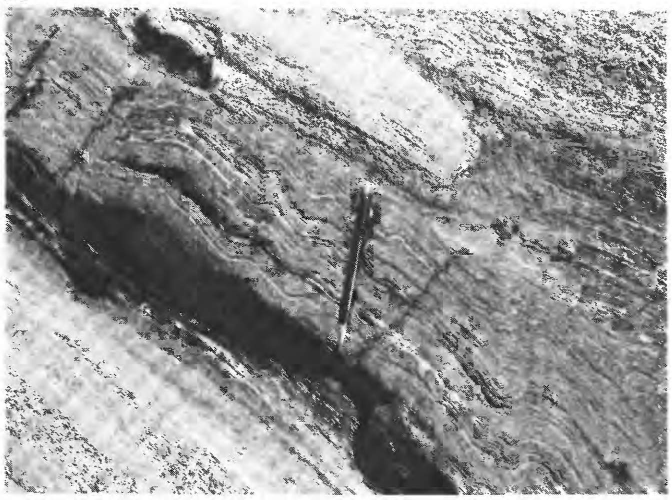
F_4 folds have been described by Stanley (1975) from the Woronoco area where they clearly deform F_3 as well as F_2 cleavages. In this area, they have consistent moderate to steep north plunges. F_4 folds from the northern part of the Rowe-Hawley zone and Connecticut Valley belt have steeply dipping north-trending axial surfaces and subhorizontal or gently north-plunging hinges; they are locally associated with a well-developed axial plane slip cleavage (fig. 9; Hatch and others, 1967; Osberg and others, 1971; Hatch, 1969, 1975; Hatch and Warren, 1981).

In this area, F_4 is thought to be responsible for deflections of S_3 cleavage on a map scale (Hatch, 1981; Hatch and Warren, 1981) and for folds in beds and in S_2 schistosity, but only rarely can it be demonstrated to deform S_3 structures in outcrop (see, for example, Osberg and others, 1971, fig. 2). There is no doubt of the existence of a post-F_3 folding event in the Woronoco area and of a post-F_3 folding event in northern Massachusetts. We have previously implied a correlation of these two events (Stanley, 1975; Hatch, 1975), and subsequent reconnaissance mapping in the eastern part of the Connecticut Valley belt and detailed mapping in the Ashfield (Hatch, 1981), Goshen (Hatch and Warren, 1981), and Westhampton (S.F. Clark, Jr., unpub. data, 1977) quadrangles has further confirmed that the two are indeed the same and can therefore be called F_4.

Around the northern part of the Granville dome, F_4 axial surfaces form a radial pattern varying systematically in trend from northeast to northwest (fig. 4). The similarity in orientation of S_4 and of the spaced schistosity in the Gile Mountain Formation, described two paragraphs above, suggests that they are the same age. If that is true, the Whately anticline is probably F_4 in age rather than F_2 (fig. 4). It is clear from cross sections *A-A′* and *D-D′* on the State bedrock map that



A



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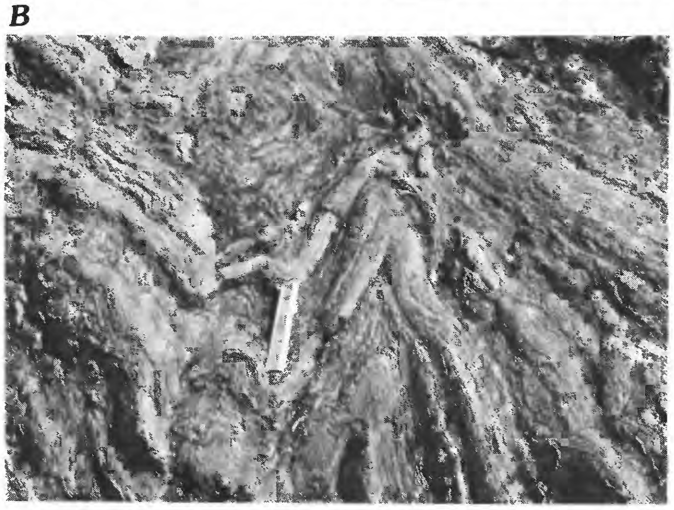


C

FIGURE 8.—Typical F_3 minor folds in thinly bedded rocks of member A of the Cobble Mountain Formation, Cobble Mountain reservoir. Note that the dominant (F_2) schistosity is folded by these folds and produces a crenulation cleavage (S_3). Pencils are 14 cm long. Ruler is 17 cm long. All views looking northeast.

A

B

FIGURE 9.—*A*, F_4 folds in well-bedded rocks of the Moretown Formation at West Cummington. *B*, Well-developed spaced cleavage (parallel to pen) is parallel to axial surfaces of F_4 folds and cuts bedding at about 30° in same outcrop as *A*. Pen is 14 cm long.

the Whately anticline folds the eastern projection of the base of the Goshen Formation and thus folds the F_2 isoclines of the Goshen to the west. The Whately anticline would thus have to be younger than F_2 and could be considered coeval with the strong vertical spaced schistosity in the Gile Mountain belt that is in turn parallel to F_4. Consequently, we show the Whately anticline as a possible F_4 structure on figure 4.

Simpson (1974), following the initial proposal by Osberg (1972), presented a different interpretation of the various cleavages and schistosities in the Goshen Formation strata mantling the Shelburne Falls dome. He suggested that the isoclinal folds with axial surface schistosity that are present in the Goshen strata immediately around the north lobe of the dome are of an earlier generation than the isoclinal folds with axial surface schistosity 2 km north of the dome and in the surrounding area. He correlated the earlier (around the dome) isoclines with a giant, regionally extensive recumbent isocline previously proposed by Osberg (1972, 1975). Osberg's and Simpson's evidence for this regional recumbent fold is the presence at a few localities of downward-facing subvertical isoclines (with axial surface schistosity) that they said required upside-down strata at the time of formation of the widespread subvertical isoclines that we call F_2. In a later section of this chapter, we present an alternative model to explain the downward-facing isoclines that does not require Osberg's large, or even smaller, pre-F_2 recumbent fold or even pre-F_2 upside-down beds. We further suggest that the schistosity described by Simpson (1974, p. 28) as "In a few of these folds, an earlier schistosity has been preserved in the more quartzitic beds, and this schistosity is wrapped around the noses of the folds," and a comparable folded schistosity described by Hall (written commun., 1977) resulted from F_2 being a continuum of folding, in which earlier folds were refolded by immediately following folds, and did not require a major regional recumbent fold as proposed by Osberg (1975). We are bothered by the aspect of Simpson's model that requires that in pre-F_2 time many small-wavelength large-amplitude isoclines with axial surface schistosity were formed in the Goshen strata within about a kilometer of the contact with the dome rocks but did not form in the same strata more than a few kilometers away from the dome. Furthermore, the hundreds of identical, equally small-wavelength large-amplitude isoclines that are ubiquitous throughout the rest of the Goshen terrane and that have the strong regional schistosity parallel to their axial surfaces apparently, by Osberg's and Simpson's model, fade out completely in the immediate vicinity of the dome and do not overprint Osberg's and Simpson's earlier isoclines and schistosity.

In conclusion, however, we wish to emphasize that the hard data that have been gathered to date are pitifully few and very ambiguous. None of us has fully resolved the question of the sequence of structures in the Goshen strata, particularly around the domes. We urge the interested student to read carefully the discussions by Osberg (1975), Simpson (1974), and ourselves (Stanley, 1975; Hatch, 1975; this paper), to list all possible models that might explain the present relations, including Osberg's, ours, and as many others as are reasonable, to carefully outline the various types of information that would bear on a selection of a "preferred" model, and to go out in the field without bias and gather those data.

SURFACE OF ACADIAN STRUCTURAL DISHARMONY

The nature of the boundary between the Goshen Formation and the underlying Ordovician and older rocks has been a subject of concern for many years to those of us working in western Massachusetts. Although the boundary has been shown on maps of the Heath (Hatch and Hartshorn, 1968), Plainfield (Osberg and others, 1971), Worthington (Hatch, 1969), and Chester (Hatch and others, 1970) quadrangles as a sedimentary contact, we long ago recognized that some displacement or differential slippage along the contact was required to explain the observed contrast in size of isoclinal folds in the Goshen and in the underlying pre-Silurian strata. On the maps of the Blandford (Hatch and Stanley, 1976), Woronoco (Stanley and others, 1982), and Goshen (Hatch and Warren, 1981) quadrangles, we showed the boundary as a "decollement." Both Hatch (1975, p. 54) and Osberg (1975, p. 67) suggested the possibility of detachment along this boundary in structural syntheses of western Massachusetts.

The somewhat unusual designation for this boundary on the State bedrock map, "surface of Acadian structural disharmony," was intended to describe the observed field relations on which we think all can agree, without implying anything as to the nature or direction of any movement on the surface about which there appears to be room for differences of interpretation. The high-amplitude, closely spaced, generally vertical, isoclinal folds in the Goshen Formation shown on the map and sections are well documented (Hatch, 1968, 1975). Furthermore, the absence of any belts of pre-Silurian rocks east of the Hawley (Cobble Mountain)- Goshen contact and of any belts of Goshen rocks west of this contact clearly shows that the Hawley-Goshen contact itself is not involved in those large isoclinal folds. Although the contact is known to be involved in only one mappable isocline (Leo M. Hall, written commun., 1977), the pre-Goshen formations, and thus possibly also the contact, contain abundant small (generally a few centimeters to 1 m in amplitude) isoclinal folds that we believe to be synchronous with the large Goshen isoclines (Hatch, 1975). The "Woronoco fold" described by Stanley (1975) and the large folds in an amphibolite in the western part of the Moretown Formation near West Cummington are the major exceptions to the generalization that the isoclines in the pre-Silurian rocks are small.

Although the field relations seem clear and thus could be classed as "data," the mechanism or process by which the rocks achieved these relations is ambiguous and thus is a matter of interpretation. First, we interpret the Goshen isoclines to have originally formed as a series of recumbent folds because they appear to have been the first set of folds imposed on the Goshen strata and because wherever the relations can be observed their axial surfaces are parallel to the basal contact of the Goshen. Second, as discussed below in the section on the domes, our interpretation is that these recumbent isoclines originally formed as west-facing structures as a result of nappe formation and other westward-shoving movements in the Bronson Hill anticlinorium. Third, we interpret from field observation of a few exposed isoclines and from their map pattern that little if any shearing or faulting occurred along the axial surfaces of individual isoclines at the time of their formation. This model is simplistically represented by figure 10.

If the geometry of figure 10 is achieved by an east-over-west couple, a net westward displacement of the Goshen strata relative to the underlying strata is produced. Furthermore, the amount of that relative displacement decreases from east to west and presumably reaches zero at some point where the isoclines die out and the Lower Devonian "rug" is effectively glued to its pre-Silurian "floor." We propose that that point was originally west of the presently exposed western Goshen contact and is thus now somewhere "up in the air," well above the present ground surface.

Although we have referred to this interface as a "surface of structural disharmony," we interpret it to be a decollement that involved considerable westward displacement. It is mechanically feasible to have a zone of weakness between older recrystallized rocks and thinly bedded water-saturated rocks of the overlying Devonian cover. Compression of a presumably westward-thinning wedge of material early in the Acadian orogeny could result in a major detachment at the base of the wedge and severe isoclinal folding of the overlying material.

DOMES OF WESTERN MASSACHUSETTS

The geology of the Connecticut Valley synclinorium in Massachusetts is complicated by four elliptical domes, the Shelburne Falls, Goshen, Woronoco, and Granville domes, that form a roughly north-south chain across the State (fig. 11). All have pre-Silurian rocks of the Collinsville and (or) Cobble Mountain Formations exposed in their cores. A fifth domal structure is well defined by the dominant F_2 schistosity northeast of the Shelburne Falls dome (fig. 2) but does not expose pre-Silurian rocks at the surface. Gravity data by Pferd (1978) and Simpson (Simpson, 1974, fig. 3) suggest that they are present at depth and participate in the domal structure. Detailed mapping by Pferd (1978) defined a series of southward-verging recumbent folds that draped over and were arched by the dome (cross section *A-A'*, State bedrock map).

The northern two domes, the Shelburne Falls and Goshen domes, have pre-Silurian cores entirely surrounded by Lower Devonian rocks. The Woronoco dome, which contains four small patches of Ordovician Cobble Mountain Formation (Ocb, State bedrock map) within its general domal configuration, is at the southern end of the continuous synclinorium. The Granville dome, which straddles the Connecticut State

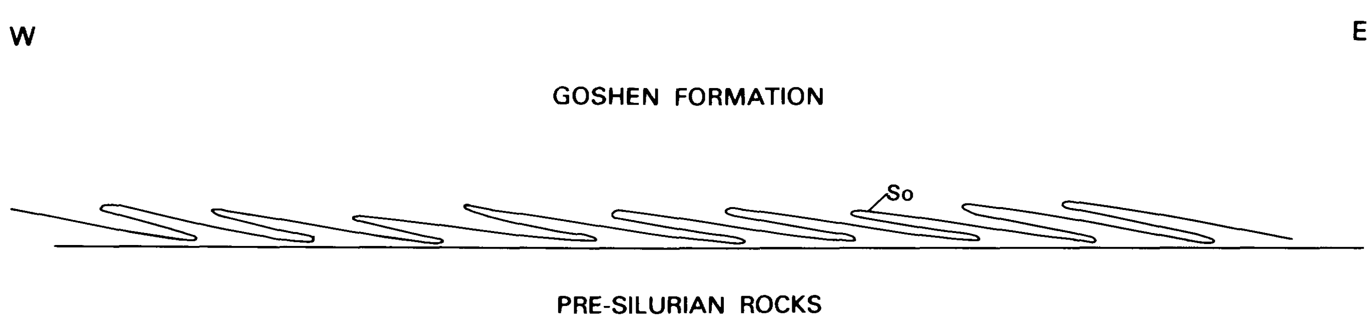
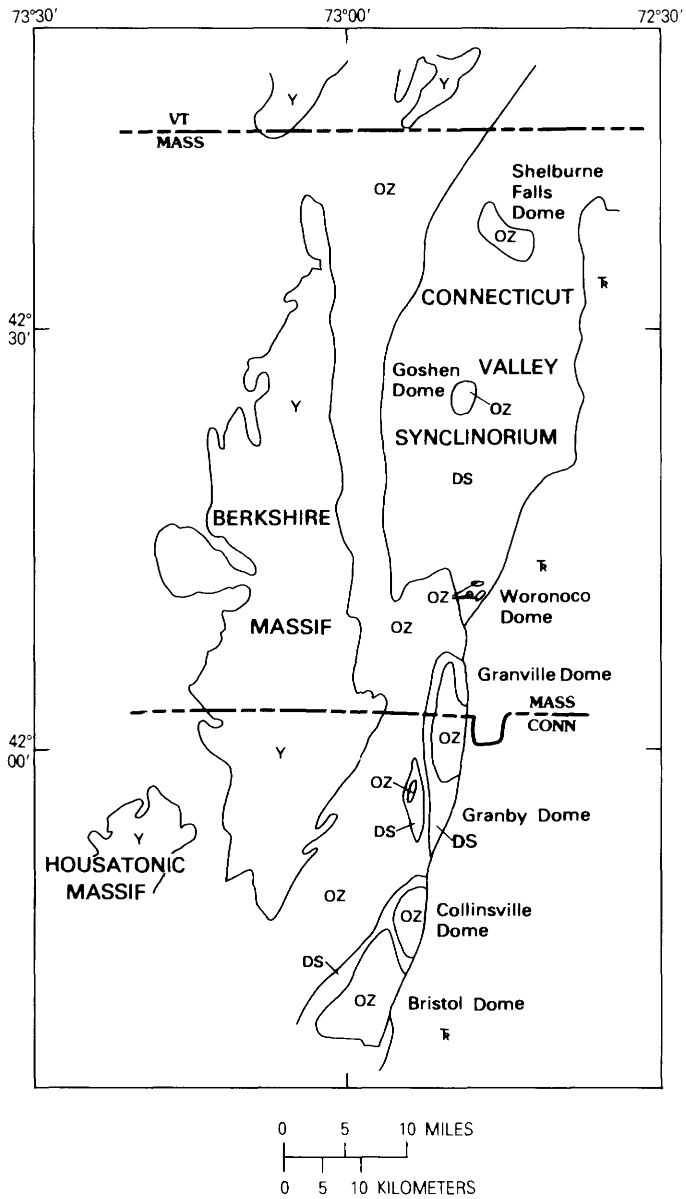


FIGURE 10.—Sketch of isoclinal folds in the Goshen Formation and their relation to underlying pre-Silurian rocks immediately after F_2 time. S_0 is bedding.



EXPLANATION

- Ʀ Triassic rocks
- DS Devonian and Silurian rocks
- OZ Ordovician to Proterozoic Z rocks
- Y Proterozoic Y rocks

FIGURE 11.—Sketch map of southwestern New England showing the domes of western Massachusetts and Connecticut in relation to principal geologic features.

line, is strongly elongate in a northerly direction and is surrounded by a narrow faulted syncline of Goshen Formation, as are the domes to the south in western Connecticut (fig. 11; Hatch and Stanley, 1973, pl. 1). Both the Woronoco and Granville domes are cut off on the east by the Mesozoic border fault.

As shown by the State bedrock map and figure 2, all four of the structures are indeed domal in that beds and the dominant schistosity (S_2) dip gently to moderately outward on all sides around their cores. Abundant reversals of facing direction of graded beds show that the F_2 isoclinal folds described above are present in the Goshen strata immediately over the core rocks and that their axial surfaces dip quaquaversally around the domes generally parallel to bedding.

Recent studies of these domes, subsequent to the early work of Emerson (1898a,b, 1917) who recognized all but the Woronoco dome, include the work of Balk (1946), Hall (written commun., 1977), and Simpson (1974) on the Shelburne Falls dome, the work of Hatch (Hatch and Warren, 1981) on the Goshen dome, the work of Clark (1977; Stanley and others, 1982) on the Woronoco dome, and the studies of Schnabel (1974) and Knapp (1977, 1978) on the Granville dome. Our intent here is to summarize our current thoughts on the evolution of these domes and the relationship of that evolution to the tectonic history of this part of Massachusetts. Key questions to be addressed are (1) when were the domes formed? and (2) what was the mechanism—vertical buoyant gravitational movement, interference of superposed folds (Ramsay type 1 pattern), or a combination thereof—by which the domes were emplaced?

The first question, when were the domes emplaced, is best considered relative to the chronology of the folding events summarized above. Both of us (Hatch, 1975, p. 56; Stanley, 1975, p. 84 and 93) have previously suggested that dome formation took place approximately during F_3 time, largely on the basis of the fact that S_3 is well developed within and parallel to the axial surfaces of the major synclines between and west of the domes in southern Massachusetts and western Connecticut. On the basis of more recent studies around the Goshen and Shelburne Falls domes to the north, and our present understanding of F_3 and F_4, we herein revise that interpretation.

Figure 3 shows, on a regional scale, the attitude of F_3 fold axial surfaces and axial-surface cleavage. Although some deflection of these structures around the domes is apparent, it is certainly not what would be expected if F_3 clearly predated all the rise of the domes. F_3 cleavage is well developed across the northwest lobe of the Shelburne Falls dome and definitely cuts bedding, S_2 schistosity, and stratigraphic contacts at a high angle on the southwest and northeast sides of that lobe. West of the Goshen dome, however, we do not recognize any spaced or slip cleavage clearly cutting an earlier schistosity and parallel bedding. Instead, the only S surface other than bedding that is obvious in this area is a schistosity that generally cuts bedding at a moderate angle; in a few exposures, relicts of an earlier orientation of mica flakes parallel to bedding in more granulose beds indicate that this schistosity is probably exceptionally well developed S_3 rather than S_2 (Hatch and Warren, 1981). It therefore appears that S_3 is deflected from its regional northeasterly trend into approximate parallelism with the west side of the Goshen dome and the rim syncline described below. This may suggest that the rim synclines are at least locally F_3 in age or that their time of formation at least partly overlapped F_3 time. Alternatively, S_3 and the axial surface of the older rim syncline may have been flattened into near parallelism by subsequent F_4 deformation. As noted on figure 4, the axial surface of the tightened rim syncline and the well developed S_3 schistosity are essentially parallel to S_4 elsewhere. Clark (1977) observed that axes of F_3 folds form an incomplete small circle about a vertical axis that he inferred to be the direction of upward movement of the core gneisses in the Woronoco dome. From this he concluded that at least some part of the rise of the dome postdated F_3. We note, however, that this same geometry could result from superposition on the dome of northeast-trending F_3 axial surfaces and thus does not unambiguously define the relative ages of F_3 and doming. To the south, F_3 folds are well developed west and north of the Granville dome. F_3 folds on the limbs of the dome have been deformed by later F_4 folds so that their axial surfaces (S_3) form an arch across the dome (Knapp, 1977, p. 85-87). The north-plunging F_3 hinges on both dome limbs show counterclockwise or east-over-west rotation sense and are not parasitic, therefore, to the dome itself. As shown on figure 4, F_4 structures are well developed north of the Granville dome. Subsequent deformation and flattening of the Granville dome during F_4 time, therefore, rotated S_3 about the axis of the dome.

The above relations indicate that all four domes underwent at least two stages of upward movement. The first, the gravitational or rim syncline ($F_{2.5}$) stage, occurred before F_3; it is most clearly shown in the Shelburne Falls dome and, to a lesser degree, in the Goshen dome. The second phase occurred as a result of superposition of F_4 onto F_3. This event formed the double-lobed configuration of the Shelburne Falls dome and the pronounced north-south elongation of the Granville dome. These gravitational and F_4 stages were separated in time by the regional development of F_3, an event that may have been time transgressive from south to north. The time separating the gravitational and F_4 events may have been short—in fact, they may have been stages in a continuous deformational sequence.

The second question about the domes is *how* did they form? As noted above, Hall (written commun., 1977) suggested a mechanism of interference of superposed folds (Ramsay type 1 pattern) to explain the Shelburne Falls dome, and, by implication, the other domes to the south. We recognize problems with the interference model and herein outline a model that combines fold interference with gravitational rise to explain both the location and form of the domes and their relationship to, and the present structure of, the F_2 isoclinal folds.

Although Hall (written commun., 1977) has argued convincingly for a fold interference (presumably of what we call F_3 and F_4) model of the Shelburne Falls dome, and although the elongation of the Granville dome parallel to F_4 indicates some degree of fold interference origin for that dome, two features of the rocks argue against this as the *only* mechanism of dome formation. First, if the western Massachusetts domes were solely the result of superposition of F_3 and F_4 anticlines, a systematic pattern of domes and basins should be present throughout the area. Not only have we failed to recognize any such pattern, but we have not recognized a single basinal structure anywhere in the area. Even though the most obvious feature of the domes is the map pattern of older units mantled by younger strata, their geometry is even more strongly defined by the attitude of beds and schistosity in both the core and mantling strata. Thus it seems reasonable that if basins associated with the domes were present, they would be unmistakably outlined by the attitude of beds and schistosity even if no separate map unit identified their centers.

The second argument against fold interference as the sole mechanism of dome formation is the distribution of recognized F_3 and F_4 folds relative to the domes. Although F_3 cleavage and minor (outcrop scale) folds are abundant throughout the Rowe-Hawley zone and Connecticut Valley belt (see fig. 3), major folds of this generation are recognized only in the vicinity of Blandford (State bedrock map; fig. 3). The large left-handed fold in that area significantly deforms all formational and member contacts from the base of the Hoosac up through the Goshen and conspicuously refolds the F_2 isoclines in the Goshen north of Blandford. The trace of the axial surface of this major F_3 fold can easily be mapped from the vicinity of Blandford northeast almost to Norwich (State bedrock map). Furthermore, the largest recognized F_4 fold in the area is the fold about a north-south axial surface in the vicinity of Russell along the north extension of the Granville dome (fig. 4). It would seem logical that if domes in western Massachusetts were formed by interference of F_3 and F_4 anticlines, the prime area for a dome would be at the intersection of the trace of the F_4 anticline through Russell and the F_3 anticline complementary to, and southeast of, the F_3 syncline through Blandford. This anticline intersection would be a few kilometers south of the village of Russell at a point that is about 9 km north-northwest of the Woronoco dome and about 18 km south of the Goshen dome. No suggestion of a dome was found in this area. We again conclude that although both F_3 and F_4 folds may have affected, to varying extent, the location, size, or shape of some or all of the domes, Ramsay type 1 interference phenomena were not the prime mechanism of either dome location or dome formation.

Having thus argued that fold interference is not the principal or only mechanism, we now turn to the chief alternative—gravitational or buoyant rise of lighter rock, presumably the gneisses that are exposed in the cores of three of the domes and probably underlie rocks of the Cobble Mountain Formation in the Woronoco dome (Griscom and Bromery, 1968, p. 423). Simpson (1974) determined a density contrast of 0.15±0.05 g/cm^3 between the core rocks (lighter) and the mantling Goshen strata (heavier) in the Shelburne Falls dome, and the similarity of both the core gneisses and the mantling Goshen strata to corresponding rocks in the southern domes suggests that a similar density contrast exists in them as well. We propose that this density difference was at least in part responsible for the rise of the domes. Following Simpson's conclusion that the gneiss forms a thin sheet (0.6-2.5 km according to Simpson, 1974, fig. 5), we explain this shape as resulting from a thrust slice floored by the Bristol thrust (Stanley and Hatch, Ch. A, this volume; cross sections on State bedrock map) rather than a rooted recumbent fold as suggested by Simpson (1974) and Osberg (1975).

Both the aeromagnetic data (Griscom and Bromery, 1968) and the gravity data (Simpson, 1974) suggest that the gneisses coring the domes extend well beyond the limits of the present surface exposures. No data are available, however, to indicate whether the gneisses presently exposed and inferred in the four domes are part of a single sheet that is essentially continuous from the Vermont-Massachusetts State line south to the western Connecticut domes and east to an inferred root zone in the Bronson Hill anticlinorium or are parts of four smaller tabular "mini-slices." On the cross sections for the State bedrock map we have opted for the former model because of the similarity of the gneisses in the domes to the Monson and Fourmile Gneisses in the Bronson Hill anticlinorium. We conclude, therefore, that the density difference between this thrust-floored sheet of gneiss and its heavier blanket of schist of the Goshen was sufficient to cause it to blister or bulge up in structural flexures or points of weakness in that mantling Goshen blanket (which was already deformed into recumbent F_2 isoclinal folds). These blisters or bulges, still driven by that density difference, were subsequently accentuated and molded into their present shape by the northwest-southeast and east-west compressive stresses of F_3 and F_4.

As noted above, our field observations clearly show that the axial surfaces of the F_2 isoclines in the Goshen Formation closely parallel the configuration of the pre-Silurian surface upon which the Goshen now rests in such critical areas as around the Shelburne Falls and Goshen domes and along the western and southern margin of the synclinorium. This relationship is shown in figure 12 by a cross section of the synclinorium at the latitude of the Goshen dome simplified to show only the

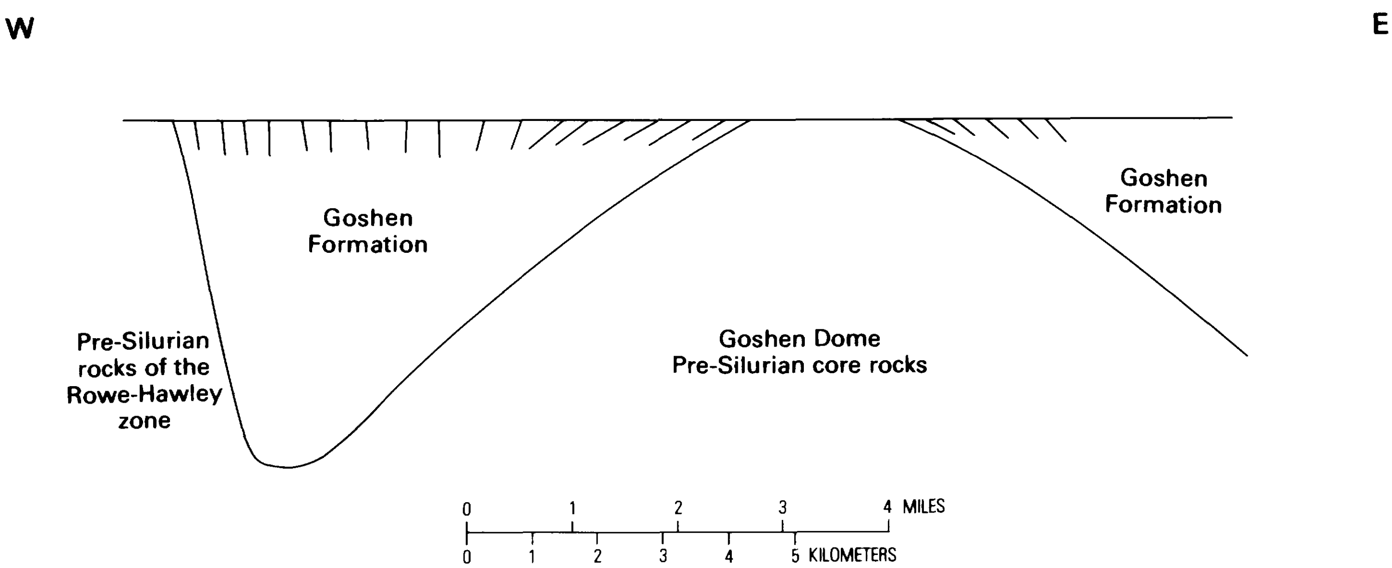


FIGURE 12.—Schematic east-west cross section across the Goshen dome showing the contact between the Goshen Formation and the pre-Silurian rocks and the attitudes of the axial surfaces of F_2 isoclinal folds in the Goshen.

surface of the pre-Silurian rocks and the attitudes of the axial surfaces of the F_2 isoclines. The figure points up two important structural features. First, the axial surfaces pass through the vertical rather than the horizontal in changing from a steeply east-dipping attitude on the west edge of the diagram to a gently west-dipping attitude on the west flank of the dome. A similar fan-like configuration is present west of the Shelburne Falls dome (Hatch and Hartshorn, 1968, cross section *B-B'*) and west of the Woronoco dome (Stanley and others, 1982). Second, about 1.5 km west of the Goshen dome, the dip of the F_2 axial surfaces changes abruptly, rather than progressively (Hatch and Warren, 1981), from about 35°W. to about 60°W. This change in orientation could easily be overlooked in the field but is, we feel, critical to understanding the structure.

After considering a number of models to explain the present shape of the pre-Silurian surface and the present configuration of the axial surfaces of F_2 folds, we concluded that the following model is most compatible with field observations. First, because the isoclines are clearly deformed by the domes (figs. 2, 12), they must predate the dome formation and thus probably originally formed as isoclines recumbent upon a then roughly horizontal pre-Silurian surface. This is clearly shown to the south in the vicinity of Blandford and Woronoco where the basal Devonian surface is folded by post-F_2 folds and deflected eastward. We further infer that, because the principal Acadian transport direction, both east and west of the synclinorium, was westward, the recumbent folds verged and faced westward and initially formed a pattern similar to that shown in figure 13*A*. This model is further supported by the fact that stratigraphically higher units in the Goshen Formation are found in synclines located progressively eastward of the Hawley-Goshen contact, as illustrated by Jackson (1975). Although it is perhaps theoretically possible that the isoclines formed originally as upright rather than recumbent folds, such a model would not account for their present gentle dips around the domes, their apparent projection into horizontality over the tops of the domes above the present ground surface, or their conformable geometry in the Blandford-Woronoco area.

Our model thus provides that following the F_2 episode of isoclinal folding, F_2 axial surfaces were stacked up in a gently dipping homoclinal array across the area of the present synclinorium and the future domes much as undeformed bedding surfaces were before F_2 (fig. 13*B*). Gentle folding of these stacked axial surfaces in slightly pre-F_3 time into an open synform, either by gravitational rise of the dome or by broad-scale east-west compression as discussed in the preceding paragraphs, would produce a configuration such as is shown in figure 13*C*. Note that, in the central part of this late synform between the incipient dome and the west margin of the basin (incipient arching of the pre-Goshen rocks to the west), the F_2 axial surfaces pass through the horizontal forming a synform with an apparent hinge area well to the east of the hinge in the pre-Silurian surface. This synform is analogous to the rim synforms around salt domes. Clearly, the present observed configuration (fig. 12) of F_2 axial surfaces requires much tighter folding than shown in figure 13*C* of what we refer to as the rim syncline. We attribute this tighter folding to F_4 (fig. 13*D*).

We have described a "rim syncline" formed by gravitational doming west of the Goshen dome (fig. 13*D*). According to our model, analogous folds must exist near all the other domes and the Whately anticline. In figure 14 we have shown possible locations of other rim synclines and companion anticlines associated with them inferred from this model. The synclines west of the domes and the Whately anticline are required by the model we outline here; the location and the very existence of the extensions of these synclines and of the anticlines are highly speculative. We assume that the axial surfaces ($S_{2.5}$) are vertical or steeply dipping. We emphasize, therefore, that figure 14 is intended chiefly to show the *kind* of fold pattern that results from our model and *not* the details of where those folds must be located. Because we interpret them to have been initiated by gravity in pre-F_3 time, we refer to those folds as $F_{2.5}$.

Figure 13*D* shows clearly that F_2 anticlines would be downward facing (upside down) on the eastern limb and upward facing (right-side up) on the western limb of the $F_{2.5}$ fold. Figure 14 shows areas in which, by our model, F_2 isoclines would be expected to face downward and upward. It also shows the known location of downward- and upward-facing isoclines reported by Osberg (1975, p. 63-67) and observed by us. All of these localities are in the areas predicted by our proposed model. We recognize, however, that figure 14 could equally well have been drawn in such a way that some of the observed folds would be incompatible with it and thus that the apparent compatibility in no way proves the model.

Although downward-facing isoclines should, from figure 14, be nearly as abundant as normal ones, very few have been reported. We point out, however, that very few upward-facing isoclinal hinges have been described either. One of the major difficulties in studying F_2 folds results from their geometry and the paucity of properly oriented surfaces of observation. Because F_2 folds are truly isoclinal, it is necessary to observe their hinges in order to determine whether

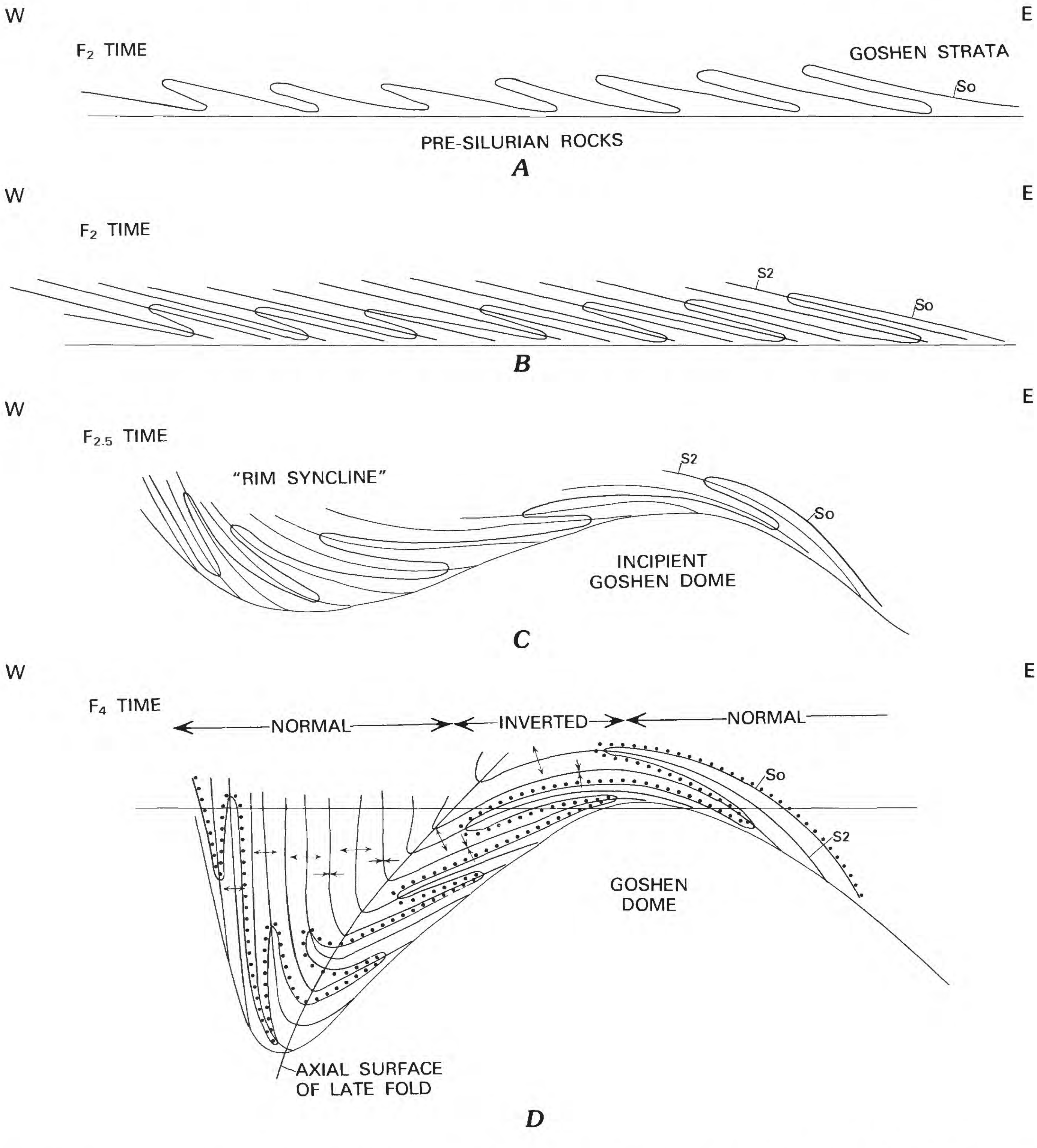


FIGURE 13.—Sequence of diagrams depicting the evolution of the present geometry of F_2 isoclinal folds in the Goshen Formation around the Goshen dome. S_0 is bedding. *A*. F_2 time. West-verging isoclines recumbent upon the then subhorizontal surface of pre-Silurian rocks. *B*, Same diagram with the addition of F_2 axial surfaces (S_2). *C*, $F_{2.5}$ time—after F_2 but before F_3 time. Initial folding of S_2 by gravitational rise of the core gneiss. *D*, Final folding of the basal Goshen surface and the F_2 axial surfaces due to east-west compression during F_3 and F_4 time. Dots represent sedimentary tops direction of a representative bed, S_0. F_2 isoclines are downward-facing between the crest of the dome and the trace of the axial surface of the late fold. F_2 isoclines are upward-facing west of the axial trace of the late fold and east of the dome.

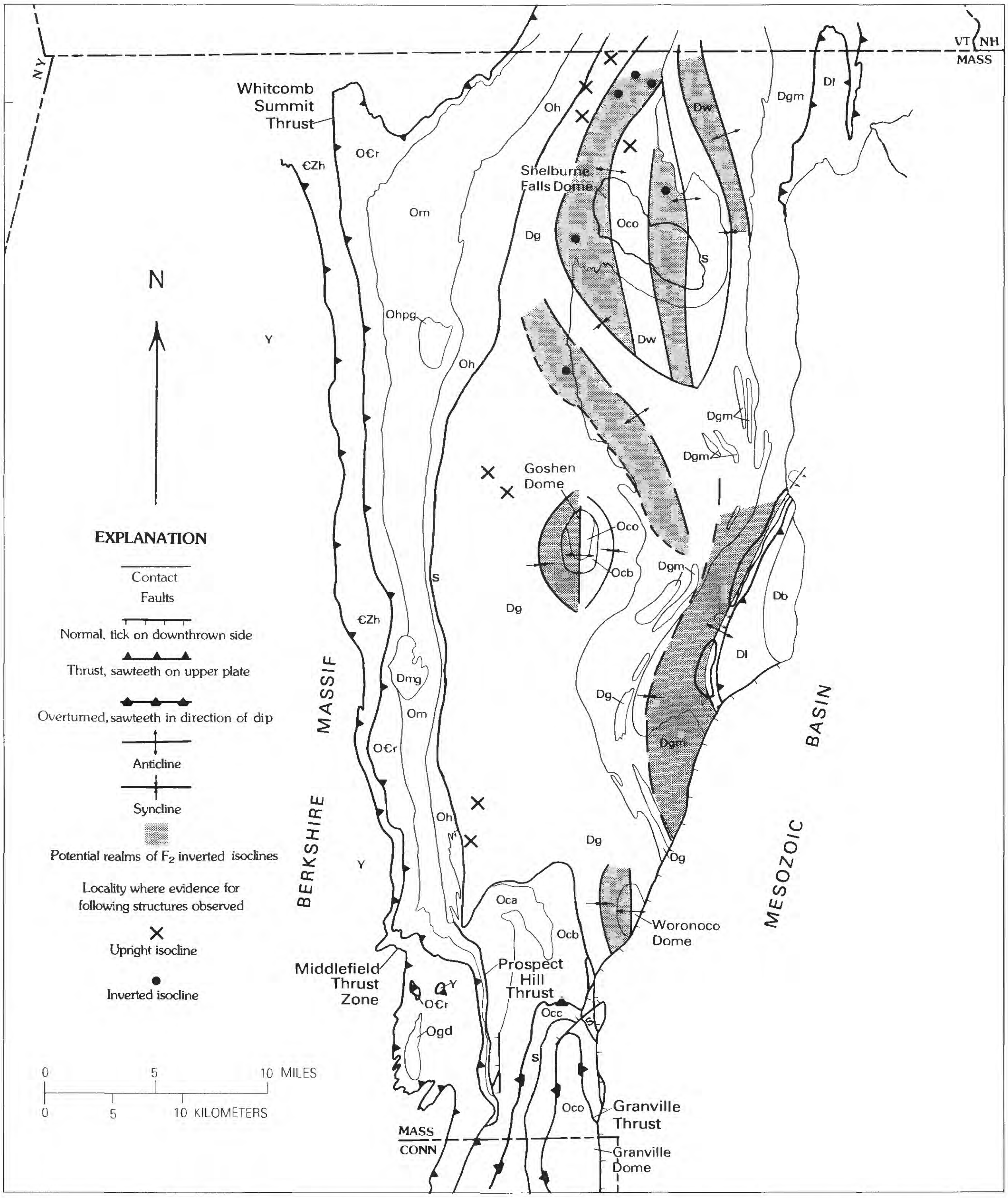


FIGURE 14.—Possible locations of inferred and speculative $F_{2.5}$ rim synclines and companion anticlines in the Connecticut Valley zone. Also shown (by pattern) are areas in which the model predicts inverted F_2 isoclines. The locations of the documented upward-facing and downward-facing F_2 isoclines are indicated by X's and solid circles, respectively. Explanation of letter symbols: Dmg, Middlefield Granite; Db, Belchertown Complex; Dl, Littleton Formation; Dgm, Gile Mountain Formation; Dw, Waits River Formation; Dg, Goshen Formation; Ohpg, gneiss at Hallockville Pond; Ogd, diorite at Goff Ledges; Oca, Ocb, Occ, members of the Cobble Mountain Formation; Oco, Collinsville Formation; Oh, Hawley Formation; Om, Moretown Formation; O€r, Rowe Schist; €Zh, Hoosac Formation; Y, Proterozoic Y rocks; S, surface of structural disharmony at the base of the Goshen Formation.

they are *structural* synforms or antiforms. Their *stratigraphic* form (younger or older beds in the core of the fold) can readily be determined from abundant graded beds. Critical data for determining whether an isocline is structurally upward or downward facing are the combination of structural form with the stratigraphic topping information, as demonstrated by Osberg (1975, p. 63-67). F_2 hinges are horizontal or plunge very gently. Consequently, they are generally well exposed only on steep surfaces that are at a high angle to the bedding and the parallel S_2 schistosity, which fabrics control most of the natural exposures. Prominent west-east joints and artificial exposures along east-west roads thus provide the best opportunities to observe F_2 hinges. Few such exposures exist, however, with the result that our data are insufficient to map out areas of upward-facing and downward-facing F_2 folds with sufficient accuracy to critically test the model we propose here.

Osberg (1975, p. 63-67) first pointed out the existence and possible significance of some downward-facing F_2 folds in the general environs of the Shelburne Falls dome. He interpreted them to indicate the presence of a major recumbent syncline, opening to the east, that predated the F_2 isoclinal folding (fig. 15). By his model, downward-facing F_2 isoclines are present in areas where the upper, inverted limb of that pre-F_2 recumbent syncline intersects the present ground surface; upward-facing F_2 isoclines would be expected in areas where the lower, upright limb is exposed. In fact, mapping of the inverted limb and thus documenting the early recumbent fold depends upon locating downward-facing isoclines—something that both Osberg and we have found difficult for the reasons discussed in the previous paragraph.

Osberg's (1975) model was designed to explain the observed downward-facing F_2 isoclines by a major pre-F_2 recumbent fold upon which steeply inclined F_2 isoclines were superposed. In contrast, ours was originally developed to explain the present attitudes of F_2 axial surfaces but produced downward-facing F_2 isoclines as an inevitable consequence of refolding the F_2 axial surfaces. Although a choice between the two models would best be based on the map distribution of demonstrably right-side-up and upside-down F_2 hinges, presently available data are far too few to enable such a choice, and the outlook for finding sufficient data in the future is grim. We therefore offer the following discussion as relevant to the problem, although not a resolution of it.

First, Osberg's model calls upon an eastward-opening large recumbent syncline that implies a west-over-east (Acadian) movement sense opposite to the Acadian movement sense indicated by most of the available data in western Massachusetts for this period of time. Our model calls upon an east-over-west movement sense to produce the initially recumbent F_2 isoclines in an area in which east-over-west movement is predominant and widely recognized.

Second, according to Osberg's model, as elaborated upon by Simpson (1974), the isoclinal folds with axial surface schistosity that immediately overlie the moderately dipping core gneisses of the Shelburne Falls dome are earlier (pre-F_2) than the isoclinal folds with similar axial surface schistosity in the same Goshen strata a kilometer or so away (out from the flank of the

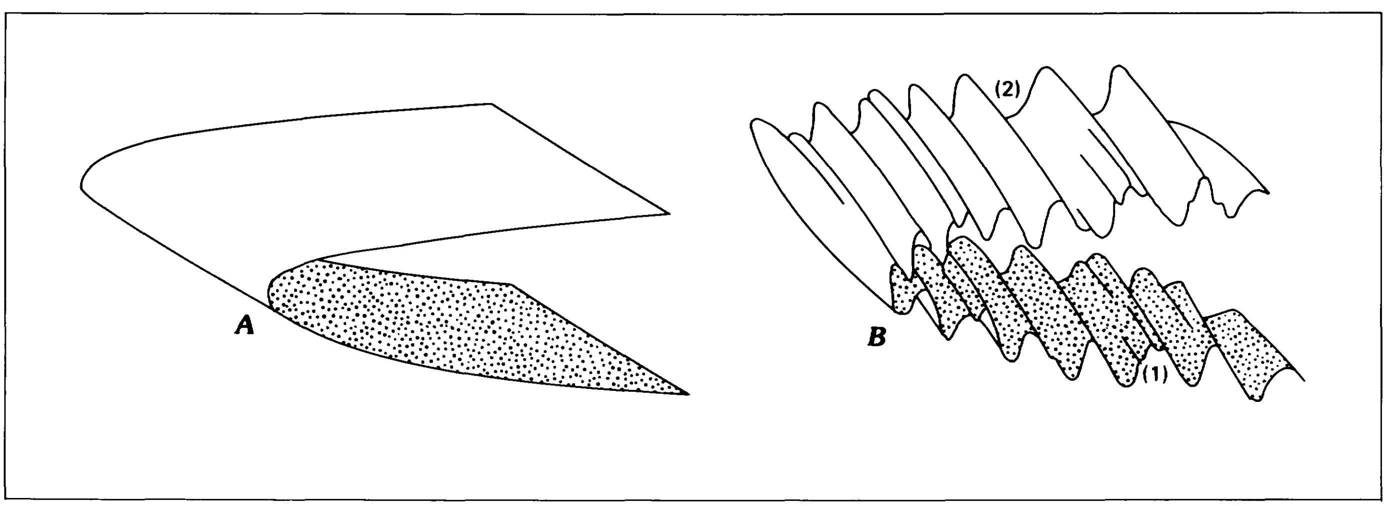


FIGURE 15.—Sketch showing Osberg's (1975) model of a major pre-F_2 recumbent syncline (*A*) subsequently refolded in F_2 time into vertical isoclines (*B*). F_2 isoclines would be downward-facing on the upper limb (2) of the recumbent fold and upward-facing on the lower limb (1). Stippling indicates the bottom of the folded surface. From Osberg (1975, fig. 59).

dome). According to our model, the two sets of isoclines are the same generation (F_2), and any differences in attitude or form result from their position relative to the later ($F_{2.5\text{-}3}$) rim syncline.

Third, by Osberg's model, we could expect to see some trace of schistosity formed parallel to the axial surfaces of the pre-F_2 folds in the hinge areas of the F_2 folds. This pre-F_2 axial-surface schistosity should wrap around the hinges of F_2 folds, whereas F_2 schistosity should be axial planar to the F_2 hinges. Although we have carefully examined all available F_2 hinges to determine the form of any cleavage or schistosity, we have not recognized any folded cleavages in any of them. Simpson, however, does report (1974, p. 28), "In a few of these [recumbent isoclines on the flank of the Shelburne Falls dome] folds, an earlier schistosity has been preserved in the more quartzitic beds, and this schistosity is wrapped around the noses of the folds." Although such a relationship could clearly be interpreted to indicate an earlier (pre-F_2) stage of (probably also isoclinal) folding, we feel that positive interpretation of cleavages in the very restricted areas of the hinges of these isoclines is sufficiently difficult that more supporting data are needed to prove a pre-F_2 folding event on this kind of evidence. Furthermore, as noted below in the discussion of east-verging folds, a progressive development or continuum of west-verging isoclines, with some rolling over of earlier formed folds by later folds, could explain Simpson's observations without a large regional recumbent structure verging in either direction.

Fourth, Hepburn (1975) described an episode of major recumbent folding from the Guilford dome area, north of the Shelburne Falls dome, in southeasternmost Vermont. He described this episode as being later than his F_1 stage of folding that produced "a well-developed schistosity (S_1) [that] apparently parallels bedding throughout the [Guilford dome] area" (Hepburn, 1975, p. 39). We correlate Hepburn's F_1 with F_2 in Massachusetts. It is thus difficult to correlate Osberg's proposed recumbent fold that predates F_2 schistosity with the major recumbent folds described by Hepburn in the same sequence of rocks in the adjacent area to the north that postdate the probable equivalent of our S_2 schistosity. The details of Hepburn's structural history of the Guilford area are similar to the sequence of events described by Stanley (1975) to the south in southern Massachusetts and northern Connecticut and expanded upon herein.

Fifth, our model offers an explanation for the abrupt increase in the value of dips of beds (and of S_2) westward from the Goshen dome from about 30° on the flank of the dome to about 60° immediately to the west. Osberg's model should produce a progressive increase in dip values with no such abrupt break.

On the basis of the available data, we suggest that our model of later rotation of west-facing recumbent F_2 isoclines into an inverted position explains their extent and predicts their known locations and is at least a viable alternative to Osberg's earlier model of a large east-verging pre-F_2 recumbent fold. We agree that small pre-F_2 folds are locally present (as, for example, on the north side of the Shelburne Falls dome) but believe that they are related to early F_2 east-over-west movement of the Devonian cover rather than to a discrete pre-F_2 episode of major recumbent folding. Locally these folds are refolded by F_2. Therefore, inverted isoclines could be formed by two processes: early east-over-west movement and refolding of F_2 by the rim synclines.

We wish to re-emphasize here the paucity and ambiguity of the available data. No model for explaining the downward-facing isoclines can currently be considered anywhere near proven, and certainly none can yet be considered disproven. Our only purpose in the preceding discussion is to present a viable alternative to Osberg's earlier model and to indicate some of the pertinent differences between them. By doing so we hope that we have pointed out the kinds of field observations that are needed to impose real constraints on the choice of model to explain the observed field relations.

Let us now return to the fundamental questions at the beginning of this section, namely how and when did the domes form? First let us list the constraints.

(1) The domes are circular to elliptical in outline, and the dominant mantling schistosity dips outward.
(2) The Collinsville and (or) the Cobble Mountain Formations form the core of the domes. These rocks are lithically and probably stratigraphically equivalent to rocks of the Bronson Hill anticlinorium (as discussed in Stanley and Hatch, Ch. A, this volume, and as previously noted by Hall and Robinson, 1982).
(3) The domes are linearly arranged along the axis of the Connecticut Valley synclinorium from southwestern Connecticut to southern Vermont (fig. 11). From the Waterbury dome in Connecticut to the Granville dome in southern Massachusetts, the domes are arranged in a right-handed en echelon pattern.
(4) Throughout the belt, the domes deform F_2 folds and coeval or older structures, such as the basal Devonian surface and the Granville thrust.
(5) F_3 axial surfaces cut across the Shelburne Falls and Colrain domes, are ambiguous across the Goshen dome, and describe a large-scale arch around the Bristol, Collinsville, Granby, and Granville domes.

(6) F_4 folds geometrically coincide with the Granville dome and the two lobes of the Shelburne Falls dome. This relationship cannot be demonstrated for the Goshen and Woronoco domes.

(7) The domes began their development after F_2 and before F_3 in what we herein refer to as $F_{2.5}$ time. Other $F_{2.5}$ folds probably formed during this time. The axial surface of the Whately anticline is parallel to S_4 to the west, so it may have developed during F_4.

Our previous discussion and the foregoing list clearly indicate that the domes were developed by a combination of vertical upward movement during $F_{2.5}$ time and subsequent F_3 and F_4 folds. We suggest that the $F_{2.5}$ movement was dominantly buoyancy driven by the lighter rocks of the Collinsville Formation of the dome cores; this explanation has been put forth by many workers on the domes of New England (Thompson, 1950; Skehan, 1961; Stanley, 1964, 1968; Hatch, 1975; Hatch and Hartshorn, 1968). We favor this mechanism over horizontal compression because of the circular to elliptical pattern of the domes and the presence of lighter felsic rocks in the cores. We further suggest that the homogeneous, garnetiferous, biotite-plagioclase gneiss, of possible plutonic origin (density about 2.67 g/cm^3, Simpson, 1974, p. 20), is the principal unit that imparts buoyancy to the core rocks. This unit is lithically like the Monson Gneiss of the Bronson Hill anticlinorium (Emerson, 1917; Hall and Robinson, 1982). It is found in the Waterbury, Bristol, and Collinsville, Conn., domes (fig. 11) and the Shelburne Falls dome; it is not exposed in the others because of present erosion level and (or) extensive surficial cover. This model is further supported by the "pear-shaped" map pattern of the Bristol dome in Connecticut. Here the base of the "pear" is underlain by the homogeneous plagioclase gneiss of the Bristol Member of the Collinsville Formation, whereas the top of the "pear" is underlain by binary mica gneiss, schist, amphibolite, and quartz-feldspar granofels of the Taine Mountain and overlying Collinsville Formations (Stanley, 1964).

Superposition of F_3 and F_4 folds on the $F_{2.5}$ structures has modified the map pattern of the domes to a greater or lesser extent. Generally, the domes are strongly elliptical where F_4 is well developed, as for example the Granville and Granby domes (Stanley, 1975, fig. 79, for example). Interestingly, it is in this area that the present east-west distance is shortest between the core rocks of the domes (western edge of the Bronson Hill plate of Robinson and Hall, 1980) and the Precambrian of the Berkshire massif. This region must have acted as a pressure point during east-west collision resulting in the elliptical map pattern of the domes. Quite simply, we suggest that the domes may represent strain ellipses that were modified from a more circular pattern largely during F_4 compression. The eccentricity of the ellipses decreases to the south in Connecticut (Waterbury dome) and to the north in Massachusetts (Goshen). The Shelburne Falls appears to be somewhat anomalous in that it is strongly elongate in a northwesterly direction. Hall (1977), however, has shown that this shape is due to the interference of F_4 on older folds to produce two north-trending lobes separated by an intervening saddle (fig. 14). Although we realize fully that present dome shape is a result of superposed strain as well as of present erosion levels, elliptical shapes with high eccentricity may well delineate "pressure" points during Acadian compression that were inherited from the original shape of the eastern edge of the Grenville plate and were subsequently modified during Taconian collision.

GRANVILLE THRUST

The Granville thrust was first proposed by Knapp (1977, p. 90-97) to explain the asymmetry of lithic members of the Goshen Formation in the syncline around the Granville dome in southern Massachusetts (fig. 16). Earlier, Stanley (1967, 1968, 1975) and Hatch and Stanley (1973) had not only proposed the correlation of the rocks of the Collinsville Formation within the domes with the rocks of the Cobble Mountain Formation to the west but, more importantly, had also demonstrated the lithic and inferred time correlation between the Goshen Formation of Massachusetts and The Straits Schist that mantles the domes in western Connecticut. They suggested further that the Goshen-Straits strata occupied a highly deformed east-facing isoclinal syncline that mantled all the domes south of the Woronoco fold. Detailed mapping by both Schnabel (1974) and Knapp (1977, 1978) in the area of the Granville dome showed conclusively that the Goshen (Straits) there was divisible into two members: (1) an outer (relative to the dome) well-bedded schist and quartzite unit typical of the Goshen to the north and (2) an inner (relative to the dome) unit of poorly bedded carbonaceous schist and quartz schist. Calc-silicate gneiss, although present in both units, is more concentrated in the inner, poorly bedded unit. Although Schnabel (1974) subdivided The Straits on the basis of the calc-silicate gneiss and considered the inner unit next to the dome to be stratigraphically the lower of the two, whereas Knapp (1977, 1978) subdivided on the basis of the bedding fabric and considered the outer, better bedded, unit to be stratigraphically lower, both agreed that the Goshen around the Granville dome consists of two and only two lithic units and thus does not have the stratigraphic symmetry demanded by a simple isoclinal syncline. This same asymmetry had been described earlier by Stanley (1964, p. 18-30) for the Collinsville and Bristol domes in Connecticut, although separate units were not mapped. Knapp (1977, fig. 7) proposed the Granville thrust along the inner contact of the Goshen (Straits) with the Collinsville Formation as an explanation for the stratigraphic asymmetry. The outer contact of the outer (Dg) unit is considered to be the base of the Goshen Formation because it locally adjoins the Russell Mountain Formation, which discontinuously underlies the Goshen north of the Granville dome. Knapp (1977, fig. 6)

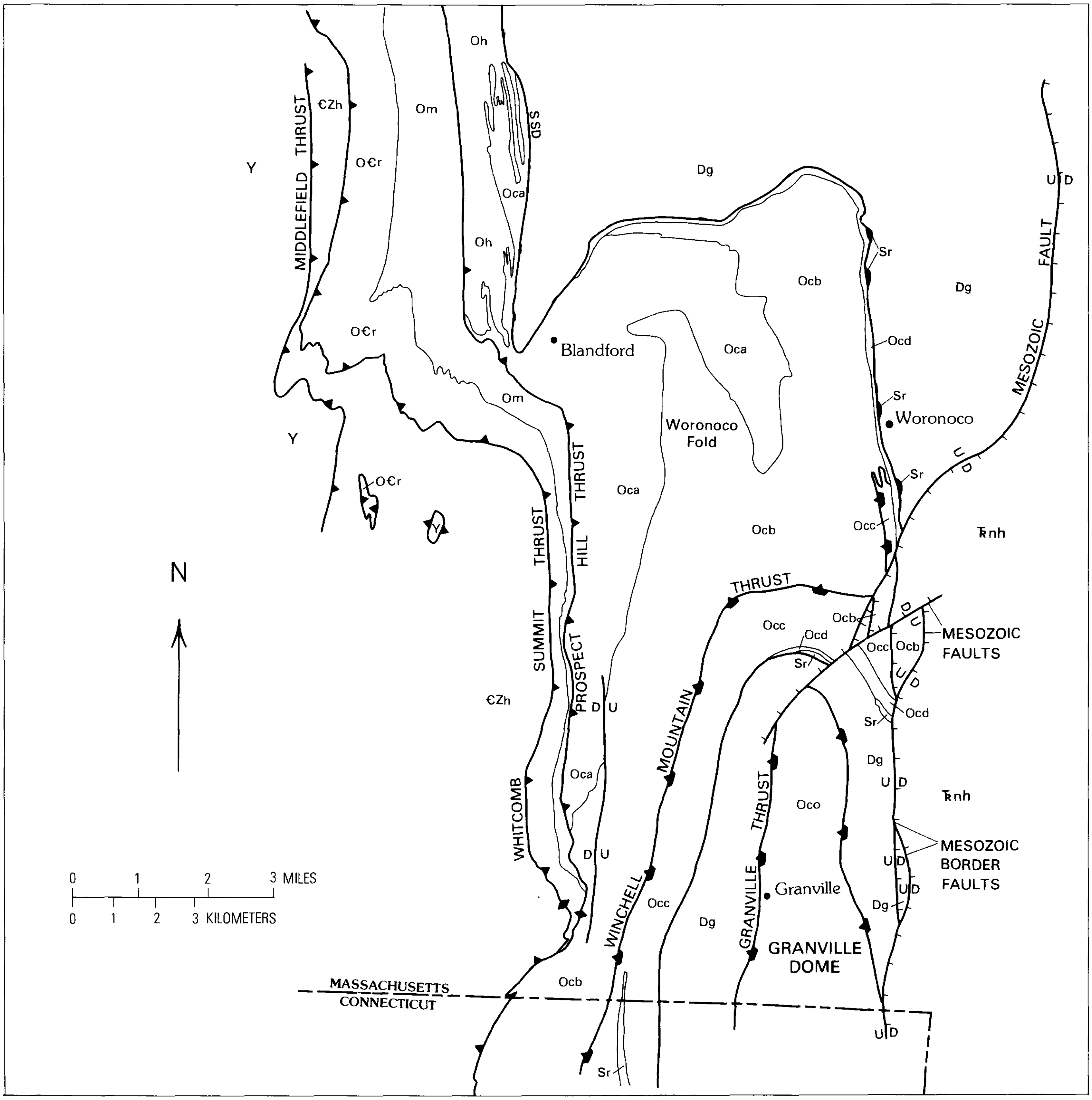


FIGURE 16.—Simplified geologic map of the Blandford-Woronoco-Granville area showing the Granville and Prospect Hill thrusts and their relations to the surrounding rocks. Explanation of letter symbols: Ŧnh, New Haven Arkose; Dg, Goshen Formation; Sr, Russell Mountain Formation; Oca, Ocb, Occ, Ocd, members of the Cobble Mountain Formation; Oh, Hawley Formation; Om, Moretown Formation; OꞒr, Rowe Schist; ꞒZh, Hoosac Formation; Y, Proterozoic Y rocks; SSD, surface of structural disharmony at the base of the Goshen Formation.

considered the possibility of a sedimentary facies change along the basal Goshen unconformity but discarded it because it would everywhere have to coincide with the axial surface of the isoclinal syncline and nowhere be visible at the surface.

What, then, is the relation between the Granville thrust and the basal Devonian surface of structural disharmony? As shown in figure 17*C*, the Granville thrust is rooted along the overturned limb of the pre-F_3 major west-facing nappe that folds the basal Devonian surface and possibly the major (F_2) isoclinal folds that are thought to be coeval with the basal detachment zone. This interpretation was used in constructing cross section *F-F'* of the State bedrock map (the pertinent part of which is shown here as figure 17*A*) where the Granville thrust is shown biting down into the Hoosac Formation. Other than the surface evidence that we have just described for the Granville thrust, we are not aware of any evidence in Massachusetts or Connecticut on the stratigraphic depth of the root zone, although the predominant east-over-west displacement of Acadian structures at this longitude suggests that it probably steps down to the east. There is no evidence for the Granville thrust north of the Granville dome; it is thus assumed to die out to the north. It is shown as absent in cross section *A-A'* and is represented by a small west-verging fold below the southern continuation of the Goshen dome in cross section *D-D'*. We believe that the Granville thrust nucleates in this fold and increases its westward displacement to the south, producing the stratigraphic asymmetry in the Goshen (Straits) Formation around the Granville dome and the domes in the eastern part of western Connecticut. Knapp (1977, p. 87-107) further discussed this problem and its application to the evolution of structures in Connecticut.

PROSPECT HILL THRUST

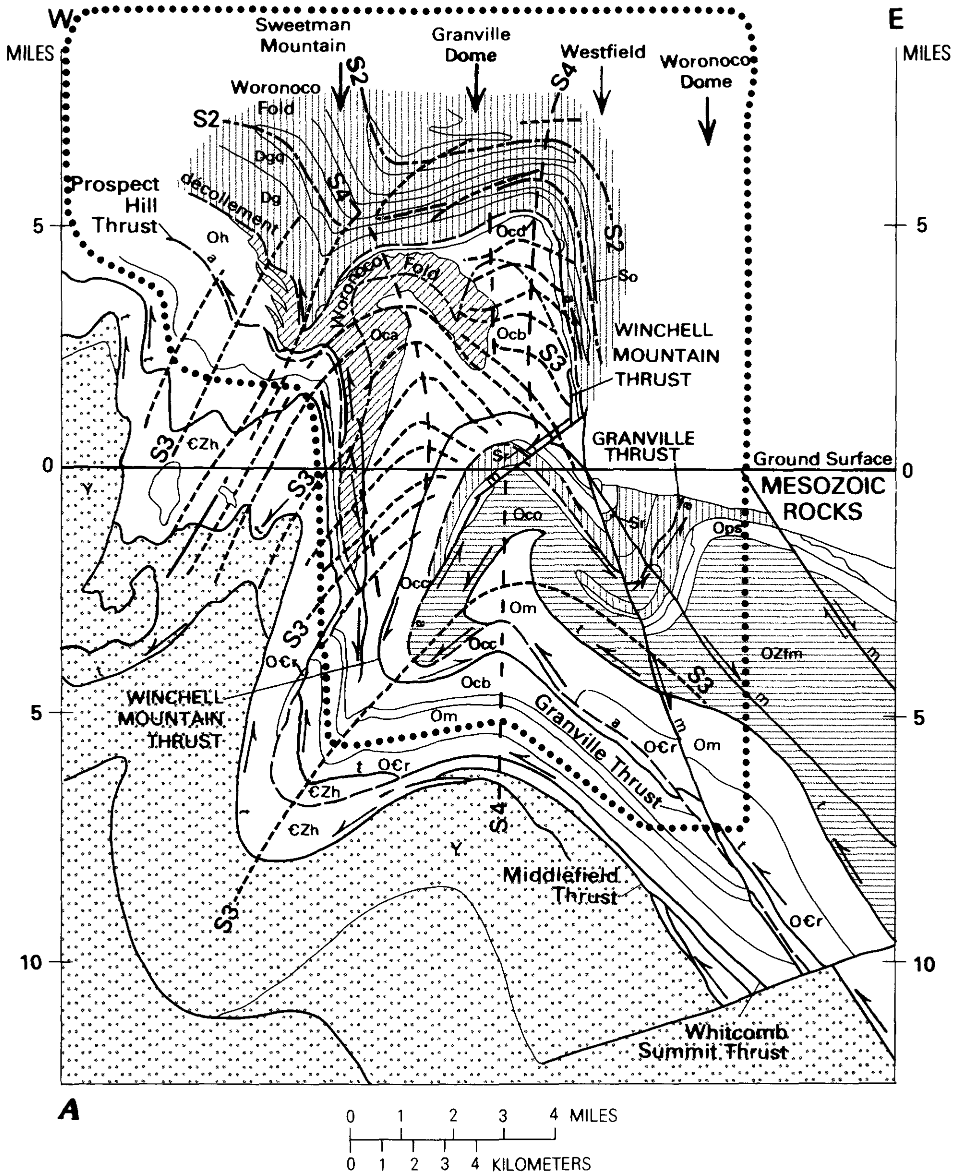
The Prospect Hill thrust is a narrow zone of thrust faults that essentially includes the basal unit, Ocar, of member A of the Cobble Mountain Formation. Although the Prospect Hill thrust is shown on the State bedrock map and figure 16 only by a fault along the western contact of unit Ocar, we believe that other, related thrust surfaces are present within Ocar. Thus, although the following discussion treats the Prospect Hill thrust as though it were a simple feature, it should be understood that in those areas where a unit Ocar is present, the Prospect Hill is more properly thought of as a zone of thrust surfaces.

The Prospect Hill thrust forms the western contact of the Cobble Mountain Formation south from the latitude of Blandford Village. Just south of the Connecticut State line it merges with the Taconian Whitcomb Summit thrust, or Cameron's line (fig. 16) (see also Stanley and Hatch, Ch. A, this volume).

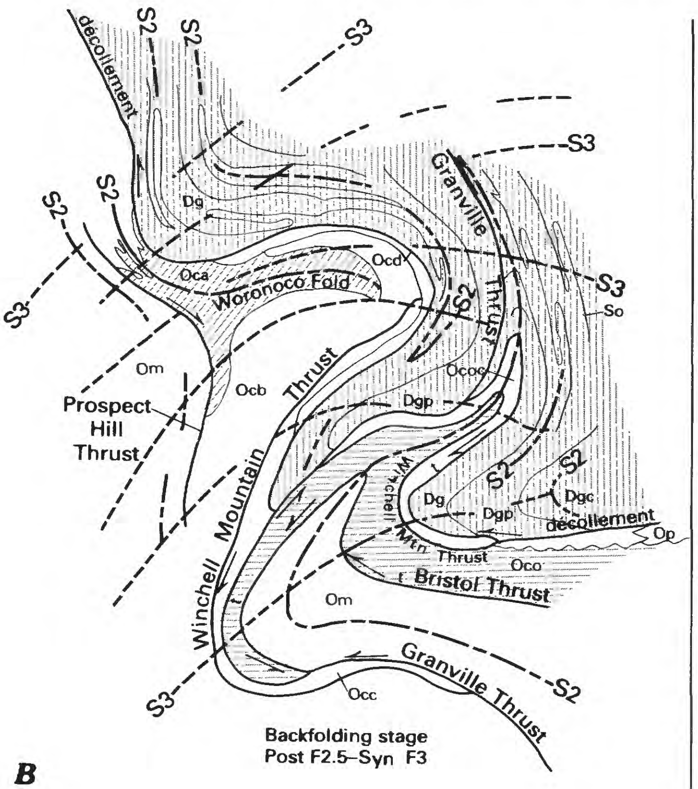
Evidence for this thrust derives in large part from the map pattern. South from the Blandford area, member A of the Cobble Mountain Formation thins progressively, and approximately at State Route 57 it is truncated in the upper plate of the thrust. Coincidentally, all of the Moretown Formation and almost all of the Rowe Schist are cut out in the lower or western plate, so that at the Connecticut State line member B of the Cobble Mountain is in contact with the tectonically thinned amphibolite that a few kilometers to the north is near the western part of the Rowe Schist.

The map pattern of the units of the Cobble Mountain Formation as compared with the pattern of the units to the west further documents both the existence and the character of the Prospect Hill thrust (fig. 16). The contact between members A and B of the Cobble Mountain Formation outlines a complex structure that Stanley (1967; 1975, p. 76) called the Woronoco fold. The fact that this structure is not reflected in the underlying units to the west indicates that the boundary, the Prospect Hill thrust, is a decollement with significant structural discontinuity across it. Unfolding the Woronoco fold results in a "gap" of approximately 10 km in the section directly west of (beneath) the thrust (fig. 18). This value, therefore, represents a minimum displacement on the Prospect Hill surface or zone. The actual net slip is not known, but it is considered to be across the mountain belt and not parallel to the trace of the thrust as constructed in figure 18.

A third line of evidence for the Prospect Hill thrust has been described by Knapp (1977, p. 39-49). At the Connecticut State line, Knapp mapped several bodies of quartz-microcline-muscovite-plagioclase-garnet blastomylonitic gneiss interlayered with silvery-gray gneiss and schist of member B of the Cobble Mountain Formation (fig. 19). All of these bodies of gneiss are within 100 m of the Prospect Hill thrust and contain distinctive layers, 3-30 cm thick, of large (2-6 cm) microcline and muscovite porphyroclasts, thin elongate quartz stringers, and finer grained matrix interlayered with finer grained blastomylonitic gneiss. These gneisses resemble some of the gneisses described by Ratcliffe (Ratcliffe and Mose, 1978; Ratcliffe and Hatch, 1979) from the Middlefield thrust zone. Knapp (1977, p. 46-49) suggested that the Prospect Hill gneisses may be Taconian intrusives that were subsequently caught up on the Acadian Prospect Hill thrust. An early Acadian age for the intrusives cannot, however, be ruled out. We here suggest that the Prospect Hill thrust started out in Taconian time as a



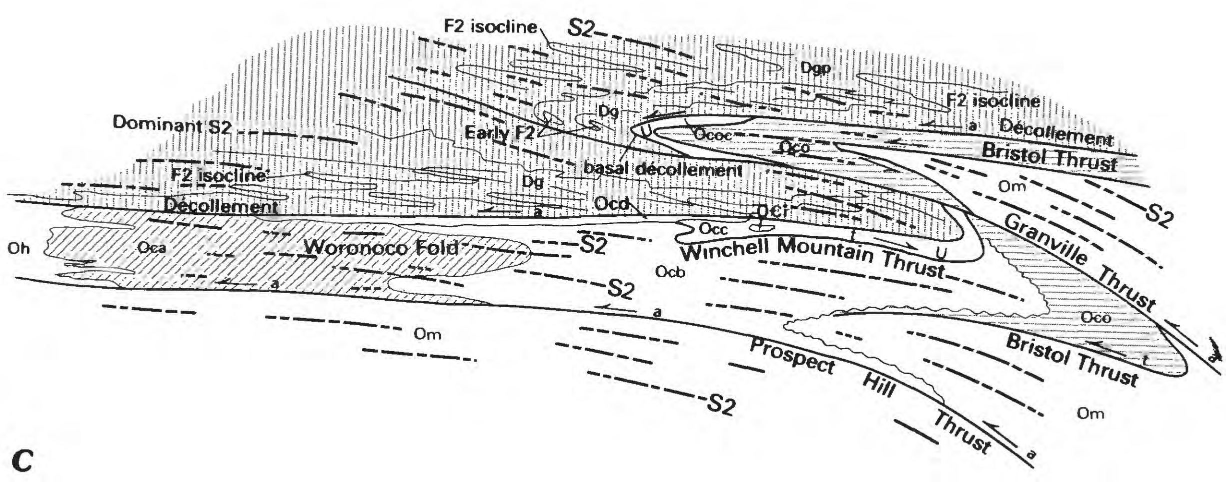
W
E
MILES
MILES
Sweetman Mountain
Granville Dome
Westfield
Woronoco Dome
Woronoco Fold
Prospect Hill Thrust
décollement
WINCHELL MOUNTAIN THRUST
GRANVILLE THRUST
Ground Surface
MESOZOIC ROCKS
Granville Thrust
Middlefield Thrust
Whitcomb Summit Thrust
A
0 1 2 3 4 MILES
0 1 2 3 4 KILOMETERS



B

splay off the Whitcomb Summit thrust zone, at which time the gneissic bodies were intruded along the thrust. Later, in the Acadian, the Prospect Hill fault was reactivated and much of the slip of the disconformable folding of the Woronoco fold was taken up along this surface. The small bodies of Taconian gneiss along the fault were sheared to produce a fluxion structure and then partly or totally recrystallized during the early stages of subsequent Acadian metamorphism to produce a foliation parallel to the dominant (S_2) regional schistosity. Stanley (1975, p. 76-77) has shown that the now deformed (by F_3 and F_4) axial surface schistosity of the Woronoco fold can be traced into the Goshen Formation where it is the dominant S_2 schistosity parallel to the axial surfaces of the F_2 isoclinal folds.

To the south, in northernmost Connecticut, Knapp (1977, p. 39) described minor F_2 folds truncated by the Prospect Hill thrust indicating some continued movement on the thrust in post-F_2 time. Displacement on this surface, however, ceased by F_3 time, and the surface was severely deformed by F_3 folds (fig. 3; Knapp, 1977, fig. 7) and subsequently folded during F_4 (fig. 4). Thus it seems well established that the decollement slippage along the Prospect Hill thrust is F_2 in age but that the fault probably also had earlier, Taconian, movement history.



C

FIGURE 17.—Systematic retrodeformation of the structure in the vicinity of State bedrock map cross section *F-F'*. *A* through *E* are cross sections that successively undeform the structure to the pre-Acadian configuration. Generalized axial surface traces S_2, S_3, and S_4 are superposed on the appropriate diagrams. S_0 is bedding. *A*, Geology as it is now envisioned at the western part of *F-F'*. Dotted line outlines geology shown in *B* through *E*. *B*, Configuration during backfolding stage after removal of most of F_4. *C*, Late F_2 time, after removal of the backfolding stage. *D*, After removal of the Granville and Prospect Hill thrusts and the Woronoco fold and the development of the basal Silurian-Devonian decollement with westward imbrication of bedding-plane thrusts and associated folds. These features are largely overprinted by major F_2 isoclines and the coeval Granville thrust of diagram *C*. *E*, Geology in pre-Acadian time, after deposition of the Goshen Formation. Crust of intermediate density shown by paired dot pattern. Mesozoic faults designated by a small "m"; Acadian thrusts by a small "a"; Taconian thrusts by a small "t". Dg, Dgq, Goshen Formation; Sr, Russell Mountain Formation; Oh, Hawley Formation; Oca, Ocb, Occ, Ocd, Cobble Mountain Formation; Oco, Ococ, Collinsville Formation; Ops, Partridge Formation; Om, Moretown Formation; O€r, O€ra, Rowe Schist; €Zh, Hoosac Formation; Y, Proterozoic Y rocks; c, beds of carbonate rock; U, pods of ultramafic rock.

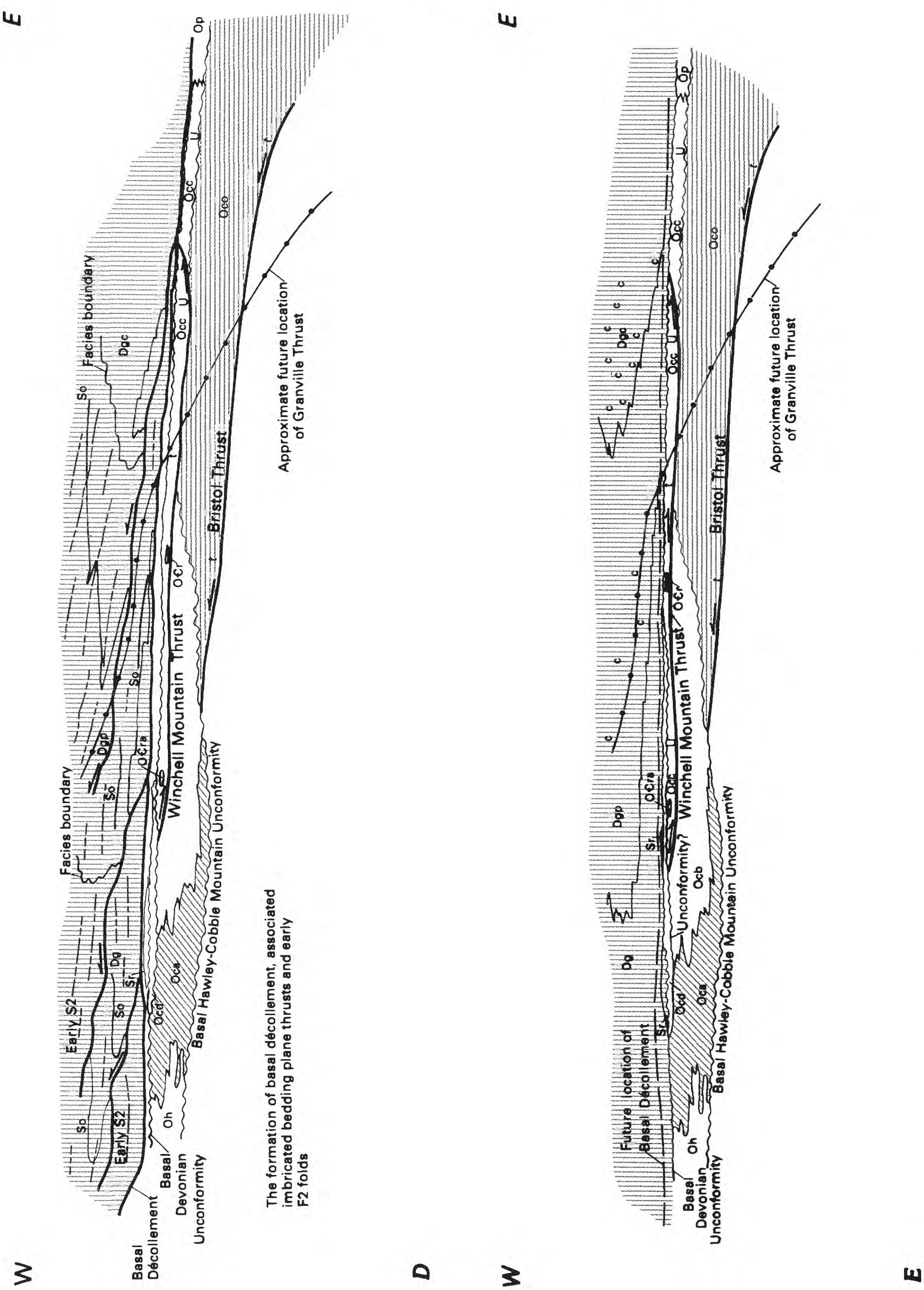


FIGURE 17.—Continued.

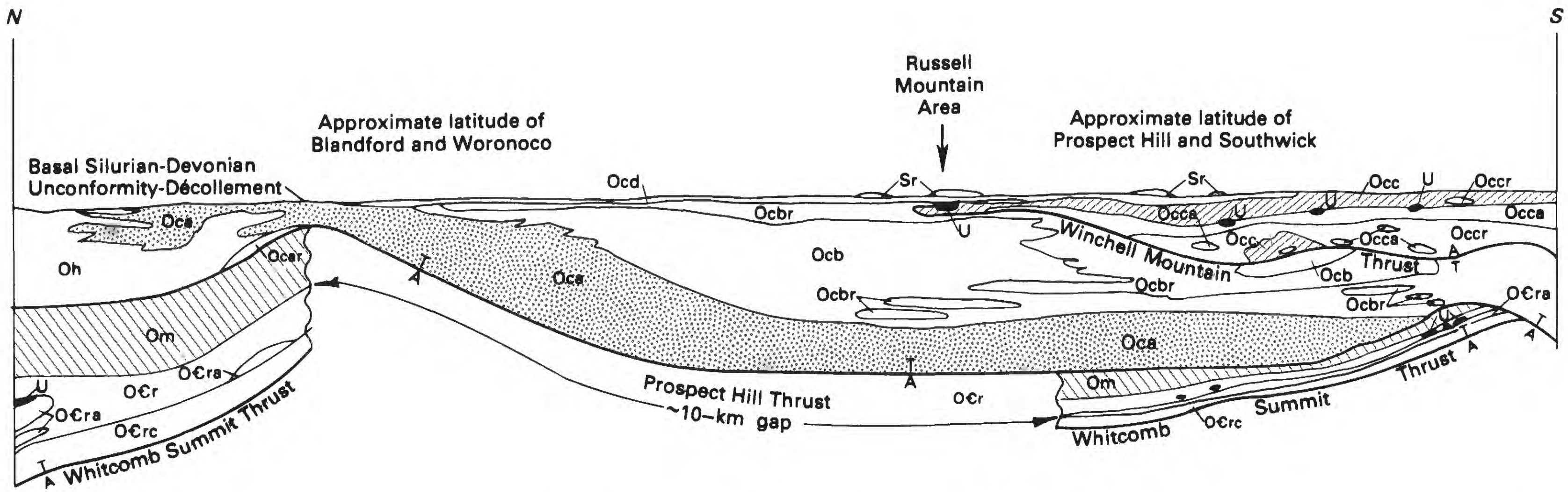


FIGURE 18.—Palinspastic diagram showing the present lithic distribution within the Cobble Mountain Formation and its relation to surrounding rocks. The Acadian Woronoco fold has been unfolded in this diagram resulting in a 10-km "gap" below the Cobble Mountain Formation. This distance is the displacement on the Prospect Hill thrust of F_2 age in the Acadian orogeny. The Winchell Mountain thrust is located at the base of member C. Member D rests unconformably on members A, B, and C. "A" and "T" indicate movement away from and toward the reader on faults. Sr, Russell Mountain Formation; Oh, Hawley Formation; Oca, Ocar, Occa, Ocb, Ocbr, Occ, Occr, Ocd, members of the Cobble Mountain Formation; Om, Moretown Formation; O€r, O€ra, O€rc, Rowe Schist; U, ultramafic rocks.

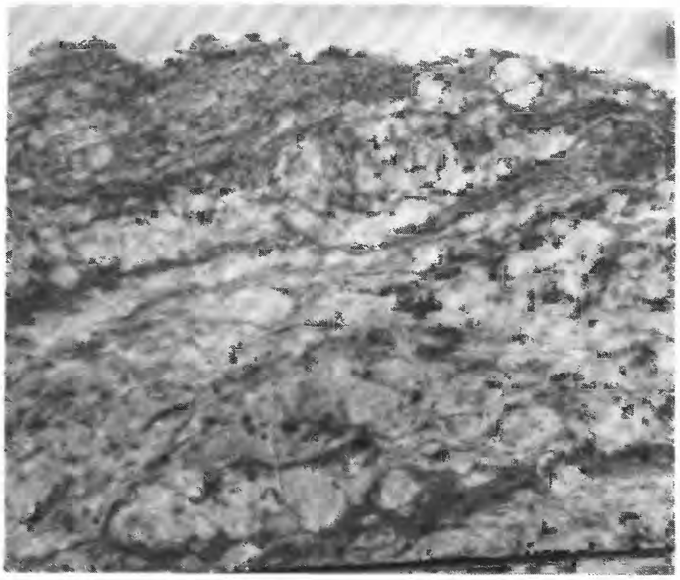


FIGURE 19.—Polished slab of mylonitized granite from sliver along the Prospect Hill thrust. Fractured potassium feldspar augen are surrounded by fine-grained recrystallized quartz, feldspar, and mica, which define the mylonitic foliation of S_2 age. Width of slab is 7 cm.

In the discussion of the "surface of structural disharmony" at the base of the Goshen Formation, we pointed out the paucity of large F_2 folds in the pre-Goshen strata (in striking contrast to the many large F_2 folds in the Goshen) as evidence for that disharmony. The Woronoco fold east of Blandford (fig. 16) is clearly a major exception to that generalization. F_2 movement on the Prospect Hill thrust and large F_2 folds below the base of the Goshen, which are not found to the north, both first appear at the latitude of Blandford. We propose that, south of Blandford, some of the F_2 isoclinal crumpling that to the north is confined to the Goshen and higher strata is taken up by the Cobble Mountain strata with the development of a decollement at the base of the Cobble Mountain Formation, the Prospect Hill thrust. The fact that the basal Goshen contact is not folded by the Woronoco fold indicates that that surface continued to be a surface of slip south of Blandford and that the slippage was simply taken up on, or shared by, two surfaces instead of one. Therefore, in southwestern Massachusetts, and possibly also to the south in western Connecticut, the east-over-west Acadian displacement is distributed on the basal Devonian decollement, the Granville thrust, the Prospect Hill thrust, and the F_2 isoclines.

EAST-VERGING FOLDS OF SECTION *F-F′*

In order to show the geometry of the rocks in the vicinity of cross section *F-F′* of the State bedrock map, the surface geology to the north has been projected into the plane of the section. The west-over-east folds of section *F-F′* are distinctly different from folds in the sections to the north. They are similar, however, to southeast-verging folds in the Pelham dome and the Keene dome of the Bronson Hill anticlinorium (Robinson, 1967, 1979) that predate formation of the domes but have not been successfully dated with respect to the regional west-verging fold nappes. The problems here are how did the east-verging folds form and when did they form relative to the overall sequence of structural events west of the Mesozoic basins. Two surfaces clearly describe the geometry: the contact between members A and B of the Cobble Mountain Formation, which outlines the Woronoco fold (fig. 16), and the basal Devonian unconformity, which deflects eastward in the Blandford-Woronoco area and mantles the domes to the south. Unfortunately, part of this geometry is cut off by the Mesozoic fault to the east. The pervasive S_2 Acadian schistosity is clearly coeval with the Woronoco fold and the isoclinal folds in the stratigraphically higher Goshen Formation (figs. 2, 17*A*; Stanley, 1975, fig. 63). Obviously, all the older Taconian thrust zones (Middlefield, Whitcomb Summit, Rowe, and Winchell Mountain) (Stanley and Hatch, Ch. A, this volume), as well as the "surface of Acadian structural disharmony" and the Granville thrust, are affected by this younger west-over-east structure.

In the west half of cross section *F- F′*, S_3 surfaces cut across member and formation contacts at a low angle producing counterclockwise (east-over-west) asymmetrical folds that characterize F_3 throughout western Massachusetts (fig. 3). S_3 surfaces strike more easterly and flatten as they are traced eastward. In the area west of the Woronoco dome and elsewhere (figs. 3, 4), S_3 is deformed by F_4 into large cleavage antiforms and synforms that are geometrically related to the Granville dome and the distorted configuration of the Woronoco fold (fig. 17*A*). S_3 is clearly coeval with the large syncline in the basal Devonian contact at Blandford, whereas the anticline in this surface north of Russell is at least in large measure an F_4 structure.

As pointed out earlier in this chapter and elaborated by Stanley (1975), the Acadian evolution of cross section *F-F′* can be understood by systematically unfolding S_3 and S_2 about their respective axes of rotation (for example, unfold S_3 about F_4). The task is made simple, fortunately, by the approximate parallelism of F_3 and F_4 hinges in this area. Figure 17 shows this evolution in reverse beginning with the present configuration of the rocks along section *F- F′* and ending with a planar and horizontal basal Devonian unconformity after the deposition of Goshen facies Dg, Dgp, and Dgc, but before the onset of Acadian deformation.

The evolution of the section, then, is read by going from figures 17*D* to 17*A*.

The mapped distribution of the members of the Cobble Mountain Formation and their relationship to the Hawley and Collinsville Formations is based on our discussion in the chapter on the Rowe-Hawley zone (Stanley and Hatch, Ch. A, this volume) and is only briefly summarized here. The black schist and mafic volcanic rocks of the Hawley are a lateral facies of Cobble Mountain member A, which is in turn equivalent to the basal part of Cobble Mountain member B. Both the Hawley and the Cobble Mountain rest unconformably on the Bristol Member of the Collinsville (all units in Oco except Ococ as shown on the State bedrock map explanation) (Stanley, 1964, 1980; Stanley and Hatch, Ch. A, this volume) in the domes where the Collinsville is in tectonic contact along the Bristol thrust with the Moretown Formation (cross sections *A-A′*, *D-D′*, *F-F′*, State bedrock map). Member C of the Cobble Mountain Formation, which is a distal facies of B, has been thrust eastward along the Winchell Mountain thrust. Member D of the Cobble Mountain unconformably overlies members A, B, and C, as well as the Winchell Mountain thrust. The basal Devonian unconformity then cuts across the section biting deeper into the older rocks as it is traced northward (Stanley and Hatch, Ch. A, this volume, fig. 27). The three facies of the Goshen are derived from map relations around the Granville dome, the Woronoco dome, and the Taconic unconformity. The poorly bedded, calc-silicate-bearing facies of the Goshen (Dgp) stratigraphically overlies the well-bedded, calc-silicate-poor facies (Dg), which in turn rests on discontinuous lenses of the Russell Mountain Formation around the Granville dome. Similar relations are found to the north near the Woronoco dome. The calcareous facies (Dgc) of the Goshen mapped around the Woronoco dome is shown on figure 17 as a facies of Dgp although the two are nowhere seen in contact (Hatch and others, Ch. B, this volume).

The early stages of Acadian deformation were marked by undetermined westward displacement of the Devonian section on the basal thrust, intense west-facing recumbent isoclinal folding, formation of the Granville thrust (which dies out to the north before cross section *D-D′*), and development of the penetrative regional S_2 schistosity. In cross section *F-F′* and in figure 17*C*, the Granville thrust is shown developing after the basal surface of disharmony, although there is no "on-the-ground" evidence for this in southern Massachusetts.

On the north flank of the Shelburne Falls dome, Leo M. Hall (written commun., 1977) mapped a small, west-facing, recumbent F_2 fold that folds the basal Goshen surface (cross section *A-A′*, State bedrock map). Hall also reported that he believed that the dominant schistosity in the Goshen (our S_2) crenulated an older schistosity. We interpret Hall's observations and similar observations by Simpson (1974) to mean that the early (D_2) Acadian deformation was intense and complex and involved the development of a series or continuum of westward-directed structures that formed as a consequence of the westward-transported nappes of the Bronson Hill zone. It is very likely that, during this event, many earlier Acadian folds and the schistosity associated with earliest movement on the basal surface were obliterated and transposed by slightly later isoclinal folds resulting from continued westward movement on the western parts of the surface. Locally, as in the case described by Hall, some of the basal surface itself may have been caught up and folded over in a recumbent fold as a result of that westward movement. Evidence for multiple folding and syn-slip surface age would be seen only at the hinges of such F_2 folds as those observed by Hall (1977) and Simpson (1974). The hypothesized recumbent fold associated with the Granville thrust in figure 17*C* would be formed by this same mechanism.

The earlier formed, westward-directed structures were then subjected to west-over-east backfolding that began in southern Massachusetts and dominated the geology in western Connecticut (Stanley, 1975; Scott, 1974; Hall, 1980, figs. 2, 3). We suggest that backfolding began after the initial formation of the domes ($F_{2.5}$) and continued during F_3. Figure 17*B* shows the inferred geometry. The gentle curvature of S_3 simply reflects the very early deformation that culminated in F_4 and resulted in the configuration of figure 17*A*.

What, then, was the cause of the west-over-east backfolding? The answer, we suggest, lies in cross section *F-F′*. As we indicated above, this section is located along a promontory in the eastern boundary of the Grenville plate that may have continued to influence subsequent Acadian structures. West of the Woronoco fold is the imbricated structure of the Berkshire massif and its thrust-bound packets of Ordovician, Cambrian, and older rocks. This architecture was formed largely during the Taconian orogeny (Stanley and Hatch, Ch. A, this volume) and was only mildly remolded during the Acadian orogeny, which dramatically increased in intensity eastward from the massif. Westward displacement before and during F_2 Acadian deformation tectonically thickened the sequence to the west, thus increasing the resistance to further movement. In contrast, rocks to the east were substantially weakened by high temperatures associated with kyanite-sillimanite-grade metamorphism and the intrusion of many large and small masses of the Williamsburg

Granodiorite. We suggest, therefore, that the "back-folding stage" resulted from incipient underthrusting of the weaker eastern crust beneath the more resistant, tectonically stacked western crust. This abortive "subduction" never reached the stage of shearing apart along a thrust, as it did east of the Bronson Hill anticlinorium, but it certainly rotated the older structures in much of western Connecticut so that the dominant Acadian schistosity now dips to the west.

FAULTS ELIMINATED FROM NEW STATE BEDROCK MAP

In his various maps of the area discussed in this chapter, Emerson (1892; 1898a,b; 1917) showed three faults that are not shown on the new State map. The following brief discussion presents our reasons for eliminating those faults.

Emerson's earlier (1892, 1898b) maps, but not his 1917 State map, showed a fault along the west margin of his Hawley Schist (largely equivalent to our Hawley Formation) extending south from the Deerfield River to the vicinity of West Cummington. As evidence for this "great fault," Emerson (1898b, p. 172) cited the alignment of two small manganese prospects with the "area of iron-manganese in Hawley" and the apparent truncation of amphibolite bands in the Hawley (Schist) as shown on all his maps. From our mapping (Osberg and others, 1971; Hatch, 1969), both the Hawley, or Forge Hill, iron deposit and the various manganese mines and prospects are stratigraphically controlled and show no evidence of faulting. Furthermore, the iron deposit is within the upper part of the Moretown Formation, whereas the manganese deposits are within one of the carbonaceous schist units near the base of the Hawley. Thus, although the iron and manganese deposits are stratigraphically close, they are in distinctly different lithologies that are characteristic of two different formations. Secondly, and perhaps more significantly, our mapping (Osberg and others, 1971; Hatch, 1969) shows remarkable continuity of minor and major map units in both the Moretown (equivalent to the upper part of Emerson's Savoy Schist) and the Hawley in the area where Emerson mapped the fault. Thus, although we cannot disprove a fault along the Moretown-Hawley contact, we feel that Emerson's basis for it has been largely eliminated. We should point out, however, that Martha M. Godchaux has described intercalation of Moretown-like schist with metavolcanic rocks similar to metavolcanic rocks of the Hawley, in the area of the Moretown-Hawley contact just north of the Deerfield River (oral commun. to Hatch, Oct. 1980). This intercalation could be interpreted as either sedimentary or tectonic.

On all three (1892, 1898b, 1917) of his maps of the area, Emerson showed a fault trending north and then northwest across the Goshen dome (his Goshen anticline). His only mention of it is in his description of the "Goshen anticline" where he refers to a "fault crack along the crest having a considerable upthrow on its west side" (1898b, p. 175). His maps show the fault as offsetting numerous contacts, including granite sills in the Goshen Formation, the core rocks-Goshen contact, and his Goshen-Conway contact. Outcrop in the vicinity of his fault was insufficient at the time of our mapping (Hatch and Warren, 1981) to document any offset of the core rock-Goshen contact or to demonstrate continuity or discontinuity of the many sills of Williamsburg Granodiorite. It was also insufficient to prove continuity or discontinuity of axial traces of the many (F_2) isoclinal folds around the dome. We saw no field evidence for brecciation, cleavage, or other structures parallel to the trace of the fault as shown by Emerson. Thus, once again, we could only fail to support Emerson's fault and certainly did not disprove it.

Emerson's maps show a third, smaller, fault offsetting his Chester Amphibolite and his Rowe-Savoy contact southeast of Florida near the major bend in the Deerfield River. Although our mapping (Chidester and others, 1967) shows the geology of the redefined Rowe Schist to be extremely complicated in that area, we once again saw no evidence for northwest-trending faulting. Instead, we suggest that the present pattern of map units is more likely the result of superposition of Acadian folding on imbricate thrust faulting described in the chapter on the Rowe-Hawley zone (Stanley and Hatch, Ch. A, this volume). Once again, however, disproving a cross fault in that area would be very difficult, and such a fault could be present.

METAMORPHISM

As indicated on the metamorphic map inset of the State bedrock map, all the stratified rocks of the Rowe-Hawley zone and Connecticut Valley belt have undergone regional metamorphism during the Acadian orogeny at grades ranging from garnet to sillimanite. Although no comprehensive discussion of the metamorphism will be attempted here, a few brief points should be made.

Some quadrangle maps (Osberg and others, 1971; Hatch, 1969; Hatch and others, 1970) show isograds in the Goshen Formation that terminate at the base of the Goshen. The isograds were so terminated for the simple reason that neither of the indicator minerals (staurolite or kyanite) that defined the isograds in the Goshen rocks was seen in the pre-Silurian strata west of the Goshen in any of those quadrangles. The termination of those isograds was not meant to imply that they were cut off or structurally terminated, but rather that the authors found no field data on which to extend them westward. Cheney and others (1980) studied the assemblages and the chemistry of the individual phases on both sides of this surface and concluded that "the complex distribution of isograds on the east limb of the Berkshire anticlinorium likely results from variation in rock composition and/or polygenetic history as suggested by Hatch and Stanley (1976) rather than syn and/or post metamorphic thrusting." Hatch (1975) demonstrated that the thermal maximum of regional metamorphism in the Connecticut Valley and Rowe-Hawley zones slightly postdated F_3 folding and associated cleavage. The apparent continuity of metamorphic isograds across the basal Goshen surface, therefore, neither precludes nor supports the possibility of net movement along that surface during F_2 isoclinal folding.

Abbott (1979) made a detailed study of the metamorphism of the Goshen pelitic schists in the vicinity of the Goshen dome. There he carefully documented the major prograde event previously described, and he identified a retrogressive event followed by a local prograde event in post-F_4 time in the vicinity of the dome.

Although the major kyanite and higher zones of the major regional event appear to be geographically coincident with the areas of exposure of the Williamsburg Granodiorite, no thermal metamorphic aureoles per se were recognized around those exposures or any other igneous bodies in the Rowe-Hawley or western Connecticut Valley zones.

Neither we, nor John Cheney in his ongoing studies of the metamorphism of the rocks of western Massachusetts, nor Sutter and Hatch (1985) have recognized any evidence of Taconian metamorphism in the rocks of the Rowe-Hawley zone.

TECTONIC SUMMARY

Figure 20 is a generalized north-south time-space diagram summarizing the Acadian tectonic events between the Mesozoic basin and the Berkshire massif. The increase in metamorphic grade and its relation to the fold generations are based on data in Hatch (1975, p. 57-60, fig. 55). The Williamsburg Granodiorite saturates much of the eastern part of the region southeast of the kyanite isograd (see State bedrock map) and is shown diagrammatically in figure 20. Justification for the chronology of each of the events has been discussed in previous sections of this paper and will not be repeated here. It is clear from our previous discussion and figure 20 that D_2 was a very tectonically active time in the Rowe-Hawley zone and Connecticut Valley belt.

Fundamental to the diagram is the assumption that each fold generation developed simultaneously throughout the belt. However, the variation in fold intensity within any one generation and the irregularity of plate margins strongly suggest that most fold generations are indeed time transgressive. We believe this to be particularly true for F_3 and F_4. To the south, in the Blandford-Woronoco area, the backfolding of F_3 is well developed; it decreases progressively northward. F_4 is also well developed to the south but is less pervasive, though present, to the north. As we have suggested in the sections on backfolding, dome formation, and the Cobble Mountain Formation, the region between the Blandford and Woronoco area and the Waterbury dome in Connecticut acted as a pressure point during both the Taconian and Acadian orogenies. We believe that this pressure point resulted from mutually opposing promontories on the eastern margin of the Grenville plate and the western margin of the Bronson Hill plate, with which the Grenville plate collided in the Taconian orogeny (Stanley and Hatch, Ch. A, this volume). Strain would clearly develop first in these regions and would progress outward from the pressure point. As a result, deformation would be most severe here and would diminish outward. These promontories probably influenced F_2 in the same way, but compression was so intense throughout the belt that the resulting structures were pervasively strained to the same level.

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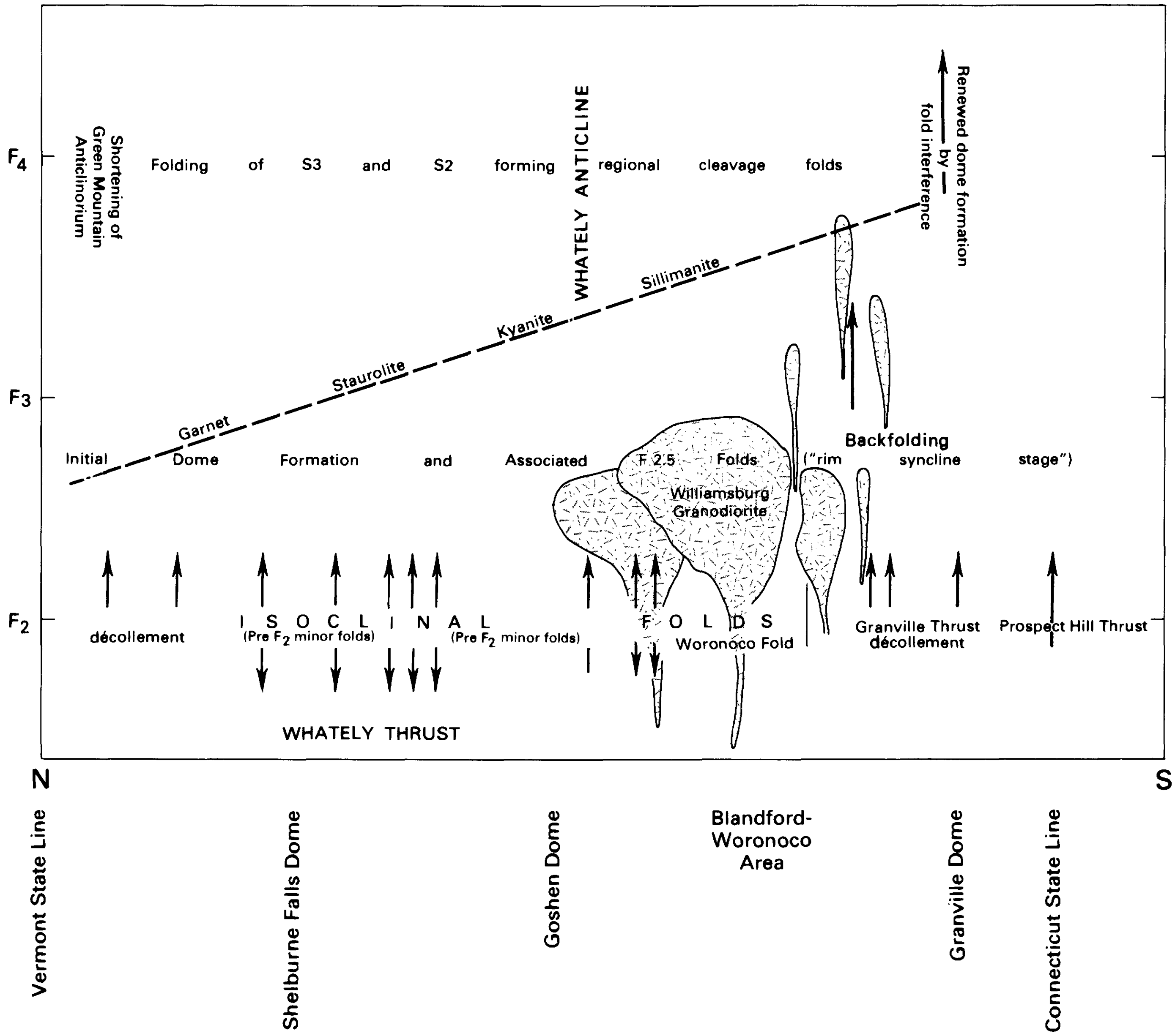

FIGURE 20.—Time-space diagram for the Acadian orogeny west of the Mesozoic basins. The vertical arrows indicate the approximate time range during which the indicated events were active.

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The Whately Thrust: A Structural Solution to the Stratigraphic Dilemma of the Erving Formation

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THE BEDROCK GEOLOGY OF MASSACHUSETTS

U.S. GEOLOGICAL SURVEY PROFESSIONAL PAPER 1366-D

Different stratigraphic sequences within the Lower Devonian strata of the Connecticut Valley belt are explained by the west-verging Whately thrust

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ILLUSTRATIONS

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THE WHATELY THRUST: A STRUCTURAL SOLUTION TO THE STRATIGRAPHIC DILEMMA OF THE ERVING FORMATION

By PETER ROBINSON,[1] NORMAN L. HATCH, JR., and ROLFE S. STANLEY[2]

ABSTRACT

Compilation for the bedrock geologic map of Massachusetts demonstrated apparent incompatibilities between the Silurian and Lower Devonian stratigraphic sequences east and west of the Mesozoic basins in Massachusetts. This problem involves rocks and structures treated in other chapters of this volume, but its specialized nature requires detailed treatment presented here in this separate chapter.

The light-gray granulites and black amphibolites of the Erving Formation are thought to have once formed a continuous blanket across the area of the present Bronson Hill anticlinorium. In the Northfield and Wendell synclines of northern Massachusetts, the Erving rests on Lower Devonian Littleton Formation and, locally, on Lower Silurian Clough Quartzite, Middle Ordovician Partridge Formation, and Ordovician or older Fourmile Gneiss, suggesting an unconformity at the base of the Erving. In the Leverett area to the south, the Erving rests on either the Partridge or the Clough. Near Quabbin Reservoir, to the southeast, the Erving rests directly on the Partridge. In the Wilbraham area in southernmost Massachusetts, the Erving rests on local lenses of Clough or on Partridge, or on pre-Partridge Middle Ordovician Ammonoosuc Volcanics. The Wilbraham area is unique in central Massachusetts in that the Erving there is overlain by presumably younger strata here assigned to the Waits River Formation of Early Devonian age.

West of the Mesozoic basins in the east limb of the Whately anticline, the apparent stratigraphically upward sequence is Partridge Formation, local Clough Quartzite, Erving Formation, Gile Mountain Formation, Littleton Formation. A few meters east of the Gile Mountain-Littleton contact at Whately a spectacular graded channel deposit near the base of the Littleton suggests that the Littleton beds stratigraphically overlie adjacent Gile Mountain beds.

The Gile Mountain-Littleton contact that disappears beneath the Mesozoic Deerfield basin north of Whately reappears from beneath the north end of the basin north of Greenfield. From thence the contact continues northward into southern Vermont, where it has been called the Chicken Yard line. Local graded beds in both units in Vermont near the line again suggest stratigraphic tops east into the Littleton.

Field relations at Windmill Mountain, Vt., on the east flank of the Athens dome, further complicate the enigma. Here a sequence of strata strikingly similar to the Erving rocks at Whately, Mass., have been assigned to the Silurian Shaw Mountain Formation. Clearly a Silurian "Erving" in Vermont is incompatible with a post-Littleton (Early Devonian) Erving in Massachusetts. Similarly, a pre-Littleton Gile Mountain Formation along the Chicken Yard line and at Whately is incompatible with the post-Littleton Gile Mountain and Waits River relations in the Bronson Hill anticlinorium.

Various stratigraphic "solutions" to the enigma all require that one or more of the observed rock types appear twice in the stratigraphic sequence. Difficulties with all such resolutions are severe. The resolution followed on the State bedrock map and the one we find least objectionable, though far from flawless, proposes that the sequence east of the Mesozoic basins is the "true" sequence and that the relations west of the basins are explained by a thrust, the Whately thrust, that carried the Littleton Formation westward at least 20 km from the vicinity of the Bronson Hill anticlinorium onto the Gile Mountain-Waits River and Erving Formations.

Acceptance of the Whately thrust model implies that the Merrimack and the Connecticut Valley belts were probably a single Silurian and Devonian sedimentary trough in which sedimentation spread westward through time. Westward movement on the Whately thrust was probably closely related in time and transport direction to the west-verging isoclinal folds (to the west) and nappes (to the east). The thrust may be a key to understanding the Connecticut Valley metamorphic low between higher grade rocks to the east and west.

INTRODUCTION

During our many years of field work in Massachusetts, culminating with the preparation of the cross sections and a "correlation of map units" for the State bedrock map (Zen and others, 1983), we became increasingly aware of apparent incompatibilities between the Silurian-Early Devonian stratigraphic sequences east and west of the Hartford and Deerfield Mesozoic basins (fig. 1). Because this is a rather specific, as well as particularly knotty, problem that involves rocks discussed in other chapters (B and C) of this volume, we have chosen to treat it by itself in this

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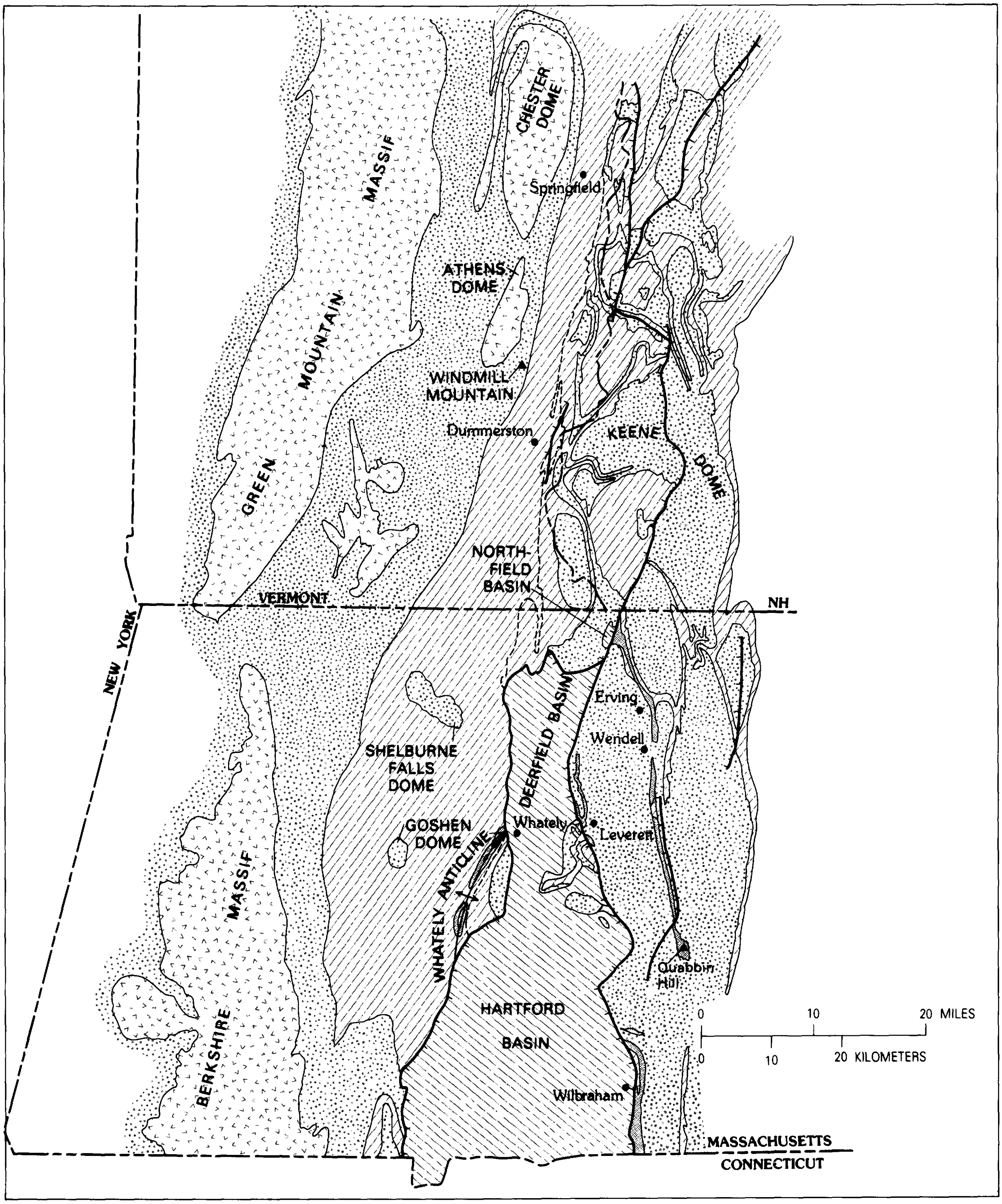

FIGURE 1.—Generalized geologic map of part of western New England showing location of areas and features discussed in text. Dashed line is Chicken Yard line separating Littleton Formation on the east from Gile Mountain Formation on the west.

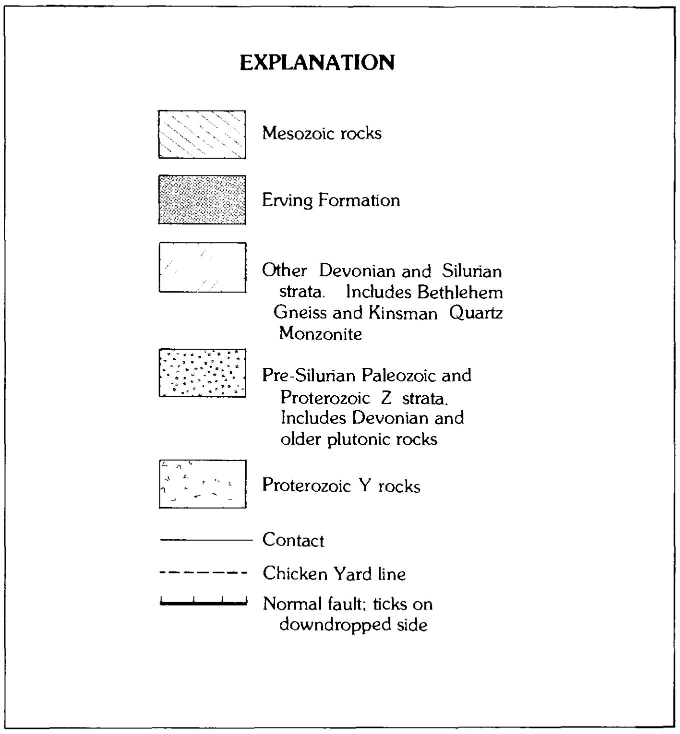

FIGURE 1.—Continued.

separate chapter where we can properly spell out the evidence, the problems, and our thoughts on a preferred resolution.

The rocks in question were deposited in a broad north-trending basin on the eastern margin of North America in the interval between the Taconian and Acadian orogenies. The Lower Devonian rocks are primarily graphitic metamorphosed shale, graywacke, and calcareous sandstone and probably were originally flysch turbidites. The Erving Formation stands out from the other Lower Devonian rocks by being lighter colored and non-graphitic and by containing conspicuous lenses of pink garnet-quartz "coticule" granulite[3] and abundant amphibolites, which are metamorphosed basaltic volcanic rocks. The stratigraphic position of the Erving is critical to the following discussions. The present aggregate thickness of the Lower Devonian strata is estimated to range from a few hundred to a few thousand meters; the variability results from a combination of differences in original thickness, subsequent erosion along local unconformities, and tectonic thinning and thickening. The underlying Silurian rocks are primarily metamorphosed quartzite, conglomerate, limestone, and calcareous shale, all apparently of shallow-water origin, with an aggregate thickness of a few tens to locally as much as a few hundred meters. All the Silurian and Lower Devonian rocks were metamorphosed and intensely deformed during the Acadian orogeny.

The apparent stratigraphic sequences of these Silurian and Lower Devonian strata east and west of the Mesozoic basins appear to be mutually incompatible (table 1). East of the basins, in the Northfield and Wendell synclines, at Leverett, at Quabbin Hill, and at Wilbraham (fig. 2), mutually consistent relations produce a composite stratigraphic sequence overlying the pre-Silurian rocks of (from base to top) Clough Quartzite, local Fitch Formation, Littleton Formation, Erving Formation, and Waits River Formation. West of the basins, near Whately and Leeds, available field evidence supports a sequence of Clough, Erving, Gile Mountain, and Littleton. The boundary between the Littleton and Gile Mountain Formations along strike in Vermont has been called the Chicken Yard line, and data for determining stratigraphic tops at a few localities suggest that the Littleton overlies the Gile Mountain and associated Waits River.

In the following pages we describe in detail the field relations in five areas in Massachusetts that are critical to the identification of the problem. We then discuss relations in southern Vermont and adjacent New Hampshire that pertain to the problem. Possible stratigraphic and structural resolutions to this apparent enigma are all subject to valid objections. In the model presented here as the least objectionable, and the one portrayed on the State bedrock map, we propose that the sequence east of the Mesozoic basins is the "true" sequence and that the relations west of the basins are explained by a thrust, the Whately thrust, that carried the Littleton Formation at least 20 km westward from the vicinity of the Bronson Hill anticlinorium onto the Gile Mountain-Waits River and Erving Formations.

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[3]The terms granulite and granofels are used interchangeably here.

TABLE 1.—*Apparent stratigraphic sequences in selected areas of west-central and western Massachusetts and eastern Vermont pertinent to the stratigraphic position of the Erving Formation. Columns 1-4 are east of the Mesozoic basins; columns 5-7 are west or northwest of them*

[— no data]

1 Northfield and Wendell synclines (fig. 3)	2 Leverett (fig. 4)	3 Quabbin Hill (fig. 5)	4 Wilbraham (fig. 6)	5 Whately anticline (figs. 7, 8)	6 Chicken Yard line area, Vermont (fig. 2)	7 Windmill Mountain, Vermont (fig. 1)	
—	—	—	—	Littleton	Littleton	—	
—	—	—	—	—	Putney	—	
—	—	—	Waits River	Gile Mountain	Gile Mountain	Waits River	
—	—	—	—	—	—	Northfield	
Erving	Erving	Erving	Erving	Erving	—	Granulite	Shaw Mtn.
Littleton	—	—	—	—	—	Amphibolite	
Clough	Clough	—	Clough	Clough	—	Quartzite	
Partridge	Partridge	Partridge	Partridge	Partridge	—	Cram Hill	

ERVING FORMATION OF THE BRONSON HILL ANTICLINORIUM

The Erving Formation is exposed in four major areas of the Bronson Hill anticlinorium east of the Mesozoic basins in Massachusetts (fig. 1). The type area near the village of Erving lies in the Northfield and Wendell synclines between the Warwick dome and the Kempfield anticline, respectively, to the east, and the Pelham dome to the west (fig. 2). The thick Erving Formation in the Leverett area on the west limb of the Pelham dome would be connected to the Erving on the northeast limb if the dome were not truncated by the Mesozoic Connecticut Valley border fault (fig. 2). The Erving Formation of the Quabbin Hill area (fig. 2) has a tenuous and poorly exposed northward connection through the extremely narrow southern extension of the Wendell syncline. South of Quabbin Hill, the Erving Formation is completely truncated by the intrusive contact of the Belchertown pluton. The Erving Formation (and the Waits River Formation) of the Wilbraham area (fig. 2) lies on the west limb of the Glastonbury dome and is truncated to the west by the Connecticut Valley border fault. Remnants of the Erving Formation also occur on the north end of the Glastonbury dome against the southern intrusive contact of the Belchertown pluton and as an inclusion in the northwestern part of the pluton (fig. 2). A strong argument can be made that the base of the Belchertown on its south and east sides closely follows the base of the Erving Formation and that in the absence of the intrusion the Wilbraham and Quabbin Hill areas would be connected on the surface. Ignoring the truncation by the Mesozoic border fault, a similar argument can be made to connect the Erving Formation of the Leverett area with the Quabbin Hill and Wilbraham areas (fig. 2). Thus, the evidence suggests that the Erving Formation once formed a continuous blanket directly east of what is now the location of the Mesozoic basins.

NORTHFIELD AND WENDELL SYNCLINES

The quartzite and gray mica schist of the New Hampshire part of the Northfield syncline were first assigned to the Clough Quartzite and Littleton Formation by Moore (1949). He did not include the Erving Formation in this package because it extends into New Hampshire only a few hundred feet and is not exposed there. Balk (1956a,b) refused to accept the correlation of Moore and included the contents of the Northfield and Wendell synclines in his Crag Mountain Formation. Earlier, B.K. Emerson (1898, 1917) had named certain prominent amphibolites near the village of Erving, Erving Hornblende Schist, and certain associated quartz-plagioclase granulites Savoy Schist or Whetstone Schist. Robinson (1963) extended Moore's Clough and Littleton into Massachusetts and established the combination of Emerson's amphibolites and granulites as the Erving Member of the Littleton Formation. This usage of the Erving as a member of the Littleton persisted through 1967 (Robinson, 1967) in the mistaken belief that these rocks might correspond to some upper part of the Littleton in the type area near Littleton, N.H. (Billings, 1937). However, the clear lithologic differences between the upper part of the Littleton in its type area and the Erving in Massachusetts were demonstrated during a field trip in 1966, and the Erving was subsequently established as a separate formation (Thompson and others, 1968).

The dominant rock type in the Erving Formation of the Northfield and Wendell synclines (fig. 3) is gray,

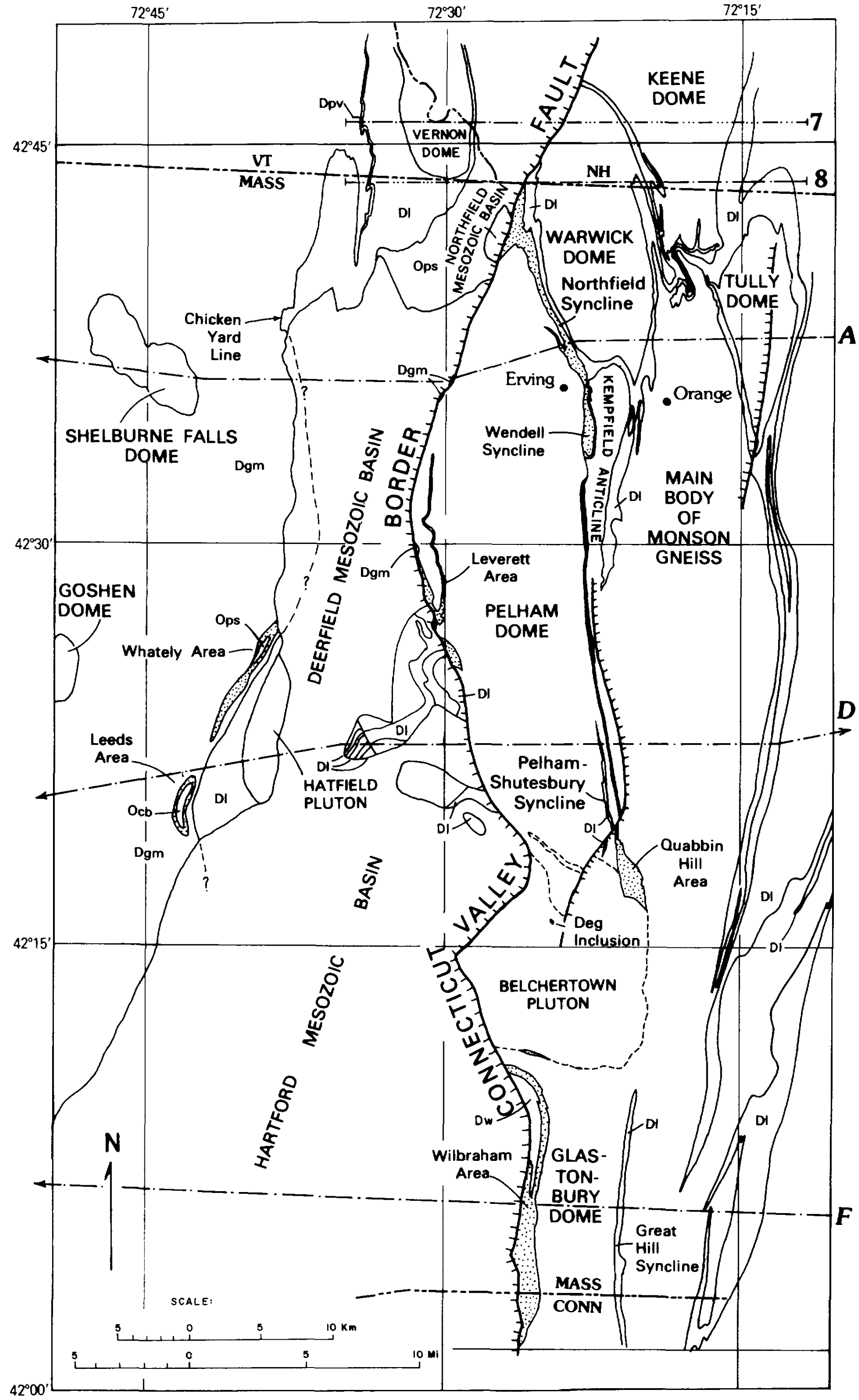


FIGURE 2.—Simplified geologic map of part of central Massachusetts showing distribution of Lower Devonian stratigraphic units. Erving Formation is stippled. Other units pertinent to discussion are indicated by letter symbols (which follow the symbols on the State bedrock map): Ocb, member B of the Cobble Mountain Formation; Ops, Partridge Formation; Dpv, Putney Volcanics; Dl, Littleton Formation; Dgm, Gile Mountain Formation; Dw, Waits River Formation; Deg, granofels and schist of the Erving Formation. Lines 7 and 8 at north edge of map are discussed in figure 11. Lines A, D, and F are the lines of cross sections A, D, and F on the State bedrock map. The queried dashed line across the west side of the Deerfield Mesozoic basin is the inferred extension of the Chicken Yard line beneath the basin.

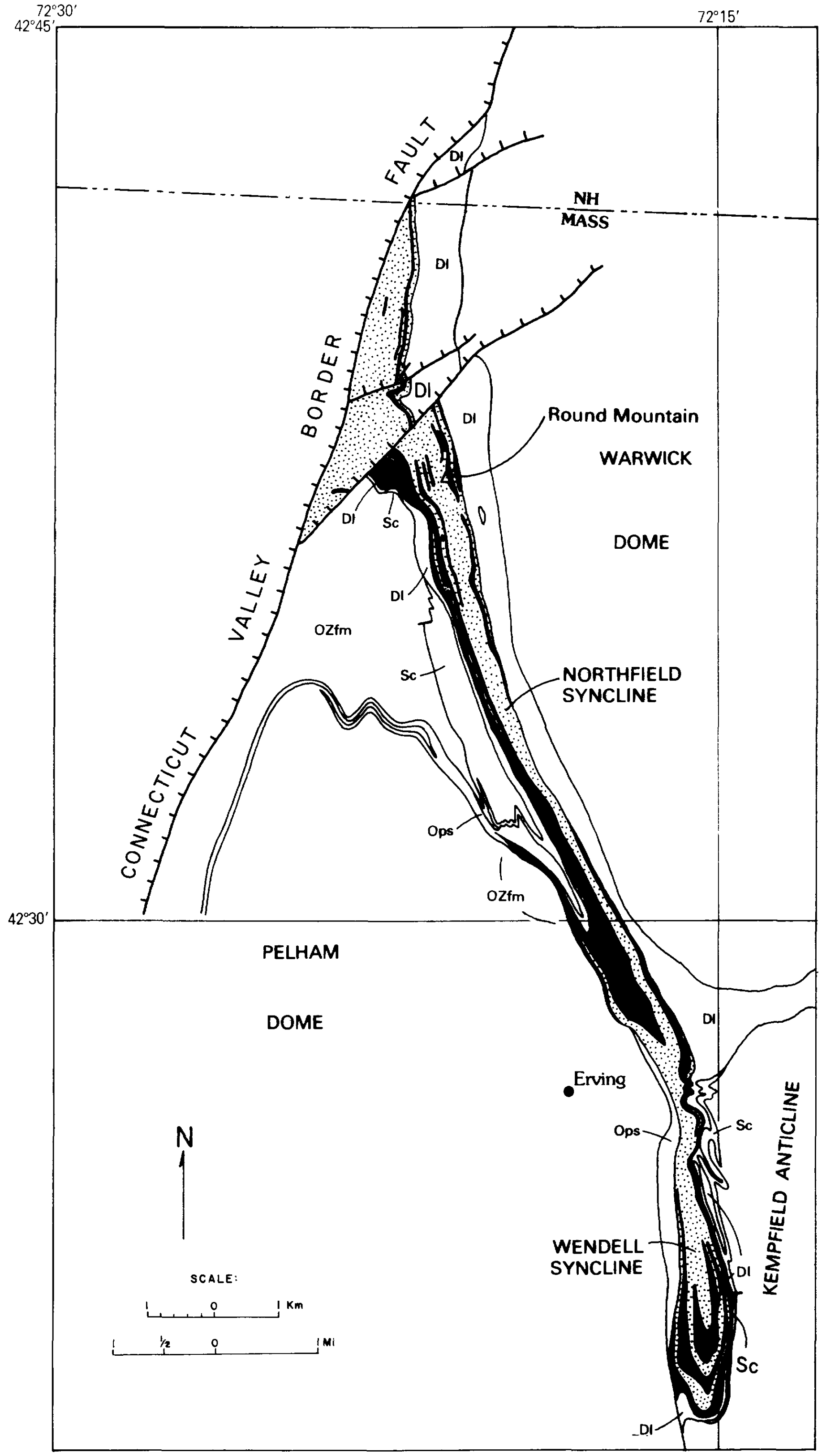


FIGURE 3.—Geologic map of the Northfield and Wendell synclines showing distribution of Erving Formation amphibolite (black) and granulite (stippled). Other symbols are the same as on the State bedrock map: OZfm, Fourmile Gneiss; Ops, Partridge Formation; Sc, Clough Quartzite; Dl, Littleton Formation. Leaders from symbol OZfm are to a band of Fourmile Gneiss that locally is in contact with Erving Formation. Contact between Fourmile and older units in the Pelham dome is not shown.

well-bedded, fine-grained quartz-plagioclase-biotite granulite interbedded with very thin to thick (1-cm to 2-m) beds of muscovite-biotite schist commonly containing dodecahedral garnet, ilmenite, and staurolite and (or) kyanite. This schist is distinct from schist of the adjacent Littleton Formation, which has abundant graphite, lacks kyanite, has ferromagnesian minerals that are slightly more iron rich, and has trapezohedral garnets. The Erving granulite contains minor but nearly ubiquitous lenses and layers of fine-grained calc-silicate granulite with conspicuous garnet and hornblende, along with fine-grained diopside, clinozoisite, or zoisite. Calc-silicate rock is locally predominant in the lowest mapped Erving granulite lenses along the east limb of the Northfield syncline; small outcrops at the Littleton contact on the east ridge of Round Mountain, Northfield, consist of dark hornblende calc-silicate rock containing white 1-cm euhedral prisms of zoisite.

The second most abundant rock type of the Erving Formation is hornblende-andesine-epidote-sphene amphibolite in sharply bounded mappable layers from 0.5 m to tens of meters thick within granulite. The predominant textural type is thinly laminated, has alternate hornblende-rich and epidote-plagioclase-rich layers, and is interpreted to have been laminated fine-grained basaltic tuff. Some light-colored layers also contain abundant diopside. Less commonly, the amphibolite is coarse grained and massive with prominent 1- to 2-cm hornblende megacrysts suggesting porphyritic flows or coarse crystal tuffs. The megacrysts commonly contain inclusions of diopside and calcite. Vague suggestions of pillows are present in one outcrop. The Erving amphibolite is relatively easily distinguished from adjacent amphibolite outcrops of the Partridge Formation and Ammonoosuc Volcanics by its mineralogical monotony and distinctive range of textures. The mapped amphibolite layers of the Erving tend to be concentrated near the base of the formation and to dominate the outcrops near the hinge of the Wendell syncline and along the west limb of the Northfield syncline. The hornblende megacrystic amphibolite is characteristic of the lowest amphibolite layers, although it does occur elsewhere.

In the Erving granulites, commonly within 1-2 m of amphibolite contacts, are 2.5- to 30-mm-thick beds or 5-mm-sized irregular patches of pink coticule granulite containing extremely fine grained euhedral manganese-bearing garnet. The matrix surrounding the garnet is mainly a quartz-biotite granulite, typically with magnetite that either is fine grained or, more commonly, occurs as discrete octahedra as much as 5 mm in diameter. The octahedra occur also within the solid garnet coticule beds, where they commonly have feldspar pressure shadows. Unlike many coticule beds found in the Ordovician Ammonoosuc Volcanics and Partridge Formation (Robinson, 1963), the Ordovician Hawley Formation (Hatch, 1969; Hatch and others, 1970), and the Lower Devonian Littleton Formation[4] (Huntington, 1975), all of which contain abundant grunerite, gedrite, or hornblende as well as magnetite, the Erving coticules typically are dominated by green biotite with or without muscovite and potassic feldspar. The origin of these coticules is still problematical, but they appear to have been derived from chemical precipitates that are somehow related in space and time to basaltic volcanic activity. Commonly the coticule beds retain delicate and complex folds not preserved in surrounding beds. The detailed characteristics of Erving coticules have been invaluable for regional correlation and have even been used to suggest close affinities with the Standing Pond Volcanics of southeastern Vermont (Robinson, 1963).

The top of the Erving Formation is not exposed in the Northfield and Wendell synclines although a thickness of at least 900 m is exposed near Northfield. Contact relations at its base are crucial to its stratigraphic interpretation (fig. 3). On the east limb of the Northfield syncline, the Erving is in sharp contact with Littleton Formation that is as much as 750 m thick. The Littleton thins southward and locally pinches out on the east limb of the Wendell syncline where the Erving rests directly on the Clough Quartzite. On the west limb of the Wendell syncline, the Erving cuts down onto the Partridge Formation and, in the Northfield syncline, locally onto the Fourmile Gneiss of the Pelham dome. Further north, the contact cuts upward across the Partridge into the Littleton, which it then follows to the north end of the Pelham dome. At one point, at the north end of the dome, the Erving rests directly on the Clough; where last seen near the Mesozoic border fault, it overlies a few meters of Littleton. Although no depositional features such as conglomerates are present at the sharp basal contact of the Erving, the contact relations suggest an unconformity. It should be remembered that the apparent unconformable relations need not be due *entirely* to post-Littleton pre-Erving erosion but could be in part due to the well-known unconformity at the base of the Clough and a younger one postulated at the base of the Littleton (Robinson, 1963).

LEVERETT AREA

The Erving Formation in the Leverett area (fig. 4), on the west limb of the Pelham dome, is dominated by

[4]Note added in proof: Robinson (1987) has shown that the coticule beds studied by Huntington (1975) correlate with the Silurian Perry Mountain Formation of southwestern New Hampshire.

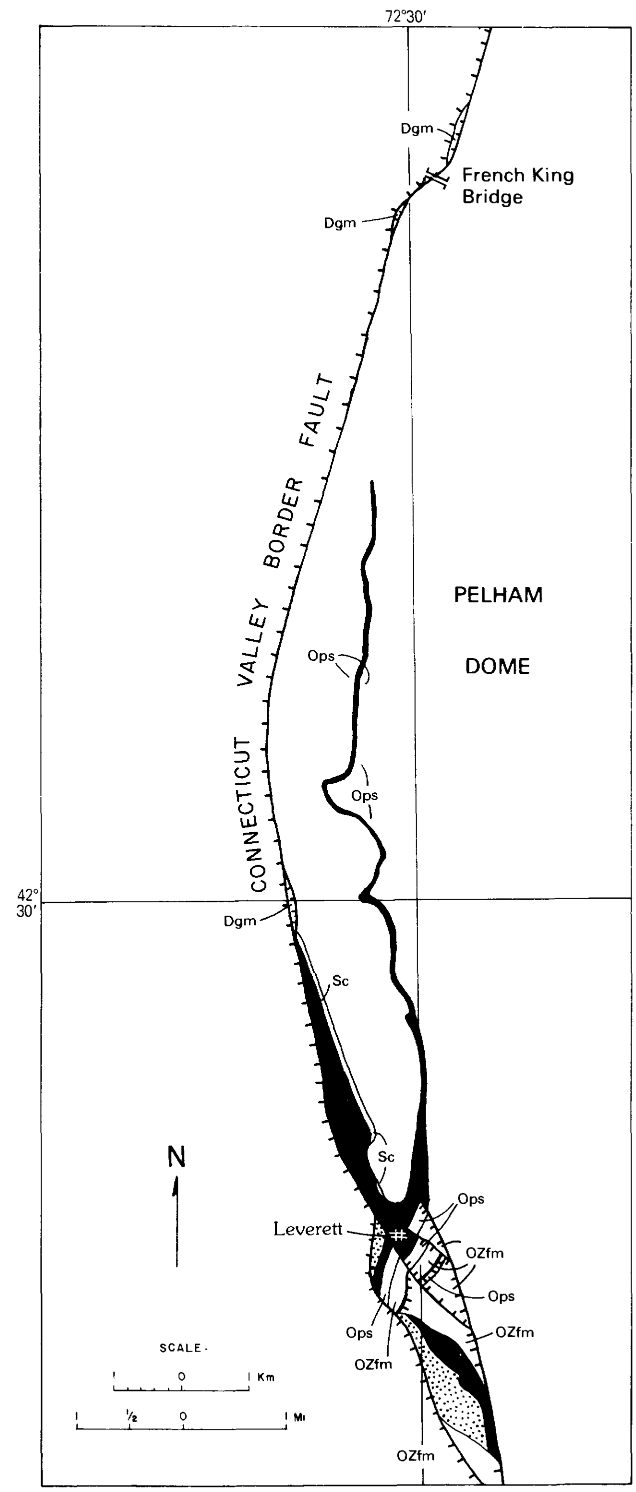


FIGURE 4.—Geologic map of the Leverett area showing distribution of Erving Formation amphibolite (black) and granulite (stippled). Other symbols are the same as on the State map: OZfm, Fourmile Gneiss; Ops, Partridge Formation; Sc, Clough Quartzite; Dgm, Gile Mountain Formation. Leaders from symbols Ops are to areas of Partridge Formation in contact with Erving. Other pre-Silurian rocks east of the Connecticut Valley border fault are not differentiated.

amphibolite but also has areas of granulite large enough to map that contain schists, calc-silicates, and coticules characteristic of the type area. The largest area of granulite, in southern Leverett, lies west of and apparently stratigraphically above the thick basal zone dominated by amphibolite. The main belt along the Connecticut Valley border fault is extensively affected by Mesozoic hydrothermal alteration, and its structure is complicated in southern Leverett by a complex of Mesozoic faults. An excellent analysis of the amphibolite from southern Leverett was published by Emerson (1917).

Due to the Connecticut Valley fault, no strata are preserved overlying the Erving in the Leverett area. The Erving rests for the most part on the Ordovician Partridge Formation and locally on the Clough Quartzite. These contact relations, including the apparent absence of the Littleton Formation, are entirely consistent with the contact relations in the Northfield and Wendell synclines where the Littleton is clearly pinching out westward on the east side of the Pelham dome, presumably as a result of a pre-Erving unconformity.

QUABBIN HILL AREA

The exposures of the Erving Formation on the shores of Quabbin Reservoir, at Quabbin Hill, and on the end of Prescott Peninsula (fig. 5*A*) are the most spectacular in the region (Halpin, 1965; Robinson, 1967), although the map pattern, structure, and petrology are complicated by late-stage folds and retrograde metamorphism associated with buttressing effects at the northeast corner of the Belchertown pluton (Halpin, 1965; Guthrie and Robinson, 1967; Robinson, 1967; Guthrie, 1972). The main amphibolite of the area close to the base of the formation underlies the peak of Quabbin Hill, where it is 400 m thick. This relict volcanic pile thins fairly abruptly northward to a continuous layer about 13 m thick on Prescott Peninsula and thins to the southwest to a string of boudins less than 1 m thick on the shore of Quabbin Reservoir west of Quabbin Hill. The granulite of the Quabbin Hill area is, for the most part, entirely typical of the Erving. The mica schist tends to be dominated by muscovite and biotite, generally lacking the abundant garnet, staurolite, and kyanite seen to the north. However, the granulite that lies above the amphibolite on Prescott Peninsula (fig. 5*B*) contains the only known graphitic schist layer in the Erving and also contains garnet, staurolite, and kyanite (Robinson, 1967). The coticule exposures, all within meters of the main amphibolite layer, are spectacular and were extensively collected by Robert Balk (written commun. and unpub. maps, 1940). The basal unit on Prescott Peninsula is about 10-40 m thick and is dominated by

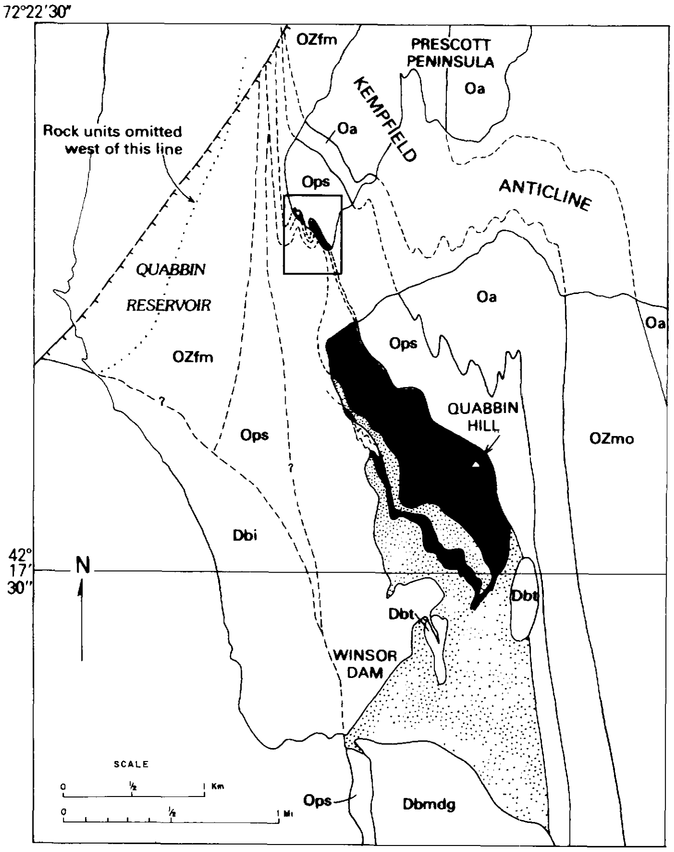
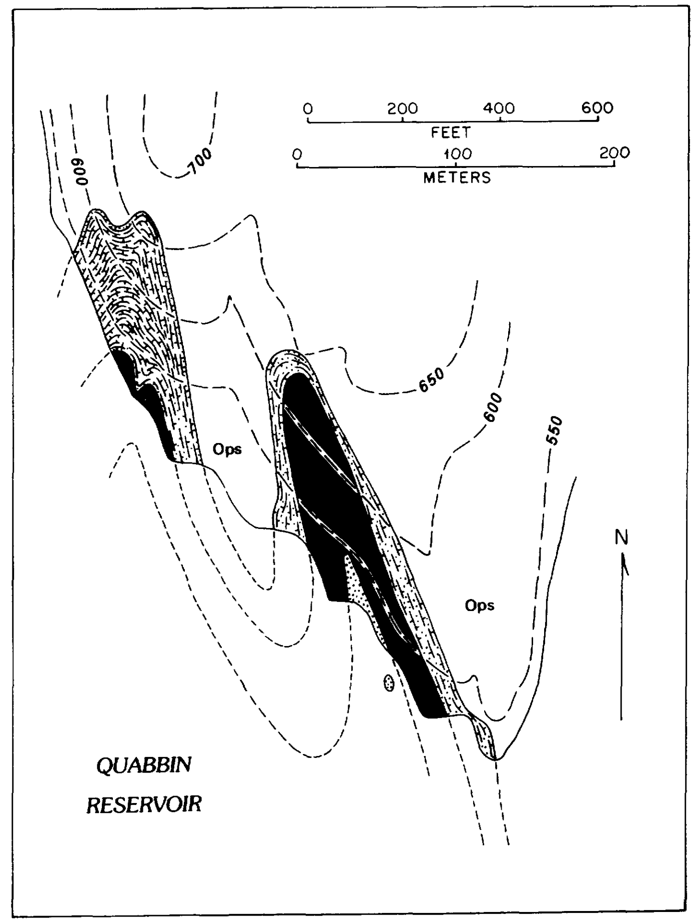


FIGURE 5A.—Geologic map of the Quabbin Hill area showing distribution of Erving Formation amphibolite (black), granulite (stippled), and calc-silicate and marble (brick pattern). Other symbols are the same as on the State bedrock map: OZfm, Fourmile Gneiss; Oa, Ammonoosuc Volcanics; Ops, Partridge Formation; Dbi, intrusive breccia; Dbt, biotite tonalite; Dbmdg, hornblende quartz monzodiorite gneiss of the Belchertown pluton; OZmo, Monson Gneiss. Box outlines area shown in figure 5B.



B

FIGURE 5B.—Geologic map showing details of the point of Prescott Peninsula. Topographic contours in feet. Symbols as in figure 5A.

calc-silicate rocks, including coarse diopside-zoisite-microcline calc-silicate rock with matted diopside and zoisite prisms 5-10 cm long. Also present, particularly in the lowest 1 or 2 m, are beds of dark hornblende-rich calc-silicate rock with white prisms 1-2 cm long of zoisite identical with the basal hornblende-zoisite rocks on Round Mountain in the Northfield syncline. As this basal calc-silicate unit is traced northwestward around the set of late folds on Prescott Peninsula (fig. 5B), it grades into a finely laminated brown-weathering muscovite-bearing marble, with the same hornblende-zoisite calc-silicate at the base. Although this basal calc-silicate unit bears some resemblance to the Silurian Fitch Formation near Orange to the north (see State bedrock map or fig. 2), the specific correlation of the hornblende-zoisite rock with rocks of the Northfield syncline makes its assignment to the Erving Formation more likely.

In the Quabbin Hill area, no rocks are exposed above the Erving Formation, which rests everywhere on the Ordovician Partridge Formation. The eastern Erving-Partridge contact is particularly well exposed on the tip of Prescott Peninsula. The location of the west contact with the Partridge is known near the west end of Winsor Dam from diamond drill cores made in the 1930's. To the south, and probably to the west, the Erving is cut off by the Belchertown pluton (Dbmdg in fig. 5A). To the north, beneath the west arm of Quabbin Reservoir, a connection (fig. 2) is postulated with the narrow southern extension of the Wendell syncline, which has a narrow belt of Erving amphibolite and minor granulite along its west side. Although the Quabbin Hill area of Erving Formation provides us with no stratigraphic information with respect to younger or older Silurian and Devonian units, its contact relations are consistent with areas to the north, and it provides an important link in lithic correlation to the south.

WILBRAHAM AREA

The Erving Formation of the Wilbraham area (fig. 6) lies mainly on the west limb of the Glastonbury dome and is cut off to the west by the Connecticut Valley border fault. The map pattern of the Erving-Waits River contact suggests that the area may contain the axial surface of a major syncline (Wilbraham Syncline of Peper, 1967) and that the rocks exposed on the western edge of the area may be on the west limb of the syncline. This interpretation is reflected in the pattern of cross section *F-F′* of the State bedrock map. Two small patches of Erving Formation are present on the north end of the Glastonbury dome, between the Ammonoosuc Volcanics and the quasi-concordant intrusive contact of the Belchertown Complex (fig. 6; Leo and others, 1977).

The Erving Formation of the Wilbraham area consists predominantly of granulite with a greater than normal abundance of mica schist beds. These rocks bear a close resemblance to the granulite and schist exposures near Quabbin Hill but appear to have reached slightly higher metamorphic grade, as shown by local patches containing sillimanite (Leo and others, 1977). Local lenses of amphibolite, both near the base and higher in the section, are similar to the amphibolite described above. Several lenses near the base labeled "Dev" (fig. 6) caused particular trouble during the mapping (Leo and others, 1977). They consist of a mixture of amphibolites and gray to rusty-weathering mica schists, some of which are similar to nearby outcrops of the Middle Ordovician Partridge Formation. However, every lens either is surrounded by more distinctive Erving rock types or lies stratigraphically above exposures of Clough Quartzite, thus virtually requiring assignment to the Silurian-Devonian part of the stratigraphy.

In the Wilbraham area, the Erving Formation rests on three isolated lenses of the Clough Quartzite, on the Partridge Formation, or directly on the Ammonoosuc Volcanics. This is the only area where Erving-Ammonoosuc contacts have been reported, although the two units come within a few hundred meters of each other near Quabbin Hill and in the Wendell syncline. As pointed out above, much of this apparent unconformity at the base of the Erving may be due to pre-Clough erosion. Because it is not known whether the Littleton Formation was ever deposited in the Wilbraham area, one can only speculate as to the extent of pre-Erving erosion. However, we do know that the Clough and the Littleton are present in the Great Hill syncline east of the Glastonbury dome (fig. 2).

The Wilbraham area is unique in Massachusetts east of the Connecticut Valley Mesozoic basin in that the *top* of the Erving Formation is present and a thickness of

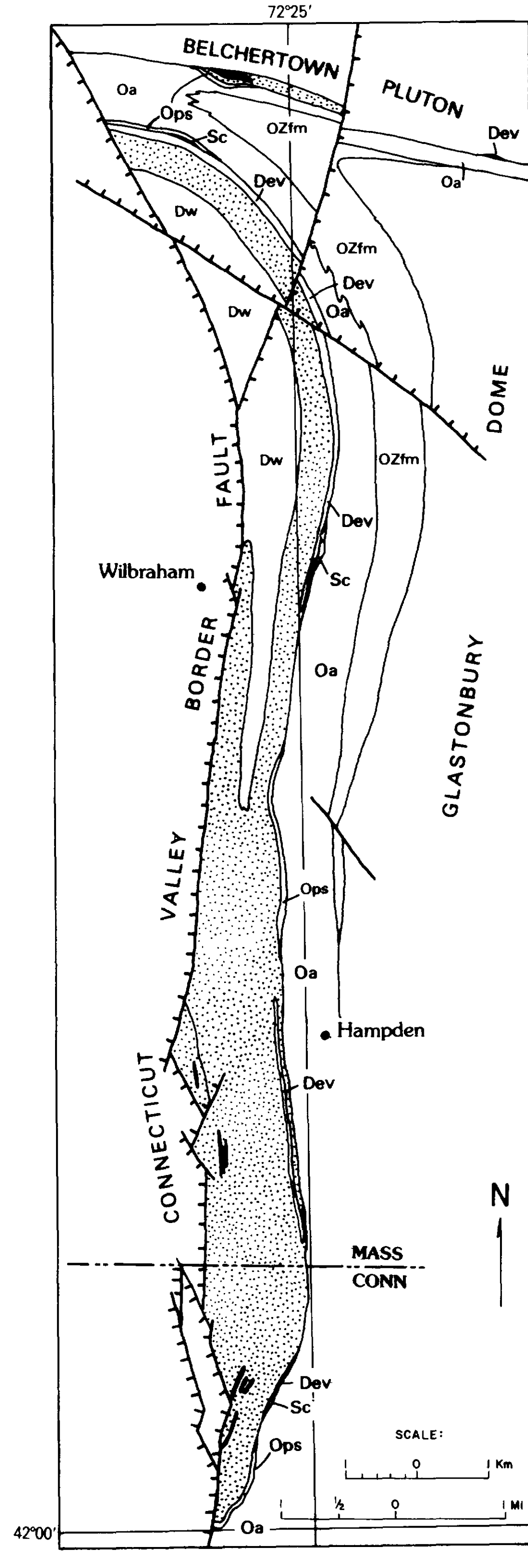


FIGURE 6.—Geologic map of the Wilbraham area showing the distribution of the Erving Formation amphibolite (black) and granulite (stippled). Other symbols are the same as on the State bedrock map: OZfm, Fourmile Gneiss; Oa, Ammonoosuc Volcanics; Ops, Partridge Formation; Sc, Clough Quartzite; Dev, undifferentiated Erving Formation schist and amphibolite; Dw, Waits River Formation.

approximately 500 m can be estimated for the formation. The overlying unit is mapped as Waits River Formation and consists of punky brown-weathering calcite-quartz-mica granulite in beds as much as 1 m thick interbedded with gray- to brown-weathering garnet-mica schist. Although these rocks generally resemble the Waits River Formation as mapped in medium- to high-grade metamorphic zones west of the Connecticut Valley in Massachusetts (Hatch and others, Ch. B, this volume), they are unlike the Waits River in having predominantly thinner carbonate beds and in containing a significant percentage of Erving-like granulites (Peper, 1977). They also do not differ greatly from mapped Gile Mountain Formation (Hatch and others, Ch. B, this volume). Exposures near the Erving-Waits River contact are few and generally poor, and both units are richer than normal in mica schist. Thus, although field relations indicate that rocks resembling the Waits River or possibly the Gile Mountain stratigraphically overlie the Erving near Wilbraham, the exact stratigraphic relation of those rocks to the section west of the Mesozoic basins is uncertain.

SUMMARY OF STRATIGRAPHIC RELATIONS OF THE ERVING FORMATION IN THE BRONSON HILL ANTICLINORIUM

In the Bronson Hill anticlinorium of Massachusetts, the Erving Formation is found above a sharp contact with units of a range of ages including the Proterozoic Z to Ordovician Fourmile Gneiss of the Pelham dome, the Middle Ordovician Ammonoosuc Volcanics and Partridge Formation, the Silurian Clough Quartzite, and the Lower Devonian Littleton Formation. These relations suggest an unconformity at the base of the Erving Formation, although some of the unconformable relations can be related to older unconformities. Of particular importance is the gradual and systematic *westward* pinching out of the Littleton Formation beneath the Erving Formation from a thickness of 675 m on the east limb of the Northfield syncline. The fact that the Erving does not appear overlying the Littleton in synclines east of the Northfield syncline suggests the Littleton may be much thicker in that direction. In the Wilbraham area, a thick section of Erving is overlain with apparent conformity by the Waits River Formation.

STRATIGRAPHIC RELATIONS IN THE WHATELY ANTICLINE

Modern stratigraphic exploration of the Whately anticline (fig. 2) was begun in 1966 by Walter E. Trzcienski (Ecole Polytechnique, Montreal) and Peter Robinson (Trzcienski, unpub. map, 1968), and work on the poorly exposed southern extension was completed by S.F. Clark, Jr. (unpub. map, 1977). The stratigraphic relations in the area were discussed on a number of informal field conferences involving the present authors, J.L. Rosenfeld, J.B. Thompson, Jr., J.C. Hepburn, and Leo M. Hall.

The Whately anticline consists of two elongate inliers, both cored by pre-Silurian rocks and ringed by Erving Formation (fig. 1). The northern inlier is herein referred to as the Whately area (fig. 7) and the southern as the Leeds area (fig. 8).

The oldest unit exposed in the Whately area is sulfidic mica schist and amphibolite, commonly containing garnet, assigned to the Middle Ordovician Partridge Formation. These rocks are well exposed in one large inlier and one small inlier. In the Leeds area, the oldest rocks are very poorly exposed owing to widespread pegmatites; they are spangly mica schist and have been hesitantly assigned to the Middle Ordovician Cobble Mountain Formation. In the northern part of the Whately area (fig. 7), the Partridge is overlain by 3-5 m of glassy white quartzite and quartz-pebble conglomerate of the Clough Quartzite. Elsewhere the Partridge Formation is generally overlain by well-laminated epidote amphibolite identical to that of the Erving Formation previously described. In the Leeds area, the Cobble Mountain is locally overlain by biotite-feldspar granulite assigned to the Erving, whereas in the Whately area, granulite of the Erving Formation occurs mainly at the top of the formation. The Erving granulite is predominantly a fine-grained, well-layered quartz-biotite-oligoclase rock with a few beds of gray mica schist and rare garnet schist. Small lens-like concordant quartz veins are characteristic in the few large outcrops northwest and southeast of the main inlier of Partridge in the Whately area. Several occurrences of coticule have been found in the granulite, close to amphibolite contacts, particularly in those areas of granulite enclosed in amphibolite in the Whately area (fig. 7) and near the southernmost contact of amphibolite in the Leeds area (fig. 8). Thus the sequence of rock types and overall contact relations between the Partridge, Clough, and Erving in the Whately area closely resemble those in the Leverett and Quabbin Hill areas, which are the nearest parts of the Bronson Hill anticlinorium to the east. Indeed we believe it is reasonable to infer that there is subsurface continuity between the rocks in the core of the Whately anticline and those on the west limbs of the Pelham and Glastonbury domes, particularly when the probable subsurface shapes of the Northampton and Belchertown areas of Belchertown tonalite are taken into account (State bedrock map cross section D-D').

Overlying the Erving Formation in the Whately area is the Gile Mountain Formation characterized by

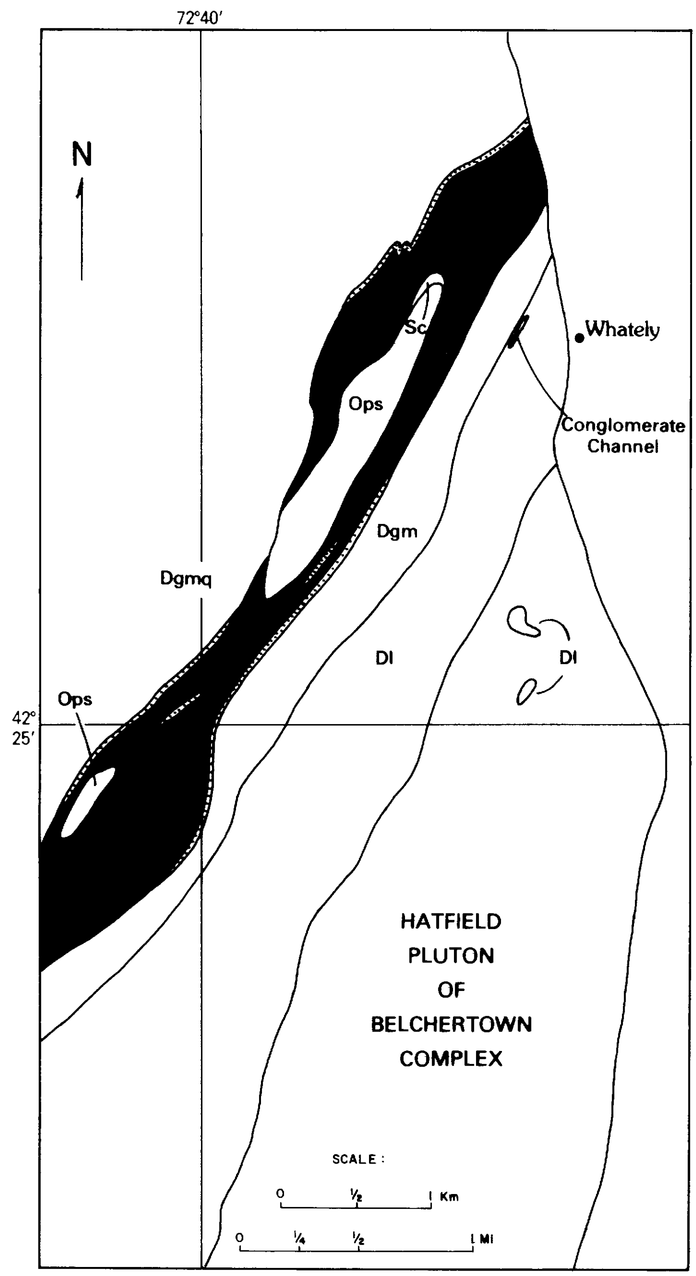


FIGURE 7.—Geologic map of the Whately area showing distribution of the Erving Formation amphibolite (black) and granulite (stippled). Other symbols are the same as on the State bedrock map: Ops, Partridge Formation; Sc, Clough Quartzite; Dgmq, quartzite-rich member of Gile Mountain Formation; Dgm, Gile Mountain Formation; Dl, Littleton Formation. For detailed map of conglomerate channel, see figure 9.

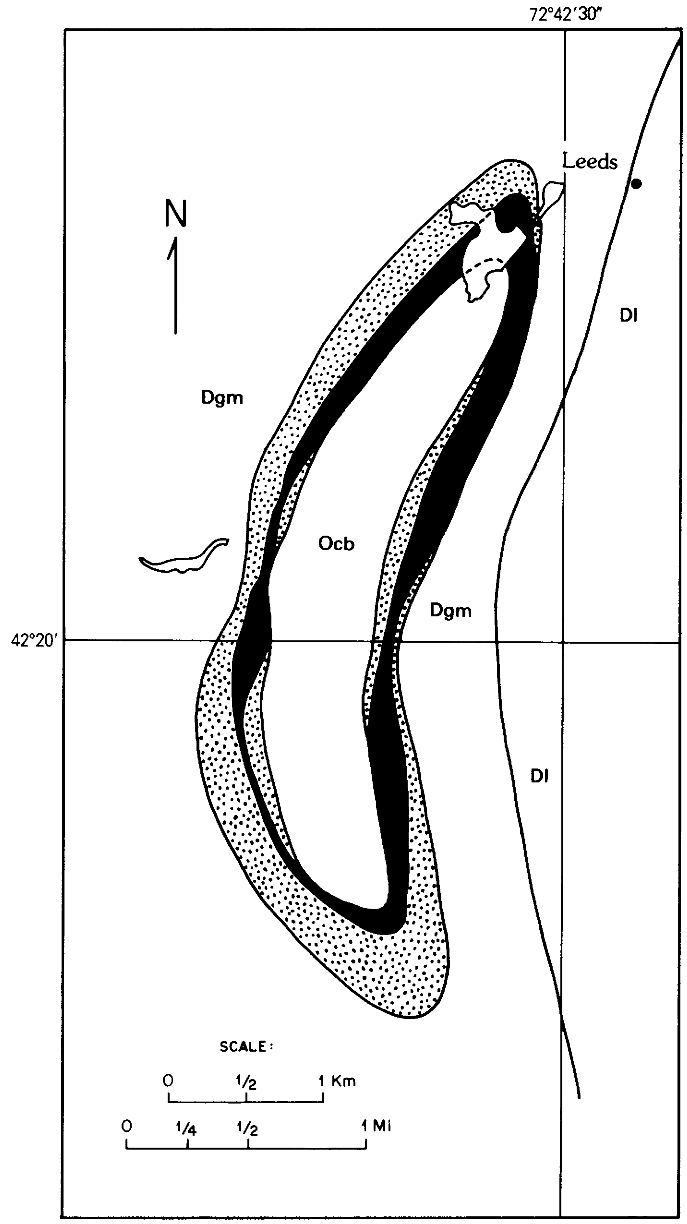


FIGURE 8.—Geologic map of the Leeds area showing distribution of the Erving Formation amphibolite (black) and granulite (stippled). Other symbols are the same as on the State bedrock map: Ocb, member B of the Cobble Mountain Formation; Dgm, Gile Mountain Formation; Dl, Littleton Formation.

well-layered gray phyllites, feldspathic micaceous quartzites, and numerous layers and lenses of punky-weathering calcite-quartz-mica granulite, and local rosettes of actinolite. These rocks are assigned to the Gile Mountain Formation on the basis of abundant quartzites (in contrast to the rocks tentatively assigned to the Waits River Formation at Wilbraham where such quartzites are scarce) and because they have been mapped continuously to the north into the Gile Mountain Formation of Vermont. The base of the Gile Mountain Formation is not well exposed and might be considered conformable, except in the western part of the Whately area where the base of the formation appears to cut through the Erving Formation and is locally in contact with the Partridge Formation.

The top of the Gile Mountain Formation is not exposed on the west limb of the Whately anticline, but on the east limb the formation appears to be overlain with sharp but subtle contact by rather homogeneous thin-bedded carbonaceous quartz schist. All who have studied these outcrops are in firm agreement that this quartz schist is typical of the Littleton Formation and would not normally be assigned to the Gile Mountain Formation.

Within the Littleton Formation, 3-5 m east of its contact with the Gile Mountain, is a spectacular graded channel deposit (figs. 7, 9) that contains a basal quartz-pebble conglomerate grading upward through quartzose grit to shaly sandstone to micaceous phyllite. The conglomerate, sandstone, and grit terminate to the south at what appears to have been the margin of a channel. The basal contact of the conglomerate and grit against the underlying black phyllite shows examples of distorted load casts (fig. 10). Although no evidence for transport direction has yet been found, the evidence for tops in the lowest part of this belt of Littleton Formation is clear, in fact probably as good as any such evidence that has been found in the Connecticut Valley synclinorium. Here the beds face to the east.

In summary, then, the most obvious stratigraphic interpretation of the Whately anticline is that the Clough Quartzite overlies the Partridge Formation, the Erving Formation overlies the Clough, the Gile Mountain overlies the Erving, and the Littleton overlies the Gile Mountain.

THE CHICKEN YARD LINE

The Chicken Yard line is a contact that enters Massachusetts from Vermont (fig. 1) in the north-central part of figure 2. It follows an intricately folded and in part imprecisely mapped course in northern Massachusetts until it disappears beneath Mesozoic cover at the northwest edge of the Deerfield basin.

The Chicken Yard line is named for exposures in a once flourishing and now overgrown chicken yard on the west side of U.S. Route 5 in Dummerston, Vt. The line is essentially the contact between homogeneous gray to black slates and phyllites of the Littleton Formation, to the east, and the somewhat more varied gray slates, phyllites, and granulites containing subordinate layers and lenses of calcite-quartz granulite that characterize the Gile Mountain Formation to the west (Thompson and others, 1968; Hepburn, 1972; Trask, 1964, 1980; Thompson and Rosenfeld, 1979). Lenses of quartz-pebble conglomerate with a shaly matrix are present locally near the western edge of the Littleton Formation in Vermont. Although the pebbles are similar to those in the conglomerate at Whately, the matrix is generally much less sorted, possibly suggesting an origin by subaqueous slumping. Adjacent quartzose phyllites with graded and cross bedding suggest that such slumping, if any, was very limited. Locally near or at the eastern edge of the Gile Mountain Formation, lenses or a continuous layer of fine-grained, poorly foliated, light-greenish-gray phyllite or granulite with quartz and feldspar phenocrysts constitutes the Putney Volcanics (Trask, 1964, 1980; Thompson and others, 1968; Hepburn, 1972).

The Chicken Yard line contact is well exposed at only a few places. At two of these places near Springfield, Vt. (fig. 1), the contact separates phyllite and granulite of the Putney Volcanics to the west from dark-gray slate or phyllite of the Littleton Formation to the east. In 1978, J.B. Thompson, Jr., convinced us from primary sedimentary structural features at a number of localities at or near the contact that bedding in the Gile Mountain and the Littleton tops to the east across their mutual contact. These relations, like those at Whately, imply that the Littleton lies stratigraphically above the Gile Mountain. We are all reasonably convinced that the Chicken Yard line extends beneath the Mesozoic cover and connects with the western contact of the Littleton at Whately (queried contact, fig. 2).

One more observation may be relevant to this discussion. The Littleton Formation at Slate Ledge, about 2 miles west of the village of Littleton, N.H. (Billings, 1937), is virtually indistinguishable from the Meetinghouse Slate about 10 miles to the northwest in Waterford, Vt. (Eric and Dennis, 1958). Thanks to tight isoclinal folding near the contact, the topping sense across the Meetinghouse-Gile Mountain contact cannot be ascertained with confidence.

In addition to the main belt of Gile Mountain Formation west of the Chicken Yard line in Massachusetts and southern Vermont, three areas of chlorite-grade schist with punky-weathering calcite-quartz granulite that occur in fault-bounded slices along the Connecticut Valley border fault in the Leverett area (figs. 2, 4) have been assigned to the Gile Mountain. Only one of these areas, the northernmost, on the west bank of the Connecticut River north of French King Bridge, has significant outcrop. The other two areas are assigned mainly on the basis of diamond drill core obtained by Northeast Utilities in connection with the proposed Montague Nuclear Plant (Northeast Utilities Preliminary Safety Analysis Report, section 2.5, Geology and Seismology). Their assignment to the Gile Mountain Formation is thus somewhat tenuous. As interpreted by Robinson on the basis of field work and personal study of the cores, these areas belong to fault slices from intermediate structural levels, structurally higher than the kyanite-grade footwall rocks east of the

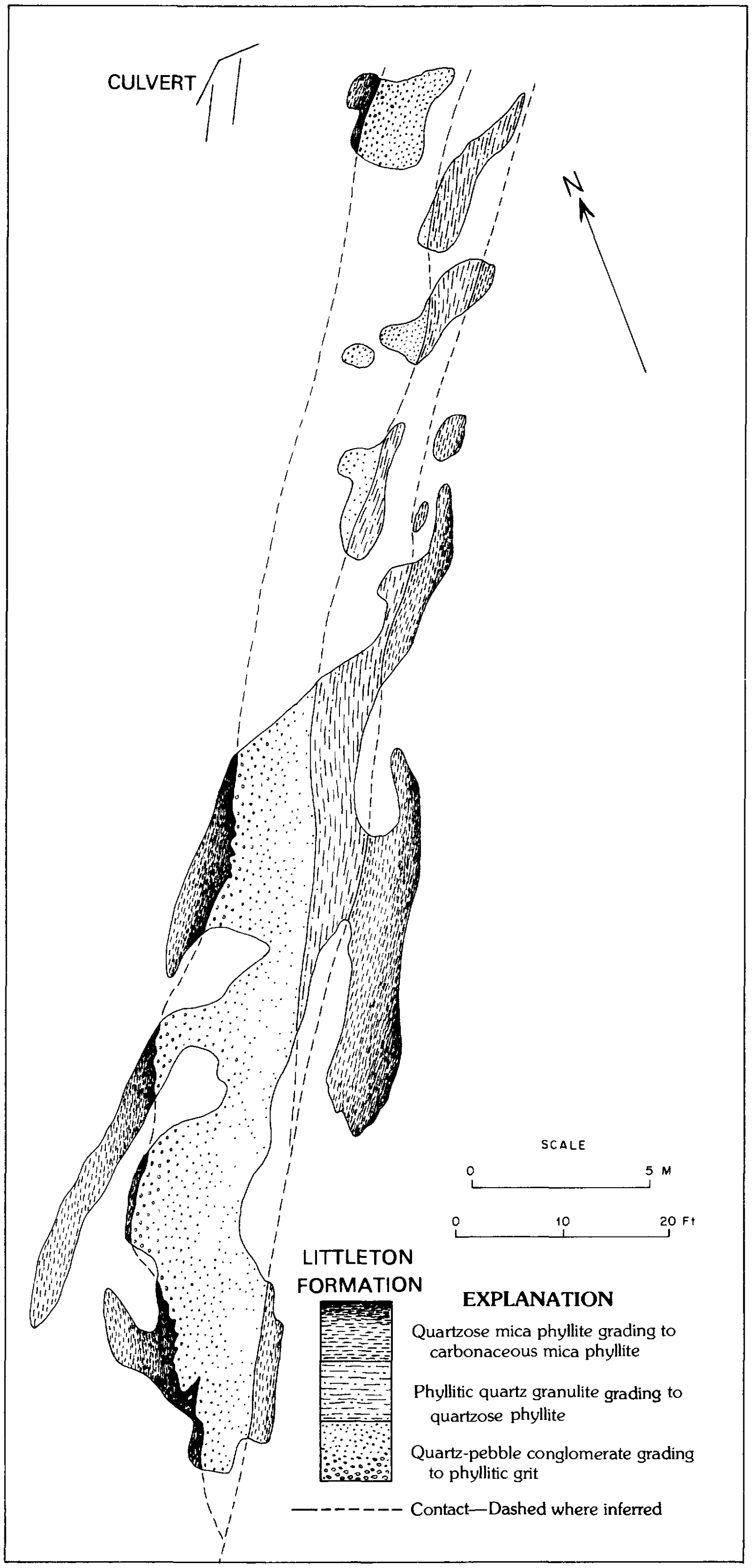


FIGURE 9.—Detailed geologic map of the channel near the base of the Littleton Formation at Whately, Mass. Outcrop area is on the south side of Haydenville Road, near a culvert, about 530 m west of the junction with Chestnut Plain Road at Whately.

FIGURE 10.—Area near south end of outcrop in figure 9 showing base of conglomerate-grit bed (lighter colored rock) with load cast features. Top of photograph is approximately east. Field of view is about 0.4 m wide.

fault zone and structurally lower than the sillimanite-grade hanging-wall rocks west of the fault zone. In fact, these data are compatible with the previously proposed concept (Thompson and others, 1968) of a giant metamorphic overhang along the east edge of the Connecticut Valley synclinorium. If the slivers are correctly assigned to the Gile Mountain, they further suggest that low-grade Gile Mountain extends east from the Chicken Yard line beneath the present surface to the Connecticut Valley border fault and probably once extended still further. Diamond drill core from the northernmost fault slice also contains light-green granulite that might reasonably be assigned to the Putney Volcanics.

WINDMILL MOUNTAIN, VERMONT

At the suggestion and with the guidance of J.B. Thompson, Jr., we have examined part of the section on the east flank of the Athens dome at Windmill Mountain, Vt. (fig. 1). Here sulfidic schist and amphibolite shown as Middle Ordovician Cram Hill Formation (carbonaceous schist and quartzite of the Missisquoi Formation of Doll and others, 1961) are overlain by a few meters of quartz-pebble conglomerate and glassy quartzite locally assigned to the Silurian Shaw Mountain Formation. East of and above the quartzite are approximately 100 m of well-layered epidote amphibolite followed by about 3 m of well-bedded gray quartz-feldspar-mica granulite and schist with a few 1-cm-thick beds of pale-pink coticule. These rocks have also been assigned to the Shaw Mountain Formation, partly on the basis of one outcrop on Windmill Mountain, where a 0.5- to 1-m-thick bed of the conglomerate is about 1 m above the base of the amphibolite. It is possible, however, that this interlayering is due to isoclinal folding or faulting, rather than to primary interstratification, and thus that the amphibolite could be significantly younger (Devonian?) than the conglomerate. The coticule-bearing granulite is overlain to the east by dark-gray garnet-mica schists with rare calcite granulite beds assigned to the Northfield Formation that shortly grade eastward into typical Waits River Formation. Thompson (oral commun., 1983) suggested, and we agree, that the sequence of rocks beneath the Partridge-Northfield Formation at Windmill Mountain bears a striking resemblance to the Clough-Erving sequence of rocks beneath the Gile Mountain Formation at Whately. If the amphibolite and coticule-bearing granulite at Windmill Mountain are the same as those at Whately, and if they are Silurian, as those at Windmill Mountain are believed to be (Boucot and Thompson, 1963), then the implication is that the Erving Formation is Silurian. This is, of course, a distinct enigma with respect to relations in the Northfield, Mass., area where the Erving overlies the Lower Devonian Littleton Formation. The reverse implication, that the upper part of the Shaw Mountain Formation is Devonian, is equally enigmatic for relations in Vermont because the formation contains Silurian fossils near Albany, Vt. (Boucot and Thompson, 1963).

A STRATIGRAPHIC DILEMMA

On straightforward stratigraphic grounds the areas described above give the sequences of units in table 1. The list alone makes it emphatically clear that either the stratigraphy or the structure is more complex than we have imagined. All stratigraphic solutions to the dilemma involve the assumption that one or more of the named rock types actually appears twice in the stratigraphy. All structural solutions involve the assumption that some units are repeated by a thrust fault.

STRATIGRAPHIC SOLUTIONS

An obvious solution suggested by table 1 would assume that there are actually two Littleton Formations, one above the Clough Quartzite and below the Erving Formation (table 1, col. 1), the other above the Gile Mountain Formation (cols. 5 and 6). There is, however, compelling, if not absolutely incontrovertible, evidence that this cannot be so. The argument is as follows.

From a three-dimensional reconstruction of pre-Mesozoic structural relationships across the Connecticut Valley border fault (fig. 2) near the New Hampshire-Vermont-Massachusetts border (Peter Robinson, Richard Dana, and Farrukh Ahmad, "unpublished" Lucite model, 1975) (fig. 11), it can be argued that the Vernon dome in the hanging wall of the fault is actually "rooted" in, and a northern lobe of, the Warwick dome in the footwall. In the stratigraphic sequence on the west limb of the Warwick dome (fig. 3), the Ammonoosuc Volcanics or Partridge Formation is overlain by Clough and then by a sequence of Littleton as much as 750 m thick, which thickens northward. On the Vernon dome (fig. 2), the stratigraphic sequence is identical except that the Littleton is apparently even thicker. The Littleton on the west limb of the Warwick dome is the Littleton that *underlies* the Erving Formation in the Northfield and Wendell synclines. As can be seen in figure 3, this Littleton thins southward and pinches out between Partridge or Clough and Erving in the Wendell syncline and on the west limb of the Northfield syncline. The Littleton on the Vernon dome *is* the Littleton that is in contact with the Gile Mountain along the Chicken Yard line, where it is supposed to *overlie* the Gile Mountain. Yet the reconstructed three-dimensional model of the Warwick and Vernon domes indicates that the two thick Littleton sequences on the west limb of the Warwick dome and the west limb of the Vernon dome must be one and the same, and they cannot be assigned to two different stratigraphic levels. Thus, unless rocks assigned to the Littleton at Whately are different from those assigned to the Littleton east of the Chicken Yard line, the two-Littleton stratigraphic solution must be ruled out.

Other two-unit solutions are still less satisfactory. Suppose, for example, that the Erving that overlies the Littleton in the Northfield syncline (table 1, col. 1; fig. 3) is not the same as the Erving that underlies Gile Mountain in the Whately anticline (table 1, col. 5; figs. 7, 8) or that underlies Waits River in the Wilbraham area (table 1, col. 4; fig. 6). If we require that the Erving of the Northfield syncline and the Erving of the Whately anticline be stratigraphically different, we are faced with the problem of a stratigraphic assignment of the Leverett-area Erving, which resembles *both* of them. To which "Erving" should we assign the Quabbin Hill rocks? They appear to have been continuous with the Wilbraham-area Erving before intrusion of the Belchertown Complex, but they also contain basal calcareous rocks, including the unique calc-silicate layer with white euhedral zoisite, like those in the Northfield-syncline Erving.

In this discussion we have assumed that the Gile Mountain and Waits River Formations were both deposited in the same general stratigraphic package of rocks. Although this relationship is compatible with the "two-Littleton solution," it cannot be true for the "two-Erving solution." In other words, if one takes the stand that the Whately and Wilbraham Erving do not correlate, then it is necessary to assume also that there is no general correlation between the Whately Gile Mountain and the Wilbraham Waits River (table 1).

STRUCTURAL SOLUTIONS

Structural solutions to the stratigraphic dilemma involve repetitions of stratigraphic units by major thrust faults within a setting of relatively complex facies relationships. Because the major stratigraphic dilemma involves the conflict in position of the Littleton Formation in the Northfield area compared with its position in the Whately anticline and Chicken Yard line areas, obvious structural solutions would place a thrust fault in one or the other of these two areas. Each of the two possibilities must then be evaluated in terms of its stratigraphic implications and geometrical reasonableness; that is, can one draw reasonable cross sections through the region and then remove the structure to obtain rational pretectonic facies relations? We have not evaluated more complex solutions involving more than one thrust fault because they would lead to wholesale abandonment of the regional stratigraphy that we think is basically sound.

The first alternative would be to consider an early or premetamorphic thrust fault in the Northfield area in which a sheet of Erving Formation and overlying units is thrust *above* the Littleton Formation. We have already shown that the Littleton in the Northfield area (fig. 3) pinches out westward beneath the Erving, therefore it would be natural to believe that a thrust at the base of the Erving would involve overthrusting from the west toward the east. Such a direction is not consistent with the transport direction of early Acadian nappes in the Connecticut Valley (Thompson and others, 1968), which are overfolded from east to west, but is consistent with the transport direction of a set of early Acadian recumbent folds recognized in the immediate vicinity of the Pelham dome. However, if such early thrusting did occur in the Northfield area, it would have to predate this recumbent folding, which clearly folds the Erving-Littleton contact surface.

The stratigraphic implications of a structural solution in which the Erving Formation is thrust over the Littleton Formation in the Northfield area are as follows. This model permits the Erving to be strati-

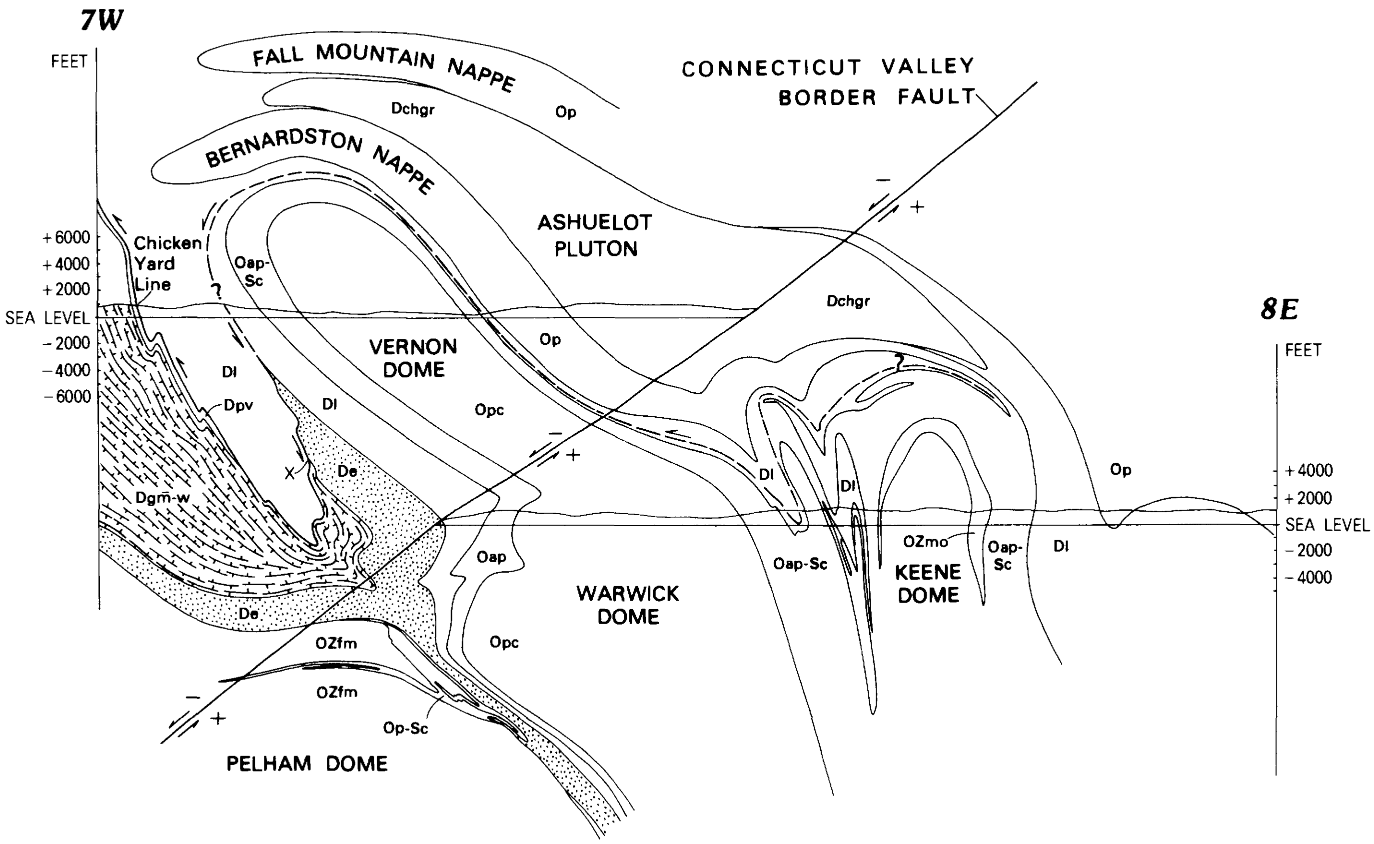


FIGURE 11.—Restored cross section in southern New Hampshire and Vermont showing pre-Mesozoic structural relations between the Warwick dome, Vernon dome, and Chicken Yard line, and inferred configuration of the Whately thrust shown by queried dashed line. Section is modified from parts of a three-dimensional Lucite model prepared by Peter Robinson, Richard Dana, and Farrukh Ahmad in 1975, which was based on cross sections prepared by Ahmad (1975). The west end of section 7W is juxtaposed against the east end of section 8E (see fig. 2) by restoration of fault movement along a net slip line trending N. 47° W. 38° W. Thus, in the Mesozoic fault movement, 7W moved down and away (-), and 8E moved up and toward (+) the viewer. Letter symbols for rock units are identical to the State bedrock map except that Ordovician Ammonoosuc Volcanics and Partridge Formation and Silurian Clough Quartzite are amalgamated (Oap-Sc, Op-Sc). OZfm, Fourmile Gneiss; OZmo, Monson Gneiss; Opc, Pauchaug Gneiss; Oap, Ammonoosuc Volcanics and Partridge Formation combined; Op, Partridge Formation; Dl, Littleton Formation; De, Erving Formation; Dgm-w, Gile Mountain and Waits River Formations; Dpv, Putney Volcanics; Dchgr, Coys Hill Porphyritic Granite Gneiss. The X in the west part of the diagram marks the point of triple join between the Littleton, Gile Mountain, and Erving Formations. (Use of trade names is for identification purposes only and does not imply endorsement by the U.S. Geological Survey.)

graphically beneath the Littleton and to correlate with the Silurian Shaw Mountain Formation of Windmill Mountain and other areas of Vermont. It also permits the Whately section to be a continuous one with Littleton at the top and permits the thick Gile Mountain-Waits River part of the section to be Silurian or lowermost Devonian. Because these units or rocks of equivalent age are absent to the east in the Bronson Hill anticlinorium, except for thin and discontinuous Fitch Formation, this interpretation implies that Erving, Gile Mountain, and Waits River were deposited in a Late Silurian to earliest Devonian sedimentary trough located west of the axis of the present anticlinorium, as suggested by Thompson and Rosenfeld (1979).

Geometrical difficulties with this thrust interpretation are severe. In the Northfield area the base of the Erving, the presumed thrust of early or premetamorphic age, cuts across not only the Littleton but also the Clough Quartzite, the Partridge Formation, *and* the Fourmile Gneiss. The distribution of rock types beneath the presumed thrust contact is erratic and does not show the consistent westward downcutting to be expected from an originally west-dipping thrust. On the west limb of the Pelham dome in the Leverett area (fig. 4), where one might expect the thrust to have cut even deeper, the Clough is preserved as well as several hundred meters of Partridge.

A further geometrical difficulty arises in the Vernon dome area (fig. 2). On the west limb of the dome itself the stratigraphy tops west, whereas along the Chicken Yard line, a few kilometers to the west, tops are east, implying a tight syncline between. If, as suggested in this thrust solution, the Chicken Yard line is a stratigraphic contact of Littleton over Gile Mountain in the *upper part* of an eastward directed thrust sheet, then that contact as well as the *lower part* of the thrust sheet should appear again to the east on the west limb of the Vernon dome. It does not, and we have been unable to prepare a consistent set of cross sections based on this structural solution.

The second structural solution to the stratigraphic dilemma, and the one we prefer, is a converse of the first. In this solution, the contact of the Erving above the Littleton Formation in the Northfield and Wendell synclines is considered to be a stratigraphic contact, probably an unconformity, whereas the base of the Littleton at Whately and along the Chicken Yard line is considered to be a thrust fault, the Whately thrust, here directed from east to west.

The stratigraphic implications of this solution are as follows. At the time the Littleton was being deposited on the present Bronson Hill anticlinorium and further east, there was nondeposition or even slight erosion to the west. Deposition subsequently expanded westward to deposit the Erving, Gile Mountain, and Waits River. In this interpretation, the quartzite on Windmill Mountain would be Silurian, whereas the overlying amphibolite and thin granulite would be Devonian and correlative with the Erving. Correlatives of the Erving, Gile Mountain, and Waits River may also have been deposited above the Littleton of the Bronson Hill anticlinorium but have since been eroded away. The Whately thrust, then, would involve thrusting of a thick eastern section of Littleton westward onto an Erving-Gile Mountain-Waits River terrane where no Littleton had previously been deposited.

There are some obvious objections to this solution. For example, there are no obvious signs of thrusting along the Littleton-Gile Mountain contact in Whately or on the Chicken Yard line, although the contact is sharp. This, of course, could be because the thrusting was pre-metamorphic or early in the metamorphism, and the rocks have been severely deformed in later events. Another objection is that the contact seems to be the locus of conglomerates and other primary sedimentary features in the Littleton Formation, as well as granulite and phyllite of the Putney Volcanics. One could well question why these features should be found along a thrust fault, although any variation from stratigraphic monotony might provide a mechanical discontinuity that could be extensively used during thrusting.

In defense of the concept of the Whately thrust, we have succeeded in making reasonable cross sections (fig. 12), a three-dimensional model (fig. 13), and a palinspastic reconstruction (fig. 14). Although we have not found any place where the Littleton Formation can be seen both stratigraphically below the Erving Formation and thrust above it, as this solution requires, we can predict the location of such places, which have mostly been removed by Mesozoic or younger erosion. Similarly, although we have not seen a place where a thrust below the Littleton truncates the Erving-Gile Mountain contact (creating a line of triple junction, point X in figs. 11 and 14), we can make a reasonable prediction of the location of this line of intersection on the basis of relations in the Northfield and Leverett areas including the three slices of Gile Mountain Formation along the Connecticut Valley border fault. We conclude that this line would trend north-northwest above the present surface exposures of the Pelham dome between the Northfield and Leverett areas. The line would be truncated by the Connecticut Valley fault somewhere north of Millers Falls and south of Northfield. On the downthrown side of the fault, this line would be entirely in the subsurface and would run somewhere along the west side of the Vernon dome. The syncline between the Vernon dome and the Chicken Yard line (see above) would repeat the Whately thrust surface (figs. 11, 13). However, because the thrust cuts downward stratigraphically to the east, on the west limb of the Vernon dome the thrust would be a contact between thrust and autochthonous sections of the Littleton Formation. Similarly, elsewhere in the Bronson Hill anticlinorium, the Whately thrust could totally lose identity as a bedding plane thrust within a relatively thick section of homogeneous Littleton rocks.

Thus, of all the stratigraphic and structural solutions to the stratigraphic dilemma of the Erving Formation, we find the concept of a Whately thrust the one that most successfully answers the various stratigraphic and geometrical objections, although we recognize its shortcomings. It is the solution followed on the State bedrock map. We hope that eventually new paleontological data, particularly in the low-grade rocks of the Connecticut Valley near the Chicken Yard line, will settle the question.

REGIONAL IMPLICATIONS OF A WHATELY THRUST

STRATIGRAPHIC

As we pointed out above, the two structural solutions to the stratigraphic dilemma yield two totally different

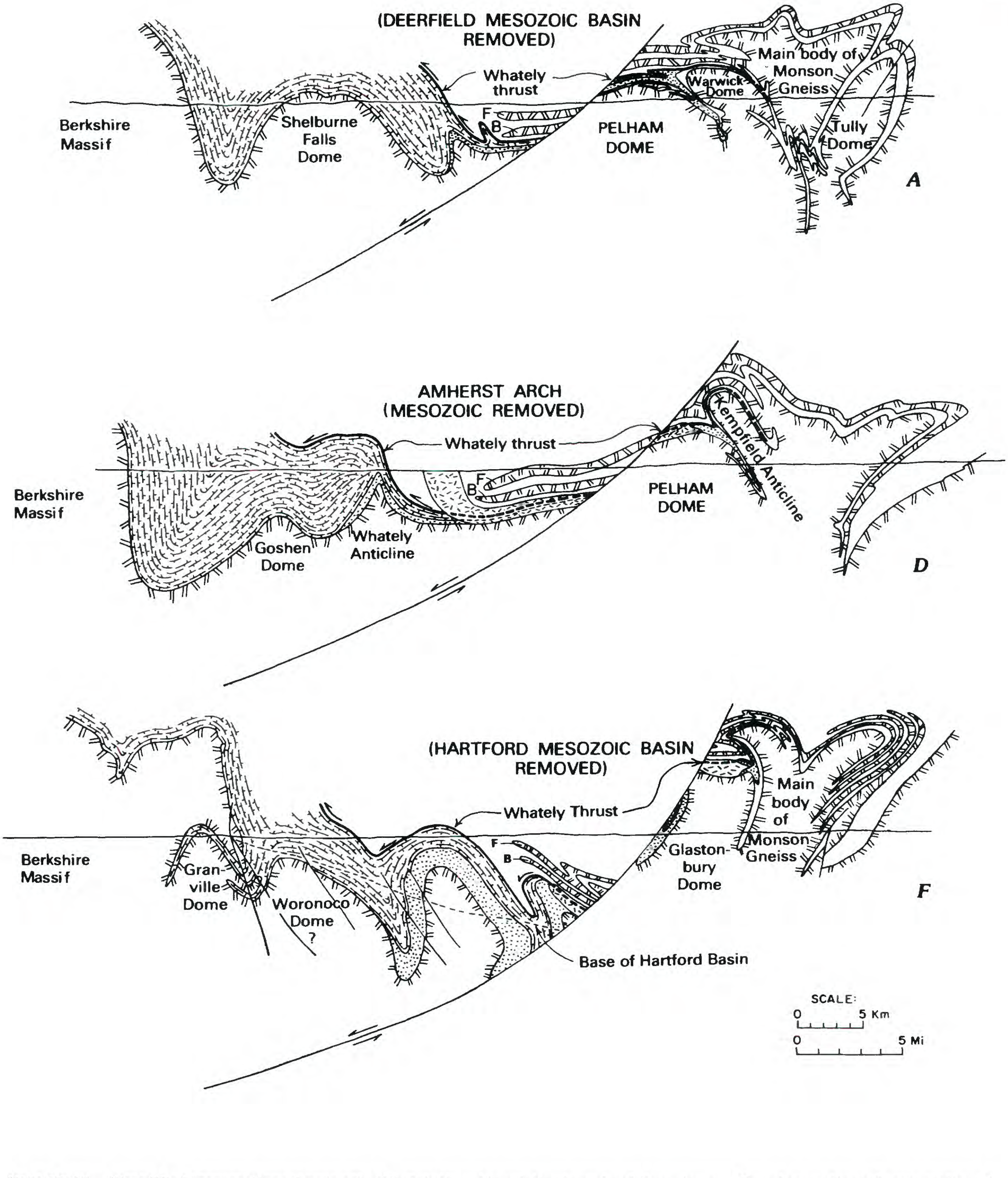


FIGURE 12.—Simplified cross sections showing the Whately thrust in relation to rocks of Devonian age. Based on portions of cross sections A, D, and F from the State bedrock map (see fig. 2). B, Bernardston Nappe, F, Fall Mountain Nappe. Hachures, Silurian and older rocks; unpatterned, Littleton Formation; stippled, Erving Formation; brick-like pattern, Gile Mountain, Waits River, and Goshen Formations and Putney Volcanics undifferentiated; dashed pattern, Belchertown-Hatfield pluton.

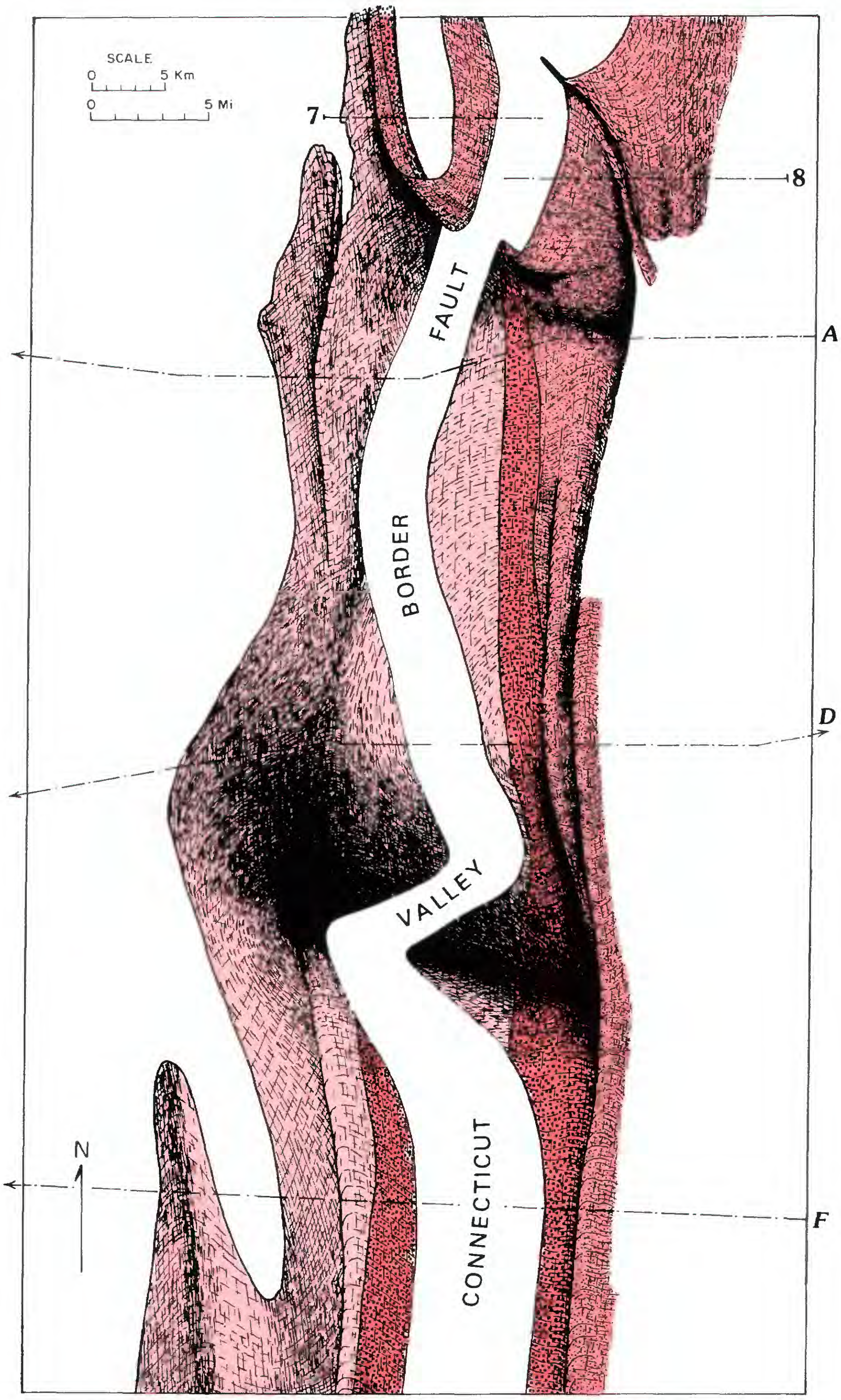


FIGURE 13.—Structural relief diagram showing the inferred geometrical disposition of the surface of the Whately thrust on both sides of the Connecticut Valley border fault. For geographic references see figure 2. *West* of the border fault the Whately thrust surface is shown only *below* the present erosion surface and is not restored above. *East* of the border fault the Whately thrust surface is fully restored, as it existed mostly above the present erosion surface, but is not extended to the east side of the Bronson Hill anticlinorium. The thrust surface consists of three parts, an east part (medium red, finely stippled) where Littleton Formation is thrust on Littleton Formation, a central part (dark red, coarsely stippled) where Littleton Formation is thrust on Erving Formation, and a western part (light red, not stippled) where Littleton Formation is thrust on Waits River and Gile Mountain Formations or Putney Volcanics.

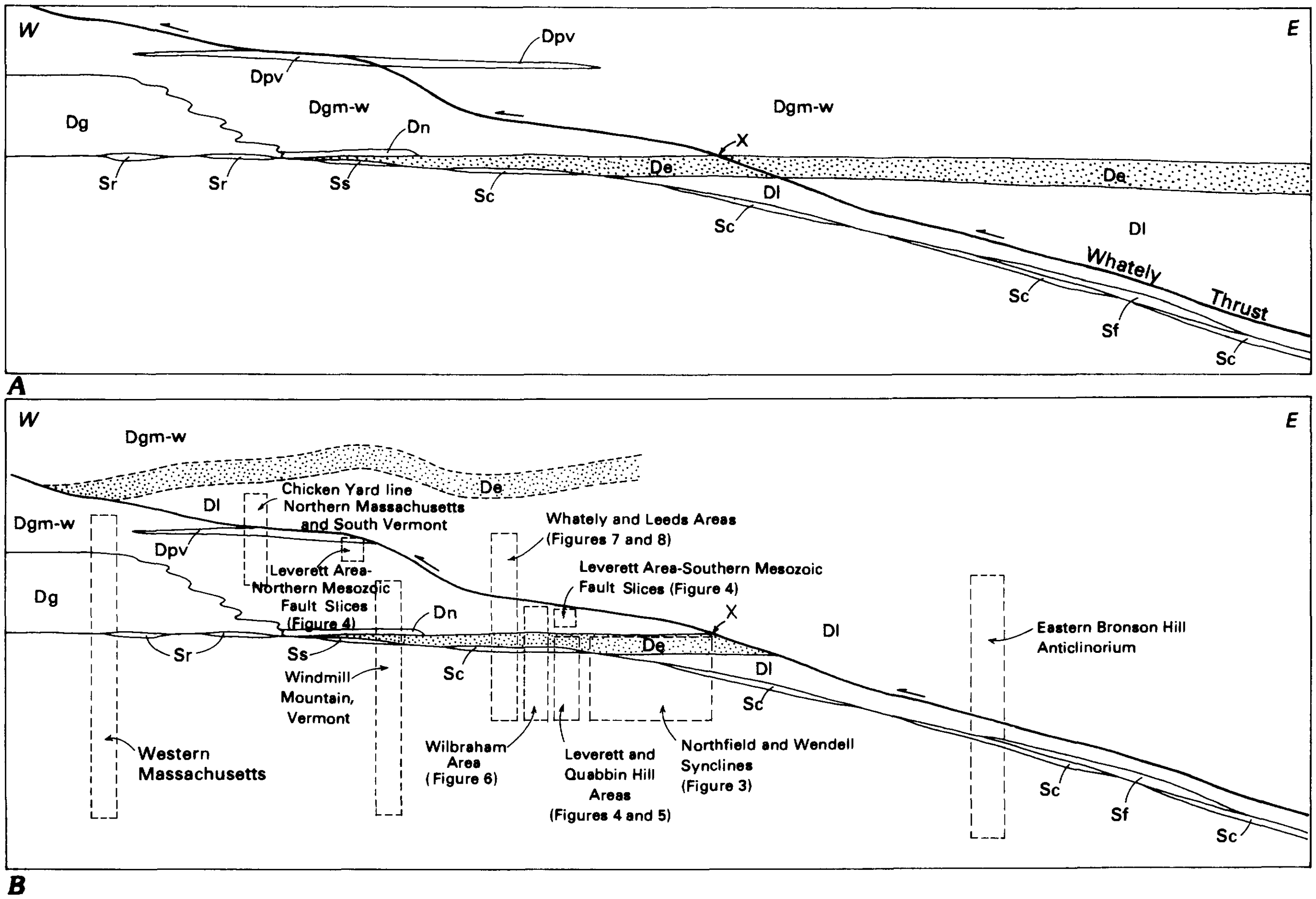


FIGURE 14.—Palinspastic reconstructions showing inferred relations between stratigraphy and the Whately thrust. Horizontal and vertical scales show relative positions only. X marks triple contact between the Littleton, Erving, and Gile Mountain Formations. *A*, Inferred relations between stratigraphy and the position of the Whately thrust before thrusting. *B*, Inferred relations between stratigraphy and the position of the Whately thrust after thrusting but with all other deformations removed. Inferred structural-stratigraphic positions of various field areas discussed in the text are indicated schematically. Stratigraphic units are as follows (pre-Silurian not differentiated): Sr, Russell Mountain Formation; Ss, Shaw Mountain Formation (Vermont); Sc, Clough Quartzite; Sf, Fitch Formation; Dl, Littleton Formation; De, Erving Formation; Dn, Northfield Formation (Vermont); Dgm-w, Gile Mountain and Waits River Formations, undifferentiated; Dg, Goshen Formation; Dpv, Putney Volcanics.

paleogeographic interpretations. The Whately thrust solution implies a single Silurian-Devonian sedimentary trough for southern New England in which thick sedimentation began in the Silurian in the east (in the Merrimack synclinorium; Robinson, 1979), spread westward over the site of the Bronson Hill anticlinorium during Littleton time, and reached the site of the Connecticut Valley synclinorium and Berkshire anticlinorium in the Devonian. The other solution, which we find geometrically difficult, in which the Erving is thrust eastward over the Littleton, implies that there were two Silurian and earliest Devonian troughs separated by a narrow uplift along the Bronson Hill anticlinorium where there was thin or no sedimentation. These two older troughs and the anticlinorium would then have been buried by the more widespread deposits of the Littleton.

TECTONIC

The westward transport of at least 20 km implied for the Whately thrust by figure 12 is in approximately the same direction as the early Acadian west-verging isoclinal folds in the Goshen Formation and the basal Goshen decollement (Hatch and Stanley, Ch. C, this volume; State bedrock map cross section *D-D′*) in the Connecticut Valley belt west of the Whately anticline and the early Acadian nappes of the Bronson Hill anticlinorium (Thompson and others, 1968; Thompson and Rosenfeld, 1979). The early Acadian nappes are considered to be synmetamorphic, and they involve a standard "Bronson Hill" sequence of Partridge Formation, Clough Quartzite, local Fitch Formation, and Littleton Formation. The axial surfaces of these nappes lie east of and structurally above the surface of the Whately thrust as defined by the Chicken Yard line. Two alternative kinematic interpretations of the thrust and the nappes should be considered:

(1) The thrust formed first, by transporting very low grade or even unmetamorphosed Littleton over the section to the west. The nappes were then emplaced above and east of this tectonically thickened section during metamorphism. The nappes may even have folded the previously formed thrust surface.

(2) The nappes formed first and were emplaced at the top of a Littleton section. This entire assemblage was then transported westward as a package.

On the whole, relations between structure and metamorphism slightly favor the first alternative. This model bears some resemblance to the model for the Taconian orogeny presented by Stanley and Ratcliffe (1985) and Stanley and Hatch (Ch. A, this volume). In this light, the Whately thrust is the Acadian analog of the Giddings Brook thrust of the Taconian.

Northward in Vermont (fig. 1), the Chicken Yard line apparently traces into a point near the hinge of the Cornish nappe (Thompson and others, 1968), and the Gile Mountain-Waits River Formations are on the west (or lower) limb of the Cornish nappe. Clearly the understanding of the structural geometry of this region, which is beyond the scope of this paper, will be crucial to the final regional interpretation.

METAMORPHIC

The Connecticut Valley metamorphic low, consisting of biotite- and chlorite-grade rocks separating kyanite-grade rocks of the "Vermont high" to the west from kyanite- and sillimanite-grade rocks of the "New Hampshire high" to the east, has long been known and puzzled over (Thompson and Norton, 1968; Thompson and others, 1968). For the most part, this metamorphic low lies very close to the position of the Chicken Yard line or Whately thrust. Thompson and others (1968) showed that the western edge of the eastern metamorphic high is in many areas a metamorphic overhang related to emplacement of the nappes, which were already being metamorphosed as they were being emplaced and continued to be metamorphosed in subsequent stages. In this scenario the Whately thrust may provide a means of transporting a large volume of Littleton Formation westward out of the Bronson Hill zone, before that region attained medium metamorphic grade. The tectonic thickening of cool, water-bearing rocks in the Connecticut Valley synclinorium caused by the Whately thrust may have prevented mergence of the Vermont metamorphic high below and the New Hampshire metamorphic high in the nappes above. If this is true, then the Whately thrust could be a key to the metamorphic, as well as the stratigraphic and tectonic, history of western New England.

OTHER TENTATIVE CORRELATIONS OF THE ERVING FORMATION

In 1960, at the instigation of John L. Rosenfeld, he and Robinson made a detailed comparison of the Erving Formation in the Northfield area and the Gile Mountain Formation and Standing Pond Volcanics at the south end of the Guilford dome near Green River, Vt. The Standing Pond is a thin but remarkably persistent amphibolite in eastern Vermont that occurs at or very near the contact between the Waits River and Gile Mountain Formations. Its apparent southern extension into northern Massachusetts is shown on the State bedrock map by the symbols Dwa and Dgma. The results of this comparison were summarized by Robinson (1963, see especially table 40). The rock types are very similar, and the coticule beds in Gile Mountain granulite near amphibolite contacts are strikingly similar to those in the Erving Formation and different from those found elsewhere in the section. It is not our task here to explore the implications of such an additional correlation, but these data do support our contention that the Erving Formation belongs in the same stratigraphic package as the Gile Mountain and Waits River Formations.

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