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Summary Results of the Glens Falls CUSMAP Project, New York, Vermont, and New Hampshire

U.S. GEOLOGICAL SURVEY BULLETIN 1887

U.S. DEPARTMENT OF THE INTERIOR

March 3, 1849

Summary Results of the Glens Falls CUSMAP Project, New York, Vermont, and New Hampshire

Edited by JOHN F. SLACK

Geologic, geophysical, and geochemical studies related to
metallic mineral resources in west-central New England

U.S. GEOLOGICAL SURVEY BULLETIN 1887

U.S. DEPARTMENT OF THE INTERIOR
MANUEL LUJAN, Jr., Secretary

U.S. GEOLOGICAL SURVEY
Dallas L. Peck, Director

March 3, 1849

UNITED STATES GOVERNMENT PRINTING OFFICE: 1990

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Library of Congress Cataloging in Publication Data

Summary results of the Glens Falls CUSMAP project, New York, Vermont, and New Hampshire : geologic, geophysical, and geochemical studies related to metallic mineral resources in west-central New England / edited by John F. Slack.
p. cm. — (U.S. Geological Survey bulletin ; 1887)
Supt. of Docs. no.: I 19.3:1887
1. Mines and mineral resources—New York (State)—Glens Falls Region. 2. Mines and mineral resources—New England. 3. Geology—New York (State)—Glens Falls Region. 4. Geology—New England. I. Slack, John F. II. Conterminous United States Mineral Assessment Program. III. Series.
QE75.B9 no. 1887
[TN25.G58]
557.3 s—dc20
[553'.0974]
89-600124
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PREFACE

For more than a decade, the U.S. Geological Survey (USGS) has been studying the mineral potential of large parts of the United States under the auspices of the Conterminous United States Mineral Assessment Program (CUSMAP). One region evaluated by this program is the Glens Falls 1° × 2° quadrangle covering eastern New York, central Vermont, and western New Hampshire. The Glens Falls quadrangle (fig. 1) extends from lat 43°00′ to 44°00′ N. and from long 72°00′ to 74°00′ W. and has a contained land area of approximately 18,400 km^2. The geology of this region is highly varied and includes Middle Proterozoic basement rocks, Late Proterozoic to middle Paleozoic metasedimentary and metavolcanic rocks, and igneous intrusions of Proterozoic, Paleozoic, and Mesozoic age.

Mineral resources of the Glens Falls quadrangle were evaluated from 1982 through 1988 by a multidisciplinary approach involving geology, geochemistry, geophysics, remote sensing, geobotany, and topical studies of individual mineral deposits. The resource assessment presented here is limited to metallic resources, although nonmetallic commodities have been (and currently are) of much greater economic importance. Hydrocarbon resources such as oil and gas, coal, and peat similarly are not evaluated.

This report contains 18 chapters that summarize the results of the Glens Falls CUSMAP project. The initial data and interpretations from the project were first presented at a public meeting held in Killington, Vt., in September 1986. The meeting, which included a 3-day field trip, was attended by nearly 80 participants from industry, academia, and Federal and State government agencies. This USGS Bulletin constitutes the formal publication of the papers given at the Killington meeting and includes several others submitted since that time.

In the opening chapter (A), J.B. Thompson, Jr., presents an overview of the geology and plate tectonic history of the Glens Falls quadrangle and makes comparisons with the geology of the western Alps. The eastern Adirondack Mountains, which form the western part of the Glens Falls map area, are described geologically in Chapter B by James McLelland. In Chapter C, D.L. Daniels gives new gravity and magnetic data for the Mesozoic intrusive complexes at Cuttingsville and Mount Ascutney, Vt., and geophysical models for their geometry at depth. The next chapter (D), by J.N. Aleinikoff and Paul Karabinos, presents new U-Pb zircon data that help constrain the age of Paleozoic metasedimentary and metavolcanic rocks in eastern Vermont. In Chapter E, H.A. Pohn shows interpreted lineaments from Landsat and side-looking radar images and makes suggestions regarding spatially associated mineral deposits. A geochemical survey of the Glens Falls quadrangle by K.C. Watts, Jr. (Chapter F), reports detailed data for heavy-mineral concentrates of stream sediments and presents interpretations related to known and postulated metallic mineral deposits. Chapter G, by C.L. Long, describes the results of airborne electromagnetic surveys of the eastern half of the Glens Falls quadrangle and discusses several geophysically anomalous areas that were studied by ground-based electromagnetic traverses.

J.F. Slack and P.B. Schruben in Chapter H present a summary discussion of metal mines, prospects, and mineral occurrences in the quadrangle, with an emphasis on the commodities and production history of the largest deposits. In the next chapter (I), M.S. Power and N.M. Milton describe the results of geobotanical investigations, including airborne spectroradiometric surveys that show a remarkable correlation of airborne anomalies with known metallic mineral deposits. In Chapter J, R.H. Moench presents the results of new geologic mapping in the northeastern corner of the Glens Falls quadrangle and proposes the existence of a major tectonic unit, the Piermont allochthon, which has implications for mineral resource potential to the northeast in northern New Hampshire and west-central Maine. S.H.B. Clark describes the sediment-hosted, stratabound zinc-lead-copper deposits of the Lion Hill area, Vt., in Chapter K and suggests a

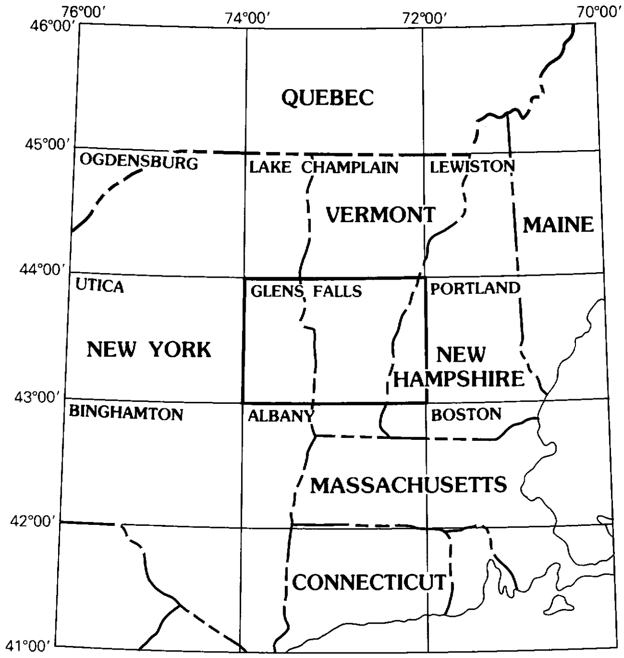

Figure 1. Location of the Glens Falls 1° × 2° quadrangle in west-central New England and surrounding quadrangles.

syngenetic-exhalative (or early diagenetic) origin for the mineralization; N.K. Foley, in the following chapter (L), presents fluid inclusion and lead isotope data for the same deposit.

In Chapter M, C.C. Cameron, P.B. Schruben, and E.L. Boudette report on studies of selected peat deposits in Vermont and western New Hampshire and on anomalous metal concentrations at two localities including high values for uranium. In Chapter N, L.J. Cox describes a geochemical survey of the Mount Ascutney intrusive complex, Vermont , and highlights results for gold, molybdenum, tungsten, and tin. G.R. Robinson, Jr., in the following chapter (O), presents geologic and geochemical data for gold-bearing epigenetic mineralization associated with the Cuttingsville intrusive complex and outlines a new model for Mesozoic syenite-related gold deposits in New England and adjacent Quebec. R.A. Ayuso and C.A. Ratté describe the geology and chemistry of bedrock uranium deposits in two areas of the Green Mountain massif in Vermont (Chapter P). J.F. Slack, P.J. Atelsek, and J.W. Whitlow in Chapter Q present the results of a detailed geochemical survey of stream sediments and heavy-mineral concentrates from the Orange County (Vt.) copper district, which includes the area surrounding the old Elizabeth mine. In the final chapter (R), J.F. Slack integrates data from all of the geologic, geochemical, and geophysical work with the results of topical studies of mineral deposits and assigns areas of low, moderate, and high resource potential for different metal commodities and deposit types in the quadrangle.

Appreciation is extended to Dr. Charles A. Ratté (State Geologist of Vermont), Dr. Eugene L. Boudette (State Geologist of New Hampshire), and Dr. Robert H. Fakundiny (State Geologist of New York), for their cooperation and encouragement during this project. The assistance of the former State Geologist of New Hampshire, Dr. Lincoln R. Page, is also acknowledged. Professor Barry L. Doolan and the Department of Geology, University of Vermont, helped in the arrangements for the public meeting in Killington, in 1986. Mr. William M. Kelly and Mr. Richard E. Nyahay of the New York State Geological Survey and Mr. Alan J. McBean, formerly of the Vermont Geological Survey, supplied data on mines, prospects, and mineral occurrences. Professors John B. Lyons and Half Zantop of Dartmouth College, Professor Wallace A. Bothner of the University of New Hampshire, Professor William S.F. Kidd of the State University of New York at Albany, Professor J. Gregory McHone of the University of Kentucky, and Professor Frederick D. Larsen of Norwich University (Vermont) also provided assistance during the project; Professor Larsen was especially helpful in supplying expertise on glacial deposits in the quadrangle.

Special thanks are given to the Ministère de l'Énergie et des Ressources du Québec and to Professor Michel Gauthier (Université du Québec à Montréal) for supplying information on mineral resource studies in the Quebec Appalachians. Professor Gauthier and his graduate students (particularly Jacques Trottier and Robert Godue) were very helpful in leading field trips to mineral deposits in southeastern Quebec and in exchanging data and ideas on new mineral deposit models for the northern Appalachians.

The production of this Bulletin was a joint effort involving the contributions of many USGS workers. I particularly wish to thank Katharine S. Schindler for her thorough and painstaking work on text editing of the manuscripts, Shirley A. Brown for coordinating the drafting of the illustrations, David A. Emery for map editing, Carolyn H. McQuaig for typesetting, and Brenda N. Woodson for transcribing manuscript files.

John F. Slack
Reston, Va.

Chapter A

An Introduction to the Geology and Paleozoic History of the Glens Falls 1° × 2° Quadrangle, New York, Vermont, and New Hampshire

By JAMES B. THOMPSON, JR.

U.S. GEOLOGICAL SURVEY BULLETIN 1887

SUMMARY RESULTS OF THE GLENS FALLS CUSMAP PROJECT, NEW YORK, VERMONT,
AND NEW HAMPSHIRE

CONTENTS

FIGURES

An Introduction to the Geology and Paleozoic History of the Glens Falls 1° × 2° Quadrangle, New York, Vermont, and New Hampshire

By James B. Thompson, Jr.[1]

Abstract

The Glens Falls 1° × 2° quadrangle of west-central New England includes little disturbed cratonal sediments, Proterozoic basement massifs, Late Proterozoic through Early Devonian metasedimentary and metavolcanic rocks, and igneous intrusions of Proterozoic, Paleozoic, and Mesozoic age. The basement massifs compose part of the ancient North American craton, although the more easterly ones are strongly overprinted by Paleozoic tectonism and metamorphism. Younger continental shelf, slope, and basin sediments, and minor subaqueous volcanic rocks of Late Proterozoic through early Paleozoic age, are preserved as autochthonous and allochthonous cover rocks in several lithotectonic zones. A volcanic arc of Ordovician age in central Vermont became welded to the North American craton and its cover rocks during the Ordovician Taconic orogeny. In the eastern half of the Glens Falls quadrangle, a separate Ordovician volcanic arc and its associated younger sediments became welded to North America during the Devonian Acadian orogeny. The lithotectonic zones in the Glens Falls quadrangle bear a remarkable similarity to those of the western Alps, both in their nature and in their order of succession.

INTRODUCTION

The Glens Falls 1° × 2° quadrangle encompasses a part of the New York-New England Appalachians and a segment of the adjacent North American craton. It extends from lat 43°00′ to 44°00′ N. and from long 72°00′ to 74°00′ W. and covers about 18,400 km^2, including portions of New York, Vermont, and New Hampshire. The northwestern corner of the quadrangle lies in the central Adirondacks of northern New York, and the southeastern corner is in the Monadnock-Sunapee region of southeastern New Hampshire. Mount Monadnock itself lies just south of the southeastern corner of the map. Thompson and others (1990) have prepared a generalized geologic map of the Glens Falls 1° × 2° quadrangle, at 1:250,000 scale.

Crystalline basement rocks of the North American craton are exposed extensively in the southeastern Adirondacks. Paleozoic deformation and metamorphism have had only local, minor effects on these rocks (Whitney and Davin, 1987). Immediately to the east in the Lake Champlain and Hudson valleys, however, the effects of Paleozoic events become strong and increase in intensity across the Taconic and Green Mountains of southern Vermont. The southeastern part of the Glens Falls quadrangle is in a migmatitic terrane of Paleozoic age in the central core of the New England Appalachians. The geology of the region can thus be conveniently described in terms of a northwest-to-southeast cross section.

Southeastward from the Adirondacks, we encounter a succession of distinctive lithotectonic zones that form natural subdivisions for discussion of the geology. As shown in figure 1 these are (1) the Adirondack border area, a little deformed Paleozoic cover; (2) the Middlebury synclinorium, mainly parautochthonous, Paleozoic carbonate rocks that exhibit westward thrusting and overturning of folds; (3) the Taconic allochthon, structurally above the rocks of the Middlebury synclinorium; (4) the Green Mountain massif, a re-emergence, in the deformed belt, of the cratonal basement; (5) the metamorphic cover sequence of the eastern Green Mountains; (6) the ductilely deformed nappe and gneiss dome terrane (Chester and Athens domes) of eastern Vermont that merges eastward into (7) the Connecticut Valley-Gaspé terrane; (8) the Bronson Hill terrane of western New Hampshire (another nappe and gneiss dome terrane) that is succeeded to the east by (9) the migmatitic Merrimack terrane that forms the western flank of the Merrimack synclinorium.

COMPARISONS WITH OTHER OROGENS

The succession of lithotectonic zones encountered between the Adirondacks and the Connecticut Valley region has features coextensive with, or analogous to, those appearing in other sections across the Appalachians. The section here, however, also has a striking resemblance, although offset in both space and time, to the succession of

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Manuscript approved for publication April 13, 1990.

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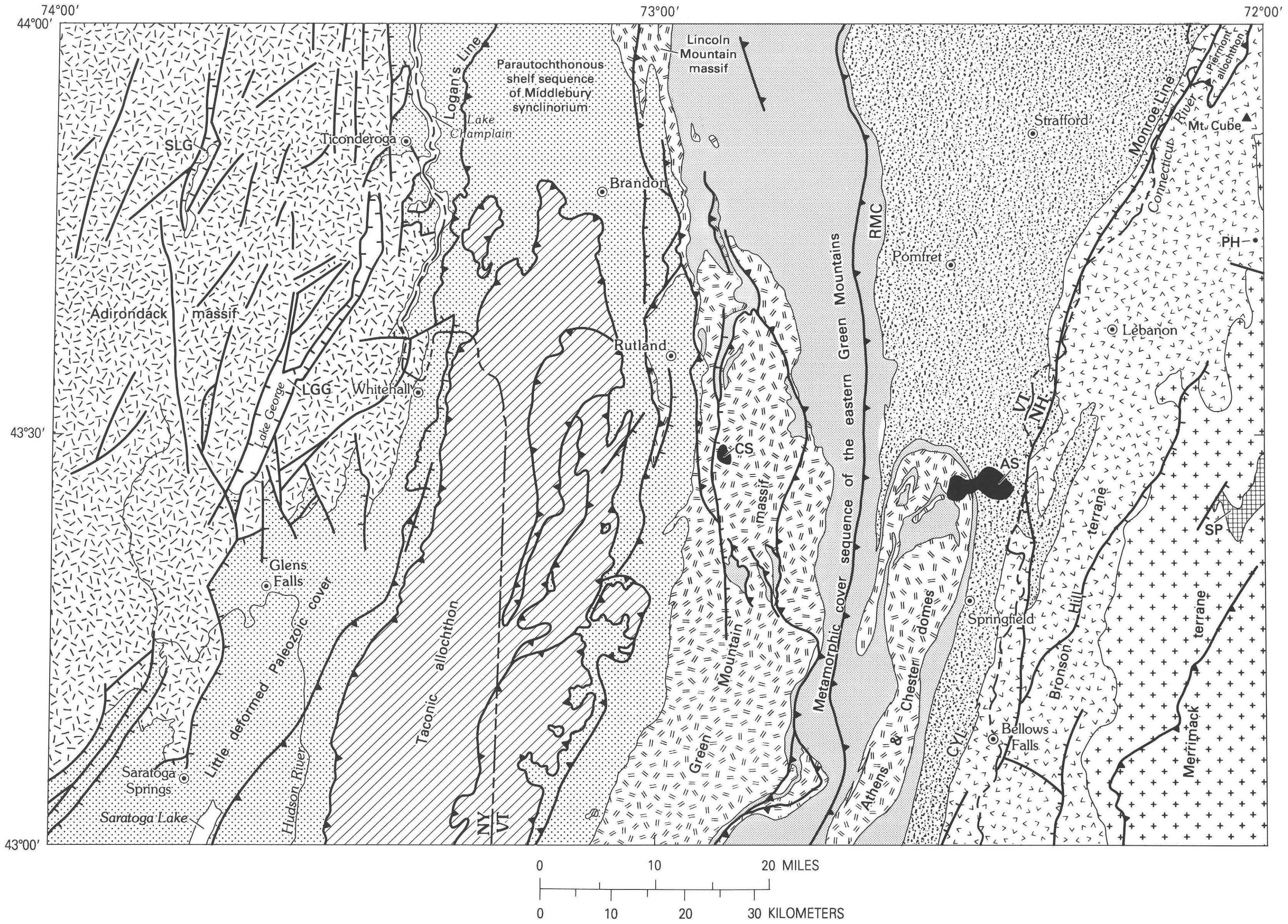


Figure 1. Major lithologic-tectonic units of the Glens Falls 1° × 2° quadrangle. As used here, terrane means a coherent set of related rocks and does not necessarily imply a distinct, accretionary, plate tectonic unit. SLG, Schroon Lake graben; LGG, Lake George graben; CS, Cuttingsville stock; AS, Ascutney stock; SP, Sunapee pluton; PH, Pollard Hill intrusives; RMC, Richardson Memorial Contact; CYL, Chicken Yard Line.

EXPLANATION

CRETACEOUS AND JURASSIC (?)	▓	White Mountain Plutonic-Volcanic Suite (syenite, granite, and gabbro)
UPPER DEVONIAN	▦	Granite plutons (two-mica granite)
LOWER DEVONIAN AND SILURIAN	⁺	Merrimack terrane (granitic gneiss, migmatite, minor pelitic schist, and quartzite)
	░	Connecticut Valley-Gaspé terrane (pelitic schist, calcareous schist, and minor amphibolite)
ORDOVICIAN AND CAMBRIAN	░	Parautochthonous shelf sequence (limestone, dolomite, and quartzite)
DEVONIAN, SILURIAN, AND ORDOVICIAN	░	Bronson Hill terrane (metavolcanic rocks, pelitic schist, and granitic gneiss)
ORDOVICIAN, CAMBRIAN, AND PROTEROZOIC Z	▨	Taconic allochthon (slate, graywacke, and minor carbonate)
	☐	Eastern Green Mountain terrane (pelitic schist, metagraywacke, and minor amphibolite)
PROTEROZOIC Y	░	Adirondack massif (felsic gneiss, quartzite, anorthosite, amphibolite, calc-silicate rocks, and metagabbro)
	░	Green Mountain massif[1] (felsic gneiss, quartzite, minor amphibolite, and calc-silicate rocks)
	┴	High-angle fault—Tick marks on downthrown side where known
	▲	Thrust fault—Sawteeth on upper plate

[1]Includes Lincoln Mountain massif and Proterozoic rocks of the Chester and Athens domes.

Figure 1. Continued.

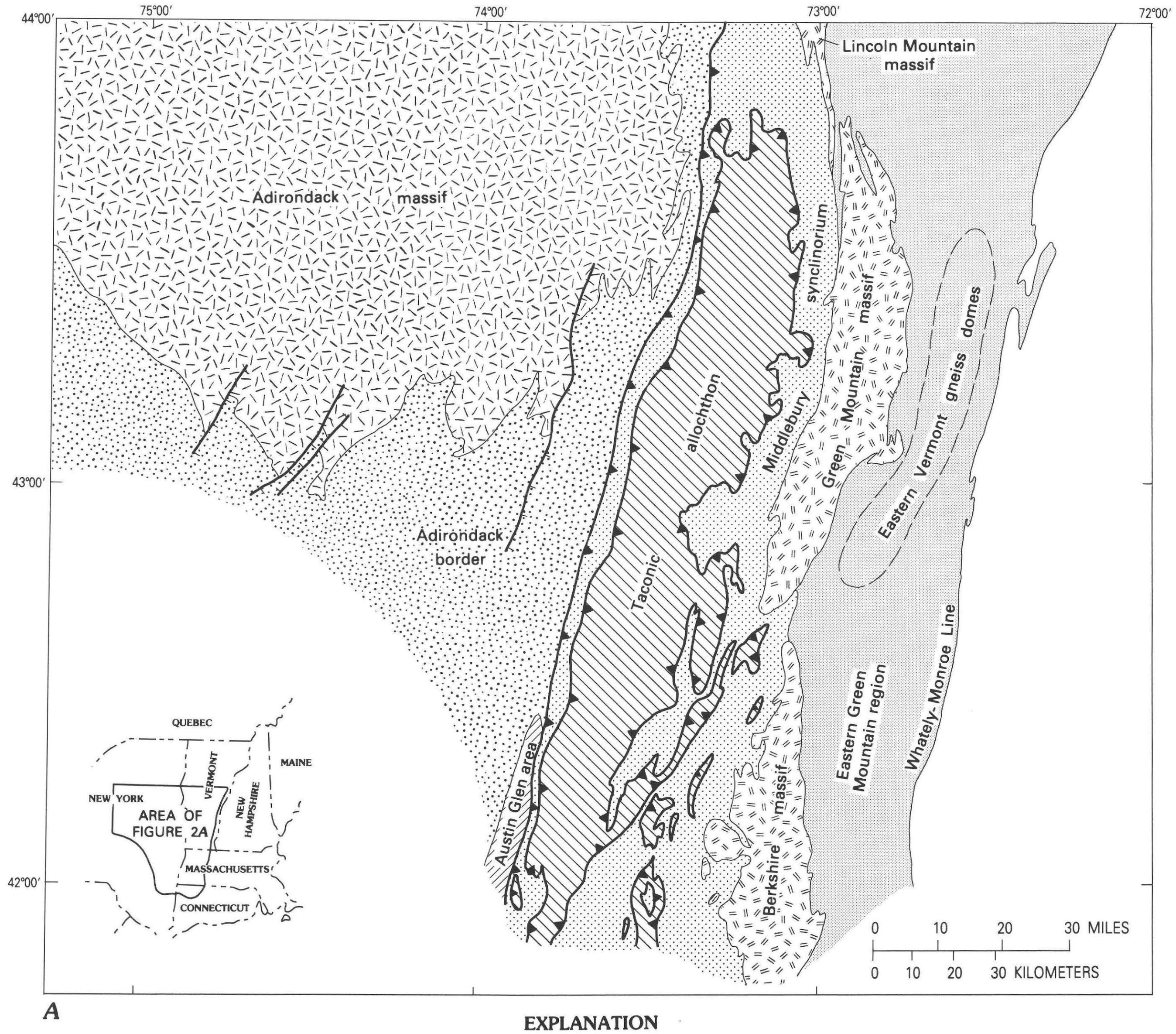
zones displayed in the western Alps. By proceeding south from the Vosges or Black Forest massif, across the pre-Alpine allochthons, the Aar massif, and the metamorphic terranes of the Pennine nappes to the Insubric Line (Spicher, 1972a), a similar succession of lithotectonic zones is encountered. Figure 2 shows simplified maps, at the same scale, of corresponding segments of these two orogens. Although current interpretations differ, the comparisons between these two areas extend to details of lithostratigraphic sequence, metamorphic textures, and mineral assemblages that are apparent at either the outcrop or thin-section scale. These similarities strongly suggest that both areas were subjected to a similar series of geological events. As knowledge of both areas improves, the differences in interpretation may perhaps become fewer or even vanish altogether. It thus seems unwise to try to understand one region without some knowledge of the other. For this reason several of the more conspicuous correspondences will be discussed.

THE ADIRONDACKS—A CRATONAL UPLIFT

The Adirondacks are one of a series of uplifts that have arched the surface of the Precambrian basement along the eastern and southern margins of the present-day North American craton. Other uplifts include the Laurentians of southeastern Quebec, the Jessamine and Nashville domes of the Cincinnati arch of the east-central United States, the Ozark uplift of southern Missouri and Arkansas, and the Llano uplift of central Texas. These massifs record warping and uplift, much of it demonstrably during Paleozoic time, that produced as much as 2,000 m of relief in the pre-Paleozoic surface of the Precambrian basement. They may thus be regarded as "foreland massifs" and as such are in many ways analogous to the foreland massifs (Black Forest, Vosges, Massif Central) peripheral to the western Alps. The rocks exposed in the Adirondacks (see McLelland, this volume) are part of the billion-year-old Grenville terrane that occupies much of the southeastern portion of the North American craton. This terrane contains a variety of rocks that were tectonized and metamorphosed under the conditions of the pyroxene granulite facies. These rocks include hypersthene-bearing granitic rocks (charnockites), quartz-feldspar-sillimanite-garnet rocks (kinzigites) that are commonly graphitic, coarsely crystalline marbles and calc-silicate rocks, and mafic rocks characterized by garnet and clinopyroxene. The metasedimentary rocks have a marked similarity to those in the Khondalite Series of India. Among the original protoliths are a variety of sedimentary and igneous rock types, including diverse gabbroic rocks and a large mass of anorthosite.

The rocks of the Adirondack basement are overlain with profound unconformity by a cover of early Paleozoic sedimentary rocks that, except in the east and southeast, dip gently outward from the central, domical uplift. The uplifting probably began early in the Paleozoic and may have continued sporadically through the Mesozoic and into modern times.

The eastern and southeastern margins of the Adirondacks, however, are cut by high-angle faulting. Much of the movement on these faults clearly postdates Middle Ordovician sedimentation, but Bosworth and Putman (1986) give evidence for Proterozoic movement on at least some of them. Although the map of Fisher and others (1970) implies



EXPLANATION

	GLENS FALLS AREA		WESTERN ALPS
Cratonal massifs:	Adirondack massif	⁄⁄	Vosges and Black Forest massifs
Valley and Ridge:	Austin Glen area	▨	Jura Mountains
Relatively undeformed sediments:	Adirondack border	⠿	Rhine graben and Swiss plain
Parautochthonous zone:	Middlebury synclinorium	░	Calcareous Alps
External allochthons:	Taconic allochthon	▧	Pre-Alpine allochthons
External massifs:	Green Mountain, Lincoln Mountain, and Berkshire massifs	⁄⁄	Aar, Gotthard, Mont Blanc, and Aiguilles Rouges massifs
Metamorphic core:	Eastern Green Mountain region	▒	Pennine Alps

Figure 2. Comparison at the same scale of (*A*) the Glens Falls 1° × 2° quadrangle and (*B*) the western Alps. In (*A*), the information from the area south of lat 43°00′ and the information from the area west of long 74°00′ is from Doll and others (1961), Fisher and others (1970), and Zen and others (1983). In (*B*), the information is from Spicher (1972a).

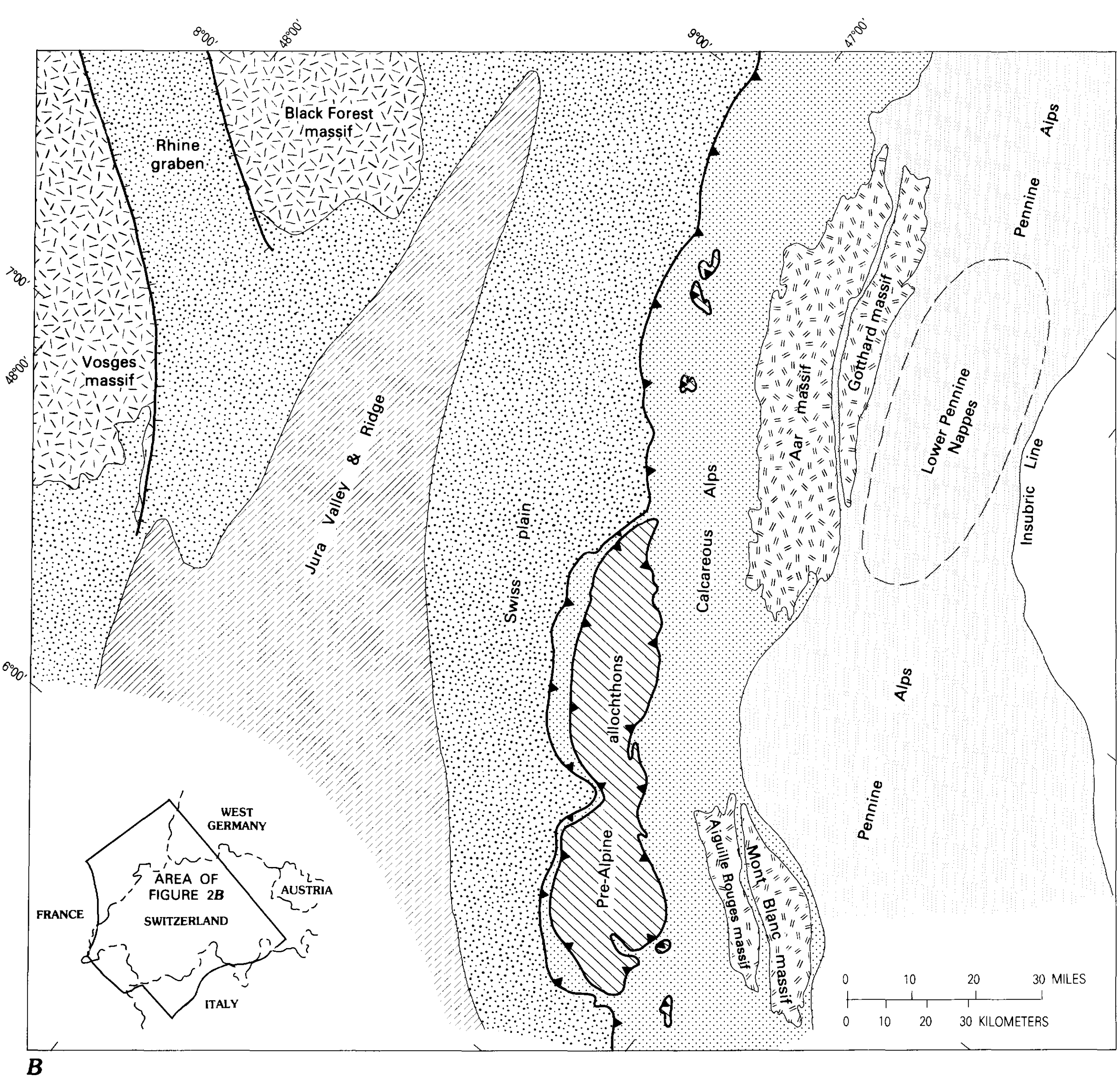


Figure 2. Continued.

that the Paleozoic movements on these faults are largely pre-Silurian, Ratcliffe (1971) suggested that Mesozoic rifting has reactivated some of the older faults. Further reactivation with movements extending into more recent times is indicated by seismic activity and changes in elevation (Isachsen, 1975; Barnett and Isachsen, 1980). The Paleozoic cover has been dropped locally as much as 2 km in well-defined graben, notably at Schroon Lake and Lake George (fig. 1). Southeast of Lake George, the craton is broken by a series of high-angle faults, downthrown on the west, into a series of tilted blocks. On each of these blocks, the Paleozoic cover dips gently east or southeast.

THE PALEOZOIC COVER TO THE ADIRONDACKS—A PLATFORM SEQUENCE

This Paleozoic cover sequence has recently been described in detail by Fisher (1984). Fisher's work deals mainly with the Glens Falls-Whitehall segment of the Adirondack border region but gives extensive references to recent work in adjacent areas and to the earlier work of Rodgers (1937), Welby (1961), and others. Palmer (1971) has provided an extensive discussion of the Cambrian stratigraphy of this area and its relation to the stratigraphy of other parts of the Appalachian region. The basal unit is the

Upper Cambrian Potsdam Sandstone, mainly a clean quartz sandstone, locally crossbedded, that has a pebble or cobble conglomerate at its base. It is overlain by the Upper Cambrian Ticonderoga Formation (Fisher, 1984), which is mainly dolomite, sandy dolomite, and minor sandstone. The Ticonderoga Formation is succeeded by approximately 200 m of dolomite, dolomitic limestone, limestone, and minor quartz sandstone of the Beekmantown Group (as used by Fisher, 1984), ranging in age from Upper Cambrian (lower part of the Whitehall Formation) through much of the Lower Ordovician. East and northeast of Ticonderoga, N.Y., along the shores of Lake Champlain, the Beekmantown carbonates are succeeded by early Middle Ordovician carbonates of the Chazy Group (Welby, 1961; Fisher, 1984; Selleck and Bosworth, 1985). The Chazyan and underlying carbonate rocks and sandstones record an onlap from the east and are regarded as constituting the remnants of a carbonate bank on the eastern (passive) margin of North America. South of Ticonderoga, the Chazyan strata are cut out by an unconformity at the base of the Middle Ordovician Black River and Trenton Groups. Although the latter contain carbonates, these grade upward and eastward into Mohawk valley shales (Canajoharie Shale, Snake Hill Formation) that clearly derive much of their substance from an eastern source. The presence of bentonites in these shales is evidence that at least some of the source material is volcanic in origin and is possibly related to the formation of an island arc offshore at roughly the end of the Chazyan sedimentation.

THE TACONIC AND GREEN MOUNTAIN REGIONS—A PALEOZOIC CONTINENTAL SHELF AND SLOPE

The relatively undisturbed Paleozoic cover rocks of the eastern Adirondacks are bounded on the east by a series of westward-verging thrust faults. The westernmost of these faults define Logan's Line, a feature that extends south into Vermont from the province of Quebec. These thrust faults also mark the western limit of strong Paleozoic deformation and may be traced south through western Vermont, along or near the western shore of Lake Champlain, into the Saratoga-Glens Falls region of New York State.

Two extensive exposures of Grenville basement rocks occur in this area as well as several smaller ones. The largest exposure is in the core of the Green Mountains of south-central Vermont. There are also large, but less extensive, exposures in the Athens-Chester-Cavendish region of southeastern Vermont about midway between the crest of the Green Mountains and the Connecticut River valley. The basement rocks of the Green Mountains show a history of metamorphic and deformational events that is like that shown by the Hercynian basement rocks in the Aar and Gotthard massifs of the western Alps. In each area, a younger, low-grade metamorphism has been superposed on an older, high-grade basement terrane. The Green Mountains and the Berkshire and Hoosatonic highlands to the south may thus be regarded as belonging to a series of external massifs extending along the Appalachians from the Blue Ridge of the southeastern United States to the Long Range of Newfoundland. This is the Blue-Green-Long axis of Rankin and others (1983).

The Paleozoic rocks between Logan's Line and the western flank of the Green Mountains occur in two distinct terranes: the parautochthonous shelf sequence of the Middlebury synclinorium and an allochthonous sequence lying on the rocks in the central part of the synclinorium. The allochthonous sequence is in a series of slices comprising the phyllites of the Taconic Mountains and the slates of their western foothills. These allochthonous rocks are believed to have been emplaced during the latter part of the Middle Ordovician at the onset of the Taconian orogeny, an event that is also recorded farther west in the clastic sedimentary rocks of the Queenston delta in western New York and Ontario.

The Green Mountains and Chester-Athens Domes—Reworked Cratonal Basement

Much of the Green Mountain basement terrane is composed of felsic gneiss and amphibolite, but metasedimentary rocks including graphitic schists, massive quartzite, and coarsely crystalline marbles and calc-silicate rocks are also present. All of these units closely resemble units found in the Grenville basement of the southeastern Adirondacks and presumably are correlative with them. This is supported by the recent radiometric dating by Karabinos and Aleinikoff (1988). Felsic gneisses are dominant in the Chester and Athens domes, but the Athens dome also contains marble, calc-silicate rocks, and extensive areas of graphitic schist.

Paleozoic deformation and metamorphism (recording Late Ordovician to Middle or Late Devonian tectonic activity) are locally strong in the Green Mountain basement terrane (Brace, 1953), but many features of the Grenville have survived, including traces of relict mineral assemblages characteristic of the pyroxene granulite facies (McLelland, this volume). The Paleozoic metamorphism has caused extensive, coarse saussuritization of calcic feldspars and chloritization of earlier garnets and pyriboles. It has also produced a blue opalescence in much of the older quartz, caused by the exsolution of fine rutile needles. In the more southeasterly occurrences, in the central gneisses and basement-cored nappes of the Chester and Athens domes, however, the basement rocks are much more strongly reworked (Thompson, 1950; Rosenfeld, 1954). The structural (highly ductile) and metamorphic (coarsely recrystallized kyanite zone) features of the Chester and Athens domes have much in common with those found in the lower basement-cored nappes of the Pennine Alps, particularly in the Simplon region and in Val Leventina, south of the Gotthard Pass (Spicher, 1972b).

The Middlebury Synclinorium—A Parautochthonous Shelf Sequence

The Middlebury synclinorium is strongly overturned to the west. Its western limb is cut by thrusts of Logan's Line, and its eastern limb is cut by thrusts along or near the western flank of the Green Mountains. The stratigraphic succession unconformably overlies Grenville basement and begins with basal clastics of Late Proterozoic (Dalton Formation, Pinnacle Formation) to earliest Cambrian age (Cady, 1945; Cady and Zen, 1960; Zen, 1964; Thompson, 1967). These are succeeded, in turn, by Lower Cambrian quartzites (Cheshire Quartzite) and by Lower and Middle Cambrian dolomites (Rutland Dolomite, Dunham Dolomite (as used by Cady, 1945), Winooski Dolomite) with minor quartzite (Monkton Formation). The Upper Cambrian Danby Formation consists of interbedded quartzite and dolomite and is at least in part correlative with the Potsdam Sandstone and Ticonderoga Formation that rest directly on the basement in the Adirondack border region. Above the Danby is a series of limestones and dolomites directly correlative with the Beekmantown and, in part, with the Chazyan carbonates of the Adirondack border. These carbonates are unconformably overlain, as in the Adirondack border region, by Middle Ordovician shales and minor carbonates. Metamorphic conditions are those of the greenschist facies, increasing in grade eastward. The metamorphism is, at least in part, of Taconian age but has an Acadian overprint eastward.

The Taconic Allochthon—A Slope Sequence

The rocks of the Taconic allochthon rest in tectonic contact upon units in the central portion of the Middlebury synclinorium, from Brandon, Vt., southward (Zen, 1967). At many localities the Ordovician shales immediately beneath the Taconic allochthon, especially on the west, contain masses of wildflysch or mélange (Forbes Hill Mélange of Fisher (1984) and Rysedorph Conglomerate) containing exotic blocks of carbonate rocks, graywackes, and (locally) of pillow lava (notably at Starks Knob in the upper Hudson Valley). Elsewhere the base of the allochthon rests on an imbricate zone that contains thrust slices of parautochthonous carbonates. The rocks of the allochthon include a variety of red, green, purple, gray, and black slates. These have been quarried extensively in the western Taconic region for slate and flagstone, commercial products that have been widely distributed over North America. In the eastern, more rugged parts of the Taconic Range, the slates pass into phyllites (commonly containing chloritoid) and fine-grained schists.

The stratigraphic sequence within the Taconic allochthon (Zen, 1961; Theokritoff, 1964) has recently been revised by W.S.F. Kidd and coworkers (Rowley and others, 1979; Kidd and others, *in* Fisher, 1984). The oldest clastic rocks (as on the western flank of the Green Mountains and in the structurally underlying parautochthonous sequence of the Middlebury synclinorium) are believed to be Late Proterozoic in age (Nassau Formation). Some of the coarser clastic members of the Nassau Formation, such as the Bomoseen Graywacke and Zion Hill Quartzite (as exposed on Bird Mountain, Vt.) and the Rensselaer Graywacke (exposed just south of the Glens Falls map), have been interpreted as having been deposited during Late Proterozoic rifting of an ancient supercontinent (Bird and Dewey, 1970). These basal clastics are succeeded by Cambrian and Ordovician slates or phyllites, associated with turbidites, cherts, and rare carbonate rocks. The Cambrian section includes the Browns Pond (Fisher, 1984), Middle Granville (of Fisher, 1984), Bull, Mettawee, and West Castleton Formations. The Hatch Hill Formation is latest Cambrian and earliest Ordovician. The Ordovician section includes the Poultney Formation and the formations of the Normanskill Group of Fisher (1984). These formations parallel the autochthonous and parautochthonous carbonate sequences in age but are of a lithologic facies more appropriate to deposition on a continental slope. The parautochthonous sequence is largely carbonate and is generally interpreted as a shelf sequence on the eastern margin of a Late Proterozoic supercontinent. The sedimentary rocks of the Taconic allochthon are believed to have originally been deposited farther east on the then newly created continental slope, although some of the coarser clastics may represent an earlier rift facies.

THE METAMORPHIC TERRANES OF EASTERN VERMONT

The metamorphosed and highly deformed cover rocks on the eastern flank of the Green Mountains and on the flanks of the Chester and Athens domes are postulated to be Late Proterozoic through early or middle Paleozoic in age, although biostratigraphic evidence to support this contention is meager (Chang and others, 1965; Thompson, 1972). These rocks are separated into two distinctive terranes by a boundary that is known to local geologists as the Richardson Memorial Contact or RMC, after C.H. Richardson, an early student of this region. The RMC has long been interpreted as an eastward-facing unconformity, but some recent investigators prefer to regard it as a tectonic contact (Westerman, 1987). In any case, the original boundary has been offset locally by later faulting.

Eastern Green Mountain Terrane

The rocks to the west of the RMC are the "talcoid" and "hydromica" schists of early 19th century geologists. These rocks include conglomeratic units (Tyson Formation, as used by Doll and others, 1961) at their western contact of the RMC and the Green Mountain basement. The rocks farther east, however, are mainly mica schists, derived from shales; greenstones, amphibolites, and gneisses of volcanic origin; quartzites, some of which are probably metamorphosed cherts; and rare carbonate rocks. The protoliths of these rocks are consistent with their having been laid down on a lower continental slope or ocean floor. The mapped units, including the Hoosac, Pinney Hollow, Ottauquechee, and Stowe Formations (probably Cambrian to earliest Ordovician) and the Moretown Formation (probably Ordovician) are interpreted here as having been derived from a coherent, eastward-topping stratigraphic sequence in the order listed. Bimodal volcanics, mapped by Doll and others (1961) as the Barnard Volcanic Member of the Missisquoi Formation, are interpreted as having been derived from an offshore island arc that developed during the Middle Ordovician. This sequence is regarded as a time equivalent of the Taconic sequence but one that represents a more oceanic environment. The biostratigraphic control, however, is admittedly meager (Chang and others, 1965).

The rocks west of the RMC show evidence of intense strain with a pervasive east-over-west shear that by itself can be responsible for an enormous amount of tectonic transport. The interpretation put forth by Stanley and Ratcliffe (1985), however, postulates that most of the boundaries between the map units are major discrete thrusts and that the above sequence is purely lithotectonic rather than stratigraphic.

Ultramafic rocks occur mostly as small lenticular bodies of serpentinite or steatite (only a few measure more than a kilometer in their longest dimension) mainly in the Ottauquechee and Moretown Formations (Chidester, 1962; Jahns, 1967). The larger ones have harzburgite or dunite cores, but the smaller ones are largely serpentinite, and many have an outer shell of steatite (talc-carbonate) rock. These lenticular bodies are typically associated with mafic metavolcanic rocks that are now greenschist or, locally, epidote amphibolite or amphibolite. Some of the more siliceous metasediments in the country rocks, especially in the Moretown Formation, may be, at least in part, metaradiolarite. The ultramafic rocks are here interpreted as intrusives, and possibly extrusives, related to ocean floor volcanism. Stanley and Ratcliffe (1985), however, argue that they are thrust slices derived from deeper oceanic crust. The ultramafic rocks of the Thetford region in southern Quebec are more or less on strike to the north.

In northern Vermont, this zone was subjected to a Taconian blueschist metamorphism (Laird and Albee, 1981). In the latitude of the Glens Falls area, however, the evidence for this early metamorphism has been almost completely obliterated by a later Acadian metamorphism of Barrovian type (Laird and others, 1984; Christensen and others, 1989). Locally, epidote amphibolite and lower amphibolite facies (garnet zone or staurolite zone) conditions have been reached.

The Connecticut Valley-Gaspé Terrane

The rocks to the east of the RMC are the "calciferous" schists of the 19th century geologists but now are assigned to the Northfield, Waits River, and Gile Mountain Formations. These formations occupy a broad area in the Connecticut Valley-Gaspé synclinorium. To the west are phyllites or schists and quartzose, micaceous carbonate rocks, formed by the metamorphism of calcareous shales and shaly limestones (calc-pelites). Conglomerates, limestones, and metavolcanic rocks (Shaw Mountain Formation) are also present near the western margin. To the east the metamorphosed calcareous shales (Waits River Formation) are less abundant and give way to interbedded schists and quartzites derived from turbidites (Gile Mountain Formation). Unnamed conglomerates are also found locally near the eastern margin. A central marker unit (Standing Pond Volcanics) contains pillow lavas, tuffs, and rocks of mixed sedimentary and volcaniclastic origin. The latter are locally metamorphosed and form spectacular garnet-hornblende garbenschiefer. Sparse fossils (Currier and Jahns, 1941; Cady, 1950; Doll, 1984; Hueber and others, 1990) have been variously dated as Middle or Late Ordovician, Silurian, or even Devonian. Recent U-Pb dating of zircons from a dike rock near Springfield, Vt., provides a minimum age of 423±4 Ma for the Standing Pond Volcanics (Aleinikoff and Karabinos, this volume). Most investigators, however, believe these rocks to be at least as young as post-Taconian (Hueber and others, 1990). The major rock types of this terrane bear a marked resemblance to the Schistes Lustrés or Bündnerschiefer of the western Alps.

The (Acadian) regional metamorphic grade reaches the staurolite-kyanite zone in the area of the Chester and Athens domes and also in the areas of what are probably buried domes farther north near Pomfret and Strafford, Vt. Farther east, toward the valley of the Connecticut River, the metamorphic grade falls off steeply toward a narrow zone of slates and low-grade phyllites.

The west bank of the Connecticut River is the political boundary between Vermont and New Hampshire but weaves back and forth across the geologic boundary that separates two distinct lithologic terranes. North of Springfield, Vt., this boundary is known as the Monroe Line. South of Springfield, it is known informally (from an exposure near East Dummerston, Vt.) as the Chicken Yard Line or CYL. The nature of both lines is enigmatic. The Monroe Line has been interpreted as a fault (Hatch, 1988) and as a westward-facing unconformity (Doll and others, 1961), and the CYL has been interpreted as either a west-directed thrust fault (the Whately thrust of Robinson and others, 1988) or as an eastward-facing unconformity (Hepburn and others, 1984). Both lines, however, have been offset by undisputed late faults, at least some of which are a northward continuation of the fault system that forms the eastern boundary of the Mesozoic lowlands just to the south in central Massachusetts and Connecticut.

The interpretation shown on the accompanying map (fig. 1) implies that the Monroe Line is a westward-facing unconformity and that the CYL is an eastward-facing unconformity. The "calciferous schists" of the Glens Falls quadrangle are therefore interpreted as occupying a Silurian (perhaps also, in part, Late Ordovician) interarc basin and thus as older than the adjacent part of the Devonian Littleton Formation of western New Hampshire. The "calciferous schists" and the Littleton Formation meet along the CYL, and both units show stratigraphic tops facing east across the line. If the relative ages of these units were reversed, as Robinson and others argue, then the CYL would be a thrust.

THE BRONSON HILL TERRANE OF WESTERN NEW HAMPSHIRE

The rocks east of the Monroe Line and the CYL belong to a zone dominated by a large-scale structural feature known as the Bronson Hill anticlinorium (Thompson and others, 1968; Robinson and others, 1979). This anticlinorium is, in fact, a series of roughly en echelon gneiss domes whose structures are complicated by a set of westward-verging nappes, partly fold nappes and partly overthrust nappes, that were emplaced during an early phase of the Acadian orogenic movements. These nappes brought hot rocks over cold and produced inverted metamorphic isograds along the western flank of the Bronson Hill anticlinorium. The cores of the nappes are occupied by Ordovician, and possibly older, metasedimentary rocks (Partridge and Quimby Formations) and minor metavolcanic rocks, all invaded locally by granitic masses. These are unconformably overlain by either metamorphosed Silurian conglomerates, quartzites, and minor schists of the Clough Formation or by schists and quartzites of the Rangeley and Perry Mountain Formations. These basal units are conformably overlain by calc-silicate rocks and schists of the Fitch or Smalls Falls and Madrid Formations; and these, in turn, are overlain by a sequence of metamorphosed shales and turbidites of the Devonian Littleton Formation. Fossils in the Silurian and Devonian units occur at several localities along the Bronson Hill zone (Boucot and Thompson, 1963) and in the area immediately northeast of the Glens Falls area (Boucot and Rumble, 1980, and references therein; Harris and others, 1983).

The axial surfaces of the nappes of the Bronson Hill zone nappes are themselves strongly deformed by the later rise of a series of gneiss domes, in a roughly en echelon array, just east of the Connecticut River. The cores of these domes are reworked Ordovician, and possibly older, granitoid plutonic rocks (Zartman and Leo, 1985), mantled by the Ordovician Ammonoosuc Volcanics (Leo, 1985; Schumacher, 1988) and, locally, by metasedimentary rocks of the Partridge and Quimby Formations. The Silurian sequence is dominated by quartzites and calc-silicate rocks (Clough and Fitch Formations, respectively) in the mantles of the domes and in the lower nappes. In the upper nappes, the Silurian section is much thicker, and mica schists are more abundant (Rangeley, Perry Mountain, Smalls Falls, and Madrid Formations). These differences in the Silurian section are consistent with a westward onlap onto the older rocks of the Bronson Hill zone, which was apparently a positive area during the Silurian sedimentation.

In the Bronson Hill zone, metamorphic grade increases eastward from slates and phyllites near the CYL to a staurolite-kyanite zone adjacent to the domes. The isogradic surfaces associated with the nappes are inverted so that rocks in the upper nappes are in the amphibolite facies (sillimanite zone). Some of the rocks in the uppermost nappes even appear to have once crossed the second sillimanite isograd, although the evidence for this crossing has largely been obliterated by retrograde processes. In any given nappe, metamorphic grade also increases eastward toward its root zone.

A small area of rock on the western flank of the Bronson Hill anticlinorium, in the extreme northeastern corner of the Glens Falls quadrangle (fig. 1), has been interpreted by Moench (this volume) to be a part of a large Piermont allochthon, a series of klippen derived from a southeasterly source. The Piermont allochthon may correspond to the Fall Mountain nappe, the uppermost of the nappes mapped farther south by Thompson and others (1968).

THE MERRIMACK TERRANE OF NEW HAMPSHIRE

The terrane extending from the east side of the Bronson Hill zone to the eastern limit of the map (Hatch and others, 1983) is called the Merrimack or Kearsarge-central Maine synclinorium. This synclinorium is believed to be the root zone of the Bronson Hill nappes and the Piermont allochthon, although investigators differ as to details. The rocks are in the amphibolite to granulite facies (sillimanite and second sillimanite zones) and are locally migmatitic. Several large, foliated, tabular bodies of gneissic granitoid rocks, notably the Bethlehem Gneiss and the Kinsman Quartz Monzonite, occupy much of the area. These gneissic granitoid rocks apparently rode west, atop the nappes, as indicated by the presence of several outliers, apparently floored (Kruger and Linehan, 1941), on the western flank of the Bronson Hill terrane. The principal outliers are at

Bellows Falls, Vt., and in the Mascoma Lake and Mt. Cube regions of New Hampshire. The rocks of these plutons are peraluminous and have a strong gneissic fabric, particularly in the more westerly occurrences. There are also smaller bodies of late or early postkinematic, peraluminous (or S-type) granitic rocks such as the Sunapee pluton (fig. 1). Large granite pegmatites are abundant, especially near the upper contact of the Bethlehem Gneiss. These and related rock types constitute the New Hampshire Plutonic Series of Billings (1956). Radiometric studies have established an Acadian (413 Ma) age for the foliated Kinsman Quartz Monzonite (Barreiro and Aleinikoff, 1985). The postkinematic Sunapee pluton, however, has a post-Acadian (366 Ma) age based on recent U-Pb zircon data (J.N. Aleinikoff, oral commun., 1989; see also Harrison and others, 1987).

WHITE MOUNTAIN PLUTONIC-VOLCANIC SUITE

Postmetamorphic dikes of Mesozoic age are numerous throughout the map area (see McHone, 1978, and references therein). Many are diabases that have a northerly or northeasterly strike, roughly parallel to the trends of high-angle faults associated with the Mesozoic rifting of eastern North America. Camptonites and devitrified felsites are also found, especially near the intrusive complexes at Cuttingsville (Eggleston, 1918) and Mount Ascutney, Vt. (Daly, 1903; Chapman and Chapman, 1940; Schneiderman, 1989). These complexes are composed largely of syenite or quartz syenite, but there is some nepheline syenite at Cuttingsville. Biotite granite is also present at Ascutney. Gabbro-diorite is found at Little Ascutney, and essexite occurs at Cuttingsville. The Ascutney complex also includes trachytic volcanic rocks. These rocks show that there was once an overlying volcanic edifice. A remarkable breccia on Little Ascutney (Schneiderman, 1989) contains fragments some meters across of rocks no longer present at or below the present-day erosion surface. Smaller intrusives, possibly volcanic necks, occur in New Hampshire east of Charlestown and north of Canaan (see Billings, 1956).

Some of the more felsic rocks, especially the syenites, are peralkaline, and some, as at Cuttingsville, are feldspathoidal. The ferromagnesian minerals in the syenites and granites are rich in iron, and some of the quartz syenites contain fayalite. Radiometric studies (K-Ar method) by Armstrong and Stump (1971) and Foland and Faul (1977) give ages of approximately 100 and 120 Ma, respectively, for the Cuttingsville and Ascutney intrusive complexes.

OUTLINE OF PLATE TECTONIC HISTORY

Plate tectonic reconstructions of the Paleozoic are difficult and fraught with uncertainty. Because no magnetically striped ocean floor of that era remains, we must base such interpretations on the stratigraphic record preserved in the cratonal Paleozoic sediments and on what can be deduced from the contorted structures of the Paleozoic orogenic belts. Among the more notable syntheses, for New England, are those of Bird and Dewey (1970), Osberg (1978), Robinson and Hall (1980), Hall and Robinson (1982), Zen (1983), and Stanley and Ratcliffe (1985).

The Late Proterozoic and early Paleozoic depositional sequences in and west of the Green Mountains may be interpreted in terms of a Late Proterozoic rifting of the ancient supercontinent, followed by the development of a proto-Atlantic or Iapetus Ocean. The western autochthonous and parautochthonous sequences are remnants of a carbonate bank that accumulated on the newly formed continental shelf during the Cambrian and Early Ordovician. During the Middle Ordovician, carbonate deposition on the shelf changed to deposition of shale. Bentonites and other evidence of nearby volcanism are also present in these rocks and suggest a reorientation of plate movements. This reorientation is consistent with the development of an eastward-dipping subduction zone, perhaps far offshore, and an island arc on the upper plate. The Middle to Late Ordovician Taconian orogeny may represent the docking of this arc, and the Taconic allochthon may be the remnant of an obducted accretionary wedge. The island arc itself may be partially preserved in the rocks of the Barnard Volcanic Member of the Missisquoi Formation, which crop out just west of the RMC. The rocks west of the Barnard volcanic arc include metamorphosed sediments, some mafic volcanics, and several ultramafic bodies. These occurrences are consistent with deposition on an ocean floor that would have separated the Barnard volcanic arc from North America. A major thrust has been postulated just west of the outcrop area of the Barnard volcanic arc. It should be emphasized, however, that all the rocks between that thrust and their contact with the Green Mountain basement show evidence of intense ductile strain, east over west, and thus can account, in themselves, for major tectonic transport. The evidence for this transport is especially striking in some of the units exposed in the mantles of the Chester and Athens domes where snowball garnets, as large as 2 cm in diameter, show maximum rotations of nearly 720°. Snowball garnets showing nearly as great a rotation have also been described by Rosenfeld (1968, 1970) in the area east of the Chester and Athens domes.

The calc-pelites and associated rocks of the Connecticut Valley-Gaspé synclinorium may, in turn, represent deposition in interarc basins between the Barnard arc, the Chain Lakes terrane (about 150 km northeast of the Glens Falls region on the Maine-Quebec border), and a second island arc, partially preserved in the Bronson Hill zone. The sedimentary sequence of the Merrimack synclinorium may have been deposited in a similar marginal basin between the Bronson Hill zone and the exotic terrane or microcontinent of Avalonia, now represented by the 600-Ma basement of eastern Massachusetts and Rhode Island. Subsequent accretionary events include docking of the Bronson Hill arc; perhaps docking of the Chain Lakes terrane to the north of the Glens Falls region (Boone and Boudette, 1989); and, eventually, docking of the Avalonian terrane to the east and south of the Glens Falls region. Sutures joining these basement terranes are quite possibly hidden by younger units of the metasedimentary terranes that separate them. These metasedimentary rocks are themselves involved in major westward-verging nappes. By the end of the Devonian, the Bronson Hill arc, the Avalonian terrane, and part of the Fennoscandian or Baltic craton were firmly welded to the North American craton (Zen, 1983) to form Laurasia. These collisional (Acadian) events are recorded in the conglomerates and coarse clastics of the Middle to Late Devonian Catskill delta just south of the southwestern corner of the map area.

At the end of the Acadian movements, Iapetus in the strict sense no longer existed. Full closure of a more southerly and outboard ocean, the Phoibic of McKerrow and Ziegler (1972), to form Pangaea probably did not occur, however, until the end of the Paleozoic during the Alleghanian orogeny. This closure was presumably brought about by the arrival of northwest Africa (the Meguma terrane of southern Nova Scotia and Georges Bank) and other terranes farther south. New England appears to have received a glancing blow at this time that affected only the coastal regions, whereas the central and southern Appalachians bore the full brunt of this collision that produced the classic fold and thrust belt of the Valley and Ridge province, not present in the Glens Falls area. Pangaea, once formed, was apparently short lived. Renewed rifting in the Triassic and Jurassic produced the Ramapo fault system and its possible extension into the Hudson-Champlain valley (Stanley, 1980). This rifting also produced the Connecticut valley fault system farther east and associated volcanism and rift valley sedimentation. With the plutonic and volcanic activity of the White Mountain Suite and the opening of the modern Atlantic (Schlee and Klitgord, 1986), events leading to the present configuration of land and sea were underway.

ACKNOWLEDGMENTS

The foregoing summary is a brief distillation of more than 40 years of involvement with the geology of the northern Appalachians. My thoughts and prejudices have therefore been influenced, over this period, by many colleagues and students. A complete enumeration is impossible here, but I wish to mention three especially, John Rosenfeld (from the beginning) and Peter Robinson and E-an Zen (for nearly as long). They may disagree—even forcefully—with some of my interpretations, but decades of friendly exchange of ideas have made it difficult to separate, completely, their thinking from mine.

I also wish to thank Norman Hatch, Douglas Rankin, John Slack, and E-an Zen for helpful suggestions and criticisms at various stages in the preparation of this manuscript.

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Chapter B

Geology of the Adirondack Portion of the Glens Falls 1° × 2° Quadrangle

By JAMES McLELLAND

U.S. GEOLOGICAL SURVEY BULLETIN 1887

SUMMARY RESULTS OF THE GLENS FALLS CUSMAP PROJECT, NEW YORK, VERMONT, AND NEW HAMPSHIRE

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Geology of the Adirondack Portion of the Glens Falls 1° × 2° Quadrangle

By James McLelland[1]

Abstract

Metamorphosed igneous and sedimentary rocks underlying the Lake George region exhibit mineral assemblages characteristic of the granulite facies. Uranium-lead zircon studies indicate that the oldest rocks are ~1320-Ma tonalites and trondjhemites that occur throughout the eastern and southern Adirondack Mountains. These calc-alkaline rocks intrude some, and perhaps all, of the metasedimentary sequence. Subsequently, anorthositic, mangeritic, and charnockitic rocks were intruded as anorogenic bimodal complexes at ~1120 to 1160 Ma. Somewhat younger granitic gneiss (~1110 Ma) and alaskites (~1070 Ma) may have been emplaced during the 1070- to 1030-Ma granulite facies metamorphism. Earlier metamorphic events appear to have accompanied emplacement of the anorthositic rocks, and, prior to this, well-foliated garnet-sillimanite xenoliths are preserved in metagabbro dated at 1144±7 Ma. All rocks in the area were intensely deformed during granulite facies metamorphism. Deformation resulted in at least three major fold sets. The earliest two fold sets are represented by approximately coaxial, east-west-trending isoclinal (F_1) and upright (F_2) folds. An upright F_3 set results in fold interference patterns. Most rocks are marked by strong linear and planar fabrics, some of which are believed to be related to thrusting.

INTRODUCTION

The southeastern extension of the Adirondack highlands region (fig. 1), which includes Lake George, has been studied by a number of workers over the past 85 years. Their contributions are summarized in figure 2, which is keyed to references given in the list of cited references. Recent investigations include those of McLelland and Isachsen (1986) in the southern half of the area, Walton and deWaard (1963) in the northern half, and Geraghty (1973), Farrar (1976), and Turner (1980) in the central portion. This study compiles and synthesizes these earlier results with field studies conducted by the author during the summers of 1983–85. A generalized map incorporating much of this information is given in Thompson and others (1990).

MAJOR LITHIC UNITS

The wide variety of Adirondack rock types can be grouped into four major Middle to Late Proterozoic units—meta-anorthosite, metagabbro and metadiorite, mainly felsic (granitic) gneisses, and metasedimentary rocks (fig. 1). Detailed descriptions of these units and their subdivisions can be found in McLelland and Isachsen (1986) and references given therein.

Meta-Anorthosites

Meta-anorthosite underlies most of the area that is located near the northern margin of the Glens Falls 1° × 2° quadrangle (fig. 1). This extensive exposure represents only a small portion of the large Marcy anorthosite massif, which continues farther to the north. The rocks of the massif are homogeneous, coarse-grained andesine meta-anorthosite and have mafic contents that average close to 10 percent and increase to 20–25 percent near the outer margins of the massif. The coarse-grained anorthosite facies was designated Marcy-type anorthosite by Miller (1919), while the finer grained, more gabbroic variety was referred to as Whiteface-type anorthosite by Kemp (1898). Average whole-rock analyses for these two types are given in table 1 (cols. 1, 2). Grain-size reduction is common throughout the massif but is particularly intense near its contacts where it is generally accompanied by well-developed garnets. Where original textures are still preserved, subophitic textural relations can be observed between plagioclase and orthopyroxene.

Sheets of gabbroic anorthosite, three of which are shown on figure 1, occur within mangeritic-charnockitic gneisses that mantle the southern margin of the Marcy massif. In many instances, these anorthosites appear to intrude the gneisses. In addition, deformed xenoliths of gabbroic anorthosite are commonly found within the mangerites and charnockites (Miller, 1919), which are also known to crosscut the anorthosites at a number of localities (Buddington, 1939). These relations suggest that the mangeritic and anorthositic rocks were emplaced at essentially

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Manuscript approved for publication April 13, 1990.

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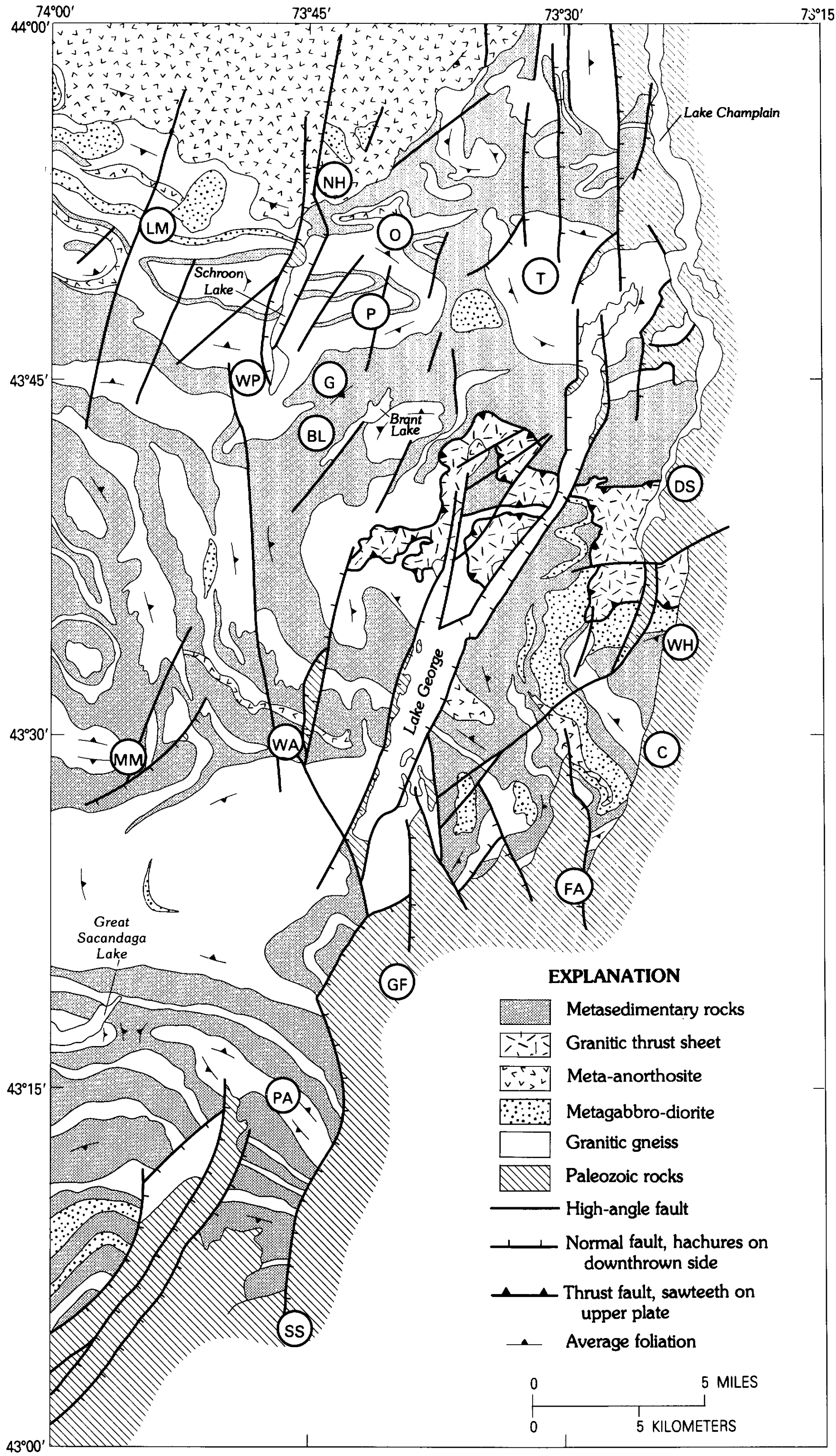

Figure 1. Simplified geologic map of the Adirondack portion of the Glens Falls 1° × 2° quadrangle. Lettered abbreviations stand for towns and topographic and geologic features; the locations of some geologic features referred to in the text are approximated by town symbols. BL, Brant Lake; C, Comstock, N.Y.; DS, Dresden Station, N.Y.; FA, Fort Ann, N.Y.; G, Grasshopper Hill; GF, Glens Falls, N.Y.; LM, Loch Muller; MM, Moose Mountain; NH, North Hudson, N.Y.; O, Owl's Head Mountain; P, Pharoah Mountain; PA, Piseco antiform; SS, Saratoga Springs, N.Y.; T, Ticonderoga dome; WA, Warrensburg, N.Y.; WH, Whitehall, N.Y.; WP, Whitney Point, N.Y.

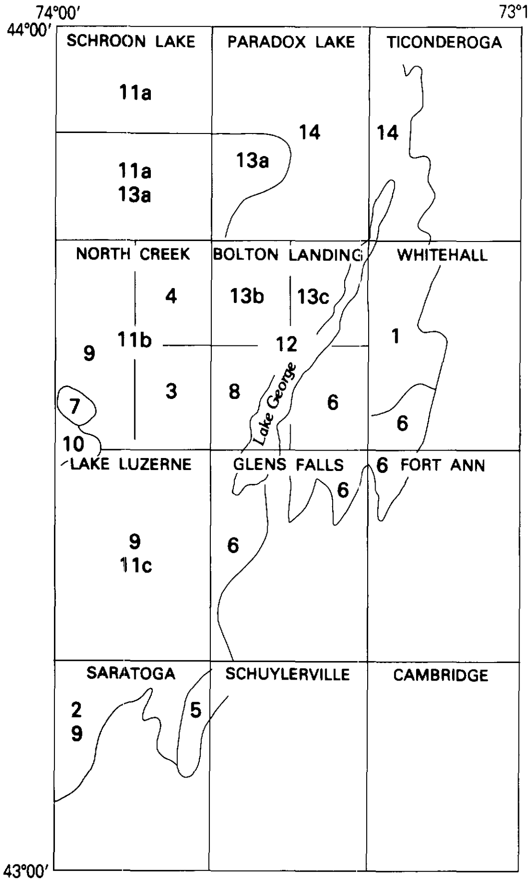

Figure 2. Index map of 15-minute quadrangles showing sources of geologic information. 1, Berry (1965); 2, Cushing and Ruedemann (1914); 3, Farrar (1976); 4, Geraghty (1973); 5, Hall (1966); 6, Hills (1961); 7, Isachsen (1966); 8, McConnell (1965); 9, McLelland (1985); 10, Metzger (1984); 11a, Miller (1914); 11b, Miller (1919); 11c, Miller (1923); 12, Newland and Vaughan (1942); 13a, Turner (1980); 13b, Turner (1963); 13c, Turner (1968); 14, Walton (1961).

the same time (McLelland, 1986; McLelland and Chiarenzelli, 1990). However, as documented by McLelland and Whitney (in press), these contemporaneous suites are not consanguineous but are members of a bimodal complex. A large magnetite-ilmenite deposit is found at Tahawus (Sanford Lake), just beyond the northwestern margin of figure 1. These oxide-pyroxene rocks (Gross, 1968) are interpreted as mafic cumulates (perhaps remobilized) that developed during the late stages of fractionation of the parent magmas of the anorthositic rocks (Ashwal, 1978).

Recent mapping by the author has revealed several large bodies of meta-anorthosite in the region between Comstock (fig. 1, C), Fort Ann (fig. 1, FA), and Lake George. The cores of these bodies resemble meta-anorthosite from the Marcy massif, and X-ray fluorescence whole-rock analyses document a similar chemistry (table 1, col. 3). A distinctive aspect of these recently recognized meta-anorthosite occurrences is their clear field association with norites and olivine metagabbros that appear to envelop and grade into the meta-anorthosite. This spatial association is similar to the concentration of large olivine metagabbro bodies near the margins of the Marcy massif where they sometimes crosscut the anorthosite (fig. 1). The close spatial association of the anorthosite and olivine metagabbros suggests that they may be petrogenetically related. Emslie (1978) presented a model for anorthosite genesis in which early formed gabbroic magmas pond at the crust-mantle interface and differentiate to leucogabbros parental to the anorthosites, as well as produce lower crustal melts of quartzofeldspathic composition. The olivine metagabbros bordering the Marcy massif may represent late examples of such ponded gabbroic magmas.

Unfortunately, the contacts between meta-anorthosite and other rocks are rarely exposed, and where they are, deformation has obscured most relations. Within the Glens Falls 1° × 2° quadrangle, anorthositic rocks are either in contact with gabbroic-noritic rocks or with mangerites. No instances are known where anorthosites in the strict sense clearly intrude metasedimentary rocks. However, it should be noted that there exist unequivocal instances in which gabbroic anorthosite and leuconorites intrude metasedimentary units (Buddington, 1939). The absence of dikes of true anorthosite suggests that, rather than representing a liquid, the anorthosites developed as crystal cumulates that were too viscous to be expelled from their magma chambers. An exceptionally good example of a crosscutting gabbroic anorthosite sheet is located west, north, and east of Warrensburg, N.Y. (fig. 1, WA), in the central part of the map area. Here a 100- to 200-m-thick sheet of gabbroic anorthosite can be mapped along strike for 12 km (Farrar, 1976). The charnockitic country rocks are clearly crosscut by this extensive sheet (fig. 1), whose composition is given in table 1 (col. 4). Another anorthositic sheet near Owl's Head Mountain (fig. 1, O), whose composition is given in table 1 (col. 5), clearly intrudes charnockitic country rocks.

Metagabbros and Metadiorites

Metagabbros are found throughout the area, but they tend to be associated with anorthositic rocks and are concentrated near the margins of the Marcy massif. Their chemistry is generally tholeiitic but ranges from olivine normative to noritic. The olivine-bearing varieties exhibit classic garnet coronas developed between olivine and plagioclase (McLelland and Whitney, 1980). Metadioritic and metatonalitic gneiss (table 1, col. 6) is shown with the same symbol as metagabbro on figure 1 but accounts for 90 percent of this unit east of Lake George (fig. 1). The metatonalites, which resemble charnockitic rocks in hand specimen, occur as thick, massive horizons that commonly grade into granodioritic and trondhjemitic lithologies (table 1, cols. 7, 8). An almost ubiquitous characteristic of the metatonalites is the presence of disrupted layers of

Table 1. Average composition of rocks of the Lake George region, in weight percent

Oxide	1	2	3	4	5	6	7
	Average Marcy anorthosite[1]	Average Whiteface anorthosite[1]	Buck Mountain Lake George	Moon Mountain Warrensburg	Owl's Head Paradox Lake	Metatonalite Whitehall	Granodiorite Piseco antiform
SiO_2	54.54	53.54	55.16	53.79	51.35	62.82	64.55
TiO_2	.67	.72	.67	.57	.52	.28	1.17
Al_2O_3	25.61	22.50	22.75	23.60	26.96	17.23	14.30
Fe_2O_3[2]	2.16	5.31	1.98	4.65	1.13	1.84	6.86
MnO	.02	.07	.02	.03	.02	.10	.10
MgO	1.03	2.21	2.95	1.05	1.47	.88	1.10
CaO	9.92	10.12	9.80	9.27	12.12	5.83	2.69
Na_2O	4.53	3.70	4.83	4.84	3.96	4.31	2.87
K_2O	1.01	1.19	.70	.94	1.21	2.13	4.65
P_2O_5	.09	.13	.15	.13	.09	.10	.45
H_2O	.55	.12	.27	.20	.31	2.78	.53
Total	100.13	99.61	99.28	99.07	99.14	98.30	99.27

Oxide	8	9	10	11	12	13	14
	Trondhjemite Skiff Mountain	Granite Gneiss Whitney Polnt	Charnockite Loch Muller	Mangerite Loch Muller	Keene Gneiss Loch Muller	Ferrogabbro Dresden Station	Alaskite Grasshopper Hill
SiO_2	72.39	70.16	67.72	62.85	58.90	41.54	72.93
TiO_2	.38	.53	.61	.71	1.63	6.37	.30
Al_2O_3	12.63	13.97	13.84	16.82	15.01	7.90	13.09
Fe_2O_3[2]	5.73	4.73	5.39	5.07	9.40	25.72	3.83
MnO	.02	.06	.07	.09	.14	.02	.02
MgO	.29	.38	.39	.41	2.11	6.24	.24
CaO	1.07	2.06	2.11	2.91	5.57	9.19	1.05
Na_2O	6.63	2.26	2.63	5.23	2.43	1.62	2.31
K_2O	.52	5.36	4.92	5.51	3.57	.74	5.50
P_2O_5	.07	.18	.12	.21	.46	.05	.05
H_2O	.36	.38	.34	.37	.37	.45	.42
Total	100.09	100.07	98.14	100.18	99.59	99.84	99.74

[1] Analyses from Buddington (1939).

[2] Total iron as Fe_2O_3.

pyroxene granulite, or amphibolite, that measure several tens of centimeters in thickness. These disrupted layers are believed to represent dismembered mafic dikes. A variety of xenoliths, including metasedimentary rocks, have been observed within the metatonalites. Most of these xenoliths appear to have undergone reaction with magma and are now represented by biotite- and feldspar-rich enclaves.

Felsic (Granitic) Gneisses

The felsic gneisses of the region consist predominantly of hornblende granitic gneiss, mangeritic-charnockitic gneiss, and leucogranitic (alaskitic) gneiss that are frequently interlayered with thin amphibolites. The most common of these felsic gneisses is pink, garnetiferous hornblende (-biotite) granitic gneiss. These rocks generally contain megacrysts of potash feldspar and are characterized by streaky patches and pods of different grain-size granitic material that give a migmatitic aspect to exposures. Throughout the Adirondack Mountains, these felsic gneisses tend to occur in large, internally homogeneous bodies that are suggestive of deformed and metamorphosed plutons. Their compositions (table 1, col. 9) are consistent with an igneous, plutonic origin. Occasional xenoliths support an intrusive history, although pervasive deformation has obscured such critical evidence as crosscutting contacts.

Charnockitic and mangeritic gneiss (table 1, cols. 10, 11) occur within the region but are less abundant than the pink, hornblende (-biotite) granitic gneiss. Occasional orthopyroxenes occur in the hornblende (-biotite) granitic gneiss, and, strictly speaking, these occurrences result in the local development of small patches of charnockite. However, most mappable bodies of charnockite have whole-rock chemical compositions that differ from those of the granitic gneisses (deWaard, 1969), especially with regard to SiO_2 content, which is consistently lower (<68 percent) in the charnockitic rocks (table 1, col. 10). Mangeritic gneiss has even lower SiO_2 contents that range from 55 to 60 percent. In almost all instances, the charnockitic and mangeritic rocks that have less than 68 percent SiO_2 exhibit the olive-green color common to orthopyroxene-bearing granulite facies rocks. Although these rocks crop out in all parts of the map area, they are most commonly developed near bodies of anorthosite that they tend to envelop. Near anorthosite-mangerite contacts, a distinctive rock occurs that appears to be a mixture of these two end members. This lithology, named Keene Gneiss by Miller (1919), is characterized by the presence of numerous andesine xenocrysts set in a mangeritic matrix and grades from a few percent xenocrysts to anorthosite with interstitial mangerite. In places, the whole-rock chemistry resembles a mixture of mangeritic and anorthositic components. A whole-rock analysis of Keene Gneiss is given in table 1, col. 12.

A structurally important body of garnetiferous charnockitic gneiss occurs northwest of Whitehall, N.Y. (fig. 1, WH), where it forms the limbs of a folded thrust sheet. The eastern portion of this thrust sheet was first mapped by Berry (1965) and later extended to the west by Turner (1980). Although the dominant rock type is charnockite, ferrogabbros containing xenoliths of anorthosite are common near the bottom of the body. Whole-rock analyses of these ferrogabbros (table 1, col. 13) show that they closely resemble late differentiates of the anorthositic suite of rocks (Buddington, 1939). The entire sheet is thought to consist of rocks coeval with the emplacement of anorthosite.

Throughout the region there occur extensive outcroppings of pink to white, magnetite-bearing alaskitic gneiss superficially similar in appearance to the pink, hornblende (-biotite) granitic gneisses. The whole-rock chemistry of the alaskitic gneisses is given in table 1, col. 14. Their SiO_2 contents (72–75 percent) are notably higher than those of the hornblende (-biotite) granites, and, as indicated in the geochronology section, the two rocks are of different ages. Quartz-sillimanite nodules typically occur within the alaskites and may represent metamorphosed quartz-kaolinite masses that resulted from hydrothermal alteration of original igneous rocks. Associated with the alaskitic gneisses of the eastern Adirondacks are stratabound, low-titanium magnetite deposits (Newland, 1908). These deposits commonly occur within albite-rich trondhjemitic layers in the alaskite. The stratabound nature of these deposits, as well as their interlayering with andradite-ferrosalite skarns and marbles, suggests that the host rock was originally sedimentary. However, it is still unclear whether the mineralization was related to exhalative fluids, as proposed by Frondel and Baum (1974) for the Sterling Hill, N.J., zinc-manganese-iron oxide deposits, or was the result of skarn formation, as described by Eugster and Chou (1979) for the iron-copper deposits at Cornwall, Pa. Some of the magnetite-apatite layers in the Adirondacks may also represent iron-oxide-rich magmas similar to those described by Panno and Hood (1983) in the St. Francis Mountains, Mo.

Metasedimentary Rocks

The metasedimentary gneisses[2] of the map area include all the rock types known to occur elsewhere in the Adirondacks (for example, quartzites, marble and calc-silicate, sillimanite-biotite-garnet-perthite-quartz gneiss (khondalites), and biotite-garnet-quartz-oligoclase± sillimanite gneiss (kinzigites)). The khondalites exhibit a bulk chemistry similar to that of Proterozoic shales, while the kinzigites resemble mixtures of graywacke and shale (McLelland and Husain, 1986). Kinzigites and khondalites commonly grade into one another along strike.

Although Alling (1917) and Walton and deWaard (1963) proposed stratigraphic sections for the eastern Adirondacks, it appears that these lithic packages cannot be traced for distances exceeding 10 to 20 km. In part this is because intrusive quartzofeldspathic rocks have been included within these sections and their presence or absence from any section significantly alters the interpretation of the package. Inclusion of these lithologies within "stratigraphic" sections of the Adirondacks reflects historic uncertainty concerning the origin of the quartzofeldspathic rocks (deWaard, 1969), particularly their ubiquitous conformity with enveloping gneisses. McLelland (1984) demonstrated that the extreme ductile strain that occurred within the Adirondacks led to the production of layered, and apparently conformable, gneisses from original plutonic, igneous rocks. Therefore, the apparently conformable contacts of most quartzofeldspathic layers cannot be taken as evidence for "metastratification." Moreover, the geochronologic evidence presented in the section on geochronology makes it clear that several contacts once believed to be depositional (for example, between the charnockite-anorthosite "basement complex" and the Paradox Lake Formation (Walton and deWaard, 1963)), must instead be intrusive.

As is clear from the predominance of igneous rocks in figure 1, the deletion of such lithologies from any stratigraphic section of the eastern Adirondacks greatly reduces the thickness and number of candidates for that stratigraphy. The situation is further complicated by the difficulty of following Adirondack metasedimentary units along strike for large distances. Field experience within the map area by the author indicates that structural repetition, or omission, coupled with original facies changes, leads to uncertainty of correlations that becomes increasingly great as the distance of extrapolation is extended. As a result, it is generally possible to conclude only that the regional metasedimentary rocks consist of quartzites, marbles, and metapelites that vary in quantity so that any one of the three may be locally dominant. Exceptions to this do exist elsewhere in the region; in the southern Adirondacks McLelland (1979) was able to follow a thick (~1,500 m) orthoquartzite for ~100 km along strike. Unfortunately such diagnostic and exten-

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[2] Mineral modifiers are used here in increasing order of abundance.

sive units are rare, and, in their absence, it is virtually impossible to establish true or reliable stratigraphic sections in these intensely deformed and highly metamorphosed rocks.

Although Adirondack stratigraphy is not well understood, it is important to note that lithotectonic sequences can be locally defined and used to map out regional structural configurations. This conclusion is implicit within the map pattern of figure 1, on which structures of various scales are clearly visible. However, such lithotectonic sequences (McLelland and Isachsen, 1986) include thrust sheets, igneous plutons, and previously deformed and dismembered metasedimentary sequences. Therefore, these sequences should not be treated, or interpreted, as stratigraphic sections.

REGIONAL STRUCTURAL FRAMEWORK

The southeastern Adirondacks are dominated by large fold structures similar to those described elsewhere in the region (McLelland, 1979, 1984; McLelland and Isachsen, 1986). In addition, the study area contains the only well-documented thrust fault thus far recognized within the Adirondack highlands.

As shown in figure 1, the structural grain within the northern and southern parts of the map area is dominated by east-west trends; north-south trends characterize the central region. The east-west structural grain is common throughout the southern and central Adirondack highlands, while the north-south trends on figure 1 are due to a deflection of foliation around the eastern margin of the Oregon dome anorthosite massif (McLelland and Isachsen, 1986) that is situated just west of the map area.

Three principal sets of folds underlie the region. The oldest of these, F_1, consists of gently plunging recumbent, isoclinal folds that fold both compositional layering and a tectonic foliation of unknown origin. This older foliation, which is defined by flat, oriented platelets of metamorphic mineral grains and assemblages, is not associated with any folds yet recognized in the area. A foliation axial planar to F_1 does exist but is generally less pronounced than the pre-F_1 foliation. Most foliation surfaces seen in the field probably consist of reoriented pre-F_1 fabrics. Mineral and ribbon lineations, as well as rodding, are commonly found oriented parallel to F_1 fold axes. The best and largest examples of F_1 folds in the map area are the Canada Lake isocline (fig. 1, northwest of Saratoga Springs, N.Y. (SS)) and the Fort Ann isocline (fig. 1, north of FA). Other F_1 folds are developed at Grasshopper Hill (fig. 1, G) and north of the Paleozoic inlier near Warrensburg (fig. 1, WA).

The F_1 isoclinal folds are themselves folded by open, upright folds of the F_2 set. These east-west-trending structures are generally large and can be traced for long distances. Their gently plunging axes are typically parallel to F_1 fold axes so that the two sets are approximately colinear and the orientation of linear features coincides with both sets of axes. Folds of the F_2 set rarely develop axial planar foliation. The best examples of these folds include the Piseco antiform (fig. 1, PA), the Moose Mountain-Fort Ann antiform (fig. 1, MM-FA), the Glens Falls synform (fig. 1, GF), the Whitehall synform (fig. 1, WH), and the Pharoah Mountain antiform (fig. 1, P).

A north-northeast-trending set of open, upright folds, F_3, results in dome and basin interference patterns where they intersect older folds. Examples of such interference patterns are developed in the Piseco antiform (fig. 1, PA) or at Crane Mountain (fig. 1, 5 km north of MM). In general, these folds do not develop axial planar foliations or linear fabrics.

McLelland (1984) proposed that the coaxiality of the F_1 and F_2 folds is due to the arcuation of early F_1 axes into ~east-west-trending sheath folds of major dimensions. The orienting mechanism is believed to be an east-west rotational ductile strain related to the emplacement of thrust slices of unknown magnitude and displacement. Ribbon lineations in the direction of tectonic transport formed in response to stretching parallel to the maximum finite extensional strain. F_2 folds are believed to be the result of constrictional strain related to the regional flow field during thrusting and sheath fold formation (Henderson, 1983), although some F_2 folds may be simply the upper and lower surfaces of subhorizontal sheath folds. Kinematic indicators, such as feldspar tails, suggest that the tectonic transport associated with these processes was dominated by an east-side-up-and-to-the-west sense of displacement (McLelland, 1984).

In the Adirondack highlands, the only clearly defined example of a thrust fault occurs within the map area and is located northwest of Whitehall. The thrust has been folded and is preserved within the large F_2 Whitehall synform (fig. 1, WH). The upper plate of the thrust consists predominantly of garnetiferous charnockite, although garnetiferous ferrogabbro occurs locally. Near the sole of the fault, strong east-west-trending rodding and ribbon lineation are developed in both the upper and lower plate. These features are accompanied by extensive ductile grain-size reduction, resulting in mylonitic rocks and a fine-grained laminar fabric in quartzofeldspathic lithologies. Detailed mapping by the author has demonstrated that the upper plate truncates lithic units and structures, particularly F_2 folds, in the lower plate. At the same time, the upper plate has been deformed by open, upright folds that have east-west axes and are therefore assigned to the F_2 generation. If this structural assignment is correct, then the relations suggest that F_2 folding overlapped in time with thrusting.

GEOCHRONOLOGY

McLelland and others (1988a) report 22 new U-Pb zircon ages for samples collected throughout the Adirondacks. Several of these samples are from the map area of figure 1 and have important implications for local and regional geology.

Two critically important ages have been obtained from samples of olivine metagabbro. The first sample is from a roadcut exposure along the easternmost margin of the Adirondacks immediately west of Dresden Station (fig. 1, DS). This metagabbro, which has been isoclinally folded and contains granulite facies coronas (McLelland and others, 1988a), yields a well-constrained U-Pb zircon age of 1144±7 Ma, which is interpreted as its emplacement age. Embedded within the metagabbro is a xenolith of strongly foliated khondalite whose fabric and mineral assemblages clearly predate the metagabbro and document the existence of an earlier high-grade metamorphism. It is the presence of this metamorphism, together with the later granulite facies event, that has made it so difficult to establish the original sedimentary sequence in the region.

The Dresden Station metagabbro has a composition much like that of other Adirondack metagabbros, and it seems reasonable to suppose that all of these rocks were emplaced at ~1140 Ma. To verify this age, a second metagabbro was collected from North Hudson (fig. 1, NH), near Schroon Lake, where it clearly intrudes the Marcy massif anorthosite. Equidimensional, multifaceted zircons, interpreted as metamorphic (Silver, 1969; McLelland and others, 1988a), are present in this sample and give an age of 1052±4 Ma. This age agrees well with the sphene and monazite cooling ages of 1030–950 Ma reported by Rawnsley and others (1987), as well as with ~1050-Ma ages determined for other metamorphic zircons from the Adirondacks (McLelland and others, 1988a). The North Hudson metagabbro also yielded a small quantity of baddeleyite that gave a ^{206}Pb/^{207}Pb age of 1110 Ma. Thus a minimum emplacement age for the anorthosite (>1110 Ma) is established, which places its intrusion prior to the ~1050-Ma granulite facies metamorphism. This result is consistent with an ~1140-Ma emplacement age for the metagabbro.

An upper age constraint for the anorthosite is derived from the previously described sheets of gabbroic anorthosite that intrude granitic, charnockitic, and mangeritic rocks along the southern margin of the Marcy massif and northwest of Warrensburg (fig. 1, WA). The charnockitic gneiss bordering the Marcy massif is part of a charnockite-granite complex that has been dated from samples collected from Whitney Point (fig. 1, WP) on the southwestern shore of Schroon Lake. Here, pink granitic rocks give a U-Pb zircon intercept age of 1125±10 Ma that is similar to the 1130- to 1150-Ma ages determined for mangerite-charnockite-granitic rocks elsewhere in the Adirondacks (McLelland and others, 1988a) as well as to the 1113±16-Ma age obtained by Silver (1969) for charnockites of the Ticonderoga dome (fig. 1, T). This date fixes a maximum age for the Marcy massif as well as for the Warrensburg sheet that intrudes hornblende granitic gneiss identical to, and correlated with, the Whitney Point sample. The foregoing results constrain the emplacement of the anorthosite to the interval 1110–1135 Ma. It is believed that emplacement occurred close to the upper part of this range and coeval with the intrusion of mangeritic and charnockitic rocks whose emplacement ages have been established at 1130–1160 Ma (McLelland and Chiarenzelli, 1990).

Zircons recovered from a metatonalitic gneiss exposed along New York State Highway 22 west of Whitehall (fig. 1, WH) were also dated by the U-Pb method. These zircons yielded an upper intercept age of 1329±36 Ma that corresponds closely to the ^{206}Pb/^{207}Pb ages of greater than 1301 Ma and greater than 1336 Ma determined for identical metatonalites from the southern Adirondacks (McLelland and others, 1988a). It therefore seems well established that these rocks represent an early calc-alkaline igneous event in the Adirondacks and may reflect a collisional event associated with the development of the early metamorphism and fabric. Note that xenoliths of metasedimentary rock occur within the metatonalite and clearly predate these intrusives.

Pink alaskitic gneiss superficially resembles pink, hornblende granitic gneiss and has often been interpreted as genetically linked with these rocks (McLelland, 1986). However, an alaskitic gneiss from Grasshopper Hill (fig. 1, G) gives a U-Pb zircon age of 1077±11 Ma, which makes it ~48 Ma younger than hornblende granitic gneiss dated at 1125±10 Ma. This younger age has been verified on five additional samples of alaskitic gneiss collected from across the Adirondack highlands (McLelland and others, 1988a). It therefore appears likely that these rocks, which previously had been interpreted as metavolcanics (McLelland, 1986), are actually metamorphosed leucogranites intruded during the ~1050-Ma granulite facies deformation and metamorphism. These rocks are presently anhydrous, and it appears they were intruded as hypersolvus granites. If this is so, then their emplacement temperatures would have been near 1,000 °C and would have contributed substantial quantities of heat to the ~1050-Ma metamorphism.

METAMORPHISM

The eastern Adirondack region is affected everywhere by a strong granulite facies metamorphism marked by temperatures of 650 to 800 °C and pressures of 7 to 8 kbar (Bohlen and others, 1985). These conditions are manifested by garnet coronas in metagabbroic and meta-anorthositic rocks (McLelland and Whitney, 1980), by garnet-sillimanite gneisses (Bohlen and others, 1983), and by a variety of calc-silicate reactions (Valley, 1985). Many charnockitic and mangeritic rocks frequently explained as metamorphic appear instead to involve inherited anhydrous igneous assemblages whose igneous orthopyroxene recrystallized during subsequent metamorphism (McLelland and others, 1988b). The granulite facies metamorphism is dated by the 1030- to 950-Ma sphene and monazite cooling ages reported by Rawnsley and others (1987), as well as by the ~1050-Ma metamorphic zircon ages in mafic igneous rocks (McLelland and others, 1988a). The orientation of high-grade minerals (for example, sillimanite) indicates that the formation of F_1 and F_2 folds accompanied metamorphism (McLelland and Isachsen, 1986).

It now appears certain that the Adirondacks have a polymetamorphic history. As originally proposed by Buddington (1939) and then by Valley and O'Neil (1982), the emplacement of anorthositic, mangeritic, and charnockitic magmas resulted in contact metamorphism throughout the Adirondacks. Uranium-lead zircon dating (McLelland and others, 1988a) indicates that emplacement took place during 1150–1130 Ma and prior to the granulite facies event dated at ~1050 Ma. McLelland and Husain (1986) have described anatexis of metapelites associated with this event, including mylonitic sillimanite-garnet-biotite-quartz-oligoclase gneisses with leucocratic quartzose two-feldspar interlayers exposed in roadcuts along Route 22 north of Fort Ann (fig. 1, FA). Other examples can be found in the corundum-spinel assemblages described from a xenolith within olivine metagabbro (McLelland and others, 1988c) near Dresden Station (fig. 1, DS), as well as from sillimanite-rich metapelites at the contact of the Buck Mountain anorthosite on the southeastern shore of Lake George, directly east of the word LAKE on figure 1.

In addition to pregranulite facies contact metamorphism, McLelland and others (1988c) have presented evidence for an early high-grade metamorphism that produced a strong fabric defined by oriented sillimanite and deformed garnets within the Dresden Station olivine metagabbro (fig. 1, DS). Uranium-lead zircon dating of the host rock yields an emplacement age of 1144±7 Ma, thus placing the fabric-forming event well before the ~1050-Ma granulite facies metamorphism. This earlier, high-grade metamorphism is believed to be responsible for the pre-F_1 fabric described previously.

The existence of two pregranulite facies metamorphic events further complicates an already complex regional history. Caution must be exercised in interpreting field and analytical data that might otherwise be attributed to a single pulse of dynamothermal metamorphism. Although anhydrous metamorphic assemblages of the region are characteristic of granulite facies conditions, it should be recognized that they may have been inherited from earlier high-grade regional or contact metamorphism, as well as from the intrusion of anhydrous igneous suites.

RELATION TO THE ADIRONDACKS AS A WHOLE

The structural geology and metamorphic grade of the region shown in figure 1 are similar to those encountered throughout most of the Adirondack highlands. However, the lithology of much of the area, especially the southern third and the region east of Lake George, closely resembles the southern Adirondacks exposed south of the Piseco antiform (McLelland and Isachsen, 1986). Together, the southern and eastern Adirondacks contain far more kinzigite and less anorthositic rock than is found in the central Adirondacks. Furthermore, metatonalites appear to be restricted to the southern and eastern Adirondacks. This configuration suggests that the southern and eastern Adirondack highlands may represent an older, already metamorphosed, sequence of rocks that was subsequently intruded by the anorthosite-mangerite-charnockite suite. Erosion has now removed this original carapace from the central part of the Adirondack dome, thus exposing the younger intrusive complex.

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Chapter C

Magnetic and Gravity Expression of Cretaceous Alkalic Plutonic Complexes at Cuttingsville and Mount Ascutney, Vermont

By DAVID L. DANIELS

U.S. GEOLOGICAL SURVEY BULLETIN 1887

SUMMARY RESULTS OF THE GLENS FALLS CUSMAP PROJECT, NEW YORK, VERMONT, AND NEW HAMPSHIRE

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Magnetic and Gravity Expression of Cretaceous Alkalic Plutonic Complexes at Cuttingsville and Mount Ascutney, Vermont

By David L. Daniels[1]

Abstract

Prominent gravity and aeromagnetic anomalies are associated with Cretaceous alkalic plutonic complexes in central Vermont. An oval negative gravity anomaly coincides with the syenite-granite body at Mount Ascutney. A model, constructed to match the gravity field, has a funnel-like cross section and a depth extent of about 4 kilometers. In contrast, a platelike cross section is inferred for the associated gabbro-diorite pluton at Little Ascutney Mountain immediately to the west. The magnetic data indicate that the southeastern contact of the gabbro-diorite must dip outward at a low angle before steepening to near vertical at about 1 kilometer from the mapped contact. The low dip suggests that the present erosion level is close to the average position of the roof of the mafic stock. A similar negative residual gravity anomaly coincides with the alkalic rocks at Cuttingsville but is broader than the surface exposure of the complex. The depth extent at Cuttingsville may be similar to or greater than that for the syenite and granite at Mount Ascutney. The gravity anomaly minimum at Cuttingsville is associated with a gold-bearing, pyrite-rich zone of hydrothermal alteration in the syenite; localization of the minimum may be due to the lower densities of the altered rocks. An aeromagnetic anomaly at Shrewsbury, north of Cuttingsville, encloses six lenticular bodies of alkalic igneous intrusive breccia, porphyry, and plutonic rocks that are probably related to the emplacement of the complex at Cuttingsville.

INTRODUCTION

Several occurrences of the White Mountain intrusive series (Foland and Faul, 1977) are found in the Glens Falls 1° × 2° quadrangle (fig. 1). The largest of these occurrences are the Cretaceous alkalic plutonic complexes at Cuttingsville and Mount Ascutney, Vt. At both sites, the dominant lithology is syenite, but each complex shows clear evidence of multiple intrusions that have compositions ranging from alkalic gabbro to syenite. Radiometric dates of 96.4 and 100±2 Ma (Armstrong and Stump, 1971) for the complex at Cuttingsville and 122.2±1.2 Ma (Foland and others, 1985) for the complex at Mount Ascutney have been obtained. At Mount Ascutney the intrusive center apparently migrated about 6 km through time from west to east and changed in composition from gabbro and diorite to syenite to granite (Daly, 1903; Chapman and Chapman, 1940). The intrusive center at Cuttingsville, however, remained relatively fixed in location through several intrusive events.

Density and magnetic properties of the rocks of these complexes contrast strongly with those of the rocks they intrude and result in prominent potential-field anomalies. This report analyzes these anomalies, which have provided new information about the shape and size of the intrusive bodies. Limited in situ measurements of magnetic susceptibility, measurements of density on collected samples, and measurements of the approximate magnitude of remanent magnetization (13 samples using a portable specimen magnetometer, see table 1), have aided in the study of the anomalies. The intrusive rocks are important metallogenically and are the subject of mineral-resource studies (Cox, this volume; Robinson, this volume; Slack, this volume). The complex at Cuttingsville has a large hydrothermally altered gold-bearing pyritic stockwork zone (Robinson, this volume), which could have resource potential if significant volumes of mineralized rock are present.

The potential-field data in this analysis are mostly new. The magnetic data were obtained from an aeromagnetic survey flown by the U.S. Geological Survey in 1983–84 along east-west flight lines spaced 0.5 mi apart and at a nominal terrain clearance of 300 ft (Long, this volume). Measurements were recorded digitally approximately every 45 m along flight lines. The International Geomagnetic Reference Field was removed from the gridded aeromagnetic data (unpublished computer program IGRFDGRID, R.E. Sweeney, 1984, U.S. Geological Survey). The gravity data come from a survey of the Glens Falls 1° × 2° quadrangle carried out between 1982 and 1986 (Daniels, 1988) and from surveys of the Mount Ascutney area (Weston Geophysical Research, Inc., 1977) and the

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Manuscript approved for publication April 13, 1990.

[1] U.S. Geological Survey.

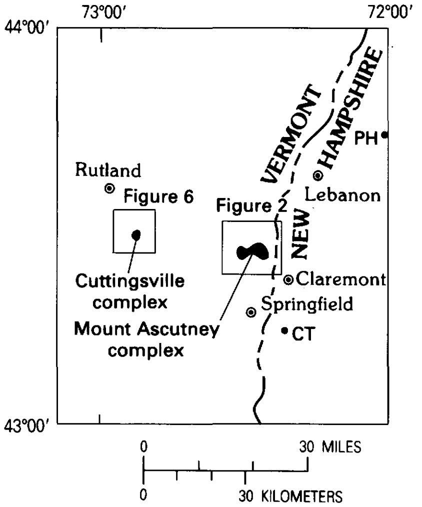

Figure 1. Index map of the eastern part of the Glens Falls 1° × 2° quadrangle showing the location of Cretaceous intrusive rocks (in black). PH, Pollard Hill area bodies (Chapman, 1939); CT, Charlestown body (Billings, 1955).

New England COCORP (Consortium for Continental Reflection Profiling) seismic traverse (W.A. Bothner, written commun., 1984). Standard formulas and a standard density of 2.67 g/cm^3 were used to reduce the gravity data. Within the study areas, the number of magnetic measurements exceeds the number of gravity measurements by a factor of about 100.

Because the syenite-granite pluton of Mount Ascutney stands well above the general terrain (>600 m), significant terrain corrections are needed for gravity stations on the mountain. Terrain corrections were calculated for all stations by using the program Bouguer (unpublished computer program based on the method of Plouff, 1977, R. Godson, U.S. Geological Survey) for zone radii from 0.895 to 166.7 km. This program calls files of digital terrain elevations, the most detailed of which have a spacing of 30 seconds of latitude and longitude. Inner-zone terrain corrections (radii of 0.053 to 0.895 km) were calculated for stations only on Mount Ascutney by using the method described by Hammer (1939). Total terrain corrections for these stations range up to 18 milligals (mGal).

In contrast to Mount Ascutney, the syenite complex at Cuttingsville does not stand above average terrain. Much of the pluton is a topographic low except for a semicircular ridge that follows the western and southern contacts. Inner-zone terrain corrections were calculated for 20 stations clustered around the complex. Total terrain corrections at Cuttingsville range up to 6.3 mGal.

COMPLEX AT MOUNT ASCUTNEY

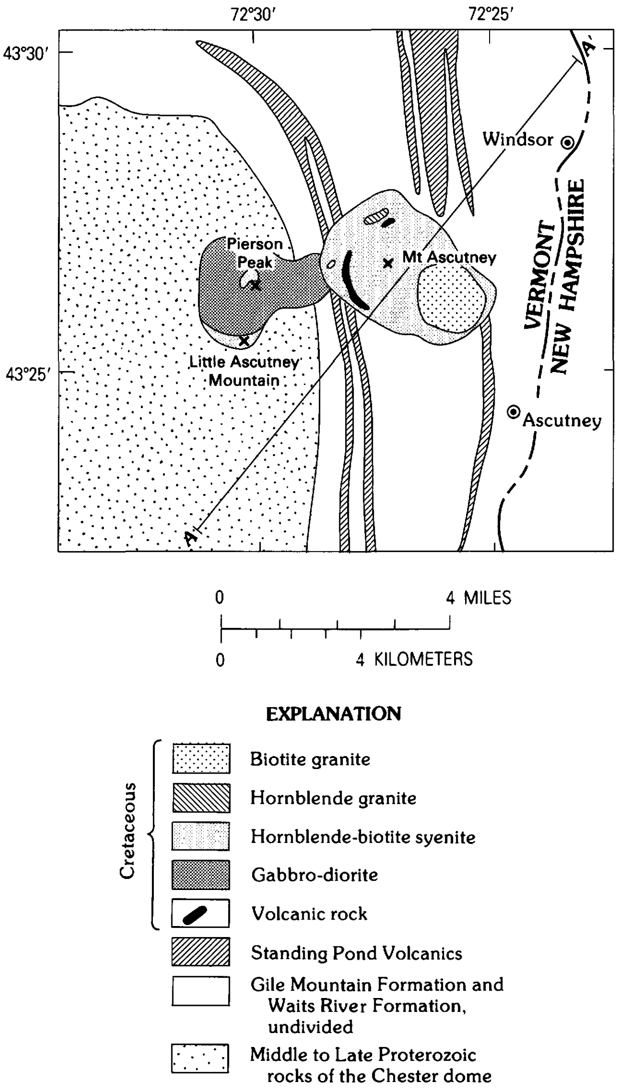

Figure 2. Geologic map of the Mount Ascutney, Vt., area (from Chapman and Chapman, 1940, and Doll and others, 1961). Profile *A–A'* modeled in figure 4.

The plutonic complex at Mount Ascutney (fig. 2) consists of (1) a western gourd-shaped gabbro-diorite stock that is cut by smaller syenite plugs at Pierson Peak and Little Ascutney Mountain and (2) an eastern oval body of syenite at Mount Ascutney that is cut by a circular plug of biotite granite (Daly, 1903; Chapman and Chapman, 1940). Geologic observations from both studies indicate that the outer walls of the syenite and granite bodies are nearly vertical at the surface. Contacts of the gabbro-diorite are not well exposed (Chapman and Chapman, 1940).

On the complete Bouguer gravity anomaly map (fig. 3), an oval, 6-mGal, closed, negative anomaly closely conforms to the geometry of the eastern syenite and granite and reflects the low density of these rocks relative to the country rocks of the Waits River and Gile Mountain Formations (table 1). By using a density contrast of 0.09 g/cm^3, a cross-sectional model of the pluton was con-

Table 1. Measurements of density and magnetic properties of Cretaceous plutonic rocks and country rocks in the Glens Falls 1° × 2° quadrangle

[Average density is weighted by lithologic proportions; number in parentheses indicates number of samples; D, Daly (1903); B, Bean (1953); N, new measurements (this study)]

	Density (g/cm^3)	Average density (g/cm^3)	Magnetic susceptibility (cgs units×10^{-3})	Approximate remanent magnetization (emu×10^{-3})	Sources of data
Mount Ascutney area					
Intrusive rocks		2.66			
Hornblende-biotite syenite	2.67 (4)		0.5 (150)	0.3 (3)	N
Granite	2.62 (1)				D
Gabbro-diorite	3.00[1] —				
Stratified metamorphic rocks		2.76			
Waits River Formation	2.76 (34)				B
Gile Mountain Formation	2.76 (39)				B
Standing Pond Volcanics	3.02 (6)				B
Chester dome gneiss	2.69 (7)				B
Bull Hill Gneiss	2.69 (2)		2.5 (8)		N
Cuttingsville area					
Intrusive rocks		2.69			
Syenite	2.66 (4)		2.3 (20)	0.5 (4)	N
Altered syenite	2.61 (4)		2.4 (2)		N
Hornblende-biotite syenite	2.68 (1)		1.0 (10)		N
Essexite	2.96 (5)		3.5 (6)	4.0 (3)	N
Middle Proterozoic rocks					
Green Mountain massif					
Mount Holly Complex		2.76			
Biotite gneiss	2.76 (4)		0.0 (4)		N
Amphibolite	3.08 (2)		.1 (2)		N
Quartzite	2.68 (1)		.0 (1)		N

[1] Estimate.

structed (fig. 4) that matches the complete Bouguer gravity anomaly data along a northeast-trending profile (*A–A′*, figs. 2, 3). A semiautomatic magnetic-gravity computer inversion program (program SAKI, Webring, 1985) was used to generate the model. The best fitting model shows an asymmetric funnel of syenite-granite that extends to a depth of about 4 km (fig. 4). In this model, the southwestern contact of the syenite and the Waits River Formation dips to the northeast in the upper kilometer of the pluton and then assumes a near-vertical dip. The maximum vertical extent of the felsic rocks occurs along the northeastern edge of the pluton. Weston Geophysical Research, Inc. (1977), calculated a depth extent of 6 km. The difference in depth extent for these two studies may be due to several factors; the Weston Geophysical study used a different modeling technique, a larger density contrast (0.3 g/cm^3), and smaller terrain corrections. Mafic rocks of the Standing Pond Volcanics (fig. 2) apparently are the primary source of the low-amplitude gravity highs on either side of the pluton (fig. 4*A*), even though the volcanics are not continuous along the eastern side. Because the strike of the Standing Pond Volcanics is not perpendicular to the profile, the representation of these rocks in the model is schematic only. The high shoulder on the southwestern side of the low also may be partly due to the attraction of the gabbro-diorite pluton (fig. 2), which is not intersected by the profile and not represented in the model. The mismatch between the observed and calculated curves on the southwestern end of the profile is probably due to the low density gneisses of the Chester dome, the effects of which were partly removed with the regional gradient.

A smaller 2- to 3-mGal positive anomaly is associated with the gabbro-diorite intrusive to the west (fig. 3). A much larger anomaly would be expected if the structure were similar to that of the syenite-granite pluton, given the probable large density contrast between the mafic intrusive rocks and the country rocks of the Chester dome (3.00 and 2.69 g/cm^3, respectively, table 1). No stations are located in the center of this body where the Bouguer gravity may reach a maximum. On the basis of the existing station distribution, the small amplitude of the anomaly indicates that the volume of mafic rock is relatively minor; a platelike shape is inferred for the intrusive. The syenite, which is exposed at Pierson Peak and Little Ascutney Mountain (fig. 2), may produce only a small negative anomaly because the density is only slightly less than the densities of the Chester dome rocks. Therefore the amount of subsurface syenite cannot be distinguished by analysis of gravity data alone.

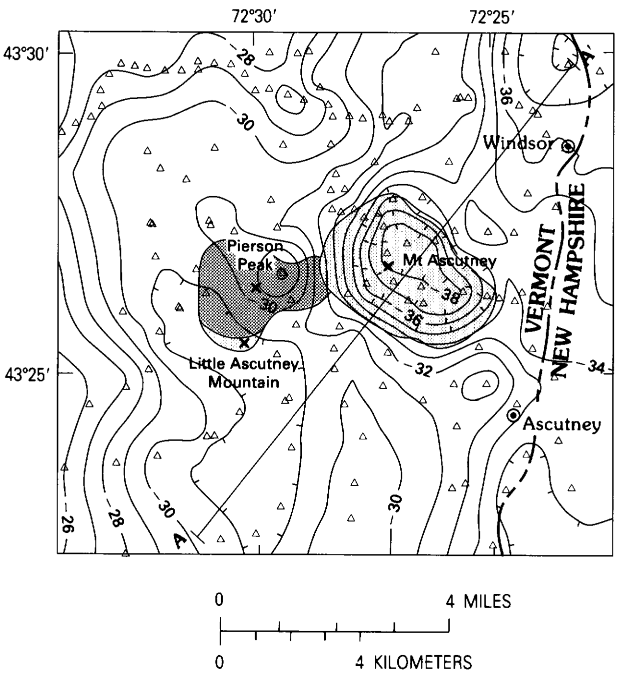

Figure 3. Complete Bouguer gravity anomaly map of the Mount Ascutney, Vt., area. Contour interval, 1 mGal. Gravity station locations shown by open triangles. Geology generalized from figure 2. Profile *A–A′* modeled in figure 4.

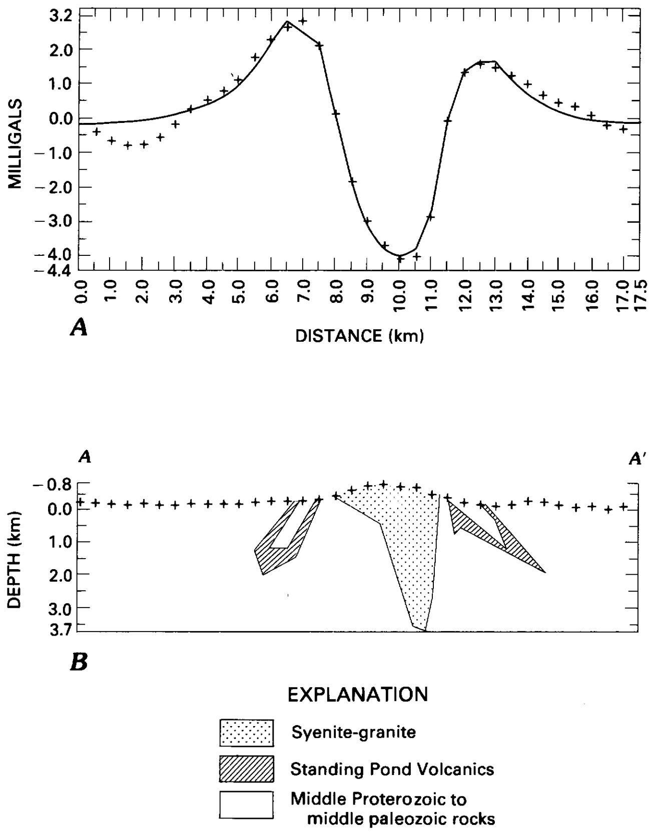

Figure 4. Model of the syenite-granite pluton at Mount Ascutney, Vt., along line *A–A′* shown on figures 2 and 3. *A*, Observed gravity (plus symbols) and calculated gravity from the model (solid line). *B*, Cross section of the model. The plus symbols show surface altitude. Density contrast of syenite-granite, -0.09 g/cm^3; density contrast of the Standing Pond Volcanics (fig. 2), $+0.2$ g/cm^3. Trend removed from the observed gravity profile has an offset of -20.9 mGal and a slope of -0.45 mGal/km. The length of the syenite-granite body perpendicular to the cross section is 2.5 km.

Both the western gabbro-diorite and the eastern granite plutons produce prominent magnetic anomalies (fig. 5). However, unlike the gravity field anomaly, the western magnetic anomaly has much greater amplitude. The intense aeromagnetic anomaly associated with the gabbro-diorite has gradient maxima that closely conform to the contacts as mapped by Chapman and Chapman (1940), except in the southeastern quadrant. Whereas the surface outline of the pluton is gourdlike and has an eastward-projecting neck, the magnetic data suggest that the subsurface outline is more rectangular. In the southeastern quadrant, the maximum magnetic gradient lies up to a kilometer beyond the contact of the pluton with the country rocks. This configuration indicates that the contact dips away from the center of the body at a low angle before steepening to near vertical in the subsurface at the position of the gradient. This geometry and the lesser altitude of the exposed gabbro-diorite along the eastern neck suggest that the eastern part of the gabbro-diorite body may be a separate intrusive center that has a roof at a lower average level than the western part. On the basis of an observed dual concentric distribution of planar structures in the mafic rocks, Chapman and Chapman (1940) postulated dual intrusive centers. A magnetic low to the north that partially wraps around the western side of the mafic rocks (fig. 5), as shown by the bend in the zero contour, indicates a limited depth extent and supports the interpretation of a platelike shape for the mafic rocks inferred from the gravity data (Andreasen and Zietz, 1969, pl. 111*A*).

Although the hornblende-biotite syenite from the summit area of Mount Ascutney has a modest but consistent magnetic susceptibility (table 1), the aeromagnetic intensity of the eastern plutons is at a maximum over the biotite granite at the southern end. This is the normal position of the maximum for a uniformly magnetized body that has a depth extent greater than its width and is magnetized by induction (Andreasen and Zietz, 1969, pl. 195*A*, inclination 75°). The bulk magnetite content of the granite, therefore, is inferred to be equal to or greater than that of the syenite.

Griscom and Bromery (1968), using magnetic data from an earlier airborne survey (Meuschke and others, 1962), noted the existence of a broad magnetic anomaly associated with the syenite-granite pluton. They interpreted this anomaly to be indicative of a stock that has steeply dipping contacts for a few thousand feet but then opens out

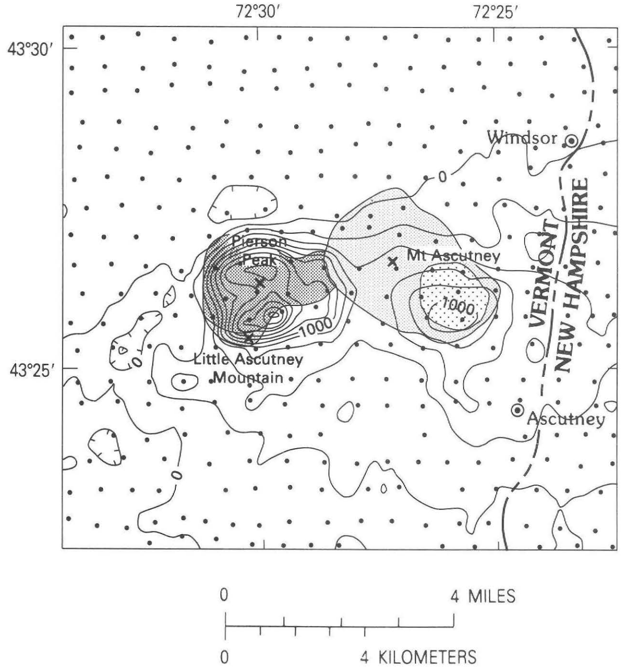

Figure 5. Residual aeromagnetic map of the Mount Ascutney, Vt., area. Contour interval, 200 nT. Flight paths shown by dots. Geology generalized from figure 2.

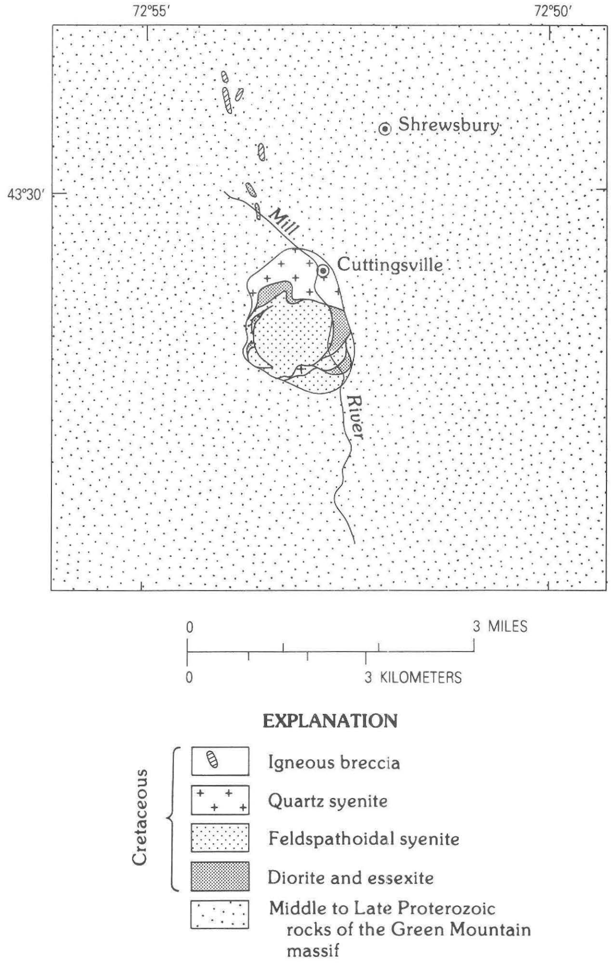

Figure 6. Geologic map of the Cuttingsville, Vt., area (from Brace, 1953, and Laurent and Pierson, 1973).

to a body three times the surface width. This interpretation would be consistent with the gravity data obtained during the present study only if the source were not syenite or granite, because these lithologies would produce a broad gravity low, which is not observed. However, a separate intrusive in the lower part of the complex composed of rocks having a density of about 2.76 g/cm^3 (equal to the average density of the surrounding Waits River and Gile Mountain Formations) would not have a gravity anomaly and would be consistent with the gravity field. The density ranges of both granodiorite and diorite meet this requirement (Telford and others, 1976), but diorite is a more likely lithology because granodiorite is absent in the surface rocks of the complex.

COMPLEX AT CUTTINGSVILLE

The intrusive rocks of the complex at Cuttingsville (fig. 6) form an oval composite pluton, which consists predominantly of quartz syenite and several varieties of feldspathoidal syenite, mostly biotite foyaite (Eggleston, 1918; Laurent and Pierson, 1973). Unlike the intrusive complex at Mount Ascutney where mafic rocks form a separate pluton, mafic rocks at Cuttingsville are dispersed in the largely felsic complex. The mafic rocks, mostly essexite and diorite that compose about 15 percent of the body, should make the average density of the complex greater than the density of the syenite-granite pluton at Mount Ascutney. Yet a negative oval gravity anomaly of similar amplitude coincides closely with the complex at Cuttingsville. The negative anomaly lies on the western flank of a large-amplitude, long-wavelength positive gravity anomaly coincident with the Green Mountain massif but probably not derived from the surface rocks (Bean, 1953; Diment, 1968). To reduce the dominance of the broad gravity high and isolate the local negative anomaly, the long wavelengths were removed by filtering the gridded gravity data (unpublished matched filtering computer program MFILT, J.D. Phillips, U.S. Geological Survey). The resulting high-pass gravity map (fig. 7) shows a 5- to 6-mGal closed low associated with the complex but centered on the eastern side.

Although the gravity anomalies at Cuttingsville and Mount Ascutney have similar amplitudes, the amplitude of the Cuttingsville anomaly should be smaller due to the smaller diameter of the complex and the density contrast between the complex and the country rocks. Based on the measurements given in table 1 and on the proportions of felsic and mafic rocks observed at the surface (Laurent and

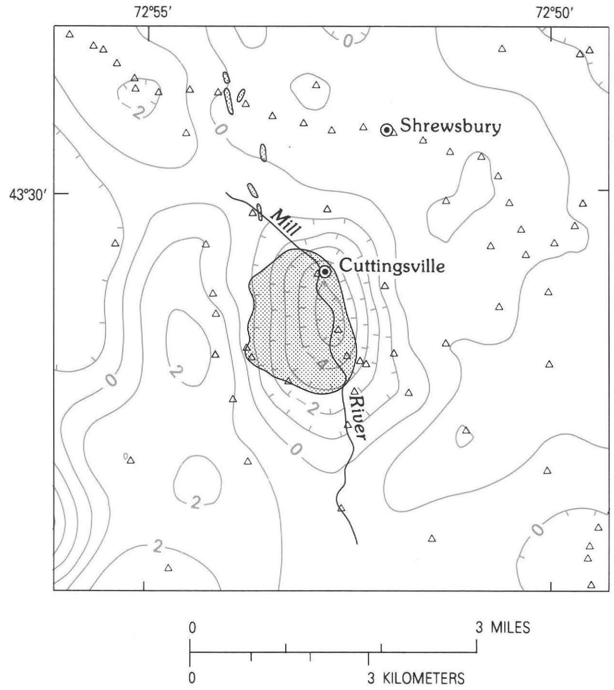

Figure 7. High-pass filtered gravity anomaly map of the Cuttingsville, Vt., area. Contour interval, 1 mGal. Gravity station locations shown by open triangles. The pattern marks area of outcrop of Cretaceous igneous rocks (geology generalized from figure 6). Hachured contour lines indicate a relative minimum.

Pierson, 1973), the average density of the complex, excluding the effects of hydrothermal alteration, is greater, about 2.69 g/cm^3. The Proterozoic Mount Holly Complex, which is intruded by the complex at Cuttingsville, is described by Doll and others (1961) to be mainly biotite gneiss, locally granitoid, and lesser amounts of amphibolite, mica schist, quartzite, and calc-silicate rocks. The principal lithologies in the Mount Holly Complex in the Rutland area, as described by Brace (1953), are massive biotite microcline gneiss and layered plagioclase garnet gneiss; the latter is the dominant lithology on the northern side of the complex at Cuttingsville. The average density of the basement in the vicinity of the alkalic complex depends on the proportions of these different lithologies. The series of highs and lows in the area of the basement gneiss shown on the residual gravity map (fig. 7) reflects the local variations in density. Four biotite gneisses from the Mount Holly Complex have an average density of 2.76 g/cm^3 (table 1).

Modeling of the intrusive rocks at Cuttingsville (program SAKI, Webring, 1985) shows that the observed residual gravity anomaly cannot be matched by using the measured densities (table 1), regardless of the depth extent. By using a short prismatic body that has an outward-dipping eastern contact, a 5-km-depth extent, and densities of 2.69 g/cm^3 for the igneous complex and 2.76 g/cm^3 for the Mount Holly Complex, only two-thirds of the amplitude of the observed anomaly is produced. However, with the addition of a low-density body (2.61 g/cm^3) to the uppermost part of the body representing the igneous complex, a reasonable match to the observed gravity anomaly is obtained. This low-density body may represent the hydrothermally altered syenite observed at the surface, which is associated with a gold-bearing pyritic stockwork, located about 0.1 km south of the town of Cuttingsville (Robinson, this volume). The altered syenites have the lowest density of all of the measured samples (table 1) and lie within the anomaly minimum. The volume of the low-density body in the model is about 3 km^3, and if this is a reasonable volume for the hydrothermally altered zone, a mineral resource potential may be indicated (see Robinson, this volume, and Slack, this volume). Two alternate possibilities must be considered, however. First, the observed gravity also can be matched if the entire igneous complex is assigned the low density of 2.61 g/cm^3 and has a 4- to 5-km depth extent; however, this is the less likely possibility, as altered rocks are observed at only one location. Second, the average density of the Mount Holly Complex may be underestimated as a result of undersampling and an abundance of amphibolite and other high-density lithologies at shallow depths. More precise modeling would be productive only if there were more closely spaced gravity stations and density measurements.

Two intense positive aeromagnetic anomalies occur in the Cuttingsville area (fig. 8). The larger anomaly coincides with the intrusive complex at Cuttingsville, and the smaller anomaly encloses a cluster of alkalic porphyry and igneous breccia pipes. Because of the higher average magnetic susceptibility of the syenite at Cuttingsville (table 1), the amplitude of the Cuttingsville magnetic anomaly is greater than that at Mount Ascutney. Syenite is probably the main magnetic source for the Cuttingsville anomaly. However, diorite and essexite, present in subordinate amounts in surface outcrops (Laurent and Pierson, 1973), may contribute significantly to the anomaly; magnetic susceptibility and remanent magnetization for these rocks are about 50 percent higher than for the syenite. Separate anomalies for the mafic phases cannot be resolved in the aeromagnetic data. As at Mount Ascutney, maximum aeromagnetic intensities are found over the southern part of the plutonic complex, again the normal position if the rocks are magnetized largely by induction (Andreasen and Zietz, 1969, pl. 195*A*). On the basis of models given in Andreasen and Zietz (1969, pl. 195*A*), the gradually decreasing magnetic intensity radially away from the contacts suggests a depth extent that exceeds that of the syenite-granite at Mount Ascutney. The maximum magnetic gradient at Cuttingsville coincides exactly with the western contact of the complex but departs to the south and lies 0.6 km outside the contact on the eastern side, indicating a progressive decrease in dip of the contact from south to east. The residual gravity anomaly (fig. 7)

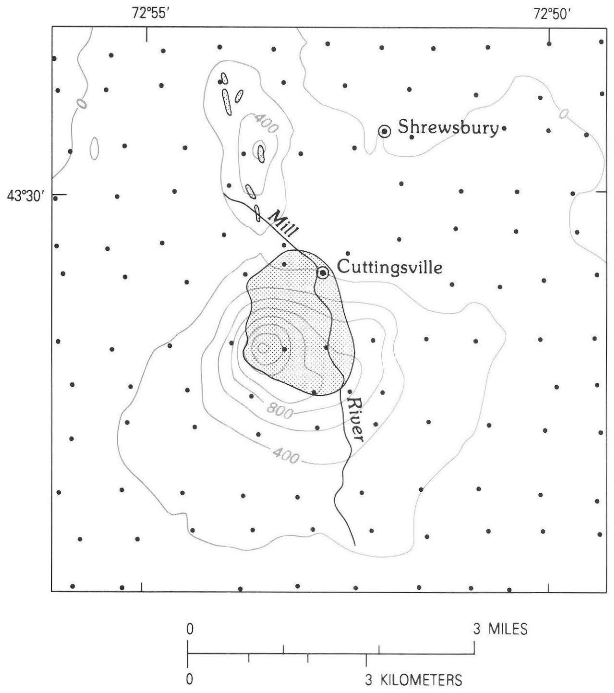

Figure 8. Residual aeromagnetic map of the Cuttingsville, Vt., area. Contour interval, 200 nT. Flight paths are shown by dots. The pattern marks area of outcrop of Cretaceous igneous rocks (geology generalized from figure 6).

follows a similar pattern and thus implies that the complex has a stocklike shape and is broader than shown by the surface geology.

SHREWSBURY BRECCIA PIPES

The second magnetic anomaly, which occurs north of Cuttingsville about 3 km west of the town of Shrewsbury, Vt. (fig. 8), encloses a group of at least six lenticular pipelike igneous bodies. These bodies are composed of intrusive breccia that contains abundant xenoliths of Proterozoic metamorphic rocks (Doss, 1986), trachyte-andesite porphyry, syenite, and essexite that are related to the emplacement of the Cuttingsville stock (Eggleston, 1918; Brace, 1953). One of the bodies, central to the group and directly beneath the crest of the aeromagnetic anomaly, was discovered during field investigations prompted by the results of the aeromagnetic data (G.R. Robinson, Jr., oral commun., 1986). Magnetic susceptibility measurements on samples of intrusive breccia collected from these outcrops yield values too low to produce the anomaly. Yet other samples of breccia in the vicinity, presumed to be float glacially transported from local outcrops, give high values with indications of strong remanent magnetization. The clear correlation between the group of lenticular bodies and the magnetic anomaly indicates that one or more of the outcrops have magnetic properties similar to those of the float samples and are, therefore, the source rocks. Modeling and magnetic depth calculations indicate that the source of the magnetic anomaly lies very near the surface. The existence of a subsurface body of syenite similar in size to the intrusive complex at Cuttingsville, which would have been a common magmatic source for each of the pipes, seems to be ruled out by the absence of a negative gravity anomaly (fig. 7). A small gravity anomaly centered on the magnetic anomaly, however, would not be detected with the existing station placement.

SUMMARY

Potential-field anomalies produced by Cretaceous alkalic igneous intrusive complexes at Mount Ascutney and Cuttingsville, Vt., suggest that the syenitic intrusives have a depth extent of at least 4 to 5 km, whereas the gabbro-diorite at Little Ascutney Mountain and Pierson Peak has a shallow root. These data indicate that the shape of the complex at Cuttingsville is stocklike and has vertical or outward-dipping contacts. The syenite-granite body at Mount Ascutney appears to taper downward in the form of an elongated funnel of 4-km-depth extent. The gabbro-diorite pluton at Pierson Peak and Little Ascutney Mountain has a platelike cross section. Its surface outline has a narrow eastern neck that expands in the shallow subsurface to a rectangular shape. It may consist of two intrusive centers, the eastern center having the lower roof.

Low-density altered syenite from a pyrite-rich hydrothermal stockwork zone coincides with a gravity minimum of the intrusive complex at Cuttingsville and may indicate a substantial volume of altered rock and related mineral potential. Six lenticular alkalic igneous bodies north of Cuttingsville and west of Shrewsbury produce a single aeromagnetic anomaly.

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Chapter D

Zircon U-Pb Data for the Moretown and Barnard Volcanic Members of the Missisquoi Formation and a Dike Cutting the Standing Pond Volcanics, Southeastern Vermont

By JOHN N. ALEINIKOFF and PAUL KARABINOS

U.S. GEOLOGICAL SURVEY BULLETIN 1887

SUMMARY RESULTS OF THE GLENS FALLS CUSMAP PROJECT, NEW YORK, VERMONT, AND NEW HAMPSHIRE

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Zircon U-Pb Data for the Moretown and Barnard Volcanic Members of the Missisquoi Formation and a Dike Cutting the Standing Pond Volcanics, Southeastern Vermont

By John N. Aleinikoff[1] *and* Paul Karabinos[2]

Abstract

We have attempted to date, by the U-Pb method, zircons from four metavolcanic rocks and one metasedimentary rock from the Paleozoic sequence of southeastern Vermont. The Barnard Volcanic Member of the Missisquoi Formation (previously correlated with the Middle to Late Ordovician Ammonoosuc Volcanics of New Hampshire) is a bimodal suite of felsic gneiss and amphibolite that have chemical compositions of felsic and mafic volcanic rocks. A sample of felsic gneiss from the Barnard from near Bridgewater, Vt., is 418±1 Ma. Prismatic morphology and pitted surfaces of the zircons suggest that they are of igneous origin and were partially damaged by metamorphism. A second felsic gneiss sample of Barnard, from southeastern Vermont, has complicated U-Pb systematics, but the data from four finer size fractions of zircon suggest an Early Cambrian age. Coarser fractions have Late Proterozoic ^{207}Pb/^{206}Pb ages of 767 and 634 Ma. Either the two samples of "Barnard" are from different formations of different ages or the older zircons are detrital or xenocrystic (although they lack common detrital morphologic characteristics such as pitting, frosting, and rounding).

A felsic rock, thought to be metavolcanic in origin, and a metapelite from the Moretown Member of the Missisquoi Formation from southeastern Vermont, both contain only detrital zircons that range in ^{207}Pb/^{206}Pb age from about 1.01 to 1.66 Ga. Although we are unable to determine the age of deposition of the Moretown (it has been mapped as conformably underlying the Barnard Member), our data indicate an Early to Middle Proterozoic source(s), probably from the Grenville province of the Adirondack Mountains and of Canada, for at least some of the detritus.

The Standing Pond Volcanics, a suite of mafic extrusive rocks, occur between the Waits River and Gile Mountain Formations. Eight size fractions of zircon from a fine-grained metafelsite dike that cuts the volcanics near Springfield, Vt., form a discordia that has an upper intercept age of 423±4 Ma. This age indicates that the Standing Pond Volcanics and the Waits River Formation must be Silurian or older.

INTRODUCTION

The geology of eastern Vermont was most recently compiled by Doll and others (1961) on the State Centennial Geologic Map. East of the Precambrian gneisses and schists in the core of the Green Mountains anticlinorium are Paleozoic rocks of the eastern cover sequence, the so-called "Vermont sequence" (White and Jahns, 1950; Thompson, this volume). These rocks have been distinguished from those of the "New Hampshire sequence" (White and Jahns, 1950), located to the east across the Monroe line or fault (Hatch, 1988a). The ages of the "Vermont sequence" rocks have mostly been assigned by their correlation with dated rocks (either by paleontologic or radiometric methods) to the north, east, and west. Plate tectonic models have been proposed for rocks of northern New England (Zen, 1983a; Stanley and Ratcliffe, 1985), including eastern Vermont. However, these models involve many undated stratigraphic units.

To provide some absolute time constraints for the tectonic models, we have undertaken a dating study of five rocks that compose part of the "Vermont sequence." These rocks include two samples of the Barnard Volcanic Member of the Missisquoi Formation, two samples of the Moretown Member of the Missisquoi Formation, and a dike cutting the Standing Pond Volcanics (unit names follow the terminology of Hepburn and others, 1984). The crystallization ages of these units have not been previously determined, although ages of metamorphism in eastern Vermont, as measured by the ^{40}Ar/^{39}Ar method, have been published (Laird and others, 1984; Sutter and others, 1985; Harrison and others, 1986). This report will attempt to constrain the ages of rocks whose petrologic and tectonic origins have been discussed at great length but for which the prerequisite geochronologic data are lacking.

Manuscript approved for publication April 13, 1990.

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REGIONAL GEOLOGIC RELATIONS IN EASTERN VERMONT

The stratigraphy of Paleozoic rocks in east-central Vermont (fig. 1) is summarized by Doll and others (1961 and references therein), as compiled from numerous published and unpublished quadrangle maps. Briefly, the Middle and Late Proterozoic schists and gneisses in the core of the Green Mountains anticlinorium are unconformably overlain by a series of undated and unfossiliferous, mostly metasedimentary, rocks thought to be Late Proterozoic or Cambrian, including (from oldest to youngest) the Cavendish, Tyson, Hoosac, and Pinney Hollow Formations. These formations are overlain by the Ottauquechee and Stowe Formations, which are Cambrian to Ordovician in age, based, in part, on Middle to Late Ordovician $^{40}Ar/^{39}Ar$ ages on amphiboles (Laird and others, 1984). Locally unconformably above the Stowe Formation is the Missisquoi Formation, which is made up of the Moretown Member (a quartz-plagioclase granulite containing interlayers of pelitic schist and felsic and mafic metavolcanic rocks), several metasedimentary members, and the Barnard Volcanic Member (a bimodal metavolcanic suite). The top of the eastern Vermont sequence is represented by the fossiliferous Shaw Mountain Formation of Silurian age and the Northfield, Waits River, and Gile Mountain Formations of presumed Silurian-Devonian age. On many maps, the Standing Pond Volcanics separate the Waits River and Gile Mountain Formations (for example, Doll and others, 1961). The uncertainty in the ages of the Moretown and Barnard Members of the Missisquoi Formation and of the Waits River and Gile Mountain Formations precipitated this study.

ANALYTICAL TECHNIQUES

Standard mineral separation techniques, including Wilfley Table, heavy liquids, and magnetic separations, were used to separate zircons from our samples. Individual fractions were sized through silk screens, rerun on the magnetic separator, and hand picked for purity. In some fractions, additional hand-picked splits were made on the basis of crystal morphology (elongate versus stubby).

Uranium and lead were extracted from the zircons by using methods modified slightly from Krogh (1973). Samples larger than 1 mg (milligram) were spiked with ^{235}U and ^{208}Pb tracers; smaller samples were spiked with a mixed ^{235}U-^{205}Pb spike. Lead blanks were about 1 ng (nanogram) for large samples and about 0.1 ng for small samples except in the (−200+250 mesh)NMR (nonmagnetic, round) fraction from sample 2711, which probably had some spot contamination. Isotopic ratios were analyzed on an automated VG 54E mass spectrometer that has a Daly detector for signals smaller than 50 millivolts. Errors for Pb/U ages have a range of about 0.45 to 0.65 percent (2σ); errors for $^{207}Pb/^{206}Pb$ ages have a range of about 0.03 to 0.38 percent (2σ). Errors for the (−200+250)NMR fraction of sample 2711 are significantly larger. Corrections for common lead were made by using the appropriate values from Stacey and Kramers (1975). The computer programs of Ludwig (1980, 1987) were used for age calculations and concordia plots.

SAMPLE DESCRIPTIONS AND ISOTOPIC RESULTS

The Barnard Volcanic Member of the Missisquoi Formation forms a narrow belt (maximum map width of about 3 km) (Doll and others, 1961) that extends from east-central Vermont south around the Chester and Athens domes and into Massachusetts where it is mapped as the metavolcanic member of the Hawley Formation (Hatch, 1967) of the Rowe-Hawley lithotectonic zone (Zen, 1983b). Sample WS–7, from a pavement in the Ottauquechee River just east of Bridgewater, Vt., was collected from an outcrop composed of alternating thin layers as much as 0.5 m thick of amphibolite and felsic gneiss (fig. 1). The rocks here have been metamorphosed to garnet grade; some textures (such as megacrysts) may be relict igneous features. Sample WS–7 is from a conformable layer interpreted to be a metamorphosed felsic volcanic rock within the bimodal sequence of the Barnard Volcanic Member. Concordance of lithologic layering is evidence that these rocks were originally flows and that our sample is not a dike (or sill), although these possibilities cannot be entirely ruled out. Sample 87Zen–1, from an outcrop just east of a north-south dirt road about 2.7 km west of Brockways Mills, Vt., is from an outcrop dominated by felsic gneiss that contains amphibolite layers up to 0.5 m thick. The lithologic layers appear to be concordant, and sample 87Zen–1 is from a felsic gneiss layer interpreted to be a metamorphosed rhyolite. The assignment of this unit to the Barnard Volcanic Member is based on its lithologic similarity to Barnard rocks near the type locality (including our sample WS–7) and by its association with rocks that are mapped as the Moretown Member. Hepburn and others (1984) have shown that the amphibolites of the Barnard Volcanic Member probably had a basaltic protolith, and Chang and others (1965) demonstrated that the felsic gneisses have chemical affinities of dacites or rhyodacites.

Zircons from both samples of the Barnard Volcanic Member are euhedral, clear to light brown, and blocky (length to width ratio (*l/w*) of 1–4). Most grains have sharp pyramidal terminations and slightly pitted surfaces. A few rounded, darker grains were present in the finer size fractions of sample WS–7. We assumed that these were xenocrystic zircons of detrital origin included in the volcanics during extrusion; they were hand picked and discarded from the analyzed fractions. On the basis of these morphologic descriptions, zircons in both samples appear to be igneous in origin.

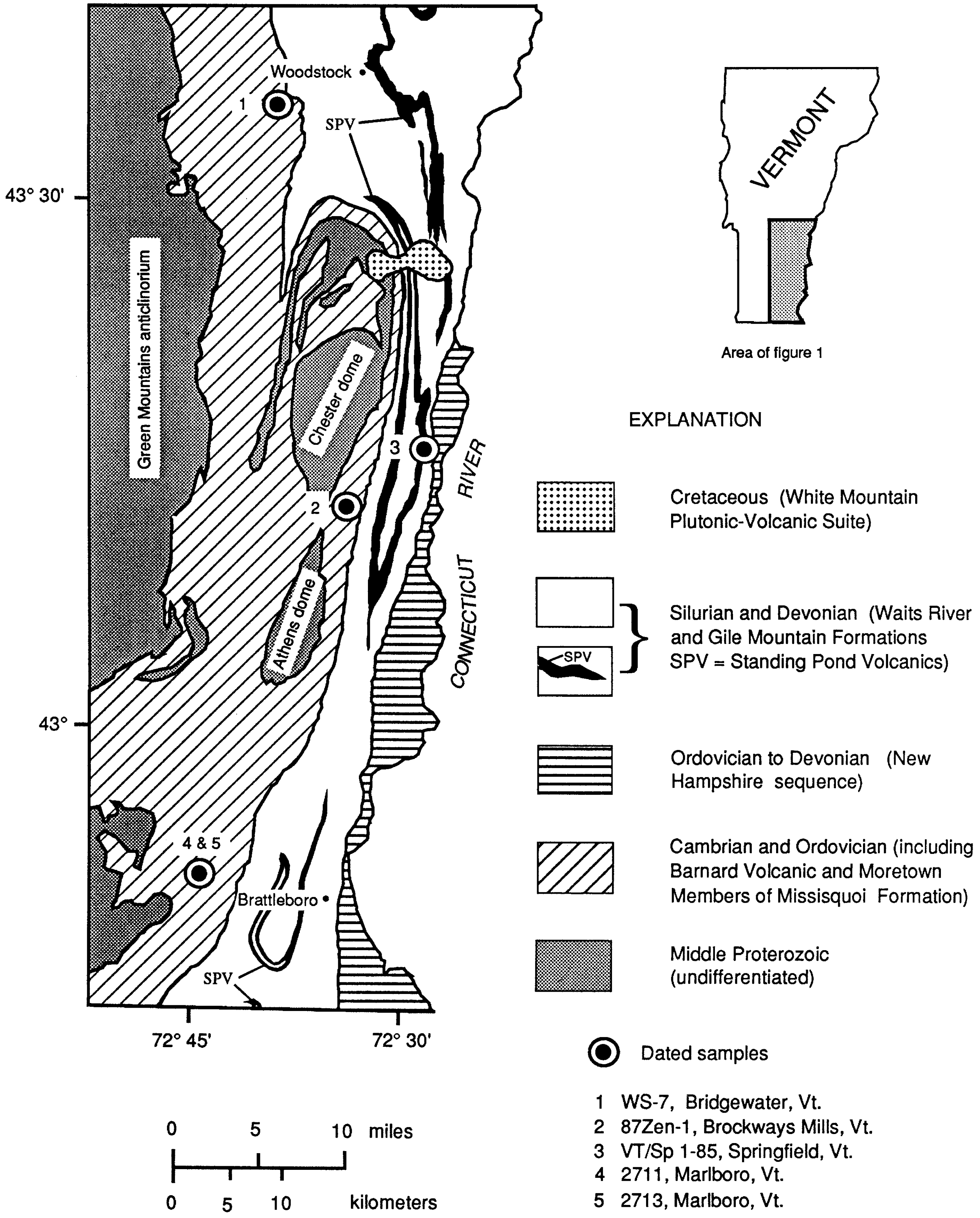

Figure 1. Generalized geologic map of southeastern Vermont showing sample locations.

Table 1. Uranium-lead concentration and composition data from zircon taken from metavolcanic and metasedimentary rocks of southeastern Vermont

[NM, nonmagnetic; Mag, magnetic; D, diamagnetic; R, round; L, elongate; A, abraded; Constants: $^{235}\lambda$=9.8485 E-10/yr; $^{238}\lambda$=1.55125 E-10/yr; $^{238}\text{U}/^{235}\text{U}$=137.88 (Steiger and Jäger, 1977)]

Fraction	Weight (mg)	Concentration (ppm)		Lead composition[1]			Ratios (2σ percent errors)			Ages (Ma)[2]		
		U	Pb	$^{206}\text{Pb}/^{204}\text{Pb}$	$^{206}\text{Pb}/^{207}\text{Pb}$	$^{206}\text{Pb}/^{208}\text{Pb}$	$^{206}\text{Pb}/^{238}\text{U}$	$^{207}\text{Pb}/^{235}\text{U}$	$^{207}\text{Pb}/^{206}\text{Pb}$	$^{206}\text{Pb}/^{238}\text{U}$	$^{207}\text{Pb}/^{235}\text{U}$	$^{207}\text{Pb}/^{206}\text{Pb}$
WS–7 (Barnard Volcanic Member, lat 43°35′04″, long 73°36′58″)												
(+150)NM	15.96	334.5	26.48	4,474.4	17.119	3.2869	0.0672(.46)	0.5107(.48)	0.0552(.14)	419	419	418
(+150)Mag	20.70	346.6	27.12	10,733	17.695	3.2852	.0667(.44)	.5069(.44)	.0552(.03)	416	416	419
(−150+200)NMD	5.34	348.4	26.49	35,481	17.986	3.3524	.0652(.56)	.4960(.57)	.0552(.06)	407	409	420
(−150+200)Mag	7.13	512.5	39.27	6,334.6	17.413	3.1283	.0644(.48)	.4896(.51)	.0551(.16)	402	405	417
87Zen–1 (Barnard Volcanic Member, lat 43°12′23″, long 72°33′07″)												
(+150)NMD	6.44	233.6	22.71	14,602	15.205	4.6971	0.0882(.57)	0.7879(.58)	0.0648(.08)	545	590	767
(−150+200)NMD	5.79	235.9	22.10	17,671	16.215	4.2753	.0839(.56)	.7039(.60)	.0608(.22)	519	541	634
(−200+250)NMD	6.88	259.0	23.10	12,208	16.844	4.0040	.0790(.56)	.6339(.56)	.0582(.06)	490	499	536
(−250+325)NMD	5.22	287.5	24.99	51,456	17.184	3.9636	.0770(.56)	.6151(.57)	.0579(.11)	478	487	526
(−325+400)NMD	4.90	297.0	27.60	40,693	17.181	3.8731	.0820(.57)	.6539(.60)	.0578(.20)	508	511	524
(−400)NM	7.36	353.6	30.65	404,350	17.198	3.7596	.0760(.56)	.6092(.56)	.0581(.06)	472	483	534
2711 (Moretown Member, felsic gneiss, lat 42°52′00″, long 72°45′42″)												
(+100)NMR	0.240	198.6	39.52	2,441.6	10.241	5.4386	0.1790(.56)	2.269(.56)	0.0919(.08)	1061	1203	1466
(−100+150)NML	1.50	537.8	82.61	3,022.4	10.969	8.1514	.1463(.47)	1.745(.52)	.0865(.20)	880	1025	1350
(−100+150)NMR	1.30	221.5	43.55	1,096.7	10.012	4.9893	.1730(.76)	2.078(.87)	.0871(.39)	1029	1142	1363
(−150+250)NMR	.265	396.6	65.84	5,108.7	9.8232	5.9747	.1514(.61)	2.069(.63)	.0991(.15)	909	1139	1607
(−200+250)NMR	.064	553.8	73.42	836.16	11.132	7.2794	.1226(7.1)	1.230(7.5)	.0728(1.6)	746	815	1008
(−250+325)NMD	11.40	431.5	65.63	9,700.9	11.205	7.8265	.1450(.43)	1.756(.45)	.0878(.10)	873	1029	1378
2713 (Moretown Member, pelitic schist, lat 42°52′15″, long 72°45′42″)												
(+150)NMR	0.125	155.5	35.72	575.1	7.9436	4.7511	0.1930(.67)	2.714(.69)	0.1020(.18)	1137	1332	1661
(−150+200)NMR	.226	284.8	48.57	6,332.9	11.205	7.3881	.1614(.49)	1.937(.50)	.0870(.07)	964	1094	1361
(−200+250)NMD	.161	274.4	39.04	2,257.2	11.415	7.1001	.1335(.57)	1.497(.59)	.0815(.16)	808	929	1230
(−250+325)NMD	10.90	336.8	51.85	9,788.2	11.079	7.8792	.1468(.47)	1.798(.60)	.0888(.38)	883	1045	1401
VT/Sp 1–85 (Dike cutting Standing Pond Volcanics, lat 43°16′14″, long 73°27′23″)												
(−100+150)LA	1.03	434.9	33.31	7,246.8	17.459	3.4754	0.0660(.12)	0.5026(.19)	0.0553(.13)	412	413	423
(+150)NM	5.81	316.9	27.82	306.09	9.7006	2.6680	.0647(.41)	.4946(.51)	.0554(.30)	404	408	429
(+150)NMA	6.57	361.9	26.51	4,540.0	17.096	3.6557	.0636(.46)	.4847(.49)	.0553(.15)	397	401	423
(−150+200)NM	13.44	305.4	22.62	5,209.4	17.240	3.6080	.0642(.48)	.4887(.48)	.0552(.08)	401	404	420
(−200+250)NM	12.56	317.1	23.43	6,441.8	17.399	3.5987	.0641(.45)	.4876(.46)	.0552(.09)	400	403	421
(−250+325)NM	12.85	322.6	23.78	8,336.7	17.578	3.5880	.0640(.45)	.4862(.45)	.0551(.06)	400	402	418
(−325)NM	11.89	309.7	22.94	5,187.5	17.211	3.5381	.0639(.44)	.4874(.46)	.0553(.11)	399	403	424
(−325)Mag	4.92	421.3	32.05	1,978.3	15.966	3.1095	.0632(.45)	.4816(.49)	.0553(.17)	395	399	423

[1] Samples larger than 1 mg corrected for 1-ng blank and mass fractionation of 0.14±.05 percent/atomic mass unit; smaller samples corrected for 0.1-ng blank lead.

[2] Common lead corrected for by using appropriate values from Stacey and Kramers (1975).

Four size fractions of zircon from sample WS–7 were analyzed for uranium and lead isotopes (table 1). The two coarse fractions, which differ only in magnetic susceptibility, are concordant within error at about 418 Ma. The two finer fractions are only slightly discordant (about 3 and 4 percent). A best-fit line through the four data points has concordia intercept ages of 418±1 and −42±67 Ma (fig. 2). Six size fractions of zircon from sample 87Zen–1 were analyzed (table 1). In marked contrast to the data for WS–7, the two coarsest fractions have $^{207}\text{Pb}/^{206}\text{Pb}$ ages of 767 and 634 Ma; $^{207}\text{Pb}/^{206}\text{Pb}$ ages in the four finer fractions range from 524 to 536 Ma. A best-fit line through these four splits yields poorly constrained intercept ages of 524±90 and −90±540 Ma (fig. 2). When forced through a present-day lower intercept, the upper intercept age is 524±4 Ma.

An unanswered question generated by the isotopic data from sample WS–7 is whether the 418-Ma date is the age of extrusion of the felsic volcanic protolith or the age of metamorphism of the felsic gneiss. On the basis of $^{40}\text{Ar}/^{39}\text{Ar}$ cooling ages on hornblende and muscovite (Laird and others, 1984; Sutter and others, 1985), Acadian metamorphism in eastern Vermont is somewhat older than about 375–385 Ma, probably about 405 Ma (Spear and Harrison, 1989). Also, because zircon is refractory, U-Pb systematics

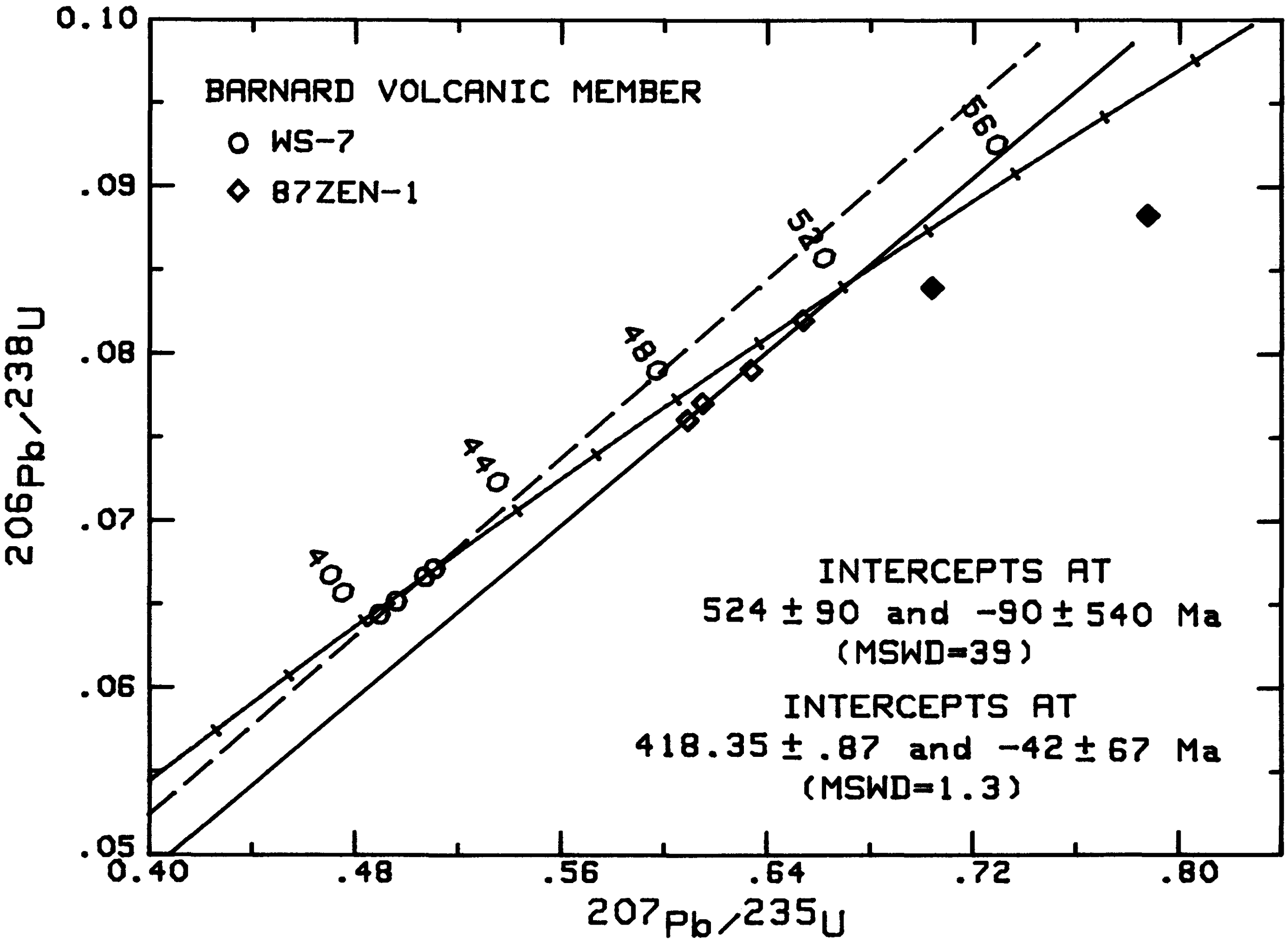

Figure 2. Concordia plot of U-Pb data from samples of the Barnard Volcanic Member of the Missisquoi Formation. Open circles are sample WS–7; open diamonds are sample 87Zen–1. Closed diamonds, which represent fractions that contain inherited radiogenic lead, are omitted from the regression. MSWD=mean square of weighted deviates.

in zircon in felsic gneiss probably would not be completely reset at garnet grade. The isotopic systematics and the prismatic, pitted morphology of the zircons from sample WS–7 thus suggest an igneous origin modified by Acadian metamorphic resorption. However, because the ^{207}Pb/^{206}Pb ages are identical within analytical error, moderate amounts of lead loss must have occurred recently.

Interpretation of the data from sample 87Zen–1 is difficult. On the basis of the U-Pb geochronology, the suggested age of about 525 Ma is older than the historically inferred Middle to Late Ordovician age for the Barnard Volcanic Member; this age is also about 100 million years older than our other Barnard sample. The Barnard Volcanic Member is shown as a continuous unit between our two sample localities (Doll and others, 1961). The geochronologic data suggest two possibilities: (1) the zircons in sample 87Zen–1 are detrital or xenocrystic and therefore record only the age of the provenance, not the age of extrusion of the felsic volcanics or (2) one of the dated felsic gneisses is not the Barnard Volcanic Member of the Missisquoi Formation, despite their similar appearance and geologic relations. If the zircons are detrital, they must have been derived from a local source because they lack the morphologic characteristics (for example, rounding) commonly shown by detrital zircons from sedimentary rocks. Possibility (2) suggests a Cambrian age for felsic gneiss from southeastern Vermont and implies that the structural and (or) stratigraphic complexities in this area are not yet fully understood. Because sample WS–7 is on strike with, and only about 15 km south of, the type locality of the Barnard, it is more likely that the rocks from which sample 87Zen–1 was collected were misidentified during geologic mapping.

The Moretown Member is considered to be part of the Missisquoi Formation (Hepburn and others, 1984), although in the past the Moretown has been ranked as a formation (for example, Ern, 1963). Most of the Moretown Member consists of quartz-rich gneiss and schist that have a characteristic pinstripe appearance (caused by alternating thin quartzofeldspathic and micaceous laminae), quartzite,

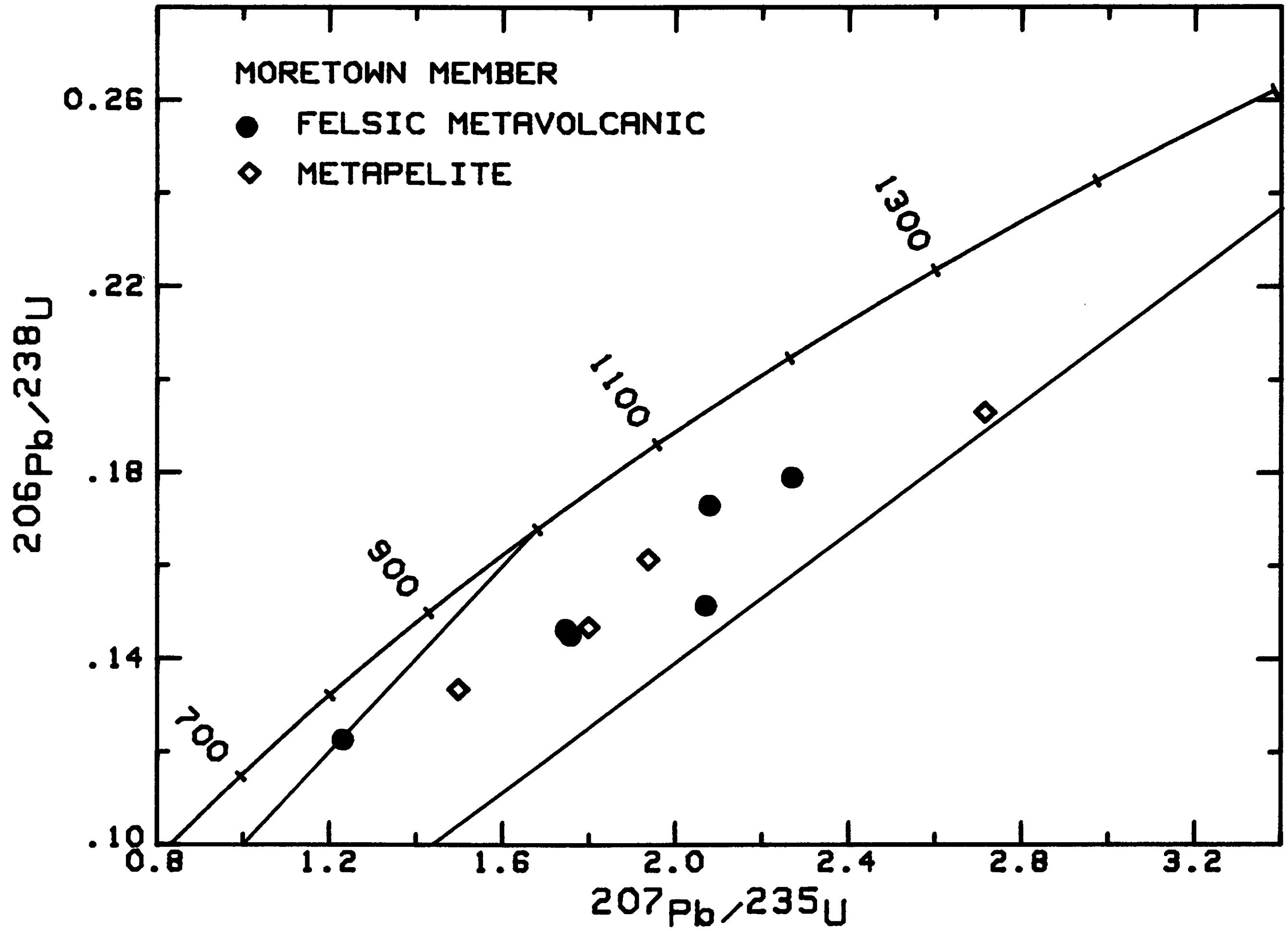

Figure 3. Concordia plot of U-Pb data from samples of the Moretown Member of the Missisquoi Formation. Closed circles are sample 2711 (felsic gneiss); open diamonds are sample 2713 (pelitic schist). Reference chords that have present-day lower intercepts have intercepts of 1.0 and 1.7 Ga.

and pelitic schist. It also contains greenstone, amphibolite, rhyolitic metatuff, and volcanogenic metasedimentary rocks (Konig, 1961; Skehan, 1961; Cady and others, 1963; Chang and others, 1965). In east-central and southeastern Vermont, the Moretown Member grades into the Barnard Volcanic Member (Ern, 1963; Hepburn and others, 1984).

We collected two samples of the Moretown Member from near Marlboro, Vt.: a felsic gneiss of possible volcanic origin (sample 2711) and a pelitic schist (sample 2713) (fig. 1). The felsic gneiss is foliated, but not thinly laminated, is equigranular (composed of quartz, feldspar, and minor biotite), and is interlayered with amphibolite. The combination of composition, texture, and its association with amphibolite suggests that the protolith of the gneiss was a felsic volcanic rock, possibly a metatuff, although we are unable to determine the exact nature of the igneous precursor. In contrast, the pelitic schist is clearly metasedimentary. Zircons from both samples appear to be detrital in origin. Although a few facets are preserved, most zircon grains are well rounded, and many are spheres that have frosted and pitted surfaces. Thus, our goal of confirming a Silurian age of the conformably overlying Barnard Volcanic Member was not attained by using these two samples. However, equally important for tectonic models is the age of the provenance of these detrital zircons.

Ten fractions of zircon from both samples (2711 and 2713) were analyzed for uranium and lead concentrations and isotopic compositions (table 1). In the felsic gneiss, two populations of the coarser fractions, elongate and round zircons, were recognized and analyzed separately. In the finer fractions, where the zircons were subdivided by magnetic susceptibility only, this shape distinction was difficult to ascertain. Coarse, round zircons were analyzed from the felsic gneiss.

Concordia systematics often are not useful in studies of detrital zircon populations, but ^{207}Pb/^{206}Pb ages provide minimum dates for the age of the provenance(s). The isotopic data for the two samples of the Moretown Member do not form linear arrays (fig. 3). The ^{207}Pb/^{206}Pb ages range from 1008±32 to 1661±4 Ma; 9 of the 10 ages are older than 1230 Ma (table 1). We conclude that the ages of the provenance(s) for the detrital zircons in these samples of the Moretown Member are Early to Middle Proterozoic. These ages clearly are older than the Paleozoic rocks of eastern Vermont; the oldest rocks known from the adjacent Green Mountains and the Adirondack Mountains to the west are about 1.4 Ga (Paul Karabinos, unpub. data, 1987; McLelland and others, 1988). The nearest known source that would account for all of the ages found in our samples is to the north in the Grenville province of Quebec (Easton, 1986).

The Standing Pond Volcanics, which were originally mapped as a member within the Waits River Formation, have been elevated to formation status (summarized in Hepburn and others, 1984). In southeastern Vermont, the volcanics commonly, but not everywhere, occur between the Waits River and Gile Mountain Formations. The exact stratigraphic location of the volcanics is indeterminate because the contact between the Waits River and Gile Mountain Formations, based on percentage of carbonate versus siliciclastic sediments, is gradational. Fisher and Karabinos (1980) and Hatch (1988b, oral commun., 1988) have found graded beds at the contact, indicating that, in some localities, the Gile Mountain overlies the Waits River. However, the two formations may be at least partially time-synchronous facies (as shown by Doll and others, 1961). The Standing Pond Volcanics probably represent a time-stratigraphic marker unit that was extruded on both formations.

The age of the Waits River-Standing Pond Volcanics-Gile Mountain Formations has been a matter of much conjecture and is at present somewhat controversial. Stratigraphic evidence (summarized in Chang and others, 1965; Hepburn and others, 1984) based on correlation with fossiliferous rocks in Quebec and lithologic similarity to the fossiliferous Early Devonian Littleton Formation in New Hampshire (Billings, 1956) suggests a Silurian to Early Devonian age for the sequence. Bothner and Finney (1986), who reoccupied field sites originally found by C.H. Richardson (1902–18), reported Middle to Late Ordovician graptolites from a metalimestone within the Waits River Formation near Montpelier, Vt. However, recent reexamination of several localities has led to the conclusion that the "graptolites" are really structural features related to deformation of the rocks (W.A. Bothner, oral commun., 1988). Hueber and others (1990) have identified Lower Devonian (Emsian) plant fossils from a unit on strike with the Gile Mountain Formation in southeastern Quebec. However, Spear and Harrison (1989) list ^{40}Ar/^{39}Ar "isochron" ages from hornblende of 425±5 Ma from the Gile Mountain Formation and 440±10 Ma from the Standing Pond Volcanics.

The Standing Pond Volcanics are composed chiefly of amphibolite. The presence of pillow structures (Dennis, 1956; Hall, 1959) in northeastern Vermont, and primary volcanic bombs in southeastern Vermont (J.F. Slack, oral commun., 1989), indicate an extrusive origin. Minor interlayered phyllite, schist, and quartzite are probably detrital admixtures. Our sample (VT/Sp 1–85) was collected from a roadcut of thickly bedded amphibolite and fine-grained mafic granulite, near Springfield, Vt. At the western end of the outcrop, a felsic, fine-grained, weakly foliated dike composed of quartz, albite, and minor chlorite and epidote (formerly trondhjemite?) cuts the nearly vertical metamorphic foliation of the Standing Pond amphibolites. Because the crosscutting relations are subtle and difficult to observe, the origin of the dike is undetermined; possibilities include a feeder to the volcanics, a layer within the volcanics, or a dike unrelated to the volcanics. Thus, an age on the dike at least provides a minimum age for the Standing Pond Volcanics. The dike was the only rock in the outcrop from which we were able to obtain zircons.

Zircons from the dike cutting the Standing Pond Volcanics are clear to light brown, euhedral, and stubby (l/w=1–2). Most grains have at least one pyramidal termination, and most faces are pitted. No distinct cores are visible. This morphology appears to indicate an igneous origin, with surface pitting caused by metamorphism. No xenocrystic zircons were found in this sample.

Eight size fractions of zircon from the Standing Pond Volcanics were analyzed for uranium and lead isotopes, including one abraded split of the (+150)NM fraction (table 1). The Pb/U and ^{207}Pb/^{206}Pb ages have little variation and range from 395 to 413 and 418 to 429 Ma, respectively. One abraded split was slightly more discordant than the unabraded split of the same size fraction, probably due to a higher uranium content in the cores (table 1). However, an abraded split composed of hand-picked, euhedral, relatively clear zircons is only slightly discordant and has Pb/U ages of 412 and 413 Ma and a ^{207}Pb/^{206}Pb age of 423 Ma. Because there is little spread in the Pb/U ages, the only meaningful concordia intercept age is determined by forcing the regression through a lower intercept of 0±50 Ma (fig. 4), a reasonable assumption given the narrow range of ^{207}Pb/^{206}Pb ages. Using this method, the upper intercept age is ~423±4 Ma. Another approach, calculating the weighted average of the ^{207}Pb/^{206}Pb ages, yields an age of 421±2 Ma. We conclude that this dike is Middle Silurian, thereby providing a minimum age for the Standing Pond Volcanics.

In summary, the new U-Pb zircon data indicate the following ages for rocks of southeastern Vermont:

(1) Barnard Volcanic Member: 418±1 Ma.

(2) A dike cutting the Standing Pond Volcanics: 423±4 Ma (or 421±2 Ma, using weighted average of ^{207}Pb/^{206}Pb ages).

(3) Moretown Member: No primary age obtained. All zircons from a felsic metavolcanic rock and a pelitic schist within this formation are detrital. The age(s) of the provenance(s) of the sediments is Early to Middle Proterozoic

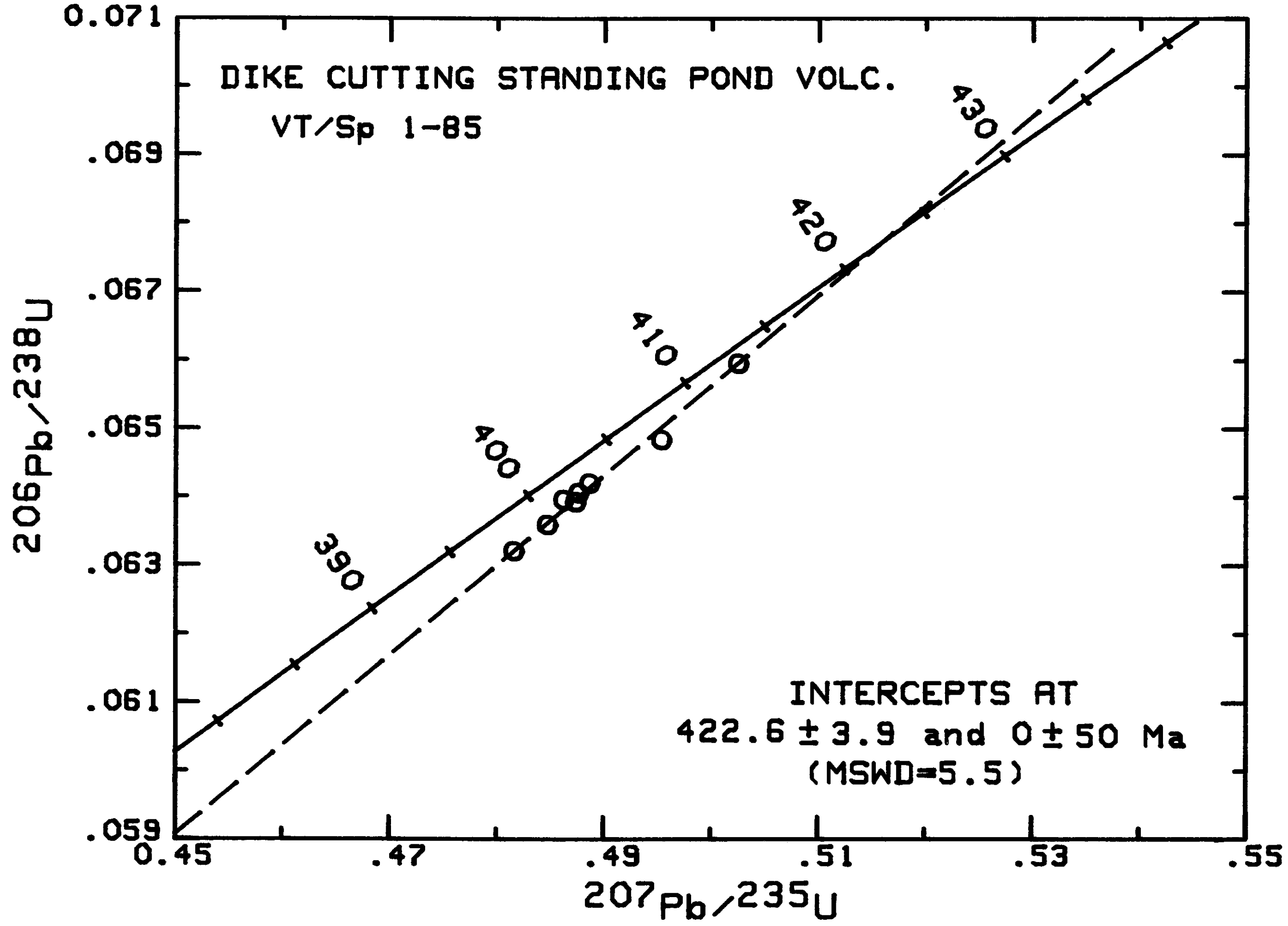

Figure 4. Concordia plot of U-Pb data from sample VT/Sp 1–85 (Standing Pond Volcanics). MSWD=mean square of weighted deviates.

(about 1.0 to 1.7 Ga). On the basis of the ages of the detrital zircons, the samples appear to be derived from the same source terrane(s).

(4) A second sample of the "Barnard Volcanic Member" may really be Early Cambrian. Alternatively, the zircons in this felsic gneiss may be detrital, and thus their apparent Early Cambrian age may indicate the age of the provenance.

REGIONAL AND TECTONIC IMPLICATIONS

The new U-Pb zircon data presented herein are inconsistent with previous interpretations of the stratigraphy of southeastern Vermont; therefore, tectonic models based on incorrect age assignments will require some modification. The probable Upper Silurian age of the Barnard Volcanic Member suggests that the underlying Moretown Member of the Missisquoi Formation may also be Silurian. However, Jacobi and Fountain (1988) show that amphibolite in the Moretown of central Massachusetts (within the Rowe-Hawley zone) is chemically distinct from amphibolites of the Hawley Formation. This difference suggests a structural discontinuity between the Hawley (Barnard) and Moretown units. Thus, the age of the Barnard Volcanic Member may be significantly younger than previously thought, and the relation of these rocks to the Moretown Member may not be a simple conformable transition, as mapped by most geologists in southeastern Vermont. Furthermore, as suggested by our results in southeastern Vermont, the Hawley Formation may be much older than the Barnard Member.

Our preferred minimum age of about 423 Ma for the Standing Pond Volcanics indicates that the entire Waits River-Gile Mountain sequence is not latest Silurian to Early Devonian but that at least part of the sequence must be Middle Silurian or older. A number of problems still hinder full understanding of the origin of these rocks. On the basis of sedimentologic evidence, the entire Waits River and Gile Mountain package apparently was deposited fairly rapidly (Hatch, 1988b, oral commun., 1988). However, the combination of paleontologic and geochronologic evidence suggests that the rocks range in age from at least Middle to Late Silurian to Early Devonian. Although more than one suite of rocks may have been deposited, no unconformity or fault has ever been recognized in the field. As shown on the State geologic map (Doll and others, 1961), the Waits River and Gile Mountain Formations may be facies variants deposited synchronously. However, Fisher and Karabinos (1980) and Hatch (1988b, oral commun., 1988) have shown conclusively that, at least locally, the Gile Mountain overlies the Waits River. Additional complications include uncertainty of fossil identification (W.A. Bothner, oral commun., 1987) and the possibility that some amphibolites mapped as Standing Pond Volcanics may belong to another stratigraphic suite (for example, Ammonoosuc Volcanics). Until more mapping, paleontology, and geochronology provide additional evidence, these problems will remain unresolved.

The Early to Middle Proterozoic age of detrital zircons in the Moretown Member of the Missisquoi Formation indicates a Proterozoic-age provenance for sediments presumably deposited in the middle Paleozoic. We are unable to determine precisely the age of the Moretown, but if it conformably grades into the Barnard Volcanic Member (as mapped throughout eastern Vermont), then we suggest that crustal rocks in the Green Mountains of Vermont and the Grenville province of Quebec were being eroded and deposited in Vermont during the Late Silurian. This model differs from the interpretation that the Moretown Member was part of an Ordovician island arc (Rowley and Kidd, 1981, fig. 1) or a forearc basin (Williams, 1978; Stanley and Ratcliffe, 1985). Stanley and Ratcliffe (1985, p. 1239) proposed that the mafic and felsic volcanic components of the Moretown Member had a "source to the east in the island arc setting of the Bronson Hill slice," whereas "the quartz-rich rocks of the Moretown originated from the emerged part of an accretionary wedge made up of older rocks now exposed in the Rowe Schist, Underhill and equivalent formations" west of the Moretown Member. Our two samples, which represent the two lithologies described by Stanley and Ratcliffe (1985), apparently were derived from the same provenance. The detrital zircons were not derived from a Paleozoic island arc, although they may have been recycled through rocks of intermediate age. If the contact between the Barnard Volcanic and Moretown Members is conformable and gradational, then the Moretown was probably deposited in the Late Silurian, not the Middle to Late Ordovician, as previously proposed. Unfortunately, our data are only able to constrain the age of the Moretown as younger than Middle Proterozoic; this unit could, therefore, be much older than the adjacent Barnard Volcanic Member if the contact between the two members is a fault or an unconformity.

ACKNOWLEDGMENTS

We thank Ken Ludwig and Kathleen Simmons for help in analyzing the small samples of zircon from the Moretown Member. Karabinos acknowledges the National Research Council for supporting him via the U.S. Geological Survey postdoctoral fellowship program during a sabbatical leave. Conversations with John Lyons, Bob Moench, Wally Bothner, Jo Laird, Norm Hatch, and Jim Thompson have helped our understanding of the present state of ideas concerning the geology of southeastern Vermont. However, we assume full responsibility for interpretations presented herein. John Lyons, Jim Thompson, John Slack, Bob Moench, and E-an Zen helped in sample selection. We thank Norm Hatch, Wally Bothner, Robert Ayuso, and Carl Hedge for thoughtful reviews.

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Chapter E

Lineament Analysis of the Glens Falls 1° × 2° Quadrangle, New York, Vermont, and New Hampshire

By HOWARD A. POHN

U.S. GEOLOGICAL SURVEY BULLETIN 1887

SUMMARY RESULTS OF THE GLENS FALLS CUSMAP PROJECT, NEW YORK, VERMONT, AND NEW HAMPSHIRE

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FIGURES

Lineament Analysis of the Glens Falls 1° × 2° Quadrangle, New York, Vermont, and New Hampshire

By Howard A. Pohn[1]

Abstract

Landsat multispectral scanner, Landsat return-beam vidicon, and side-looking airborne radar images were used in the lineament analysis of the Glens Falls 1° × 2° quadrangle. The multispectral scanner and return-beam vidicon images show little correlation with mineral deposits. The side-looking airborne radar images, however, show an excellent correlation between molybdenum deposits and diffuse, yet identifiable, long lineaments that strike N. 60° to 65° W. Garnet deposits in the northwestern part of the quadrangle are spatially coincident with individual short lineaments, and some iron and lead deposits in the central part of the quadrangle are associated with swarms of short lineaments. The northwestern-striking system of long lineaments is believed to be the surface manifestation of a deep-seated fracture system that occurs throughout New England and indeed throughout most of the Eastern Overthrust Belt of the Appalachian orogen.

INTRODUCTION

Studies of topographic lineaments in the northern Appalachians date from 1904 when Hobbs first studied patterns of lineaments in southern New England. More recent lineament studies, which include the Glens Falls area, are by Barosh (1976), Isachsen (1976), Wise (1976), Goldstein and Wise (1982), Palmquist and Pees (1984), and Fakundiny (1986). These recent studies used Landsat images, raised relief maps, or both. Most recently Shake and McHone (1987) used Landsat images and raised relief maps to study the lineament pattern for the entire Glens Falls quadrangle and the adjacent region north and east of the quadrangle.

Landsat multispectral scanner (MSS), Landsat return-beam vidicon (RBV) (Freden and Gordon, 1983), and side-looking airborne radar (SLAR) (Moore, 1983) images were used in the lineament analysis of the Glens Falls 1° × 2° quadrangle, located in eastern New York, central Vermont, and western New Hampshire. The Landsat images were analyzed on uncontrolled mosaics at scales of 1:1,000,000 and 1:600,000 (resolution 80 m). The SLAR images were analyzed from strip data at a scale of 1:400,000 (resolution 6–10 m) and from a controlled mosaic at a scale of 1:250,000. Sun elevations varied from 27° to 48° on the MSS and RBV images. The X-band (3-cm) SLAR data were acquired from flights flown in a north-south direction and have a west look-direction and a depression angle (from the horizontal) of 9° to 31°. The SLAR data were obtained by Goodyear Aero-Service.

The terrain in the quadrangle varies from low rolling glaciated hills in the northeastern part, to moderately rugged mountains in the central part, to the Hudson Valley lowlands in the west-central part, to glaciated alpine terrain in the northwestern part.

MAP COMPILATION

Two lineament maps were produced. One map was generated from the combined Landsat RBV and MSS data (fig. 1), and the other map was generated from the radar data (fig. 2). An examination of the Landsat lineament map shows that there is virtually no correlation of mapped lineaments and known mineral deposits. The low frequency of east-west lineaments on the data sets may be due to a look-angle bias, but neither the SLAR nor the Landsat data show east-west features, and it is probable that the paucity of lineaments that have this trend is real.

Three classes of lineaments were selected for mapping. The first class includes sharply defined, linear to slightly curvilinear to highly curved or circular mappable elements. The second class includes diffuse linear or curvilinear features that consist of slightly irregular tonal lineaments, which lack the distinct, sharply defined boundaries of the first class of lineaments. The third class includes general trends that represent broad or diffuse tonal lineaments that are interrupted by crossing features but which nonetheless are intermittently traceable for tens of kilometers. These first two classes of lineaments probably represent shallow fracture systems that may or may not have demonstrable offset. Some of these lineaments represent faults, other lineaments may represent erosional scarps between rock types of differing erosional characteristics,

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Manuscript approved for publication April 13, 1990.

[1] U.S. Geological Survey.

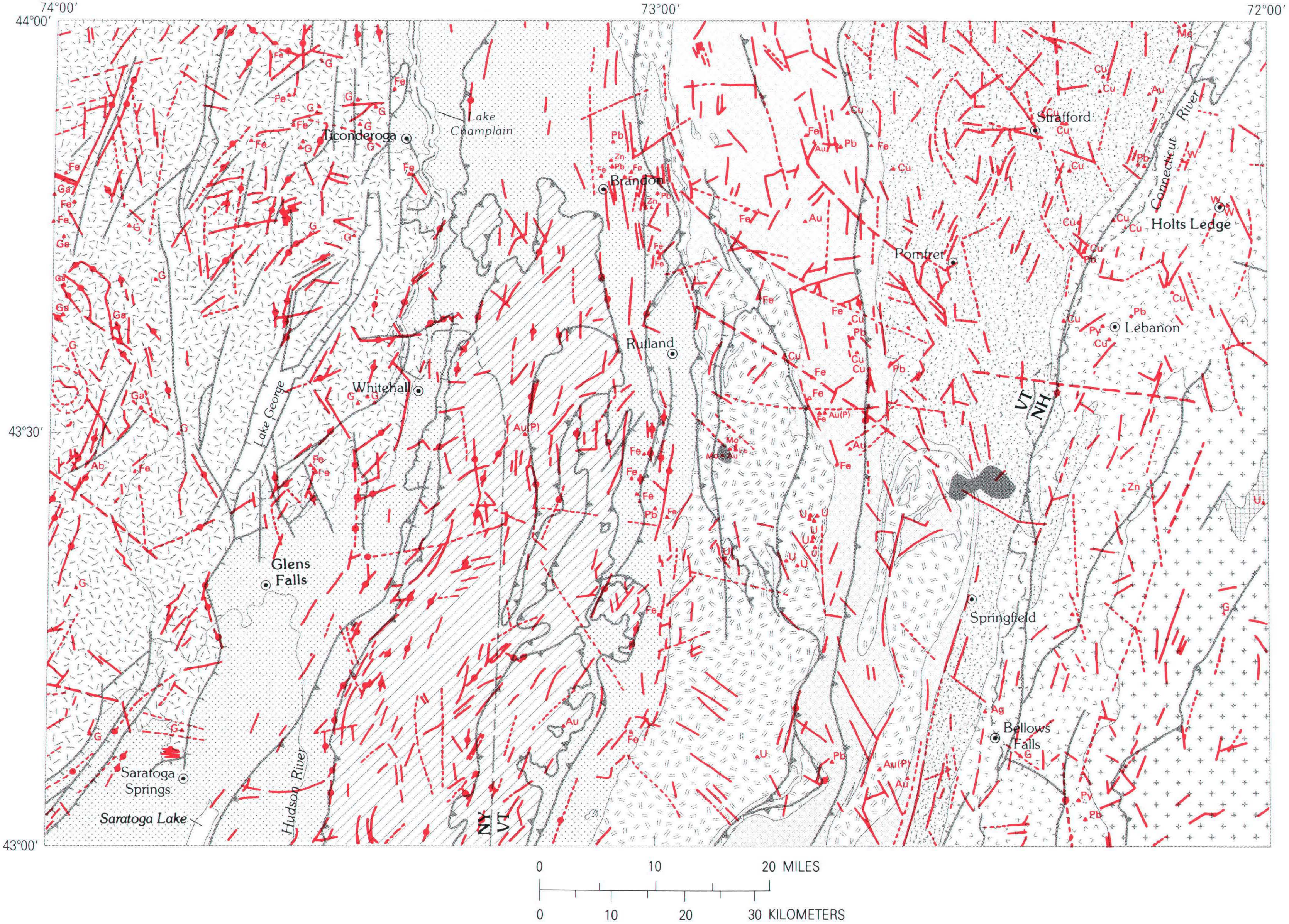

Figure 1. Landsat lineament map of the Glens Falls 1° × 2° quadrangle, New York, Vermont, and New Hampshire. This lineament map overlay was compiled from nonstable base materials. Therefore some lineaments and map boundaries may not conform to the base map.

EXPLANATION

CRETACEOUS AND JURASSIC (?)	▩	White Mountain Plutonic-Volcanic Suite (syenite, granite, and gabbro)
UPPER DEVONIAN	▦	Granite plutons (two-mica granite)
LOWER DEVONIAN AND SILURIAN	☐	Merrimack terrane (granitic gneiss, migmatite, minor pelitic schist, and quartzite)
	▒	Connecticut Valley-Gaspé terrane (pelitic schist, calcareous schist, and minor amphibolite)
ORDOVICIAN AND CAMBRIAN	▒	Parautochthonous shelf sequence (limestone, dolomite, and quartzite)
DEVONIAN, SILURIAN, AND ORDOVICIAN	▒	Bronson Hill terrane (metavolcanic rocks, pelitic schist, and granitic gneiss)
ORDOVICIAN, CAMBRIAN, AND PROTEROZOIC Z	▨	Taconic allochthon (slate, graywacke, and minor carbonate)
	▒	Eastern Green Mountain terrane (pelitic schist, metagraywacke, and minor amphibolite)
PROTEROZOIC Y	▒	Adirondack massif (felsic gneiss, quartzite, anorthosite, amphibolite, calc-silicate rocks, and metagabbro)
	▒	Green Mountain massif[1] (felsic gneiss, quartzite, minor amphibolite, and calc-silicate rocks)

—┴—┴— High-angle fault—Tick marks on downthrown side where known

—▲—▲— Thrust fault—Sawteeth on upper plate

——— — — Broad (solid line) or diffuse (long-dashed line) tonal linear or curvilinear feature, not necessarily continuous

——— Linear or curvilinear feature

- - - - - - Diffuse linear or curvilinear feature

——●—— Linear or curvilinear feature coincident with or along extension of mapped fault (Thompson and others 1990)

▲ Mineral deposit (symbols below represent major mineral production)

Au	Gold	Ga	Garnet
Ag	Silver	Mo	Molybdenum
As	Arsenic	Pb	Lead
Ab	Asbestos	Py	Pyrite
Cu	Copper	U	Uranium
Fe	Iron	W	Tungsten
G	Graphite	Zn	Zinc

[1] Includes Lincoln Mountain massif and Proterozoic rocks of the Chester and Athens domes.

Figure 1. Continued.

and still other lineaments may be indicative of joint sets or prevalent joint directions. The third class of lineaments, as interpreted by the author, may represent the surface manifestation of deeper seated (basement) features that through reactivation have penetrated the cover rocks.

Many of the faults mapped on the companion geologic map (Thompson, this volume; Thompson and others, 1990) were recognized on the Landsat or SLAR images. Some of the lineaments found on the Landsat or SLAR images but not recognizable on the geologic map may represent faults (or parts of faults) that have yet to be mapped. The major lineaments (broad or diffuse tonal lineaments) that were mapped on the images are not present on the geologic map; however, a number of lithotectonic units appear to terminate or change orientation near these lineaments.

Lineaments related to glacial features, which include oriented positive relief features and broad glacial valleys and grooves, vary in orientation from north-south to N. 30° W. Lineament patterns are generally distributed uniformly except along the major river valleys and areas of low relief. In these areas, the lineaments are obscured by cultural patterns. The Hudson Valley in the southwestern part of the map and the Lake Champlain valley and environs in the north-central part of the map are examples of culturally saturated areas. Lineaments in these areas are difficult to map and, for the most part, have been omitted from the figures.

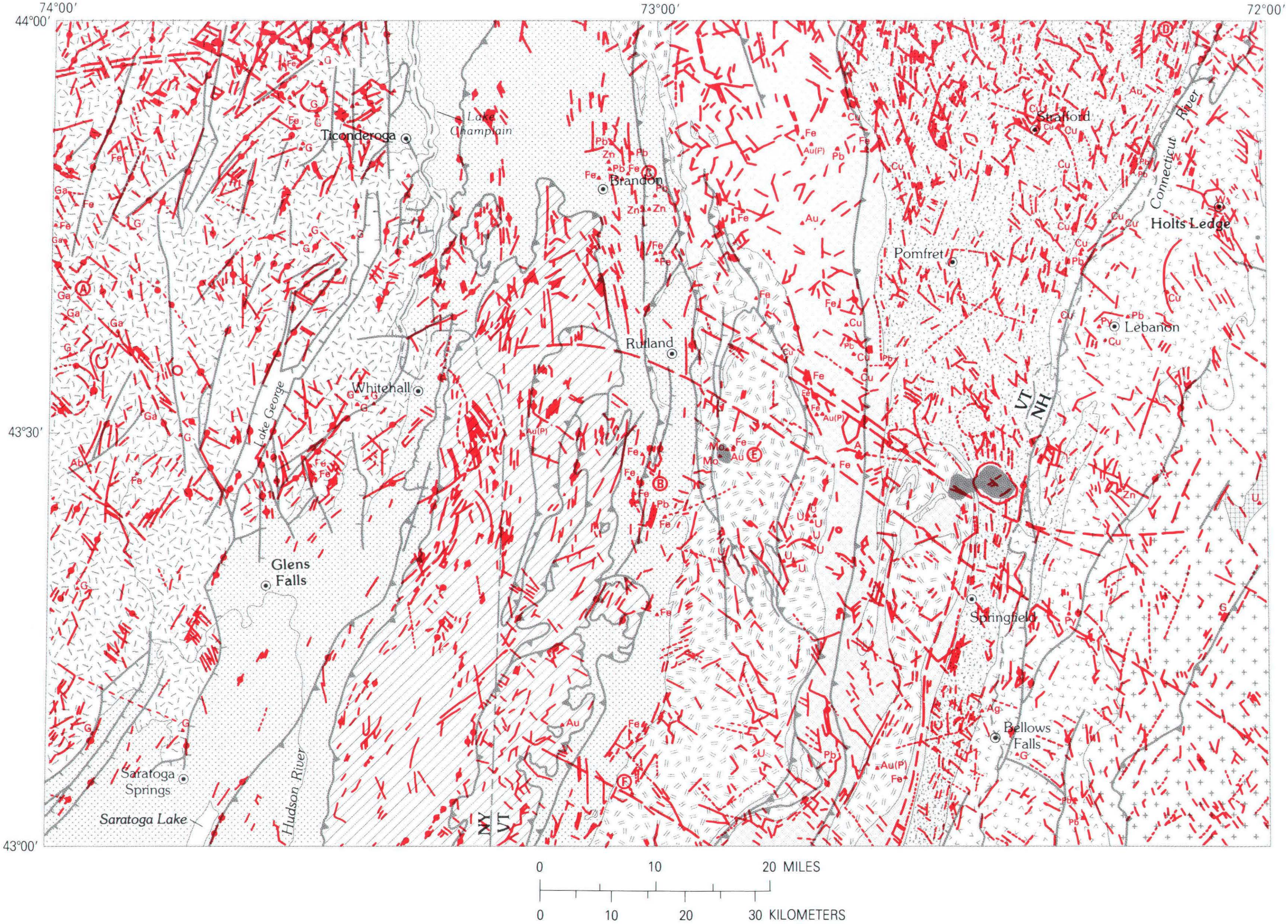

Figure 2. Radar lineament map of the Glens Falls 1° × 2° quadrangle, New York, Vermont, and New Hampshire. Circled letters indicate locations discussed in text. This lineament map overlay was compiled from nonstable base materials. Therefore some lineaments and map boundaries may not conform to the base map. See figure 1 for explanation of geologic map units and symbols.

RELATION OF MINERAL DEPOSITS TO LINEAMENTS

There is no correlation between the location of mineral deposits and lineaments except in three instances. In fact, few mineral deposits lie on or even within 2 km of lineaments. The three exceptions are (1) molybdenum and tungsten deposits in the extreme northeastern and central parts of the quadrangle (fig. 2, D and E, respectively), (2) garnet deposits in the west-central part of the quadrangle (fig. 2, A), and (3) iron and lead deposits in the central part of the quadrangle (fig. 2, B).

Perhaps the most conspicuous association of mineral localities with lineaments involves the molybdenum deposits in the extreme northeastern and central parts of the quadrangle (fig. 2, D and E, respectively). These molybdenum deposits are closely associated with two diffuse, yet identifiable, long lineaments that strike N. 60° to 65° W. These lineaments are part of a system of subparallel lineaments, observed by the author, that occur throughout New England (note a third subparallel diffuse lineament in the south-central part of figure 2 at location F). The great length and continuity of these lineaments suggest that they may be related to deep-seated (basement) faults that are part of a continental fracture system. The northeasternmost molybdenum occurrence (fig. 2, D) lies directly on one such lineament, and the nearby molybdenum- and tungsten-bearing quartz veins at Holts Ledge, N.H., have the same strike (Slack and others, 1987).

The molybdenum and gold deposits associated with the Cuttingsville, Vt., alkalic stock (center of fig. 2; see Robinson, this volume) are only 3 to 5 km south of and along a lineament subparallel to the major diffuse lineament that extends eastward from the map area nearly to the Atlantic Ocean (H.A. Pohn, unpub. data). Other molybdenum occurrences found immediately to the south of the Glens Falls quadrangle in the Albany 1° × 2° quadrangle are within 5 km of the extension of the broad diffuse lineament in the south-central part of figure 2 at location F. Still other molybdenum deposits throughout New England appear to be closely associated with this same northwest-trending lineament system. A recent study (H.A. Pohn, unpub. data) indicates that 62 percent of all molybdenum deposits in the Eastern United States (locations from J.F. Slack, oral commun., 1986; Schmidt, 1978) occur within 5 km of a major N. 60°–65° W.-striking lineament. Although 5 km may appear to be a large distance, if the "root" of the lineament in the deep crust is the source of the molybdenum (for example, through anatexis and subsequent granite evolution), then the 5-km horizontal distance is relatively insignificant when compared to the tens of kilometers of vertical migration that has occurred from the deep crust.

Commercially important garnet deposits occur only in the west-central part of the Glens Falls quadrangle within the perimeter of the Adirondack Mountains (fig. 2, A). Every garnet deposit in this region is located either on, within 0.5 km of, or along the extension of mapped SLAR lineaments that trend approximately N. 25° W., N. 65° W., or in one instance, N. 10° W. Recent field work by J.M. McLelland (oral commun., 1989) indicates that all of these garnet deposits are associated with high-angle faults that commonly are intruded by pegmatite. Although several different origins have been proposed for the deposits (Miller, 1938; Levin, 1950), it seems likely that the high-angle faults played an important role in their genesis, either directly by providing access for fluids or indirectly by localizing igneous intrusive bodies. Clearly the garnet-related lineaments are only part of a larger Adirondack lineament system that displays these three preferred directions.

Stratabound iron and lead deposits in the central part of the quadrangle (fig. 2, B) are spatially coincident with a swarm of short lineaments that strike N. 12° E. These metal deposits (Clark, this volume; Clark and Slack, 1988) may be related to other iron, lead, and zinc deposits in the quadrangle (fig. 2, C). The iron, lead, and zinc deposits at location C may have been affected by the N. 60° W. lineament system, as follows. Note that the southernmost broad lineament in the central part of the quadrangle is arcuate at its western end. If this arcuate segment is due to a counterclockwise rotation of the original lineament, then a restoration of the rotated segment to the presumed original straight lineament (N. 60° W.) would rotate the approximately north-south lineaments in the north-central iron, lead, zinc district to the same general orientation as the N. 12° E. lineaments associated with the iron and lead occurrences south of the quadrangle center. This restoration to linearity does not account for all of the 38 degrees of rotation observed on the arcuate segment, but additional rotation can be accounted for between the several broad lineaments in the central part of the map. This hypothesis, although speculative, appears to warrant additional study.

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Chapter F

Regional Exploration Geochemistry of the Glens Falls 1° × 2° Quadrangle, New York, Vermont, and New Hampshire

By KENNETH C. WATTS, JR.

U.S. GEOLOGICAL SURVEY BULLETIN 1887

SUMMARY RESULTS OF THE GLENS FALLS CUSMAP PROJECT, NEW YORK, VERMONT, AND NEW HAMPSHIRE

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Regional Exploration Geochemistry of the Glens Falls 1° × 2° Quadrangle, New York, Vermont, and New Hampshire

By Kenneth C. Watts, Jr.[1]

Abstract

Semiquantitative emission spectrographic analyses of 1,286 nonmagnetic stream sediment concentrates from active stream channels collected between 1982 and 1986 from the Glens Falls 1° × 2° quadrangle are the basis of this report. These data are reduced and summarized statistically, on the basis of the total data set and on data subsets from seven separate lithotectonic terranes. For most reported elements, median values for the individual terranes are within a sufficiently narrow range to allow contouring across terrane boundaries. By contouring across terranes, crosscutting patterns of possible regional tectonic and metallogenic significance are revealed. Isopleth maps prepared from the spectrographic data depict the distribution of reported values at or above the 50th percentile for lead, copper, iron, cobalt, tungsten, tin, niobium, boron, barium, calcium, strontium, and manganese. Other ore-forming and pathfinder elements that are significantly anomalous in stream sediment concentrates locally, but which are geographically too scattered for contouring, include arsenic, silver, gold, bismuth, beryllium, chromium, molybdenum, nickel, antimony, and zinc. R-mode factor analysis was instrumental in separating the geochemical data into associated element suites based on statistical intercorrelations. Many of the element concentrations not amenable to contouring are shown as plots of R-mode factor scores that depict the distribution of element suites.

Seven lithotectonic terranes outlined in the Glens Falls quadrangle are from west to east: (Terrane 1) the Proterozoic high-grade metamorphic (Grenville-age) terrane of the Adirondack massif, (Terrane 2) the autochthonous carbonate-siliciclastic and parautochthonous black slate and graywacke sequences of the Cambrian-Ordovician shelf terrane in the Champlain-Hudson and Vermont valleys, (Terrane 3) the Late Proterozoic to Ordovician transitional metasedimentary sequence of the Taconic allochthon of eastern New York and western Vermont, (Terrane 4) Proterozoic high-grade metamorphic gneisses and amphibolites (Grenville age) and Late Proterozoic to Cambrian cover rocks of the Green Mountain massif and the Chester and Athens domes of Vermont, (Terrane 5) a largely eugeosynclinal Cambrian to Early Devonian, predominantly metasedimentary terrane of eastern Vermont, (Terrane 6) an Ordovician to Early Devonian island-arc terrane that composes the Bronson Hill anticlinorium and the Oliverian gneiss domes, mainly east of the Connecticut River in western New Hampshire, and (Terrane 7) a syn- to postkinematic granitic terrane dominated by the New Hampshire Plutonic Suite of Devonian age. Postkinematic, alkalic plutonic complexes of Mesozoic age at Cuttingsville and Mount Ascutney, Vermont, also are considered central to some geochemical interpretations.

The isopleth patterns of the key anomalous elements cover tens of square kilometers of the Glens Falls quadrangle. They generally can be related to specific geologic features such as: (1) the traces of regional fault zones (for example, Whitcomb Summit thrust, Ammonoosuc fault), (2) the regional strike of lithostratigraphic sequences (for example, Giddings Brook thrust slice, Dorset Mountain thrust slice, basal Late Proterozoic-Cambrian cover rocks), (3) the axes of arches (for example, Strafford-Willoughby) and domes (for example, Strafford, Pomfret, Mascoma, and Alstead), (4) the outcrop areas of syn- and postkinematic igneous intrusions (for example, Sunapee granite [informal name], Brocklebank pluton, and Mount Ascutney and Cuttingsville plutons), (5) Landsat and radar lineaments (for example, a northwest-trending set), and (6) aeromagnetic anomalies (for example, east of Mount Ascutney).

R-mode factor analysis, here based on a seven-factor model, uses interelement correlations to reduce the basic log-transformed analytical data for 31 elements to fewer variables. The grouped element associations, or factors, reflect sample mineralogy and, ultimately, geochemical similarities in element characteristics. Accordingly, the element associations typically contain more information and are easier to interpret than are single-element anomalies. The seven factors derived from the R-mode analysis are as follows (in order of decreasing strength of association): Factor 1—V, Nb, Ti, B, Mn, Cr, Cu, Fe, and Mg—is a rutile (sphene)-tourmaline association in the samples. Factor 2—La, Mg, Sc, Y, B, Ca, and Ba—also is a rock-accessory association probably dominated by heavy minerals from carbonate or calc-silicate sources. Factor 3—Ag-Au—is related to gold in placers and quartz veins largely devoid of accessory sulfide minerals. Factor 4—Ni, Co, As,

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Manuscript approved for publication April 13, 1990.

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Fe, Cu, and Zn—is a sulfide mineral factor, which probably also includes arsenopyrite, an arsenate mineral. Factor 5—Sn, Pb, Sb, Ba, and Cu—suggests sources from polymetallic veins. Factor 6—Ca, Sr, Mn, Co, Pb, and Y—is a carbonate mineral or apatite association. Factor 7—W-Bi-Mo-Be—is a scheelite-rare metal association typical of quartz-scheelite-molybdenite veins.

Regional geochemical anomaly trends, derived from the element isopleth maps and the perspective plots, are of three general types: (1) crosscutting, (2) terrane restricted, and (3) along terrane boundaries. An increase in anomaly-to-background contrast, variety of anomalous elements, and size of the area yielding anomalous samples characterize the intersections of these geochemical trends. In all, 13 recognized zones of geochemical anomalies form regional trends.

Geochemical attributes of the anomalous element trends are further analyzed by a U.S. Geological Survey computer program called Relative Element Magnitude (REM). This program compares and ranks, relatively, the key chemical elements within specific anomalous zones on the basis of intensity (mean anomaly-to-background contrast) and size of the areal distribution. From these computations, the dominant elements (those highest in REM) for the anomaly trends and selected intersection areas are (1) northwesterly, crosscutting—lead; (2) northeasterly, crosscutting—tin and lead; (3) east-west, crosscutting—lead; (4) northerly, terrane-restricted—tin, lead, and lanthanum; (5) northerly, terrane boundary—lead and copper; and (6) two trend intersections—copper.

The following conclusions are based on the number of anomalous samples and on the range and geometric mean for key ore-forming and pathfinder elements, as determined for each of the seven tectono-stratigraphic terranes:

(1) Heavy-mineral samples from Terrane 5 are relatively enriched in silver, arsenic, gold, copper, cobalt, niobium, nickel, vanadium, zinc, and strontium compared with those from the other terranes. The sources of these anomalous metals are probably both vein and bedded deposits.

(2) Samples from Terrane 3 are relatively enriched in iron, barium, zinc, and antimony compared with those from other terranes. The sources are likely to be pyrite±barite±sphalerite veins, disseminations, or beds.

(3) Samples from Terrane 6 are relatively enriched in molybdenum and tungsten and are related to quartz-scheelite-molybdenite veins or to stratabound deposits in metavolcanic or metasedimentary host rocks.

(4) Samples from Terrane 7 are relatively enriched in beryllium and tin and probably are related to veins, greisens, or pegmatites.

(5) Samples from Terrane 4 are relatively enriched in calcium compared with other terranes, ranging up to 50 percent in some samples, with a geometric mean of 9.5 percent. This calcium may be related to carbonate veins known in the area or to calc-silicate rocks.

(6) Lead anomalies are found in all terranes, but particularly in Terranes 3, 4, 5, and 6. The highest geometric mean for lead (124 ppm) is in Terrane 3; the range, however, is highest in Terrane 5 (up to 15,000 ppm).

(7) High chromium (up to 7,000 ppm; geometric mean 103 ppm) and nickel, which has a low range but has the highest geometric mean in the Glens Falls quadrangle (46 ppm) in Terrane 3, probably reflect detrital minerals in the black slate and graywacke of the Taconic allochthon. The chromium-nickel association in clastic eugeosynclinal rocks has been used elsewhere as evidence of a volcanic component.

INTRODUCTION

This reconnaissance geochemical survey of drainage basin alluvium was conducted in the Glens Falls 1° × 2° quadrangle between 1982 and 1986 and covers about 18,130 km^2. The purpose of the regional survey was to identify regional geochemical anomalies that may guide subsequent detailed mineral-resource investigations. Throughout the world, regional geochemical surveys show that mining districts commonly are surrounded by broad, geochemically anomalous zones that produce slightly to moderately enriched metal contents in geochemical samples. These metal-enriched zones are termed "ore field" anomalies by some exploration geologists. Some of the world's most highly productive mineral deposits are located within these broad, regionally anomalous zones.

Tabulated analytical results and a sample locality map from this survey were recently released (Day and others, 1986). The results discussed here are based on emission spectrographic analyses of 1,286 nonmagnetic heavy-mineral samples that were collected during the regional reconnaissance program. Analytical results from coatings of iron and manganese oxides on stream pebbles, another form of sample medium that has possible exploration applications, are available in Welsch and Watts (1986).

Isopleth maps that depict element distribution patterns over tens of square kilometers were prepared for several key elements whose enrichments are widespread enough to allow contouring. Typically the elongation of isopleth patterns is consistent in trend for two or more elements, and the isopleth distributions are coextensive with regional fault zones, lithostratigraphic units, structural domes, arch axes, and syn- and postkinematic igneous bodies. Genetic ties between the geologic features and the geochemical anomalies are therefore suggested. The elements that form widespread anomalies include those that (1) form metallic deposits (Pb, Cu, Fe, Co, W, and Sn), (2) are found in gangue minerals (Ba), and (3) characterize geochemically specialized rocks or favorable ore-forming environments (Nb, B, and Sr).

Anomalous concentrations of Ag, Au, Bi, Mo, As, Sb, Be, Ni, Cr, and Zn, all elements of mineral-resource interest, are present in the Glens Falls quadrangle, but their scattered distribution does not allow contouring. To some extent at least, the distribution of these element anomalies is reflected in the plots of element suites as determined from R-mode factor analysis.

Although results of the small-scale (1:250,000) geochemical coverage of the Glens Falls quadrangle no doubt contribute to the recognition of broad metallized zones, problems were encountered with the sampling and interpretation. The efficient coverage of large areas in reconnaissance by an alluvial drainage basin survey was hampered in the Glens Falls quadrangle by (1) transported overburden of glacial origin, (2) locally low relief and poorly developed drainage systems that made sampling difficult and, in some cases, impossible, and (3) anthropogenic contamination of the stream alluvium. These factors resulted in local data gaps and conceivably influenced the geochemical signature enough to introduce errors into the interpretations. These problems notwithstanding, the resultant geochemical data are probably most influenced by bedrock geology of the drainage basins. On this basis, interpretations of the geochemistry with respect to regional and local geology suggest new exploration targets for undiscovered metallic mineral deposits.

GEOLOGY

To relate the regional geochemistry to a geologic framework, the Glens Falls 1° × 2° quadrangle is separated into seven lithotectonic terranes (see articles by Thompson and McLelland in this volume for details on the geology of the quadrangle). On figure 1, each of the seven terranes is designated by a number. From west to east, the terranes are (1) Proterozoic (Grenville-age) basement rocks of the Adirondack massif, which include felsic gneiss, quartzite, and minor amphibolite, graphitic schist, and calc-silicate rocks; marble, metagabbro, charnockite, mangerite, and meta-anorthosite are present within the Adirondack highlands; (2) a Cambrian-Ordovician shelf sequence composed largely of autochthonous carbonate and quartzite and parautochthonous, Middle Ordovician Normanskill-equivalent black slate and phyllite in the Vermont valley and carbonate, quartzite, and Normanskill-equivalent shale and graywacke in the Champlain-Hudson valley; (3) an allochthonous-parautochthonous transitional sequence of slope-basin deposits consisting of black, green, and red slate, phyllite, graywacke, and minor carbonate and quartzite of Late Proterozoic to Ordovician age that constitutes the Taconic allochthon of eastern New York and western Vermont; (4) Proterozoic core rocks (also Grenville age) consisting of felsic gneiss, schist, and lesser amphibolite, quartzite, and calc-silicate rocks and Late Proterozoic to Cambrian(?) cover rocks of schist, metagraywacke, and minor amphibolite and dolomite, all within the Green Mountain anticlinorium and the Chester and Athens domes of Vermont; (5) a largely eugeosynclinal Late Proterozoic to Early Devonian sequence of carbonaceous, calcareous, and quartzose mica schist, minor amphibolite and greenstone, minor steatite and serpentinite (the talc belt), and minor felsic metavolcanic rocks east of the Green Mountains in Vermont; (6) a largely Ordovician to Early Devonian sequence, mostly east of the Connecticut River in western New Hampshire, of calcareous, carbonaceous, and quartzose metasedimentary rocks, felsic and mafic metavolcanic rocks, and silicic plutonic rocks, including the mantled gneiss domes (Oliverian) and Ammonoosuc Volcanics that together compose the Bronson Hill anticlinorium; and (7) an area dominated by syn- to postkinematic granites of the Devonian New Hampshire Plutonic Suite, consisting of the Kinsman Quartz Monzonite, the Sunapee two-mica granite, and the Bethlehem Gneiss and associated pegmatites.

Two of the larger postkinematic alkalic intrusives of the Mesozoic White Mountain Plutonic-Volcanic Suite in New England are at Cuttingsville and Mount Ascutney, Vt. The Mesozoic plutons crop out over small areas but are geochemically and metallogenically significant. The small Devonian(?) Brocklebank granite pluton of the New Hampshire Suite, just to the west of the Orange County copper district in Vermont (see Doll and others, 1961, and Slack and others, this volume), is of key relevance to geochemical interpretations in the northeastern part of the study area.

Much of the area of the geochemical survey is covered by glacial, glaciofluvial, glaciolacustrine, and eolian materials that were deposited during the Pleistocene. The surficial deposits left by this glaciation have concealed much of the bedrock, except in the higher elevations, in riverbeds and banks, in the local ledge of outcrop on hillsides, and in manmade exposures in roadcuts. Bedrock in the higher elevations, if unexposed, is typically covered by only a thin veneer of ablation till. The large stream and river valleys contain the greatest thickness of glacial deposits (Larsen, 1972, 1987). During sampling, some small north-facing tributaries were seen to have thick-plastered accumulations of compact basal till filling a large part, if not all, of the drainage basin. In these cases, the majority of sample material clearly is composed of till-derived materials. However, most of this till is thought to be from local bedrock sources, so that the impact on the interpretations of the geochemistry is probably minimal, as far as the regional scale is concerned.

METHODS

Sample Collection, Preparation, and Analysis

The 1,286 nonmagnetic heavy-mineral concentrates were beneficiated from 9- to 13-kg samples of bulk stream sediment collected from small streams that have typical lengths of 2 to 5 km, but locally as much as 7 km (fig. 1). Throughout this report, these samples are referred to as

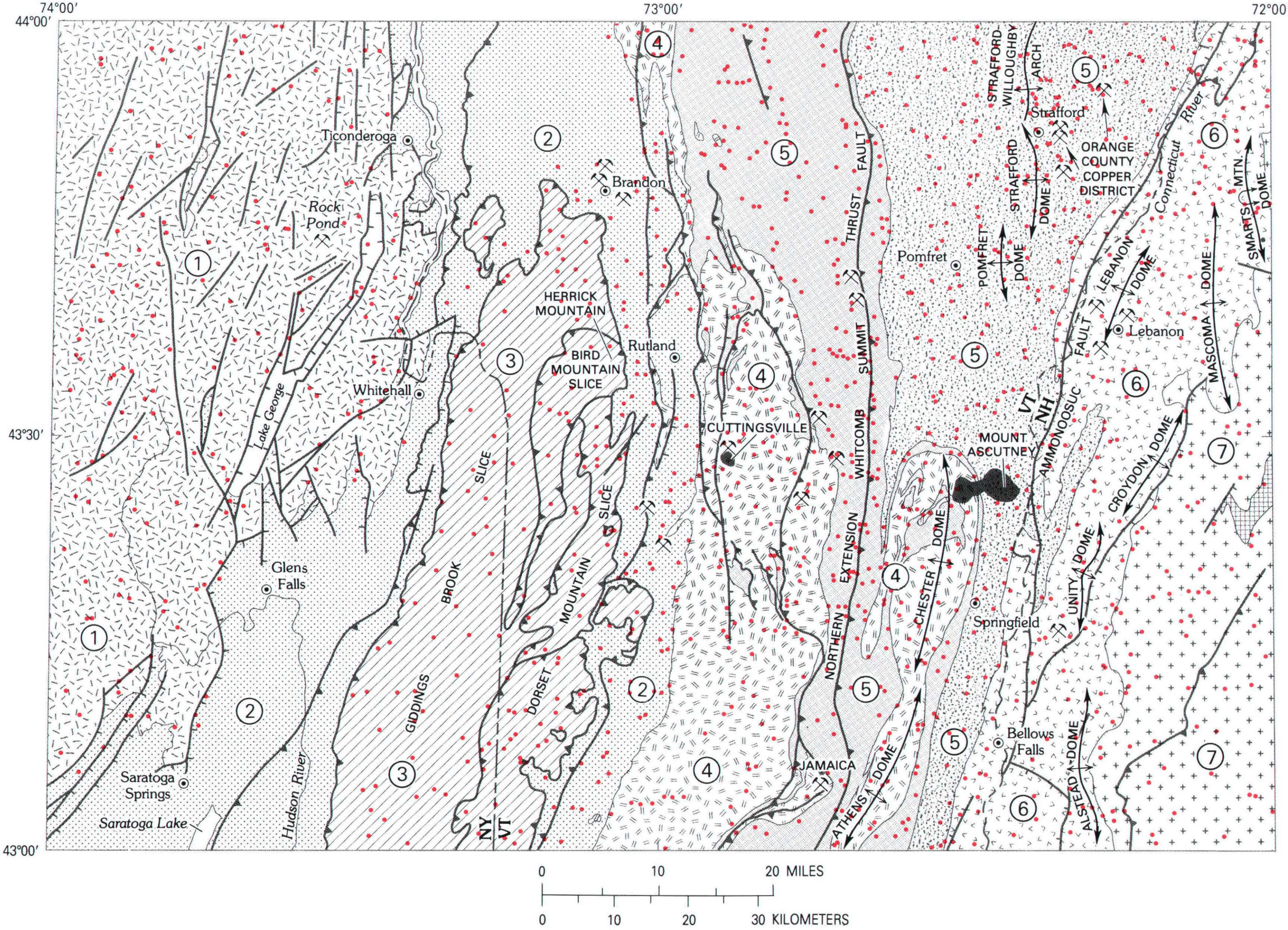

Figure 1. Sample localities (dots), tectono-stratigraphic terranes (circled numbers 1–7), and some selected geologic and geographic features. Note that Terrane 5 here includes both the eastern Green Mountain terrane and the Connecticut Valley-Gaspé terrane as shown on the geologic base (Thompson, this volume).

EXPLANATION

CRETACEOUS AND JURASSIC(?)	■	White Mountain Plutonic-Volcanic Suite (syenite, granite, and gabbro)
UPPER DEVONIAN	▦	Granite plutons (two-mica granite)
LOWER DEVONIAN AND SILURIAN	+	Merrimack terrane (granitic gneiss, migmatite, minor pelitic schist, and quartzite)
	▒	Connecticut Valley-Gaspé terrane (pelitic schist, calcareous schist, and minor amphibolite)
ORDOVICIAN AND CAMBRIAN	░	Parautochthonous shelf sequence (limestone, dolomite, and quartzite)
DEVONIAN, SILURIAN, AND ORDOVICIAN	▢	Bronson Hill terrane (metavolcanic rocks, pelitic schist, and granitic gneiss)
ORDOVICIAN, CAMBRIAN, AND PROTEROZOIC Z	▨	Taconic allochthon (slate, graywacke, and minor carbonate)
	□	Eastern Green Mountain terrane (pelitic schist, metagraywacke, and minor amphibolite)
PROTEROZOIC Y	▧	Adirondack massif (felsic gneiss, quartzite, anorthosite, amphibolite, calc-silicate rocks, and metagabbro)
	▢	Green Mountain massif[1] (felsic gneiss, quartzite, minor amphibolite, and calc-silicate rocks)
	——┴┴	High-angle fault—Tick marks on downthrown side where known
	—▲——▲—	Thrust fault—Sawteeth on upper plate

[1]Includes Lincoln Mountain massif and Proterozoic rocks of the Chester and Athens domes.

Figure 1. Continued.

stream sediment concentrates. Areas covered by extensive overburden of stratified glacial materials of fluvial or lacustrine origin, which can have very complex and distal sources, were not sampled. For example, two large areas of glaciolacustrine deposits in the Champlain valley (Connally, 1970, 1973) were unsampled. To avoid stratified glacial deposits, sample sites were located above the 150-m elevation contour and on tributary streams, not major rivers.

The samples consist of bulk composites gathered from various positions in active streambeds by using long-handled garden shovels as close to the bedrock-sediment interface as possible. Concentration was on site by use of conventional gold panning techniques. Samples were then shipped to U.S. Geological Survey (USGS) laboratories for standard heavy-mineral processing using bromoform (specific gravity 2.86). After treatment with bromoform, samples were electromagnetically separated into three fractions.

The nonmagnetic fraction provided the principal data for this report. After the electromagnetic separations were made, the nonmagnetic fraction was split into a mineralogic fraction, which was archived, and into an analytical fraction. The mineralogic fraction was studied by binocular microscope and, in some cases, by X-ray diffraction and scanning-electron microscope (SEM) methods. The analytical fraction was pulverized to a fine powder before analysis. The powders were then analyzed by routine emission spectrographic methods (Grimes and Marranzino, 1968). Results of these spectrographic analyses were measured within geometric intervals (for example, boundaries at 1,200, 830, 560, 380, 260, 180, 120, and 83 parts per million (ppm)) but were reported as the approximate geometric midpoints (1,000, 700, 500, 300, 200, 150, and 100 ppm). The spectrographic data are reported as a series of six steps per order of magnitude. Reported values fall within one adjoining reporting interval 83 percent of the time and two adjoining reporting intervals 96 percent of the time (Motooka and Grimes, 1976).

Data Reduction and Analysis

As a preliminary evaluation procedure, the analytical data were statistically reduced and summarized (tables 1–8) by using standard USGS STATPAC (statistical package) computer programs (Van Trump and Miesch, 1977). The median value for each element for samples from the entire Glens Falls quadrangle (table 1) was compared with the median values for each element in each terrane (tables 2–8). Comparisons of the data indicate that, for most elements, differences are within the analytical variations inherent in the spectrographic reporting method (Motooka and Grimes, 1976) and are probably consistent with the sampling precision as well; therefore, for practical purposes, the regional median value is probably a valid geochemical threshold for all terranes. The threshold value represents the overlap of reported values considered "high background" and "weakly anomalous." The ratio of reported values to this median or threshold is a measure of "anomaly intensity" or anomaly-to-background contrast. As a result, the relative intensity of the chemical elements in each terrane can be evaluated, and terranes can be compared geochemically to the region as a whole.

Some of the key elements also were examined for statistical populations by using cumulative frequency plots

Table 1. Statistical summary of analytical data for nonmagnetic concentrates from all terranes

[Elements sought in emission spectrographic analysis but not shown in table include cadmium, scandium, zirconium, and thorium. Iron, magnesium, calcium, and titanium, in percent; all other elements, in parts per million. L is detected but below lowest spectrographic standard, G is above upper limit, N is not detected; —, no data]

Element	Valid observations	1,286 samples Percentile distribution[1]							Range	Geometric mean[2]	Qualified observations		
		25	50	75	85	90	95	99			L	G	N
Antimony......	15	—	—	—	—	—	—	200	200–1,500	389	25	0	1,246
Arsenic........	22	—	—	—	—	—	—	700	500–7,000	1,141	0	0	1,264
Barium........	755	—	100	200	500	700	1,500	10,000	50–>10,000	233	109	12	410
Beryllium......	202	—	—	—	2	2	5	50	2–200	3.9	17	0	1,067
Bismuth.......	27	—	—	—	—	—	—	150	20–1,000	105	9	0	1,250
Boron.........	1,233	50	150	300	500	700	1,000	3,000	20–5,000	149	30	0	23
Calcium.......	1,285	5	5	10	12.5	20	20	50	.10–50	5.7	1	0	0
Chromium.....	1,229	70	100	150	200	200	300	300	20–7,000	102	0	0	57
Cobalt.........	505	—	—	10	20	30	50	100	10–300	18	0	0	781
Copper........	1,113	10	15	50	100	150	200	1,000	10–30,000	28	161	0	12
Gold..........	12	—	—	—	—	—	—	30	20–1,000	102	2	2	1,270
Iron...........	1,286	.50	1	2	2	2	5	7	.10–50	.83	0	0	0
Lanthanum.....	865	—	50	150	200	200	500	1,000	50–2,000	110	77	3	341
Lead..........	952	<20	30	100	150	200	700	2,000	20–30,000	68	121	0	213
Magnesium....	1,270	.10	.50	1	2	2	5	5	.05–20	.37	16	0	0
Manganese.....	1,286	150	200	700	1,000	1,500	1,500	2,000	20–7,000	304	0	0	0
Molybdenum...	8	—	—	—	—	—	—	—	15–500	72	2	0	1,276
Nickel.........	143	—	—	—	—	10	10	70	10–200	18	0	0	1,143
Niobium.......	877	—	70	100	150	200	200	500	50–700	97	85	0	324
Silver.........	22	—	—	—	—	—	—	7	2–1,500	15	0	0	1,264
Strontium......	917	—	200	500	700	700	1,000	1,500	200–2,000	408	1	0	368
Tin............	536	—	—	50	70	150	300	1,500	20–>2,000	66	38	4	708
Titanium.......	397	2	>2	>2	>2	>2	>2	>2	.10–>2	1.6	0	889	0
Tungsten......	157	—	—	—	100	100	200	1,000	100–20,000	245	61	0	1,068
Vanadium.....	1,253	50	150	200	300	300	300	500	20–2,000	109	32	0	1
Yttrium........	1,285	100	200	300	500	500	500	1,000	20–1,500	184	1	0	0
Zinc...........	12	—	—	—	—	—	—	500	500–5,000	718	3	1	1,270

[1] In spectrographic reporting intervals nearest the percentile.

[2] On the basis of reported values only.

(figs. 2 and 3). Most of the plots show two or more statistical populations, as delineated by breaks in the curves (see Lepeltier, 1969). The curves showing multiple statistical populations result from variations in the heavy-mineral suites in the samples and in their metal content and are not simply "anomalous" and "background" statistical populations. These variations must be evaluated from mineralogic, geochemical, and geologic contexts to relate them to possible mineral deposit locations or, indirectly, to specific geologic, geochemical, and (or) mineralogic environments.

The data were further statistically examined for element interrelations by correlation analysis and R-mode factor analysis by using a varimax rotation (Harman, 1960). The log-transformed data sets comprising all samples collected from the Glens Falls quadrangle (table 9), and subsets from each of the seven lithotectonic terranes were examined by R-mode analysis. Some reshifting of element associations or strength-of-factor loading of elements in each factor are seen when the data subsets are compared with the total data set, but the differences for a regional study are probably not sufficiently significant to justify separate treatment of each terrane.

A computer program called Relative Element Magnitude (REM) developed by George Van Trump, Jr. and H.V. Alminas of the USGS (Van Trump and Alminas, 1978) was used to further define the element associations, intensity of geochemical contrast, and relative area encompassed by element anomalies. The REM analysis was done within linear zones or trends of regional anomalies or within polygonal anomalous areas referred to as "windows" (see tables 10–25).

The REM computations (see tables 10–24) are based on the measurement of two parameters for each element and include all elements sought in the spectrographic analyses. The first parameter determined by the REM analysis is the intensity factor (or anomaly-to-background contrast), which is the ratio of mean value of anomalous samples to threshold value (median for the element) computed for each window area (window area is the block of terrane being analyzed, which is delineated on the basis of the widest extent of an isopleth pattern). The second parameter is the areal factor, which is the ratio of the number of anomalous samples to the total number of samples in the window area. Therefore, the actual square-kilometer area that an anomalous zone

Table 2. Statistical summary of analytical data for nonmagnetic concentrates from Terrane 1

[Elements sought in emission spectrographic analysis but not shown in table include cadmium, scandium, zirconium, and thorium. Iron, magnesium, calcium, titanium, in percent; all other elements, in parts per million. L is detected but below lowest spectrographic standard, G is above upper limit, N is not detected; —, no data]

Element	Valid observations	202 samples Percentile distribution[1]							Range	Geometric mean[2]	Qualified observations		
		25	50	75	85	90	95	99			L	G	N
Antimony	1	—	—	—	—	—	—	—	200	—	11	0	190
Arsenic	0	—	—	—	—	—	—	—	—	—	0	0	202
Barium	103	—	50	200	200	300	500	700	50–1,000	158	25	0	74
Beryllium	24	—	—	—	<2	2	2	2	2–7	2.1	14	0	164
Bismuth	1	—	—	—	—	—	—	—	20	—	1	0	200
Boron	180	50	100	200	200	500	500	1,000	20–1,000	99	22	0	0
Calcium	202	5	5	7	10	10	10	20	.5–20	5.5	0	0	0
Chromium	175	20	50	70	100	100	200	300	20–500	60	0	0	27
Cobalt	37	—	—	—	10	15	25	50	10–150	18	0	0	165
Copper	116	<10	10	10	10	20	50	100	10–200	14	74	0	12
Gold	0	—	—	—	—	—	—	—	—	—	0	0	202
Iron	202	.3	.5	1	2	2	5	5	.1–7	.68	0	0	0
Lanthanum	184	50	100	200	500	500	1,000	2,000	50–2,000	139	1	0	14
Lead	92	—	<20	20	30	30	50	300	20–500	28	53	0	57
Magnesium	200	.59	2	5	5	5	5	10	.05–10	1.3	2	0	0
Manganese	202	150	200	500	500	700	1,000	1,500	50–2,000	227	0	0	0
Molybdenum	1	—	—	—	—	—	—	—	20	—	1	0	200
Nickel	13	—	—	—	—	—	30	70	30–100	43	0	0	189
Niobium	94	—	<50	60	70	70	70	100	50–150	62	28	0	80
Silver	1	—	—	—	—	—	—	—	7	—	0	0	201
Strontium	90	—	—	200	200	200	300	500	200–500	227	1	0	111
Tin	49	—	—	<20	20	50	70	200	20–1,500	41	23	0	130
Titanium	79	2	>2	>2	>2	>2	>2	>2	.1–>2	1.2	0	123	0
Tungsten	6	—	—	—	—	—	—	150	100–300	154	0	0	196
Vanadium	190	70	100	150	150	150	200	200	20–200	94	12	0	0
Yttrium	202	150	200	300	500	500	500	700	200–1,500	201	0	0	0
Zinc	1	—	—	—	—	—	—	—	500	—	0	0	201

[1] In spectrographic reporting intervals nearest the percentile.

[2] On the basis of reported values only.

may encompass is a function of the sample density and can be approximated by multiplying average sample density by number of anomalous samples. These two parameters, intensity and area (in terms of proportion of anomalous samples), are then multiplied for each element and give a product called the element magnitude (EM). Then, all individual EM values are summed to give the total magnitude for the window area; the individual EM for each element is then ratioed to the total and expressed in relative percent, as the REM. For these computations (although there is little difference in median values between terranes), if a window area is totally within a lithotectonic terrane, the median value for that terrane (tables 2–8) was used as a threshold, or a lower limit of anomalous values; if the window area cuts across terranes, the median value for the total Glens Falls quadrangle was used (table 1).

For both the R-mode computations and the REM computations, elements having qualified reporting values (that is, detected, not detected, and detected but below lowest spectrographic reporting interval or above highest detection limit) were cleared of these qualified readings by substitution of reasonable values. These assigned values were based on one spectrographic reporting interval above the upper detection limit for values above the upper limit, one interval below for readings of detected reportings (but below the lower limit), and two reporting intervals below the detection limit for undetected reportings. Zinc and tungsten, as two examples, have detection limits far above the expected normal range of values; in these cases, samples that contain an element reported as undetected were assigned values of crustal abundance for that element.

Map Preparation

The isopleth maps were prepared from computer programs developed by Jack B. Fife of the USGS. The plotting routine of these programs uses a circular search window that is passed over a gridded area, and the distance-weighted average of all sample values within the circle is calculated by using a cell averaging technique. These values are posted at grid intersections and are then computer contoured. Because the grid value depends on distance, weighted sample sites closest to a grid intersection influence

Table 3. Statistical summary of analytical data for nonmagnetic concentrates from Terrane 2

[Elements sought in emission spectrographic analysis but not shown in table include cadmium, scandium, zirconium, and thorium. Iron, magnesium, calcium, and titanium, in percent; all other elements, in parts per million. L is detected but below lowest spectrographic standard, G is above upper limit, N is not detected; —, no data]

Element	Valid observations	54 samples Percentile distribution[1] 25	50	75	85	90	95	99	Range	Geometric mean[2]	Qualified observations L	G	N
Antimony......	2	—	—	—	—	—	—	250	200–300	—	0	0	52
Arsenic........	0	—	—	—	—	—	—	—	—	—	0	0	54
Barium........	24	—	<50	300	500	500	700	1,000	50–1,000	319	10	0	20
Beryllium......	4	—	—	—	—	—	2	415	2–7	2.7	0	0	50
Bismuth.......	0	—	—	—	—	—	—	—	—	—	0	0	54
Boron.........	46	70	200	500	1,000	1,500	2,000	2,000	20–2,000	259	0	0	8
Calcium.......	54	5	5	5	10	10	10	10	.1–10	3.7	0	0	0
Chromium.....	54	70	100	150	150	150	150	150	20–150	90	0	0	0
Cobalt.........	3	—	—	—	—	—	10	40	10–50	25	0	0	51
Copper........	50	10	10	2.0	70	100	150	1,000	10–2,000	19	4	0	0
Gold..........	0	—	—	—	—	—	—	—	>1,000	—	0	1	53
Iron...........	54	.5	.5	1	1	1.5	2	7	.1–7	.52	0	0	0
Lanthanum.....	40	50	50	150	150	200	200	1,000	50–1,000	95	5	0	0
Lead..........	49	20	50	100	175	200	700	1,000	20–1,500	61	3	0	2
Magnesium....	54	.1	.5	1	1	2	2	5	.05–5	.28	0	0	0
Manganese.....	54	100	100	150	200	200	200	500	20–500	112	0	0	0
Molybdenum...	0	—	—	—	—	—	—	—	—	—	0	0	0
Nickel.........	0	—	—	—	—	—	—	—	—	—	0	0	0
Niobium.......	15	—	—	50	150	150	200	200	50–200	109	8	0	31
Silver.........	1	—	—	—	—	—	—	—	7	—	0	0	53
Strontium......	25	—	—	200	500	500	700	1,000	200–1,000	318	0	0	29
Tin............	15	—	—	30	60	70	200	2,000	20–1,500	74	2	0	36
Titanium.......	23	2	>2	>2	>2	>2	>2	>2	1–>2	1.9	0	31	0
Tungsten......	0	—	—	—	—	—	—	—	—	—	0	0	0
Vanadium.....	52	50	70	100	150	150	150	150	20–150	58	2	0	0
Yttrium........	54	200	300	500	700	700	700	1,000	100–1,000	336	0	0	0
Zinc...........	0	—	—	—	—	—	—	—	—	—	0	0	0

[1] In spectrographic reporting intervals nearest the percentile.

[2] On the basis of reported values only.

the posted value more than those values at a greater distance, and so the chosen grid size affects the amount of smoothing of the data; the larger the grid cell, the greater the smoothing. For these isopleth maps, the grid cell size is about 3.4 km on a side.

The contour intervals on the geochemical maps are nearly logarithmic (Grimes and Marranzino, 1968) and consist of emission spectrographic reporting intervals of values above the median. The methods of reporting spectrographic analyses and the truncated, or censored, distribution of the data for some elements are not amenable to a rigorous percentile-based contour interval. Where the element distribution allows, however, the spectrographic reporting intervals used for contouring correspond approximately to the 50th, 75th, 85th, 90th, and 99th percentile of abundance distribution for each element.

Because there is little significant difference in median values between terranes, the regional median was selected as the threshold value (lowest contour interval) for each key element. Had there been significant differences in median values between terranes, separate threshold values would have been assigned. As a result, contour patterns cut across terrane boundaries and by so doing suggest patterns of geochemical overprinting caused by igneous activity and (or) mineralization.

R-mode scores for the seven factors determined for the Glens Falls region are shown on perspective diagrams (see figs. 16–22). These diagrams were prepared in a similar manner to that of the isopleth maps of the individual elements, by using the same cell size but the full range of positive factor scores. Such diagrams are useful in showing the distribution of an element association which can in some cases be of greater significance than the distribution of any single element.

SAMPLE MINERALOGY

The samples show a varied and complex mineralogy that includes rutile, tourmaline, zircon, apatite, pyrite, pale-colored epidote-group minerals (for example, clinozoisite), muscovite, chlorite, sphene, carbonate minerals, diopside, sillimanite, wollastonite, andalusite, staurolite, kyanite, tremolite-actinolite, phlogopite, and leucoxene.

Table 4. Statistical summary of analytical data for nonmagnetic concentrates from Terrane 3

[Elements sought in emission spectrographic analysis but not shown in table include cadmium, scandium, zirconium, and thorium. Iron, magnesium, calcium, and titanium, in percent; all other elements, in parts per million. L is detected but below lowest spectrographic standard, G is above upper limit, N is not detected; —, no data]

Element	Valid observations	171 samples Percentile distribution[1] 25	50	75	85	90	95	99	Range	Geometric mean[2]	Qualified observations L	G	N
Antimony	6	—	—	—	—	<200	<200	700	200–1,500	460	13	0	152
Arsenic	7	—	—	—	—	—	—	1,500	500–7,000	1,088	0	0	164
Barium	128	200	700	2,000	5,000	10,000	>10,000	>10,000	70–>10,000	895	13	12	18
Beryllium	28	—	—	—	—	—	—	—	2–50	2.2	2	0	141
Bismuth	2	—	—	—	—	—	—	30	30–50	39	5	0	164
Boron	171	100	200	200	500	500	500	2,000	20–2,000	166	0	0	0
Calcium	171	5	5	7	10	10	15	20	.5–20	4.7	0	0	0
Chromium	170	70	100	150	200	200	300	1,000	20–2,000	103	0	0	1
Cobalt	45	—	—	10	20	30	50	100	10–100	24	0	0	126
Copper	155	10	20	70	100	200	300	3,000	10–7,000	34	16	0	0
Gold	3	—	—	—	—	—	—	30	30–200	56	1	0	167
Iron	171	.5	1	2	2	2	7	20	.10–30	.96	0	0	0
Lanthanum	135	50	100	150	200	200	500	1,000	50–2,000	115	11	0	25
Lead	166	30	100	300	700	1,000	1,500	7,000	20–10,000	124	4	0	1
Magnesium	168	.2	.5	1	1	1.5	2	2	.05–3	.38	3	0	0
Manganese	171	100	150	200	200	200	500	700	50–7,000	151	0	0	0
Molybdenum	2	—	—	—	—	—	—	15	15–30	21	1	0	168
Nickel	22	—	—	—	—	30	70	100	10–150	46	0	0	149
Niobium	100	—	50	70	100	150	150	150	50–200	78	13	0	58
Silver	5	—	—	—	—	—	—	7	5–15	6.7	0	0	166
Strontium	144	200	500	500	700	700	700	1,000	200–1,000	406	0	0	27
Tin	101	—	20	200	300	700	1,000	2,000	20–2,000	128	8	0	61
Titanium	64	2	>2	>2	>2	>2	>2	>2	.1–>2	1.6	0	107	0
Tungsten	0	—	—	—	—	—	—	—	—	—	—	0	171
Vanadium	166	50	70	100	150	150	150	150	20–200	65	4	0	1
Yttrium	171	200	300	300	500	500	500	1,000	100–1,000	283	0	0	0
Zinc	7	—	—	—	—	—	<500	700	500–1,000	638	3	0	161

[1] In spectrographic reporting intervals nearest the percentile.

[2] On the basis of reported values only.

Barite is abundant in samples from the Taconic allochthon of New York and Vermont but is scarce elsewhere. Less commonly seen, but locally recognized in trace to large amounts, are anatase (again, chiefly from the Taconics), spinel (either gahnite or Cr-spinel), scheelite, chalcopyrite, gold, and arsenopyrite. Pyrite is by far the most abundant sulfide mineral in the samples, and it may be the source of many of the geochemical anomalies.

Relating elevated metal contents found in heavy-mineral samples to mineralogic association may be useful in metallogenic interpretations. Elevated metal contents may be related to some of the following: (1) major or minor constituents of ore and gangue minerals; (2) metal-rich intergrowths, inclusions, or coatings; and (3) constituents of common rock-forming and rock-accessory minerals resulting from diadochic substitutions and other forms of camouflaging of minor metals in crystal lattices. High metal contents in common minerals may reflect unusual bulk compositions of host rocks brought about by premetamorphic hydrothermal alteration or by mineralization. As an example, zinc-rich staurolite may be derived from a metamorphosed massive sulfide deposit (Spry and Scott, 1986). In other cases, elements such as tin and niobium may characterize granitoid rocks that have rare- and precious-metal potential (Robinson, this volume; Cox, this volume). Thus, many element anomalies are probably associated with common rock-forming and accessory minerals and involve, in some cases, elements not usually considered to be important in the exploration for metallic resources (for example, barium, calcium, boron, strontium, and manganese). It will be shown that regional distribution patterns of these elements, as much as the other key ore-forming elements, help define possible patterns of regional metallization in the Glens Falls quadrangle.

MAP DISTRIBUTION AND GEOCHEMICAL SIGNIFICANCE OF KEY ELEMENTS

The individual elements (see figs. 4–15) are discussed in this section on the basis of their mineralogic and geochemical characteristics and anomaly distribution within each of the terranes. The section on regional anomaly trends describes element associations, regional isopleth patterns, and possible implications for tectonics and mineralization.

Table 5. Statistical summary of analytical data for nonmagnetic concentrates from Terrane 4

[Elements sought in emission spectrographic analysis but not shown in table include cadmium, scandium, zirconium, and thorium. Iron, magnesium, calcium, and titanium, in percent; all other elements, in parts per million. L is detected but below lowest spectrographic standard, G is above upper limit, N is not detected; —, no data]

Element	Valid observations	207 samples Percentile distribution[1]							Range	Geometric mean[2]	Qualified observations		
		25	50	75	85	90	95	99			L	G	N
Antimony......	2	—	—	—	—	—	—	<200	500–700	592	1	0	204
Arsenic........	0	—	—	—	—	—	—	—	—	—	0	0	207
Barium........	113	—	70	200	500	500	500	1,500	50–7,000	197	27	0	67
Beryllium......	4	—	—	—	—	—	—	—	2	—	0	0	203
Bismuth.......	3	—	—	—	—	—	—	20	20–200	84	1	0	203
Boron.........	198	50	150	500	700	1,000	2,000	5,000	20–5,000	166	9	0	0
Calcium.......	207	5	10	20	20	20	20	50	.2–50	9.5	0	0	0
Chromium.....	195	50	70	100	150	150	200	300	20–700	81	0	0	12
Cobalt.........	91	—	—	10	20	20	30	50	10–300	15	0	0	116
Copper........	187	10	15	30	50	70	150	500	10–3,000	23	20	0	0
Gold..........	0	—	—	—	—	—	—	—	—	—	0	0	207
Iron...........	207	.50	1	1.5	2	2	2	2	.10–10	.76	0	0	0
Lanthanum.....	155	50	70	200	300	300	300	500	50–500	127	14	0	38
Lead..........	175	20	50	150	150	150	500	2,000	20–7,000	68	23	0	9
Magnesium....	202	.20	1	2	2	2	5	6	.05–10	.60	5	0	0
Manganese.....	207	150	200	700	1,000	1,500	1,500	1,500	50–7,000	335	0	0	0
Molybdenum...	1	—	—	—	—	—	—	—	70	—	0	0	206
Nickel.........	1	—	—	—	—	—	—	—	20	—	0	0	206
Niobium.......	139	<50	50	70	100	150	150	200	50–700	76	16	0	52
Silver.........	2	—	—	—	—	—	—	—	7–30	14	0	0	205
Strontium......	170	200	500	500	700	700	1,500	1,500	200–2,000	401	0	0	37
Tin............	93	—	—	30	50	70	100	300	20–500	41	7	0	107
Titanium.......	50	2	>2	>2	>2	>2	>2	>2	1–>2	1.8	0	157	0
Tungsten......	7	—	—	—	—	—	—	150	100–500	150	1	0	199
Vanadium.....	197	70	150	200	500	500	500	500	20–500	124	10	0	0
Yttrium........	207	500	500	500	500	500	700	700	70–1,000	328	0	0	0
Zinc...........	0	—	—	—	—	—	—	—	—	—	0	0	207

[1] In spectrographic reporting intervals nearest the percentile.

[2] On the basis of reported values only.

Lead

The greatest range of lead values is in samples from Terrane 6 (20–30,000 ppm, table 7). The geometric mean for the entire quadrangle is 68 ppm lead (table 1). The highest geometric mean is in samples from Terrane 3 (124 ppm, table 4), which is nearly twice the regional background value; samples from other terranes have geometric means near that of the regional background (tables 1–3, 5–8). A cumulative frequency plot (fig. 2) shows separate statistical populations of lead, occurring at breaks in the curve at 100, 150, 300, and 700 ppm. Background ranges of reported values (up to 70 ppm) reflect the lead contents of common rock-forming minerals. Lead values at slightly higher concentrations are due to intergrowths or coatings and to minor occurrences of lead minerals. Greater lead contents reflect increasing increments of ore minerals in the samples. Lead values above 700 ppm suggest that samples were collected near a high lead source derived from the local drainage basin bedrock. Lead enrichment may be attributed to at least three important factors: (1) the low chemical mobility of lead in the surficial zone; (2) the presence of lead in high-density primary and secondary minerals, leading to its concentration in the heavy-mineral samples; and (3) the widespread occurrence of lead in trace amounts in nearly all sulfide deposits.

The only lead mineral so far identified in the samples is minium (Pb_3O_4), which is found in one concentrate sample from Terrane 3. However, unrecognized primary sulfide minerals as well as their secondary oxidized derivatives are undoubtedly present. SEM investigations of some lead-rich samples from the Glens Falls quadrangle have shown that lead and sulfur occur in fine-grained mineral aggregates. Minium is a secondary mineral formed from the oxidation of galena and cerussite (Rankama and Sahama, 1950, p. 731) or a product of manmade contaminants. If the minium occurrence is the result of contamination, it is probably not a contributor to the widespread zones of anomalous lead typical of parts of the study area. Pyrite is very abundant in some samples and is probably a significant contributor to the lead anomalies. Locally, radiogenic lead in zircon may be responsible for some lead anomalies; this should be suspected in cases of isolated lead highs where there is no enrichment of other sulfide-forming elements (for example,

Table 6. Statistical summary of analytical data for nonmagnetic concentrates from Terrane 5

[Elements sought in emission spectrographic analysis but not shown in table include cadmium, scandium, zirconium, and thorium. Iron, magnesium, calcium, and titanium, in percent; all other elements, in parts per million. L is detected but below lowest spectrographic standard, G is above upper limit, N is not detected; —, no data]

Element	Valid observations	441 samples Percentile distribution[1]							Range	Geometric mean[2]	Qualified observations		
		25	50	75	85	90	95	99			L	G	N
Antimony	2	—	—	—	—	—	—	—	200	—	0	0	439
Arsenic	14	—	—	—	—	—	—	1,500	500–7,000	1,122	0	0	427
Barium	208	—	<50	100	150	200	300	700	50–5,000	128	34	0	199
Beryllium	46	—	—	—	—	2	2	7	2–30	2.9	1	0	394
Bismuth	10	—	—	—	—	—	—	125	20–500	112	3	0	428
Boron	421	50	150	500	500	700	1,000	3,000	20–5,000	148	5	0	15
Calcium	440	5	7	10	20	20	20	50	.2–50	6.4	1	0	0
Chromium	427	70	150	200	200	300	300	500	20–7,000	118	0	0	14
Cobalt	263	—	10	20	30	50	70	150	10–300	19	0	0	178
Copper	401	10	20	70	150	200	300	700	10–30,000	36	40	0	0
Gold	9	—	—	—	—	—	—	150	20–1,000	124	0	0	432
Iron	441	.5	1	2	2	2	5	7	.1–50	.92	0	0	0
Lanthanum	212	—	<50	70	100	150	200	500	50–700	88	25	0	204
Lead	340	20	30	100	150	200	500	1,500	20–15,000	65	25	0	76
Magnesium	434	.07	.2	.5	1	1	1.5	2	.05–3	.24	7	0	0
Manganese	441	150	500	1,500	1,500	1,500	2,000	5,000	50–7,000	404	0	0	0
Molybdenum	1	—	—	—	—	—	—	—	300	—	0	0	440
Nickel	92	—	—	—	10	10	15	70	10–200	14	0	0	349
Niobium	331	50	70	200	200	300	300	500	50–500	124	12	0	98
Silver	12	—	—	—	—	—	—	10	2–1,500	18	0	0	429
Strontium	342	200	500	700	700	1,000	1,500	2,000	200–2,000	475	0	0	99
Tin	177	—	—	50	50	70	150	300	20–1,000	50	2	0	262
Titanium	138	2	2	>2	>2	>2	>2	>2	.5–>2	1.7	0	303	0
Tungsten	38	—	—	—	<100	<100	150	500	100–1,000	186	49	0	354
Vanadium	438	50	150	300	300	300	300	500	20–700	118	3	0	0
Yttrium	440	70	100	200	200	300	300	700	20–700	114	1	0	0
Zinc	4	—	—	—	—	—	—	500	500–5,000	967	0	1	436

[1] In spectrographic reporting intervals nearest the percentile.

[2] On the basis of reported values only.

copper and zinc). Near Jamaica, Vt., anomalous, possibly radiogenic, lead defines a zone that encompasses an area of known uranium deposits (Ayuso and Ratté, this volume) and extends beyond into surrounding areas, perhaps outlining additional uranium resources.

Lead at levels above the regional threshold of 30 ppm is nearly absent from Terrane 1, yet this terrane contains at least four potential sources for lead-rich samples: (1) the Dixon Schist, a highly pyritic, quartz-feldspar schist that is host to many graphite deposits in the eastern Adirondacks (Alling, 1918, p. 48–51; McLelland, this volume); (2) stratabound Proterozoic magnetite deposits (Newland, 1908); (3) possible Balmat-type zinc±lead occurrences in carbonate sequences of Grenville rocks; and (4) Mesozoic fault zones.

Alling (1918, p. 69–70) reports intergrowths of pyrite, pyrrhotite, and galena in a gossanous feldspathic schist ("arkosite") at the Rock Pond graphite property (fig. 4). He also reports the presence of sphalerite and gahnite at other graphite deposits in the area. The host rock at the Rock Pond property is probably equivalent to the Dixon Schist and, from the description by Alling (1918), suggests a premetamorphic, syngenetic or diagenetic origin for the mineralization; there is also an indication of later epigenetic pyritization. Lead is not found in sediment samples from this area, but one sample contains anomalous boron and tin that are probably the products of Mesozoic mineralization.

The area delineated by lead isopleths along the eastern edge of Terrane 1 (fig. 4) may have sources in the Dixon Schist, which crops out in the area. A more likely possibility based on the close association of lead with the trends and intersections of Mesozoic fault zones suggests that the faults are the sources of high lead.

Only one small lead anomaly seems to have an association with the stratabound Proterozoic magnetite deposits in Terrane 1. This anomaly occurs to the west of the main zone of iron deposits in a setting similar to that of the previously mined magnetite bodies (Newland, 1908; McLelland, this volume). Although lead can be accommodated in the structure of magnetite, magnetite cannot be the source of the lead anomalies because it is removed from the samples during processing and before analysis.

The highest lead in Terrane 2 is at the northern end (fig. 4). Only one or two samples are involved in the

Table 7. Statistical summary of analytical data for nonmagnetic concentrates from Terrane 6

[Elements sought in emission spectrographic analysis but not shown in table include cadmium, scandium, zirconium, and thorium. Iron, magnesium, calcium, and titanium, in percent; all other elements, in parts per million. L is detected but below lowest spectrographic standard, G is above upper limit, N is not detected; —, no data]

Element	Valid observations	154 samples Percentile distribution[1]							Range	Geometric mean[2]	Qualified observations		
		25	50	75	85	90	95	99			L	G	N
Antimony......	2	—	—	—	—	—	—	300	300–1,500	—	0	0	152
Arsenic........	1	—	—	—	—	—	—	—	2,000	—	0	0	153
Barium........	104	—	150	300	300	300	500	700	100–2,000	232	0	0	50
Beryllium......	53	—	—	2	5	10	20	70	2–100	5.4	0	0	101
Bismuth.......	10	—	—	—	—	—	50	500	20–700	121	0	0	144
Boron.........	154	70	100	200	250	500	500	700	20–1,000	117	0	0	0
Calcium.......	154	2	5	5	10	10	10	20	.1–20	3.3	0	0	0
Chromium.....	149	100	150	200	200	300	300	300	50–300	149	0	0	5
Cobalt.........	37	—	—	—	10	12.5	20	70	10–70	16	0	0	117
Copper........	152	15	20	50	70	150	200	1,500	10–3,000	30	2	0	0
Gold..........	0	—	—	—	—	—	—	<20	—	—	1	0	152
Iron...........	154	.50	1	2	2	2	2	7	.1–7	.81	0	0	0
Lanthanum.....	87	—	50	70	150	150	200	500	50–500	85	19	0	48
Lead..........	94	—	20	50	150	200	700	10,000	20–30,000	61	6	0	54
Magnesium....	154	.10	.20	.20	.50	.50	.50	1.0	.05–1	.18	0	0	0
Manganese.....	154	175	500	700	1,000	1,000	1,500	1,500	50–2,000	354	0	0	0
Molybdenum...	3	—	—	—	—	—	—	300	15–500	155	0	0	151
Nickel.........	7	—	—	—	—	—	—	—	10	—	0	0	147
Niobium.......	148	70	100	150	200	200	300	500	50–500	108	4	0	2
Silver.........	2	—	—	—	—	—	—	3	5–1,500	—	0	0	152
Strontium......	104	—	200	700	700	700	1,000	1,500	200–1,500	488	0	0	50
Tin............	69	—	—	70	150	200	700	1,500	20–>2,000	86	0	2	83
Titanium.......	27	2	>2	>2	>2	>2	>2	>2	1.5–>2	2	0	127	0
Tungsten......	91	—	100	300	500	1,000	1,500	2,000	100–20,000	297	8	0	55
Vanadium.....	153	100	150	200	200	500	500	700	20–700	144	1	0	0
Yttrium........	154	70	150	200	200	300	300	500	30–1,000	130	0	0	0
Zinc...........	0	—	—	—	—	—	—	—	—	—	0	0	154

[1] In spectrographic reporting intervals nearest the percentile.

[2] On the basis of reported values only.

anomaly. The basal Cambrian-Ordovician sequence that hosts the stratabound zinc-lead-copper deposits near Brandon (Clark, this volume) is encompassed by the isopleth pattern, and similar base-metal occurrences may be contributing to the anomaly.

Southwest of Cuttingsville, Vt., lead isopleths delineate the interface of the basal Cambrian Cheshire Quartzite and underlying Proterozoic rocks of the Green Mountain massif. As a result, it is difficult to determine if Paleozoic or Proterozoic terranes, or both, are the sources of the lead-rich samples there. Because the geochemical patterns seem to crosscut rock types, epigenetic veins are the most likely mode of occurrence. Stratabound lead, similar to the sandstone-hosted lead deposits at Laisvall, Sweden (Rickard and others, 1979), may be present in the Cheshire Quartzite (Slack and Mory, 1983; Slack, this volume). Two rock-chip samples of Cheshire Quartzite collected from the anomalous zone have coarse, presumably syngenetic-diagenetic pyrite that has a slightly elevated silver content as high as 3 ppm, barium up to 1,000 ppm, molybdenum up to 5 ppm, and zinc up to 90 ppm; these relatively low values suggest that, excepting iron, the pyrite is largely free of metals. There is no significant enrichment of lead, as the highest amount found is only 20 ppm. Such low values imply that sedimentary pyrite is not the source of the lead anomalies. The lead isopleths in this zone are part of larger regional patterns of lead anomalies of northwesterly and easterly trend that may relate to epigenetic mineralization.

The zone of high lead in Proterozoic basement rocks delineated by isopleths southwest of the Cuttingsville stock (see Robinson, this volume) also includes within the same isopleth pattern the mineralization at Cuttingsville and suggests discordant sources for the lead and an origin similar to that of the Cuttingsville mineralization. This interpretation is supported by the lack of major lead enrichments in Proterozoic rocks elsewhere in the Green Mountain massif (fig. 4; Slack and others, 1985; J.D. Peper, U.S. Geological Survey, written commun., 1989).

Terrane 3 shows some of the strongest anomaly-to-background contrast and the most areally broad anomaly patterns for lead of any terrane in the Glens Falls quadrangle (fig. 4). In effect, the isopleths separate this terrane into a northern anomalous zone within the Giddings Brook-Bird Mountain thrust slices (see Zen, 1961, 1967) and a southern

Table 8. Statistical summary of analytical data for nonmagnetic concentrates from Terrane 7

[Elements sought in emission spectrographic analysis but not shown in table include cadmium, scandium, zirconium, and thorium. Iron, magnesium, calcium, and titanium, in percent; all other elements, in parts per million. L is detected but below lowest spectrographic standard, G is above upper limit, N is not detected; —, no data]

		56 samples Percentile distribution[1]									Qualified observations		
Element	Valid observations	25	50	75	85	90	95	99	Range	Geometric mean[2]	L	G	N
Antimony......	0	—	—	—	—	—	—	—	—	—	0	0	56
Arsenic........	0	—	—	—	—	—	—	—	—	—	0	0	56
Barium........	51	150	300	500	500	500	500	1,500	50–2,000	260	0	0	5
Beryllium......	42	2	5	10	50	50	100	200	2–200	8.2	0	0	14
Bismuth.......	2	—	—	—	—	—	—	500	20–1,000	141	0	0	54
Boron.........	56	70	150	300	500	500	700	1,000	20–1,000	137	0	0	0
Calcium.......	56	2	5	10	10	15	20	20	.20–20	4.4	0	0	0
Chromium.....	55	70	100	150	150	200	200	300	50–300	98	0	0	1
Cobalt.........	21	—	—	10	10	15	20	30	10–30	12	0	0	35
Copper........	50	10	20	30	50	50	100	1,000	10–1,500	22	6	0	0
Gold..........	0	—	—	—	—	—	—	—	—	—	0	0	56
Iron...........	56	.50	.85	1	2	2	2	5	.20–5	.66	0	0	0
Lanthanum.....	47	50	70	300	700	700	700	1,000	50–1,000	143	1	0	8
Lead..........	30	<20	20	20	30	30	150	500	20–500	32	14	0	12
Magnesium....	55	.10	.18	.20	.20	.50	.50	.70	.05–.70	.15	1	0	0
Manganese.....	56	300	700	700	1,000	1,000	1,000	1,500	100–1,500	486	0	0	0
Molybdenum...	0	—	—	—	—	—	—	—	—	—	0	0	56
Nickel.........	4	—	—	—	—	—	—	—	10	—	0	0	52
Niobium.......	42	50	50	100	100	150	150	200	50–200	78	7	0	7
Silver.........	0	—	—	—	—	—	—	—	—	—	0	0	56
Strontium......	35	—	200	200	200	200	200	300	200–500	205	0	0	21
Tin............	28	—	—	150	300	500	1,500	2,000	20–2,000	182	0	0	28
Titanium.......	15	>2	>2	>2	>2	>2	>2	>2	1–2	1.7	0	41	0
Tungsten......	13	—	—	<100	150	200	500	1,000	100–1,000	263	2	0	41
Vanadium.....	56	150	150	200	200	300	500	500	20–500	153	0	0	0
Yttrium........	56	150	500	700	700	700	1,000	1,500	30–1,500	312	0	0	0
Zinc...........	0	—	—	—	—	—	—	—	—	—	0	0	56

[1] In spectrographic reporting intervals nearest the percentile.

[2] On the basis of reported values only.

zone in which anomalies occupy both the low and high Taconic sequences (Giddings Brook-Dorset Mountain slices). The isopleths crosscut the different thrust slices, and yet their shape and abutment against major faults suggest that lead enrichment of the source rocks occurred prior to thrust faulting. In the southern part of this terrane, anomalous lead occurs in samples from the leading (western) edge of the Taconic allochthon and in a broad north-south-trending zone along the eastern edge of the Dorset Mountain thrust slice. It is possible that some of the lead in these samples is derived from pyritic and graphitic phyllites of the parautochthonous Ira (Hortonville) Formation.

At the northern end of the Taconic allochthon near Herrick Mountain, a large zone of lead anomalies occurs in the Giddings Brook-Bird Mountain slices of the lower Taconic sequence. The northwesterly trending isopleth pattern here is part of the quadrangle-wide, crosscutting pattern of northwesterly trending geochemical anomalies and radar lineaments. Galena is found locally in quartz veins of the Cambrian slates in this area (Dale 1899, p. 260).

A linear, north-trending isopleth pattern for lead parallels the contact of Proterozoic basement and Late Proterozoic-Cambrian cover rocks on the eastern edge of the Green Mountain massif (fig. 4, Terrane 4). From the isopleth pattern, the lead sources seem restricted to the lower plate of the proposed northern extension of the Whitcomb Summit thrust fault (Stanley and Ratcliffe, 1985). Because the lead anomalies follow the strike of the basal Late Proterozoic-Cambrian cover rocks, lead may be restricted to specific units of the lithostratigraphic sequence, a situation perhaps analogous to the geologic setting at Laisvall, Sweden, and at other stratabound, sandstone-hosted lead deposits of the world. Based on the studies of Rickard and others (1979) and Bjorlykke and Sangster (1981) on these relatively lead-dominant deposits, further investigation within the zone delineated by the lead isopleths may be warranted.

At the southern extreme of Terrane 4, lead isopleths form a north- to west-trending pattern coincident with, but extending beyond, an area near Jamaica, Vt., that has major bedrock uranium deposits (see Ayuso and Ratté, this

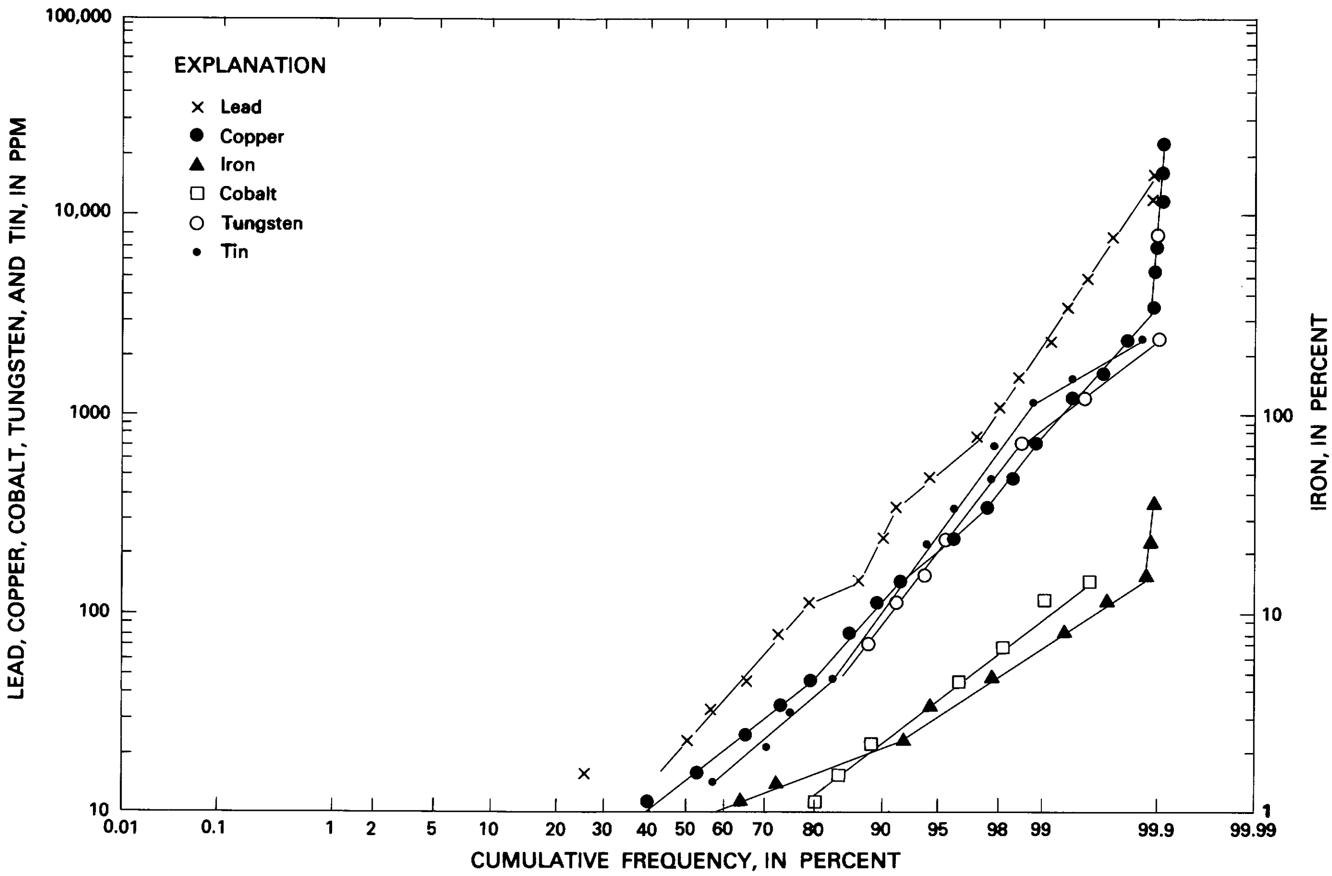

Figure 2. Cumulative frequency plots for lead, copper, iron, cobalt, tungsten, and tin in the Glens Falls quadrangle.

volume). The lead here may be radiogenic in origin and is possibly associated with the uranium deposits. Isopleths of anomalous lead that extend beyond the zone of known deposits may delineate similar, undiscovered uranium deposits in the area.

Terrane 5 yielded an abundance of lead-rich samples. A narrow zone near the base of the Late Proterozoic-Cambrian sequence on the eastern side of the Green Mountain massif is one of the best defined and extensive zones of anomalous lead. Another narrow zone delineated by isopleths in the northwestern corner of this terrane probably defines a different genetic source of lead-rich bedrock. The lead isopleths correlate spatially with abundant greenstone bodies shown on the State geologic map of Doll and others (1961), which implies a volcanogenic origin for the lead. Cobalt and copper, both typical of pyritiferous volcanogenic ores, are also found in this area and reinforce the interpretation of a volcanogenic association for the lead.

Lead anomalies also occur in Terrane 5 near the contact of the Ottauquechee and Pinney Hollow Formations east of the large "suture zone," proposed to be the northern extension of the Whitcomb Summit thrust fault (Stanley and Ratcliffe, 1985). These lead anomalies and those associated with the Chester-Athens domes may have sources in the Barnard Volcanic Member of the Missisquoi Formation, which forms broad outcrop bands within the lead-anomalous area (see Doll and others, 1961).

Lead isopleths delineate zones on the eastern and western sides of the Mount Ascutney pluton, near the Pomfret dome, and along the Strafford-Willoughby arch (fig. 4). For the most part, the lead isopleths in the northeastern part of Terrane 5 are linear and northerly in trend and parallel the axis of the Strafford-Willoughby arch. However, in contrast to the patterns of some key elements, the highest lead contents are from samples collected slightly to the west of the arch axis.

High lead values also form isolated, but somewhat aligned, patterns that parallel the Connecticut River. The isopleth patterns are probably associated with mineralized sources in the Ammonoosuc Volcanics, which also parallel the river along the trend of the Bronson Hill anticlinorium (see Leo, 1985). However, lead deposited as a result of epigenetic mineralization within the Ammonoosuc fault zone is an additional possible lead source.

One discrete lead high southeast of the Pomfret dome, in the Hartland 7½-min quadrangle, Vt., is centered over an area where felsic metavolcanic(?) rock of the Standing Pond Volcanics that has several percent pyrite can be seen in a roadcut. Although the lead anomaly is clearly associated with the Pomfret dome, rock-chip samples of the

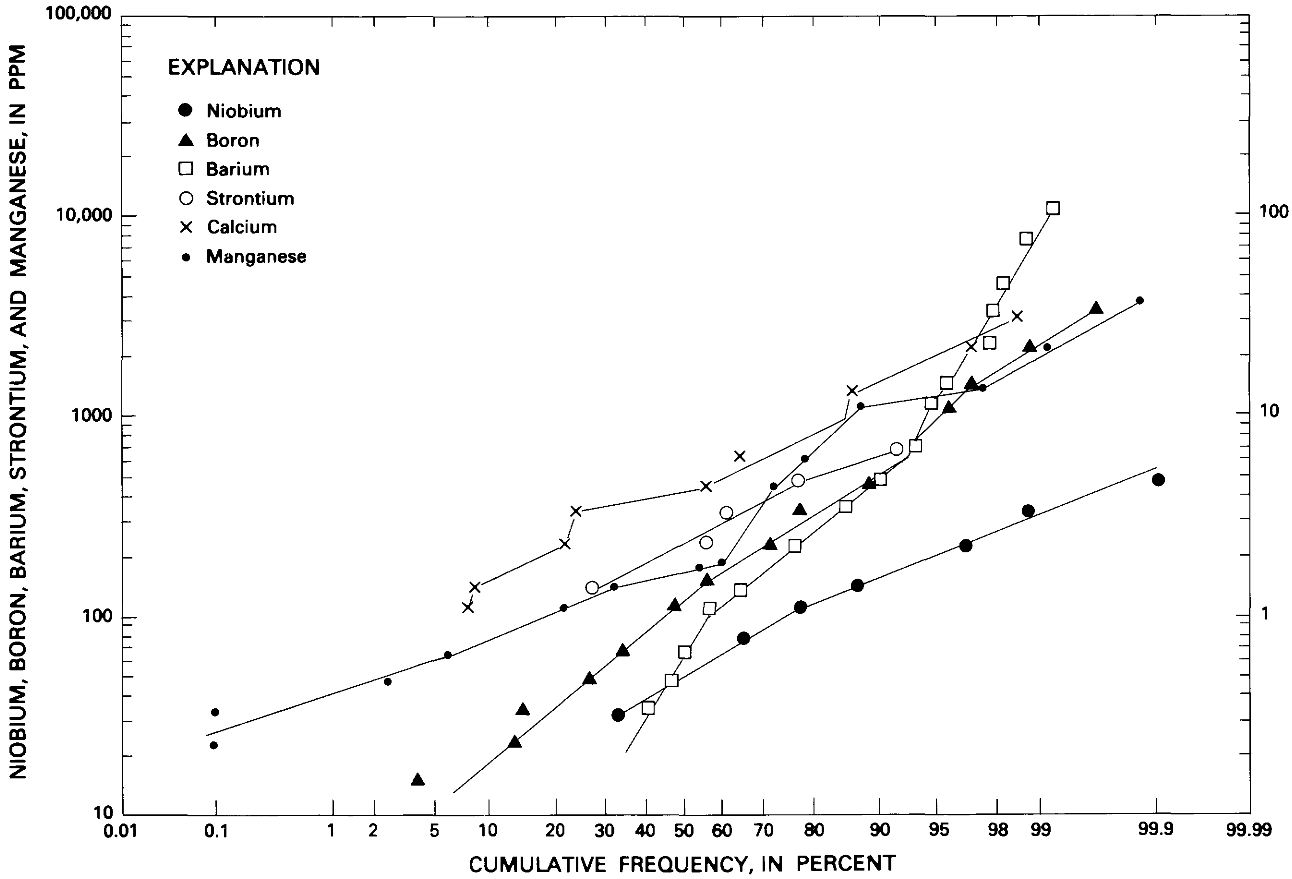

Figure 3. Cumulative frequency plots for niobium, boron, barium, strontium, calcium, and manganese in the Glens Falls quadrangle.

pyritic felsite were not found to be highly enriched in metals. However, possible lateral and vertical variations in the metal content of the rock sequence exposed in the dome were not checked.

In Terrane 6, lead is spatially associated with the Oliverian domes, which are aligned in a northeasterly trend along the Bronson Hill anticlinorium and are the apparent sources of the lead anomalies. These domes include the Lebanon, Mascoma, Unity, and Alstead domes and are regarded as the roots of the island arc volcanic chains that existed off the east coast of North America during the early Paleozoic (Rowley and Kidd, 1981; Leo, 1985; Stanley and Ratcliffe, 1985). The similarity in shape and precise coincidence of the isopleths with the axis of the Alstead dome are particularly striking and suggest that lead in the bedrock source area is related genetically to the dome. The lead anomalies of these domes may have sources in the Ammonoosuc Volcanics, which are exposed on the domal flanks (Billings, 1956; Leo, 1985). However, sources in contemporaneous or later epigenetic veins in the core gneisses cannot be ruled out. One occurrence of abundant galena (with molybdenite, chalcopyrite, arsenopyrite, and sphalerite) in epigenetic veins has been observed on the southern margin of the Lebanon dome along Interstate I–89 (Slack and Schruben, this volume).

Two samples in Terrane 7 that have high lead contents are from areas near the contact of the synkinematic Bethlehem Gneiss and the postkinematic Sunapee two-mica granite (see Thompson, this volume). The contact appears to be the source area of the anomalies. Specific sources for the lead may be pegmatites localized along the contact or veins or bedded deposits within a band of parallel metasedimentary rocks. Tin is the only metal that shows a continuous isopleth pattern here and, most likely, has vein sources; most other elements show an alignment of isolated highs.

Copper

Geometric means for copper from Terrane 3 (34 ppm, table 4) and Terrane 5 (36 ppm, table 6) are slightly above the geometric mean of 28 for the entire Glens Falls quadrangle (see table 1). The range of reported values is highest in Terranes 3 (10–7,000 ppm, table 4) and 5 (10–30,000 ppm, table 6). A cumulative frequency plot for copper (fig. 2) shows several statistical populations with

Table 9. Varimax-rotated R-mode factor loadings based on log-transformed data for nonmagnetic heavy-mineral concentrates

[Correlation of each element with each factor; a perfect correlation is 1.0000; factor loadings ≥0.30 are considered to be significant and are highlighted in bold type]

Element	Factor						
	1	2	3	4	5	6	7
Ag	−0.0454	−0.0298	**0.9146**	0.0845	0.0296	0.0496	0.0671
As	−.1479	−.0684	.0816	**.5825**	.0792	.1221	−.0088
Au	.0032	−.0078	**.9108**	−.0167	−.0199	−.0105	.0076
B	**.6637**	**.3807**	−.0044	.1128	.1032	.1199	−.0655
Ba	.0382	**.3102**	−.0031	.1379	**.3667**	.1788	.1707
Be	.1635	.0711	−.0810	.1144	−.0858	−.2810	**.4701**
Bi	−.1005	−.0365	.0413	−.0085	.1738	.1190	**.5498**
Ca	−.0346	**.3395**	.0043	−.0219	.0091	**.7963**	−.0773
Co	.2544	.1095	.0440	**.6619**	−.0326	**.3627**	−.0149
Cr	**.6409**	−.1232	.0098	.1135	.2246	−.1929	−.1003
Cu	**.5123**	−.0437	.0298	**.3913**	**.3550**	.2810	.1002
Fe	**.4950**	**.3479**	−.0183	**.5763**	.0185	.2667	.1127
La	.0887	**.7876**	−.0840	.0723	−.0360	.0385	.0512
Mg	**.3518**	**.6727**	−.0529	.0226	−.0913	.0791	−.0832
Mn	**.6471**	.2617	−.0223	.1804	−.1429	**.4407**	.1575
Mo	−.0645	.0654	.0289	.0380	.0246	−.0420	**.5401**
Nb	**.8137**	−.2738	−.0375	.1063	.0504	.0753	.1099
Ni	.1828	.0305	−.0179	**.7537**	.0069	−.0775	.0620
Pb	.1115	.1212	.0278	.1361	**.6953**	**.3436**	−.0805
Sb	.0193	.1054	−.0170	.0633	**.5142**	−.2106	−.0624
Sc	−.0942	**.6560**	.0240	.0382	.1367	.0097	−.0365
Sn	.1990	−.0247	.0030	−.0382	**.7047**	.0399	.2474
Sr	.2528	.0203	−.0490	.1503	.1427	**.7710**	.0427
Ti	**.7433**	.0981	.0333	−.1330	.1621	−.0409	−.0254
V	**.8660**	.1307	−.0257	.0340	−.0263	.1588	.1029
W	.2370	−.2288	.0624	−.0913	−.0359	.1329	**.6155**
Y	−.0700	**.6453**	.0318	−.2297	.2640	**.3013**	.0205
Zn	−.0328	.0165	−.0324	**.3314**	.0927	−.1882	−.0142
Percent of variance	20.3	9.6	7.1	6.0	5.5	5.1	4.5
Cumulative percent	20.3	29.9	37.0	43.0	48.5	53.6	58.1

breaks in the curve at 50, 150, 300, and 700 ppm. Above 700 ppm, samples probably contain chalcopyrite or some other ore mineral of copper.

High copper contents in rock-forming and rock-accessory minerals may be an indirect indication of nearby copper-bearing sulfide deposits, even in the absence of cupriferous ore-forming minerals. Although the crystal structures of common rock minerals can accommodate only a moderate amount of copper, intergrowths of ore minerals and rock-forming or rock-accessory minerals can raise copper contents substantially. As an example, Taylor and Slack (1984) report anomalously high copper (to 350 ppm) in magnesian tourmaline (dravite) where it is associated with massive sulfide deposits of volcanogenic or chemical-exhalative origin.

Copper distribution patterns in the Adirondack highlands east of Lake George are nearly identical to those of lead and probably are derived from the same bedrock sources. The copper isopleths (fig. 5) delineate in particular the intersections of Mesozoic faults; this delineation suggests that the permeability at fault intersections may have facilitated hydrothermal transport and deposition from metal-bearing fluids.

The copper anomaly in the southwestern corner of Terrane 1 is most likely associated with a highly pyritic zone. Abnormally acid conditions of the surface and ground waters there (tabulated National Uranium Resource Evaluation data; Koller, 1979; Watts, in press), high concentrations of iron and other metals in heavy-mineral samples, and thick iron-manganese oxide coatings on streambed pebbles (Welsch and Watts, 1986) suggest that there may be large amounts of pyrite in the area. The iron-manganese oxide coatings result from intense remobilization of iron and manganese in the weathering zone; the remobilization can be caused environmentally (for example, acid rain influence on surface water pH or bog conditions) but is probably more closely related to drainage basin lithologies and the amount of pyrite present.

A small zone of copper at the northern extreme of Terrane 1 (fig. 5) may be related to the Proterozoic magnetite deposits that occur to the west. Following current theories on the origin of the magnetite ores (McLelland,

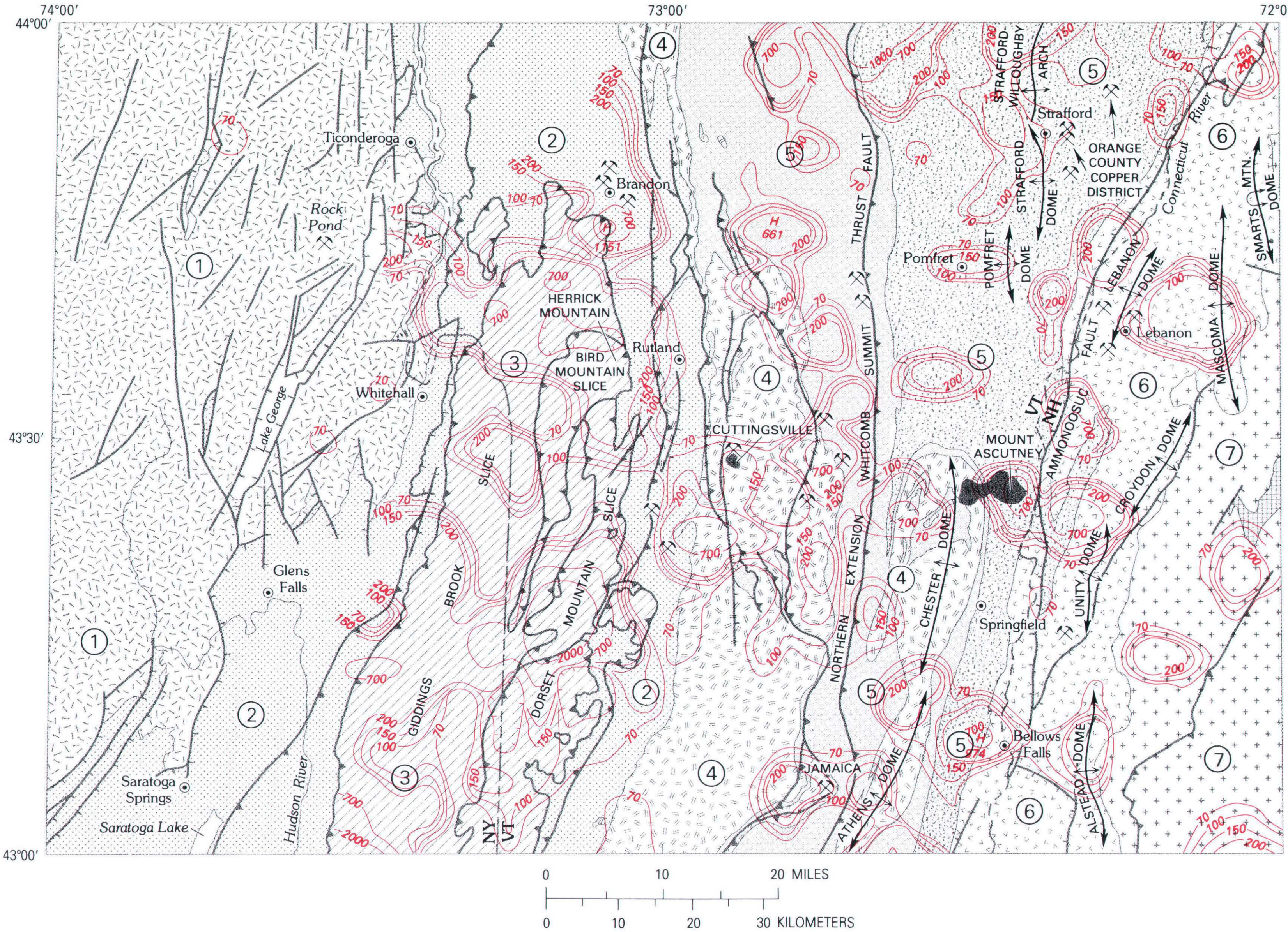

Figure 4. Isopleth map of lead in the Glens Falls quadrangle. Isopleths define areas where samples contain at least 70, 100, 150, 200, 700, or 2,000 parts per million lead. Hachured areas, if shown, contain at least one sample that has a reported value below the isopleth value. See figure 1 for explanation of geologic map units and symbols.

1985; Slack and others, 1987), the copper was introduced into the bedrock as an exhalative product during Middle Proterozoic volcanism. Subsequently, it underwent weathering and was then metamorphosed along with the enclosing country rocks. The nature of the copper occurrence suggests that it is unlikely to be present in exploitable grades.

Copper is not anomalously concentrated in most samples from Terrane 2. A few slightly copper-enriched samples in the northern end of the terrane, near the Brandon area, may be related to stratabound zinc-lead-copper deposits such as those at Lion Hill (Clark, this volume), although these sediment-hosted deposits are not notably rich in copper. Near the southwestern end of Terrane 2, zones of iron-stained shelf rocks present in scattered outcrops also may be sources of minor copper.

In Terrane 3, copper isopleths form a northwesterly trend in the northern half of the Glens Falls quadrangle within the Bird Mountain and Giddings Brook thrust slices (fig. 5). The isopleth patterns encompass a zone of deformed pyritic black and green slates. The isopleth highs center over the general area west of Herrick Mountain. One locality within the center of these copper highs was found to contain an outcrop of iron-stained breccia in black slates mapped by Fowler (1950) as the Schodack Formation (West Castleton Formation). Surface crusts of iron-stained material on fractures from this area contain >1 percent arsenic, 2 parts per billion (ppb) gold, and very little else. Throughout Terrane 3, the copper isopleths locally terminate at the thrust boundary separating the Dorset Mountain-Bird Mountain thrust slices from the Giddings Brook slice, which is here considered to be evidence that copper was concentrated in the source rocks prior to thrust faulting. The generally north-south trend of isopleths parallel to the regional strike of beds and the abutment of these isopleths at thrust faults suggest that the copper sources are stratabound in the rock sequence and are probably syngenetic or diagenetic in origin.

A large zone in Terrane 4 northeast and southwest of Cuttingsville is delineated by copper isopleths that probably reflect sources in northeasterly trending veins. The isopleths form an uninterrupted pattern that connects with the Cuttingsville stock and suggests that similar mineralization both in age and type to that at Cuttingsville is the underlying source of elevated contents of copper throughout the zone. Because the mineralization at Cuttingsville is Mesozoic in age (Robinson, this volume), the entire area of anomalies in the northern part of the Green Mountains may be caused by a geochemical overprint produced during Mesozoic tectonism and mineralization.

The Chester and Athens domes of Terrane 4 are the source of a few copper-enriched samples probably derived from sources in the Barnard Volcanic Member of the Missisquoi Formation that crops out on each side of these domes (see Doll and others, 1961). Some of the anomalous samples appear to come from the juncture of the two domes, and it may be the juncture itself that localized the copper. The Barnard Volcanic Member is a potential host for stratabound massive sulfide deposits (Slack, this volume), and the copper-enriched samples in the area of the Chester and Athens domes could be derived from volcanogenic mineralization within this stratigraphic unit.

In Terrane 5 the isopleth trends for copper (fig. 5) may be the combined result of several structural and lithologic associations, such as (1) greenstone bodies along the northwestern edge of the terrane, (2) metal deposited in the northern extension of the Whitcomb Summit thrust fault, (3) the Barnard Volcanic Member and Moretown Formation, (4) talc-carbonate bodies in the upper plate of the Whitcomb Summit thrust fault, (5) skarns or veins related to granites (for example, Brocklebank pluton) beneath the Strafford-Willoughby arch (see Bean, 1953), (6) stratabound metal remobilized during regional metamorphism into epigenetic veins, (7) stratabound metal concentrations in the Gile Mountain and (or) Waits River Formations and (or) the Standing Pond Volcanics, and (8) epigenetic copper deposited in the Ammonoosuc fault zone and surrounding country rocks.

The close association of north-trending copper isopleths with the axis of the Strafford-Willoughby arch and the approximate symmetry of the isopleths about the axis suggest that the arch in part controlled the configuration of the copper isopleths. The isopleths outline portions of the Orange County copper district on the eastern flank of the arch (White and Eric, 1944; Annis and others, 1983) and cover an area of equal size; similar geochemical contrast exists on the western flank of the arch. Because volcanogenic copper deposits of this district are known on the eastern flank, analogy suggests that similar stratabound deposits may occur on the western flank. The arching and subsequent erosion may have exposed copper-bearing stratigraphic units on the western side of the arch similar to those in the copper district. A cross section drawn by Howard (1969, pl. 3) shows a structural syncline 5 to 10 km west of the main Strafford-Willoughby arch that exposes the same sequence of rocks that are present at the Elizabeth mine (see Howard, 1969, and Annis and others, 1983). On the basis of this geologic similarity and the geochemical anomalies, it is suggested that stratabound massive sulfide deposits like those at the Elizabeth mine may exist on the western limb of the arch, in a relatively unexplored area.

From the area of the Orange County copper district southward, a trend of copper isopleths parallels the Connecticut River. Sources for this copper may be felsic or mafic metavolcanic rocks, which crop out parallel to the river. A copper high occurs to the southeast of the Pomfret dome along this same trend. Copper sulfides are found disseminated and within schistose layers in chloritized and sericitized amphibolite on the eastern flank of the Pomfret dome (Thresher, 1972, p. 4). The mineralized rock there

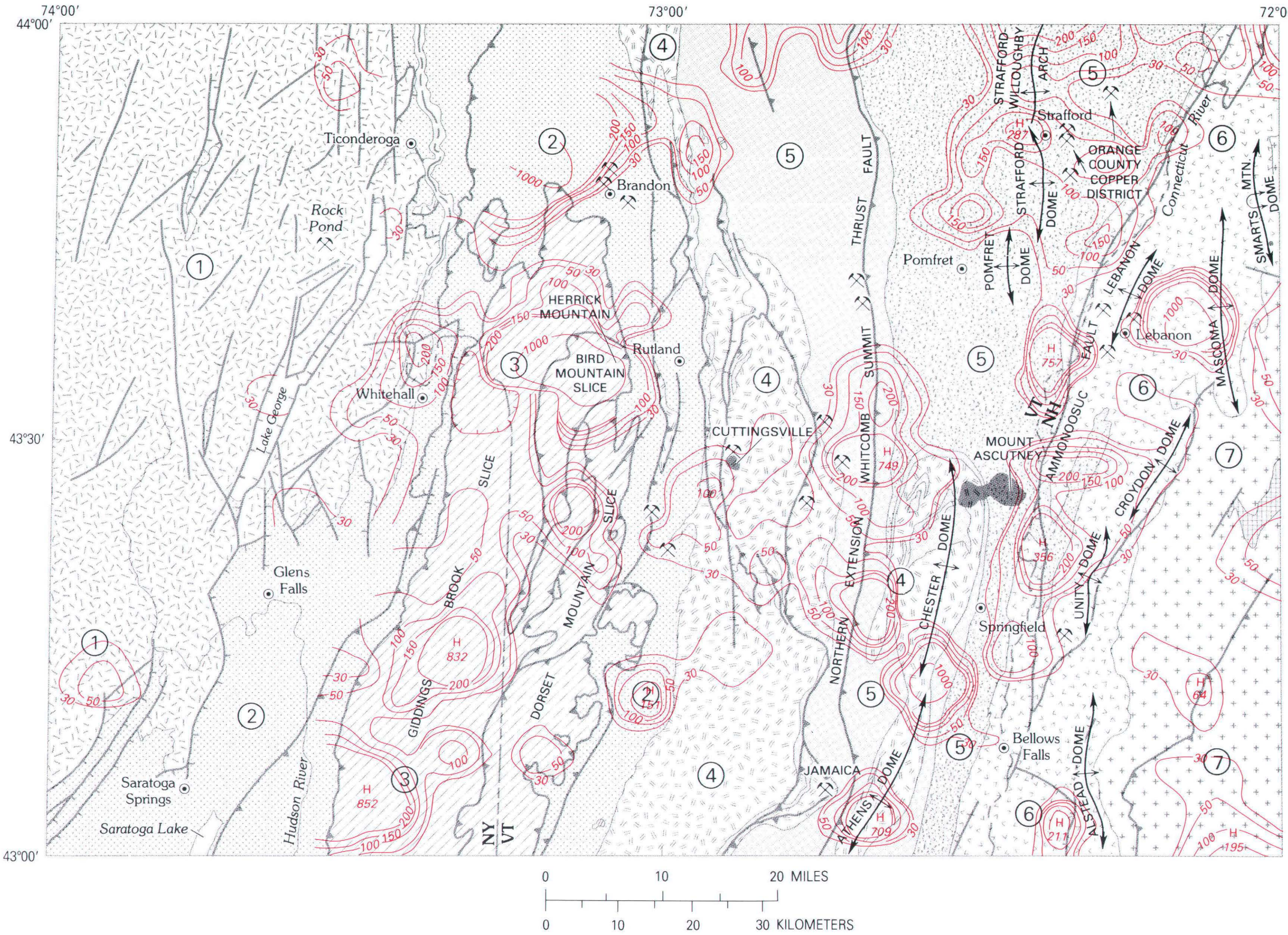

Figure 5. Isopleth map of copper in the Glens Falls quadrangle. Isopleths define areas where samples contain at least 30, 50, 100, 150, 200, or 1,000 parts per million copper. Hachured areas, if shown, contain at least one sample that has a reported value below the isopleth value. See figure 1 for explanation of geologic map units and symbols.

contains 5 to 25 percent sulfides by volume, of which about one-fourth is chalcopyrite and the remainder pyrrhotite.

Copper anomalies in Terrane 6 are confined to a wedge-shaped, northeasterly trending structural block bounded by the Ammonoosuc fault and another Mesozoic fault of northeasterly trend, suggesting that the copper was deposited in the host rocks prior to Mesozoic faulting. An exception to this conclusion may be at the northernmost end of Terrane 6, where the copper isopleths are close enough to the Ammonoosuc fault to suggest it as the source of the copper-rich samples.

An isolated copper high near the Alstead dome, in the southern part of Terrane 6, is the only other area of copper-rich samples. The copper source here may be the Ammonoosuc Volcanics, which are exposed on the flanks of the dome (see Leo, 1985). In places, the Ammonoosuc Volcanics contain small, massive pyritic sulfide bodies, which locally are cupriferous or contain minor copper minerals (see Slack and others, 1987; Moench, this volume; Slack and Schruben, this volume).

An isolated copper high occurs just to the west of the major thrust fault in Terrane 7. The same area has yielded high concentrations of other elements and is discussed in detail in the sections on other element anomalies.

Iron

Nonmagnetic concentrates have a similar high range of iron content in Terranes 3 (0.10–30 percent, table 4) and 5 (0.10–50 percent, table 6). Geometric means in samples from the two terranes are 0.96 percent and 0.92 percent iron, respectively, which are higher than the geometric mean for the entire Glens Falls quadrangle (table 1). The cumulative frequency plot (fig. 2) shows a break in the curve at 2 percent iron. Samples that have 20 to 50 percent iron typically contain a large amount of pyrite, which in such abundance probably is derived from a source of massive pyrite. Iron-rich carbonate minerals and chlorite probably also contribute to some of the iron anomalies (fig. 6). Pyrite, iron-rich carbonate minerals, and chlorite were identified in samples of nonmagnetic concentrates from Terrane 5.

Because pyrite is a principal source of high iron in the samples, the map for iron partially provides a regional view of the distribution of pyritic bedrock. Analyses of pyrite-rich samples indicate that abnormally high amounts of Cu, Co, As, Sn, Mo, Au, or Ag are contained in some pyrite, whereas other types of pyrite are nearly devoid of minor metals. Comparison of the iron map (fig. 6) with the maps of other anomalous metals, for example, near the northern boundary of Terranes 1 and 3, shows zones where pyrite probably is enriched in minor metals overlapping zones where pyrite probably is abundant but is deficient in minor elements.

The two most likely sources of pyrite in the drainage sediments of Terrane 1 are the Proterozoic Dixon Schist, which is characteristically pyritic, and Mesozoic fault zones. Iron anomalies form a north- to northeast-trending isopleth pattern in Terrane 1 that is traceable for several kilometers from the southwestern corner of the terrane northward (fig. 6). The pattern of iron anomalies crosscuts many different rock types and suggests that at least some of the iron occurs in postkinematic crosscutting structures. One area of higher-than-normal iron, east of Lake George, also has samples enriched in lead and copper in a zone that also includes Cambrian Potsdam Sandstone and numerous high-angle faults.

Iron staining, as secondary limonite after pyrite, is widespread in the shelf sequence rocks of Terrane 2, but iron is not enriched in samples from the area. The failure of the large amounts of limonite and pyrite to show geochemically may be partly a result of low relief, which impedes the mechanical dispersion of pyrite and (or) limonite. Many of the sample sites are located on drainages that have sources in the adjacent highland areas of the Green Mountains (Terrane 4) or the Taconic allochthon (Terrane 3) and thus may not totally reflect the amount of iron in Terrane 2.

Much of the iron in Terrane 3 is derived from pyrite in the sooty black slates of the Lower Cambrian Hatch Hill Formation, the West Castleton Formation, the carbonate breccias of the Browns Pond Formation of Rowley and others (1979), and the cherty black slates of the Mount Merino Formation. Some of the pyrite formed by remobilization during folding and metamorphism. Thin laminations and nodular metacrysts of pyrite along joint surfaces and in quartz gash veins are particularly common in the northern part of Terrane 3, in the Giddings Brook and Bird Mountain thrust slices (Slack and others, 1987).

There may be zones on the leading (western) edge of the Taconic allochthon that contain pyrite, apparently enriched locally in gold and arsenic. For example, one heavy-mineral concentrate sample high in pyrite from the leading edge of the allochthon, in the southern part of Terrane 3 (near Schuylerville, N.Y.), contains anomalous gold and arsenic, but neither native gold nor arsenic minerals were seen in microscope scans of the concentrate.

The areas of iron-enriched samples in Terrane 4 (fig. 6) follow a nearly east-west trend and are postulated to be the result of iron introduced during Mesozoic tectonism, because the iron isopleths crosscut the strike of most lithologies and the Proterozoic-Paleozoic structural fabric.

Iron carbonate minerals and pyrite are probably additional sources of iron anomalies in Terrane 5. Iron isopleths follow the trend of the Strafford-Willoughby arch in a pattern similar to those of several other key ore and pathfinder elements (fig. 6). A zone of iron depletion just to the east of the Ely mine, in the Orange County copper district (see Slack and others, this volume), may be the result of iron leaching from the surficial zone. This area

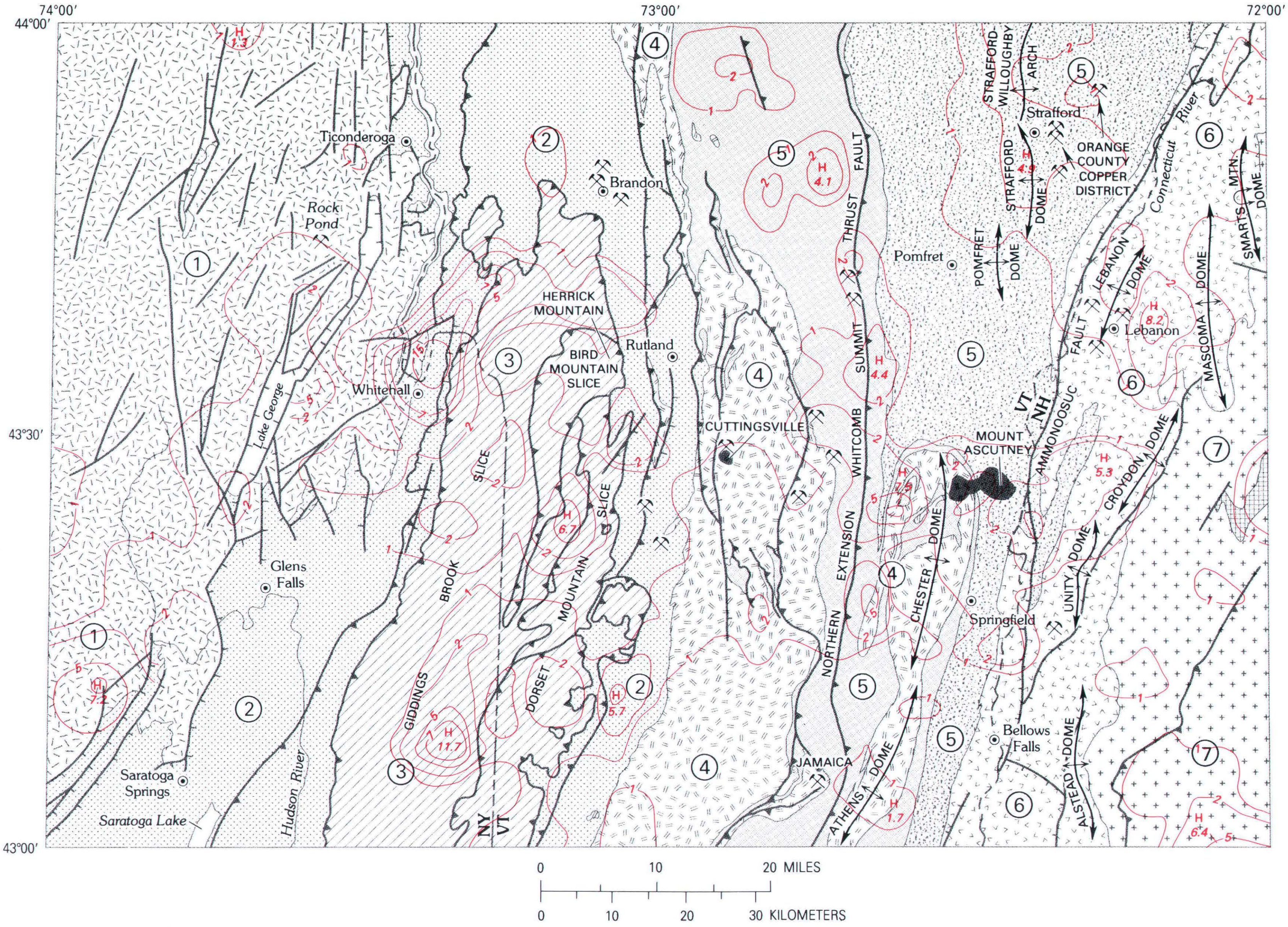

Figure 6. Isopleth map of iron in the Glens Falls quadrangle. Isopleths define areas where samples contain at least 1, 2, 5, or 7 percent iron. Hachured areas, if shown, contain at least one sample that has a reported value below the isopleth value. See figure 1 for explanation of geologic map units and symbols.

should be investigated for possible massive sulfides because the iron depletion there may be the result of highly acid conditions resulting from sulfide oxidation.

Another area of high iron occurs in a zone to the east of the northern extension of the Whitcomb Summit thrust fault (see Stanley and Ratcliffe, 1985). Several other metals show a similar pattern of enrichment. Sources for the iron are probably either pyrite or iron carbonate minerals in veins or in chemical-exhalative deposits within the metasedimentary sequence. Talc-carbonate rock and gold-bearing quartz veins also are known to occur in the area (Slack and Schruben, this volume), and anomalous gold and arsenic were found in concentrate samples from this zone (Day and others, 1986).

Iron-rich samples were collected east of the Lebanon dome in Terrane 6, in contrast to an apparent zone of low iron centered over the axis of the dome. Whether this configuration is the result of hydrothermal depletion in the dome and enrichment on its flanks is only speculative. Also in Terrane 6, iron forms a broad isopleth pattern a few kilometers east of the Mount Ascutney pluton in the same general area where composite rock-chip samples from an outcrop of felsic Ammonoosuc Volcanics containing several percent of stratabound disseminated pyrite and sphalerite have as much as 2.4 percent zinc, 100 ppm arsenic, and 0.13 ppm gold.

In Terrane 7, anomalous iron appears to have sources in the contact zone between the Sunapee granite and the Bethlehem Gneiss (fig. 6). The contact zone may have localized the deposition of sulfides because metals other than iron are enriched in the same area.

Cobalt

Terranes 4 and 5 show the highest range of cobalt values (tables 1–8). The highest geometric means are in Terranes 2 (25 ppm, table 3) and 3 (24 ppm, table 4), which are somewhat higher than the regional mean of 18 ppm cobalt for the total data set (table 1). By far the largest number of valid determinations, reported at or above the detection limit of 10 ppm of cobalt (263 samples), is in samples from Terrane 5 (table 6); these samples account for over one-half of the total number of reported values in the Glens Falls quadrangle (505 samples, table 1). A cumulative frequency plot (fig. 2) indicates that reported cobalt values belong to one statistical population, which probably reflects, for the most part, varying amounts of cobalt in pyrite. Cobalt is a valuable indicator element for some types of sulfide deposits.

Cobalt anomalies occupy the same areas as those for iron in the southwestern corner of Terrane 1 (fig. 7). Such areas may contain large quantities of presently oxidizing sulfides, particularly pyrite, near the surface.

A large area delineated by cobalt isopleths along the eastern edge of Terrane 1 is nearly coextensive with the distribution patterns of anomalous lead, copper, iron, and tin and probably has the same source(s). As with other anomalous elements in this terrane, the pyritic and graphitic Dixon Schist is a possible source. Mesozoic fault zones are another.

Most cobalt in Terrane 3 is derived from samples within the Dorset Mountain thrust slice (high Taconic sequence). The isopleth patterns here are north-south in trend and are parallel to the strike of the metasedimentary beds; the patterns suggest that high cobalt values here may have a syngenetic-diagenetic origin during basin sedimentation.

A few anomalous cobalt values occur in nonmagnetic concentrate samples collected from the vicinity of the Cuttingsville pluton in Terrane 4. The anomalies show a connection with a far more extensive area of anomalies a few kilometers to the south (fig. 7). The continuous distribution of elevated cobalt within areas of different geology suggests that the cobalt may be derived from discordant epigenetic pyrite mineralization (see Robinson, this volume).

A large area delineated by cobalt isopleths in the northwestern corner of Terrane 5 coincides with a belt of greenstone bodies and terminates where the greenstone bodies do not crop out. The cobalt anomalies in these areas may reflect volcanogenic pyrite mineralization related to the greenstones.

The linear pattern of north-south cobalt isopleths in Terrane 5 suggests an overall lithostratigraphic control on the distribution. At the northern end of the cobalt zone, the distribution of isopleths suggests that the lower plate units of the northern extension of the Whitcomb Summit thrust fault are the source of the cobalt, whereas patterns to the south suggest that the upper plate contains the cobalt.

A separate northeasterly trending belt of cobalt isopleths in the southern part of Terrane 5 coincides with the flanks of the Chester and Athens domes (fig. 7). Metavolcanic rock units that occupy these areas are the most likely sources of the cobalt. Both the Standing Pond Volcanics on the eastern side of the domes and metavolcanic rocks of the Barnard Volcanic Member of the Missisquoi Formation on the western side are possible cobalt sources.

In the northern part of Terrane 5, isopleths of anomalous cobalt enclose the outcrop area of the small Brocklebank granite pluton. Also in the same general area, northwesterly trending isopleths delineate the western flank of the Strafford-Willoughby arch in general alignment with the regional direction of latest glaciation (see Larsen, 1987). However, if the cobalt isopleth trend is an artifact of glacial dispersion, then the distance of glacial dispersion is not likely to be more than a few kilometers from the bedrock source.

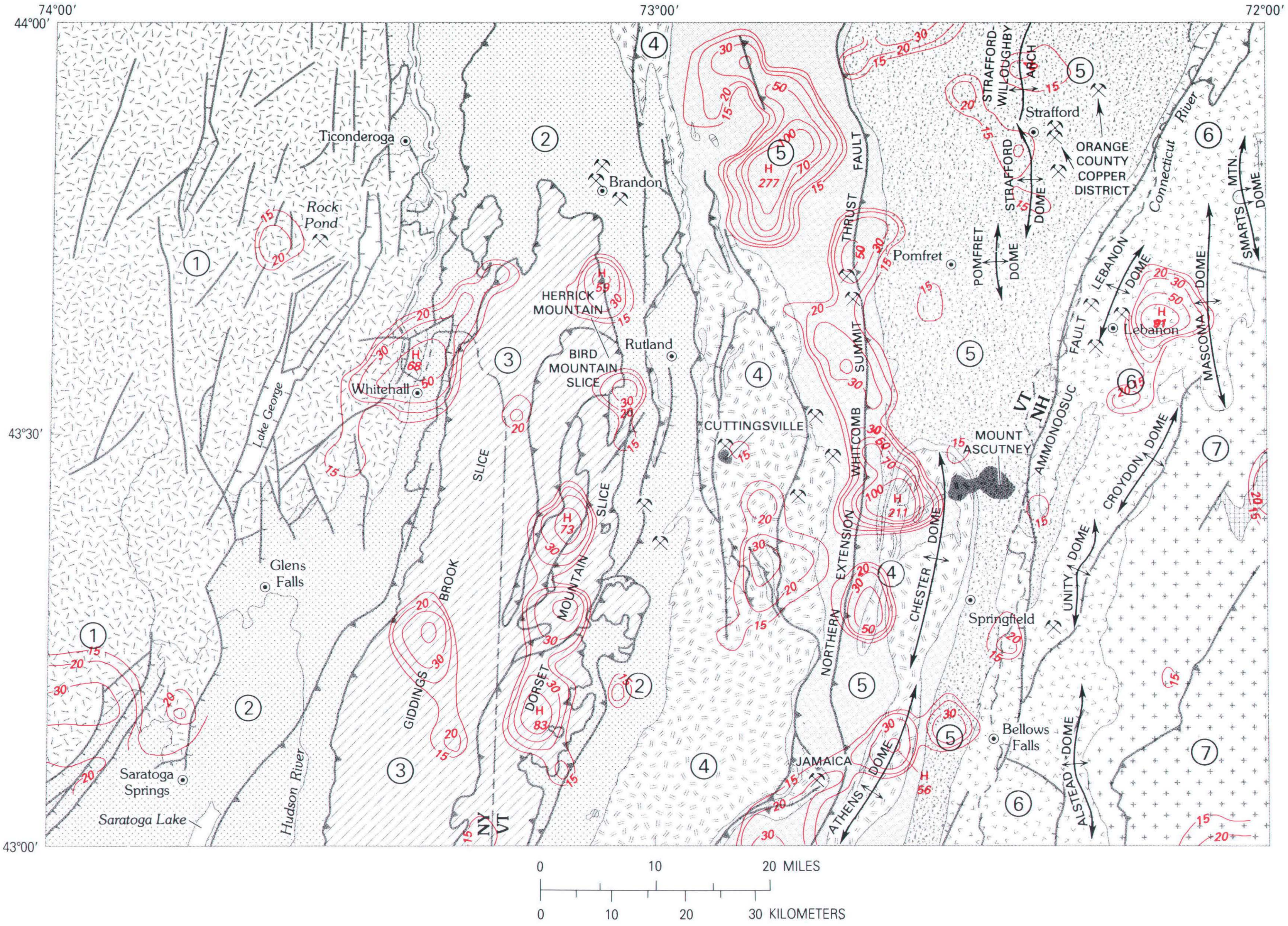

Figure 7. Isopleth map of cobalt in the Glens Falls quadrangle. Isopleths define areas where samples contain at least 15, 20, 30, 50, 70, or 100 parts per million cobalt. Hachured areas, if shown, contain at least one sample that has a reported value below the isopleth value. See figure 1 for explanation of geologic map units and symbols.

A large area delineated by cobalt anomalies south-southeast of the Lebanon dome in Terrane 6 may be derived from mineralized Ammonoosuc Volcanics that crop out on the flanks of the dome. If the cobalt source is stratabound pyrite, the anomaly may be due to doming along the axis of the Bronson Hill anticlinorium; this doming may have exposed pyritic massive sulfide deposits and (or) pyritic volcanosedimentary unit(s) to erosion.

Tungsten

The greatest range (100–20,000 ppm) and the highest geometric mean (297 ppm) for tungsten in the Glens Falls quadrangle are in Terrane 6 (table 7), compared to a geometric mean of 245 ppm tungsten for the total data set (table 1). Because the lowest spectrographic reporting value (100 ppm) is substantially above the normal crustal (background) abundance of 1.5 ppm tungsten (Taylor, 1964, p. 1281), many of the data are truncated at the lower end of normal tungsten abundance (that is, in unmineralized rocks). These shortcomings notwithstanding, the geometric mean within the range of detectable values allows a comparison of the relative extent of tungsten mineralization in the different terranes of the quadrangle. In summary, from the map distributions (fig. 8) and tables 1 to 8, tungsten is most enriched in samples from Terranes 6 and 7, is moderately enriched in a few samples from Terranes 1 and 5, is slightly enriched in parts of Terrane 4, and is present, but below detection limits (<100 ppm) in samples collected from Terranes 2 and 3.

Scheelite was identified in a few samples by scans under the microscope using an ultraviolet lamp and is probably the chief source of tungsten anomalies in the Glens Falls quadrangle. The scheelite fluoresces bright blue, which indicates a low content of the molybdate molecule (Kerr, 1946, p. 73–74), a conclusion supported by the lack of molybdenum reported in spectrographic analyses of samples high in tungsten. There are three notable exceptions: (1) one sample high in tungsten in the Mt. Cube area, New Hampshire, has 500 ppm molybdenum, (2) one sample in the Orange County copper district has 300 ppm molybdenum, and (3) one sample in the Bellows Falls area, along the Vermont-New Hampshire border, contains 500 ppm molybdenum. A cumulative frequency plot (fig. 2) shows two statistical populations of tungsten. The break at 700 ppm is probably the lower limit of a population reflecting samples derived from nearby bedrock containing scheelite.

Lithologic composition is probably the chief factor controlling the distribution of scheelite in the bedrock. Calcium-rich host rocks, in particular, favor the precipitation of scheelite (Kerr, 1946, p. 33–34), and some of the scheelite sources in the Glens Falls quadrangle are here interpreted to be either in mafic metavolcanic or calcic metasedimentary rocks. Other sources of anomalous tungsten are probably epigenetic quartz veins or skarns related to granitoids, in particular those of the New Hampshire Plutonic Suite (Paleozoic) and the White Mountain Plutonic-Volcanic Suite (Mesozoic).

The distribution patterns of tungsten, as well as other metals related to heavy minerals in the Glens Falls quadrangle, may be partly the result of mechanical dispersion southeastward, in the direction of glacial movement (Larsen, 1972, 1987). Analyses of basal till samples in selected areas of the Glens Falls quadrangle (K.C. Watts, Jr., unpub. data) show minor amounts of tungsten in the till, but tungsten is considered by the author unlikely to have traveled far from its bedrock source; scheelite is a brittle mineral and is unstable in the surficial zone. During mechanical transport, it comminutes rapidly with distance, breaking down into very small particles within only a few kilometers, in a process called sliming (Krauskopf, 1970, p. 74–76). Accordingly, scheelite in a heavy-mineral fraction, particularly where present in coarse fragments, is probably near its bedrock source. Thus it is highly unlikely, for example, that in the northern part of Terrane 5, scheelite was glacially transported south several kilometers from skarns or veins associated with granites of the Barre or Knox Mountain plutons to the north of the Glens Falls quadrangle, as proposed by Slack and others (this volume); their model requires this glacial transport plus fluvial transport for as much as 6.5 km, the maximum length of sampled streams, at the same time retaining a visible size range for the scheelite (some grains are 1–3 mm in length). Moreover, the distribution patterns of tungsten anomalies correlate well with the axis of the Strafford-Willoughby arch (fig. 8), a correlation that suggests that a controlling influence on the tungsten isopleths is the arch itself.

The few samples that have detectable tungsten in the Adirondack Mountains (Terrane 1) are from an area of north-northeasterly trending faults (fig. 8). These faults are of probable Mesozoic age (McLelland, this volume), and the fault zones may well be the source(s) of the tungsten. A Proterozoic igneous protolith of specialized, rare metal granites, associated with Proterozoic rifting, is also a possible source of these anomalies. Studies by McLelland (1985, 1986, and this volume) suggest that rift-related igneous-tectonic activity that may have links to the metal suite in the area was part of the geologic events shaping the Adirondacks during the Late Proterozoic. Perhaps of greater significance for the genetic source of the tungsten is the location of the anomalous zone on the projection of an east-west geochemical lineament that probably resulted from crustal extension during the Mesozoic.

Minor tungsten was detected north of Cuttingsville in Terrane 4. The tungsten probably has been dispersed mechanically from skarn deposits or related epigenetic veins associated with a northerly plunging subsurface projection of the Cuttingsville syenite pluton. This interpreta-

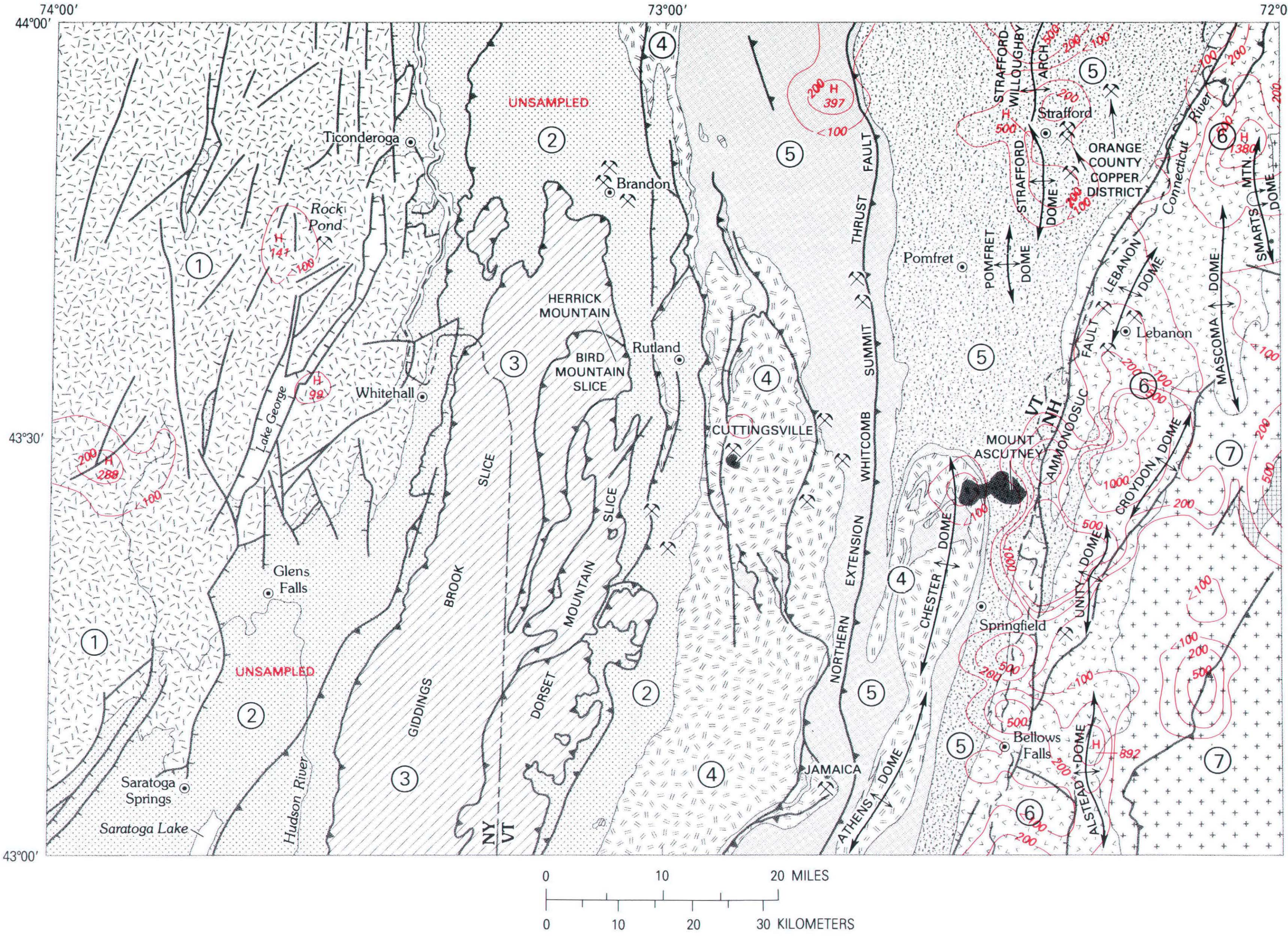

Figure 8. Isopleth map of tungsten in the Glens Falls quadrangle. Isopleths define areas where samples contain at least detectable (at 100 parts per million), 200, 500, or 1,000 parts per million tungsten. Hachured areas, if shown, contain at least one sample that has a reported value below the isopleth value. See figure 1 for explanation of geologic map units and symbols.

tion is supported by field observations of minor wolframite associated with quartz-pyrite veins in a highly altered stockwork zone along the northeastern border of the Cuttingsville stock (Slack and others, 1987; Robinson, this volume). If the anomaly is due to minor scheelite, then the scheelite may be secondary after the observed wolframite, as wolframite is too magnetic a phase to occur in the nonmagnetic heavy-mineral fraction of the analyzed concentrates. Geophysical data (Daniels, this volume) and airborne spectral anomalies of vegetation in this same area (Power and Milton, this volume) also suggest the presence of mineralized rock in the subsurface.

Several areas in Terrane 5 yield nonmagnetic concentrate samples that have detectable (100 ppm) or higher tungsten. The tungsten anomalies can be related to several possible types of mineralization or geologic associations that include (1) greenstone bodies cropping out west of the northern extension of the Whitcomb Summit thrust fault in the northern part of Terrane 5, (2) contact metasomatic skarns or veins associated with granites beneath the axis of the Strafford-Willoughby arch, and (3) skarns or veins surrounding the Mount Ascutney pluton, particularly on the eastern and western margins (see Cox, this volume) (fig. 8).

Stratabound sources of tungsten are possible in Terrane 6. The two most favorable lithostratigraphic units are the Ammonoosuc Volcanics and the calcareous metasedimentary rocks of the Fitch Formation. To some degree the isopleth patterns of anomalous tungsten follow the general outcrop area of these rock units; a correlation of high tungsten values with the Ammonoosuc Volcanics also was found along strike to the northeast by Canney and others (1987). Geologically similar areas where occurrences of stratabound tungsten have been described include the Sangre de Cristo Mountains of Colorado, United States (Moench and Erickson, 1980), Broken Hill, Australia (Barnes, 1983), and Mittersill (Felbertal), Austria (Höll and others, 1987; Thalhammer and others, 1989). On the other hand, the large area of tungsten anomalies east of the Connecticut River at the latitude of Mount Ascutney is most likely related to skarns or veins associated with subsurface intrusive bodies, either of the same age as the Mount Ascutney pluton (Cretaceous) or older (perhaps Devonian). The complex aeromagnetic highs in this region (Daniels, this volume) support this interpretation.

Tungsten is also enriched in samples from around the Oliverian domes of Terrane 6. This enrichment may be the result of erosional exposure of the Ammonoosuc Volcanics on the flanks of the domes; tungsten anomalies near the Smarts Mountain and Alstead domes may be examples of erosionally exposed tungsten bodies in the Ammonosuc Volcanics. The Mascoma dome, another of the Oliverian domes, is known to contain scheelite in quartz veins (Meyers and Stewart, 1956, p. 52). Slack and others (1987) present geologic and geochemical data on the most prominent of the tungsten-bearing veins at Holts Ledge, on the northwestern margin of the Mascoma dome. The overall trend of tungsten isopleths in Terrane 6 is north to northeasterly; this trend is similar to that of posttectonic quartz veins containing lead, zinc, silver, and copper that parallel the host-rock foliation (Alan McBean, Mineral Resource Data System files, Vermont Geological Survey, 1985–86).

Two tungsten-rich samples in Terrane 7 occur near the contact of the Sunapee granite and the Bethlehem Gneiss. The source of these tungsten-rich samples may be pegmatite bodies located near the contact or granite-related hydrothermal mineralization within the contact aureole itself (see Slack, this volume). A small northeast-trending band of metasedimentary rocks composing the Littleton Formation is another possible bedrock tungsten source.

Tin

Reported values of tin have the highest range (20–2,000 ppm) in samples from Terranes 3, 6, and 7 (tables 4, 7, and 8), although the geometric mean is higher in Terrane 7 (182 ppm compared to 128 and 86 ppm). Samples from these terranes are enriched in tin far above the levels in samples from the remainder of the Glens Falls quadrangle (geometric mean = 66 ppm for the total data set, table 1). A cumulative frequency plot (fig. 2) shows three statistical populations of tin and breaks in the curve at about 50 and 1,500 ppm. Reported values in the 500 to greater than 1,000 ppm range probably reflect discrete tin minerals (for example, cassiterite) or tin-bearing sulphosalts (for example, stannite); the exact minerals are yet to be determined although preliminary SEM studies have found some lead-tin compounds, so far unidentified mineralogically. Tin also may be accommodated structurally in titaniferous minerals such as sphene and rutile, in which quadrivalent tin replaces quadrivalent titanium (Rankama and Sahama, 1950, p. 732). Apatite, a widespread accessory mineral in the Glens Falls quadrangle, can in addition contain up to several hundred parts per million tin in geochemically specialized environments (Mulligan, 1973, p. 56). Rock silicates probably account for all of the tin values below the median and perhaps for some slightly above the median. Also, some samples high in pyrite may be slightly to moderately enriched in tin. The tin content of sulfide minerals, in general, appears to be higher in regions that have economic tin deposits (Mulligan, 1973, p. 30) and is especially concentrated in chalcopyrite (Fleischer, 1955).

Tin anomalies along the western edge of Terrane 1 (fig. 9), which also are associated with tungsten, molybdenum, beryllium, and barium, are best explained by two possibilities: (1) sources from a Proterozoic gneiss derived from a tin-specialized igneous protolith (McLelland, *in* Slack and others (1987) discusses metamorphosed K_2O-rich tuffs in the Adirondack highlands that may have tin affinities) and (2) sources from zones of Mesozoic mineraliza-

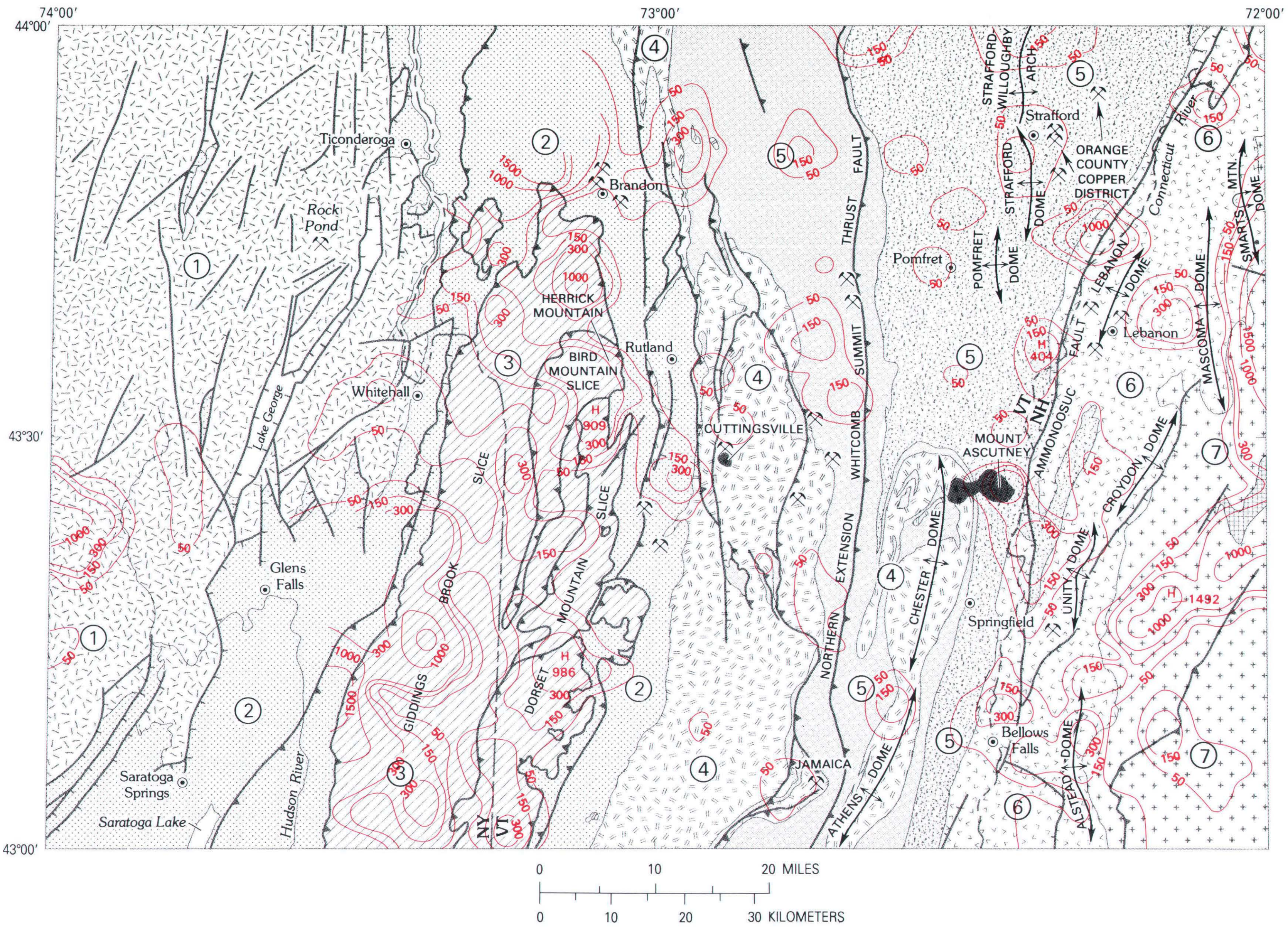

Figure 9. Isopleth map of tin in the Glens Falls quadrangle. Isopleths define areas where samples contain at least 50, 150, 300, 1,000, or 1,500 parts per million tin. Hachured areas, if shown, contain at least one sample that has a reported value below the isopleth value. See figure 1 for explanation of geologic map units and symbols.

tion, in which tin was introduced either epigenetically into Mesozoic fault zones or formed in association with an unexposed Mesozoic igneous body (the anomalous area is on projection with a westerly trending geochemical lineament considered here to be related to Mesozoic tectonics). Other tin anomalies in Terrane 1 are near well-defined Mesozoic faults, suggesting these fault zones as their source.

Two areas in Terrane 2 yielded tin-enriched samples. One area, near Brandon, Vt., in the northern part of this terrane, is also a source of lead anomalies (fig. 4). The source of these metals may be epigenetic veins, as there is no laterally continuous zone of anomalies parallel to the strike of bedding that would suggest a stratabound source. The second tin-rich area in Terrane 2 is farther south near lat 43°30′. Here isopleths are part of a regional anomaly pattern of crosscutting, northwesterly trend, again similar to the isopleth patterns for lead. The tin and lead in this area are probably derived from epigenetic veins.

Samples from Terrane 3 have some of the highest reported tin values in the Glens Falls quadrangle (fig. 9). The geologic setting of this terrane is not typical of areas favorable for the formation of tin deposits, as the Taconic allochthon lacks felsic igneous rocks of any kind, either plutonic or volcanic (for example, see Zen, 1967). Isopleths south of lat 43°30′ outline two separate, but parallel, tin-enriched zones within the Giddings Brook and Dorset Mountain thrust slices. The Giddings Brook slice contains the highest tin values, which are concentrated, much like those for barium, near the leading (western) edge of the Taconic allochthon. The tin isopleths parallel the stratigraphic grain and appear to be cut by the thrust faults, suggesting a Paleozoic and possibly synsedimentary source for the tin. Anomalous tin has been reported from black slates that host stockwork gold deposits in Czechoslovakia (Aichler and others, 1988) and in the Lower Cambrian, potassium-rich shales (fucoid beds) of northwestern Scotland, which also host stratiform barium-zinc-lead deposits of diagenetic-synsedimentary origin (Bowie and others, 1966; Coats and others, 1980; Russell and others, 1984). The close geographic correlation of the high tin anomalies and the barite-related barium anomalies, particularly in the Giddings Brook thrust slice, suggests a mutual bedrock source for the two metals. For example, several nonmagnetic concentrate samples from the Giddings Brook thrust slice that have barium contents of $\geq$5,000 ppm contain tin in the range 700–1,000 ppm.

Tin associations in the Taconic allochthon also may include sedimentary fluorapatite, which is present in some tin-rich samples, and sulphosalt mineral(s), suggested by strong tin-antimony-lead correlations in the geochemical data. Two factors imply that a synsedimentary tin-fluorapatite association may exist in the slate-carbonate sequences of the Taconics: (1) fluorine complexes are considered by many workers to be important transporting agents of tin and (2) tin deposits in metamorphosed sedimentary sequences are known to contain fluorine-rich apatite (Mulligan, 1973, p. 51).

Despite the strong possibilities for a stratabound Paleozoic source of the anomalous tin, the regional tin distribution also suggests that tin may have been introduced epigenetically with the large-scale Mesozoic tectonism. Accordingly, another possible source of tin in this region may be epigenetic vein deposits in Mesozoic fault zones; some high-angle faults that have recently been recognized in parts of the Taconic allochthon (see Thompson, this volume, fig. 1) have northerly or northeasterly trends similar to the trends of the tin isopleths.

Terrane 4 has yielded only a few samples that have elevated tin contents. One small area that contains anomalous tin at the northern end of Terrane 4 is coextensive with a zone also anomalous in lead and copper. The source of these anomalies may be minor galena-bearing quartz veins.

A few tin highs north of Cuttingsville (fig. 9) are probably of Mesozoic origin and have sources either from Mesozoic fault zones or mineralized rocks on the periphery of the Cuttingsville pluton. These may be analogous to the vein-type tin deposits at the Jackson tin mine, northeast of the Glens Falls quadrangle in the White Mountains of New Hampshire (see Moench and others, 1984). Tin is also associated with anomalous copper near the juncture of the Athens and Chester domes, the significance of which is unknown.

Tin isopleths in Terrane 5 follow the axial trend of the Strafford-Willoughby arch. Coincident tungsten anomalies (fig. 8) are also present here, and gold is concentrated locally (see Slack and others, this volume). A granitoid body is inferred beneath the surface of the Strafford-Willoughby arch, and the tin may be related to it.

Tin concentrations near Mount Ascutney probably are derived from rock-accessory titaniferous minerals and not cassiterite. Apatite also may accommodate the tin. The anomalies reflect the affinities of tin to the White Mountain Plutonic-Volcanic Suite of igneous rocks, which are geochemically specialized in rare and possibly precious metals (see Cox, 1987, and this volume, and Robinson, this volume).

Isopleths of high tin also follow the boundaries of Terranes 5 and 6 along the trend of the Ammonoosuc fault zone (fig. 9). These tin anomalies may be derived from epigenetic mineralization along the Ammonoosuc fault.

Some tin isopleths in Terrane 6 are associated spatially with a few Oliverian domes. In some cases, the isopleths follow closely the domal axes and suggest that the tin may have stratabound sources in the Ammonoosuc Volcanics, as proposed for the tungsten anomalies. Stratabound tin deposits in mafic volcanic host rocks have been suggested by some investigators to be of exhalative origin (for example, Plimer, 1980), although without widespread agreement (see Thalhammer and others, 1989).

Some of the highest tin values in the quadrangle are found in a northeasterly trending zone that straddles Terranes 6 and 7. The tin anomalies form a continuous isopleth pattern of northeasterly trend that extends from the Alstead dome on the southwest, easterly to the boundary of the Glens Falls quadrangle. An even larger tin zone to the east in the Portland 1° × 2° quadrangle adjacent to Glens Falls is suggested by the broadening and strengthening of tin concentrations reflected by the isopleths at the boundary between the two quadrangles (see also Canney and others, 1987). The source(s) of tin-rich minerals within the northeast trend may be one of, or a combination of, the following: (1) a belt of calcareous metasedimentary rocks that follows the same trend (see Thompson, this volume); (2) a tin-rich granite phase associated with the New Hampshire Plutonic Suite of intrusive rocks, in particular, the Late Devonian Sunapee two-mica granite; (3) aligned pegmatite bodies; or (4) northeast-trending quartz veins, similar to the quartz-tourmaline-scheelite-molybdenite veins described at Holts Ledge, N.H. (Slack and others, 1987).

Because Devonian granites of the New Hampshire Plutonic Suite are the principal rocks occupying Terrane 7, it is worth noting that similar late- to postkinematic Devonian granites are hosts to several important tin deposits throughout the northern Appalachians (Mulligan, 1973; Ruitenberg and Fyffe, 1982; Richardson, 1988), and at least one notable molybdenum-bearing base-metal deposit at Mount Pleasant, New Brunswick, has possible economic grades of tin (Kooiman and others, 1986). Stannite, a Cu-Fe-Sn sulphosalt, is a significant tin mineral at Mount Pleasant (Mulligan, 1973, p. 26). One other example of a possible tin association with the New Hampshire Plutonic Suite is in Terrane 5, where tin anomaly patterns follow the trend of the Strafford-Willoughby arch, which may be underlain by Devonian granite.

Niobium

Samples from Terrane 4 have the highest range of niobium contents (50–700 ppm, table 5), and those from Terrane 5 have the highest geometric mean (124 ppm, table 6). The overall regional mean for the Glens Falls quadrangle is 97 ppm (table 1).

The levels of niobium in the heavy-mineral concentrates suggest that it is mostly in the lattice of rock-forming or rock-accessory minerals, rather than in discrete minerals of niobium. Niobium strongly follows the geochemical behavior of titanium and zirconium. Typically it is found in ordinary rocks within minerals containing titanium and zirconium (Rankama and Sahama, 1950, p. 605–609). In economic deposits, niobium occurs with tantalum, tin, beryllium, fluorine, and lithium. In the Glens Falls quadrangle, common heavy minerals that can accommodate niobium are rutile, sphene, and zircon. Ilmenite, which typically concentrates niobium, may be enriched in niobium in several areas of the quadrangle, but it was separated from the nonmagnetic fraction during sample processing and was not analyzed. A cumulative frequency plot of niobium (fig. 3) shows two probable statistical populations. Values below 100 ppm represent normal (background) concentrations of niobium in the common rock-forming minerals. Above 100 ppm, a gradation of niobium contents from slightly enriched to highly enriched probably is associated also with rock-accessory minerals like rutile, sphene, and zircon, but high niobium contents in these common minerals (that is, >200 ppm) suggest unusual source rocks. The types of rocks that can contain minerals rich in niobium include specialized granites and pegmatites and, most notably, alkalic igneous rocks that also are a source of abundant zircon and rutile (Parker and Adams, 1973, p. 446–447; Herz, 1976).

Discrete pegmatite minerals of niobium include those of the perovskite group, the pyrochlore group, the niobite-tantalite group, and niobium-rich zircon. Zircon in pegmatites can contain as much as 2 weight percent Nb_2O_5 (Rankama and Sahama, 1950, p. 607–608), and pegmatites are considered to be a source of some niobium anomalies in the Glens Falls quadrangle. Anomalies from pegmatite sources can be expected to form isolated highs and not continuous isopleths covering several square kilometers, such as characterize large areas of the quadrangle. In general, the pegmatite occurrences that most likely contain discrete niobium minerals are in the southeastern corner of the quadrangle where rare metal-bearing pegmatites are known in association with the Bethlehem Gneiss (Morrill, 1960).

Mineralogic scans of the panned concentrates show that rutile is widespread in the Glens Falls quadrangle. It can be assumed as well that it is the chief mineralogic source of niobium in the heavy-mineral samples, but there is not a one-to-one correlation between the amount of niobium and the amount of rutile in the samples. Concentrate samples highly enriched in niobium are more limited areally than is anomalous rutile; this limitation suggests that only rutile from geochemically specialized environments is highly enriched in niobium. Niobium-rich samples (fig. 10) are concentrated in a zone in the northeastern part of the quadrangle and form isopleth highs coextensive with the Strafford-Willoughby arch. These isopleth highs show an outward gradational decrease from the arch axis and form a broad zone parallel to the trace of the Ammonoosuc fault. Metamorphic grade of the host rocks in this area correlates to some extent with the niobium distribution, as the isopleths locally crosscut metamorphic zones; in general, niobium anomalies are most pervasive and intense within the garnet and kyanite-staurolite zones (see Doll and others, 1961). This, however, may be the result of temperature gradients in which niobium was introduced concurrent with granite intrusion beneath the Strafford-Willoughby arch. The localization of niobium in the vicinity of the arch

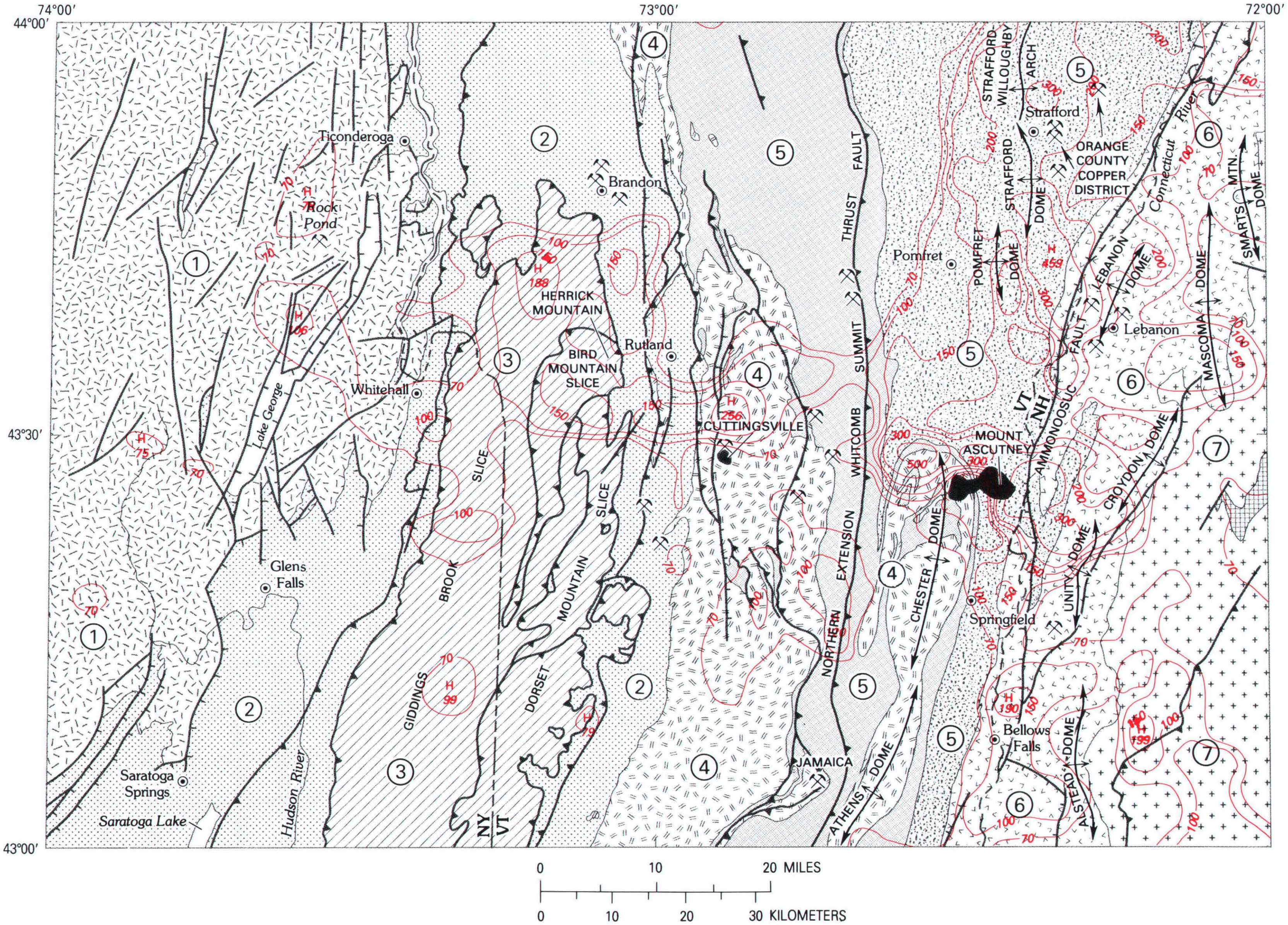

Figure 10. Isopleth map of niobium in the Glens Falls quadrangle. Isopleths define areas where samples contain at least 70, 100, 150, 200, 300, or 500 parts per million niobium. Hachured areas, if shown, contain at least one sample that has a reported value below the isopleth value. See figure 1 for explanation of geologic map units and symbols.

therefore is not considered to be a result of folding and metamorphism of the lower Paleozoic rocks in the area.

Niobium is scarce and not widely distributed in samples from Terrane 1 (fig. 10). Some nonmagnetic concentrate samples that are slightly enriched in niobium were collected near Mesozoic faults, which suggests possible sources for niobium in the fault zones. The area along the eastern edge of Terrane 1 is associated with a westerly trending niobium high located at the intersections of Mesozoic faults. These niobium anomalies are probably related to Mesozoic tectonism.

Alkalic igneous protoliths of Proterozoic gneisses in the Adirondacks also may be a niobium source. High K_2O felsic metatuffs that may be enriched in niobium are described by McLelland, *in* Slack and others (1987). A zone of niobium-rich samples in the west-central portion of Terrane 1 is associated with other elements of the lithophile-rare metal suite and may be related to these K_2O-rich Proterozoic rocks.

Niobium is nearly absent (<50 ppm) in samples from Terrane 2. One small area of high niobium in the northern part of this terrane may reflect dipersion of niobium-rich minerals from a source in the most northerly portion of the Taconic allochthon (Terrane 3) or, more likely, from a geochemical overprint imposed during the Mesozoic.

The northern extreme of the Taconic allochthon yielded several samples enriched in niobium. Anatase was identified in a few of the concentrates from the area and may accommodate niobium by crystal lattice substitution. It has not been established, however, that anatase from this area is niobium rich. The anatase is distinguished by its lack of rounding, euhedral form, and blue color in contrast to rutile, which is typically well worn and red to reddish brown in color. The niobium isopleths cut across the thrust slices in Terrane 3 and may indicate that niobium enrichments are fault and fracture controlled and originated later than the Taconian orogeny. The north-south-trending niobium isopleths along the western edge of the Taconic allochthon also may be related to northerly trending Mesozoic fault zones.

In Terrane 4, an area of high niobium occurs just north of the Cuttingsville pluton (fig. 10), in about the same area as a tungsten high. Like tungsten (fig. 8), the niobium is probably related to the Cuttingsville pluton. The isopleth high is within a sinuous, east-west-trending anomaly at about lat 43°30′. This niobium anomaly encloses both the Cuttingsville and Mount Ascutney plutons and may delineate a zone of structural weakness into which the niobium was introduced. The anomaly cuts across several different lithostratigraphic units and metamorphic facies (Doll and others, 1961), indicating that it is a postorogenic overprint. Niobium values near the Cuttingsville pluton also may reflect a zone of high-niobium zircon, which may indicate anomalous amounts of zircon in the samples, as the concentration of niobium in igneous rocks increases roughly in proportion to increases in zircon content (Rankama and Sahama, 1950, p. 606), and zircon is notably abundant in alkalic igneous rocks.

The broad, northerly trending isopleths of elevated niobium in Terrane 5 delineate a zone that in general follows the Strafford-Willoughby arch. The isopleths are most strongly expressed slightly to the east of the arch axis. The niobium is probably related to rutile, which is abundant in samples from this area. It is possible, although not the preferred interpretation, that the niobium-rich rutile here is a reflection of the regional metamorphism of aluminosilicate rocks enriched in titanium at the time of deposition. Other possibilities are that the niobium was introduced during Late Devonian granitic or hydrothermal activity or during Mesozoic rifting and igneous activity in zones of weakness parallel and adjacent to the Connecticut River.

Isopleths of high niobium coincide with the northern, western, and eastern peripheries of the Mount Ascutney pluton in Terrane 5 (see also Cox, 1987). Niobium lows are delineated by the isopleths on the southern margin of the pluton and within the outcrop area of the pluton. The patterns suggest differential niobium enrichment and depletion as zoned dispersion from hydrothermal alteration and mineralization associated with the emplacement of the pluton. Varying amounts of niobium-rich zircon, derived directly from alkalic igneous bodies, also may be a source of these patterns. One zone of high niobium southeast of Mount Ascutney may be related to a concealed pluton of similar composition to the Mount Ascutney pluton or, alternatively, to glacial dispersion of niobium-rich tills southward from Mount Ascutney (K.C. Watts, Jr., unpub. data). The distance of this glacial dispersal would have to be about 11 to 15 km to account for the observed pattern.

Niobium is very highly concentrated in a multielement anomaly a few kilometers directly east of the Mount Ascutney pluton in Terrane 6. Elsewhere in Terrane 6, the isopleths representing lesser, but still anomalous, amounts of niobium follow closely the trend of the Ammonoosuc fault. The niobium isopleths in Terrane 7 probably reflect the distribution of pegmatites containing niobium-bearing minerals (Morrill, 1960).

Boron

Boron values are highest in Terranes 4 and 5 (20–5,000 ppm, tables 5 and 6). The highest geometric mean is in Terrane 4 (166 ppm, table 5), and the geometric mean for the entire Glens Falls quadrangle is 149 ppm (table 1). Tourmaline, which has a B_2O_3 content typically of 9 to 11.5 weight percent (Rankama and Sahama, 1950, p. 487), is the probable source of most of the anomalous boron in these samples. Tourmaline is common in mineralogic scans and, on the basis of color, is generally of two types: (1) brown, probably magnesium-rich tourmaline, derived from accessory occurrences in certain lithologies, from epigenetic veins, and from inferred chemical metasedimentary rocks, where it may be a percent-range constituent, and (2) black, presumably iron-rich tourmaline derived largely from granites or pegmatites. Tourmaline is a common constituent of the samples and is highly amenable to collection in panned concentrates because of its high specific gravity (3.0–3.2) and resistance to weathering.

Boron distributions occur in four statistical populations based on a cumulative frequency plot (fig. 3) that shows curve breaks at 150, 500, and 1,500 ppm. Variations in the abundance of tourmaline probably account for nearly all of these populations, although muscovite and other aluminous minerals may be minor contributors also. Boron values above 1,500 ppm may suggest percent-range concentrations of tourmaline in bedrock sources, such as in tourmalinite deposits. These concentrations of boron are possible indicators of ancient fumarolic centers, genetically linked to a submarine-exhalative environment (Slack, 1982; Taylor and Slack, 1984). Documented examples of such tourmalinites in the Glens Falls quadrangle occur at and near the Elizabeth mine in the Orange County copper district (Annis and others, 1983) and to the west of North Hartland, Vt., in the Hanover 15-min quadrangle (Lyons, 1955; J.F. Slack and M.P. Annis, unpub. field data).

Mesozoic faults and boron isopleths have similar patterns in Terrane 1 (fig. 11). The large anomalous area at the northern end of this terrane and the area of moderately high boron north-northwest of Lake George are marked by many high-angle faults, and hypogene deposition of boron in these fault zones may be the cause of the isopleth patterns. There is a strong impression of a northerly trending control over the anomalies parallel to Lake George; this also is the most recently imposed (Mesozoic) structural grain in the region (see McLelland, this volume).

One of the most significant zones of boron anomalies occupies the southern one-half of the Glens Falls quadrangle in Terrane 2. Brown tourmaline in abundance and lesser amounts of dark brown or black(?) tourmaline were seen in microscope scans of samples from this zone. The boron isopleths define a north-south trend that appears to continue southward and increase in concentration beyond the boundaries of the Glens Falls quadrangle. Discordant veinlets within geochemically anomalous zones locally contain abundant veinlets of brown tourmaline (and pyrite) in outcrops of Winooski Dolomite (Slack and others, 1987, stop 4.3). It is not established, however, that all of the tourmaline within this broad zone is in epigenetic veinlets; some tourmaline here may be of diagenetic-metamorphic origin. Diagenetic boron in the clay fraction of limestones can contain as much as 2,000 ppm boron in illite (Rankama and Sahama, 1950, p. 493), the amount of which has been proposed to be a function of the salinity of the marine environment. During diagenesis and metamorphism, the boron accumulated in the clay is released and then precipitated as dravite (or uvite) around nuclei of detrital tourmaline (Reynolds, 1965, p. 1111). Therefore, some of the tourmaline observed in Terrane 2 may be derived from metamorphism of the shelf carbonates and siliciclastic rocks of the area. However, two factors argue against a strictly synsedimentary source for the tourmaline and for a largely epigenetic origin of the tourmaline localized by faults or fractures: (1) the abrupt northward termination of tourmaline isopleths and their eastward swing at about lat 43°30′, even though the same sedimentary sequence of rocks continues northward along strike, and (2) the low grade of metamorphism (subgreenschist facies) in the shelf-sequence rocks of this area.

If a large proportion of tourmaline in the southern part of Terrane 2 is of epigenetic origin, then an explanation for the extensive distribution of boron is necessary. One interesting possibility is that the anomalous zone delineates a fossil imprint of upwelling boron-rich hydrothermal solutions that moved along high-angle longitudinal faults, which formed concurrently with the foundering of the continental shelf, early in the Taconian orogeny (see Zen, 1967, p. 64–65, for a description of these events). Another, less likely, possibility is that the tourmaline has migrated (through glacial dispersal or other processes) down valley from the Cambrian Cheshire Quartzite, in which it is locally an important accessory mineral (see Slack and others, 1988).

The northern end of the Taconic allochthon in Terrane 3 has produced samples enriched in boron within the same general area that contains concentrations of ore-forming elements including copper, lead, and barium. This anomaly, and similar ones in the central and southwestern Taconics, are all within the Giddings Brook tectonic slice. The boron anomalies in the Taconic allochthon probably reflect the presence of epigenetic tourmaline and may be related to ore-forming processes.

Much of the boron identified in Terrane 4 is derived from black tourmaline, which locally is a constituent of pegmatites and quartz veins in Grenvillian basement gneisses. There is also some brown tourmaline that may be derived from discordant zones associated with Paleozoic or Mesozoic tectonism. The boron isopleths in Terrane 4 form a wide, east-west-trending zone near lat 43°30′, which may be related to rifting during the Mesozoic. Accordingly, possible explanations for boron anomalies within this east-west zone, which are considered to be unrelated to the occurrence of black pegmatitic tourmaline, include (1) boron (tourmaline) introduced during Mesozoic tectonism or (2) tourmaline in accessory detrital grains within infolded metasedimentary cover rocks. Isopleth continuity with boron zones surrounding the Cuttingsville pluton and the limitation of boron to the east-west zone at lat 43°30′ favor the first possibility. Geologic and geochemical data from

Figure 11. Isopleth map of boron in the Glens Falls quadrangle. Isopleths define areas where samples contain at least 200, 300, 500, 700, 1,000, or 3,000 parts per million boron. Hachured areas, if shown, contain at least one sample that has a reported value below the isopleth value. See figure 1 for explanation of geologic map units and symbols.

the Devils Den area, in the Wallingford 15-min quadrangle (Slack and others, 1985), are consistent with the second possibility.

Boron enrichments near the plutons at Cuttingsville and Mount Ascutney are considered to be related to the plutons. In all probability, the tourmaline was introduced either during Mesozoic igneous activity or is a product of contact metamorphism of boron-rich country rocks. No tourmaline-rich lithologies are known in either area, however.

One zone of high boron about 15 km south of Mount Ascutney has been postulated, as the result of a geochemical survey of glacial tills (K.C. Watts, Jr., unpub. data), to be the position of a glacially concealed pluton, similar to the syenite body at Mount Ascutney. A small aeromagnetic anomaly in this area (D.L. Daniels, oral commun., 1989) is consistent with this interpretation. The postulated igneous body also would occupy the westerly trending geochemical lineament near lat 43°30′ (fig. 11) considered here to be a locus of Mesozoic tectonic and igneous activity and related mineralization.

On the northwestern extreme of Terrane 5, near the northern border of the Glens Falls quadrangle, boron highs coincide with areas underlain by greenstones in the Stowe and Pinney Hollow Formations (see Doll and others, 1961). This spatial association suggests that the greenstones or associated country rocks are sources of the boron; in this case, the boron may be volcanogenic in origin and related to submarine-hydrothermal concentrations of tourmaline (see Slack, 1982).

Boron forms a linear isopleth pattern of northerly trend on the boundary between Terranes 4 and 5, from a distance about 25 km south of lat 43°30′, to a similar distance north of that latitude. Because of the linear pattern and consistent relation to the same general lithostratigraphic sequence, the boron in this area is probably stratabound in the basal Cambrian rocks or in the cover rocks; the source may be an unconformity. The same zone is delineated by anomalous lead in nonmagnetic heavy-mineral concentrates (fig. 4).

Another zone of boron enrichment occupies the upper plate of the northern extension of the Whitcomb Summit thrust fault, just to the east of the area discussed above, particularly at the latitude of the Chester dome. All of the tourmaline from this zone is brown and probably is cogenetic with the metal concentrations (lead, copper, and so forth) found in samples from the same general area.

The axis of the Strafford-Willoughby arch (Terrane 5) is also a locus for boron anomalies (fig. 11). If a granitoid pluton underlies the arch, then the high volatility of boron may have allowed it to migrate upward and be deposited in veins or skarns above the subjacent pluton. If boron was present in the overlying cover rocks as tourmaline in bedded chemical sediments prior to folding and metamorphism, there may have been further concentration of boron as a result of metamorphic remobilization and redistribution into epigenetic veins. Alternatively, the boron may be derived only from stratabound sources and owes its isopleth configuration to the exposure, through erosion, of tourmaline-rich beds within the arch. Glacial dispersion from source areas along the Strafford-Willoughby arch may be the cause of the southeastward shift in isopleth direction at the southern end of the boron anomaly, as the most recent glaciation in the region was directed southeastward (Larsen, 1987).

A few samples from near the Lebanon and Mascoma domes of Terrane 6 have elevated boron contents and, in some cases, also tungsten. The boron in these areas may be from concentrations of tourmaline in the contact aureoles of the domes or from tourmaline in discordant quartz-scheelite-molybdenite veins, such as those at Holts Ledge along the northwestern margin of the Mascoma dome (Meyers and Stewart, 1956; Slack and others, 1987).

A strong boron high occurs between the Ammonoosuc fault and the Unity dome also in Terrane 6. This same area is highly anomalous in several elements and is distinguished by a zone of aeromagnetic highs (D.L. Daniels, oral commun., 1987). The area may have been a center of igneous activity and hydrothermal mineralization, as suggested by the occurrence here of anomalous ore-forming elements. The boron-enriched samples in Terrane 7 probably reflect tourmaline in the pegmatites common to rocks of the area, particularly the upper portions of the Bethlehem Gneiss (see Thompson, this volume; Morrill, 1960).

Barium

Terrane 3 is the area most enriched in barium in the Glens Falls quadrangle and has a geometric mean of 895 ppm (table 4), nearly a factor of four greater than the geometric mean of 233 ppm for the quadrangle as a whole (table 1). A cumulative frequency plot (fig. 3) shows three statistical populations; curve breaks occur at 100 and 700 ppm. Values below 100 ppm are low background contents, attributable to minor barium in the crystal structure of minerals or intergrowths in minerals not inclined to accumulate barium. Values in the 100 to 700 ppm range likewise are largely attributable to the diadochic substitution of barium for potassium in the structures of highly to moderately dense potassic minerals (Rankama and Sahama, 1950, p. 471). For example, muscovite is one possible mineral repository of moderate amounts of barium. Muscovite can range as high as 2.88 in specific gravity and is abundant in many of the heavy-mineral samples; it may be a mineralogic source of much of the slightly to moderately enriched barium, such as that within the east-west zone at about lat 43°30′. It commonly is believed that muscovite is removed by heavy-mineral panning, but the muscovite lost during the panning process is chiefly a function of grain shape and not density.

Barium may enter the crystal structure of apatite, if the apatite is associated with alkalic igneous rocks (Cruft, 1966, p. 391). If barium contents are generally high in the geologic environment, the barium content of apatite will likewise be high. Apatite probably accounts for some of the moderately enriched barium found in samples from drainage basins; this barium has debris input from the alkalic plutons at Mount Ascutney and Cuttingsville. In general, in the Glens Falls quadrangle, moderately but not highly barium-enriched panned concentrate samples probably reflect barium in rock-forming and rock-accessory minerals.

The barium content of metamorphic feldspars in Terrane 1 (fig. 12) associated with gneissic rocks postulated to be metamorphosed K_2O-rich tuffs is as high as 2 weight percent BaO (McLelland, *in* Slack and others, 1987, p. 31). Barium contents in the panned concentrate samples from drainage basins that contain these metatuffs may reflect the occurrence of the barium-rich feldspars. Apatite like that associated with the stratabound magnetite deposits at Mineville, N.Y., just to the north of the Glens Falls quadrangle (McKeown and Klemic, 1956) could also contribute to the barium anomalies.

The highest concentrations of barium in the heavy-mineral samples are a reflection of barite and barium-carbonate minerals. The statistical population of barium values ranging from about 700 ppm to percent-level concentrations is related to the occurrence of these minerals. Most samples that have barium values in this high range were collected in the Taconic allochthon; their presence indicates that Terrane 3 is a barium-enriched province. Samples of heavy-mineral concentrates from the Taconic allochthon exceeding 10,000 ppm barium suggest that percent-range concentrations of barium may be present locally in the bedrock.

Barium isopleths outline a zone along the western edge of Terrane 1, which is coextensive with the distributions of anomalous tungsten and tin (associated beryllium and molybdenum anomalies are not shown by isopleth maps); together these elements constitute a felsic-igneous (lithophile) suite. From these elevated barium contents, and the associated metal suite, the bedrock source of the samples is considered to be igneous and enriched in potassium (the barium is most likely substituting for potassium in feldspar or in a rock-accessory mineral such as apatite).

Barium isopleths in the northern part of Terrane 1 delineate the area of the Marcy massif of metagabbro-anorthosite, which can be expected to have accessory apatite. Some apatite, particularly that associated with the low-titanium magnetite deposits of the region (Newland, 1908), is known to contain significant amounts of rare earth elements, as, for example, in the stratabound magnetite deposits of the Mineville, N.Y., area (McKeown and Klemic, 1956). Accordingly, the barium anomaly patterns may prove useful for locating deposits of rare-earth-bearing apatite in this area (see McLelland, this volume, for geology and rock descriptions).

Except for the part of Terrane 2 southeast of Brandon, Vt., where the barium may be in barite or barium-carbonate minerals associated with Lion Hill-type base-metal sulfide deposits (Clark, this volume), barium isopleths in Terrane 2 seem to represent an offlap from Paleozoic barite derived from Terrane 3 and (or) from postkinematic (that is, Mesozoic) barium overprints on the shelf-facies rocks.

The Taconic allochthon constitutes a previously unrecognized barium-rich terrane in the Glens Falls quadrangle (fig. 12). The areas of highest barium content are in the Giddings Brook thrust slice and closely coincide with zones of high Sn, Pb, Cu, and Zn, and locally anomalous Ag, Sb, and As (some of these metals are not shown on isopleth maps).

From a few kilometers south of lat 43°30′ to the southern boundary of the Glens Falls quadrangle, barium is strongly concentrated in the Giddings Brook thrust slice (low Taconic sequence), whereas barium is not as notably anomalous in the Dorset Mountain thrust slice (high Taconic sequence). Sharp breaks of isopleths at one of the key thrust faults between areas of low and high barium is convincing evidence that barium was deposited in the host rocks prior to thrust faulting and that the presence, or absence, of barium is probably stratigraphically controlled. This does not preclude the coexistence of epigenetic veins of barite and (or) barium-carbonate minerals, some of which may have resulted from remobilization of stratabound barium-rich rocks. Dale (1899, p. 259) reports barite in small concretions within green slates of the Granville area, New York, and in joint fillings elsewhere in Taconic rocks. These minor occurrences, however, are inadequate to account for the high barium values in panned concentrate samples from the Taconics.

The proposed synsedimentary origin(s) of barium concentrations in the Taconic allochthon is supported by analyses of rock-chip samples of black slate and phyllite from the Hatch Hill and West Castleton Formations. Because these samples were collected over a broad geographic area within Terrane 3 during rapid reconnaissance, the exact stratigraphic position of some of the rock samples is not precisely known. A few samples of the black slate outcrops may be from the much younger Mount Merino Formation consisting of chert, black slate, and carbonate breccias located above the second green-black slate contact in the western Taconic sequence (Rowley and others, 1979). The black slate samples classified as pyritic, or iron stained, have a lower geometric mean of barium contents than those classified as unmineralized (that is, lacking obvious staining or sulfides). Fifteen samples of iron-

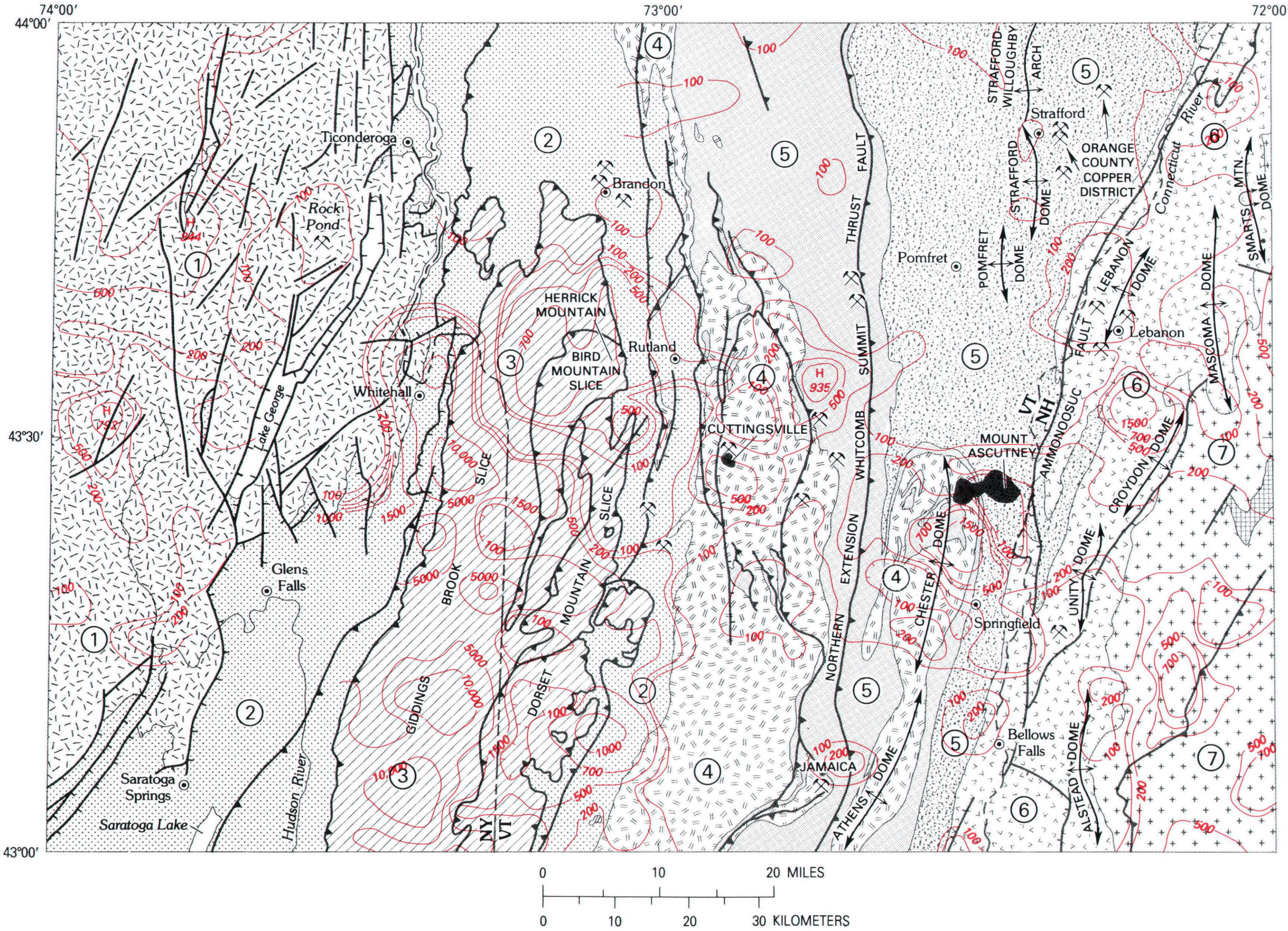

Figure 12. Isopleth map of barium in the Glens Falls quadrangle. Isopleths define areas where samples contain at least 100, 200, 500, 700, 1,500, 5,000, or 10,000 parts per million barium. Hachured areas, if shown, contain at least one sample that has a reported value below the isopleth value. See figure 1 for explanation of geologic map units and symbols.

stained or pyritic black slate show a range in barium contents of 200 to 3,000 ppm and a geometric mean of 474 ppm. Eleven unmineralized samples of black slate have a range of 300 to 2,000 ppm and a geometric mean of 820 ppm. From these reconnaissance data, it appears that sulfide-rich samples are not systematically enriched in barite (or barium-carbonate minerals). When interpreting these data, it should be kept in mind that lateral or vertical zonation of lithofacies rich in barite from facies rich in sulfides can be expected, on the basis of observations in similar geologic environments elsewhere. For example, Carne and Cathro (1981, p. 70, fig. 4) and Large (1980, p. 88, fig. 14) show in their models for sedimentary-exhalative types of mineralization a setting possibly applicable to the Taconics in which lateral and vertical zonation of barite relative to sulfides is a common feature, on a deposit, district, and even regional scale.

In some cases the depletion of an indicator element can be as useful in mineral exploration as the enrichment of that element. Such may be the case for barium near the Cuttingsville and Mount Ascutney plutons. The isopleth map (fig. 12) shows that barium is relatively depleted in zones north of the Cuttingsville pluton, where vein or skarn-related tungsten occurrences are suggested (fig. 8). Barium is depleted also in parts of the area surrounding the Mount Ascutney pluton. In these cases, barium depletion may be the result of hydrothermal leaching of barium from potassic minerals (for example, alkali feldspars) in the host rocks. Nearby barium highs accordingly may be areas where the mobilized barium was redeposited, such as in carbonate veins. Slightly enriched barium near these alkalic plutons also may reflect the common association of barium with apatite in alkalic igneous rocks (Rankama and Sahama, 1950, p. 475).

A trend of barium anomalies of low to moderate contrast in Terrane 5 follows the outcrop of the basal Late Proterozoic-Early Cambrian strata on the eastern side of the Green Mountain massif. This barium trend is similar to those for lead and boron and may indicate the presence of galena-barite and (or) barium-carbonate mineralization in veins or stratabound layers within the basal clastic metasedimentary sequence immediately overlying the Proterozoic Grenville basement. Alternatively, this barium trend may be related to the unconformity itself.

Isopleths of weakly anomalous barium coincide precisely with the axis of the Strafford-Willoughby arch. One possible explanation is that barium was hydrothermally leached from granitic feldspars beneath the arch, redistributed upward into the peripheral cover rocks, and deposited in epigenetic veins. The barium anomalies here are not very high (100 ppm) and are most likely related to barium-carbonate minerals, an interpretation supported by the high calcium values (fig. 13) in the vicinity.

Samples from along the trend of the Ammonoosuc fault are enriched in barium in concentrations similar to those associated with the Strafford-Willoughby arch. The presence of only a few grains of barite in a panned concentrate sample will raise the barium content near the percent range, but this range of barium values is not observed here. Therefore, the barium in the area of the Ammonoosuc fault is probably related to barium-carbonate minerals or to potassic rock-accessory minerals but not to barite.

The barium anomalies in Terrane 6 may have been introduced concurrent with deposition of the Ammonoosuc Volcanics. This proposed mineralizing event may account for the apparent association of barium highs with the Lebanon and Alstead domes where the volcanics are exposed. Isopleths of high barium follow a northeasterly trend within the wedge-shaped structural block bounded by the Ammonoosuc fault and the northeast-trending Mesozoic fault just to the east. This trend is very similar to those for anomalous tungsten and calcium and may reflect scheelite-quartz-carbonate assemblages in veins or beds.

Barium-enriched samples in Terrane 7 also occur, as with other elements, just to the west of the thrust fault shown on the geologic base map. This barium may be associated with barite but is more likely related to apatite or carbonate minerals.

Calcium

The geometric mean for calcium is highest in samples from Terrane 4 (9.5 percent, table 5), almost a factor of 2 higher than that for the total Glens Falls data set (5.7 percent, table 1). The second highest geometric mean is in Terrane 5 (6.4 percent, table 6); terrane 6 has the lowest mean (3.3 percent, table 7). The range of calcium values is lowest (0.1–10 percent, table 3) in the autochthonous carbonate-siliciclastic sequence composing Terrane 2. The low calcium values in this autochthonous sequence demonstrate that the highest calcium values in the concentrates are unrelated to low-density calcite in calcareous metasedimentary units, although dolomite from these units may locally contribute some calcium to the anomalies. The most likely sources of the calcium anomalies are high-density carbonate minerals from veins, calc-silicate skarns, and calcium-bearing igneous and metamorphic minerals such as apatite, wollastonite, and clinozoisite.

The cumulative frequency plot for calcium (fig. 3) shows curve breaks at 1.5, 2.2, about 3.2, 4.5, about 10, and about 15 percent. Values of 10 percent or greater probably represent extreme influx into drainage basins of relatively high-density carbonate vein material, calc-silicate minerals, or apatite.

Calcium reaches its highest levels in Terrane 1 in the northwestern corner of the Glens Falls quadrangle (fig. 13). The 5 percent contour for calcium in this area nearly coincides with the contact zone between meta-anorthosite

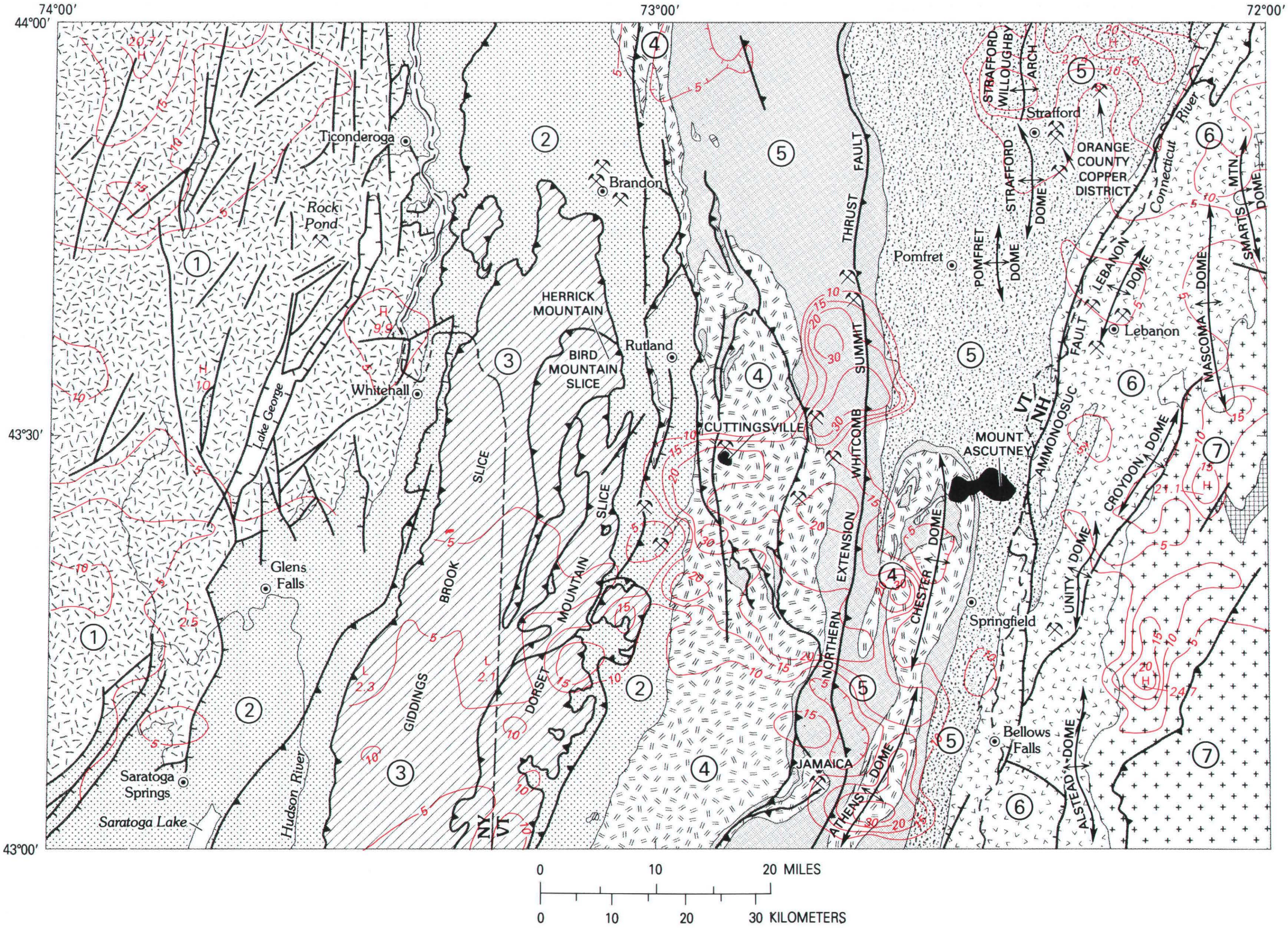

Figure 13. Isopleth map of calcium in the Glens Falls quadrangle. Isopleths define areas where samples contain at least 5, 10, 15, 20, or 30 percent calcium. Hachured areas, if shown, contain at least one sample that has a reported value below the isopleth value. See figure 1 for explanation of geologic map units and symbols.

and metagabbro of the Marcy massif (McLelland, this volume) and the surrounding metamorphic gneisses. This relation suggests that the isopleths of high calcium values probably delineate areas of concentrated wollastonite. Accordingly, the distribution of calcium reflected by this heavy-mineral data may prove useful in locating commercial wollastonite deposits, based on areas within the 10 to 20 percent calcium isopleths.

As discussed above, calcium highs appear to be largely unrelated to calcareous lithologies. Calcium is not enriched in samples from Terrane 2 with one possible exception, an area immediately to the south of lat 43°30′. This anomaly, however, is considered to be part of a crosscutting pattern related to postkinematic mineralization, such as epigenetic carbonate or apatite in veins, or widespread hydrothermal alteration of the country rock resulting in the formation of pervasive epidote-group minerals (for example, clinozoisite), or other calcic alteration products.

Except in the extreme southern part of Terrane 3, the Giddings Brook thrust slice is depleted in calcium. In contrast, the Dorset Mountain slice is comparatively enriched in calcium, forming much the same isopleth pattern as that for strontium (fig. 14).

Calcium is highly concentrated near the Cuttingsville pluton in Terrane 4 (fig. 13). The calcium anomalies here probably reflect the distribution of epigenetic carbonate minerals in veins and stockworks as observed by Robinson (this volume) near Cuttingsville. The calcium highs continue several kilometers southward from the pluton, in an uninterrupted pattern, and may delineate the further extent of a geochemical overprint concurrent with Mesozoic tectonism. Alternatively, this southward trend may reflect glacial dispersion of hydrothermal carbonate minerals from the area of the Cuttingsville pluton.

Some of the calcium anomalies in Terrane 4 also may have sources in Proterozoic (Grenvillian) calc-silicate rocks and in accessory igneous-metamorphic apatite. Comparison of the calcium highs with the Proterozoic geology of the anomalous drainage basins (Doll and others, 1961) may be used to discriminate among these sources.

An additional calcium high in Terrane 4 is centered over the axis of the Athens dome. The same area is also anomalous in lead and therefore may indicate a carbonate-base-metal association in the Proterozoic basement rocks of the Athens dome.

Isopleths of high calcium in Terrane 5 are largely restricted to the lithostratigraphic sequence east of the northern extension of the Whitcomb Summit thrust fault, except in the northern part of the area. The highest calcium values ($\geq$30 percent) occur in a zone west of the thrust fault and north of lat 43°30′. These isopleths probably reflect the areal distribution of carbonatized rock, carbonate veins, and talc bodies, which appear to be closely associated with some gold-arsenic anomalies (see R-Mode Factor Analysis section).

One other area of Terrane 5 also produced panned concentrate samples that have noteworthy amounts of calcium. This area is centered on the axis of the Strafford-Willoughby arch (fig. 13), like the anomalous areas observed for many of the other key elements. Significantly, tungsten anomalies (fig. 8) coincide areally with calcium anomalies, suggesting that calcium highs may be related to the occurrence of scheelite-bearing mineralized zones. This conclusion is based on the close spatial correlation of the highest calcium values with anomalous tungsten. Calcium enrichments in this area also may reflect the distribution of dolomite in the Waits River Formation; the dolomite has a somewhat high specific gravity (2.85) and thus may constitute a significant proportion of the panned concentrate samples.

In Terrane 6, the distribution patterns of high calcium (fig. 13) are similar to those of tungsten (fig. 8). Calcium highs are centered over the Mascoma dome, where scheelite-bearing quartz veins are known on the northwestern side at Holts Ledge (Slack and others, 1987). Calcium anomalies also are associated with the Lebanon and the Smarts Mountain domes, occupying much of the same areas as those delineated by tungsten anomalies. The source of at least some of the anomalous calcium therefore is probably in scheelite. Calc-silicate minerals and (or) carbonate gangue perhaps are associated with this scheelite.

Isopleths of very high calcium outline portions of Terrane 7. One zone of high calcium occurs immediately to the west of the thrust fault in the southeastern corner of the quadrangle. The calcium isopleths form a wedge-shaped pattern of highs just to the west of the Sunapee two-mica granite. The mineralogic source of the calcium may include scheelite, particularly in the area of the closed high in the southeastern corner of the map area, but probably also includes calc-silicate minerals, vein carbonate minerals, or pegmatite minerals.

Strontium

Panned concentrate samples from Terranes 4 and 5 have the same range of strontium (200–2,000 ppm, tables 5 and 6); this range shows the most variation in strontium content in the Glens Falls quadrangle. Terranes 5 (geometric mean=475 ppm, table 6) and 6 (geometric mean=488 ppm, table 7) have similar average contents of strontium, which are slightly above the geometric mean for the quadrangle as a whole (408 ppm, table 1). A cumulative frequency plot (fig. 3) shows a subtle break in the curve at about 500 ppm strontium. Samples that have strontium contents above 700 ppm, particularly those in the 1,000 to >2,000 ppm range, probably are derived from unusual source rocks that have possible mineral resources.

Similarities in chemical properties and ionic size allow Sr^{2+} to isomorphically substitute for Ca^{2+} in the crystal structures of rock-forming and rock-accessory phases, particularly apatite, carbonate minerals, and calc-silicate minerals. Apatite is undoubtedly one of the chief hosts of high strontium in both fresh and altered igneous rocks and in metasedimentary rocks of the Glens Falls quadrangle. Strontium-bearing carbonate or sulfate minerals also may account for high strontium values in both hydrothermally altered rocks and metasedimentary rocks. Celestite ($SrSO_4$) was identified in one sample from the Taconic allochthon and may be widespread. However, carbonate minerals that have high strontium contents are probably more common in the Taconics (and elsewhere) than is celestite. In sedimentary rocks, a direct correlation exists between contents of Sr and P_2O_5 in primary sedimentary phosphorite (Bliskovskiy and others, 1967, p. 1186); this correlation may have significance for the interpretation of strontium anomalies in the Taconic allochthon relative to the bedded phosphorite units that occur in parts of the allochthon (Rowley and others, 1979). With regard to igneous rocks, strontium is moderately enriched in apatite from felsic igneous rocks; is somewhat higher in concentration in apatite accessory to alkalic rocks, ultrabasic rocks, and carbonatites (Cruft, 1966, p. 397); and is highest in hydrothermal apatite (Rankama and Sahama, 1950, p. 473). Cruft (1966, p. 391) stresses that a strontium geochemical province will be reflected in strontium-rich apatite, regardless of the origin of the apatite.

Isopleths of moderately enriched strontium, probably related to rock-accessory apatite, outline the outcrop area of the calcium-rich meta-anorthosite and metagabbro of the Marcy massif in the northwestern corner of Terrane 1 (fig. 14). The strontium distribution here may be used to locate apatite-rich bodies, although the use of calcium anomalies may be a more direct method. The scattered, minor amounts of strontium in samples collected near Mesozoic faults elsewhere in this terrane may, on the other hand, be related to carbonate minerals in the fault zones.

In the carbonate-quartzite shelf sequence of Terrane 2, strontium anomalies are probably a reflection of carbonate or sulfate minerals in veins. Slight enrichments of strontium in samples from this area may be the result of diagenetic release of strontium to pore waters in the sediments (Veizer and others, 1971, p. 16), followed by precipitation in epigenetic veins. Primary strontium concentration is a controlling factor on the distribution of strontium in sediments, so that areas where strontium is slightly enriched in metasedimentary rocks may indicate locations where strontium was concentrated during primary sedimentation and diagenesis.

A well-defined, northerly trending pattern of strontium isopleths delineates the Dorset Mountain thrust slice in Terrane 3 (fig. 14). The strontium here apparently has lithostratigraphic sources and is confined within the limits of the Dorset Mountain (high Taconic) slice; it is relatively depleted in the Giddings Brook slice, where barium is highly enriched. The strontium-rich samples in the high Taconic sequence probably have paleoenvironmental implications with respect to controlling parameters such as pH and salinity; strontium content, for example, is particularly high in hypersaline carbonate sequences (Veizer and Demovič, 1974, p. 110). One other possibility is that the strontium-rich samples are from zones in the Dorset Mountain thrust slice that contain abundant carbonate or quartz-carbonate veins formed perhaps during deformation, as the stratigraphic sequence in the high Taconics contains few sedimentary carbonate units compared to the Giddings Brook slice (for example, Zen, 1967). The thrust faults also are known to have brought tectonic slivers of carbonate units upward (Rowley and Kidd, 1981; Stanley and Ratcliffe, 1985), but the source of the strontium anomalies cannot be ordinary carbonate rocks, as strontium (and calcium) anomalies are absent from the area of the shelf rocks (figs. 13 and 14). Some stratigraphic units in the high Taconic sequence may contain sedimentary fluorapatite, which could be another source of high strontium. Fine-grained fluorapatite was identified in a few panned concentrate samples from different parts of the Taconic allochthon, but its overall distribution is unknown. Pebbles of black phosphate in a calcarenite matrix have been described in the lower Hatch Hill Formation, the Browns Pond Formation, and the Mettawee Slate by Rowley and others (1979). Samples from the Mount Merino Formation also were found to contain high phosphorus attributable to fluorapatite nodules such as those seen in outcrop at the type locality for the Mount Merino Formation, south of the Glens Falls quadrangle. All of these units are in the Giddings Brook thrust slice, however, and not the Dorset Mountain slice.

Celestite ($SrSO_4$) was found in one sample from Terrane 3, but the extent of its distribution has not yet been determined. If widespread, celestite, like barite, may be a key indicator mineral reflecting possible hypersaline, synsedimentary brines or hydrothermal fluids that entered the Cambrian-Ordovician depositional basin and perhaps deposited ore metals such as lead and zinc.

Isopleths of strontium anomalies form the same east-west trend in Terrane 4 as those of other alkaline earth and rare metals and may be similarly related to a crosscutting geochemical overprint of probable Mesozoic age. North of about lat 43°30′, the strontium isopleths broadly follow the core of the Green Mountain massif. At about lat 43°30′, the strontium isopleths are crosscut by a broad (15 × 18 km) zone of east-west-trending, weakly to moderately anomalous strontium values. The isopleths within this zone form peripheral patterns of increased strontium concentration in samples from the areas surrounding the Cuttingsville and Ascutney plutons (fig. 14). Some of the high strontium may reflect normal values for alkalic igneous rocks, based on the concentrations of >1,000 ppm strontium identified in samples of unaltered intrusive rocks from both plutons (Laurent and Pierson, 1973; Foland and others, 1985;

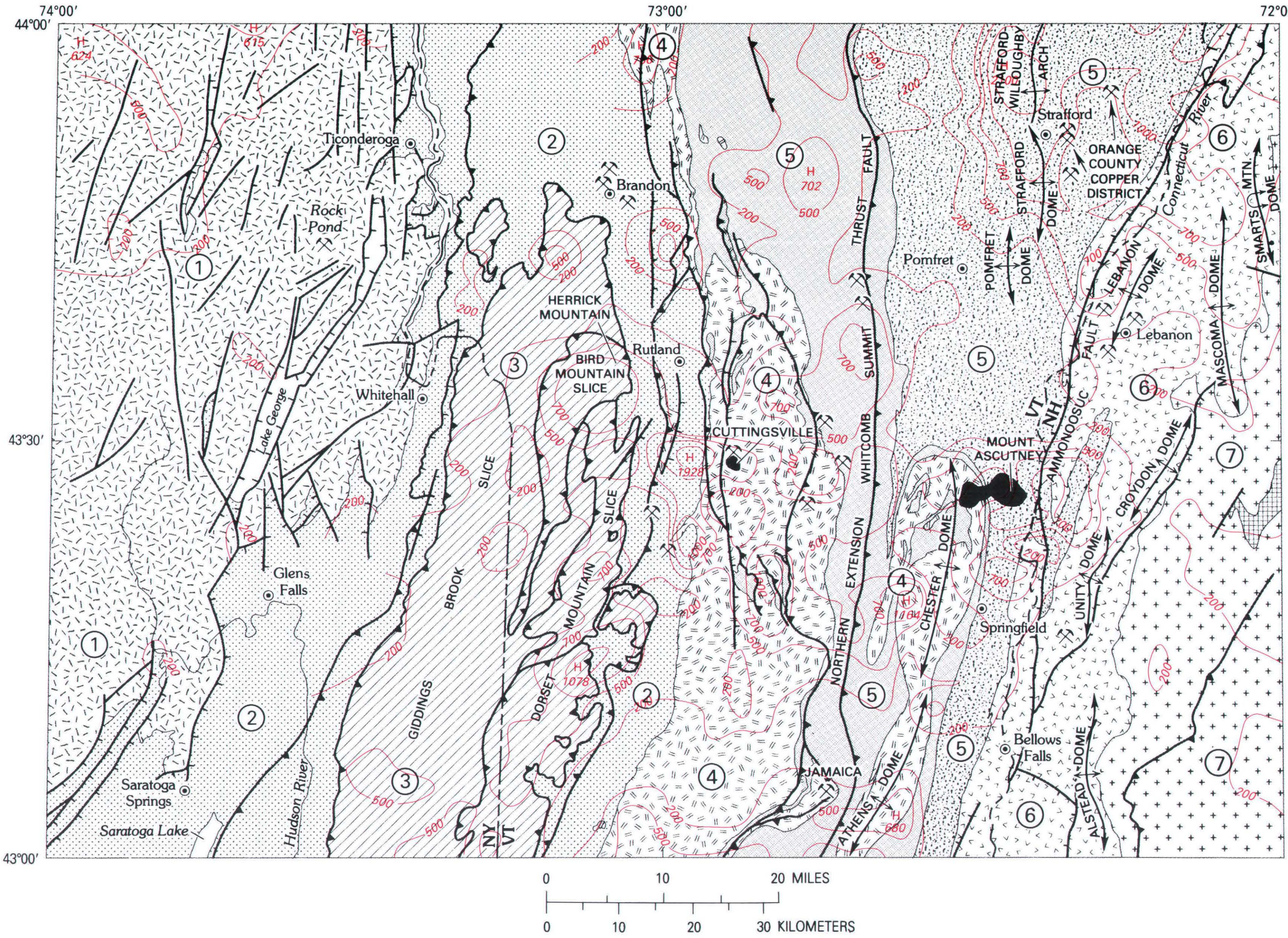

Figure 14. Isopleth map of strontium in the Glens Falls quadrangle. Isopleths define areas where samples contain at least 200, 500, 700, 1,000, or 1,500 parts per million strontium. Hachured areas, if shown, contain at least one sample that has a reported value below the isopleth value. See figure 1 for explanation of geologic map units and symbols.

Robinson, this volume), in which case the broad strontium isopleth pattern may demarcate the extent of emplacement of alkalic intrusive bodies, including lamprophyre dikes, during the Mesozoic. Hydrothermal or contact metasomatic strontium in either hydrothermal apatite, or in carbonate minerals within epigenetic veins, may be additional sources of the anomalies. Hydrothermal mobilization of strontium during igneous activity at depth, and redeposition in surface fractures or in altered country rocks as a hypogene halo, may account for the concentric shape of the anomaly patterns that surround the two plutons. A strontium source in the unmineralized metamorphic country rocks of these areas is unlikely, on the basis of the low strontium contents (typically <500 ppm) reported by Foland and others (1985) for Precambrian gneiss and Paleozoic schist surrounding the Mount Ascutney pluton.

Northerly trending isopleths of high strontium delineate some zones of Terrane 5. The lower plate of the northern extension of the Whitcomb Summit thrust fault east of the Green Mountain massif is delineated by strontium highs that extend in a north-south trend for the full distance of the quadrangle. The strontium here is probably a constituent of carbonate minerals, derived from zones of carbonatized (steatized) ultramafic rock or carbonate-bearing veins in or near the ultramafic belt that generally follows the Ottauquechee Formation (see Doll and others, 1961). The areas delineated by the strontium isopleths are possible environments for the occurrence of gold and base-metal deposits. The areas that have high strontium also contain anomalous iron and calcium (figs. 6 and 13); these data may support the interpretation that relatively dense carbonate minerals derived from areas underlain by carbonatized rock are the source of the anomalies.

Strontium anomalies delineate the axis of the Strafford-Willoughby arch in Terrane 5 and form a gradient of decreasing strontium concentrations outward from the arch axis toward its flanks. The distribution and trend of the strontium isopleths suggest genetic ties between strontium enrichment and the formation of the arch. The similarity in areal distribution of strontium anomalies with those of calcium (fig. 13) suggests hydrothermal carbonate minerals and (or) hydrothermal vein apatite as the source of the anomalies. An attractive hypothesis is that hydrothermal alteration at depth beneath the arch, as described for areas peripheral to the Cuttingsville and Mount Ascutney plutons, mobilized strontium and redeposited it in epigenetic veins within the metasedimentary cover rocks. Accordingly, the concentration gradient of strontium may have been caused by heat centered immediately beneath the arch, and the strontium may delineate a hypogene halo, perhaps related to granite intrusion and (or) hydrothermal mineralization beneath. Recent geochemical studies by J.F. Slack (oral commun., 1989) show a strong correlation between strontium and calcium contents of unaltered and unmineralized carbonate-rich rock samples of the Waits River Formation from eastern Vermont. This correlation suggests a local lithologic control on the distribution of strontium, independent of any mineralization.

There is some indication that strontium was introduced into the Ammonoosuc fault or adjacent country rocks, perhaps by hydrothermal activity. This interpretation is based on the elongate, northerly trending strontium isopleths that border the Connecticut River, between Terranes 5 and 6 (fig. 14), which suggest a geochemical association with the fault. It is equally possible, however, that the strontium concentrations here reflect carbonate-bearing veins in Paleozoic metavolcanic rocks bordering the river.

Anomalous strontium is also associated with the Mascoma dome in Terrane 6. Postkinematic quartz-carbonate veins and (or) hydrothermally altered Ammonoosuc Volcanics are likely sources of the anomalous strontium in this area.

Calcic minerals in the gneissic rocks or apatite in pegmatites chiefly within the Bethlehem Gneiss are the most probable source(s) of high strontium in Terrane 7 (fig. 14). Little mineral-resource significance can be attached to these anomalies.

Manganese

The highest geometric mean for manganese is in Terrane 7 (486 ppm, table 8), closely followed by Terrane 5 (404 ppm, table 6). The lowest mean is in Terrane 2 (112 ppm, table 3), where manganese anomalies are absent, at least in nonmagnetic heavy-mineral concentrates. The geometric mean for the Glens Falls quadrangle as a whole is 304 ppm (table 1). Breaks in the cumulative frequency plot (fig. 3) occur near 150, 200, about 450, 1,000, and 1,300 ppm, and so several statistical populations are present. Manganese contents range from 20 to 7,000 ppm over the Glens Falls quadrangle, and values of up to 7,000 ppm are present in Terranes 3, 4, and 5 (tables 4–6).

Manganese in nonmagnetic concentrates can have a variety of mineralogic hosts, the most important of which in the Glens Falls quadrangle is probably carbonate minerals. The various oxyhydroxides of manganese, which are useful in geochemical exploration, are generally sparse in the nonmagnetic fraction of the heavy-mineral concentrates. Hydrothermal apatite and apatite from pegmatites also can be sources of anomalous manganese, as can epidote- and garnet-group minerals. Manganese contents in hydrothermal or pegmatitic apatite may be as high as 9 percent (MnO), and accessory apatite from syenite can contain up to 0.2 percent (Rankama and Sahama, 1950, p. 646).

In Terrane 1, the highest concentration of manganese is in the southwestern corner (fig. 15). This area has acid ground waters that may be caused by an oxidizing massive pyrite (or pyrrhotite) body (Watts, in press). Cobbles

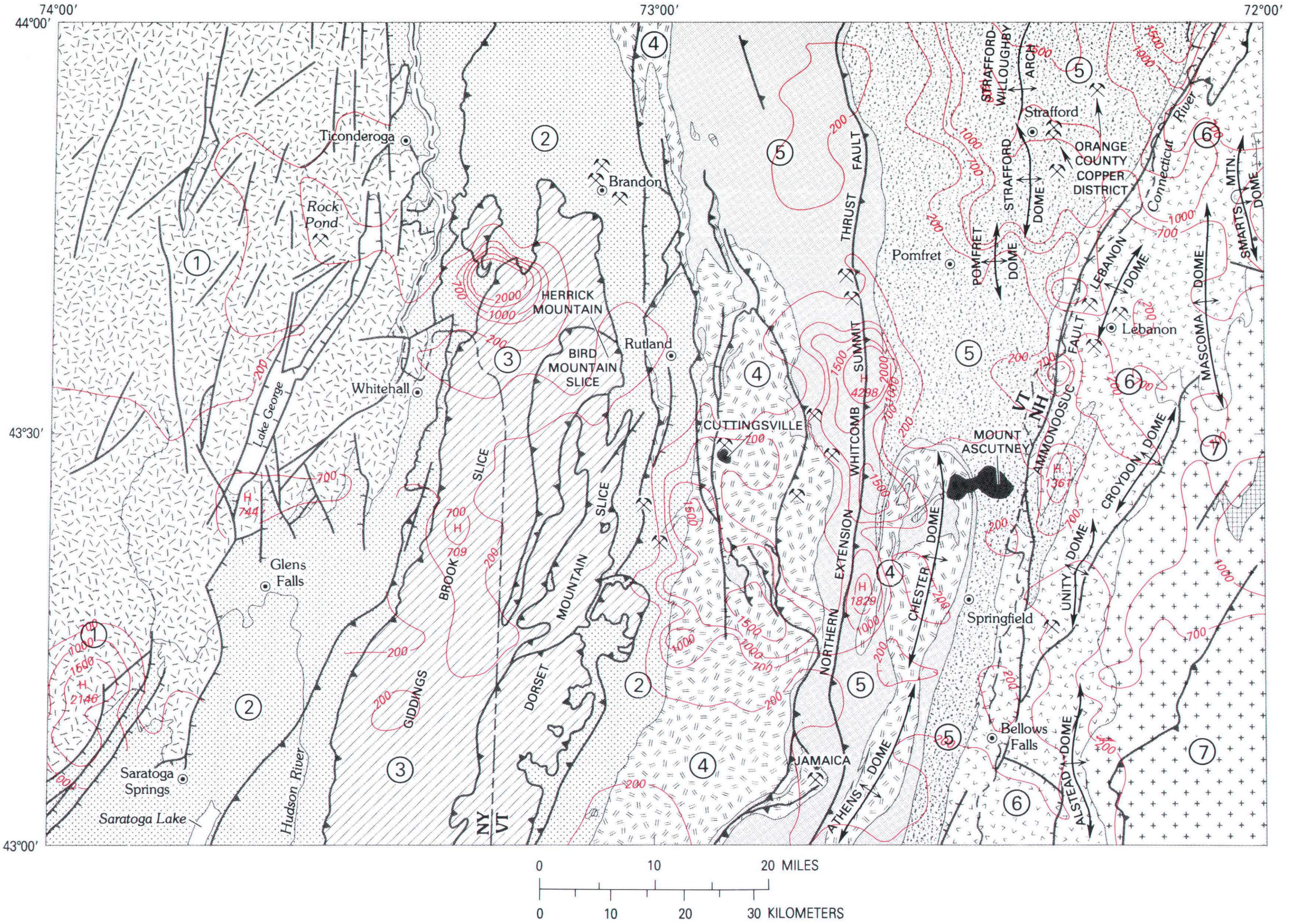

Figure 15. Isopleth map of manganese in the Glens Falls quadrangle. Isopleths define areas where samples contain at least 200, 700, 1,000, 1,500, or 2,000 parts per million manganese. Hachured areas, if shown, contain at least one sample that has a reported value below the isopleth value. See figure 1 for explanation of geologic map units and symbols.

and pebbles in the streambeds of this anomalous area are heavily coated with hydrous oxides of manganese and iron. Analyses of the coatings (Welsch and Watts, 1986) show them to be slightly enriched in cobalt and zinc, relative to the same materials collected elsewhere. Anomalous manganese in nonmagnetic heavy-mineral concentrates from this area suggests a primary manganese-rich source, which provided the manganese for mobilization in the supergene zone and later precipitation in the streambeds. Sulfide bodies may constitute such a bedrock source of high manganese.

Very little manganese was detected in Terrane 2. Manganese-oxide staining was seen on outcrops in the area, but apparently nonmagnetic heavy minerals such as carbonate or calc-silicate minerals that may have high manganese are sparse or absent in samples from this carbonate-siliciclastic shelf sequence.

A broad zone of manganese anomalies occurs in the northern part of Terrane 3 (fig. 15). The manganese occupies a larger area than the anomalies for lead, copper, or barium, for example, but the manganese isopleths overlap with those for the other elements. A clue to the source of these anomalies may be in the observations of Dale (1899, p. 190, 260) who noted thin layers of manganese carbonates (rhodochrosite) within the red slates of the northern Taconics.

A large isopleth pattern in Terrane 4 extends from the area surrounding the Cuttingsville pluton at about lat 43°30′ for a distance of about 25 to 35 km to the south. This manganese anomaly probably delimits a zone of either carbonate veins or calc-silicate rock. Calcium has a similar anomaly pattern in this area, and as discussed in the section on calcium, Mesozoic (or at least post-Proterozoic) carbonate-bearing veins may be a source of the manganese anomalies.

A short distance south of lat 43°30′, manganese forms a broad linear anomaly, largely restricted to an area between the Chester dome and the northern extension of the Whitcomb Summit thrust fault. An isopleth high immediately to the west of the Chester dome is clearly restricted to the lithostratigraphic sequence in the upper plate of the thrust. Coticule, or fine-grained quartz-spessartine (Mn-garnet) rock, also is present in the area, at the base of the Moretown Formation (J.B. Thompson, Jr., Harvard University, oral commun., 1985). The manganese-rich zone continues north of lat 43°30′ several kilometers in the upper plate of the thrust fault, crosses over to the west, and follows the distribution of the lower plate sequence. The isopleths further widen in the lower plate to the west and coincide with the increase in outcrop area of the Stowe and Pinney Hollow Formations and the increased areal density of greenstone bodies (see Doll and others, 1961). The distribution of manganese anomalies suggests that volcanogenic mineralization may be present in the Barnard Volcanic Member of the Missisquoi Formation in the area of the Chester dome and may be associated with nearby greenstones in the Stowe and Pinney Hollow Formations.

Also in Terrane 5, manganese forms a broad halo in and surrounding the Orange County copper district (fig. 15). An isopleth maxima for manganese encompasses the Gove and Elizabeth mines located just to the east of the Strafford-Willoughby arch. The manganese isopleths in general follow the geometry of the arch and have a flat concentration gradient on the east and an abruptly steep gradient on the west. The manganese anomalies may represent the lateral extent of early Paleozoic ore-forming processes in the district (see Slack and others, this volume). The effect of the arching may have been to expose chemical metasedimentary rocks rich in manganese, such as the coticule beds described at the Elizabeth mine (Annis and others, 1983) or the bedded magnetite iron formation and coticule present locally within the Standing Pond Volcanics (J.F. Slack, oral commun., 1986). The anomalous manganese zone thus may be interpreted as part of a broad hydrothermal manganese halo related to the chemical-exhalative processes believed to have formed the strata-bound sulfide deposits of the Orange County copper district (Annis and others, 1983; Slack and others, this volume).

The similar distribution of isopleths of anomalous manganese, calcium, and strontium (and locally tungsten) suggests the alternative possibility that manganiferous carbonate veins may be a source of manganese anomalies along the trend of the Strafford-Willoughby arch. Remobilization of primary sedimentary manganese as a result of a heat gradient beneath the Strafford-Willoughby arch could have deposited manganiferous carbonate into dilatant zones in the cover rocks, derived perhaps from manganiferous chemical-exhalative sources (for example, coticules) within the metasedimentary sequence.

A southern extension of the large manganese isopleth pattern that follows the trend of the Strafford-Willoughby arch coincides precisely with the axis of the Pomfret dome (fig. 15). This southern anomaly may likewise result from the exposure to erosion of manganiferous coticule beds similar to those in the Orange County copper district to the north.

A trend of manganese isopleths follows the eastern side of the Ammonoosuc fault, southeast of the Orange County copper district in Terrane 6. One isopleth encloses the area of the Reservoir Hill lead-silver-gold mine on the western flank of the Lebanon dome (fig. 15) (see Slack and Schruben, this volume). This manganese anomaly reflects the presence of primary manganese enrichments related to the formation of the Reservoir Hill deposit.

A very strong manganese high occurs to the east of the Ammonoosuc fault at the latitude of Mount Ascutney, also in Terrane 6. Samples of metamorphosed volcanic tuff collected from the area of this manganese high contain anomalous zinc, gold, arsenic, and lead; heavy-mineral concentrates from here also are strongly enriched in other elements. The manganese high may delineate a zone of volcanogenic mineralization within the lower Paleozoic rock sequence, particularly the Ammonoosuc Volcanics. Complex aeromagnetic anomalies (D.L. Daniels, oral commun., 1986), anomalous drainage geochemistry, and anomalous bedrock geochemistry support the conclusion that this zone has been a locus of igneous activity and hydrothermal mineralization during the early Paleozoic.

Isopleths representing higher-than-median manganese values for the Glen Falls quadrangle cover most of the area of Terrane 7. The isopleths nearly delineate the boundary of Terrane 7 with adjacent terranes and also indicate that manganese contents of samples collected to the east of the Glens Falls quadrangle may contain even higher values of manganese. The manganese patterns (geometric mean for Terrane 7=486 ppm, table 8) seem to indicate a decreasing gradient of manganese outward (westward) from the core of the main body of Bethlehem Gneiss (the Cardigan pluton) that may have petrologic significance (see Billings, 1956; Thompson, this volume).

R-MODE FACTOR ANALYSIS

To further evaluate the geochemical data, R-mode factor analysis was used to examine element associations on the basis of interelement correlations. The R-mode factor analysis uses a varimax solution of log-transformed data based on the total Glens Falls geochemical data set for all 1,286 heavy-mineral samples. In simple terms, R-mode factor analysis (for example, Harman, 1960) is a method of reducing a large number of variables in a multivariate data set to simpler relations having fewer variables. These fewer variables are then related to underlying geologic, geochemical, or environmental causes that determine the interelement relations. For example, if an element is confined to a single factor, its occurrence may be explained by a single underlying geologic, geochemical, or environmental control, and if an element appears in more than one factor, then there is more than one process controlling its distribution. The number of factors is selected by trial and error, beginning with the maximum number allowed by the computer program (10) and then working downward to a set of element associations that can be explained on the basis of geochemical behavior and (or) geologic association. As a practical rule of thumb, however, nearly the same result can be achieved by selecting factors on the basis of eigenvalues near to, and above, 1.000 (see Harman, 1960, for a discussion of eigenvalues).

The seven-factor model used here accounts for 58.1 percent of the data variance (table 9). A 28-factor model would be required to account for 100 percent of the variance because of the high variability inherent in heavy-mineral data and because the data base is derived from a large geographic area. The seven-factor model produces element associations that are easily explained on the basis of geologic, mineralogic, and geochemical similarities of the elements.

Elements that constitute an association for each factor were identified by inspection of the factor loadings (correlation of a variable with each factor) shown in table 9. Positive factor loadings at values greater than or equal to 0.30 indicate that an element is probably a significant member of the assemblage constituting the factor and indicates a common chemical behavior and mineralogic occurrence for the element and other elements within the factor association.

The percentage of element variance for each factor, in effect the degree to which an element is influenced by a factor, can be determined from table 9 by squaring the factor loading and multiplying by 100. For example, from table 9 about 33 percent of the data variance for iron resides in Factor 4, 25 percent resides in Factor 1, and 12 percent resides in Factor 2.

The perspective contour plots of the seven R-mode factors (see figs. 16–22) show the map distribution of element associations on the basis of R-mode factor scores (or the amount a factor is represented at each sample site). Factor peaks on the perspective diagrams range from 0.0 factor score units to the maximum for each range. The perspective contour plots of the factor scores form map distribution patterns similar to the isopleths of the key anomalous elements. The element associations indicate mineral correlations in the samples, and the R-mode factors contribute to a better interpretation of the geochemical anomalies. Moreover, elements that are not amenable to contouring, because of sparse anomaly distribution, are represented in the associations of the R-mode factors and give a clearer indication of the type of geologic environment that may have caused the geochemical anomalies. Elements constituting each R-mode factor are listed in order of decreasing strength of loading.

Factor 1

The association of V, Nb, Ti, B, Mn, Cr, Cu, Fe, and Mg accounts for the largest proportion of data variability in the 1,286 sample set (table 9). The element association suggests the coexistence in the samples of rutile (Ti and Nb) and (or) sphene (V, Nb, Ti, and Cr), tourmaline (B, Fe, Mg±Ti, and Cu), and probably pyrite (Fe±Mn and Cu). This proposed coexistence is supported by mineralogic scans that have documented the occurrence of these minerals in the heavy-mineral concentrates.

The most prominent peaks of high Factor 1 scores on the perspective plot (fig. 16) are along the Connecticut River, particularly in the northeastern corner of the Glens Falls quadrangle. An abrupt break between peaks to the east of the Connecticut River and lows in terranes to the west

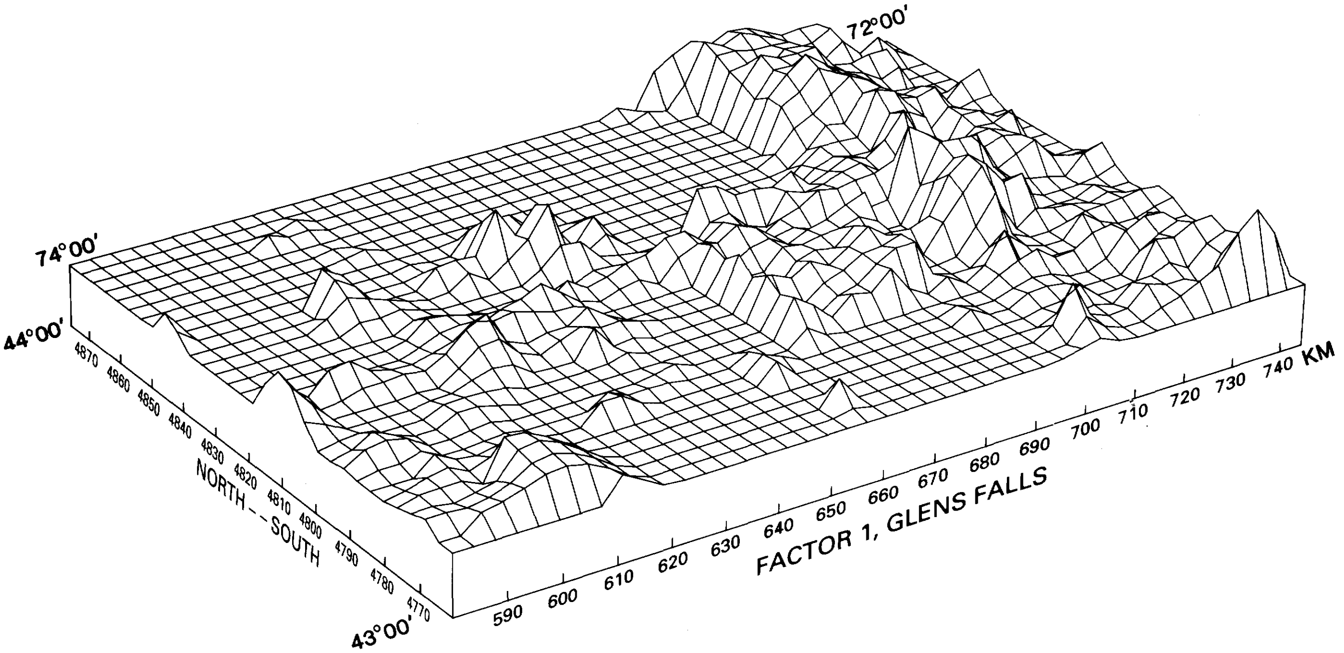

Figure 16. Perspective plot showing the distribution of Factor 1 (V, Nb, Ti, B, Mn, Cr, Cu, Fe, Mg) in the Glens Falls quadrangle. Perspective view is from S. 35° E. and 25° above the horizontal. Range of Factor 1 scores is −3.98 to 2.19. Perspective plot shows values of positive score units above 0.0.

emphasizes the contrasting geology on the two sides of the Ammonoosuc fault (see Thompson, this volume).

The perspective plot of high Factor 1 scores also shows an east-west trend of alkaline earth-rare metal anomalies at about lat 43°30′, west of the Connecticut River (discussed in more detail in the section Regional Anomaly Trends). The similarity of the distribution of Factor 1 scores to the single element plots of both niobium (fig. 10) and boron (fig. 11) along the east-west trend indicates that rutile and tourmaline are contributors to this trend.

Microscope scans of nonmagnetic concentrate samples from throughout the quadrangle show that rutile is far more widely distributed in samples than are spectrographic readings of anomalous niobium. Thus, niobium anomalies probably reflect rutile from a geochemically specialized environment and not an ordinary environment where only small amounts of niobium are available for incorporation into rutile. The northerly trending map distributions along the Connecticut River, for example, could be related to metasomatism along the Strafford-Willoughby arch, and the east-west trend at latitude 43°30′ may be a significant indication of niobium-rich rutile associated with Mesozoic tectonism and igneous activity. Alkalic rocks, for example, are typically high in rutile (Herz, 1976). The important point is that rutile from hydrothermally or metasomatically altered igneous rocks is morphologically and chemically distinct from rutile of other origins (Force, 1980, p. 488). Because niobium and titanium are more enriched in the postkinematic rocks of the White Mountain Plutonic-Volcanic Suite than in the surrounding metamorphosed Paleozoic rocks (K.C. Watts, Jr., unpub. data, 1988), the anomaly pattern may be the result of niobium enrichment during the Mesozoic.

Mineralized source areas probably control the distribution of tourmaline and pyrite in this association in the Glens Falls quadrangle. Both minerals are, of course, accessories in many of the rocks of the region, but the reasons they are concentrated may be linked, indirectly, to processes that formed mineral deposits.

Factor 2

The association of La, Mg, Sc, Y, B, Ca, and Ba accounts for the second largest proportion of data variance (table 9). Many rock-forming or rock-accessory minerals may contribute to this association, but the common denominator is that most of these elements (excluding boron) can substitute for calcium in mineral structures. From a consideration of the mineralogic makeup of the samples, a principal residence may be in calc-silicate minerals. Garnet also is known to concentrate these elements and may be another source of this association. However, for reasons of magnetic susceptibility, most garnet was not partitioned into the nonmagnetic fraction analyzed for this study. Nevertheless, partitioning occurred in several instances.

On the basis of the distribution of high scores for Factor 2 (fig. 17), areas of high-grade metamorphic gneisses are likely sources of this element association, particularly in the Adirondack highlands (Terrane 1). In the Green Mountain massif (Terrane 4) and east of the Green Mountains (Terrane 5), on the other hand, Factor 2 association is prominent only in the southern half of the Glens

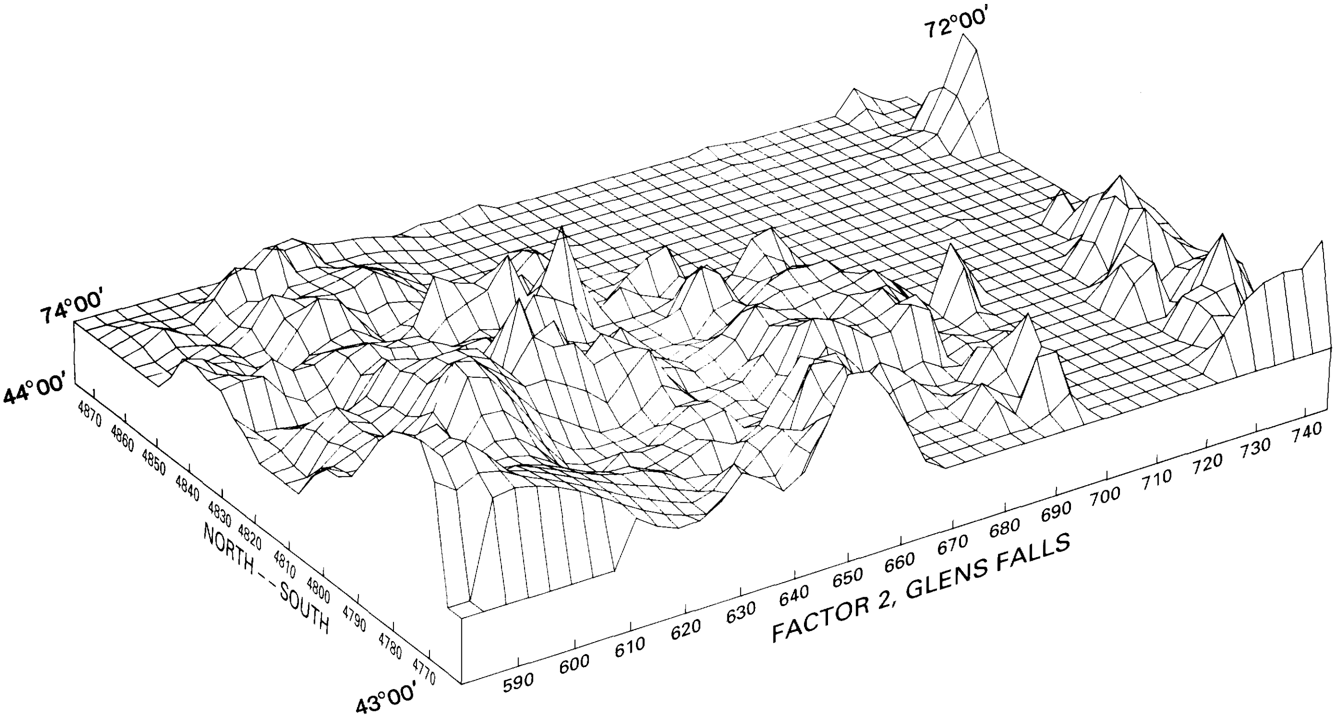

Figure 17. Perspective plot showing the distribution of Factor 2 (La, Mg, Sc, Y, B, Ca, Ba) in the Glens Falls quadrangle. Perspective view is from S. 35° E. and 25° above the horizontal. Range of Factor 2 scores is −2.41 to 3.37. Perspective plot shows values of positive score units above 0.0.

Falls quadrangle. The high factor scores in Terrane 3 are puzzling because rocks in this area are of low metamorphic grade. This metal association may be related to a detrital mineral component of the clastic rocks or to calc-silicate minerals in the carbonate units of this terrane.

The high scores for Factor 2 also show a well-defined peak in the northeastern corner of Terrane 6, apparently in association with the Smarts Mountain dome. Here, calc-silicate minerals, perhaps of hydrothermal origin, may be the source. The strong Factor 2 peaks in the southeastern corner of the Glens Falls quadrangle in effect delineate the outcrop area of granitoids of the New Hampshire Plutonic Suite (for example, Bethlehem Gneiss; Kinsman Quartz Monzonite) and therefore are probably a reflection of granitic lithologies.

The distribution of Factor 2 peaks in the central to western part of the Glens Falls quadrangle (fig. 17) shows a northwesterly trend. The boundary between highs to the south and areas lacking a Factor 2 association to the north nearly coincides with a northwesterly trend of single-element geochemical anomalies and the trend of Landsat lineaments (see Pohn, this volume).

Factor 3

Silver and gold constitute Factor 3. Because other elements that may be associated with sulfide deposits are excluded from the silver-gold factor, two modes of occurrence (perhaps interrelated) may account for this factor: (1) gold-silver placers in which leaching has removed sulfide and gangue constituents and (2) relatively pure gold-quartz veins in the bedrock of sampled drainage basins.

The distribution trend of the peaks for Factor 3 scores is northeasterly (fig. 18), which is parallel to the strike of most postkinematic (probably Mesozoic) base- and precious-metal veins in the region (Cox, 1970; Slack and Schruben, this volume). The Factor 3 peaks in Terrane 5 coincide with areas known to contain gold placers and gold-bearing quartz veins near Plymouth and Bridgewater, Vt. (see Slack and others, 1987; Slack and Schruben, this volume), and, in general, follow the belt of ultramafic rocks on the eastern side of the Green Mountains (Doll and others, 1961); the close association of the ultramafic belt with gold placers in Vermont was noted many years ago (Smith, 1908, p. 88). As the Factor 3 peaks continue northward from the Plymouth-Bridgewater area, detailed map plots of gold and silver anomalies show a close association with the occurrence of greenstones in the Pinney Hollow and Stowe Formations and with drainage basins whose bedrock geology is dominated by the Moretown Formation. These formations are thus interpreted to be the source of the gold and silver anomalies, either from quartz veins or placer deposits derived from them.

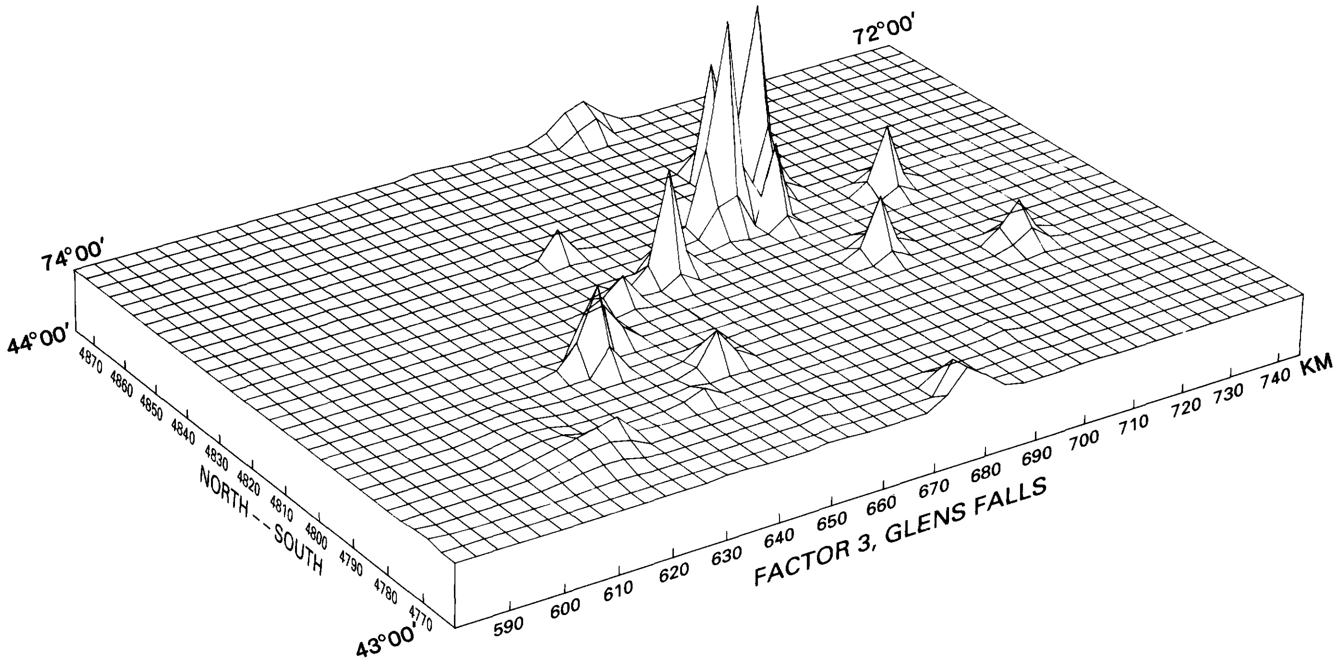

Figure 18. Perspective plot showing the distribution of Factor 3 (Ag, Au) in the Glens Falls quadrangle. Perspective view is from S. 35° E. and 25° above the horizontal. Range of Factor 3 scores is −1.66 to 14.5. Perspective plot shows values of positive score units above 0.0.

In the Taconic allochthon (Terrane 3), Factor 3 peaks are probably a reflection of quartz-breccia veins containing gold and (or) silver, perhaps analogous to the Mount Colfax occurrence near Cambridge, N.Y. (Dale, 1899, p. 186; Slack and others, 1987). R-mode factor analysis on a subset of panned concentrate data from Terrane 3 (K.C. Watts, Jr., unpub. data) generated a factor that has the association gold-beryllium-tin. In the Glens Falls region, the geochemical data suggest that these metals are characteristic of mineralization that has genetic affinities to Mesozoic tectonic and igneous activity. These observations together with the northeasterly trend of Factor 3 peaks suggest that gold deposits may be associated with Mesozoic structures along this northeasterly trend, not only in Terrane 3 but in the other terranes crosscut by aligned Factor 3 peaks.

Factor 4

A coexistence in some samples of pyrite and base-metal sulfide minerals is suggested by the element association Ni, Co, As, Fe, Cu, and Zn (loading slightly below 0.35, but probably significant) that constitutes Factor 4. Although pyrite is the most common sulfide mineral in the heavy-mineral concentrates and probably contributes the most to this association, arsenopyrite, chalcopyrite, and gahnite also are present in some samples. The perspective contour plot (fig. 19) shows that Factor 4 is most prominent in Terranes 3 and 5; these are also areas where outcrop observations show widespread occurrences of pyrite. Factor 4 accounts for about 57 percent of the Ni, 44 percent of the Co, 34 percent of the As, 33 percent of the Fe, 15 percent of the Cu, and 11 percent of the Zn variance in the entire Glens Falls quadrangle (table 9).

A massive oxidizing pyrite zone has been postulated, on the basis of acid ground and surface waters and on element anomalies, in the southwestern corner of Terrane 1. A strong peak on the perspective plot for Factor 4 (fig. 19) supports the interpretation that oxidizing massive sulfides are the cause of these abnormally acid waters in the area.

In Terrane 3, separate zones of north-trending Factor 4 peaks may correspond to fault slices of the low and high Taconic sequences. The strong Factor 4 peaks along the western edge of the Taconic allochthon may be due in part to arsenopyrite; it was seen in one outcrop of mélange during the geochemical sampling. A composite chip sample of deformed and limonitic black slate from the West Castleton Formation in the Giddings Brook tectonic slice exposed along Lewis Brook, at Gorhamtown, Vt., contains 1,500 ppm arsenic in a bulk rock sample and greater than 10,000 ppm arsenic in iron-stained zones along fractures.

Factor 4 peaks in Terrane 5 occupy most of the same areas as the single-element anomalies that constitute this factor. Peaks occur (1) immediately east of the Green Mountain core, (2) in the Orange County copper district, (3) along the Strafford-Willoughby arch, and (4) parallel to the Connecticut River. A small peak on the perspective map, in the southeastern corner of the Glens Falls quadrangle (fig. 19), coincides with a zone of multielement anomalies and a

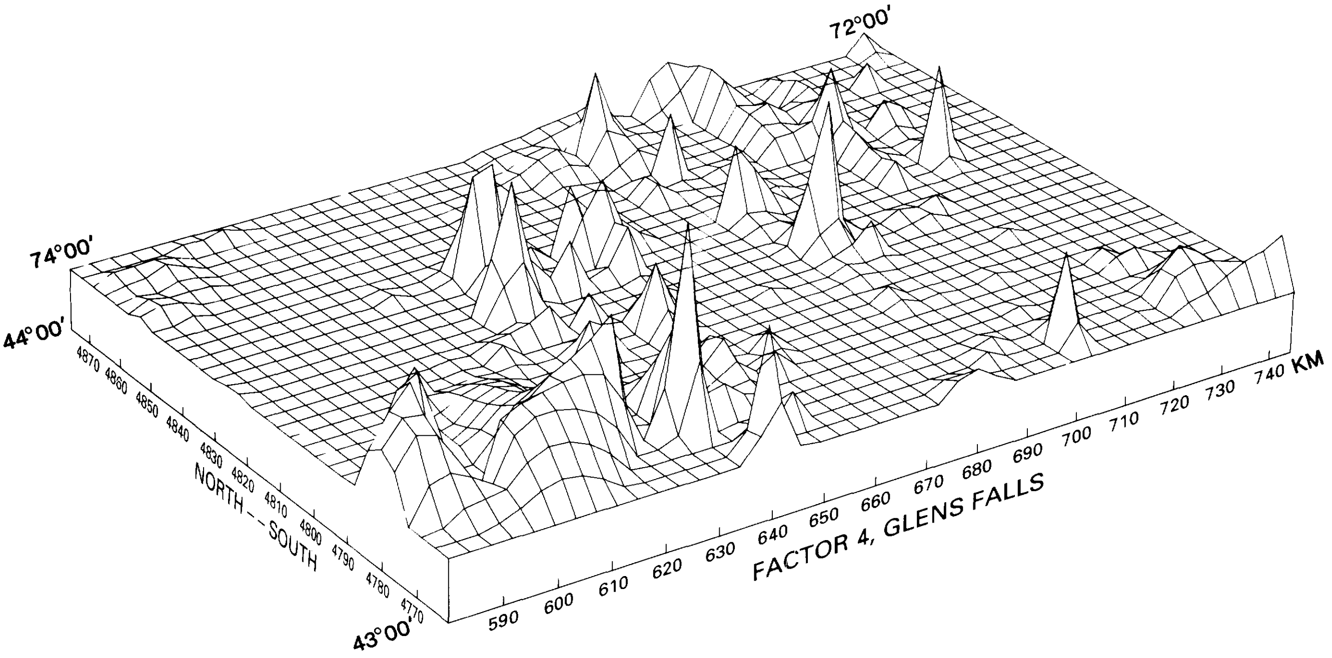

Figure 19. Perspective plot showing the distribution of Factor 4 (Ni, Co, As, Fe, Cu, Zn) in the Glens Falls quadrangle. Perspective view is from S. 35° E. and 25° above the horizontal. Range of Factor 4 scores is −2.20 to 7.82. Perspective plot shows values of positive score units above 0.0.

possible source area of acid ground water (National Uranium Resource Evaluation data; Koller, 1979; Watts, in press); this coincidence suggests that the anomaly source may be oxidizing pyrite (or pyrrhotite).

Factor 4 can be a significant tool in locating sulfide-metal occurrences in the Glens Falls region, particularly massive, stratabound sulfides that may contribute large quantities of detrital sulfides to drainage basin alluvium.

Factor 5

The element suite Sn, Pb, Sb, Ba, and Cu appears to be a sulphosalt(?)-barite-barium carbonate(?) association largely restricted to the Taconic allochthon of Terrane 3 (fig. 20). Table 9 shows that about 50 percent of the Sn, 48 percent of the Pb, 26 percent of the Sb, 13 percent of the Ba, and 13 percent of the Cu variance in the Glens Falls quadrangle is accounted for by Factor 5.

Barite and carbonate minerals enriched in barium were identified in panned concentrate samples from the Taconic allochthon. Sulphosalt minerals were not seen in cursory microscope scans, but a more careful search may find them. Tin is present in anomalous amounts in a large proportion of the samples from Terrane 3, and reported values exceed 2,000 ppm in some samples. Cassiterite has not been observed in the samples and probably is absent. As yet, no discrete tin minerals have been identified from Terrane 3, but the high tin concentrations suggest the presence of a discrete tin mineral (for example, stannite).

Factor 6

Factor 6, which includes Ca, Sr, Mn, Co, Pb, and Y, is a reflection of carbonate-vein material, chemical metasedimentary rocks, possible carbonatized zones, and probable igneous and (or) sedimentary apatite. The factor is not derived from calcareous metasedimentary rocks. Part of the factor may be contributed by the weathering of lead-bearing carbonate veins and their residual concentration of lead (and perhaps cobalt) at the surface in secondary manganese oxides. The perspective plot of Factor 6 scores (fig. 21) shows no relation to the distribution of calcareous lithologies of the Waits River Formation (see Doll and others, 1961) or to the shelf carbonate rocks composing Terrane 2. The presence of yttrium in the metal association supports the interpretation that carbonate veins contribute to the factor, because yttrium commonly substitutes for calcium in mineral structures and is particularly enriched in vein calcite (Rankama and Sahama, 1950, p. 524–525).

One of the most prominent features shown by the high Factor 6 scores is a north-trending pattern east of the Green Mountains in Terrane 5. The carbonate mineral signature expressed by Factor 6 in this area is no doubt due to the occurrence of carbonate veins and talc-carbonate rock. The veins also may be sources of some metal anomalies in the area, as suggested by the presence of lead in the Factor 6 association.

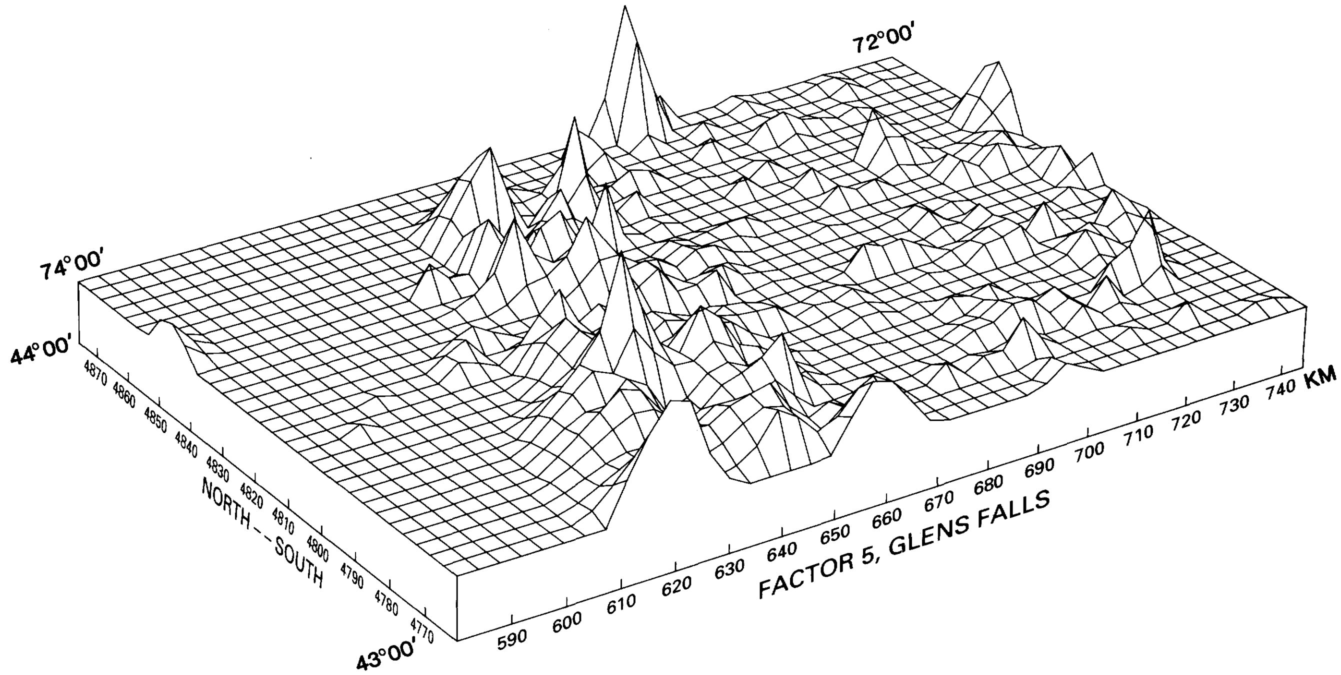

Figure 20. Perspective plot showing the distribution of Factor 5 (Sn, Pb, Sb, Ba, Cu) in the Glens Falls quadrangle. Perspective view is from S. 35° E. and 25° above the horizontal. Range of Factor 5 scores is −2.18 to 6.93. Perspective plot shows values of positive score units above 0.0.

Factor 7

Factor 7 comprises the suite W, Bi, Mo, and Be and most likely reflects mineralization related to felsic igneous activity. The association is dominated by scheelite-related tungsten. Table 9 shows that Factor 7 accounts for about 38 percent of the W, 30 percent of the Bi, 29 percent of the Mo, and 22 percent of the Be variance in the Glens Falls quadrangle. The remainder of the variance for these metals is present in other associations not accommodated in the seven-factor model.

Terranes 6 and 7 show the most pronounced affiliation with the Factor 7 suite of elements, whereas other peaks in the Glens Falls quadrangle are somewhat scattered (fig. 22). The metal suite of Factor 7 is notably present in minerals within posttectonic quartz veins at Holts Ledge, N.H., on the northwestern side of the Mascoma dome (Slack and others, 1987, p. 6), which gives a clue to the geologic environment of the source of these metals in other parts of the Glens Falls quadrangle. Crosscutting veins are thus one, if not the principal, source of the Factor 7 metal association. A minor source probably is pegmatites associated with the Bethlehem Gneiss in New Hampshire, which locally contain molybdenite and beryl (Morrill, 1960).

REGIONAL ANOMALY TRENDS

The regional isopleth patterns of the key ore-forming and pathfinder elements (figs. 4–15) are interpreted as delineations of metallogenic-tectonic zones and mineralized stratigraphic sequences. Three general types of zones are recognized, including those that (1) crosscut the lithotectonic terranes, (2) conform to the north-south structural and stratigraphic grain within the lithotectonic terranes, and (3) bound the lithotectonic terranes. The isopleth maps (figs. 4–15) show anomaly trends that correlate with the distribution of R-mode associations represented as peaks on the perspective contour maps (figs. 16–22). However, as discussed, metals involved in the R-mode factor suites also include those found to be locally anomalous and significant metallogenically but not individually amenable to contouring.

A method of analysis called Relative Element Magnitude (REM) (Van Trump and Alminas, 1978) was applied to discrete areas or anomaly trends that were determined from overlays of the geochemical isopleth maps. These areas or "windows" were examined on the basis of two parameters: (1) anomaly-to-background contrast for each element sought in the emission spectrographic analyses and (2) relative proportion of anomalous samples for each element.

Crosscutting

As with the other inferred geochemical anomaly trends, boundaries of the crosscutting regional zones (fig. 23) were drawn on the basis of maximum areal distribution

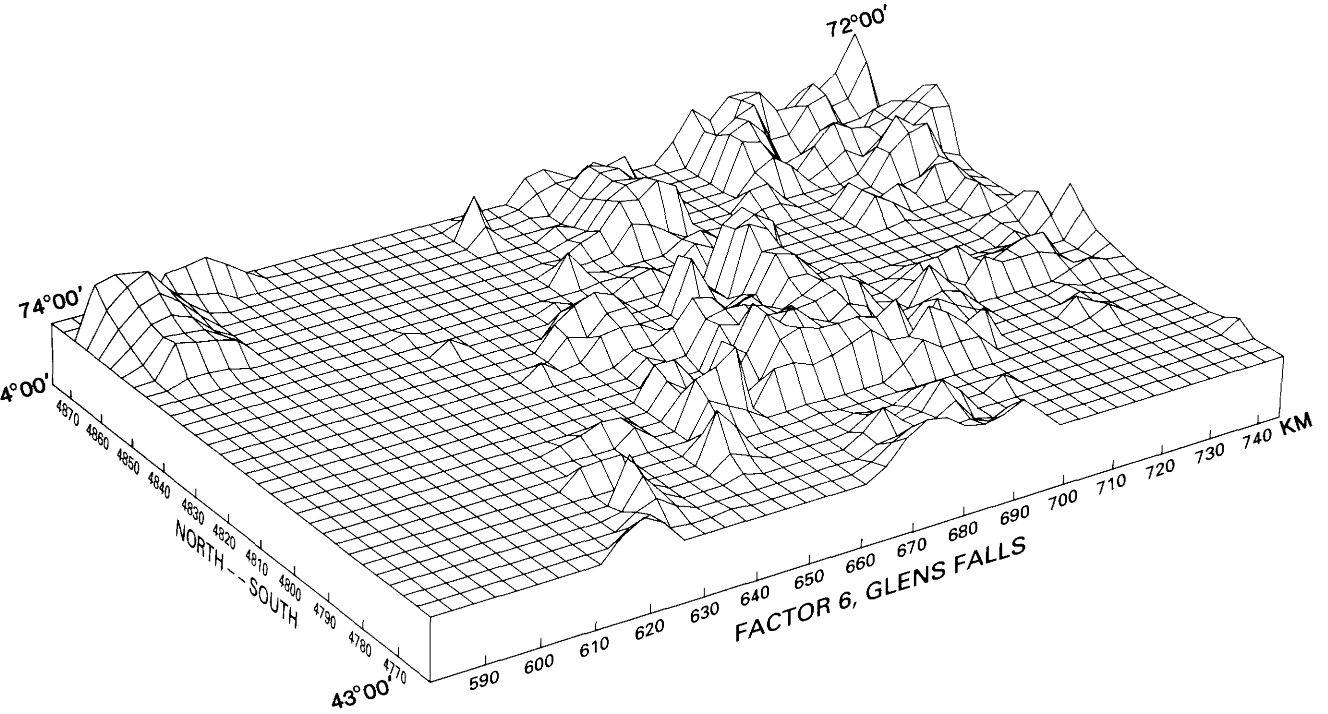

Figure 21. Perspective plot showing the distribution of Factor 6 (Ca, Sr, Mn, Co, Pb, Y) in the Glens Falls quadrangle. Perspective view is from S. 35° E. and 25° above the horizontal. Range of Factor 6 scores is −3.59 to 2.40. Perspective plot shows values of positive score units above 0.0.

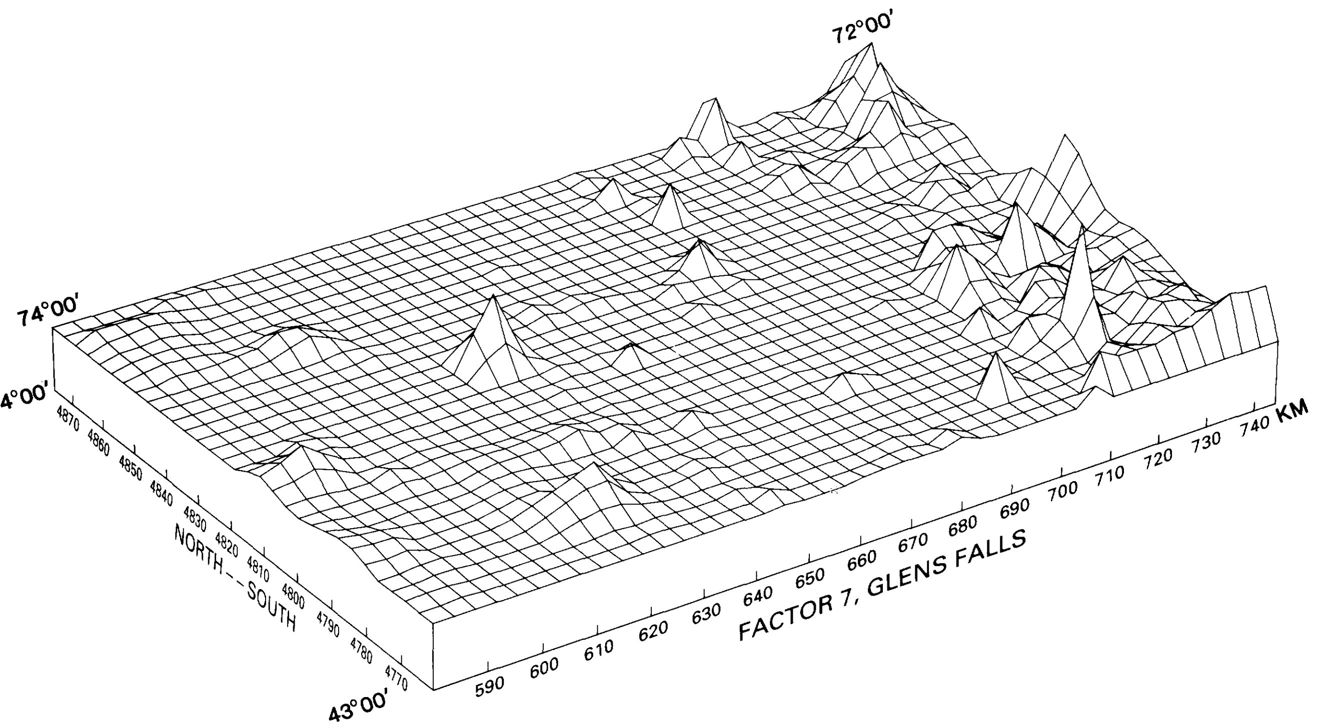

Figure 22. Perspective plot showing the distribution of Factor 7 (W, Bi, Mo, Be) in the Glens Falls quadrangle. Perspective view is from S. 35° E. and 25° above the horizontal. Range of Factor 7 scores is −1.40 to 7.14. Perspective plot shows values of positive score units above 0.0.

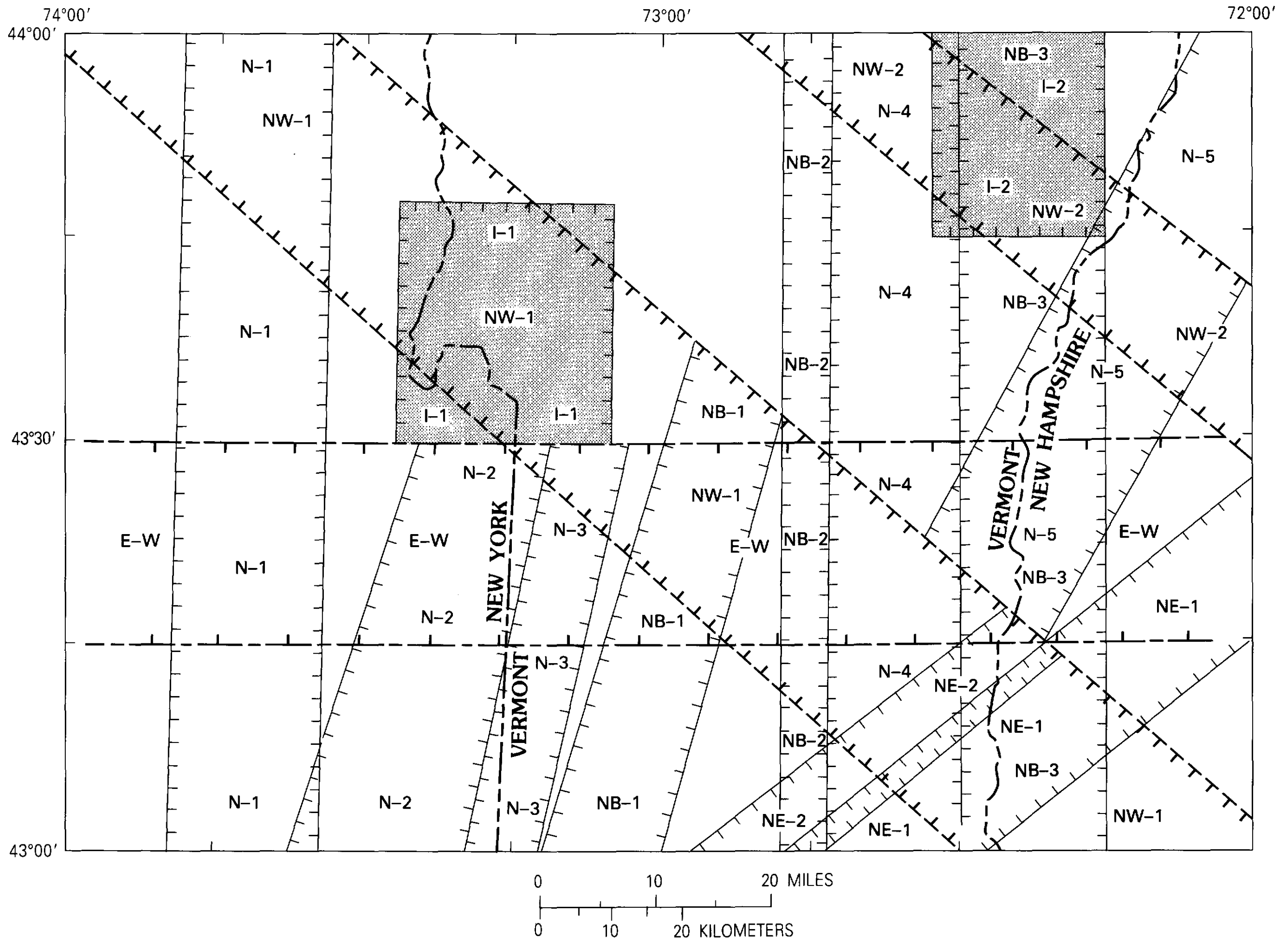

Figure 23. Regional geochemical trends and Relative Element Magnitude (REM) windows in the Glens Falls quadrangle. Shaded areas are intersection areas I–1 and I–2.

and overall trend derived from stacked overlays of the isopleth maps. Associations of elements in the geochemical anomalies are reflected in complementary regional isopleth trends. Areal overlap of regional isopleths of the various metals may indicate regional metal zoning. Some of the inferred trends are drawn on the basis of the signature of a single dominant chemical element; others are based on a combination of anomalous elements. The isopleth maps, and trends derived from them, were also overlain for comparison on 1:250,000-scale geologic (Thompson and others, 1990), aeromagnetic and gravity (D.L. Daniels and C.L. Long, unpub. data), and interpretive lineament maps (Pohn, 1986). Although the geochemical data compose the sole basis for delineating the crosscutting zones, much of the other data is supportive. The principal crosscutting isopleth trends thus delineated are northwesterly, northeasterly, and westerly. These trends cut at least two and, in some cases, several of the lithotectonic terranes of the region.

The most conspicuous of the northwesterly trends of crosscutting anomalies consists of an alignment of geochemical highs near the central part of the Glens Falls quadrangle (fig. 23, NW–1). These geochemical highs delineate a discontinuous, rather than a continuous, zone. The zone boundary was drawn on the basis of the isopleth maps for lead, copper, and tin (figs. 4, 5, and 9). The alignment of geochemical highs, and the slight elongations of isopleth configurations in a northwest-southeast direction (particularly on the tin isopleth map at the northern end of Terrane 3), suggests a genetic link.

Barium lows (fig. 12) that have complementary adjacent highs are also characteristic of the northwest-trending geochemical trend NW–1. Some of the barium lows coincide with positive anomalies for other metals along the zone. Barium may have been depleted in these zones as a result of hydrothermal leaching of rock feldspars and perhaps was redistributed to nearby areas now outlined by the regional barium highs.

From the REM computations (table 10), the NW–1 zone is dominated by lead, followed by copper, tin, and boron. A few samples within the NW–1 trend contain anomalous gold, tungsten, molybdenum, and bismuth, suggesting the possible presence of epigenetic veins. The relative ranking of these metals is supported by the isopleth maps in which lead dominates in anomaly-to-background contrast and drainage basins that have copper anomalies occupy the largest area within the NW–1 zone.

In addition to aligned geochemical highs, the NW–1 zone also coincides nearly precisely with a northwestern trend of Landsat and radar lineaments (see Pohn, this volume). These lineaments transect the lithotectonic terrane boundaries obliquely on a N. 60° W. to N. 65° W. trend and are coextensive with the geochemical anomaly trend. The lineaments are considered to be the surface expression of a system of deep-seated, basement fractures or faults of continental extent that have printed through the younger cover rocks, perhaps as a result of tectonic reactivation (Pohn, this volume). The geochemical anomalies independently showing this trend support that interpretation based on distribution and anomaly-to-background contrast and further suggest the existence of former solution pathways within the broad NW–1 zone, which have allowed metals to migrate locally into the cover rocks.

A second northwest-trending alignment of geochemical highs is present in the northeastern part of the Glens Falls quadrangle (fig. 23, NW–2). The boundaries for this trend are based on geochemical criteria similar to those used to outline trend NW–1. Lead in particular was useful in defining these boundaries (fig 4). As with trend NW–1, trend NW–2 appears to coincide with a zone of Landsat and radar lineaments (Pohn, this volume) and, moreover, encompasses several types of epigenetic mineral deposits, including lead-zinc-silver veins in the Thetford and Norwich areas of eastern Vermont (Slack and Schruben, this volume) and the tungsten-molybdenum veins at Holts Ledge along the northwestern margin of the Mascoma dome in New Hampshire (see Slack and others, 1987).

REM computations for the NW–2 geochemical trend (table 11) show the sequence of dominant elements to be Pb, Cu, Sn, Mn, Sr, B, and Nb. One anomalous value places silver higher in the REM ranking than it should occupy and is therefore discounted. Table 11 indicates that copper is more broadly distributed than the other anomalous elements within the NW–2 zone and that lead is higher in geochemical contrast. This is a common case in geochemical exploration, as lead tends to form residual concentrations in the surficial zone because of its low supergene mobility from an oxidizing sulfide system, whereas the high supergene mobility of copper allows it to migrate significantly, both laterally and vertically. Table 11 also shows that bismuth, antimony, arsenic, and zinc constitute part of the suite in a modest number of samples from zone NW–2.

Table 10. Relative Element Magnitude (REM) values for crosscutting regional trend NW–1 (360 samples)

[These are relative figures used to indicate the position of chemical elements in terms of prominence. Numbers are rounded and do not represent precision greater than is present in the initial data]

Element	Anomalous samples	Anomaly mean	Intensity[1]	Area, percent anomalous	EM[2]	REM, cumulative percent
Pb	221	296	9.9	61.4	605.6	14.6
Cu	235	122	8.2	65.3	532.6	12.8
Sn	160	130	8.7	44.4	386.1	9.3
B	224	721	4.8	62.2	299.2	7.2
Ba	189	537	5.4	52.5	281.8	6.8
Mn	266	745	3.7	73.9	275.1	6.6
La	265	161	3.2	73.6	236.8	5.7
Mg	247	1.4	2.8	68.6	194.5	4.7
Ca	287	12	2.4	79.7	191.1	4.6
Sr	245	511	2.6	68.1	173.9	4.2
Co	151	28	3.9	41.9	165.3	4.0
Be	42	19	12.4	11.7	144.8	3.5
Fe	203	2.3	2.3	56.4	122.6	3.1
Y	264	350	1.7	73.3	128.3	3.1
Cr	201	195	1.9	55.8	108.6	2.6
V	214	208	1.4	59.4	82.4	2.0
Nb	196	102	1.5	54.4	79.6	1.9
Ni	15	73	10.5	4.2	43.6	1.1
Au	2	724	48.3	.6	26.8	.6
W	24	281	4.0	6.7	26.8	.6
Mo	4	154	15.4	1.1	17.1	.4
Ag	4	12	7.8	1.1	8.7	.2
Bi	3	123	8.2	.8	6.8	.2
As	4	1,350	4.5	1.1	5.0	.1
Zn	4	638	2.1	1.1	2.4	.1
Sb	3	333	2.2	.8	1.9	0
Total					4,153.3	

[1] Intensity=Mean/Threshold.

[2] EM=Intensity × Area.

The northeasterly trend of crosscutting geochemical patterns (fig. 23, NE–1) is best displayed in the southeastern corner of the Glens Falls quadrangle. In particular, tin isopleths (fig. 9) clearly define a northeasterly trending zone that encompasses areas near the Alstead dome, extending approximately along the contact of the Sunapee two-mica granite and the Bethlehem Gneiss, coextensive with a band of calcareous metasedimentary rocks (see Thompson, this volume). This same pattern is replicated by an alignment of discontinuous lead, copper, and tungsten highs (figs. 4, 5, and 8) and is also consistent with the north-to-northeast trends of known Mesozoic faults and metalliferous veins in this part of New England (Cox, 1970). In support of the map distributions, REM computations (table 12) show the dominance of tin in this trend of anomalies, followed by lead and copper.

Chromium and magnesium rank high in REM for zone NE–1 as well and suggest that the mineral fuchsite (chromium mica) may have some influence on the geochemical signature of samples from this zone. Although fuchsite was not observed in cursory microscope scans of the panned concentrate samples, it has been described in the Clough Quartzite near Acworth, N.H. (Clifford, 1957), which is within the NE–1 zone. The significance of possible widespread fuchsite within zone NE–1 lies in its common association with gold deposits (for example, Mother Lode district, California; Larder Lake and Porcupine districts,

Table 11. Relative Element Magnitude (REM) values for crosscutting regional trend NW–2 (131 samples)

[These are relative figures used to indicate the position of chemical elements in terms of prominence. Numbers are rounded and do not represent precision greater than is present in the initial data]

Element	Anomalous samples	Anomaly mean	Intensity[1]	Area, percent anomalous	EM[2]	REM, cumulative percent
Pb	70	809	26.9	53.4	1,440.2	23.9
Ag....	1	1,500	1,000.0	.8	763.4	12.7
Cu....	104	127	8.5	79.4	672.5	11.2
Sn	78	158	10.5	59.5	626.7	10.4
Mn ...	109	906	4.5	83.2	377.1	6.3
Sr.....	119	612	3.1	90.8	277.9	4.6
B	96	502	3.3	73.3	245.3	4.1
Nb....	102	199	2.8	77.9	221.6	3.7
Cr	102	197	2.0	77.9	153.4	2.6
V.....	98	300	2.0	74.8	149.6	2.5
Co....	79	17	2.4	60.3	144.5	2.4
Fe	90	2	2.0	68.8	139.3	2.3
Be....	26	10	6.6	19.6	130.8	2.2
Ca....	101	7.6	1.5	77.1	116.6	1.9
W	59	161	2.3	45.0	103.3	1.7
La	74	81	1.6	56.5	91.4	1.5
Ba....	58	201	2.0	44.3	88.9	1.5
Ni	59	13	1.8	45.0	80.7	1.3
Au....	2	729	48.6	1.5	74.2	1.2
Mg ...	52	.6	1.3	39.7	51.9	.9
Y.....	20	310	1.5	15.3	23.7	.4
Bi	4	113	7.5	3.0	22.9	.4
Sb	3	667	4.4	2.3	10.2	.2
As	2	1,250	4.2	1.5	6.4	.1
Zn	1	700	2.3	.8	1.8	0
Total					6,014.3	

[1] Intensity=Mean/Threshold.

[2] EM=Intensity × Area.

Table 12. Relative Element Magnitude (REM) values for crosscutting regional trend NE–1 (79 samples)

[These are relative figures used to indicate the position of chemical elements in terms of prominence. Numbers are rounded and do not represent precision greater than is present in the initial data]

Element	Anomalous samples	Anomaly mean	Intensity[1]	Area, percent anomalous	EM[2]	REM, cumulative percent
Sn	33	343	22.9	41.8	956.1	30.5
Pb	54	124	6.2	68.3	422.8	13.5
Cu....	26	127	6.4	32.9	209.5	6.7
Cr	54	301	3.0	68.3	205.7	6.6
Mg ...	47	.4	2.3	59.5	134.8	4.3
Fe	30	3.0	3.5	38.0	134.0	4.3
Nb....	60	88	1.7	75.9	133.2	4.2
W	23	309	4.4	29.1	128.6	4.1
Ca....	44	11	2.3	55.7	125.8	4.0
Bi	3	450	30.0	3.8	113.9	3.6
La	18	246	3.5	22.8	79.9	2.5
Mo ...	1	500	50.0	1.3	63.3	2.0
Sr.....	34	279	1.4	43.0	60.1	1.9
Y.....	33	667	1.3	41.8	55.7	1.8
Be....	21	10	2.0	26.6	53.7	1.7
Co....	11	25	3.6	13.9	49.7	1.6
B	15	373	2.5	19.0	47.3	1.5
V.....	30	183	1.2	38.0	46.4	1.5
Ni	2	105	15.0	2.5	38.0	1.2
Ba....	15	573	1.9	19.0	36.3	1.2
Mn ...	19	911	1.3	24.0	31.3	1.0
Ag....	1	10	6.7	1.3	8.4	.3
Zn	1	500	1.7	1.3	2.1	.1
Total........................					3,136.6	

[1] Intensity=Mean/Threshold.

[2] EM=Intensity × Area.

Ontario) and its possible utility as a pathfinder for gold (Whitemore and others, 1946). Table 12 suggests that as many as 54 panned concentrate samples from zone NE–1 may contain fuchsite, although its presence has not yet been documented.

A pattern of northeasterly trending cobalt isopleths (fig. 7) in the south-central part of the Glens Falls quadrangle are the basis for drawing the boundaries for crosscutting geochemical trend NE–2. Cobalt in this zone is most likely a constituent of pyrite. The bedrock source of these anomalies appears to be a northeasterly trending zone of cobaltiferous pyrite derived from veins or fractures of similar trend or from a coextensive pyritic lithology. REM computations for the zone (table 13) show the sequence of dominant elements to be Pb>Sn>Ba>Cu>Ca>Co. The northeasterly trending cobalt pattern in part coincides with lead anomalies near Jamaica, Vt. (fig. 4), where significant bedrock uranium deposits are known (Ayuso and Ratté, this volume). The cobalt and lead anomalies coincide with, but also extend beyond, the areas of known uranium, and their distributions suggest that they may be used as indicators of additional uranium occurrences.

An east-west trend of isopleths covers a regional zone from about lat 43°30′ southward (fig. 23, E–W) and crosscuts all pre-Mesozoic lithotectonic terranes. Many of the elements found within the zone are typically associated with alkalic igneous rocks and their related mineral deposits formed in a rift tectonic setting. Barium and strontium isopleths (figs. 12 and 14) are the basis for drawing the boundaries of the east-west trend; these two elements have marked affinities for alkalic igneous rocks. As delineated by the maximum extent of the barium and strontium isopleths, the anomalous east-west zone is about 22 km wide in a north-south direction and of undetermined, but at least tens of kilometers length, in an east-west direction. The trend of the anomalies is about the same as the direction of Mesozoic crustal extension in the region, as shown by the geometry of the St. Lawrence rift system and the Ottawa-Bonnechere graben (fig. 24) along which are emplaced the alkalic igneous complexes of the Monteregian Hills of southern Quebec (Philpotts, 1969; Kumarapeli, 1976). These igneous complexes are notably enriched in niobium, a metal also characteristic of the east-west anomaly trend, particularly in zones surrounding the two Mesozoic plutons (fig. 10). By analogy, the east-west-trending zone of anomalies in the Glens Falls quadrangle may demarcate a structural

Table 13. Relative Element Magnitude (REM) values for crosscutting regional trend NE–2 (44 samples)

[These are relative figures used to indicate the position of chemical elements in terms of prominence. Numbers are rounded and do not represent precision greater than is present in the initial data]

Element	Anomalous samples	Anomaly mean	Intensity[1]	Area, percent anomalous	EM[2]	REM, cumulative percent
Pb....	31	217	7.2	70.4	509.1	21.7
Sn....	22	88	5.9	50.0	293.9	12.5
Ba....	11	311	10.4	25.0	259.1	11.1
Cu....	12	139	7.0	27.3	189.8	8.1
Ca....	29	16	2.3	65.9	154.2	6.6
Co....	17	35	3.5	38.6	134.1	5.7
Y.....	30	383	1.9	68.2	130.7	5.6
Mn ...	34	327	1.6	77.3	126.1	5.4
Cr....	25	240	1.6	56.8	90.9	3.9
Bi....	3	167	11.1	6.3	75.8	3.2
La....	22	71	1.4	50.0	71.4	3.0
Fe....	21	1.4	1.4	47.7	65.9	2.8
W....	2	1,000	14.3	4.5	64.9	2.8
B.....	11	245	1.6	25.0	40.9	1.7
Sr.....	12	642	1.3	27.3	35.0	1.5
Nb....	10	97	1.4	22.7	31.5	1.3
Ag....	2	8.5	5.7	4.5	25.8	1.1
Be....	4	4	2.7	9.1	24.2	1.0
Mg ...	5	2	1.0	11.4	11.4	.5
V.....	4	163	1.1	9.1	9.8	.4
Total........................					2,344.5	

[1] Intensity=Mean/Threshold.

[2] EM=Intensity × Area.

weakness that controlled the emplacement of the alkalic stocks at Cuttingsville and Mount Ascutney. This zone also may have provided access for broadly coeval hydrothermal fluids to deposit a variety of metals in posttectonic veins.

Additional support for the existence of an approximately 22-km-wide, east-west-trending zone of weakness near lat 43°30′ is from field studies. The reported observations of mafic-alkalic dikes and irregular intrusive bodies within the geochemically delineated, east-west-trending zone indicate that at least some dikes of probable Mesozoic age are oriented east-west rather than northeast as commonly noted elsewhere in the Glens Falls region (McHone, 1978, 1988); therefore, Mesozoic igneous activity may have been concentrated in the east-west zone. For example:

1. Perry (1929[?], p. 48) described an east-west-trending camptonite dike on the western side of Soltudus Mountain near lat 43°30′20″ long 72°43′06″.
2. Richardson (1929[?], p. 236–242) reported a trachydacite dike, 75 m wide, apparently easterly in trend at about lat 43°27′18″ long 72°32′21″; an east-west-trending diabase dike about 120 m wide that contains syenite xenoliths at approximately lat 43°28′06″ long 72°32′15″; amygdaloidal diabase of unknown trend, approximately lat 43°28′30″ long 72°33′15″; and a syenite body near lat 43°26′33″ long 72°33′12″.

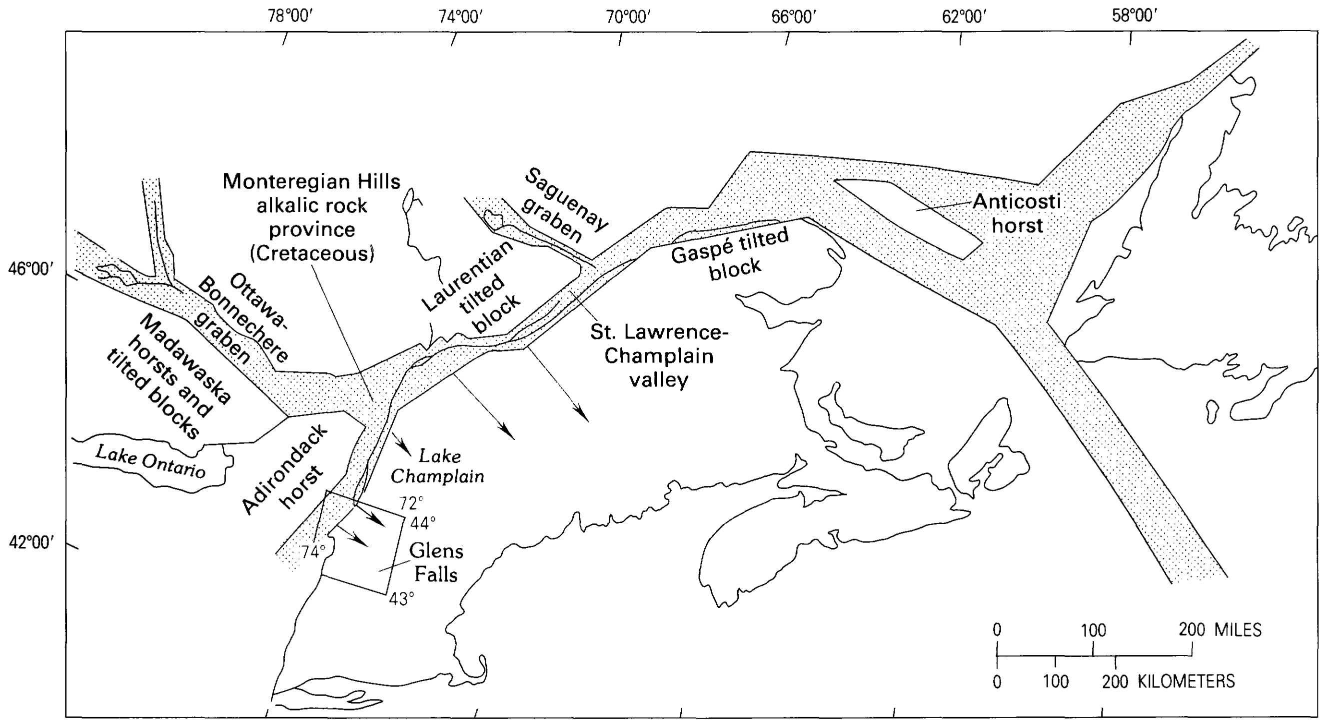

Figure 24. Rift zones in the St. Lawrence system showing suggested crustal movements (arrows) and rift features in relation to the Glens Falls quadrangle. Figure (after Kumarapeli and Saul, 1966) used with permission of The National Research Council of Canada.

3. Dale (1899, p. 222–223) described an east-west-trending augite-camptonite dike in the bed of Lewis Brook, west of Herrick Mountain, probably between lat 43°34′18″ and 43°34′27″ and long 73°10′00″ and 73°11′00″.

REM computations for crosscutting regional trend E–W (table 14) show that lead and tin are the two most definitive metals, followed by barium, copper, and manganese. The relatively high ranking of strontium and calcium in the REM computations results from the large areal distribution of samples that have metal contents above threshold values.

The association of anomalous concentrations of several metals, including those associated with the Cuttingsville and Mount Ascutney plutons, within the east-west zone of multielement anomalies is consistent with observations throughout New England that rocks of the Mesozoic White Mountain Plutonic-Volcanic Suite are temporally and probably genetically related to base-, precious-, and rare-metal occurrences (Cox, 1970; Cox, this volume; Robinson, this volume). The Cuttingsville pluton is located at the intersection of trends E–W and NW–1, and the Mount Ascutney pluton is located at the intersection of trend E–W and a north-south terrane-boundary trend. By analogy, the Mesozoic niobium-bearing alkalic-carbonatite complexes of the Monteregian line of intrusive bodies at Oka and St. Andre, Quebec (fig. 24), are also localized along an east-west graben related to Mesozoic rifting (Philpotts, 1969; Kumarapeli, 1976). All of these observations suggest that the east-west zone of geochemical anomalies in the Glens Falls quadrangle constitutes an important metallogenic signature and defines a locus of extensional tectonism, alkalic igneous activity, and hydrothermal mineralization.

Table 14. Relative Element Magnitude (REM) values for crosscutting regional trend E–W (349 samples)

[These are relative figures used to indicate the position of chemical elements in terms of prominence. Numbers are rounded and do not represent precision greater than is present in the initial data]

Element	Anomalous samples	Anomaly mean	Intensity[1]	Area, percent anomalous	EM[2]	REM, cumulative percent
Pb	210	373	12.4	60.2	748.2	14.8
Sn	148	210	14.0	42.4	592.5	11.7
Ba	249	775	7.7	71.3	552.7	10.9
Cu	248	109	7.3	71.1	515.7	10.2
Mn ...	276	853	4.3	79.1	337.1	6.7
B	234	684	4.6	67.0	305.8	6.0
La	277	187	3.7	79.4	297.2	5.9
Mg ...	239	1.9	3.8	68.5	259.9	5.1
Sr.....	281	505	2.5	80.5	203.1	4.0
Ca	307	11.3	2.3	88.0	198.7	3.9
Co	189	23	3.3	54.1	178.3	3.5
W	48	850	12.1	13.7	167.1	3.3
Fe	261	2.0	2.0	74.8	149.9	3.0
Y.....	264	358	1.8	75.6	135.2	2.7
V.....	233	246	1.6	66.8	109.4	2.2
Nb....	186	124	1.8	53.3	94.5	1.9
Cr	200	148	1.5	57.3	84.7	1.7
Be	34	5.4	3.6	9.7	35.1	.7
Bi	3	577	38.4	.9	33.0	.7
Ni	12	45	6.4	3.4	22.1	.4
As	4	3,175	10.6	1.1	12.1	.2
Ag....	4	11.7	7.8	1.1	9.0	.2
Sb	5	680	4.5	1.4	6.5	.1
Au....	3	83	5.6	.9	4.8	.1
Zn	4	688	2.3	1.1	2.6	.1
Mo ...	2	43	4.2	.6	2.4	0
Total........................					5,057.6	

[1] Intensity=Mean/Threshold.

[2] EM=Intensity × Area.

Terrane Restricted

The terrane-restricted anomaly trends are probably related to stratabound syngenetic and (or) diagenetic mineral occurrences in some of the lithostratigraphic sequences, accompanied by epigenetic quartz-metal veins resulting from metamorphic remobilization and redistribution. The shape and direction of the isopleth patterns within these terranes are coextensive with the distribution of lithostratigraphic units, arches, domes, and northerly trending faults.

Patterns of iron and boron isopleths (figs. 6 and 11) are the basis for the boundary of terrane-restricted trend N–1 in Terrane 1. The northerly trends of these two element patterns cut many different rock types within the high-grade metamorphic terrane of the Adirondacks and lack any evidence of control by lithology or metamorphic grade. These elements therefore appear to have had a postkinematic (Mesozoic) introduction. A close association of the anomalies with the distribution of postkinematic normal faults suggests that the fault zones are possible sources of at least some of the pyrite and tourmaline with which the anomalous elements are associated.

Lanthanum ranks high in REM computations (table 15) for terrane-restricted trend N–1 (fig. 23) followed by barium, boron, and manganese. With respect to areal distributions, the following elements rank in decreasing order: Ca, La, Mn, Cr, and Y. The broad distribution of these elements in high concentrations suggests a lithostratigraphic control for the elements, probably reflecting bulk composition and metamorphic grade of the host rocks. Tin, beryllium, and tungsten on the other hand are most likely derived from Mesozoic igneous sources or fault zones. REM computations indicate that the areal distribution of tin, beryllium, and tungsten is small, as shown also by the isopleth maps (figs. 8 and 9). Tin and tungsten occur within trend N–1 only on the western edge of Terrane 1, where anomaly-to-background contrast is moderately high.

Copper, cobalt, tin, barium, and strontium (figs. 5, 7, 9, 12, and 14) form distinctly north-trending isopleth patterns in Terrane 3 that owe their configuration largely to the regional outcrop patterns of source beds. Source units in the Paleozoic sequence are considered to be the Hatch Hill

Table 15. Relative Element Magnitude (REM) values for terrane-restricted trend N–1 (58 samples)

[These are relative figures used to indicate the position of chemical elements in terms of prominence. Numbers are rounded and do not represent precision greater than is present in the initial data]

Element	Anomalous samples	Anomaly mean	Intensity[1]	Area, percent anomalous	EM[2]	REM, cumulative percent
La....	48	394	3.9	82.8	325.9	15.5
Ba....	35	200	4.0	60.3	241.7	11.5
B.....	37	364	3.6	63.8	231.9	11.0
Mn ...	48	375	1.9	82.8	155.2	7.4
Nb....	38	56	1.9	65.5	123.3	5.9
Cr....	42	85	1.7	72.4	122.4	5.8
Pb....	35	30	2.0	60.3	119.5	5.7
V.....	39	176	1.8	67.2	118.1	5.6
Ca....	52	6.2	1.2	89.7	111.4	5.3
Cu....	33	19	1.9	56.9	109.5	5.2
Y.....	42	298	1.5	72.4	107.8	5.1
Mg ...	35	2.9	1.4	60.3	87.1	4.1
Sr.....	31	219	1.5	53.4	78.2	3.7
Co....	11	26	3.8	19.0	71.4	3.4
Sn....	10	44	2.9	17.2	50.6	2.4
Be....	7	2.7	1.8	12.1	21.8	1.0
Ni....	4	48	2.4	6.9	16.4	.8
W....	3	117	1.7	5.2	8.6	.4
Fe....	3	5.0	1.0	5.2	5.2	.2
Total........................					2,105.8	

[1] Intensity=Mean/Threshold.

[2] EM=Intensity × Area.

Table 16. Relative Element Magnitude (REM) values for terrane-restricted trend N–2 (46 samples)

[These are relative figures used to indicate the position of chemical elements in terms of prominence. Numbers are rounded and do not represent precision greater than is present in the initial data]

Element	Anomalous samples	Anomaly mean	Intensity[1]	Area, percent anomalous	EM[2]	REM, cumulative percent
Sn....	32	483	24.1	69.6	1,679.3	31.9
Cu....	31	309	15.4	67.4	1,040.2	19.8
Ba....	40	7,251	10.4	87.0	900.7	17.1
Pb....	23	741	7.4	50.0	370.6	7.0
Mg ...	39	.9	1.8	84.8	156.5	3.0
Fe....	27	2.2	2.2	58.7	127.2	2.4
Mn ...	31	265	1.8	67.4	118.8	2.3
V.....	38	96	1.4	82.6	113.0	2.1
Nb....	38	66	1.3	82.6	108.3	2.1
La....	31	150	1.5	67.4	101.1	1.9
Ca....	36	5.9	1.2	78.3	93.0	1.8
Cr....	27	150	1.5	58.7	88.0	1.7
Y.....	30	383	1.3	65.2	83.3	1.6
B.....	28	264	1.3	60.9	80.4	1.5
Ni....	4	45	6.4	8.7	55.9	1.1
Sr.....	21	500	1.0	45.6	45.6	.9
Co....	5	28	4.0	10.9	43.5	.8
Au....	1	200	13.3	2.2	29.0	.6
Ag....	1	7.0	4.7	2.2	10.1	.2
Sb....	1	700	4.7	2.2	10.1	.2
As....	1	700	2.3	2.2	5.1	.1
Be....	1	2.0	1.3	2.2	2.9	.1
Total					5,263.0	

[1] Intensity=Mean/Threshold.

[2] EM=Intensity × Area.

Formation and the Browns Pond Formation(?) of Rowley and others (1979) and Rowley and Kidd (1981). As suggested by W.S.F. Kidd (oral commun., 1988), carbonate breccias of the Browns Pond could be metallogenically important because they may be slump breccias related to synsedimentary faulting that provided conduits for hydrothermal fluids (see Zen, 1961; Rowley and others 1979; Rowley and Kidd, 1981, for descriptions of the rock units). Recent analyses of rock-chip samples from the Browns Pond Formation(?) contain as much as 2.5 percent zinc. In addition, cherty black slate units of the Middle Ordovician Mount Merino Formation, higher up in the Taconic sequence, may also host massive base-metal sulfide deposits (Slack and Watts, 1989; Slack, this volume). The sources for many of the metal anomalies in the Taconic allochthon are likely to be confined to specific lithostratigraphic units in which the metals were deposited before major thrust faulting. This conclusion is based on the isopleth patterns that parallel the strike of the beds and abut against key thrust faults, thus restricting the anomalies to specific thrust slices.

Anomalies in the Taconic allochthon were separated for individual analysis into two REM windows corresponding to the Giddings Brook thrust slice (fig. 23, N–2) and the Dorset Mountain thrust slice (fig. 23, N–3). A third window area, in the northern Taconics (north of about lat 43°30′), includes the northern part of the Giddings Brook thrust slice and some of the Bird Mountain slice (see Zen, 1967, for geologic relations).

REM computations based on data from the Giddings Brook slice (trend N–2; table 16) listed in order of decreasing element magnitude are Sn, Cu, Ba, Pb, Mg, Fe, and Mn. In contrast, the EM sequence for the Dorset Mountain slice (N–3; table 17) is, in decreasing order, Sn, Pb, Cu, Fe, Co, La, and Ca. The comparison of REM data from the two thrust slices shows that tin is the dominant anomalous element in panned concentrate samples from both slices but that the EM values are nearly twice as high for tin and over twice as high for copper in the Giddings Brook (N–2) slice. Significantly, barium EM values are over a factor of seven higher in the Giddings Brook slice (N–2) than in the Dorset Mountain slice (N–3). These anomalies in the Giddings Brook slice are not only more intense but also more widely distributed than are those for the same elements in the Dorset Mountain slice. In contrast, iron and cobalt, both elements predominantly associated with pyrite, are relatively enriched in the Dorset Mountain thrust slice (N–3). Areally, but also in terms of intensity, calcium has higher EM values in the Dorset Mountain thrust slice than in the Giddings Brook slice. This difference supports interpretations of the map distributions of strontium and calcium anomalies and their relation to carbonate minerals and

Table 17. Relative Element Magnitude (REM) values for terrane-restricted trend N–3 (56 samples)

[These are relative figures used to indicate the position of chemical elements in terms of prominence. Numbers are rounded and do not represent precision greater than is present in the initial data]

Element	Anomalous samples	Anomaly mean	Intensity[1]	Area, percent anomalous	EM[2]	REM, cumulative percent
Sn	34	294	14.7	60.7	891.7	24.8
Pb	36	1,089	10.9	64.3	700.0	19.4
Cu	22	218	10.9	39.3	428.6	11.9
Fe	39	3.0	3.0	69.6	207.1	5.8
Co	25	32	4.6	44.6	204.1	5.7
La	29	321	3.2	51.8	166.1	4.6
Ca	52	8.1	1.6	92.9	150.7	4.2
Ni	10	55	7.9	17.9	140.3	3.9
Cr	37	201	2.0	66.1	133.0	3.7
Ba	33	1,482	2.1	58.9	124.7	3.5
Sr	38	597	1.2	67.9	81.1	2.3
B	26	269	1.3	46.4	62.5	1.7
As	3	3,067	10.2	5.4	54.8	1.5
Y	20	350	1.2	35.7	41.7	1.2
Mn	21	155	1.0	37.5	38.7	1.1
Mg	12	.8	1.7	21.4	36.4	1.0
Be	12	2.0	1.3	21.4	28.6	.8
Sb	3	767	5.1	5.4	27.4	.8
Nb	3	53	1.1	23.2	24.6	.7
Ag	2	10	6.7	3.6	23.8	.7
V	10	73	1.0	17.9	18.6	.5
Bi	2	40	2.7	3.6	9.5	.3
Au	1	30	2.0	1.8	3.6	.1
Mo	1	15	1.5	1.8	2.7	.1
Total					3,600.3	

[1]Intensity=Mean/Threshold.
[2]EM=Intensity × Area.

Table 18. Relative Element Magnitude (REM) values for terrane-restricted trend N–4 (259 samples)

[These are relative figures used to indicate the position of chemical elements in terms of prominence. Numbers are rounded and do not represent precision greater than is present in the initial data]

Element	Anomalous samples	Anomaly mean	Intensity[1]	Area, percent anomalous	EM[2]	REM, cumulative percent
Pb	149	336	11.2	57.5	645.2	15.8
Cu	174	124	8.3	67.2	555.1	13.6
Ba	130	311	10.4	50.2	521.0	12.7
Ag	9	215	143.3	3.5	497.8	12.2
Co	126	38	5.4	48.6	262.5	6.4
Mn	120	1,300	4.3	46.3	200.8	4.9
B	103	698	4.6	39.8	183.9	4.5
Y	189	247	2.5	73.0	180.5	4.4
Sn	74	86	5.8	28.6	164.5	4.0
La	138	135	2.7	53.3	143.9	3.5
Ca	155	16.7	2.4	59.9	143.2	3.5
Fe	134	2.7	2.7	51.7	140.0	3.4
Nb	127	114	1.6	49.0	80.1	2.0
Cr	106	259	1.7	40.9	70.8	1.7
V	111	220	1.5	42.9	62.9	1.5
Sr	106	668	1.3	40.9	54.7	1.3
Au	7	296	19.7	2.7	53.3	1.3
Bi	7	193	12.9	2.7	34.7	.9
Ni	9	70	10.0	3.5	34.7	.9
As	10	1,810	6.0	3.9	23.2	.6
Mg	29	2.3	1.2	11.2	12.9	.3
W	12	153	2.2	4.6	10.1	.2
Be	6	3.3	2.2	2.3	5.1	.1
Zn	3	1,286	4.3	1.2	5.0	.1
Sb	1	200	1.3	.4	.5	0
Total					4,086.6	

[1]Intensity=Mean/Threshold.
[2]EM=Intensity × Area.

calcareous lithologies. The relatively greater abundance of barium (related to barite and barium-carbonate minerals) in the Giddings Brook slice may be critical to the assessment of possible stratabound Pb-Zn-Ag-Ba deposits in rock units of the low Taconic sequence. The barium may be an indicator of lateral and vertical zonations with respect to stratabound sulfide deposits.

Terrane-restricted trend N–4 (fig. 23) comprises a north-striking zone in eastern Vermont. It includes the zone on both sides of the northern extension of the Whitcomb Summit thrust fault (Stanley and Ratcliffe, 1985), the "talc belt" of carbonatized (steatized) and serpentinized ultramafic rocks, and the lithostratigraphic sequence of potentially mineralized host rocks such as the Moretown Formation, the Barnard Volcanic Member of the Missisquoi Formation, the Pinney Hollow Formation (which is known to host gold-quartz veins (Slack and Schruben, this volume)), and the Stowe Formation. The boundaries of this terrane-restricted trend were drawn on the basis of the combined distributions of anomalous Pb, Cu, Fe, Co, B, Ca, Sr, and Mn (figs. 4–7, 11, 13–15). The REM computation (table 18) shows a ranking of Pb>Cu>Ba>Ag>Co>Mn>B>Y. These metals are probably derived from two types of sources: (1) metamorphosed chemical sediments and (or) stratabound sulfide occurrences and (2) epigenetic vein deposits. The presence of yttrium in the REM sequence may indicate the influence of carbonate veins on the geochemical signature, as a high yttrium content is typical of calcite in vein deposits (Rankama and Sahama, 1950, p. 524–525). Gold and arsenic are also relatively abundant within this zone and with a greater sampling density and analytical sensitivity for these metals, gold and arsenic may prove to be more widespread and abundant than the available data suggest.

The northeast-trending pattern of tungsten isopleths in Terrane 6 is the basis for defining the boundaries of terrane-restricted trend N–5 (fig. 23, table 19). This is the only terrane-restricted trend that is oriented slightly more northeasterly than northerly. If this trend were studied in detail, it could prove to be crosscutting rather than terrane-restricted, as the trend is parallel to the strike of discordant quartz-base-metal veins in the region. In part, the trend follows the outcrop distribution of the Ammonoosuc Volcanics and the calcareous Fitch Formation in western New Hampshire (see also Canney and others, 1987). This pos-

Table 19. Relative Element Magnitude (REM) values for terrane-restricted trend N–5 (132 samples)

[These are relative figures used to indicate the position of chemical elements in terms of prominence. Numbers are rounded and do not represent precision greater than is present in the initial data]

Element	Anomalous samples	Anomaly mean	Intensity[1]	Area, percent anomalous	EM[2]	REM, cumulative percent
Pb....	89	755	37.7	67.4	2,545.1	40.1
Sn....	59	215	14.3	44.7	640.8	10.1
Cu....	105	135	6.7	79.5	535.2	8.4
W....	74	725	7.2	56.1	406.4	6.4
Sr.....	111	585	2.9	84.1	245.8	3.9
B.....	103	265	2.6	78.0	206.8	3.3
Be....	35	11	7.4	26.5	195.4	3.1
Ba....	105	355	2.4	79.5	188.1	3.0
Mg...	89	.5	2.7	67.4	179.9	2.8
La....	86	115	2.3	65.1	149.2	2.4
V.....	110	269	1.8	83.3	149.2	2.4
Fe....	99	1.8	1.8	75.0	134.8	2.1
Nb....	99	172	1.7	75.0	129.2	2.0
Co....	55	18	2.6	41.7	107.7	1.7
Ca....	98	7.0	1.4	74.2	103.9	1.6
Y.....	83	242	1.6	62.9	101.3	1.6
Cr....	98	202	1.3	74.2	100.0	1.6
Mn...	87	752	1.5	65.9	99.1	1.6
Bi....	8	170	11.3	6.1	68.7	1.1
Mo...	2	258	25.7	1.5	39.0	.6
Sb....	2	900	6.0	1.5	9.1	.1
Ni....	5	12	1.7	3.8	6.5	.1
As....	1	2,000	6.7	.8	5.0	.1
Ag....	1	5.0	3.3	.8	2.5	0
Total					6,349.0	

[1] Intensity=Mean/Threshold.
[2] EM=Intensity × Area.

sible association with metavolcanic and carbonate rocks suggests analogies with stratabound tungsten deposits at Felbertal (Mittersill), Austria, and elsewhere in the Tyrols (Höll and others, 1987; Thalhammer and others, 1989), as well as with occurrences in the southwestern United States (Moench and Erickson, 1980). The consistent strength of the trend, however, lends support to the interpreted correlation with posttectonic quartz veins (some with lead, zinc, silver, and copper) that, in central and eastern Vermont, trend N. 5° to 15° E. approximately parallel to bedding (field data in USGS Mineral Resource Data System files collected by Alan McBean, Vermont Geological Survey, 1985–86). The reported metal association from these veins is, in part, supported by the metals shown to be important in the REM computations for terrane-restricted trend N–5. Lead, tin, and copper are indicated by the REM computations to be significant metals within this trend. Of these elements, copper has the broadest areal distribution, and lead has the highest anomaly-to-background contrast (intensity). This is a typical geochemical signature in heavy-mineral concentrates from stream sediments for an anomaly related to polymetallic veins.

Table 20. Relative Element Magnitude (REM) values for terrane-boundary trend NB–1 (88 samples)

[These are relative figures used to indicate the position of chemical elements in terms of prominence. Numbers are rounded and do not represent precision greater than is present in the initial data]

Element	Anomalous samples	Anomaly mean	Intensity[1]	Area, percent anomalous	EM[2]	REM, cumulative percent
Pb....	72	305	10.2	81.8	831.8	21.2
B.....	79	1,117	7.4	89.8	668.6	17.1
La....	75	173	3.5	85.2	295.2	7.5
Mg...	68	1.7	3.3	77.3	258.4	6.6
Ba....	62	338	3.4	70.4	238.1	6.1
Mn...	58	681	3.4	65.9	224.4	5.7
Y.....	87	449	2.2	98.9	222.1	5.7
Sr.....	69	545	2.7	78.4	213.6	5.5
Ca....	77	12.1	2.4	87.5	211.8	5.4
Cu....	55	46	3.1	62.5	192.0	4.9
Sn....	38	67	4.4	43.2	191.7	4.9
Fe....	52	1.5	1.5	59.1	88.6	2.3
V.....	53	196	1.3	60.2	78.8	2.0
Co....	30	13	1.9	34.1	64.9	1.7
Nb....	32	115	1.6	36.4	59.7	1.5
Cr....	42	120	1.2	47.7	57.4	1.5
W....	4	138	2.0	4.5	8.9	.2
Mo...	1	70	7.0	1.1	7.9	.2
Be....	1	2.0	1.3	1.1	1.5	0
Total........................					3,915.7	

[1] Intensity=Mean/Threshold.
[2] EM=Intensity × Area.

Related to Terrane Boundaries

Terrane boundary trend NB–1 is delineated on the basis of boron isopleths (fig. 11). The relatively high EM values for magnesium (table 20) support the premise that the associated tourmaline, which is brown or dark brown under the microscope, is magnesium rich and probably of evaporative or hydrothermal origin. Lead has the overall highest REM values in the zone, mostly on the basis of anomaly-to-background contrast. The lead anomalies along the western edge of Terrane 4, with possible stratabound sources in the Cheshire Quartzite, account for the high REM for lead. Relatively high REM values for lanthanum and yttrium in the zone are probably derived from the abundant red, rounded zircon seen in the panned concentrate samples. High values for barium and manganese reflect carbonate minerals in the samples, as do the less enriched strontium and calcium, which are near the middle ranking of REM values.

Terrane-boundary trend NB–2 (fig. 23, table 21) is delineated on the basis of lead isopleths (fig. 4). Relatively high REM values for lead support the conclusion that any mineral deposits present in the area of this trend are likely to be dominated by lead. Calcium, cobalt, manganese, and copper also show high REM values for this area. The high REM values for calcium reflect the large area encompassed by the calcium anomalies rather than large amounts of calcium enrichment. Lead and cobalt REM values result, in

Table 21. Relative Element Magnitude (REM) values for terrane-boundary trend NB–2 (69 samples)

[These are relative figures used to indicate the position of chemical elements in terms of prominence. Numbers are rounded and do not represent precision greater than is present in the initial data]

Element	Anomalous samples	Anomaly mean	Intensity[1]	Area, percent anomalous	EM[2]	REM, cumulative percent
Pb	44	204	6.8	63.8	434.3	12.4
Ca	66	19.1	3.8	95.6	364.9	10.4
Co	47	37	5.3	68.1	359.2	10.2
Mn ...	57	647	3.2	82.6	267.4	7.6
Cu	44	58	3.8	63.8	246.9	7.0
La	48	173	3.5	69.6	240.6	6.9
B	33	689	4.6	47.8	219.8	6.3
Sr.....	66	450	2.2	95.6	215.2	6.1
Ba	33	424	4.2	47.8	202.9	5.8
Sn	34	62	4.1	49.3	201.9	5.8
Mg ...	42	1.6	3.2	60.9	192.5	5.5
Fe	42	2.5	2.5	60.9	152.2	4.3
Y.....	56	329	1.6	81.2	133.3	3.8
V.....	33	214	1.4	47.8	68.1	1.9
Nb....	26	90	1.3	37.7	48.6	1.4
Cr	26	125	1.2	37.7	47.1	1.3
Ni	3	57	8.1	4.3	35.2	1.0
As	3	1,667	5.6	4.3	24.1	.7
W	6	187	2.7	8.7	23.2	.7
Ag....	2	7.0	4.7	2.9	13.5	.4
Be	6	2.0	1.3	8.7	11.6	.3
Au....	1	30	2.0	1.4	2.9	.1
Bi	1	30	2.0	1.4	2.9	.1
Total........................					3,508.4	

[1]Intensity=Mean/Threshold.

[2]EM=Intensity × Area.

Table 22. Relative Element Magnitude (REM) values for terrane-boundary trend NB–3 (110 samples)

[These are relative figures used to indicate the position of chemical elements in terms of prominence. Numbers are rounded and do not represent precision greater than is present in the initial data]

Element	Anomalous samples	Anomaly mean	Intensity[1]	Area, percent anomalous	EM[2]	REM, cumulative percent
Cu	90	99	6.6	81.8	541.5	14.5
Pb	38	461	15.4	34.5	531.2	14.2
Sn	35	250	16.6	31.8	529.5	14.2
W	33	1,125	16.1	30.0	482.3	12.9
Nb....	92	200	2.8	83.6	238.6	6.4
Cr	99	222	2.2	90.0	200.0	5.4
B	65	425	2.8	59.1	167.3	4.5
V.....	77	339	2.3	70.0	158.2	4.2
Mn ...	62	552	2.8	56.4	155.4	4.2
Ba	51	322	3.2	46.4	149.1	4.0
Sr.....	54	543	2.7	49.1	133.2	3.6
Fe	52	2.2	2.2	47.3	103.2	2.8
Co....	44	11	1.6	40.0	64.9	1.7
Be	9	11	7.5	8.2	61.2	1.6
La	38	78	1.6	34.5	54.0	1.4
Mg ...	32	.8	1.6	29.1	45.3	1.2
Ca	48	5.1	1.0	43.6	44.7	1.2
Y.....	23	270	1.3	20.9	28.2	.8
Zn	1	5,000	16.7	.9	15.1	.4
Ni	11	10	1.4	10.0	14.3	.4
Au....	1	100	6.7	.9	6.1	.2
Ag....	2	4.0	2.7	1.8	4.8	.1
Bi	1	30	2.0	.9	1.8	0
Sb	1	300	2.0	.9	1.8	0
Total........................					3,731.8	

[1]Intensity=Mean/Threshold.

[2]EM=Intensity × Area.

contrast, from enrichments of these metals above background values. Carbonate-sulfide veins or lithostratigraphic sources are the likely contributors to trend NB–2. The isopleth patterns parallel the strike of the basal lithostratigraphic cover sequence along the eastern edge of the Green Mountain massif, which suggests that the cover sequence is the source of the anomalies and that the source is probably concordant.

Boundary-related trend NB–3 (fig. 23, table 22) outlines a north-south pattern of Pb, Cu, W, Sn, and Nb anomalies (figs. 4, 5, 8–10). These anomalies parallel the Connecticut River and follow the same general trend as the island arc rocks of the Bronson Hill anticlinorium (Rowley and Kidd, 1981; Leo, 1985) and the trend of the east Vermont sequence that hosts stratabound copper deposits in the Orange County copper district. A variety of element sources are suggested by the relation of the isopleths to the geology. These sources include (1) mineralized zones exposed by, or contemporaneous with, the arches and domes or (2) hydrothermal deposits along the Ammonoosuc fault; the regional geochemical data provide few definitive clues as to which of these is the most likely possibility. The prominence of REM copper values in this zone (table 22) may be significant to the interpretation. Distribution patterns of the copper anomalies cover a larger area than those of lead, tin, or tungsten, but copper is less enriched relative to background values than are the other metals. The high ranking of copper in the REM calculations may be an indication that the entire NB–3 zone is a broad area of above-normal copper enrichment. It is probably significant that the same lithostratigraphic sequence hosting the deposits in the Orange County copper district underlies at least the western side of the NB–3 zone. The doming and arching within this zone may have exposed copper-bearing horizons comparable to those in the Orange County copper district to the east.

If the broad regional isopleths of the various key elements geochemically define major hydrothermal conduits or other structural and stratigraphic features related to mineralization, it follows that the intersections of these trends may be particularly favorable areas in which to search for mineral deposits. In a number of cases where the geochemical trends converge or intersect, favorable geochemical characteristics are noted including an increase in (1) anomaly-to-background contrast, (2) the number of anomalous elements, and (3) the size of the anomalous area. In addition, there commonly are outcrop signs of mineral-

Table 23. Relative Element Magnitude (REM) values for trend intersection I–1 (45 samples)

[These are relative figures used to indicate the position of chemical elements in terms of prominence. Numbers are rounded and do not represent precision greater than is present in the initial data]

Element	Anomalous samples	Anomaly mean	Intensity[1]	Area, percent anomalous	EM[2]	REM, cumulative percent
Cu	32	295	14.7	71.1	1,048.9	24.5
Sn	28	268	13.4	62.2	833.3	19.5
Mn	39	456	3.0	86.7	263.7	6.2
Pb	22	507	5.1	48.9	247.8	5.8
Nb	45	113	2.3	100.0	225.8	5.3
Fe	34	2.6	2.6	75.6	196.7	4.6
V	44	139	2.0	97.8	194.3	4.5
Ba	15	3,526	5.0	33.3	167.9	3.9
Mg	35	1.0	2.1	77.8	160.0	3.7
Co	12	35	5.0	26.7	133.3	3.1
Cr	22	273	2.7	48.9	133.3	3.1
Ni	6	68	9.8	13.3	130.2	3.0
B	34	324	1.6	75.6	122.2	2.9
Y	37	432	1.4	82.2	118.5	2.8
La	19	205	2.0	42.2	86.7	2.0
Ca	31	5.6	1.1	68.9	77.8	1.8
Sr	18	733	1.5	40.0	58.7	1.4
Be	12	2.0	1.3	26.7	35.6	.8
Ag	2	5.0	3.3	4.4	14.8	.3
As	2	850	2.8	4.4	12.6	.3
Zn	2	525	1.7	4.4	7.8	.2
Sb	2	250	1.7	4.4	7.4	.2
Mo	1	30	3.0	2.2	6.7	.2
Total					4,283.8	

[1] Intensity=Mean/Threshold.

[2] EM=Intensity × Area.

Table 24. Relative Element Magnitude (REM) values for trend intersection I–2 (96 samples)

[These are relative figures used to indicate the position of chemical elements in terms of prominence. Numbers are rounded and do not represent precision greater than is present in the initial data]

Element	Anomalous samples	Anomaly mean	Intensity[1]	Area, percent anomalous	EM[2]	REM, cumulative percent
Cu	90	446	22.3	93.7	2,090.1	37.2
Mn	92	1,353	4.5	95.8	432.3	7.7
B	93	577	3.8	96.9	372.6	6.6
Nb	96	225	3.2	100.0	321.9	5.7
Pb	74	121	4.0	77.1	310.1	5.5
Mg	93	.6	2.9	96.9	277.1	4.9
Sn	81	64	3.2	84.4	270.1	4.8
Fe	93	2.1	2.1	96.9	204.2	3.6
V	93	305	2.0	96.9	197.2	3.5
Ba	68	77	2.6	70.8	181.9	3.2
Sr	91	956	1.9	94.8	181.2	3.2
Co	93	13	1.3	96.9	123.4	2.2
Cr	90	196	1.3	93.7	122.6	2.2
La	47	107	2.1	49.0	104.4	1.9
Ni	74	12	1.2	77.1	91.7	1.6
Y	57	145	1.4	59.4	85.9	1.5
Ca	45	11.3	1.6	46.9	75.6	1.3
Be	26	4.6	2.3	27.1	62.5	1.1
Bi	3	367	18.3	3.1	57.3	1.0
W	23	215	2.1	24.0	51.6	.9
Ag	1	10	5.0	1.0	5.2	.1
Zn	1	700	1.4	1.0	1.5	0
Total					5,621.0	

[1] Intensity=Mean/Threshold.

[2] EM=Intensity × Area.

ized rock at the intersections of the trends, although locally these are subtle. As an example, the Orange County copper district is located at the intersection of trends NW–2 and NB–3, and this intersection displays all of the above listed characteristics. The copper district is one of two areas of trend intersection examined by using the REM program (Van Trump and Alminas, 1978) to further focus on the geochemical characteristics of intersection areas. The other area is at the intersection of trends NW–1, N–2, and N–3 in the northern Taconic allochthon and is labeled I–1 (fig. 23).

Some interesting characteristics of intersection area I–1 are brought out by the REM computations (table 23). The high REM values for copper and tin and the broad distribution of the manganese anomalies, reflected in the high REM of manganese, are notable here. The areally broad manganese anomaly suggests analogies with Meggen, West Germany; Tynagh, Ireland; and Heath Steele, New Brunswick, where manganese forms peripheral halos both laterally and vertically to sediment-hosted, stratabound base-metal deposits (Whitehead, 1973; Russell, 1975; Gwosdz and Krebs, 1977). If sedex mineral deposits are present in the Glens Falls quadrangle, such geochemical guidelines as manganese halos can be very useful in the search for economic deposits.

Lead, niobium, iron, vanadium, and barium also rank relatively high in REM in area I–1. The significance of lead, iron, and barium is without doubt their implication of sulfide- and sulfate-rich sources in bedrock of the drainage basins. High niobium and vanadium REM values in this area suggest that titanium minerals in the panned concentrates are derived from rutile or anatase in the detrital fraction of slates from the zone.

The Orange County copper district and surrounding region constitute the area of REM window I–2 (fig. 23, table 24). The ores of the copper district, described elsewhere (McKinstry and Mikkola, 1954; Howard, 1969; Annis and others, 1983), are volcanogenic massive sulfide deposits of Besshi-type (see Slack and others, this volume). Rocks within the area of window I–2 consist of pelite and calc-pelite, minor quartzite and metagraywacke, and interbedded amphibolite of the Standing Pond Volcanics. Anomalies of several elements define a north-south belt within area I–2 that closely follows the trend of the Strafford-Willoughby arch. The anomaly patterns show the same relation to the arch, and there appears to be a general gradient of concentrations outward from the axis of the arch, suggesting a genetic relation between the arch and the anomalies. Large bodies of granitic rocks of the Devonian

New Hampshire Plutonic Suite are exposed along this arch to the north of the quadrangle (Doll and others, 1961), and a small exposure of similar postkinematic granite, the Brocklebank granite (informal name), crops out about 13 km northwest of the Elizabeth mine (see Howard, 1969), exactly on the axis of the Strafford-Willoughby arch. The emplacement of granite beneath the arch, therefore, is one possible explanation for the coincidence of metal anomalies with the arch.

Tungsten is an example of an element that may have been introduced as a result of granite emplacement beneath the arch. Tungsten isopleths outline a zone centered precisely on and surrounding the Brocklebank granite (fig. 8). The Strafford dome to the south, on the other hand, is locally intruded by small dikes of granite pegmatite (Slack and others, 1987) and, like the Pomfret dome, coincides with a gravity low (Bean, 1953), suggesting granitic bodies at depth (see also Lyons, 1955). These relations permit the speculation that the emplacement of plutons in this area was accompanied by the introduction of metals, or the mobilization of preexisting metals, by hydrothermal processes active along zones of structural dilatancy(?), created by the intersection of northwesterly and north-trending zones of weakness in the crust. Hydrothermal activity at depth below the arch may have resulted in the upward transport and redeposition of alteration products such as barium, base metals like lead, volatiles such as boron, and skarn±vein metals like tungsten and tin. Some enriched metals related initially to synsedimentary-exhalative deposition (for example, copper and zinc) also may be derived from the lithostratigraphic sequence within the arch by mobilization and reconcentration as a result of granite intrusion below the arch and (or) by prior regional metamorphism.

The REM computations for window area I–2 show that copper, manganese, and boron occur with the massive sulfide deposits of the Orange County copper district (see also Slack and others, this volume). Anomaly patterns for these elements correspond to the areas of known deposits and suggest areas where there may be additional deposits. As an example, manganese and boron are probably derived from chemical-exhalative metasedimentary rocks that may be expected in peripheral lateral or vertical zones surrounding synsedimentary sulfide horizons. The boron anomalies are clearly related to the occurrence of tourmaline, which is a common gangue mineral in the stratabound sulfide deposits of the district (Annis and others, 1983). The manganese anomalies may have their sources in beds of coticule (fine-grained spessartine quartz rock), on the basis of their presence in the immediate wall rocks of the massive sulfide bodies at the Elizabeth mine (Annis and others, 1983); coticules also occur locally within chemical metasedimentary rocks of the Standing Pond Volcanics (J.F. Slack, oral commun., 1989). Other manganese anomalies in the copper district may have sources in carbonate veins formed during metamorphic mobilization of manganese from manganese-rich stratigraphic units, perhaps as a result of granite emplacement beneath the Strafford-Willoughby arch.

INFLUENCE OF GLACIAL DISPERSION

Much of the Glens Falls 1° × 2° quadrangle is covered by glacial till and by glaciofluvial, glaciolacustrine, and eolian materials deposited during the Pleistocene. The effects of this transported debris on the regional geochemical patterns must be considered, particularly with respect to possible influences that this overburden may have on configurations of the geochemical isopleths and on the concentrations of metals in the geochemical samples. Samples were not collected in areas where fluvial and eolian debris are known to be widespread, and so these types of materials, although not totally removed from influence, do not affect the geochemical data as much as the areally more widespread glacial till. Two large areas in the Champlain valley shown as unsampled on the geochemical maps are data gaps because of an extensive glaciolacustrine cover (see Connally, 1970, 1973).

Glacial transport of clasts and pebbles, minerals, and chemical elements has been variously estimated to range anywhere from about 2 to 5 km for copper, lead, and zinc in the 180-micron fraction of till in New Brunswick (Szabo and others, 1975) to as much as 50 km for nickel, chromium, and magnetite in the 630-micron fraction in southeastern Quebec (Shilts, 1973). The principal that emerges from these and other regional geochemical studies in the northeastern United States and Canada is that distance of transport by glaciers of ore-related materials is a function of particle size, mineralogy, and mode of transport, whether supraglacial, englacial, or subglacial.

The order of magnitude for transport distance probably is illustrated by the findings of Szabo and others (1975), who report down-ice dispersion trains from the Mt. Pleasant, New Brunswick, tin-molybdenum-base-metal deposit, as follows: (1) 7 km down-ice for cobble-sized clasts; (2) 2 to 5 km for copper, lead, and zinc minerals in the fine silt-clay range (180-micron) fraction; (3) about 16 km for copper, lead, and zinc in the coarse (500- to 2,000-micron) sand fraction; and (4) 10 km for lead, zinc, arsenic, and sulfur in the heavy-mineral part of the silt-sand fraction. These distances are comparable to those in southeastern Vermont observed for heavy minerals in glacial till (K.C. Watts, Jr., unpub. data) and for corresponding granite clasts in till of the same area (Larsen, 1972, 1987).

The distribution of pebble- to cobble-sized clasts dispersed as indicator trains from lithologically distinctive granitoid plutons in Vermont and western New Hampshire was used by Larsen and his students (Larsen, 1972, 1987) to determine the regional direction of the latest glacial imprint and the relative distances of glacial transport. Clasts were counted down-glacier from the Knox Mountain and

Barre granitoid plutons (see Doll and others, 1961) 16 to 24 km north of the Glens Falls quadrangle and from the Brocklebank, Braintree, Cuttingsville, and Mount Ascutney plutons and from the Lebanon dome within the Glens Falls quadrangle (see Doll and others, 1961, for the locations of the small Brocklebank and Braintree plutons). From these (Larsen, 1972, 1987) and other studies, the direction of ice movement is shown to have been south to southeasterly. Maximum distances of glacial transport of pebble- to cobble-sized materials determined from contour maps constructed from clast counts (Larsen, 1987) are approximately as follows: (1) Knox Mountain pluton, 32 km; (2) Barre pluton, 16+ km; (3) Brocklebank pluton, 10 km; (4) Braintree pluton, undetermined; (5) Cuttingsville pluton, 2 to 3 km; (6) Mount Ascutney pluton, 15 km; and (7) Lebanon dome, 10 km.

The Knox Mountain pluton, which has an areal outcrop of approximately 840 km^2, is much larger than the other exposed plutons of the region, whose areal outcrop is generally less than 26 km^2. As may be expected, the clast dispersion train of the Knox Mountain is over twice that of the other plutons, amounting to a width of about 20 km and an area of approximately 620 km^2 (Larsen, 1987). Accordingly, it can be concluded that the size of the bedrock source has a definite influence on the areal size of the down-glacier dispersion train, with one possible exception. The Barre granitoid pluton, which has an outcrop area of less than 26 km^2, has a clast indicator train disproportionate to its size, being about one-half the length of the train of the Knox Mountain pluton (Larsen, 1987). A possible explanation for the abnormally large dispersion train from a relatively small pluton may lie in the location of the Barre pluton on the axis of the Strafford-Willoughby arch. Evidence suggests that the arch is underlain by intrusive granitoid bodies. Thus, the abnormal length of the dispersion train down-glacier from the Barre pluton may be the result of refurbishing of granite debris of similar appearance and composition from concealed plutons in the down-glacier direction. As an example, the small Brocklebank pluton is centered on the Strafford-Willoughby arch and has demonstrably contributed granite clasts down-glacier (Larsen, 1987). The train of granite debris is recognized as having this specific plutonic source only because the pluton is exposed and the debris can be traced back to its bedrock source.

Till geochemical studies in the area of the Mount Ascutney pluton (K.C. Watts, Jr., unpub. data) were undertaken to assist in interpretation of the stream sediment data for the entire Glens Falls region. The Mount Ascutney area was selected because (1) there exist pebble-count data collected by Larsen (1972, 1987) and (2) heavy-mineral concentrates there are geochemically anomalous. Nearly 100 samples of basal till, and the B-horizon soils developed on them, were collected at 1.6-km intervals chiefly down-ice (south-southeast) from the Mount Ascutney pluton. The distribution of several elements in fractions of the tills and soils supports the transport direction and distance of the indicator train determined from the pebble counts. The most definitive patterns that correlate with the dispersion train were provided by isopleth maps of the magnetite-ilmenite heavy-mineral fraction of the tills, which show a magnetite low and an identically shaped ilmenite high relative to tills derived from the surrounding regionally metamorphosed country rocks. Mechanical glacial dispersion from the Mount Ascutney pluton is south to southeast for a maximum distance of 15 km, which is nearly identical to the pattern shown by the pebble-count data. Thus, one may conclude from these observations and the results reported by Szabo and others (1975) that the magnitude of transport distances is similar for clasts and heavy minerals in till.

On the basis of these studies, the influence of glacial dispersion on the alluvial geochemical patterns is concluded to be only of local significance. The available data suggest that, overall, the south- to southeasterly trending isopleth patterns seen on a regional scale in parts of the Glens Falls quadrangle probably owe their direction more to underlying geologic control than to superimposed glacial modification. In general, the effect of glacial till on stream sediment geochemistry is more likely to be one of dilution, rather than enhancement, of metal values. Specific arguments that support these conclusions with respect to areas of the Glens Falls quadrangle are

(1) Isopleths of anomalous copper, tungsten, niobium, boron, and strontium follow the axis of the Strafford-Willoughby arch; this correspondence suggests a close genetic association between the arch and the metal anomalies. The metals are not likely to have been glacially transported from near the Barre or Knox Mountain plutons, because (a) glacial dispersion of the metals precisely along the axis of the arch would be highly coincidental, particularly as the arch lacks a topographic definition that may have influenced the direction of glacier flow; (b) the clast indicator train from the Knox Mountain pluton (Larsen, 1987) barely enters the Glens Falls quadrangle in its far northeastern corner and trends away from it in a southeasterly direction; (c) the clast indicator train from the Barre pluton extends only 3 km into the Glens Falls quadrangle (Larsen, 1987); (d) metal-rich heavy minerals in sediments derived from till are apparently dispersed throughout the region in patterns similar to those of clasts, so that any information on transport of clasts is probably applicable to heavy-mineral grains; and (e) scheelite, which is one of the minerals responsible for the anomalous geochemical patterns, is a brittle mineral (Krauskopf, 1970) that probably does not withstand long distances of glacial transport if the source of the tungsten is in distal skarns near the Barre or Knox Mountain plutons.

(2) The east-west trend of element anomalies at about lat 43°30′ and the north to northeasterly trend dominated by tungsten and tin in Terrane 6 are clearly discordant to the grain of regional ice movement. The

geochemical patterns that define these trends cannot be derived from glacial dispersion and must therefore reflect local bedrock geology.

SUMMARY AND CONCLUSIONS

The significance of many of the regional geochemical anomalies is still largely speculative, but these new data form a basis for more work and suggest avenues for further research. The geochemical survey has identified a number of anomalous zones within the Glens Falls quadrangle, whose relation to the regional geology suggests that individual anomalous areas, and mineral deposits within these areas, are not isolated features but are part of a large, interrelated network of geochemical anomalies that have tectonic and metallogenic implications. Moreover, indicator elements commonly ignored in exploration geochemistry, such as Fe, B, Ba, Sr, Ca, and Mn, are shown here to be important pathfinders in defining the extent of the anomalous zones. Within these broader anomalous zones are smaller areas where local sample data indicate concentrations of metals of economic and exploration interest, including Pb, Cu, Zn, As, Co, Au, Ag, W, Sn, Bi, and Sb.

The regional geochemical trends shown on the isopleth maps fall into three main categories: crosscutting, terrane restricted, and terrane boundary. Directions of the crosscutting trends are northwest, northeast, and east-west. The terrane-restricted trends are generally northerly, as are the orientations of the terrane-boundary trends. One trend, best defined by tungsten (fig. 23, N–5), is north to northeasterly and is largely confined to Terrane 6, although viewed over a larger area it may be considered crosscutting. This trend is most likely a reflection of postkinematic quartz-metal veins like those described at scattered localities, such as at Holts Ledge, N.H., on the northwestern side of the Mascoma dome (Meyers and Stewart, 1956; Slack and others, 1987). The large size of the area covered by the isopleths suggests that more veins of similar type are present here but as yet are undiscovered.

Northwest-trending crosscutting anomalies (fig. 23, NW–1 and NW–2) are interpreted as a reflection of hydrothermal mineral deposits associated with deep-seated (basement) zones of crustal weakness. Landsat studies indicate that this same northwest orientation is characteristic of continental-scale fracture systems throughout the Appalachian orogen (Pohn, this volume). These major structural zones are considered to be of ancient origin and to have undergone recurrent tectonic reactivation, including reactivation subsequent to the Acadian orogeny (Devonian). The reactivation may have been accompanied by hydrothermal activity, which the geochemical anomalies suggest was related to the deposition of metals. Geochemical maps also imply that the hydrothermal activity was accompanied by the depletion and the lateral or vertical redeposition of barium, resulting in barium lows in areas of positive anomaly highs of other elements and by nearby barium highs. Because barium behaves like potassium during hydrothermal alteration, zones of barium depletion may reflect hydrothermally leached areas, and those zones enriched in barium may be related to areas of potassic alteration.

The northeasterly trends of crosscutting, regional anomalies (fig. 23, NE–1 and NE–2) are probably reflections of vein and fault orientations and the trend of granitic contact zones, such as the contact between the Sunapee two-mica granite and the Bethlehem Gneiss in the southeastern corner of the Glens Falls quadrangle. Posttectonic, discordant base- and precious-metal veins of probable Mesozoic age are known to have a similar trend in much of west-central New England (Cox, 1970; Slack and others, 1987; Slack and Schruben, this volume).

The east-west trend of crosscutting anomalies (fig. 23, E–W) is best defined by strontium and barium of probable alkalic igneous affinity, accompanied by ore-forming and pathfinder elements such as copper, iron, niobium, and boron. Elements that define the trend (for example, niobium, boron, strontium, and barium) are probably constituents of common rock-accessory and rock-forming minerals from alkalic mafic dikes and irregular bodies and possibly also hydrothermal deposits localized along dike-host rock contacts or in fracture zones. The alignment of Mesozoic plutons at Cuttingsville and Mount Ascutney within the geochemically delineated east-west zone may reflect an intracratonic rift that localized the emplacement of the plutons in a manner similar to that for the alkalic Monteregian plutons in southeastern Quebec, Canada, which likewise are intruded into a rift-related feature, the east-west-trending Ottawa-Bonnechere graben (fig. 24).

Geochemical anomaly trends confined to terranes probably have stratabound sources in most cases. Particular localization of anomalous metals seems to have taken place within, and on the flanks of, arches and domes and within lithostratigraphic units in the eastern part of Terrane 5 and in Terranes 6 and 7. In Terrane 5, anomalies are associated spatially with the Strafford-Willoughby arch and the Strafford and Pomfret domes. Some of the anomalies, such as those for tungsten along the Strafford-Willoughby arch, are considered to have skarn and (or) vein sources. In contrast, the configuration of the copper isopleths with respect to the Strafford dome suggests stratabound mineralization. The portion of the isopleth pattern on the eastern flank of the Strafford dome is undoubtedly related to syngenetic and (or) diagenetic volcanogenic mineralization of the Orange County copper district (Slack and others, this volume); the nearly identical pattern on the western flank of the dome suggests a similar source for the copper anomalies located there. It may be significant that a structural cross section drawn by Howard (1969) shows the same lithostratigraphic

Table 25. Relative rank of geochemical trend or window for each key element on the basis of Relative Element Magnitude (REM)

Key element	Geochemical trend or areal window							
Pb	N–5	NW–2	NB–1	E–W	N–4	N–3	NW–1	NB–3
Cu	I–2	I–1	N–2	NW–2	N–4	NB–3	N–5	NW–1
Fe	N–3	I–2	I–1	NB–2	E–W	N–4	NW–2	N–5/NE–1
Co	NB–2	N–4	N–3	E–W	NW–1	NW–2	NE–2	I–1
W	NB–3	N–5	E–W	NW–2	NE–2	I–2	NW–1	NB–2
Sn	N–2	NE–1	N–3	I–1	NE–5	NW–2	E–W	NB–3
Nb	I–2	NB–3	I–1	NW–2	NE–1	N–5	N–1	N–2
B	NB–1	I–2	E–W	NW–1	NW–2	N–1	NB–2	N–5
Ba	N–2	E–W	N–4	NW–1	NE–2	N–1	NB–1	NB–2
Sr	NW–2	N–5	NB–2	NB–1	E–W	I–2	NW–1	NB–3
Ca	NB–2	NB–1	E–W	NW–1	NE–2	N–3	N–4	NE–1
Mn	I–2	NW–2	E–W	NW–1	NB–2	I–1	NB–1	N–4
REM rank	1	2	3	4	5	6	7	8

sequence on the western flank of the Strafford dome as the one that hosts the mineral deposits of the copper district on the eastern flank. Further exploration on the western flank of the Strafford dome for stratabound copper deposits therefore seems warranted.

Like the anomalies associated with the Strafford-Willoughby arch, the anomalies in Terranes 6 and 7 are spatially associated with the Oliverian domes and imply a genetic relation. However, as with the Strafford dome, it may be that the doming of the Oliverian structures in New Hampshire only facilitated the exposure and erosion of premetamorphic mineralization.

Of the geochemical anomalies associated with terrane boundaries, those along the boundaries of Terranes 2 and 4 and 4 and 5 appear to be related either to unconformities associated with basement and cover rocks or to synsedimentary mineralization in specific lithologic units of the early Paleozoic rocks. The anomalies parallel to the Connecticut River on the other hand, between Terranes 5 and 6, may have a more complicated origin associated with regional tectonism and volcanism. East and southeast of Mount Ascutney, a large anomalous area occurs at the intersection of north-south-trending anomalies parallel to the Connecticut River and the east-west crosscutting anomaly trend (fig. 23, E–W). The anomaly-to-background contrast here is among the highest in the Glens Falls quadrangle, and the anomalies cover a broad bedrock source. Complex aeromagnetic highs also are present in this area (D.L. Daniels, oral commun., 1986) and are interpreted to reflect amphibolite-rich zones in the Ammonoosuc Volcanics, which are known to locally host stratabound sulfide deposits (see Moench, this volume, and Slack, this volume). However, considered together with the high metal values, there is a strong suggestion that the area east and southeast of Mount Ascutney has been a focal point for intense igneous and hydrothermal activity possibly in both the Paleozoic and the Mesozoic, perhaps because of the location of this area at the intersection of regional tectonic trends. From an overall view of the Glens Falls quadrangle, intersections of geochemical trends are probably the best areas in which to conduct mineral-resource investigations, because these areas are characterized by increases in the number of anomalous metals, anomaly-to-background contrasts, and the number of anomalous samples.

The REM computations have allowed more quantitative study of the geochemical trends and anomalous window areas. Tables 10 to 24 show the relative ranking, on the basis of anomaly intensity and number of anomalous samples, of each element sought in the spectrographic analyses in which at least one reported value was above the median. Table 25 compares the geochemical trends and window areas only on the basis of the key elements shown on the isopleth maps (figs. 4–15). The following conclusions are derived from tables 10–24: (1) Lead EM values in the north- to northeasterly trending tungsten zone of Terrane 6 (fig. 23, N–5; EM=2,545) are nearly twice the next highest EM value for lead (EM=1,440), which is within the northwest-trending zone (NW–2) in the northeastern corner of the Glens Falls quadrangle. (2) Copper EM values for intersection area I–2, the Orange County copper district and surrounding areas (EM=2,090), are about twice the next highest value for copper in the northern Giddings Brook thrust slice (I–1) of the northern Taconic allochthon and of the southern Giddings Brook slice (N–2). Copper has about the same EM values for both these slices (1,049 and 1,040, respectively). (3) The Dorset Mountain slice of the high Taconic sequence has the greatest EM value for pyrite-related iron (N–3; EM=207), although this value is not substantially higher than that for the southern Giddings Brook slice (N–2; EM=197) or the Orange County copper district (NB–3; EM=204). (4) The eastern boundary of the Green Mountain massif (NB–2) has the highest EM values for cobalt (359), followed by the serpentine-talc belt (N–4; EM=262) in the upper plate of the northern extension of the Whitcomb Summit thrust fault and the Dorset Mountain thrust slice in area N–3 (EM=204). High cobalt EM values are probably an indication of cobaltiferous pyrite mineralization. (5) Tungsten EM values in the terrane-boundary trend parallel to the Connecticut River (NB–3) and in the north to northeast trend of Terrane 6 (N–5) are roughly threefold or more higher (EM=482 and 406, respectively) than in any other areas of the Glens Falls quadrangle. The tungsten is clearly related to the presence of scheelite and may be in stratabound beds or discordant quartz veins trending northeasterly. (6) Tin is nearly twice as high in EM value (1,679) in the southern Giddings Brook slice of the Taconic allochthon (N–2) than in the next highest area, the northeast-trending tin zone NE–1 in the southwestern corner of Terrane 7 (EM=956). The Dorset Mountain slice (N–3) and the northern Giddings Brook-Bird Mountain slices (I–1) of the Taconic allochthon have high EM values for tin (EM=892 and 833, respectively). Tin EM values for samples from the Giddings Brook slice are nearly twofold higher than the values for either of the other two thrust slices. (7) Niobium EM values (322) are highest in the window area that includes the Orange County copper district (I–2) and largely reflect high-contrast niobium anomalies coextensive with the Strafford-Willoughby arch. The three next highest EM values for niobium are from the terrane-boundary area along the Connecticut River (NB–3; EM=239), the Giddings Brook thrust slice (I–1; EM=226), and the northwesterly crosscutting trend in the northeastern corner of the Glens Falls quadrangle (NW–2; EM=222). High niobium EM values in the Giddings Brook slice may have the same bedrock sources as the tin in the same area and may also be mineralogically related because of the similar geochemical behavior of the two elements. (8) The highest EM values for boron are from the zone of northerly trending boron anomalies on the boundary of Terranes 2 and 4 (NB–1; EM=669). The second highest area of boron EM values is the window area that includes the Orange County copper district (I–2; EM=373). This boron is considered to be derived from tourmaline-rich lithologic units associated with the stratabound massive sulfide deposits of the district (see Annis and others, 1983). The third area of high EM values for boron is the east-west crosscutting zone (E–W; EM=306) that is considered to be the result of a geochemical overprint synchronous with Mesozoic tectonism. (9) By far the largest EM value for barium is in the southern Giddings Brook thrust slice of the Taconic allochthon (N–2; EM=901). This high EM value reflects both the high intensity and broad areal distribution of the barium anomalies. Barium is far more enriched in the Giddings Brook slice than in the other thrust slices of the Taconics (EM=168 and 125, respectively). The high barium is related to barite and barium-carbonate minerals. Celsian (barium feldspar) is suspected, though not confirmed, to be a contributor to the barium anomalies as well. The second highest EM value for barium is associated with the east-west trend of crosscutting Mesozoic(?) anomalies (E–W; EM=553). High EM values from this zone are more a reflection of the broad areal extent of barium than a reflection of high anomaly-to-background contrast. This low geochemical contrast suggests that the barium is in rock-forming or rock-accessory minerals, rather than in a discrete mineral of barium such as barite. The distribution of barium enrichment in minerals other than barite nevertheless may still be used indirectly in the search for metal deposits in which barium participated in hydrothermal reactions. The third zone (N–4) of high EM values for barium (EM=521) is associated with the talc belt and the related lithostratigraphic package on the eastern side of the Green Mountain massif. It is similar to the east-west zone of crosscutting anomalies with respect to REM values. These barium anomalies are not intense enough to be related to barite but may reflect slight enrichments of barium in carbonate minerals that are abundant in panned concentrate samples from this zone. (10) EM values for the anomalous calcium trend at the eastern boundary of Terranes 4 and 5 (NB–2) are substantially higher than those of the other areas (EM=365). Lead is anomalous in the same general area, suggesting that the anomalous calcium is derived from carbonate veins or carbonate-hosted replacement deposits near the basement-cover unconformity. (11) The EM values for strontium are within a fairly close range in the seven areas where they are high (174–279). The source of the strontium at these concentration levels may be carbonate veins or carbonate-rich sedimentary units. In the two areas having the highest EM values, strontium isopleths (fig. 14) closely conform with the isopleths of anomalous tungsten (fig. 8) and calcium (fig. 13). (12) Manganese may be a good indicator element for chemical metasedimentary rocks such as those in the Orange County copper district, on the basis of a high EM value (432) in intersection area I–2. Other areas of high EM values for manganese are in the northwest and east-west crosscutting anomaly trends (NW–1, NW–2, E–W) that have EM values of 275, 377, and 337, respectively. In these cases, manganese is concluded to be a member of the suite of elements that is enriched within the crosscutting regional anomaly zones. Possible syngenetic-diagenetic concentrations of manganese in thin beds of rhodochrosite in the Bird Mountain-Giddings Brook slices of the Taconic allochthon (I–1) are reported by Dale (1899, p. 190) and are reflected by the relatively high EM value for manganese (264) in the area. These manganese concentrations may be peripherally related to contemporaneous sediment-hosted base-metal sulfide deposits.

The Taconic allochthon of New York and Vermont and the associated sequence of deformed and thrust-faulted black, red, and green slates, graywacke, and locally interbedded carbonate, quartzite, and chert are highly anomalous in many elements. Geochemically and geologically the Taconics are comparable to the areas of large sediment-hosted Pb-Zn-Ag-Ba deposits in the Selwyn basin, Yukon, and at Rammelsberg and Meggen in West Germany. Anomalies in the Taconic allochthon include high values for Pb (to 10,000 ppm), Cu (to 7,000 ppm), Ba (to >10,000 ppm), and Sn (to 2,000 ppm), accompanied by scattered high values for Zn, Ag, Co, As, Au, Sb, Bi, Mo, and Be. Tin and barium are most prevalent in heavy-mineral concentrates of stream sediments from the low Taconic sequence (Giddings Brook and Bird Mountain thrust slices); cobalt and arsenic are in contrast predominantly anomalous in the high Taconics (Dorset Mountain thrust slice). Copper and lead are anomalous in samples from both the low and high Taconic sequences, although copper probably dominates in the Giddings Brook and Bird Mountain thrust slices, and lead is more prevalent in the Dorset Mountain slice. The other metals occur in anomalous amounts in scattered samples, more or less equally in both the high and low Taconic sequences.

In addition to the geochemical data and geologic setting, the identification of barite, celestite, and fluorapatite in the heavy-mineral samples is further support for the proposed existence of sediment-hosted mineral deposits in the Taconic allochthon. These geochemical and mineralogic indicators of possible mineralization in an area that has no history of previous metal mining, and only limited prospecting, suggest that the Taconic allochthon constitutes a new metallogenic province identified by this regional geochemical survey.

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Chapter G

Airborne and Ground Electromagnetic Studies of the Eastern Half of the Glens Falls 1° × 2° Quadrangle, New York, Vermont, and New Hampshire

By CARL L. LONG

U.S. GEOLOGICAL SURVEY BULLETIN 1887

SUMMARY RESULTS OF THE GLENS FALLS CUSMAP PROJECT, NEW YORK, VERMONT, AND NEW HAMPSHIRE

CONTENTS

FIGURES

Airborne and Ground Electromagnetic Studies of the Eastern Half of the Glens Falls 1° × 2° Quadrangle, New York, Vermont, and New Hampshire

By Carl L. Long[1]

Abstract

In the fall of 1983 and the spring of 1984, a very low frequency airborne electromagnetic survey was conducted in the eastern half of the Glens Falls 1° × 2° quadrangle, Vermont and New Hampshire. The primary objective of the supplemental airborne electromagnetic measurements was to contribute new structural and lithologic information that cannot be obtained from aeromagnetic measurements. Data from the airborne and ground surveys delineated many known and unknown conductive and resistive features associated with lithologies and structures that may be suggestive of permissive or favorable ground for the occurrence of mineral deposits.

INTRODUCTION

The survey (fig. 1) was flown at half-mile (0.8 km) spacing at an optimal 300 ft (91 m) above the terrain and used airborne very low frequency (VLF) in conjunction with a primary aeromagnetic survey. The VLF receiver used in this survey measured signals from a communications transmitter located in Annapolis, Md., at a frequency of 21.4 kHz. The VLF technique measures the local electric and magnetic characteristics of the secondary field induced by a vertically polarized ground wave transmitted by the VLF station. The amplitude of the electric and magnetic field components at a local receiver site are affected by several factors, such as transmitted power output, distance between transmitter and receiver site, propagation events in the ionosphere, local topography, and the resistivity of Earth between the transmitter and receiver. The U.S. Geological Survey's system measures inphase (R or real) and quadrature (Q or imaginary) parts of three orthogonal spatial components of the VLF magnetic field and the inphase part of the vertical electric field (EVR), as well as the inphase and quadrature parts of the horizontal electric field (EHR and EHQ, respectively). The measured local EHQ component can be normalized by the time-varying EVR. This normalization eliminates variations in the propagation path but leaves local perturbations of the ground wave caused by the topography and resistivity of Earth. Wavetilt is the ratio of EHQ to EVR. Over lithologically homogeneous terrane, the wavetilt is related to the apparent resistivity of the ground, neglecting displacement currents (Hoekstra and others, 1975) by Earth. A complete description of airborne VLF theory, principles, and techniques may be found in articles by Arcone (1978, 1979), Frischknecht (1971), and Hoekstra and others (1975). The VLF method can be very useful in mapping conductors in crystalline rock in areas where the overburden is thin or nonconductive and the terrain relief is not extreme. Inspection of the data and field checking indicated that the VLF survey of the eastern half of the Glens Falls 1° × 2° quadrangle delineated some resistive zones associated with intrusive rocks and many conductive zones thought to be associated with carbonaceous, graphitic, or sulfidic zones.

DISCUSSION OF VLF RESULTS

The VLF data are presented here as airborne and ground profiles (figs. 2–5, 8, and 9) and as a contoured apparent resistivity map (fig. 6). An aeromagnetic map (fig. 7) is provided for comparison with the resistivity map. The profiles of figures 2 and 3 show five channels of VLF and one channel of aeromagnetic data that have been plotted for a short section of flight lines 136 and 137. A major conductor is present between longitudes 72°35′ and 72°36′ and is marked by an asymmetric anomaly in the vertical magnetic field (HVR and HVQ); a more symmetric anomaly is seen in the horizontal or inline magnetic field (HIR and HIQ). The HIR anomaly peaks near the midpoint or crossover of the vertical field anomaly. The horizontal electric field (EHQ) is marked by a less distinctive anomaly, which indicates the conductive zone. The similarity of the shapes of the profiles on lines 136 and 137 indicates continuity of electrical characteristics along strike between the two lines. The conductor coincides approximately with a small magnetic anomaly (channel 11) whose position, relative to the conductor, changes from line to line, as shown in figures 2 and 3. The six airborne profiles of HVR

Manuscript approved for publication April 13, 1990.

[1] U.S. Geological Survey.

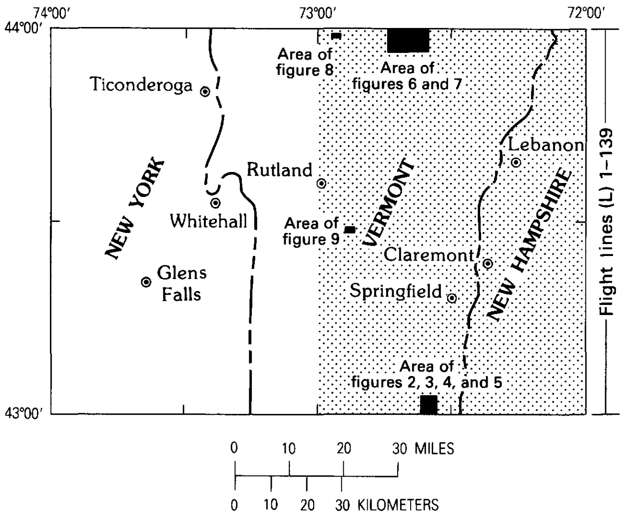

Figure 1. Index map showing area of low-level survey (stippled pattern) of the eastern half of the Glens Falls 1° × 2° quadrangle, New York, Vermont, and New Hampshire. The solid areas indicate areas of the profiles and maps shown on figures 2–9.

(fig. 4) illustrate the continuity of a north-northeast-striking conductor over a distance of at least 4 km. A ground VLF traverse (fig. 5) was made over a portion of line 136, extending eastward from the anomaly. The tilt angle of the ground measurements is roughly equivalent to the HVR on the airborne profile (fig. 2). The ground profile identifies many more minor features than the airborne profile, but overall it shows a large asymmetric anomaly that coincides with the airborne anomaly. Although a detailed geologic map is unavailable for this area, the State map of Vermont (Doll and others, 1961) shows that a carbonaceous mica schist (part of the Missisquoi Formation) occurs near the crossover point in both anomalies.

The apparent resistivity and aeromagnetic maps (figs. 6 and 7) cover an area over the Braintree plutonic complex (see Ern, 1963). The geology of this area consists of adjacent dioritic and granitic bodies emplaced in the northwestern portion of the Randolph 15-minute quadrangle. Evidence of the granitic body is readily seen on the resistivity map by the 4,000-ohm-m high at the upper

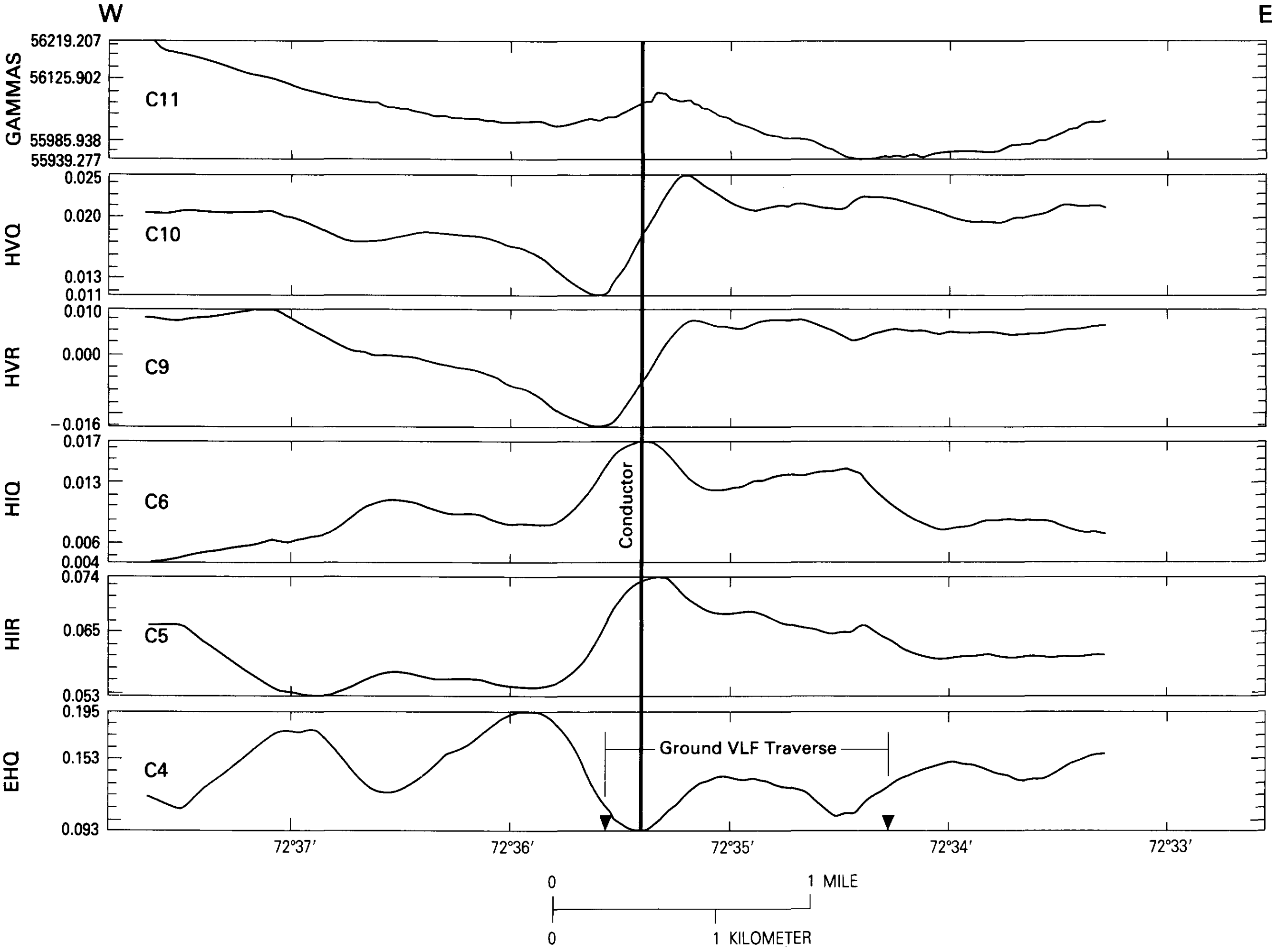

Figure 2. Airborne profiles for part of flight line 136 (L136) of vertical (HVR, HVQ) and horizontal (HIR, HIQ) very low frequency magnetic fields, horizontal very low frequency electric field (EHQ), and the intensity of Earth's magnetic field (gammas). C, channel.

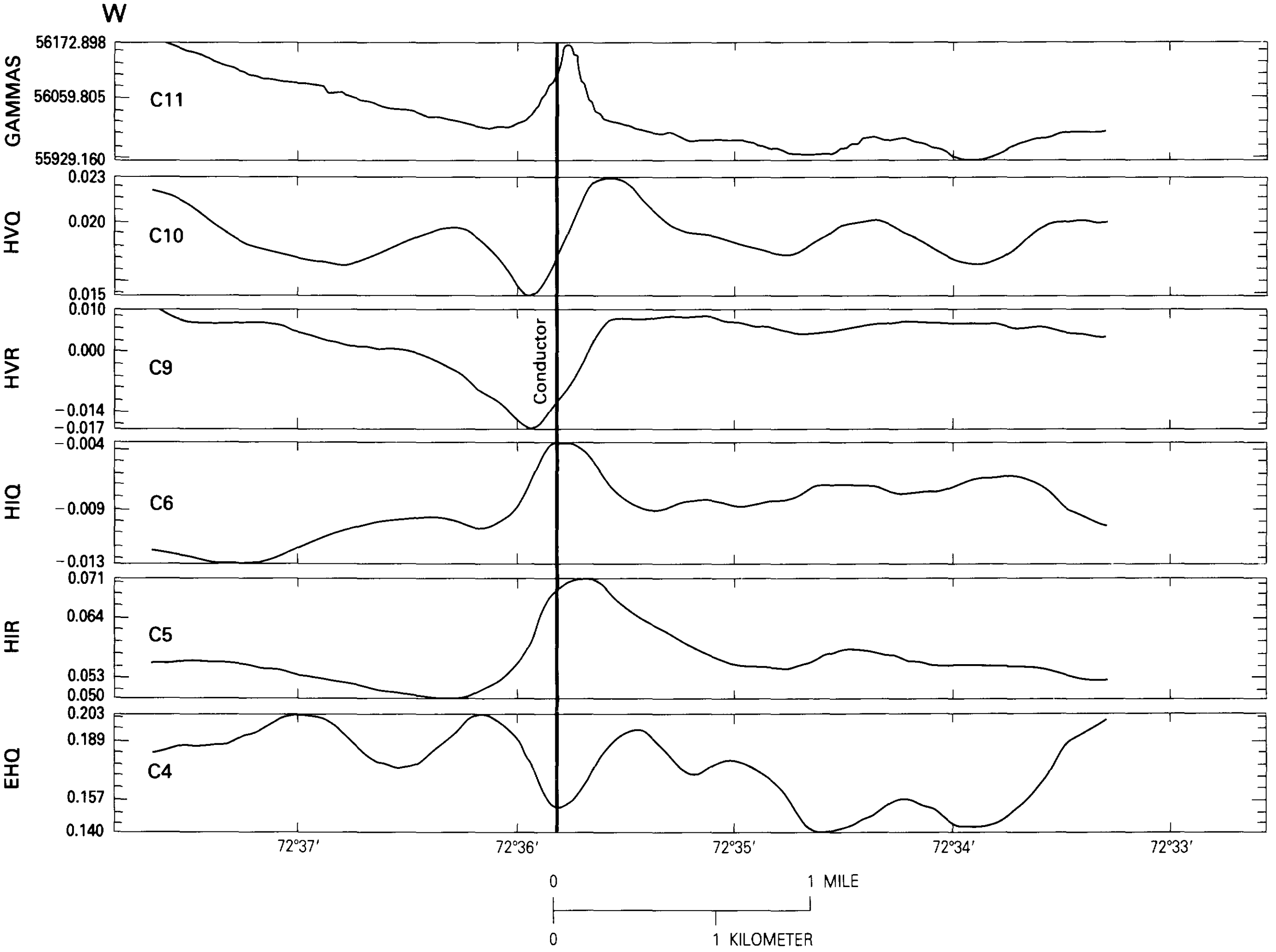

Figure 3. Airborne profiles for part of flight line 137 (L137) of vertical (HVR, HVQ) and horizontal (HIR, HIQ) very low frequency magnetic fields, horizontal very low frequency electric field (EHQ), and the intensity of Earth's magnetic field (gammas). C, channel.

northwestern side of the map (fig. 6) and by the −100-nT magnetic low on the aeromagnetic map (fig. 7). Steep gradients in the resistivity and magnetic contours on the eastern sides of the high and low anomalies may be indicative of a steep intrusive contact, possibly with the Northfield Formation. The plutonic body is surrounded by rock of low apparent resistivity and relatively high magnetic susceptibility forming similar U-shaped trends that possibly give some indications of the aerial extent of contact metamorphism. About 2.4 km west of Perth, Vt., a significant apparent resistivity low coincides with a 400-nT aeromagnetic high that probably indicates another buried intrusive body that has a different lithology than the Braintree pluton. Near the eastern side of the maps, a broad north-south-trending resistivity low and a coincident magnetic high indicate a significant change in the lithology of the Waits River Formation in that area.

The computed intensity of magnetic field (gammas), VLF apparent resistivity, and total VLF magnetic field (fig. 8) are shown for a short section of flight line 2 and include data from a ground profile run over a small portion of the same line. A conductive zone (apparent resistivity low) is seen between longitudes 72°55′ and 72°56′ and is marked by a broad symmetrical VLF apparent resistivity low. Decreasing resistivities and the crossover in the tilt angle and quadrature of the ground profile coincide with airborne anomalies of resistivity and total VLF magnetic field. A small broad magnetic anomaly is coincident with the conductive zone. The presence of a black graphitic quartz-muscovite schist (Monastery Formation) mapped in the area by Osberg (1952) may explain the conductive zone seen on the profiles.

A ground traverse was also made across the eastern portion of the Cuttingsville intrusive complex (fig. 9). A major conductive source is indicated by the 213-m-wide apparent resistivity low (<400 ohm-m) near the gabbro-syenite contact (Robinson, this volume) on the western end of the profile and corresponds to a 5,000-nT magnetic high.

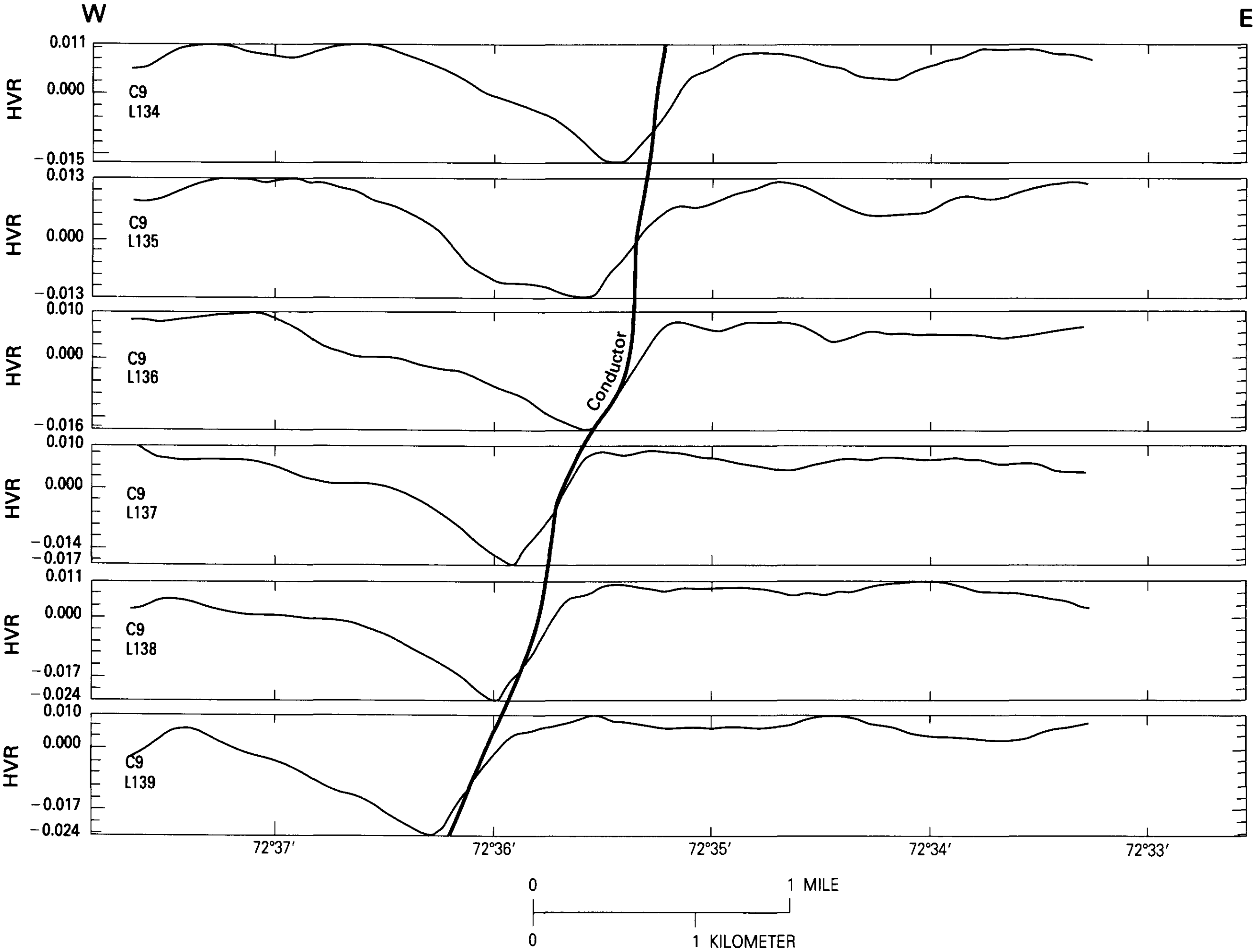

Figure 4. Airborne profiles of vertical (HVR) very low frequency magnetic fields for parts of flight lines 134–139 (L134–L139). C, channel.

Both of these anomalies coincide with the central portion of the aeromagnetic high that reflects the presence of the Cuttingsville intrusive complex (Daniels, this volume). The conductive zone (fig. 9), which reaches a minimum of <100 ohm-m, may be indicative of hydrothermally altered rock, a fracture zone, or a mineralized zone. The large magnetic high seems to reflect an unusual concentration of magnetic material in the syenite body coincidental with the conductive zone. At the eastern end of the traverse, but still inside the intrusion, is another small conductive zone (400 ohm-m and about 61 m across) that has a coincidental magnetic high (2,000 nT). Additional studies will be needed to rigorously correlate geophysical anomalies with mineralization in the Cuttingsville area.

CONCLUSIONS

The examples included in this report demonstrate how VLF electromagnetic surveys can be used in regional mapping. The VLF method adds another parameter to the existing geophysical data base by measuring the electrical response of the rocks. As can be seen from the examples presented, VLF can trace conductive units from flight line to flight line and can outline resistive bodies. Integrated with other geophysical, geochemical, and geologic studies, results such as these can add new structural and lithologic information to a geologic map and thus aid in the definition of prospective exploration targets.

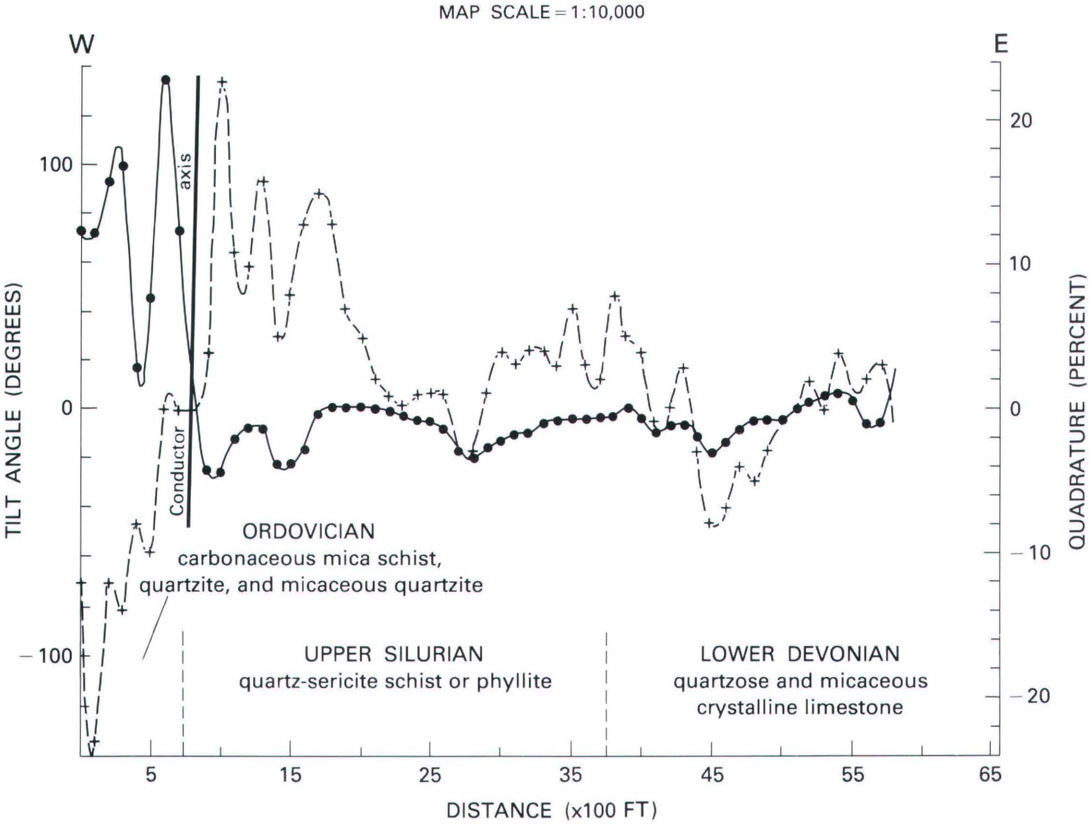

Figure 5. Ground very low frequency profiles of tilt angle (crosses) and quadrature (closed circles) for part of flight line 136 (L136).

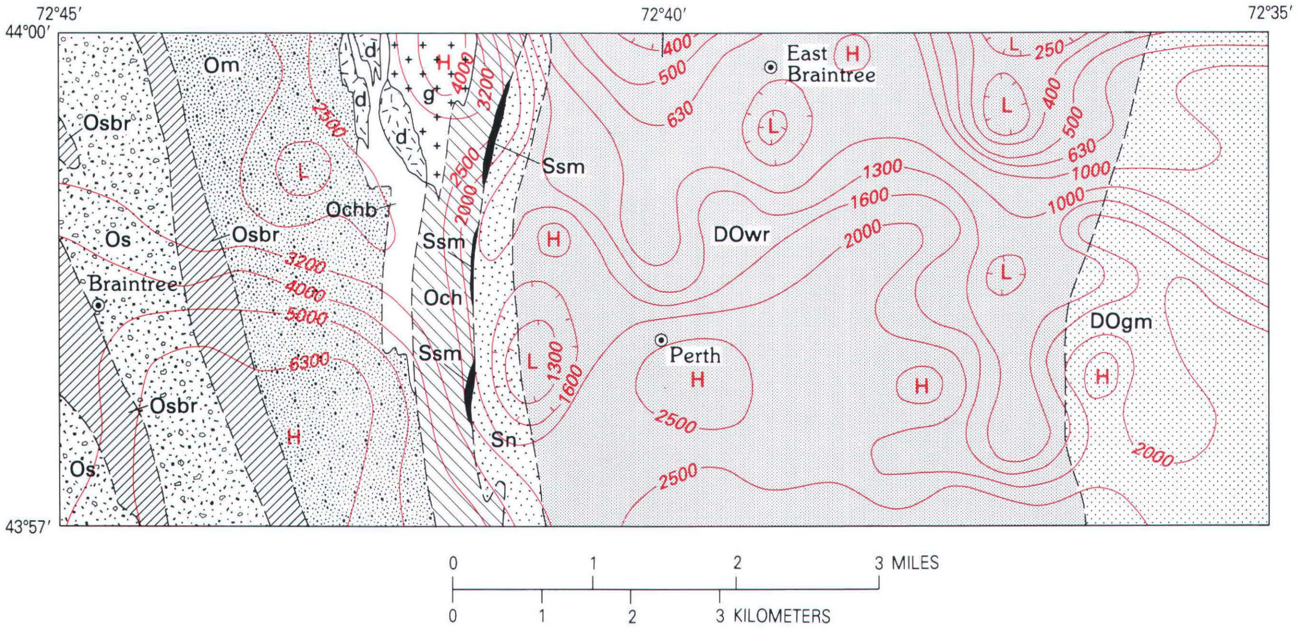

Figure 6. Very low frequency resistivity map of the northwestern corner of the Randolph 15-minute quadrangle. Resistivity contours are superimposed on the geology (after Ern, 1963). The contour interval is logarithmic and has 10 divisions per decade. d, diorite; g, granite; DOwr, Waits River Formation; DOgm, Gile Mountain Formation; Sn, Northfield Formation; Ssm, Shaw Mountain Formation; Och, Cram Hill Formation; Ochb, Harlow Bridge member (Cram Hill Formation); Om, Moretown Formation; Os, Stowe Formation; Osbr, Brackett member (Stowe Formation).

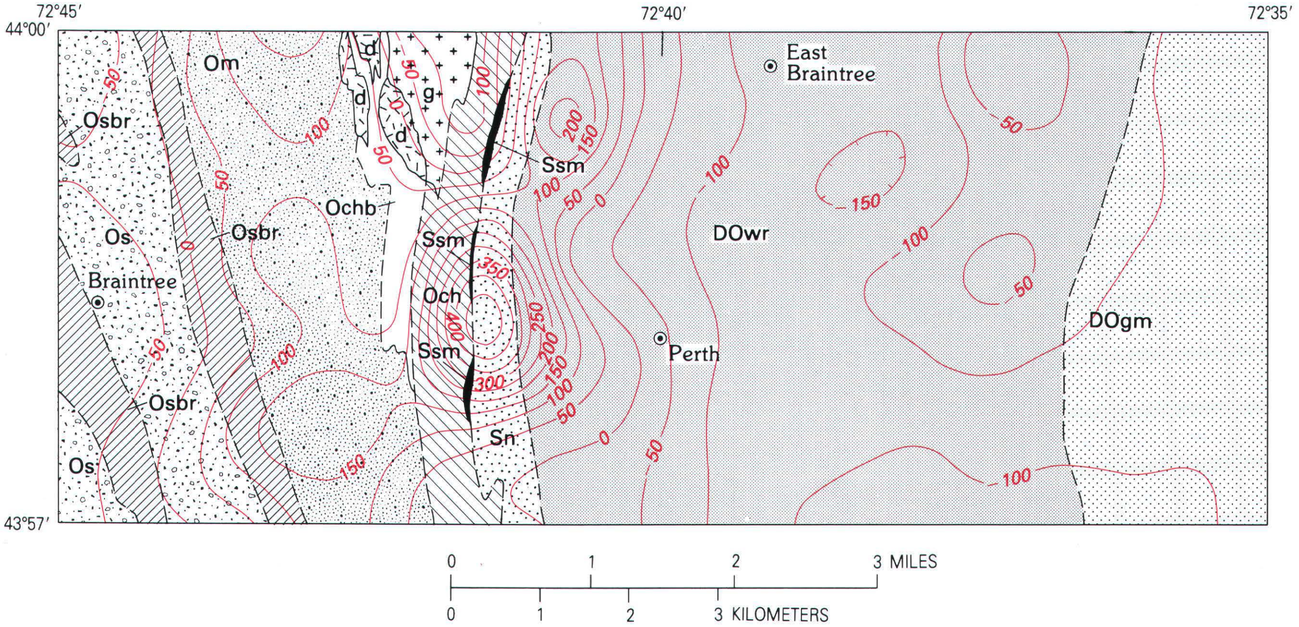

Figure 7. Aeromagnetic map of the northwestern corner of the Randolph 15-minute quadrangle. Magnetic contours are superimposed on the geology (after Ern, 1963). The contour interval is 50 nanoteslas (nT), but the 1984 International Geomagnetic Reference Field has been removed. d, diorite; g, granite; DOwr, Waits River Formation; DOgm, Gile Mountain Formation; Sn, Northfield Formation; Ssm, Shaw Mountain Formation; Och, Cram Hill Formation; Ochb, Harlow Bridge member (Cram Hill Formation); Om, Moretown Formation; Os, Stowe Formation; Osbr, Brackett member (Stowe Formation).

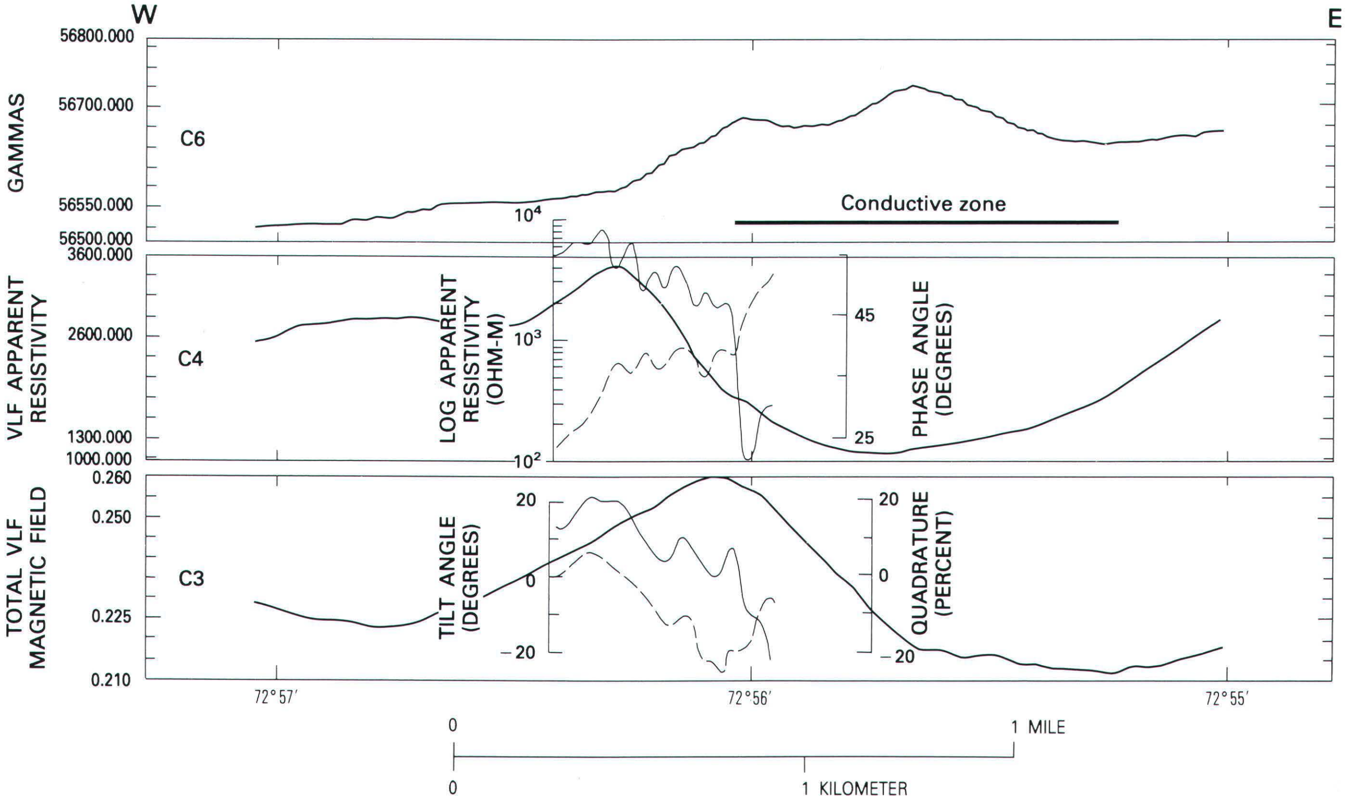

Figure 8. Airborne profiles of the intensity of Earth's magnetic field (gammas), very low frequency (VLF) apparent resistivity, and total very low frequency magnetic field for part of flight line 2 (L2). Superimposed on the total very low frequency magnetic field and the very low frequency apparent resistivity is a ground very low frequency traverse of roughly equivalent data. Apparent resistivity (solid line) and phase angle (dashed line) are shown on the very low frequency apparent resistivity profile. Tilt angle (solid line) and quadrature or ellipticity (dashed line) are shown on the total very low frequency magnetic field profile. C, channel.

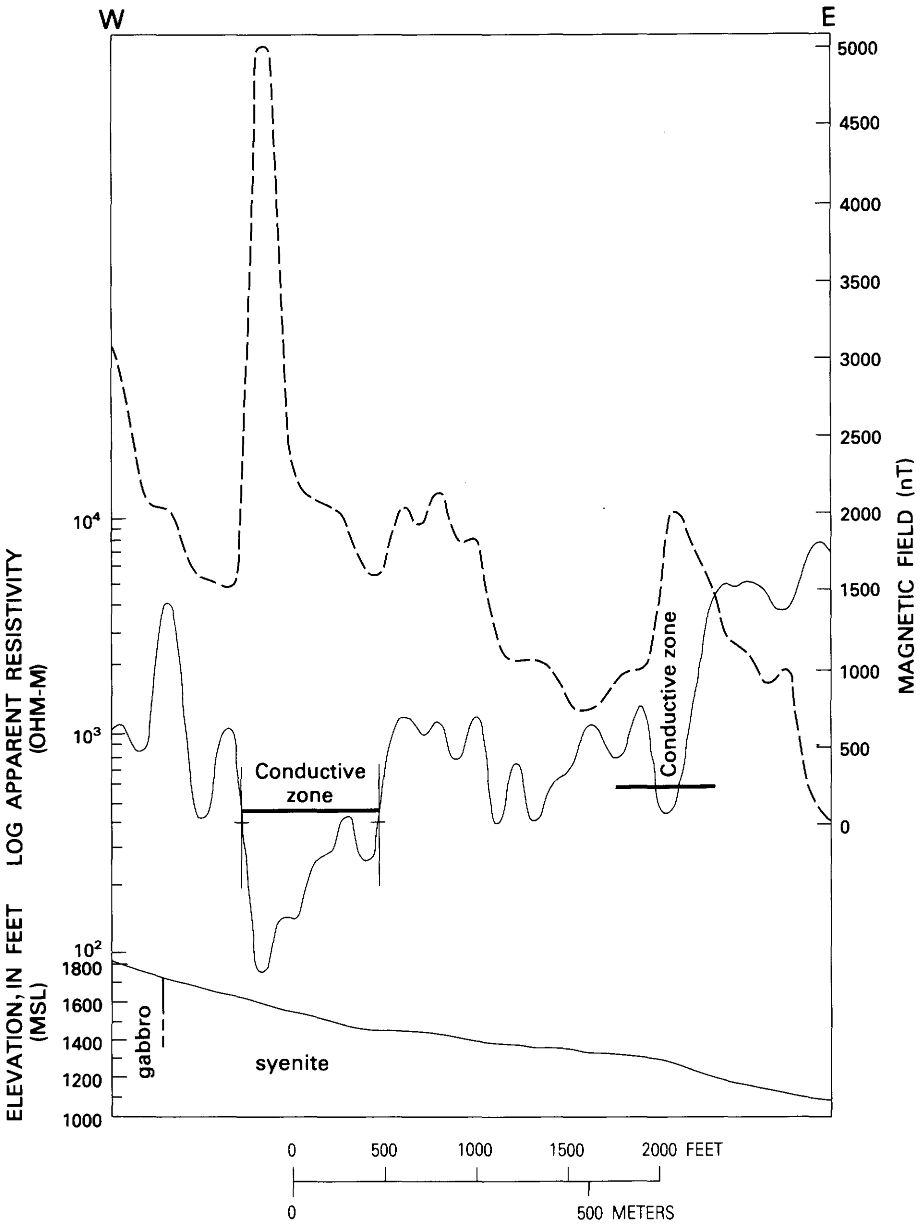

Figure 9. Ground very low frequency profiles of log apparent resistivity (solid line) and ground magnetics (dashed line) for part of flight line 73, over central and eastern portions of the Cuttingsville intrusive complex. MSL, mean sea level.

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Chapter H

Metallic Mineral Deposits in the Glens Falls 1° × 2° Quadrangle, New York, Vermont, and New Hampshire

By JOHN F. SLACK and PAUL G. SCHRUBEN

U.S. GEOLOGICAL SURVEY BULLETIN 1887

SUMMARY RESULTS OF THE GLENS FALLS CUSMAP PROJECT, NEW YORK, VERMONT, AND NEW HAMPSHIRE

CONTENTS

FIGURE

TABLE

Metallic Mineral Deposits in the Glens Falls 1° × 2° Quadrangle, New York, Vermont, and New Hampshire

By John F. Slack[1] *and* Paul G. Schruben[1]

Abstract

Historically important metallic mineral deposits in the Glens Falls 1° × 2° quadrangle include stratabound iron ores in the eastern Adirondack Mountains of New York and stratabound copper±zinc±silver ores in eastern Vermont. Small deposits of stratabound sulfides (copper±zinc±lead±silver), epigenetic veins (lead±zinc±silver±gold±arsenic±tungsten±uranium), placers (gold), and residual concentrations (iron±manganese) occur in many parts of the quadrangle. Some hydrothermal deposits of molybdenum±tin±tungsten±gold are associated with granitic plutons of Paleozoic and Mesozoic age in eastern Vermont and western New Hampshire.

INTRODUCTION

The Glens Falls 1° × 2° quadrangle of eastern New York, central Vermont, and western New Hampshire has a wide variety of metallic and nonmetallic mineral deposits. In the past, most of the production has been of nonmetallic commodities, including granite, marble, slate, graphite, talc, asbestos, garnet, feldspar, mica, stone, and sand and gravel; metals produced are Cu, Fe, Mn, Zn, Pb, Ag, Au, As, and W. Metal production in the quadrangle ceased in 1958 when the Elizabeth mine in Vermont closed. For the Glens Falls CUSMAP project, data on 590 metallic and nonmetallic mines, prospects, and mineral occurrences have been compiled (P.G. Schruben, A.J. McBean, J.F. Slack, and R.E. Nyahay, unpub. data). This report covers a subset of 60 metal deposits in the quadrangle that have recorded past production and (or) geochemical or exploration significance. For general information on metallic mineral deposits in the quadrangle, the reader is referred to Beck (1842), Jackson (1844), Hager (1861), Hitchcock (1874), Newland (1919), Meyers and Stewart (1956), Pearre and Calkins (1957a, b), Luedke and others (1959), Morrill (1960), Morrill and Chaffee (1964), and Grant (1968).

REGIONAL SETTING

Other papers in this volume give information on the regional setting of mineral deposits in the Glens Falls quadrangle. Thompson and McLelland provide a general overview of the geology of the region and a discussion of the major terranes within the quadrangle. Daniels presents a summary of newly acquired aeromagnetic and gravity data for Mesozoic plutons, including detailed information surrounding metal deposits in the Cuttingsville, Vt., area. Pohn, in a remote sensing study based on Landsat and radar images, correlates some lineaments with the distribution of mineral deposits and occurrences. Watts interprets anomalous concentrations of selected elements from regional geochemical surveys to be related in part to lineaments and to the distribution of mines and prospects.

DEPOSIT TYPES

Metallic mineral deposits in the Glens Falls quadrangle contain a variety of commodities and may be grouped into several deposit types. In this report, the deposits are assigned to five general categories on the basis of morphology and genesis. The scheme adopted here facilitates map portrayal and discussion, although some deposits are not easily classified according to the plan used. The five deposit categories, discussed in detail below, are (1) stratabound and (or) stratiform, (2) orthomagmatic, (3) granitoid associated, (4) discordant vein, and (5) residual and (or) placer. They are shown in three sizes, small, medium, and large, on the accompanying map (fig. 1) and are described in table 1. Most of the deposits (37) are small and lack production records, although many have been prospected to some extent. The 21 medium-sized deposits are those that have evidence of past mining activity and recorded production or have current economic significance. Only two of the deposits are classified as large, the Hammondville iron mines in New York and the Elizabeth mine in Vermont, each of which has significant past production of more than 1 million tons of ore.

Manuscript approved for publication April 13, 1990.

[1] U.S. Geological Survey.

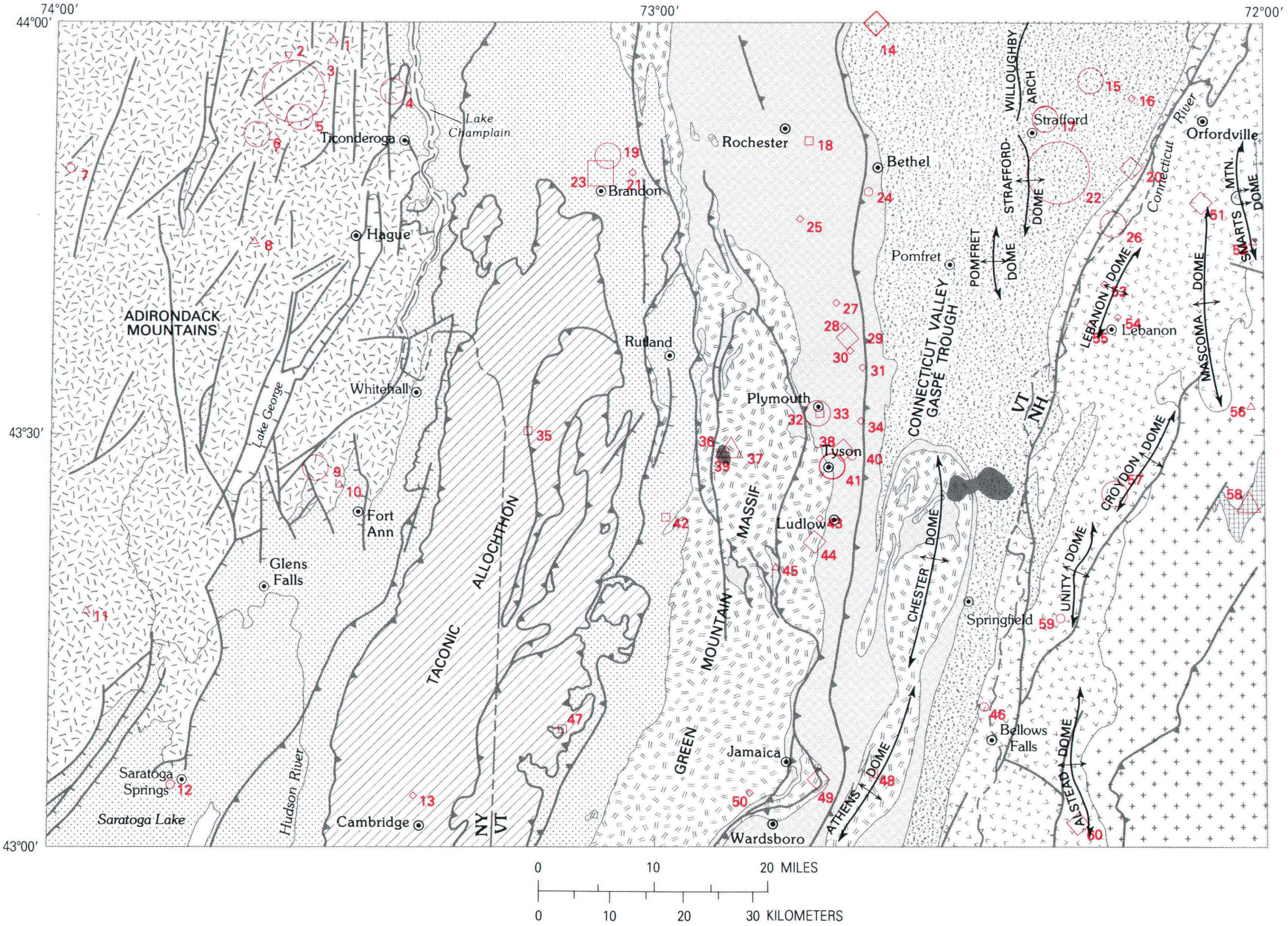

Figure 1. Map showing important metallic mineral deposits in the Glens Falls 1° × 2° quadrangle. Large symbols indicate deposits that have >1 million tons production of ore; medium-sized symbols indicate deposits that have significant past production; small symbols indicate prospects or occurrences. Center of symbol corresponds to location of mine, prospect, or occurrence. See table 1 for descriptions of numbered deposits.

EXPLANATION

CRETACEOUS AND JURASSIC (?)	■	White Mountain Plutonic-Volcanic Suite (syenite, granite, and gabbro)
UPPER DEVONIAN	▦	Granite plutons (two-mica granite)
LOWER DEVONIAN AND SILURIAN	⁺	Merrimack terrane (granitic gneiss, migmatite, minor pelitic schist, and quartzite)
	░	Connecticut Valley-Gaspé terrane (pelitic schist, calcareous schist, and minor amphibolite)
ORDOVICIAN AND CAMBRIAN	⁘	Parautochthonous shelf sequence (limestone, dolomite, and quartzite)
DEVONIAN, SILURIAN, AND ORDOVICIAN	⸪	Bronson Hill terrane (metavolcanic rocks, pelitic schist, and granitic gneiss)
ORDOVICIAN, CAMBRIAN, AND PROTEROZOIC Z	▨	Taconic allochthon (slate, graywacke, and minor carbonate)
	□	Eastern Green Mountain terrane (pelitic schist, metagraywacke, and minor amphibolite)
PROTEROZOIC Y	⁄	Adirondack massif (felsic gneiss, quartzite, anorthosite, amphibolite, calc-silicate rocks, and metagabbro)
	⸫	Green Mountain massif[1] (felsic gneiss, quartzite, minor amphibolite, and calc-silicate rocks)
	─┴─	High-angle fault—Tick marks on downthrown side where known
	─▲─	Thrust fault—Sawteeth on upper plate

[1] Includes Lincoln Mountain massif and Proterozoic rocks of the Chester and Athens domes.

Figure 1. Continued.

Stratabound and (or) Stratiform Deposits

Stratabound and (or) stratiform deposits occur in layered metavolcanic and (or) metasedimentary rocks and are broadly to perfectly conformable with their host rocks. The stratabound deposits are confined to single stratigraphic units, but are not necessarily bedded, and may, in some cases, be discordant in detail. Stratiform deposits, which are variants of stratabound deposits, have a concordant layered form and typically are bedded and have contacts parallel to the lithologic layering of enclosing host rocks. The largest metal producer of the 20 stratabound-stratiform deposits in the quadrangle is the Elizabeth mine in the Orange County copper district of east-central Vermont (fig. 1, no. 22). This mine produced approximately 3.2 million tons of ore, chiefly for copper but also including some zinc, silver, gold, copperas (iron sulfate), and pyrrhotite (McKinstry and Mikkola, 1954; Howard, 1969). Other notable stratabound-stratiform base-metal deposits in the quadrangle are at Ely, Vt. (fig. 1, no. 15); Orange and Gove, Vt. (fig. 1, no. 17); Lion Hill, Vt. (fig. 1, no. 19); Croydon, N.H. (fig. 1, no. 57); and Waterman (Blood Mountain), Vt. (fig. 1, no. 26).

A number of stratabound and stratiform, low-titanium iron deposits in the eastern Adirondack Mountains of New York State (Newland, 1908) also have recorded metal production. The largest of these comprises the Hammondville deposits (fig. 1, no. 3), which were mined intermittently until 1893 and yielded about 2 million tons of iron ore (Newland, 1908, 1919). Because the grade of the iron ore of this type of deposit commonly was very high, the Hammondville mines in actuality have produced the greatest amount of metal in the Glens Falls quadrangle. Similar but smaller iron mines that have significant recorded production in New York are the Crown Point mines (fig. 1, no. 4); the Skiff Mountain mine (fig. 1, no. 5); the Paradox Lake mines (fig. 1, no. 6); and the Fort Ann group of mines (fig. 1, no. 9), which includes the Potter, Podunk, and Mount Hope properties. Iron mines in a somewhat different geologic setting occur in Vermont along the eastern side of the Green Mountains at Plymouth Union (fig. 1, no. 32) and Tyson Furnace (fig. 1, no. 41), where small stratabound deposits of magnetite and (or) hematite were worked intermittently during the last century. The small Spathic iron mine east of Tyson (fig. 1, no. 40) is unusual in that it has iron in siderite (iron carbonate) and pyrrhotite rather than in oxides.

Orthomagmatic Deposits

Metal deposits of orthomagmatic affinity typically occur within mafic, ultramafic, or anorthositic intrusive rocks or their border zones. The concentrations of metals may take one of several forms: disseminations, pods, lenses, layers, or veins. In the Glens Falls quadrangle, only one mineral occurrence of this type is known, at the Moose Mountain prospect in the eastern Adirondacks of New York (fig. 1, no. 2), where gabbro intrudes anorthosite. This small deposit was mined around the turn of the century, but only a few hundred tons of titanium ore were removed (Newland, 1908).

Table 1. Important metallic mineral deposits and occurrences in the Glens Falls 1° × 2° quadrangle, New York, Vermont, and New Hampshire

[Number refers to map number in figure 1. Commodities are listed in decreasing order of production or importance: py=pyrite, F=fluorite, pyr=pyrrhotite. Deposit form code: om=orthomagmatic, ss=stratabound and (or) stratiform, dv=discordant vein, rp=residual and (or) placer, ga=granitoid associated]

Number	Deposit or occurrence name	Commodities	Deposit form	Latitude	Longitude
1	Spar Bed Hill pegmatite	U,Cu	ga	43°58′55″ N.	73°32′24″ W.
2	Moose Mountain deposit	Fe,Ti,py	om	43°57′33″ N.	73°37′05″ W.
3	Hammondville mines	Fe	ss	43°54′55″ N.	73°36′36″ W.
4	Crown Point (Vineyard-Butler) mines	Fe	ss	43°54′54″ N.	73°26′42″ W.
5	Skiff Mountain mine	Fe	ss	43°53′03″ N.	73°36′00″ W.
6	Paradox Lake mines	Fe	ss	43°51′45″ N.	73°40′13″ W.
7	Minerva mine	Fe,py	ss	43°49′09″ N.	73°58′31″ W.
8	Brant Lake area pegmatites	U	ga	43°44′35″ N.	73°38′45″ W.
9	Fort Ann group of mines	Fe,py	ss	43°27′44″ N.	73°33′55″ W.
10	Fort Ann area pegmatites	U	ga	43°26′38″ N.	73°31′24″ W.
11	Overlook pegmatite	U	ga	43°17′42″ N.	73°56′12″ W.
12	Saratoga Springs occurrence	Zn,F	ss	43°04′45″ N.	73°47′50″ W.
13	Little Colfax mine	Au	dv	43°04′05″ N.	73°23′50″ W.
14	Braintree mine	As,Ag	dv	43°59′50″ N.	72°38′36″ W.
15	Ely mine	Cu,Zn,pyr(?)	ss	43°55′45″ N.	72°17′10″ W.
16	Fairlee mine	Au,Ag,Cu,Pb	dv	43°54′27″ N.	72°12′57″ W.
17	Orange and Gove mines	Cu,Zn	ss	43°53′02″ N.	72°21′41″ W.
18	White River placer	Au	rp	43°51′30″ N.	72°45′00″ W.
19	Lion Hill (Nicklaw-Oram) mines	Zn,Pb,Cu	ss	43°50′19″ N.	73°05′02″ W.
20	Thetford Hill prospect	Pb,Zn,Cu,Ag	dv	43°49′25″ N.	72°13′15″ W.
21	Forest Dale mine	Pb	dv	43°49′10″ N.	73°02′35″ W.
22	Elizabeth mine	Cu,Zn,Ag,Au,pyr	ss	43°49′10″ N.	72°20′17″ W.
23	Brandon mine	Fe	rp	43°49′09″ N.	73°05′49″ W.
24	Unnamed mine	Au	ss	43°47′47″ N.	72°39′04″ W.
25	Joe Manning mine	Au	dv	43°45′47″ N.	72°45′50″ W.
26	Waterman (Blood Mountain) mine	Cu,Zn	ss	43°45′21″ N.	72°14′57″ W.
27	Chateauguay mines area	Au,Fe	dv	43°39′43″ N.	72°42′12″ W.
28	Pioneer-Carbeneau mines	Au,Pb,Cu	dv	43°38′18″ N.	72°41′29″ W.
29	Taggart mine	Au,Pb,Cu,Ag	dv	43°37′16″ N.	72°41′06″ W.
30	McKinsey mines	Au,Pb,Cu	dv	43°36′10″ N.	72°40′41″ W.
31	Joslyn mine	Au,Pb,Cu	dv	43°35′03″ N.	72°39′38″ W.
32	Plymouth Union mine	Fe	ss	43°31′49″ N.	72°44′15″ W.
33	Ottauquechee-Black Rivers placers	Au	rp	43°31′45″ N.	72°43′45″ W.
34	Hinchman prospect	Au,Pt	dv	43°31′10″ N.	72°39′48″ W.
35	Poultney River placer	Au	rp	43°30′30″ N.	73°12′45″ W.
36	Copperas Hill mine	Au,pyr,Cu,Ag,Zn	ga	43°29′28″ N.	72°52′25″ W.
37	Cuttingsville occurrence	Mo,W,Cu,Au	ga	43°29′09″ N.	72°52′58″ W.
38	Rooks mine	Au	dv	43°29′08″ N.	72°41′29″ W.
39	Granite Hill prospect	Mo,Cu	ga	43°28′50″ N.	72°53′54″ W.
40	Spathic mine	Fe,Cu	ss	43°28′43″ N.	72°40′32″ W.
41	Tyson Furnace mine	Fe	ss	43°27′58″ N.	72°42′33″ W.
42	Vail mine	Fe,Mn	rp	43°24′16″ N.	72°59′00″ W.
43	Okemo No. 4, 6 uranium occurrences	U	dv	43°24′11″ N.	72°43′50″ W.
44	Grant Brook uranium occurrence	U	dv	43°22′29″ N.	72°44′13″ W.
45	Weston area pegmatite	U	ga	43°21′22″ N.	72°47′57″ W.
46	Unnamed occurrence	Au	ss	43°10′33″ N.	72°27′40″ W.
47	Batten Kill placer	Au	rp	43°09′00″ N.	73°09′00″ W.
48	Williams-Saxtons-West Rivers placers	Au	rp	43°05′45″ N.	72°38′30″ W.
49	East Jamaica occurrence	U	dv	43°05′09″ N.	72°43′48″ W.
50	West Jamaica occurrence	U	dv	43°04′20″ N.	72°50′44″ W.
51	Holts Ledge prospects	W,Mo,Bi,As	dv	43°46′48″ N.	72°06′00″ W.
52	Pollard Hill occurrence	Mo,F,Sn,W,Zn	ga	43°43′56″ N.	72°01′06″ W.
53	Mascoma mine	Cu,As	dv	43°41′01″ N.	72°15′43″ W.
54	Reservoir Hill mine	Pb,Ag,Au	dv	43°38′36″ N.	72°14′21″ W.
55	Lebanon I-89 roadcut occurrence	Cu,Mo,Pb,Zn,As,Au	dv	43°38′11″ N.	72°17′03″ W.

Table 1. Important metallic mineral deposits and occurrences in the Glens Falls 1° × 2° quadrangle, New York, Vermont, and New Hampshire—Continued

[Number refers to map number in figure 1. Commodities are listed in decreasing order of production or importance: py=pyrite, F=fluorite, pyr=pyrrhotite. Deposit form code: om=orthomagmatic, ss=stratabound and (or) stratiform, dv=discordant vein, rp=residual and (or) placer, ga=granitoid associated]

Number	Deposit or occurrence name	Commodities	Deposit form	Latitude	Longitude
56	Aaron's Ledge mine	Au,mica	ga	43°32′06″ N.	72°01′10″ W.
57	Croydon (Corbin Park) mine	Zn,Cu,Pb,Ag,Au	ss	43°25′54″ N.	72°14′53″ W.
58	Sunapee I-89 occurrence	U	ga	43°25′06″ N.	72°01′36″ W.
59	Neal mine	Cu,pyr,Au	ss	43°16′54″ N.	72°20′06″ W.
60	Surry Mountain mine	Pb,Cu,Au	dv	43°02′01″ N.	72°18′26″ W.

Granitoid-Associated Deposits

Granitoid-associated deposits occur in or near granitic intrusive rocks of variable age and chemistry. Such deposits typically consist of ore minerals in disseminations or local fracture fillings, either within the main body of the intrusion or in adjacent country rocks. No major granitoid-associated deposits are known in the Glens Falls quadrangle, but several small deposits and occurrences have been identified. These include disseminated molybdenum in the alkaline igneous stocks of Mesozoic age on Granite Hill near Cuttingsville, Vt. (fig. 1, no. 39); molybdenum, tungsten, and tin associated with similar Mesozoic intrusions at Pollard Hill, N.H. (fig. 1, no. 52); gold reportedly related to a Devonian granitic pegmatite at Aaron's Ledge mine, N.H. (fig. 1, no. 56); gold associated with the Cuttingsville intrusive complex (Robinson, this volume) in a carbonate-hosted pyrrhotite replacement body on Copperas Hill (fig. 1, no. 36) and in a quartz-pyrite stockwork (fig. 1, no. 37); and uranium within the Late Devonian Sunapee two-mica granite (fig. 1, no. 58). Significant uranium occurrences are also known in older (Proterozoic) pegmatites at Spar Bed Hill, N.Y. (fig. 1, no. 1), near Brant Lake, N.Y. (fig. 1, no. 8), near Fort Ann, N.Y. (fig. 1, no. 10), at Overlook, N.Y. (fig. 1, no. 11), and near Weston, Vt. (fig. 1, no. 45). The only metal production recorded for these granitoid-associated deposits is from the Copperas Hill pyrrhotite mineralization at Cuttingsville (fig. 1, no. 36), which was mined for copperas (iron sulfate) in the last century.

Discordant Vein Deposits

Discordant veins, the most common type of metal deposit in the quadrangle, are present in a wide variety of host rocks and contain diverse metal commodities. Such veins are characteristically discordant to bedding or layering and are localized by structural features like faults and fractures. The veins may be further subdivided on the basis of their contained metals into (1) Au, Pb, Zn, Ag, As, and (or) Cu; (2) W, Mo, Bi, As, Cu, and (or) Au; and (3) U. Examples of the first category are widespread and include veins at Thetford Hill, Vt. (fig. 1, no. 20); Taggart, Vt. (fig. 1, no. 29); and Surry Mountain, N.H. (fig. 1, no. 60); none of these deposits has significant recorded production. Veins in the second category are less common and include those at Holts Ledge, N.H. (fig. 1, no. 51), and Lebanon, N.H. (fig. 1, no. 55). Important uranium-bearing veins and stratabound segregations (Ayuso and Ratté, this volume) are present at Okemo Mountain, Vt. (fig. 1, no. 43), and at Grant Brook near Jamaica, Vt. (fig. 1, no. 44).

Residual and (or) Placer Deposits

Deposits in this category commonly form by secondary processes and occur in unconsolidated materials such as soils (residual) and stream gravels (placer). In the Glens Falls quadrangle, these types of deposits contain two diverse metal associations; one consists of placer gold and the other consists of residual iron and (or) manganese. The placer gold deposits lack any recorded production. All are in Vermont and include those along the White River (fig. 1, no. 18); the Ottauquechee-Black Rivers (fig. 1, no. 33); the Poultney River (fig. 1, no. 35); the Batten Kill River (fig. 1, no. 47); and along the Williams, Saxtons, and West Rivers (fig. 1, no. 48). Henson (1982) lists other rivers and streams in Vermont where gold has been recovered by panning methods.

Residual iron and (or) manganese deposits occur west of the Green Mountains mainly in Bennington, Addison, and Rutland Counties, Vt. (Hager, 1861; Penrose, 1890; Burt, 1931[?]). These deposits consist of beds and local veins of iron and (or) manganese oxides and hydroxides within unconsolidated Tertiary sediments. Mining of the ores was mainly from 1889–1913, during which time Vermont ranked third in the Nation in the production of ocher (Burt, 1931[?]). The major operations were at the Brandon mine northeast of the town of Brandon (fig. 1, no. 23) and at the Vail mine near South Wallingford (fig. 1, no. 42).

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Chapter I

Detection of Geobotanical Anomalies Associated with Mineralization in the Glens Falls 1° × 2° Quadrangle

By M.S. POWER and N.M. MILTON

U.S. GEOLOGICAL SURVEY BULLETIN 1887

SUMMARY RESULTS OF THE GLENS FALLS CUSMAP PROJECT, NEW YORK, VERMONT, AND NEW HAMPSHIRE

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Detection of Geobotanical Anomalies Associated with Mineralization in the Glens Falls 1° × 2° Quadrangle

By M.S. Power[1] *and* N.M. Milton[1]

Abstract

Geobotanical studies were conducted to evaluate the efficacy of using biogeochemical surveys and spectral analyses to detect the presence of mineral occurrences in heavily forested terrain. Biogeochemical sampling was used to determine which tree species are metal accumulators and to quantify the foliar concentrations of several elements. An airborne remote sensing approach, based on spectral analyses, was also tested to determine if plant canopies associated with known mineral deposits or with metal accumulations produce anomalous spectral signatures. Biogeochemical results for study areas in the Wallingford, South Strafford, and West Topsham quadrangles in Vermont and the Mt. Cube quadrangle in New Hampshire show that the leaves of birch species accumulate greater concentrations of copper, zinc, and cobalt compared with leaves of other canopy tree species. Airborne spectroradiometer data show changes in canopy spectral signatures between forests growing on and off mineralized substrates. These techniques link mineralization in the substrate with canopy reflectance characteristics and provide quantitative and qualitative assessment of mineral resources in heavily vegetated terrain.

INTRODUCTION

Geobotanical studies derive information from vegetation that can be used to make inferences about substrate conditions. Vegetation may respond to soil geochemical conditions at the leaf level (increased metal concentrations), the species level (changes in species distribution patterns), and the community level (changes in forest canopy reflectance). In this mineral assessment study, we examine the correspondence of known mineral deposits and occurrences with biogeochemical analyses of leaf metal content, and with airborne spectroradiometric data of canopy reflectance, to determine the feasibility of using these techniques to locate potential mineral resources in forested terrain.

Biogeochemical assays provide a quantitative measure of metal content in plants (Brooks, 1972). Comparison of elemental contents of leaves from plants growing over and away from orebodies indicates which metals preferentially are taken up and sequestered within plant tissues and identifies which plant species can be used for biogeochemical surveys in specific geographic areas. To be useful for such a survey, a plant must be a metal accumulator and have a nearly ubiquitous distribution.

The remote sensing technique is based on greenhouse experiments in which spectral measurements are recorded in the laboratory for plants dosed with heavy metals (Horler and others, 1980; Chang and Collins, 1983; Milton and others, 1988). Concentrations of heavy metals are associated with spectral reflectance shifts of 10 to 20 nm (nanometer) to shorter wavelengths in the long wavelength edge of the chlorophyll *a* absorption band centered at 680 nm (fig. 1), which are evident in both reflectance and derivative spectra. Similar spectral shifts have been detected in airborne spectroradiometer data acquired over forests growing on mineralized soils (Collins and others, 1983; Milton and others, 1983).

PROCEDURES

Four areas in Vermont and New Hampshire were chosen for biogeochemical sampling on the basis of known or potential mineralization. Two of these areas were flown by using an airborne spectroradiometer 1 week prior to the biogeochemical sampling. Two massive sulfide deposits within the Vermont copper belt, at the Elizabeth mine (South Strafford 7½-minute quadrangle) and the Eureka mine on Pike Hill (West Topsham 7½-minute quadrangle), and the mineralized Cuttingsville stock (Wallingford 15-minute quadrangle) compose the Vermont study areas. Mt. Cube (Mt. Cube 15-minute quadrangle) composes the New Hampshire study area. For detailed geologic descriptions of these areas, the reader is referred to Annis and others (1983, Elizabeth mine), Slack and others (this volume, Eureka mine), Robinson (this volume, Cuttingsville stock), and Moench (this volume, Mt. Cube area).

The natural vegetation for these sites is characterized by the beech-maple-hemlock community of the eastern

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Manuscript approved for publication April 13, 1990.

[1] U.S. Geological Survey.

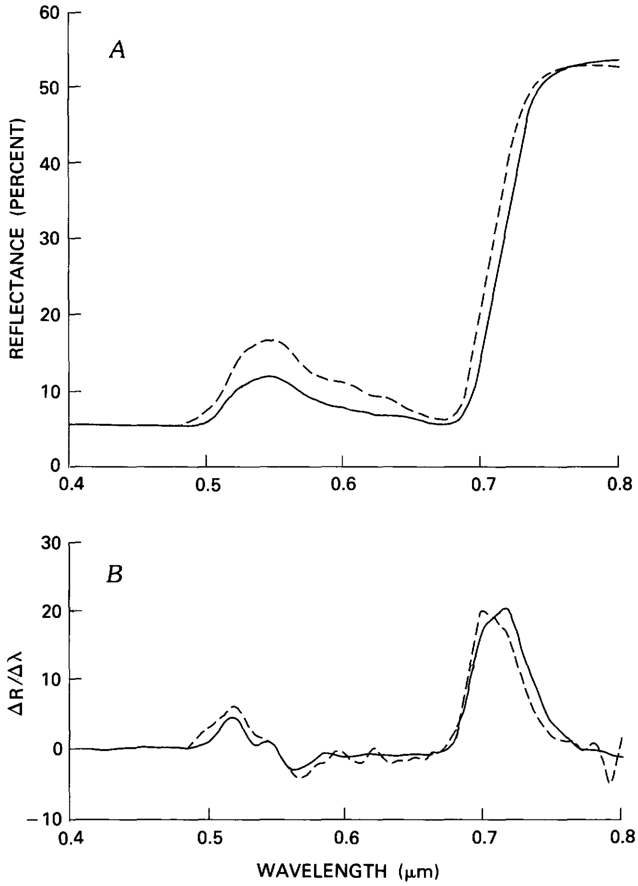

Figure 1. *A*, Spectral reflectance curves of greenhouse-grown *Hosta ventricosa* plants; $NiSO_4$-dosed plants (dashed line) and control plants (solid line). *B*, First-derivative curves of reflectance spectra.

deciduous forest province (Gleason and Cronquist, 1964). Sugar maple and beech predominate. Other deciduous species present in the study areas include northern red oak, ash, ironwood, basswood, and four species of birch. Coniferous species, such as hemlock, balsam fir, red spruce, and white pine, intermingle with the deciduous species. Scientific names of plants are given in table 1.

Table 1. Scientific and common names of plants

[Nomenclature follows Gleason and Cronquist (1963). Asterisk indicates that species was collected for biogeochemical analysis]

**Abies balsamea*	balsam fir
**Acer saccharum*	sugar maple
**Betula cordifolia*	heart-leaf paper birch
**Betula lutea*	yellow birch
**Betula papyrifera*	paper birch
**Betula populifolia*	gray birch
**Fagus grandifolia*	beech
**Fraxinus* sp.	ash
Hosta ventricosa	hosta
**Ostrya virginiana*	ironwood
**Picea rubens*	red spruce
**Pinus strobus*	white pine
**Quercus rubra*	northern red oak
Tilia americana	basswood
Tsuga canadensis	hemlock

Biogeochemical sampling was conducted at all four sites. Sampling techniques included the identification of canopy and subcanopy trees, the mapping of changes in species distribution on traverses that cross orebodies, and the collection of leaves for elemental analysis from canopy and subcanopy trees along traverses. Sampling traverses were established at each site to cover zones of known or potential mineralization and background or unmineralized areas. At each mineralized zone, a traverse was laid out consisting of 6 to 12 sample sites. At each sample site, leaves from several canopy and subcanopy trees were collected within a 3-m radius. Each sample contained 30 to 100 leaves, collected from 2 to 6 trees of the same species. The number of leaves collected depended on foliar size and texture. This reconnaissance sampling technique was used to acquire an average concentration of elements per species per site and to determine which species were metal accumulators.

To establish background values, leaves of each species were collected along one traverse away from a mineralized area. These results were supplemented with published data of Connor and Shacklette (1975) and unpublished data collected by Frank Canney (oral commun., U.S. Geological Survey, 1981).

Prior to analysis, leaves were washed and rinsed in deionized water to remove surface contaminants, oven dried, ground, and ashed. Ashed samples were analyzed either by Jerry Motooka (U.S. Geological Survey, Denver) by using induction-coupled-plasma (ICP) spectroscopy or by Nancy Milton (U.S. Geological Survey, Reston) by using atomic absorption spectroscopy.

An airborne spectroradiometer survey was conducted for two of the biogeochemical sites, and the digital spectral data were processed by Geophysical Environmental Research (GER), Inc. The Mark II spectroradiometer is configured to acquire 512-channel data, at bandwidths of 1.48 nm, between 260 and 1020 nm. This instrument was flown over Pike Hill and Cuttingsville at an altitude of approximately 600 m in early September 1983, prior to leaf senescence. Flight lines for these two sites were gridded and flown without prior knowledge of the location of specific mineralized areas. An average of 280 spectral measurements was collected for each flight line. Each spectrum represents energy integrated over an instantaneous field of view for an area of 18×18 m and therefore consists of radiance data from several different species of canopy trees.

Aircraft data were computer processed by using waveform analysis to remove atmospheric noise. The results produced by waveform analysis constitute two sets of spectral curves. The first set represents background (normal) radiance values for surface features, such as forest canopies, fields, roads, and water. The second set of curves represents deviations from the background values in the first set. If the deviations, which consist of slope changes or amplitude differences, coincide with a closed forest canopy, the presence of spectral anomalies in the canopy is indicated. The anomalous spectra suggest potential areas of mineralization (Collins and others, 1983).

RESULTS AND DISCUSSION

Biogeochemical Technique

Of the 12 species of trees sampled for biogeochemical analysis, birches show the highest concentration of metals. At the Pike Hill massive sulfide locality, heart-leaf paper birch growing near the Eureka mine contains anomalously high concentrations of copper and zinc (maximum values of 350 and 4,500 ppm, respectively) compared with background concentrations (100 and 1,000 ppm, respectively) away from the orebody (fig. 2). Yellow birch has high concentrations of zinc (2,500–5,300 ppm) but has low concentrations of copper (80–280 ppm, which approximate background levels). Vegetation mapping indicates that sugar maple, although nearly ubiquitous in its distribution, is absent near the Eureka mine site. Metal values for sugar maple are uniformly low, even where it is found growing with birches that contain anomalous metal concentrations. Birches increase in dominance at base-metal sulfide mines and constitute most of the natural reforestation on abandoned mine sites. High concentrations of metals, however, appear to be deleterious to these trees, as the main stems of affected birches consist of stunted sucker shoots that grow from older root systems. Anomalous metal concentrations in the substrate may affect plant morphology directly through the accumulation of toxic levels of metals in plant tissues or indirectly through inhibition of water and (or) nutrient availability.

At the Elizabeth mine site, three birch species grow along the open pit where copper- and zinc-bearing massive sulfide ore was mined until 1958. Gray birch there accumulates the greatest amounts of zinc and copper (maximum values of 11,000 and 370 ppm, respectively (figs. 3*A*, *B*)). Paper birch also exhibits high concentrations of zinc (8,250 ppm) but contains amounts of copper that bracket the background values of 100 ppm. Near the pit, all three birch species contain elevated concentrations of cobalt (fig. 3*C*) and nickel (fig. 3*D*). White pine has high concentrations of nickel only (200 ppm). Concentrations of cobalt in birches range from 5 to 145 ppm, compared with the background value of 20 ppm. Nickel values in birches vary from 50 (background) to 250 ppm. At the Elizabeth mine background values for copper are similar to the background values for copper at Pike Hill (100 ppm); however, background values for zinc at the Elizabeth mine (2,000 ppm) are twice those at Pike Hill (1,000 ppm). As at Pike Hill, the birches at the Elizabeth mine exhibit stunted growth of sucker shoots.

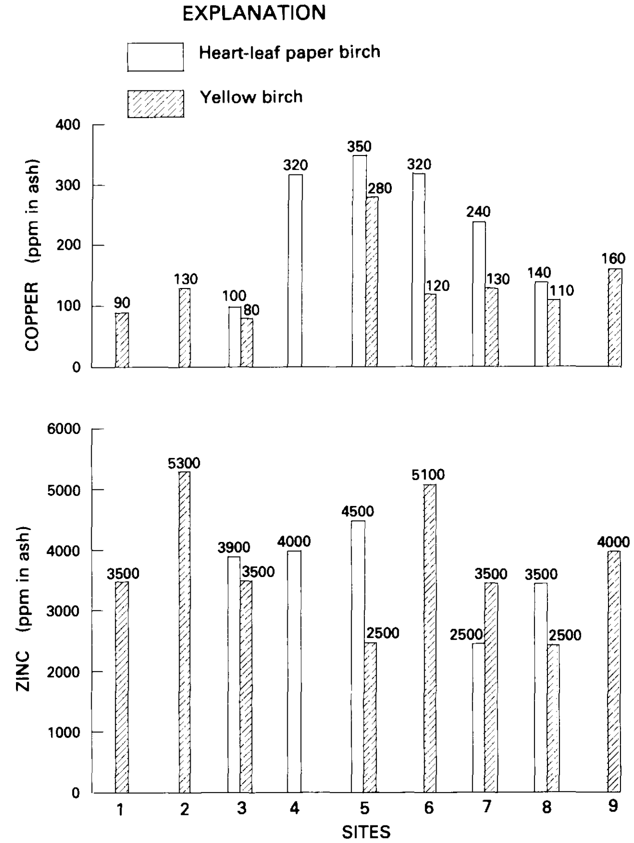

Figure 2. Biogeochemical analyses from induction-coupled-plasma spectroscopy of plant leaves collected along a transect that includes the Eureka mine on Pike Hill near East Corinth, Vt. The background value for copper is 100 ppm, and the background value for zinc is 1,000 ppm. Location of transect sites: 1, southwestern slope of Pike Hill; 2, western slope of Pike Hill; 3, western side of mine dump; 4, mine dump; 5, eastern side of mine dump; 6, along mine trench; 7, along mine trench; 8, along mine trench; 9, northeastern slope of Pike Hill.

The area of the Cuttingsville stock is covered primarily by a mature forest of sugar maple, ash, northern red oak, and beech. A few yellow and paper birches are present, but there are no gray or heart-leaf paper birches. ICP analyses of elements in ashed leaves from these species indicate relatively low concentrations, approaching background values. No metal accumulation occurred in the species sampled.

The site at Mt. Cube, N.H., varies in vegetation cover. The flanks of the hill are covered with a mixed hardwood-coniferous canopy as at the other sites in Vermont. On top of the hill, however, heart-leaf paper birch, red spruce, and balsam fir predominate. These three species were sampled along a transect that parallels the ridge, but only heart-leaf paper birch shows anomalous amounts of zinc (4,390 ppm) and copper (180 ppm) (fig. 4) and arsenic (28 ppm) and cadmium (33 ppm). Site numbers 2 and 7

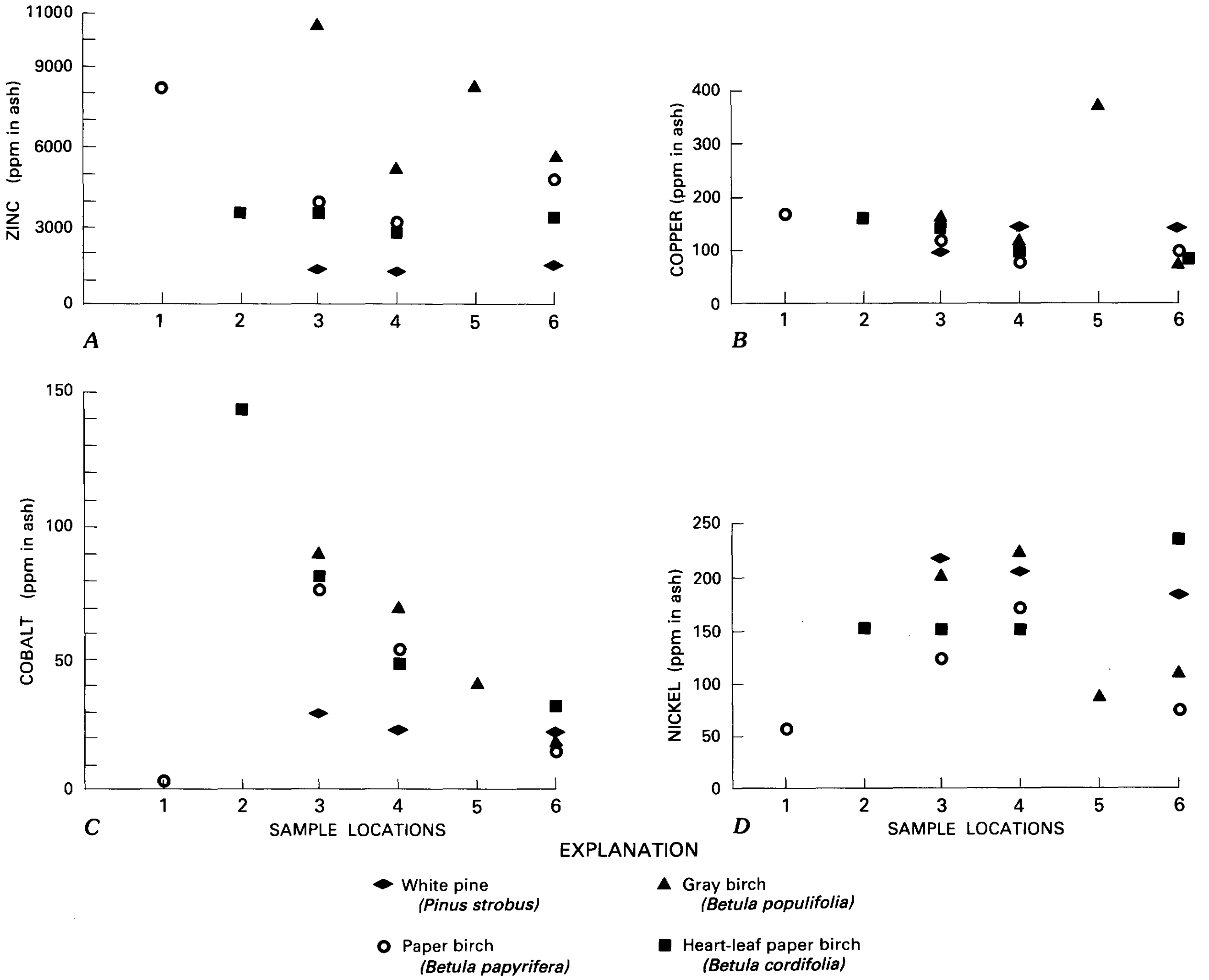

Figure 3. Biogeochemical analyses from atomic absorption spectroscopy of plant leaves collected at the Elizabeth mine, Vermont. Background values are as follows: zinc, 2,000 ppm; copper, 100 ppm; cobalt, 20 ppm; nickel, 50 ppm. Sample locations: 1 and 2, southern end of south pit; 3, western rim of south pit; 4, middle of south pit; 5, dump at northern end of south pit; 6, 0.8 km east of sulfide.

(fig. 4) show background values for zinc and copper. Background values for arsenic and cadmium are 8 and 9 ppm, respectively. Low but measurable amounts of tin (23 ppm) and tungsten (12 ppm) also are found in these leaves. Spruce and fir do not contain anomalous amounts of any of these elements and show no trace of tin or tungsten in ashed samples. There are no known mineral deposits at Mt. Cube, but these biogeochemical results warrant further study of this area.

Of the 12 tree species sampled, the 4 birch species exhibit the greatest amount of metal accumulation and occur throughout the study area, thus making them the species of choice for biogeochemical sampling. Sugar maple was eliminated as a potential biogeochemical indicator species due to the uniformly low concentrations of metals at all sites. The branch height of northern red oak precluded leaf collection, and it, too, was eliminated from consideration. Beech, ash, and ironwood either lack evidence of metal accumulation or occur too infrequently for systematic sampling. Coniferous tree species, such as balsam fir and red spruce, generally show lower metal concentrations; white pine is the exception at the Elizabeth mine, where a nickel accumulation was noted.

In this study, biogeochemical surveys were conducted nearly simultaneously with acquisition of the remote sensing data. A more useful exploration technique is to collect biogeochemical samples from areas previously identified as spectrally anomalous in airborne spectroradiometer data. However, even without a remote sensing component, biogeochemical surveys can provide an alternative or a complement to soil geochemical surveys in vegetated terrain.

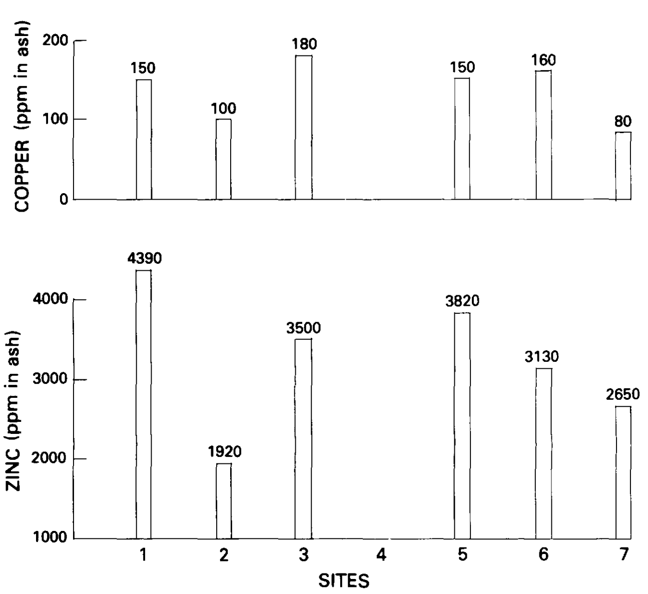

Figure 4. Biogeochemical analyses from induction-coupled-plasma spectroscopy of heart-leaf paper birch leaves collected at Mt. Cube, N.H. Sites 2 and 7 show background values for both elements.

Remote Sensing Technique

Airborne spectroradiometer data for the Pike Hill and Cuttingsville localities display anomalous spectral signatures for closed forest canopies near areas of known mineralization. Spectral reflectance curves derived from the raw radiance data lack evidence of spectral shifts in the chlorophyll *a* absorption band due to interference by atmospheric absorption. However, spectral curves derived from waveform analysis show the location of spectral anomalies for both the Pike Hill and Cuttingsville sites. Anomalies are classified as strong or weak on the basis of empirical observations of deviations from background waveform curves. After the anomalies were plotted on aerial photographs, the results were compared with geological and geochemical data provided by other investigators (Slack and others, this volume; Robinson, this volume).

At the Pike Hill locality, 13 flight lines were gridded and flown over the Eureka mine and surrounding terrain. Figure 5 compares locations of anomalous spectra and locations of mineralized drainage basins as sampled and reported by Slack and others (this volume). One anomaly coincides with biogeochemical data collected on a transect along the Eureka mine, where high copper and zinc values occur in leaves of birch trees. Other spectral anomalies correspond to high elemental values in stream sediment and panned concentrate samples (Slack and others, this volume). Strong spectral anomalies associated with high metal values in panned concentrate and stream sediment samples occur north and east of Pike Hill. The panned concentrate samples have anomalous values for copper, zinc, and cobalt; copper shows the highest concentration. The stream sediment samples contain high values for copper and cobalt and lower values for zinc, gold, and lead. Other areas of correspondence between spectral and geochemical data include four drainage basins south and west of Pike Hill that show anomalous concentrations of gold.

At Cuttingsville, 16 flight lines were gridded and flown. Figure 6 compares the locations of anomalous spectra and the locations of hydrothermal alteration and mineralization mapped by Robinson (this volume). Spectral anomalies coincide with three areas of hydrothermal alteration. Two anomalies fall along a replacement pyrrhotite body (fig. 6, A), which includes the site of an abandoned mine on the southwestern side of Copperas Hill. Anomalous signatures delineate the western edge of the Cuttingsville stock, one of which coincides with a molybdenum prospect (fig. 6, B). Another anomaly is located just south of a pyritic stockwork along the Mill River (fig. 6, C). Anomalous spectra are also associated with xenolith-rich dikes to the west of Shrewsbury (fig. 6, D).

The delineation of potentially mineralized areas by spectroradiometric surveys can be done only over a closed forest canopy. Spurious signals result over (1) a thin canopy, where spectral reflectance measurements are a composite of plant and soil or (2) over cultural features, such as roads and pastures. Therefore some areas of known mineralization, reported by Slack and others (this volume) for Pike Hill and by Robinson (this volume) for Cuttingsville, are not identified by using this technique. Conversely, some anomalous spectra, plotted on figures 5 and 6, cannot be explained at present by geological mapping or geochemical sampling. However, many of these spectra coincide with areas for which no field data were collected, and buried mineralization may be indicated.

In spite of these limitations, the results of this geobotanical remote sensing technique indicate that a correlation exists between metal anomalies in the substrate and spectral anomalies in the plant canopy. Although this correlation has been demonstrated quantitatively and qualitatively in greenhouse experiments (Milton and others, 1988), the precise mechanism or pathway by which metals affect plant reflectance is not well understood. The value of this technique is based on the ability to survey large forested areas and pinpoint spectral anomalies, which can then be used to guide collection and analysis of ground data.

The remote sensing technique produces results for a wide range of soil types, and the lack of metal specificity is an advantage for locating areas of potential mineralization. Of special interest in this study is the correlation of spectral anomalies and the presence of gold, either indirectly, as with the panned concentrates and stream sediments in the Pike Hill area (Slack and others, this volume), or directly, as with the syenite-associated deposits at Cuttingsville (Robinson, this volume). Separation of spurious spectral signatures from those indicative of soil metal anomalies

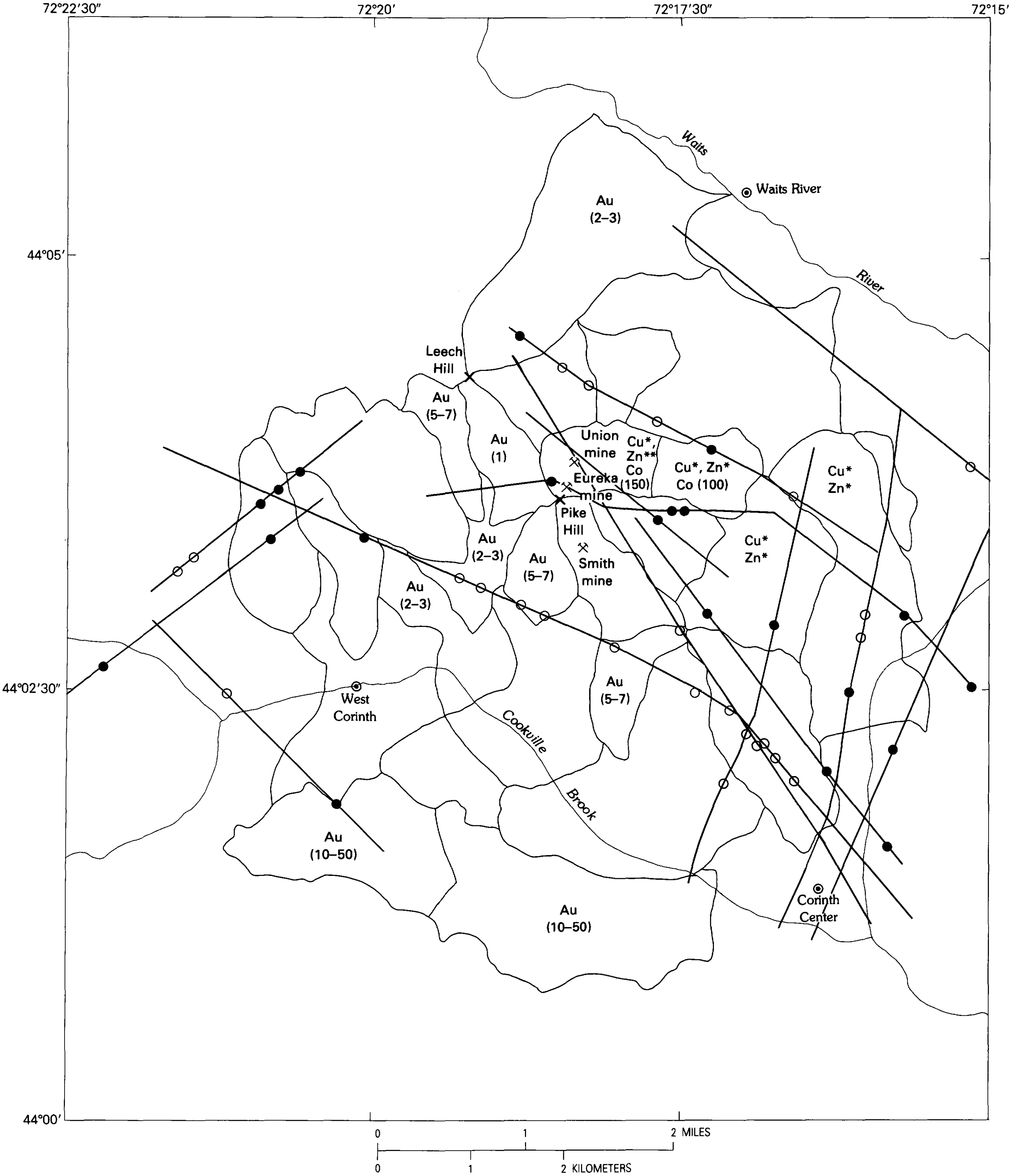

Figure 5. Results of airborne spectroradiometer survey conducted in the Pike Hill area, Vermont. Map shows outlines of stream drainage basins correlated with panned concentrate anomalies (numbers in parentheses are in parts per million) collected by Slack and others (this volume). Cu*, 10,000–50,000 ppm; Zn*, 5,000–10,000 ppm; Zn**, 2,000–3,000 ppm. Superimposed is the flight line grid. Closed circles represent strong spectral anomalies from the forest canopy, and open circles represent weak spectral anomalies.

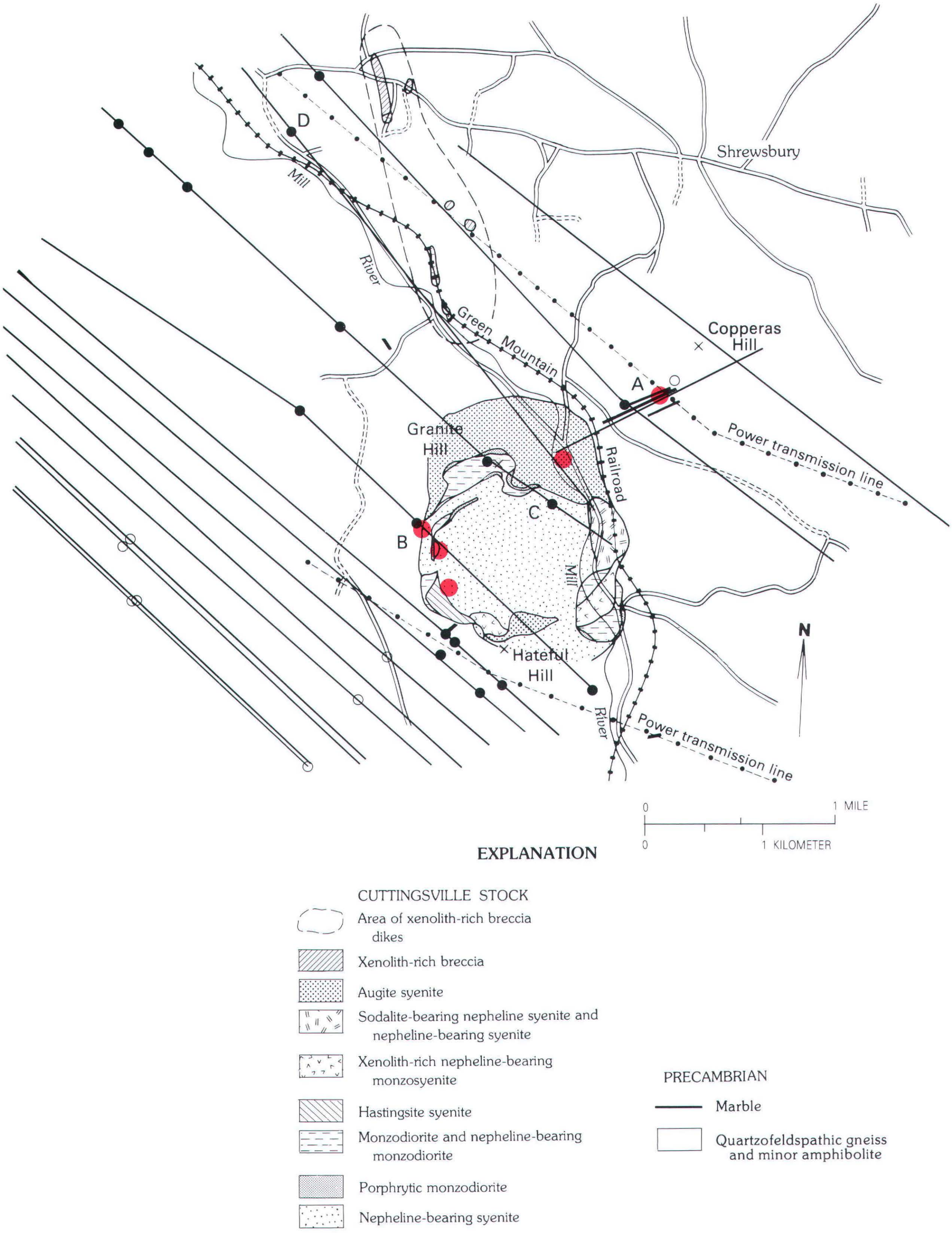

Figure 6. Results of airborne spectroradiometer survey conducted in the Cuttingsville area, Vermont. Superimposed is the flight line grid. Red dots represent areas of hydrothermal alteration and (or) mineralization. Closed circles represent strong spectral anomalies, and open circles represent weak spectral anomalies. Letters A–D refer to anomalies discussed in the text. Map modified from Laurent and Pierson (1973).

may ultimately be possible by using more sophisticated data processing techniques.

CONCLUSIONS

Biogeochemistry and airborne spectroradiometry can be used to detect anomalous vegetation associated with mineralization. Biogeochemical analyses show that birch species accumulate greater concentrations of metals than other deciduous tree species. This information, combined with vegetation mapping, shows that birch, although not ubiquitous, is the optimal species for mapping contours of metal availability. Sugar maple, although nearly ubiquitous in distribution, has low foliar concentrations of metals, at or near background values. Other deciduous trees need to be sampled more extensively to determine degrees of metal accumulation. Of the coniferous species, white pine accumulated nickel at one massive sulfide locality. Conifers generally have low concentrations of metals in their needles.

Airborne spectroradiometric data indicate that anomalous spectral signatures are produced in forest canopies associated with elevated metal concentrations in the substrate. Causal relations between biogeochemical data and spectral analyses are not well understood. Spectral anomalies may correspond to direct effects of metal accumulation or to indirect environmental effects, such as water stress or lack of nutrient availability.

Airborne spectroradiometric surveys appear to have significant potential for selecting exploration targets for more detailed geologic and biogeochemical sampling. Anomalous spectral signatures from closed canopy forests on Pike Hill correspond with drainage basins that contain high stream sediment values for copper, zinc, cobalt, and gold. Several spectral anomalies around the edge of the Cuttingsville stock coincide with areas of known hydrothermal alteration and mineralization, some of which host anomalous amounts of gold. Spectral anomalies need to be verified by detailed ground investigations, such as biogeochemistry and soil geochemistry, and by geologic mapping.

Although these techniques can be used independently, they are more effective when used as part of a multidisciplinary approach. Together they offer a quantitative and qualitative link between substrate, plant, and forest canopy. The link between substrate metal content and canopy reflectance provides an important contribution to the assessment of mineral resources in heavily vegetated terrain.

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Chapter J

The Piermont Allochthon, Northern Connecticut Valley Area, New England—Preliminary Description and Resource Implications

By ROBERT H. MOENCH

Comparative massive-sulfide mineral resources of allochthonous and autochthonous metavolcanic rock assemblages in northern New England

U.S. GEOLOGICAL SURVEY BULLETIN 1887

SUMMARY RESULTS OF THE GLENS FALLS CUSMAP PROJECT, NEW YORK, VERMONT, AND NEW HAMPSHIRE

CONTENTS

FIGURES

The Piermont Allochthon, Northern Connecticut Valley Area, New England—Preliminary Description and Resource Implications

By Robert H. Moench[1]

Abstract

New geologic mapping in the Mt. Cube 15-minute quadrangle, New Hampshire and Vermont, and in several quadrangles to the northeast, indicates that an extensive belt of rocks previously considered to lie stratigraphically below the Middle Ordovician Ammonoosuc Volcanics actually is mainly Silurian in age and allochthonous. This belt, called the Piermont allochthon, may extend 300 kilometers or more northeastward from the Mt. Cube quadrangle into northwestern Maine. The allochthon is composed of metamorphosed siliciclastic turbidites and locally abundant subaqueous bimodal volcanic rocks that accumulated without apparent interruption from Late Ordovician, through Silurian, and into Early Devonian time. A model is proposed in which deposition occurred 25 to 50 kilometers to the southeast (present distance) of the present position of the allochthonous rocks on a postulated platform that is no longer recognizable because of Acadian compression. The platform, with subbasins, is thought to have existed just northwest of the margin of the deep basin that was ancestral to the Kearsarge-central Maine trough. According to this model, subaqueous sedimentation and bimodal volcanism on the platform were accompanied by emplacement of mafic and bimodal dike swarms, sheeted intrusive bodies, and small Silurian to Early Devonian granitic plutons and by hydrothermal alteration and sea-floor sulfide mineralization. The tectonic setting was extensional.

Sea-floor mineralization on the postulated platform produced many small, volcanogenic, copper-dominant sulfide deposits, which formed where the original host volcanics were interbedded with abundant shaly sediments that reduced permeability and thereby inhibited hydrothermal circulation. The relatively larger polymetallic deposits at Ledge Ridge, Maine, and in the Clinton River district, southeastern Quebec, formed near a major Silurian volcanic center; volcanogenic copper-zinc deposits at Thrasher Peaks, Maine, formed in association with an Early Devonian volcanic center.

During an early stage of the Acadian orogeny, the platform was greatly compressed, and its cover of sedimentary, volcanic, and intrusive rocks was transported to the present position of the Piermont allochthon. Here, the allochthon truncates an autochthonous sequence containing two metavolcanic belts having potential for the occurrence of massive sulfide deposits—the Upper Cambrian(?) Jim Pond Formation of the Boundary Mountains anticlinorium and the Middle Ordovician Ammonoosuc Volcanics of the Bronson Hill anticlinorium. Production from two small massive sulfide deposits in the Ammonoosuc has yielded a total of about 600,000 tons of copper-zinc ore.

Potential for the occurrence of undiscovered massive sulfide deposits in all of the metavolcanic belts, regardless of age, is determined by (1) adequate thickness of originally permeable volcanics to host energetic hydrothermal circulation, (2) evidence of a volcanic hiatus and chemical sedimentation, (3) evidence of hydrothermal circulation in the form of premetamorphic alteration, (4) evidence of a heat source to drive circulation, and (5) evidence of tectonic extension. Later processes, such as shearing that accompanied incorporation of the Jim Pond Formation into an accretionary mélange, may reduce the massive-sulfide favorability of a metavolcanic belt.

The many small, crosscutting, locally gold-bearing quartz-carbonate-sulfide veins of the "Ammonoosuc gold field" and elsewhere are inferred to have originated during brittle fracturing and hydrothermal circulation that probably mainly accompanied Mesozoic emplacement of high-level plutons. The mineral content of these veins may have come from nearby rocks and stratabound sulfide deposits.

INTRODUCTION

The bedrock geology of northern New England, as depicted on geologic maps of Maine, New Hampshire, and Vermont published prior to 1968, has been extensively revised as a result of mapping done since the 1960's and, particularly in New Hampshire, since 1978 as part of the Sherbrooke-Lewiston and Glens Falls CUSMAP (Conterminous United States Mineral Assessment Program) projects. Most of the new work in Maine is incorporated on the new bedrock geologic map of Maine (Osberg and others, 1985). A preliminary version of the bedrock geologic map of the Sherbrooke and Lewiston 1° × 2° quadrangles is available (Moench, 1984) but is being revised.

Manuscript approved for publication April 13, 1990.

[1] U.S. Geological Survey.

In 1985, I was asked to map in the Mt. Cube and adjacent 15-minute quadrangles, New Hampshire and Vermont (fig. 1), and to extend newly mapped metamorphic stratigraphy in the western part of the Lewiston 1° × 2° quadrangle southwestward into the Glens Falls quadrangle. I quickly learned, however, that new information was to flow the other way. In short, evidence uncovered by mapping in the Mt. Cube quadrangle and farther northeast indicates that rocks of an extensive belt previously considered to be Ordovician or older in age and autochthonous are actually mainly Silurian and allochthonous. The allochthonous rocks constitute the Piermont allochthon (Moench and others, 1987), shown on figure 1. I wish to emphasize that my present view of the allochthon is based on limited field work—3 months in 1985 and about 10 weeks during 1986–89. Accordingly, both the physical delineation of the allochthon on figure 1 and the origin that I propose in this report will be subject to considerable revision. The name of the allochthon is taken from the village of Piermont, N.H., because both the age of the rocks and their tectonic setting were recognized nearby.

The new age assignments are based on detailed correlations of the allochthonous formations in the Piermont area with the Upper Ordovician, Silurian, and Lower Devonian sequences of western Maine (Moench and Boudette, 1970, 1987; Moench and Pankiwskyj, 1988) and are supported by U-Pb zircon studies by J.N. Aleinikoff. The Fairlee pluton west of Piermont, N.H. (fig. 1, F), long considered to be a member of the Ordovician Highlandcroft Plutonic Suite, but intrusive into rocks that I correlate with Silurian formations of western Maine, has yielded an Early Devonian U-Pb zircon age of 410±5 Ma (Moench and Aleinikoff, 1987; age about 395 Ma reported but revised by J.N. Aleinikoff, oral commun., 1989, on the basis of new data and recalculations). On the basis of this age and the field relations, the Fairlee pluton appears to have intruded the Piermont allochthon during or after its emplacement. Aleinikoff (written commun., 1989) also obtained isotopically indistinguishable U-Pb zircon data that yield ages of about 412 to 416 Ma from two bodies of hypabyssal intrusive rocks and one rhyolite metatuff bed, all within the former Albee Formation (not used in this paper) in the type area of the Albee on Gardner Mountain (fig. 1; Billings, 1935), about 20 km west of Littleton, N.H. The dated samples are from rocks now assigned to the volcanic-bearing facies of the Perry Mountain Formation.

Zircon from the metatuff bed yielded the most definitive age—upper intercept at 414±4 Ma and $^{207}Pb/^{206}Pb$ ratio at 416±3 Ma (J.N. Aleinikoff, written commun., 1989)—because it dates the associated stratified rocks, whereas the data from the intrusive rocks show only that the country rocks can be no younger than about 414 Ma, even though the intrusive and volcanic rocks are logically interpreted to be comagmatic. Although 414 Ma is 5 to 15 m.y. younger than the expected age of the Perry Mountain Formation (Silurian, Wenlockian(?)), the actual paleontologic age of the Perry Mountain and the numerical time scale for the Late Silurian are poorly constrained.

As presently mapped (fig. 1), the Piermont allochthon extends along the northern Connecticut valley from near Orfordville village, in the Mt. Cube quadrangle, northeastward to the Gore Mountain plutons (fig. 1, GM) and tentatively across the Canadian border. Regional relations suggest, however, that the allochthon may actually extend at least 300 km from the Piermont area to Caucomgomoc Lake in northwestern Maine at the northeastern terminus of the Frontenac Formation (Osberg and others, 1985), northeast of the area of figure 1. The maximum inferred width of the allochthon in the area of figure 1 is about 19 km.

Moench and others (1987) and Moench (1989) describe the classic work by Billings (1935, 1937) and his associates in northern New Hampshire and discuss the stratigraphic and structural revisions that have resulted from mapping done during 1983–88. This report summarizes the critical stratigraphic and structural relations that were uncovered near Piermont in 1985 and discusses the geologic and resource implications farther northeast.

GEOLOGY OF THE PIERMONT AREA

Figure 2 depicts the bedrock geology of the Piermont 7½-minute quadrangle (south of lat 44°00′ and east of long 72°07′30″ on fig. 2) and adjacent parts of the Fairlee and Newbury 7½-minute quadrangles, respectively west and north of the Piermont quadrangle. Structural interpretations are shown on figure 3. Metamorphic rank is amphibolite facies east of the Ammonoosuc fault (Spear and Rumble, 1986) and greenschist facies west of the fault (Rumble and Finnerty, 1974).

Stratigraphy

Following Moench and others (1987), the stratified metamorphic rocks of the Piermont area (fig. 2) can be separated into the autochthonous Bronson Hill sequence, characteristic of the Bronson Hill anticlinorium (fig. 1), and the Piermont allochthon sequence (see fig. 4).

Piermont Allochthon

The map pattern shown in figure 2 is broadly comparable to the patterns mapped by Hadley (1942, 1950) and White and Billings (1951) in the Mt. Cube and Woodsville 15-minute quadrangles, respectively. Rumble (1969) revised some of Hadley's (1942) mapping east of the Connecticut River. The fundamental difference between my mapping and that of Hadley, White and Billings, and Rumble is that I recognize seven formations (plus members)

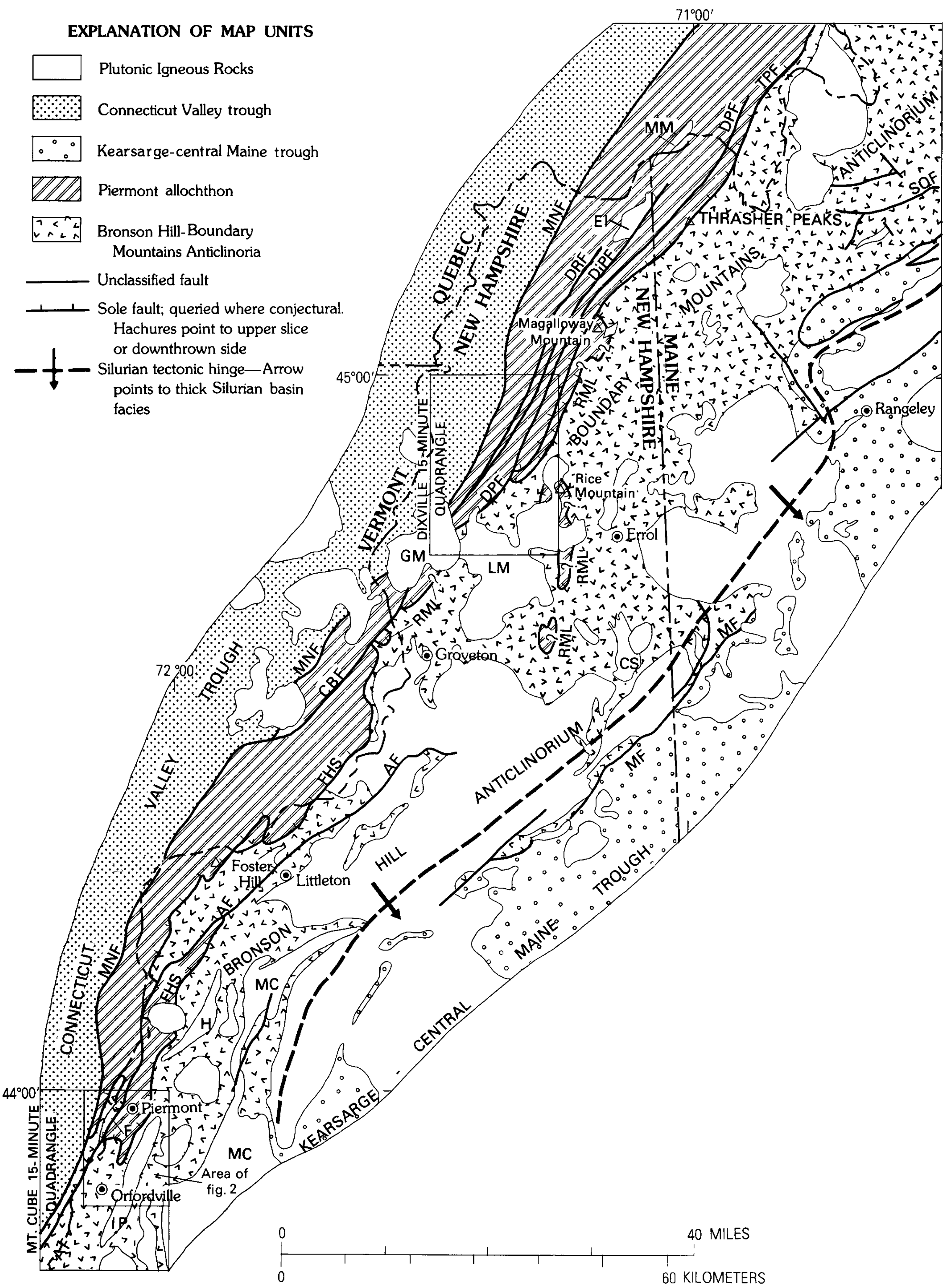


Figure 1. Major tectonic features of northern New Hampshire and adjacent areas. Named faults: AF, Ammonoosuc fault; CBF, Connary Brook fault; DPF, Deer Pond fault; DRF, Dead Water Ridge fault; DiPF, Diamond Pond fault; FHS, Foster Hill sole fault; MF, Mahoosuc fault; MNF, Monroe fault; RML, Rice Mountain line; SQF, Squirtgun fault; TPF, Thrasher Peaks fault. Plutons referred to in text (north to south): MM, sheeted intrusions at Marble Mountain; EI, East Inlet; GM, Gore Mountain; LM, Long Mountain; CS, Chickwolnepy; MC, Mount Clough; H, Haverhill; F, Fairlee; IP, Indian Pond.

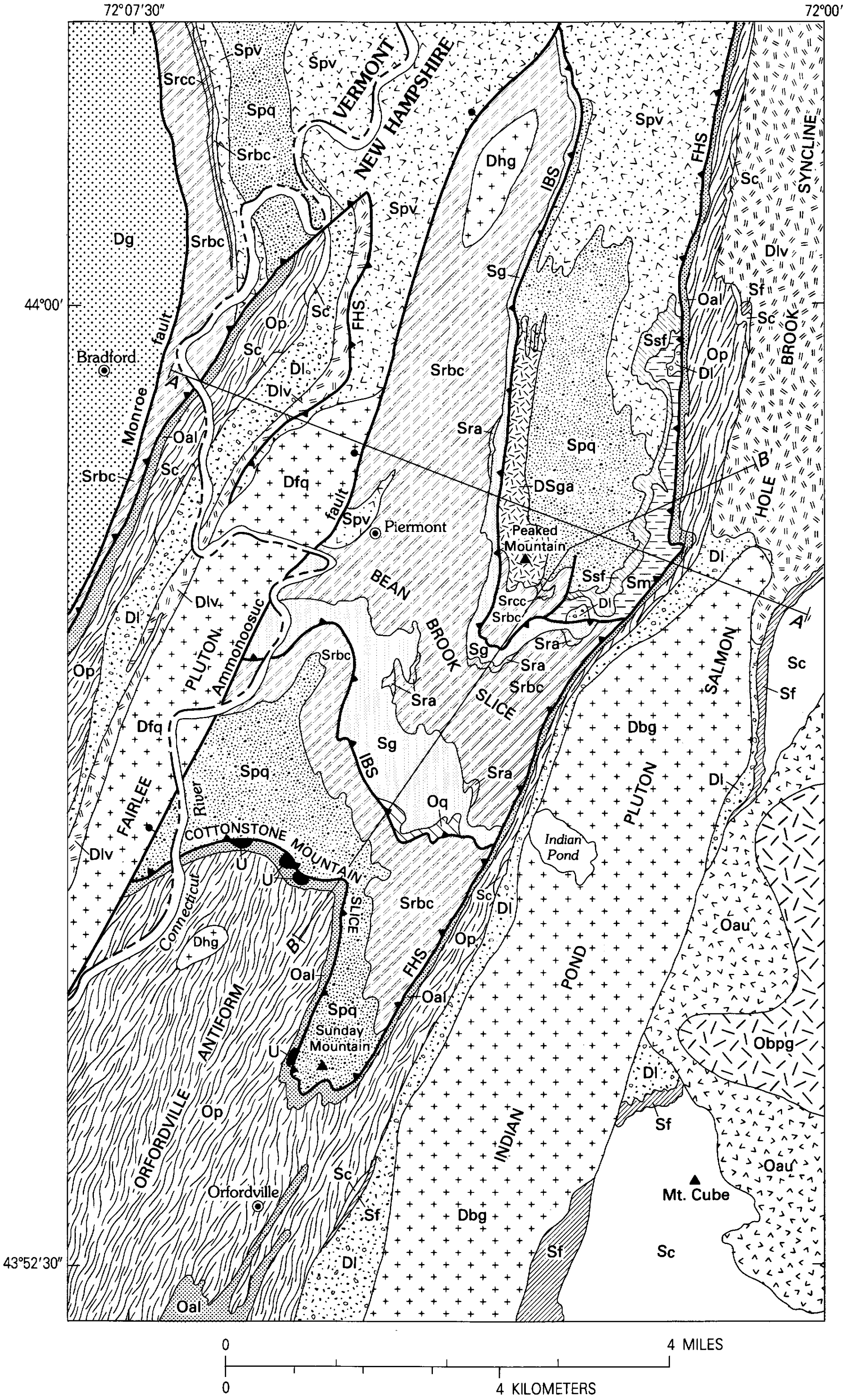


Figure 2. Geology of the Piermont 7½-minute quadrangle and adjacent parts of the Fairlee and Newbury 7½-minute quadrangles, Vermont and New Hampshire. Modified from Hadley (1942, 1950), White and Billings (1951), and Rumble (1969). See figure 3 for cross sections. West of the Ammonoosuc fault, the stratified metamorphic rocks are mainly within the biotite zone, and east of the fault, the rocks are mainly within the staurolite (±kyanite) zone.

EXPLANATION OF MAP UNITS

INTRUSIVE ROCKS

Dhg Dfq Dbg	Granitic rocks (Devonian)—Dhg, Haverhill Granodiorite (Hadley, 1942); Dfq, Fairlee Quartz Monzonite; Dbg, Bethlehem Gneiss
DSga	Sheeted metagabbro and metadiabase (Devonian or Silurian)
U	Ultramafic talc schist (Devonian(?))—Possible intrusions along Foster Hill sole fault; plotted from Rumble (1969)
Obpg	Baker Pond Gneiss (Hadley, 1942) (Ordovician)—Foliated granite

STRATIFIED METAMORPHIC ROCKS

Dg	Gile Mountain Formation (Lower Devonian)—Mainly gray pelitic schist or slate and metagraywacke

Piermont Allochthon

Dl	Littleton Formation (Lower Devonian)—Mainly gray pelitic schist
Sm	Madrid Formation (Upper Silurian(?))—Gray, slightly calcareous metasandstone and calcareous schist
Ssf	Smalls Falls Formation (Upper Silurian)—Black sulfidic schist and minor quartzite
Spq Spv	Perry Mountain Formation (Silurian)—Spq, sharply interbedded muscovitic schist and quartzite; Spv, volcanic-bearing member, muscovitic and chloritic schist interbedded with variably feldspathic quartzite and dark- to light-colored ash metatuff
	Rangeley Formation (Lower Silurian)
Srbc	Members B and C, undivided. Interbedded dark-gray pelitic schist, variably feldspathic quartzite, and sparse thin beds of quartz-granule conglomerate
Srcc	Interbedded quartz-pebble conglomerate and dark-gray pelitic schist mapped locally near lower contact of Member C
Sra	Member A. Lenses of poorly sorted, schist-matrix, polymictic pebble-cobble conglomerate and sedimentary breccia; metasedimentary and metafelsite clasts
Sg	Greenvale Cove Formation (Lower Silurian(?))—Purplish-gray, laminated, garnetiferous muscovite-biotite-quartz-plagioclase schist and calc-silicate rock
Oq	Quimby Formation (Upper Ordovician(?))—Black sulfidic schist and metagraywacke

Bronson Hill Sequence (Autochthonous)

Dl Dlv	Little Formation (Lower Devonian)—Dl, mainly gray pelitic schist or slate; Dlv, volcanic member, mainly felsic metatuff; commonly interbedded with gray pelitic schist or slate
Sf	Fitch Formation (Upper Silurian)—Calcareous metasandstone and schist, impure metalimestone, and calc-silicate rock
Sc	Clough Quartzite (Lower Silurian)—Quartzite, quartz conglomerate, and quartz-mica schist
Op	Partridge Formation (Middle Ordovician)—Gray to black sulfidic schist and metagraywacke
Oau Oal	Ammonoosuc Volcanics (Middle Ordovician)—Upper member(?) (Oau), felsic to mafic metavolcanic rocks; Lower member (Oal), basaltic amphibolite or greenstone and subordinate felsic metavolcanic rocks and associated metasedimentary rocks

EXPLANATION OF MAP SYMBOLS

————?—	Contact—Queried where conjectural in section
——•——	Postmetamorphic normal fault—Bar and ball on downthrown side
—————	Premetamorphic unclassified fault
—▲—▲—▲—	Premetamorphic sole fault (Early Acadian)—Sawteeth on side of inferred upper plate. Foster Hill sole fault (FHS) interpreted as sole of Piermont allochthon; Indian Brook sole fault (IBS) interpreted as sole of Bean Brook slice

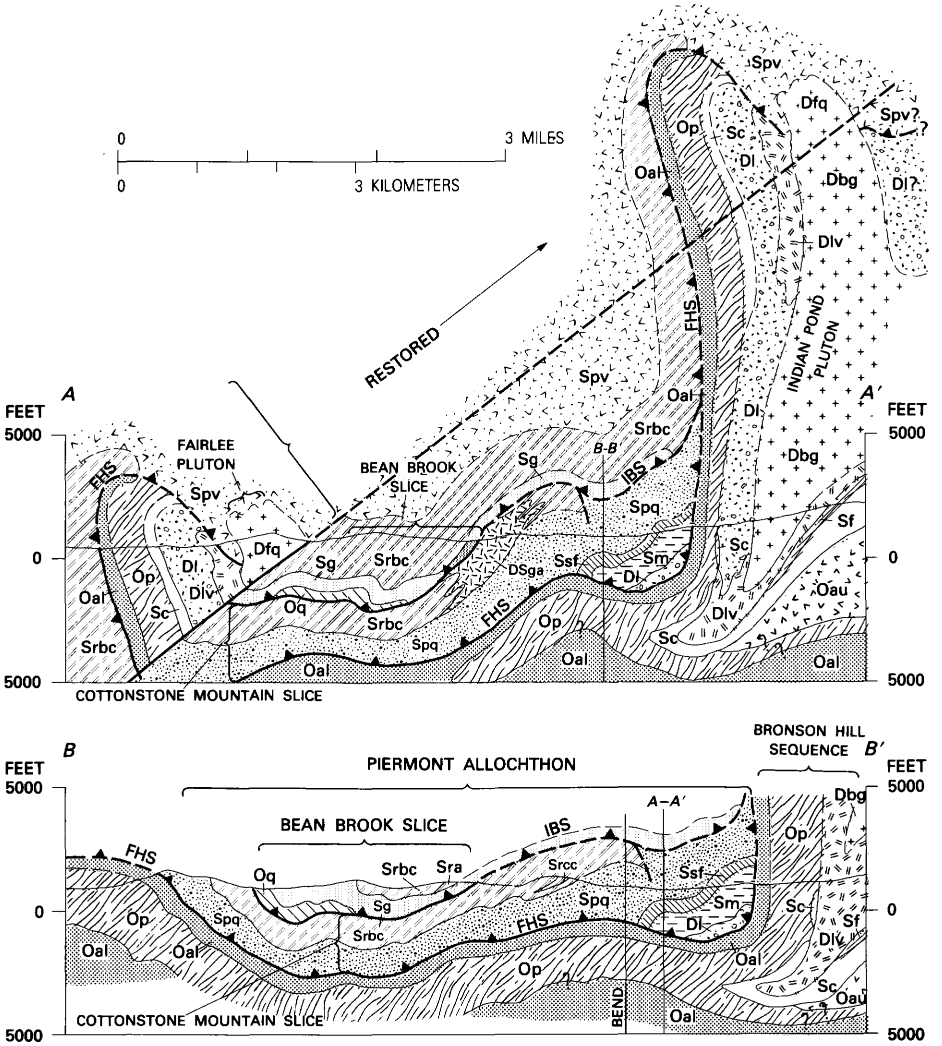


Figure 3. Interpretive cross sections showing the structure of the Bronson Hill sequence and the Piermont allochthon in the area of figure 2. Explanations of rock units and structural features are on figure 2.

that constitute the Piermont allochthon. Hadley (1942, 1950) mapped all of these formations as the Ordovician Albee Formation, a stratigraphic name that is not used in this paper. On figure 2, rocks originally mapped as Albee are reassigned to stratigraphic equivalents of the Quimby (Upper Ordovician(?)), Greenvale Cove (Lower Silurian(?)), Rangeley (Lower Silurian), Perry Mountain (Silurian), Smalls Falls (Upper Silurian), Madrid (Upper Silurian(?)), and Littleton (Lower Devonian) Formations, whose type localities, except the Littleton, are near Rangeley, Maine (fig. 1; Moench and Boudette, 1987). Although the Rangeley Formation of the allochthon is much thinner than the type Rangeley (1,000 m versus 3,000 m, respectively), conglomeratic rocks of members A and C of the type Rangeley are clearly mappable within the allochthon. Hadley (1942) designated a rusty-weathering, dark-colored, conglomerate-bearing unit to be the Piermont Member of the Albee Formation, which on figure 2 approximately corresponds to the Rangeley Formation. Like member C of the type Rangeley Formation (Moench and Boudette, 1987), the upper part of the Rangeley of the allochthon locally contains interbedded schist and quartz conglomerate (fig. 2, Srcc) considered to be a marine basin facies of the Clough Quartzite of the Bronson Hill sequence.

The contacts between all formations of the Piermont allochthon are conformable and either sharp or abruptly gradational. The sediments thus appear to have accumulated without break from Late Ordovician(?) to Early Devonian time. With the exception of carbonate-bearing rocks in the Greenvale Cove and Madrid Formations, conglomerates in the Rangeley Formation, and volcanics in the Perry Mountain Formation, these rocks are mainly siliciclastic turbidites. Environments of deposition ranged from strongly or moderately euxinic (Quimby, Rangeley, and Smalls Falls Formations) to well aerated (Greenvale Cove, Perry Mountain, and Madrid Formations).

Within the allochthon, only the uppermost 100 m of the Upper Ordovician(?) Quimby Formation are preserved. The Quimby contains volcaniclastic metagraywacke, probably derived from contemporaneous and older Ordovician

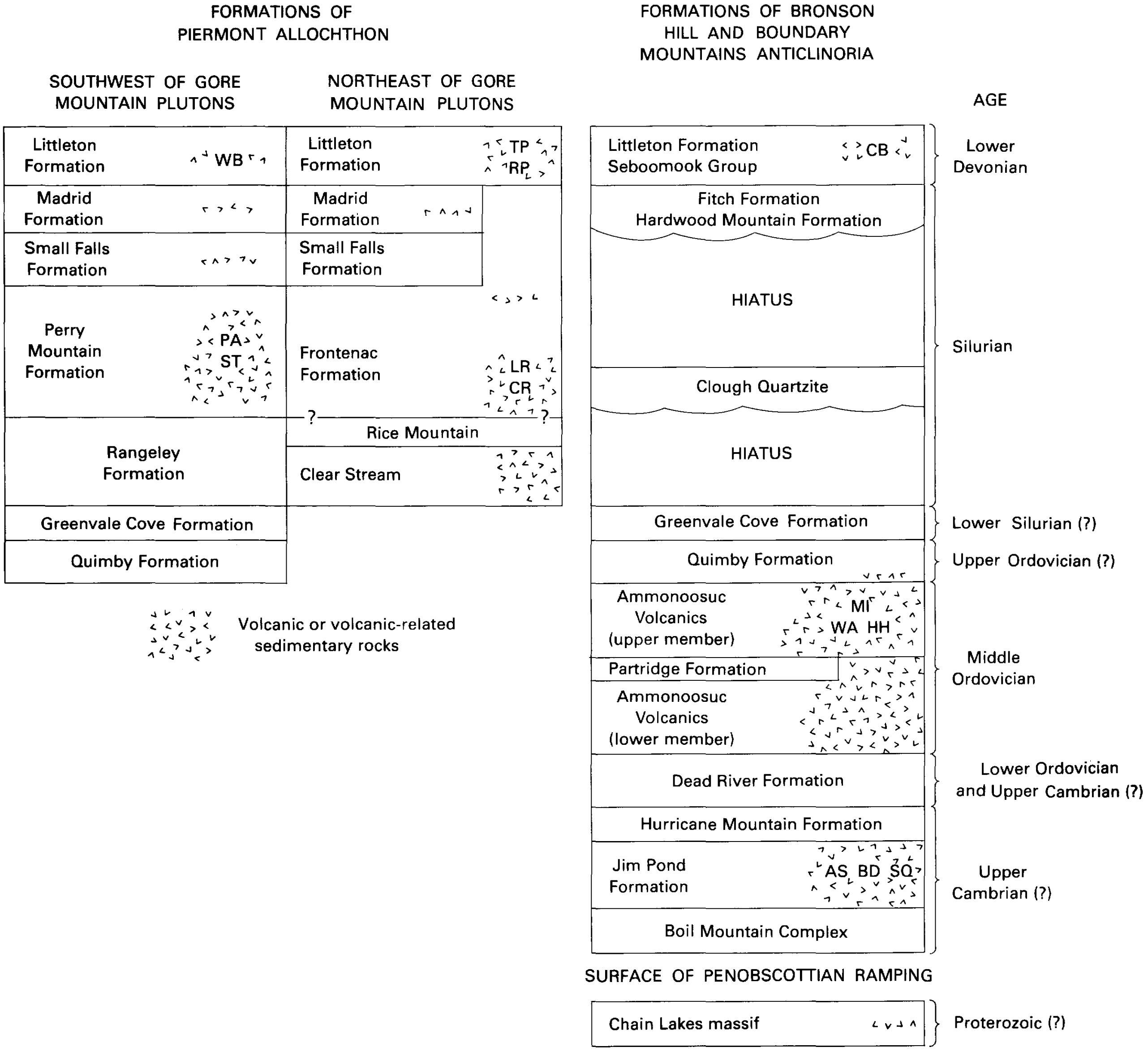


Figure 4. Stratigraphic columns for the Piermont allochthon and parts of the Bronson Hill and Boundary Mountains anticlinoria and selected named volcanogenic sulfide deposits. Deposit names: WB, Washburn Brook; PA, Paddock; ST, Stevens; TP, Thrasher Peaks; RP, Rump Pond; LR, Ledge Ridge; CR, Clinton River district; CB, Coppermine Brook; MI, Milan; WA, Warren; HH, Hampshire Hills; AS, Alder Stream; BD, Border; SQ, Squirtgun flowage (northeast of area).

volcanics that shed into a deep euxinic basin. As in western Maine, the Lower Silurian(?) Greenvale Cove Formation of the allochthon is about 200 m thick. It is interpreted to represent the first sediments shed from an emerging western Silurian source area (Boucot, 1968, fig. 6–1). Chemical data for the type Greenvale Cove (for example, high sodium, boron, and phosphorus) suggest hypersaline conditions. Because the Greenvale Cove Formation of the allochthon locally contains conglomerate not found in the western Maine Greenvale Cove, it accumulated closer to the source than that of western Maine. The conglomerate contains deformed, rounded cobbles of metasandstone in a matrix of finer grained material of similar composition.

Because it is so thin relative to the type Rangeley Formation of western Maine and because it lacks evidence of shoreline or nearshore conditions, the Rangeley of the allochthon is inferred to represent a transitional platform facies that accumulated in water of moderate depth along or just northwest of a Silurian tectonic hinge. The hinge is the approximate axis of abrupt southeastward thickening of Silurian deposits—particularly the polymictic conglomerate-rich members A and B of the Rangeley

Formation (Moench and Boudette, 1987)—along the northwestern margin of the ancestral basin of the Kearsarge-central Maine trough (fig. 1; also see Hatch and others, 1983, and Moench and Pankiwskyj, 1988).

The Silurian Perry Mountain Formation is about 600 m thick in both the Piermont area of the allochthon and western Maine; it seems to be somewhat thicker farther north in the allochthon, where it contains more abundant metavolcanic rocks. The variably lenticular to planar bedding in the Perry Mountain of the allochthon suggests that it was deposited in a shallower marine environment (perhaps near wave base) than the Perry Mountain of western Maine, which has uniform planar bedding. In western Maine, the strongly euxinic Upper Silurian Smalls Falls Formation has a maximum thickness of 800 m; in the Piermont area of the allochthon, it has a maximum thickness of about 100 m, where it is inferred to represent the western margin of the euxinic basin. Farther north, rocks assigned to the Smalls Falls range from about 40 m to perhaps as much as 100 m thick. Although poorly exposed, the calcareous, Upper Silurian(?) Madrid Formation of the allochthon near Piermont has a maximum thickness of about 200 m, somewhat thinner than in western Maine.

Bronson Hill Sequence

Formations of the Piermont allochthon are sharply juxtaposed by the Foster Hill sole fault (fig. 2, FHS) against northeast-trending belts of five autochthonous formations (plus members) of the Bronson Hill sequence: the mainly basaltic facies of the lower member of the Ammonoosuc Volcanics (Middle Ordovician), the Partridge Formation (Middle Ordovician), the Clough Quartzite (Lower Silurian), the Fitch Formation (Upper Silurian), and the Littleton Formation (Lower Devonian).

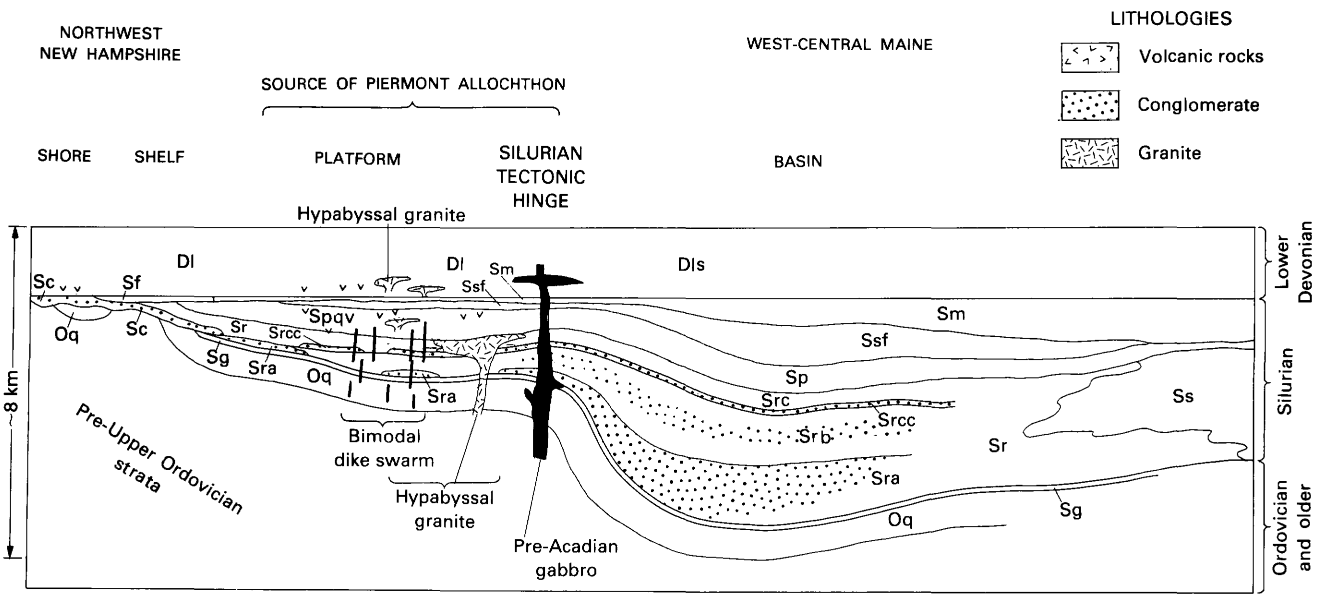
In earlier publications (Moench and others, 1987; Moench, 1989), I assigned the Ammonoosuc (lower member) and Partridge of the area of figure 2 respectively to the Ammonoosuc (upper member) and Quimby Formation (Upper Ordovician(?)). The assignments used in this paper are based on mapping in 1989, when I traced the belt of predominantly basaltic rocks at Sunday Mountain (fig. 2, Oal) directly into previously mapped rocks of the lower member of the Ammonoosuc Volcanics, 20 km to the north. There, the lower member is stratigraphically overlain by black schist of the Partridge Formation, which is succeeded by felsic and mixed metavolcanic rocks of the upper member of the Ammonoosuc Volcanics. About 15 km northeast of the area of figure 2, the upper member is overlain by black sulfidic schist and metagraywacke assigned to the Quimby Formation, which differs only in minor ways from the Partridge Formation. The Ammonoosuc Volcanics, widely regarded as the volcanic products of an Ordovician island arc, is a bimodal assemblage of subaqueous volcanic rocks composed of metamorphosed tholeiitic basaltic greenstone or amphibolite and low-potassium keratophyre (Aleinikoff, 1977; Leo, 1985). Whereas predominantly basaltic rocks of the lower member (Oal) are exposed west of the Indian Pond pluton (fig. 1, IP), mixed felsic and basaltic rocks (Oau) probably of the upper member are exposed to the east. The Partridge and Quimby Formations are euxinic sequences of black sulfidic phyllite and volcanic-derived metagraywacke. North of the area of figure 2, thinly bedded metachert, garnet-rich silicate iron formation, and other rocks of exhalative origin are common along the lower contact of the Partridge.

About 60 km northeast of the area of figure 2, the Ammonoosuc Volcanics are underlain by interbedded schist and quartzite assigned to the Dead River Formation (Lower Ordovician and Upper Cambrian(?)). The Dead River, an early Paleozoic flysch sequence, is remarkably similar to part of the Perry Mountain Formation of the Piermont allochthon, and in previous papers (Moench and others, 1987; Moench, 1989), I apparently incorrectly assigned some rocks to the Dead River that belong in the Perry Mountain.

Pre-Silurian rocks of the Bronson Hill sequence are unconformably overlain by the Lower Silurian Clough Quartzite, which is succeeded by calcareous rocks of the Upper Silurian Fitch Formation exposed mainly east of the Indian Pond pluton but also locally exposed to the west. Although the lower contact of the Clough marks a major regional unconformity west of the Silurian tectonic hinge (fig. 1), near and southeast of the hinge, coeval quartz conglomerates occur conformably with a thick Silurian basin sequence as shown schematically on figure 5. The lenticular bodies of Clough exposed west of the Indian Pond pluton (fig. 2) may have filled stream channels west of the inferred Early Silurian shore. Latest Silurian shelf sedimentation, marked by deposition of the calcareous Fitch Formation, was followed by general subsidence in Early Devonian time when the entire region was buried by muddy sediments and local volcanic rocks of the Littleton Formation.

Intrusive Rocks

Plutonic rocks mapped and described by Hadley (1942, 1950) and White and Billings (1951) include the Ordovician Baker Pond Gneiss and the Devonian Fairlee Quartz Monzonite, Bethlehem Gneiss (all foliated), and Haverhill Granodiorite (mainly unfoliated). The Fairlee pluton (fig. 1, F) and the Mount Clough pluton of Bethlehem Gneiss (fig. 1, MC) to the east have been dated at 410±5 Ma by the U-Pb zircon method (Aleinikoff and Moench, 1987; Moench and Aleinikoff, 1987; reported ages later revised by J.N. Aleinikoff, oral commun., 1989). Metamorphosed sheeted gabbro and diabase at Peaked Mountain (fig. 2) are shown by Hadley (1942) to be in an area where basic dikes and sills are abundant in the Albee



EXPLANATION

Dlc	Littleton Formation and Seboomook Group (Lower Devonian)—Mainly southeast-derived, gray mud-silt-sand turbidites
Dl	Littleton Formation (Lower Devonian)—Locally includes volcanic facies
Sf	Fitch Formation (Upper Silurian)—Fossiliferous, calcareous shelf sediments
Sm	Madrid Formation (Upper Silurian(?))—Calcareous, feldspathic sand turbidites; locally contains volcanic rocks west of Silurian tectonic hinge
Ssf	Smalls Falls Formation (Silurian)—Black sulfidic shale and northwest-derived, quartz-feldspar sand turbidites; local calcareous rocks; locally contains volcanic rocks west of Silurian tectonic hinge
Sp	Perry Mountain Formation (Silurian)—Green shale and northwest-derived, quartz-feldspar sand turbidites east of Silurian tectonic hinge
Spqv	Perry Mountain Formation (Silurian)—Undivided volcanic-bearing facies and green shale-sandstone facies west of Silurian tectonic hinge
Sc	Clough Quartzite (Lower Silurian)—Transgressive, nearshore and onshore, quartz-rich sand and gravel
Sr	Rangeley Formation (Lower Silurian)—Dark-gray, sulfidic shale and northwest-derived quartz-feldspar sand and conglomeratic turbidites, gravel flows, and debris flows
Src	Member C. Approximately 600 m of turbidites, in lower part interbedded with quartz-pebble conglomerate
Srcc	Basin-facies quartz conglomerate; coeval with Clough Quartzite
Srb	Member B. Thick (to 1,200 m) proximal turbidites and medial quartz-rich polymictic conglomerate southeast of Silurian tectonic hinge
Sra	Member A. Thick (to 1,200 m) subaqueous polymictic fanglomerate southeast of Silurian tectonic hinge; thin (few meters) lenticular, polymictic debris-flow deposits west of hinge
Ss	Sangerville Formation (Silurian)—Green and gray shale, volcaniclastic sand and granule-conglomeratic turbidites, and calcareous flysch
Sg	Greenvale Cove Formation (Lower Silurian(?))—Northwest-derived calcareous sand and silt
Oq	Quimby Formation (Upper Ordovician(?))—Gray to black, sulfidic shale and volcaniclastic graywacke turbidites

Figure 5. Schematic section showing facies relations among Silurian rocks from northwestern New Hampshire to west-central Maine prior to the Acadian orogeny. Modified from Hatch and others (1983) and Moench and Pankiwskyj (1988).

Formation. This body and related mafic dikes and sills, which may be Silurian or Devonian in age, are considered to be part of the Piermont allochthon. The small bodies of talc schist within the volcanic rocks at Sunday Mountain (fig. 2, U) might be ultramafic intrusions strung out along the Foster Hill sole fault (fig. 2, FHS). They have not been proven to be intrusive, however, and they might be products of sea-floor alteration and magnesium enrichment.

Structure

As depicted in the cross sections (fig. 3), the structure of the Piermont area consists of the Piermont allochthon and an underlying, tentatively inferred east-facing, synclinal nappe of autochthonous formations of the Bronson Hill sequence. The existence of the nappe is based on the correlation of predominantly basaltic rocks assigned to the lower member of the Ammonoosuc Volcanics (fig. 2, Oal) south of Orfordville and at Sunday Mountain. The steeply dipping portion of the inferred nappe is the Salmon Hole Brook syncline, which was originally mapped by Billings (1935) more than 30 km to the northeast. The Indian Pond pluton of Bethlehem Gneiss (fig. 1, IP) is here interpreted as a doubly plunging synclinal remnant of a folded granitic sheet within the nappe. The northern end of the pluton, according to my interpretation, rises northward above ground along the trace of the syncline and then descends to join the Haverhill pluton of Bethlehem Gneiss (fig. 1, H), which lies along the trace of the Salmon Hole Brook syncline in the Moosilauke quadrangle (fig. 2; Billings, 1935). To the east, I infer that the same granitic sheet arches over the Bronson Hill anticlinorium where it reappears as the Mount Clough pluton of Bethlehem Gneiss (fig. 1, MC).

The present warped configuration of the Piermont allochthon and of the postulated underlying nappe is a result of later deformation associated with the formation of the Bronson Hill anticlinorium and the Orfordville antiform (fig. 2). The Orfordville antiform is equivalent to the Orfordville anticlinorium of Spear and Rumble (1986), which extends at least 30 km to the southwest. I prefer the term antiform, however, because the feature is much smaller than the Bronson Hill anticlinorium and because the oldest stratified rocks do not occur consistently along the axial trace.

The Ammonoosuc fault (figs. 1, 2) juxtaposes predominantly greenschist-facies rocks west of the fault against predominantly amphibolite-facies rocks to the east. Brittle features and silicification are characteristic of the fault, and dips of 35° to 40° west are reported (Billings, 1937; Hadley, 1942; White and Billings, 1951). Although originally mapped as a thrust fault, I interpret the Ammonoosuc fault as a normal fault, downthrown on the west. It is one of a family of major normal faults that mark, or are parallel to, the eastern side of the Triassic basin of southern New England. About 80 km northeast of the area of figure 2, the Ammonoosuc fault is cut by the Early Jurassic alkalic Pilot-Pliny plutons. If an average dip of 37° can be assumed above ground, restoration of the rocks exposed west of the fault to a reasonable prefault position to the east yields a vertical displacement of about 5 km (fig. 3, section *A–A′*). This interpretation is viable because of the new stratigraphic assignments on both sides of the fault and because of isotopic age data that permit correlation of the Fairlee and Indian Pond plutons (Aleinikoff and Moench, 1987; Moench and Aleinikoff, 1987).

Formations of the Piermont allochthon are juxtaposed against the Bronson Hill sequence along the Foster Hill sole fault (fig. 3). This fault, first recognized in the Foster Hill area 10.5 km west of Littleton, N.H. (Moench and others, 1987), is inferred to mark the regional sole of the Piermont allochthon. The fault is knife sharp at all localities and has been deformed and metamorphosed along with the stratified rocks on both sides. It is no more conspicuous than the premetamorphic faults of western Maine (see descriptions in Moench, 1970) and is likewise interpreted as a surface of almost no friction. If it were not for the apparent regional discordance of this boundary, the Foster Hill fault at most localities probably would be mapped as a normal stratigraphic contact.

The Piermont allochthon is composed of two major fault slices, the Cottonstone Mountain slice and the Bean Brook slice (figs. 2, 3). Structural data and stratigraphic identifications indicate that the Cottonstone Mountain slice is composed of an inverted sequence of formations (Rangeley member C, Perry Mountain, Smalls Falls, Madrid, Littleton), whereas the overlying Bean Brook slice contains an upright sequence of older formations (Quimby, Greenvale Cove, Rangeley members A and B). The Indian Brook sole fault (figs. 2, 3, IBS), which has not been seen in outcrop, is at the base of the Bean Brook slice. There is no proof that the allochthon divides into two or more slices farther north.

STRATIGRAPHIC RELATIONS TO THE NORTH

Between Piermont area (fig. 2) and Gore Mountain plutons (fig. 1, GM)

The Piermont allochthon of this area is bounded on the west by the Monroe fault (fig. 1, MNF), described by Hatch (1988), and on the east by the Foster Hill sole fault (fig. 1, FHS), except where this fault is truncated by the Connary Brook fault (fig. 1, CBF). Tentatively thought to be included in the allochthon of this area are two small bodies of black sulfidic phyllite and bimodal metavolcanic rock, in two faulted synclines just south of the Gore Mountain plutons, that are tentatively correlated with isotopically dated rocks at Rice Mountain about 10 km northwest of Errol, N.H. The lower contact of these bodies is marked by the Rice Mountain line (fig. 1, RML), which may be either a sole fault or an unconformity.

Between the Piermont area and the Gore Mountain plutons, the Piermont allochthon is composed largely of facies of the Rangeley and Perry Mountain Formations; the Quimby and Greenvale Cove Formations have not been found, but small remnants of the Smalls Falls, Madrid, and

Littleton Formations are present. As shown in figure 4, metavolcanic rocks occur in all of the recognized Silurian units of the area, except the Rangeley and Greenvale Cove Formations, but only locally in amounts that exceed 20 percent of the exposed sequence. Margeson (1982) and Hafner-Douglass (1986) have described some of these volcanics in detail, in rocks that I call the volcanic-bearing facies of the Perry Mountain Formation. In this facies, sharply interbedded, variably feldspathic, commonly chalky-weathering metasandstone and green or greenish-gray phyllite are interspersed with well-stratified keratophyric ash- and lapilli-metatuff, locally abundant basaltic flows and tuffs, now greenstone or amphibolite, and scattered small bodies of fine-grained felsic metaporphyry and metagranophyre. Uranium-lead isotopic age data already cited are consistent with the assigned Silurian age of the volcanic rocks in the Perry Mountain Formation. Also included in the volcanic-bearing facies are rocks characterized by abundant bleached, pyritic, quartz-muscovite-feldspar schist (some with quartz "eyes") and interpreted as metamorphosed and hydrothermally altered sediments and rhyolite tuff. This assemblage was previously mapped as Ammonoosuc Volcanics (Billings, 1935; Hall, 1959) that occurs in an area that lies astride long. 72°00′ just east of the Monroe fault (fig. 1). A basalt-rich facies also may be mapped that contains similar rocks and as much as 50 percent massive basaltic greenstone.

Recognized Smalls Falls Formation exposed at the southern end of Foster Hill (fig. 1) has a measured thickness of about 40 m. Here, the Smalls Falls lies conformably above and below rocks assigned to the Perry Mountain and Madrid Formations, respectively. These units occur in a southeast-younging sequence that is juxtaposed against a northwest-younging sequence within the Ammonoosuc Volcanics. This is the type locality of the Foster Hill sole fault (Moench and others, 1987, stop 8), which has been located within 1 m but is not exposed. At this locality, the Smalls Falls is composed of black pyritic or pyrrhotitic phyllite, sparse thin beds of feldspathic metasandstone, and sparse felsic metatuff. An outcrop of Smalls Falls, 8 km north of Piermont village, contains several thick beds of quartz-feldspar crystal metatuff.

The Madrid Formation at Foster Hill forms a faulted synclinal belt, about 30 m wide, on the northwestern or inferred hanging wall side of the Foster Hill fault. Here, the Madrid is weakly calcareous and is composed of (1) brown-weathering slate and interlaminated slate and metasiltstone; (2) brown-weathering, rather friable, probably tuffaceous metasandstone that is locally interbedded with white calc-silicate rock; and (3) calcareous greenstone, locally having deformed quartz-filled amygdules.

A large synclinal body of Littleton Formation is exposed within the allochthon about 10 km west of Groveton, N.H. (fig. 1). It is composed of typical Littleton gray phyllite but contains abundant laminations of fine-grained manganiferous garnet (coticule) and disseminated magnetite. The gray phyllite envelopes a body of silicate iron formation that hosts a small copper deposit (fig. 6, Washburn Brook). These rocks are interpreted to represent muddy Littleton sediments that have been enriched in iron (±manganese) and locally altered and mineralized by volcanic-exhalative solutions. Identical rocks occur less abundantly in the Littleton Formation northeast of the Gore Mountain plutons but without known copper deposits.

Northeast of Gore Mountain plutons (fig. 1, GM)

Here, as farther south, the inferred northern extension of the Piermont allochthon is juxtaposed on the northwest by the Monroe fault (fig. 1, MNF) against predominantly metasedimentary rocks of the Connecticut Valley trough. The southeastern margin of the allochthon in this area sharply truncates rocks of the Boundary Mountains anticlinorium along the Thrasher Peaks fault and the younger Deer Pond fault (fig. 1, TPF and DPF). The Thrasher Peaks fault may be a northeastern extension of the Foster Hill sole fault that shows increasing evidence of brittle faulting toward the northeast. Near the point of truncation of the Thrasher Peaks fault by the Deer Pond fault, about 24 km southwest of Magalloway Mountain, the Thrasher Peaks fault is a subvertical sharp surface that lacks evidence of cataclasis, although rocks of the northwestern side are nearly vertically rodded. Northwest of Magalloway Mountain, the fault dips 50° northwest, shows abundant evidence of cataclasis, and has slickensides that plunge downdip. In northwestern Maine, at a locality 100 km northeast of Magalloway Mountain, Marvinney (1986) describes evidence of brittle deformation and right-slip displacement of at least 8 km along the Thrasher Peaks fault. This fault has thus experienced complex movement patterns and probably more than one reactivation.

The Boundary Mountains anticlinorium is cored by the possibly Proterozoic Chain Lakes massif, composed mainly of polymetamorphic diamictite of controversial age and origin (see discussions in Boone and Boudette, 1989; Boone and others, 1990; Dunning and Cousineau, 1990; Olszewski and others, 1990). On its southern side the massif is structurally overlain by the Upper Cambrian(?) Boil Mountain Complex and Jim Pond Formation, together thought to represent ophiolite that formed by spreading above a subduction zone (Boudette, 1982; Coish and Rogers, 1987) and ramped over the Chain Lakes massif during the early Paleozoic Penobscottian orogeny (Boone and Boudette, 1989). The Boil Mountain and Jim Pond are succeeded to the south by, and partly incorporated within, a euxinic mélange of the Upper Cambrian(?) Hurricane Mountain Formation, which is conformably overlain farther south by flysch turbidites of the Lower Ordovician and Upper Cambrian(?) Dead River Formation (fig. 4). The

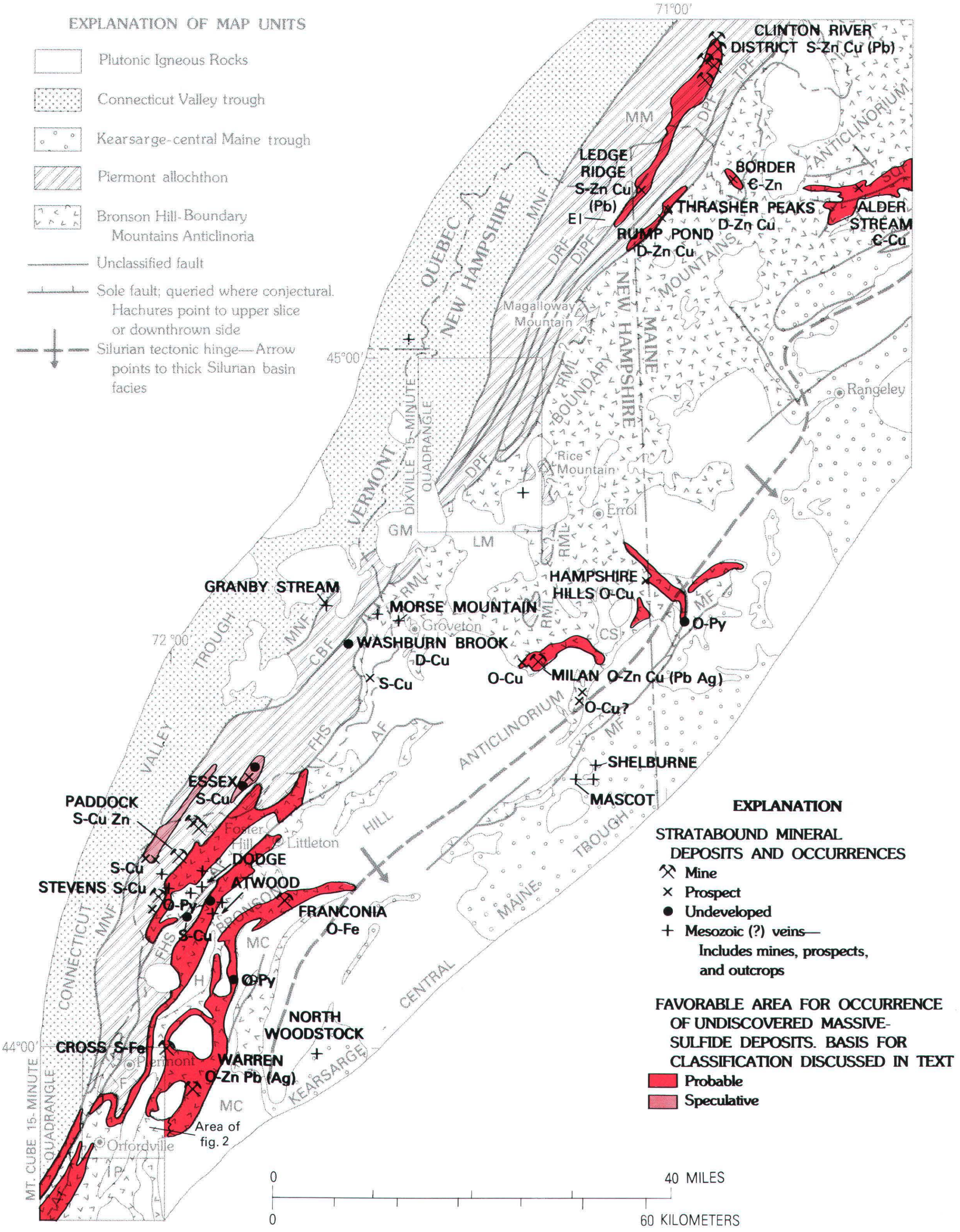


Figure 6. Major tectonic features of northern New Hampshire and adjacent areas, locations of stratabound and vein-type mineral deposits, and favorable areas for the occurrence of undiscovered volcanogenic sulfide deposits. Named mineral deposits are cited in text. Age of stratabound mineral deposit: D, Early Devonian; S, Silurian; O, Middle Ordovician; Є, Late Cambrian(?). Composition of mineral deposit: Cu, copper dominant; ZnCu(Pb), zinc-copper with lead; Zn-Cu(PbAg), zinc-copper with lead and silver; ZnCu, zinc-copper without lead; ZnPb, zinc-lead; ZnPb(Ag), zinc-lead with silver; Fe, quartz-magnetite (hematite) iron formation; Py, mainly pyrite or pyrrhotite. Named faults: AF, Ammonoosuc fault; CBF, Connary Brook fault; DPF, Deer Pond fault; DRF, Dead Water Ridge fault; DiPF, Diamond Pond fault; FHS, Foster Hill sole fault; MF, Mahoosuc fault; MNF, Monroe fault; RML, Rice Mountain line; SQF, Squirtgun fault; TPF, Thrasher Peaks fault. Plutons referred to in text (north to south): MM, sheeted intrusions at Marble Mountain; EI, East Inlet; GM, Gore Mountain; LM, Long Mountain; CS, Chickwolnepy; MC, Mount Clough; H, Haverhill; F, Fairlee; IP, Indian Pond.

Middle Ordovician Ammonoosuc Volcanics and Partridge Formation, which are succeeded by the Upper Ordovician(?) Quimby and Lower Silurian(?) Greenvale Cove Formations (fig. 4), lie above the Dead River Formation south of the axial zone of the Boundary Mountains anticlinorium, near Rangeley, Maine, and at the northeastern end of the Bronson Hill anticlinorium (fig. 1). The Jim Pond-Dead River sequence of the Boundary Mountains anticlinorium is intruded by Late Ordovician granitic plutons and is unconformably overlain by thin, transgressive quartzite and quartz conglomerate similar to the Lower Silurian Clough Quartzite and by thin, fossiliferous, calcareous shelf deposits of the Upper Silurian Fitch(?) and Hardwood Mountain Formations (fig. 4). These units are succeeded by thick Lower Devonian muddy sediments, arenites, and local volcanics of the Littleton Formation and Seboomook Group (fig. 4).

According to my interpretation, the extension of the Piermont allochthon northeast of the Gore Mountain plutons includes all of the rocks mapped as Frontenac Formation in southern Quebec (McGerrigle, 1935; Marleau, 1968), northernmost New Hampshire, and adjacent Maine (Green, 1968), and all of the rocks mapped as Kidderville Formation by Hatch (1963) in the Dixville quadrangle, about 20 km northwest of Errol, N.H. (fig. 1). The Frontenac Formation is a complex assemblage of well-stratified, pale-green or greenish-gray phyllite, tuffaceous metasandstone, basaltic greenstone (commonly pillowed), and locally thick felsic metavolcanic rocks. A major bimodal volcanic center in the Frontenac hosts the polymetallic sulfide deposits of the Clinton River district and at Ledge Ridge. Extensive metadiabasic sills and longitudinal dikes also intrude the Frontenac Formation, some or all of which may be comagmatic with the mafic volcanic rocks of the Frontenac (Ebinger, 1985; W.A. Bothner and R.H. Moench, unpub. chemical data). The basaltic and intrusive greenstones consistently have iron- and titanium-rich compositions (negatively correlated with magnesium) that are characteristic of basalts erupted in regions undergoing tectonic extension (Eisenberg, 1982, 1983; Chevé and others, 1983; Ebinger, 1985; Marvinney, 1986; W.A. Bothner and R.H. Moench, unpub. data).

Recent mapping by the author has shown that the Kidderville Formation of Hatch (1963) in the Dixville quadrangle is separable into three fault-bounded belts:

(1) A western belt, between the Dead Water Ridge and Diamond Pond faults (fig. 1, DRF and DiPF), of locally pillowed metabasalt that to the north is intruded by the epizonal granitic (to dioritic on the western side) East Inlet pluton (fig. 1, EI), dated at 430±4 Ma by the U-Pb zircon method (Eisenberg, 1982; Lyons and others, 1986). West of the pluton, metabasalt of this belt is conformably overlain by interbedded green phyllite and tuffaceous metasandstone of the Frontenac Formation. Available U-Pb isotopic data, discussed below, permit the inference that the East Inlet is a feeder to felsic metavolcanic rocks of the Frontenac Formation.

(2) A central belt, between the Diamond Pond and Deer Pond faults (fig. 1, DiPF and DPF), that is separable into a probably west-younging sequence composed of (ascending stratigraphically from east to west): (a) rusty-weathering interbedded dark-gray schist and quartzite resembling the Rangeley Formation, (b) interbedded greenish-gray schist, feldspathic metasandstone, and minor metabasalt resembling much of the volcanic-bearing facies of the Perry Mountain Formation, and (c) rocks similar to those of b, but containing abundant metabasalt and less abundant schistose felsic metatuff. I interpret b and c as distal facies of the proximal bimodal metavolcanic rocks that host the massive sulfide deposits farther north at Ledge Ridge and in the Clinton River district (fig. 6). A felsic dike that intrudes b in the Dixville quadrangle and felsic metatuff exposed about 2 km east of the East Inlet pluton have yielded identical U-Pb zircon ages of 418±4 Ma (Lyons and others, 1983, 1986; data later recalculated by J.N. Aleinikoff, written commun., 1988). Felsic metatuff from the same belt, but perhaps 1 km downsection, has yielded a U-Pb zircon age of 432±10 Ma (previously reported as 414±4 Ma by Aleinikoff and Moench, 1985; revised age based on recalculations and new data by J.N. Aleinikoff). The felsic dikes are thus reasonably considered to be comagmatic with felsic metavolcanic rocks of the Frontenac Formation, and the data indicate that the felsic and bimodal volcanic rocks of this belt are Silurian in age. Fragmental metarhyolite from the same belt about 3 km northeast of the northern tip of the East Inlet pluton yielded U-Pb zircon data that favor a Late Silurian age, certainly no older than 430 Ma (J.N. Aleinikoff, written commun., 1988).

(3) An eastern belt, between the Deer Pond and Thrasher Peaks faults (fig. 1, DPF and TPF), of gray slate, characteristic of the Littleton Formation but containing abundant thick-stratified felsic crystal-metatuff lapilli-metatuff, magnetite-bearing gray phyllite and iron formation (similar to the Littleton west of Groveton) and richly pyritic, black phyllite, all of which are assigned to the Littleton Formation. At Mud Pond Ridge (near the eastern boundary of the Dixville quadrangle), dark-gray phyllite of the Littleton Formation is conformably underlain by thickly bedded, medium-gray, fine-grained metasandstone and calc-silicate rocks characteristic of the Madrid Formation of western Maine, but also containing schistose felsic metatuff, amphibolite, and a distinctive layer of white felsite. Although stratigraphic relations are unclear, these rocks appear to be underlain by laminated sulfidic calc-silicate rocks and black sulfidic schist that resemble the upper part of the Smalls Falls Formation of western Maine.

East of the Thrasher Peaks fault in the Dixville quadrangle is a peculiar black, sulfidic schist-matrix, quartzite-chip breccia that extends at least 20 km along the presumed footwall of the Thrasher Peaks fault. The same breccia also occurs along the footwall of the Foster Hill fault south of the plutons, west of Groveton, N.H., where it extends into footwall rocks of possible Dead River Formation. This enigmatic breccia may be a slide scar that formed before or during emplacement of the Piermont allochthon. Near Magalloway Mountain, the breccia grades eastward away from the Thrasher Peaks fault to interbedded coaly-black, sulfide-rich phyllite and feldspathic metasandstone and local feldspar-rich grit that intertongues with felsic metatuff, pillowed metabasalt, and thinly bedded, pyritic tuffaceous argillite. Although the black sulfidic rocks of this assemblage are sufficiently sulfidic and graphitic to qualify as Smalls Falls Formation, the whole assemblage is here tentatively correlated with somewhat older Silurian rocks at Rice Mountain (fig. 1), described below.

The Littleton Formation of belt 3 extends into northwestern Maine, where gray slate of this same belt is recognized as the Ironbound Mountain Formation of Marvinney (1986), which is the basal unit of the Seboomook Group of Pollock (1987), and has been shown to be conformably underlain by basaltic greenstone and metasedimentary rocks of the Frontenac Formation (Marvinney, 1986). Assuming that an Early Devonian age applies to the whole belt of gray slate into northwestern Maine, a Late Silurian age is implied for at least the uppermost rocks of the Frontenac Formation, in accord with U-Pb zircon data, already cited, that indicate a Silurian age for the bimodal metavolcanic rocks of the Frontenac Formation of belt 2.

Although the Frontenac Formation is in need of thorough stratigraphic study and additional dating, the relations and isotopic age data presented in the previous paragraphs should dispel notions that the Frontenac and the associated massive sulfide deposits are Early Devonian (Green, 1968; Marleau, 1968), Ordovician (Harwood, 1969; Chevé, 1978; St-Julien and Slivitsky, 1987), or Cambrian (Moench, 1984); they are Silurian.

The Piermont allochthon also contains a Lower Devonian massive-sulfide-bearing sequence. The host rocks of the Thrasher Peaks prospect (fig. 6)—originally mapped within the Middle Ordovician Dixville Formation (Green, 1968; Harwood, 1973) and then within the Upper Cambrian(?) Jim Pond Formation (Moench, 1984; Osberg and others, 1985)—are now assigned to the Lower Devonian Littleton Formation of the Piermont allochthon. The present interpretation is based partly on surface and drill-hole observations by R.A. Cavalero (information furnished by the Boise Cascade Corporation, written commun., 1988). He observed that Seboomook-like gray slate and felsic tuff breccia (originally mapped as Seboomook in the prospect area by Green, 1968, fig. 5) are interstratified with, rather than lying unconformably above, schistose felsic metatuff and basaltic greenstone that host the sulfide deposits. Also on Thrasher Peaks are lenticular bodies of polymictic conglomerate thought by Green (1968) and Harwood (1973) to be Silurian in age but here considered to be Lower Devonian mafic volcanic conglomerate. These bodies are associated with agglomeratic and pillowed greenstone of the sequence; they contain a wide variety of clasts, but basaltic greenstone clasts are most abundant. Other evidence supporting the new interpretation is a U-Pb zircon age of 414±5 Ma obtained by Eisenberg (1982, p. 42, table 3) from a small body of hypabyssal microgranite (intrusive felsite) at Thrasher Peaks (fig. 1). This body is believed to be comagmatic with the metavolcanic host rocks of the Thrasher Peaks prospect; it may be a rhyolite dome. The Thrasher Peaks fault was mapped but not observed on the northwestern side of the Thrasher Peaks sequence and is now placed at a shear zone mapped by Eisenberg (1982, fig. 4) and Harwood (1973) on the southeastern side.

Areas that are bounded on the southeast or outlined by the Rice Mountain line (fig. 1, RML) east of the Long Mountain pluton (fig. 1, LM) are underlain by units called the sulfidic shale and sandstone at Rice Mountain and the conformably underlying volcanic rocks at Clear Stream, which is a few kilometers south of Rice Mountain (fig. 1). Although previously mapped as the Rice Mountain and Clear Stream Members of the Middle Ordovician Dixville Formation, representing the upper two of the three members of the Dixville (Green, 1964), felsic metavolcanic rocks near the upper contact of the Clear Stream unit have yielded an Early Silurian age of 434±8 Ma (Aleinikoff and Moench, 1985). The volcanic rocks at Clear Stream are composed mainly of basaltic amphibolite, locally pillowed and commonly containing abundant siliceous, manganiferous iron formation. Well-stratified polymictic metaconglomerate is exposed in the lower 300 m of the Clear Stream unit, 20 km southwest of Errol, N.H. The dated felsic metavolcanic rocks occur in a thin zone at the top of the Clear Stream unit, 9 km west of Errol. The sulfidic shale and sandstone at Rice Mountain have bedding styles that are remarkably similar to those of much of the Lower Silurian Rangeley Formation of western Maine. At Rice Mountain, the sulfidic shale (schist) is dark gray and only weakly graphitic, but elsewhere (in the belt at Magalloway Mountain and the two bodies bordered by RML, northwest of Groveton) it is coaly black, strongly graphitic, and richly pyritic or pyrrhotitic and is more akin to typical Smalls Falls Formation. Accordingly, the correlations adopted in this paper are under review.

Also problematic are the Dead River Formation and the overlying Dixie Brook Member of Green (1964), where these units lie discordantly below the volcanic rocks at Clear Stream. In large parts of northern New Hampshire, the Dead River-Dixie Brook sequence is identical to the type Perry Mountain-Smalls Falls sequence of western Maine. If these sequences are in fact equivalent, the Piermont allochthon extends 20 to 25 km farther east at the latitudes of Errol and Groveton than shown on figure 1. This would mean, furthermore, that the new eastern extension of the allochthon divides into two slices, as in the Piermont area: a main lower slice composed largely of the Perry Mountain and Smalls Falls Formations and an upper slice composed of slightly older Silurian rocks of the former Clear Stream and Rice Mountain Members of Green (1964) that are coeval with the Rangeley Formation. The Rice Mountain line (fig. 1, RML) would be an older-over-younger overthrust fault.

The lower contact of the Clear Stream unit, where exposed west of Errol, is knife sharp and is discordant relative to underlying schist and quartzite mapped as Dead River Formation and black schist and quartzite of the Dixie Brook Member of Green's (1964) Dixville Formation (Middle Ordovician). This contact may thus be either a fault or an unconformity, and the Clear Stream and Rice Mountain units may or may not be allochthonous.

INTRUSIVE ROCKS

Throughout the known length of the Piermont allochthon, stratified rocks are intruded by swarms of metadiabase dikes and sills, which also occur outside the allochthon in areas underlain by pre-Silurian rocks. Within the allochthon are at least four relatively large bodies of sheeted metagabbro and metadiabase. One is shown at Peaked Mountain on figure 2, near the southern end of the allochthon. The sheeted intrusion at Marble Mountain (fig. 1, MM; Chevé, 1978) contains dikes of hypabyssal felsite as well as metagabbro and metadiabase (E.L. Boudette, oral commun., 1985). The intrusion at Marble Mountain is thought to be approximately coeval with a bimodal dike swarm that occurs between the Diamond Pond and Deer Pond faults in the Dixville quadrangle (fig. 1, DiPF and DPF). One felsic dike of the bimodal swarm and a felsic metatuff in the Frontenac Formation have yielded identical Late Silurian U-Pb zircon ages of 418±4 Ma (Lyons and others, 1983; ages later recalculated by J.N. Aleinikoff) and are almost certainly comagmatic. The dikes and sheeted intrusions, which have not been adequately studied, remain a key problem in establishing the origin of the allochthon (Hafner-Douglass, 1987). Tentatively, the four sheeted bodies, many of the dikes in the allochthon, and the country rocks are considered to be allochthonous.

Within the allochthon, mafic dikes, sills, and sheeted intrusive bodies occur most abundantly in the Perry Mountain and Frontenac Formations. Petrochemical data from the Woodsville area, about 25 km southwest of Littleton, indicate that these rocks have high-iron and high-titanium tholeiitic compositions characteristic of extensional tectonic settings (Hafner-Douglass, 1986, 1987). Metabasalt and metadiabase associated with the Frontenac Formation farther north, on both sides of the international border, have the same chemical characteristics (Eisenberg, 1982, 1983; Chevé and others, 1983; Ebinger, 1985; Marvinney, 1986; W.A. Bothner and R.H. Moench, unpub. data). Ebinger (1985), in particular, observed that many metadiabase sills and dikes appear to have been intruded into poorly lithified Frontenac sediments perhaps contemporaneously with sedimentation and basaltic eruptions.

Isotopic data already cited suggest that the Early Silurian East Inlet pluton (fig. 1, EI), which is fine grained and granophyric, is subvolcanic to the earliest felsic eruptions of tuffs in the Frontenac Formation. More studies are needed, however. If the Frontenac is allochthonous, so probably is the East Inlet pluton. Although the East Inlet is a low-density granite body that intrudes thick, high-density greenstone, it lacks obvious expression in the gravity field (Bothner, 1985) and must be thin in vertical dimension; probably it is rootless.

Other granitic bodies that may be allochthonous include several small plutons of microgranite that intrude the Littleton Formation between the Deer Pond and Thrasher Peaks faults (Green, 1968). Evidence that these rocks are coeval with the felsic volcanics of the Littleton of the allochthon has been presented above. Several small granitic bodies that intrude the Frontenac Formation (the Kidderville Formation of Hatch, 1963) appear to have been emplaced at a somewhat deeper level; at least one is pegmatitic. These bodies may be relatively deep expressions of the Littleton-age microgranites.

Although many problems remain, all of these intrusive rocks are considered to be integral parts of the Piermont allochthon. An extensional setting for emplacement of the mafic dike swarms and sheeted bodies is indicated by their significant width across strike and by their titanium- and iron-rich compositions (at low magnesium contents) characteristic of basaltic rocks in regions of tectonic extension.

INTERPRETATION

I interpret the Piermont allochthon to be a single, fault-bounded lithotectonic belt of continuously deposited Upper Ordovician to Lower Devonian subaqueous sedimentary and volcanic rocks and associated bimodal intrusives that extends at least from the Piermont area to the Gore Mountain plutons in New Hampshire (fig. 1) and perhaps as far northeast as the Caucomgomoc Lake area in northern Maine. The allochthon is characterized by the absence of nearshore Silurian or Lower Devonian deposits that have basal unconformities. Such nearshore deposits occur in the autochthon southeast of the southeastern boundary faults of the allochthon. On the basis of thickness and facies comparisons, I think the allochthonous rocks originally accumulated some 25 to 50 km (present distance) to the southeast of their present position. As shown on figure 5, I propose that the site of deposition was a platform that existed between the Silurian shore and the Silurian tectonic hinge. Although the space that is now available for the platform is only 10 or 15 km wide, in Silurian time the platform, probably containing subbasins, may have been 50 km wide, or wider. Moench (1970) proposed that the ancestral basin of the Kearsarge-central Maine trough southeast of the hinge is extensional in origin. According to my present analysis, extension normal to strike on the platform northwest of the hinge was expressed by the emplacement of mafic and bimodal dike swarms and sheeted intrusive complexes, parallel to strike, and small granitic bodies. This intrusive magmatism probably occurred contemporaneously with sedimentation, bimodal volcanism, and sea-floor mineralization on the postulated platform.

After a wide region was buried by muddy sediments and associated volcanic rocks of the Littleton Formation and the Seboomook Group of Pollock (1987), the proposed platform was greatly compressed during the Acadian orogeny, and its sedimentary-volcanic-intrusive cover was transported northwestward, probably several tens of kilometers. The mechanism of transport and whether or not the allochthon was originally rooted at its source are unsolved problems. During transport, the rocks of the allochthon were complexly folded and internally thrust faulted and may have been ramped over major contemporaneous or earlier formed nappes. The allochthon was emplaced against the northwestern sides of the Bronson Hill and Boundary Mountains anticlinoria before or during intrusion of the Early Devonian (410±5 Ma) syntectonic Fairlee pluton. After the emplacement of the allochthon, the allochthonous and autochthonous rocks were tightly compressed, cleaved, folded again, intruded by many Devonian granitic to gabbroic plutons and metamorphosed during the Acadian orogeny and subsequently displaced by the probably Triassic Ammonoosuc normal fault.

RESOURCE IMPLICATIONS

As shown on figure 6, many sulfide mineral deposits occur in northern New England and southeastern Quebec. On the basis of their form and volcanic host-rock associations, the majority of the deposits can be classified as stratabound and volcanogenic in origin. Such deposits occur both within the autochthonous sequences of the Bronson Hill and Boundary Mountains anticlinoria and within the Piermont allochthon and its proposed northern extension (fig. 4). In the early 1970's, the largest known massive sulfide deposit in the region, Ledge Ridge, was discovered in an area covered by glacial drift by J.S. Cummings, Incorporated. On the basis of 24 core holes, this deposit is computed to contain 3.7 million tons of mineralized rock at average grades of 2.3 percent zinc, 0.95 percent copper, 0.85 percent lead, 0.60 ounce/ton silver, and 0.015 ounce/ton gold (Cummings, 1988, p. 253). Additional holes with intercepts at vertical depths of 800 to 1,470 feet indicate that the deposit contains 5 million tons having about the same metal contents (Cummings, 1988, p. 253–254). Gold values are locally high, as shown by drill-core data obtained by J.F. Slack, in cooperation with the Boise Cascade Corporation (J.F. Slack, written commun., 1987). The Ledge Ridge deposit is hosted by weakly metamorphosed, thick, felsic volcanic rocks that I infer to lie above a thick sequence of pillowed greenstone, all within the Frontenac Formation of the proposed northern extension of the Piermont allochthon. The nature and origin of the southern group of much smaller sulfide deposits within the Piermont allochthon (Stevens, Paddock) has long been uncertain. However, the discovery of the allochthon and the recognition of the widely exposed metavolcanic rocks within it have illuminated the regional setting and origin of all these deposits and have provided an improved basis for resource assessment in the Sherbrooke-Lewiston CUSMAP area.

In addition to the stratabound sulfide deposits, several types of iron formation are present, all but one of which are considered to be volcanogenic. The one exception is at the Cross iron mine, near the junction of the Lewiston and Glens Falls 1° × 2° quadrangles (fig. 6). Here, well-foliated, massive and disseminated specular hematite and minor magnetite and barite occur within the Lower Silurian Clough Quartzite, which elsewhere commonly has iron-rich mineral assemblages developed along bedding planes (Rumble and Dickenson, 1986, p. 39). This deposit may have originated by concentration of oxidized iron during subaerial weathering that preceded and perhaps accompanied deposition of the Clough. The occurrence of barite in such a setting is unusual, however, and suggests the need for further study.

Among the several recognized types of volcanogenic iron formations in this area (fig. 6) are (1) banded, quartz-magnetite layers in the Ammonoosuc Volcanics, as at the Franconia iron mine (Hitchcock, 1878, p. 56–57; Annis, 1982); (2) a conformable 10-cm layer of massive pyrite in well-stratified felsic metavolcanic rocks of the Ammonoosuc Volcanics about 9 km southeast of the Hampshire Hills prospect (Moench, 1979, field observations); and (3) a body several tens of meters thick of quartz-chlorite-magnetite rock (±garnet and staurolite) that hosts a stratabound pyrrhotite-chalcopyrite deposit at the Hampshire Hills prospect, which is in an epiclastic facies of the Ammonoosuc Volcanics (Pyke, 1985), and similar rock at the Washburn Brook copper occurrence, which is in a magnetite-bearing facies of the Littleton Formation (Moench, 1984, field observations). Although these iron formations are not an iron resource, they provide evidence of chemical sedimentation during volcanic hiatuses; such hiatuses are essential ingredients for the formation of massive sulfide deposits.

Interspersed with the group of massive sulfide deposits southwest of Littleton, N.H., are several quartz-carbonate-sulfide veins, some gold bearing, that constitute the so-called "Ammonoosuc gold field" (Hitchcock, 1878, p. 7–34). The veins are clearly postorogenic but otherwise are undated and may be approximately coeval with the silver-bearing quartz-galena veins at the Shelburne and Mascot mines in eastern New Hampshire (fig. 6). These latter veins are inferred by Cox (1970) to have originated during emplacement of the Mesozoic plutons of the White Mountain Plutonic-Volcanic Suite. In western New Hampshire, the most productive vein is at the Dodge mine (fig. 6), said to have yielded $50,000 in gold (1878 price) prior to 1878 (Hitchcock, 1878, p. 11). According to Hitchcock's description, the Dodge vein strikes northeast, dips northwest, and cuts gray slate of the Littleton Formation. It is as much as 5 m thick and is composed mainly of quartz, which contains masses of slate, and crystals of pyrite, ankerite, and galena; spangles of gold are reported, particularly along slate-quartz contacts. The Atwood mine (fig. 6), said to have produced $300 in gold (1878 price), has a quartz vein that is at least 1.3 m thick and cuts Clough Quartzite. The Atwood vein contains sparse pyrrhotite and chalcopyrite; gold is associated with the pyrrhotite (Hitchcock, 1878, p. 9).

Farther north at Granby Stream (fig. 6) is a subvertical, 1-m-thick quartz vein that strikes N. 85° W. and cuts sharply across basaltic amphibolite of the Perry Mountain Formation(?); the amphibolite has pinpoints of chalcopyrite, and the vein contains coarsely crystallized clots of chalcopyrite and scattered flakes of molybdenite (Johansson, 1963, p. 83; R.H. Moench, 1986, field observations). To the east is an occurrence at Morse Mountain, which is in a large roadcut about 3.2 km north of Groveton, N.H. At one place in the roadcut, thin veinlets and breccia fillings of drusy quartz, calcite, sphalerite, galena, and sparse chalcopyrite occur along a minor, northwest-dipping normal fault; veinlets of solid pyrite and chalcopyrite splay off the footwall of the fault. Other sulfide-bearing veinlets also occur in the roadcut.

The close association of stratabound sulfide deposits and distinctly younger veins southwest of Littleton may not be a coincidence. Following regional brittle fracturing, probably mainly in Mesozoic time, the vein minerals may have precipitated from hydrothermal fluids that circulated through widespread, open fracture systems; the metals in the resulting veins could have come from preexisting nearby sources. Thus, quartz veins in the "Ammonoosuc gold field" may have acquired gold from a possibly detrital (placer) source in the Clough Quartzite (perhaps the "auriferous conglomerate" of Hitchcock, 1878, p. 20–21) and (or) from volcanogenic sulfide deposits in the Ammonoosuc Volcanics or in the volcanic-bearing facies of the Perry Mountain Formation. If so, the discordant metal-bearing veins may suggest the presence of nearby stratabound mineral deposits. A similar model for the source of gold in quartz veins in New Mexico has been proposed recently by Fulp and Woodward (1990).

TERRANE FAVORABILITY FOR VOLCANOGENIC MASSIVE SULFIDE DEPOSITS

Factors that control the size and distribution of the stratabound mineral deposits (fig. 6) can be explained by a set of recognized host-rock characteristics. These characteristics provide a geologic basis for estimating the likelihood that undiscovered deposits may occur within a given area and can be grouped into five broad recognition criteria, assembled from many sources by me and augmented by J.F. Slack (*in* McCammon, 1986, p. 1–12). The criteria are as follows:

(1) Occurrence of a thick sequence of originally permeable subaqueous volcanic rocks that could have hosted a large, energetic hydrothermal circulation system. A minimum thickness of rocks of about 1/2 km is probably required. Pillow basalts or thickly stratified pyroclastic rocks would be ideal, but large amounts of interbedded shaly sediments would inhibit circulation across bedding and reduce the potential for the formation of large deposits. A subaqueous environment may be indicated by pillow basalts, turbidites, or chemical sediments. The necessary minimum depth of water is uncertain, but the occurrence of bedded massive sulfide deposits in the region is evidence that the water depth was great enough to prevent boiling and precipitation of metals in the underlying hydrothermal conduits.

(2) Evidence of a volcanic hiatus that was sufficiently long enough to permit formation of a large sulfide deposit on the sea floor, accompanied by indications of chemical sedimentation. Evidence for a volcanic hiatus and chemical sedimentation may include zones of thinly stratified, fine-grained epiclastic sediments, chert, coticule rock (fine-grained, spessartite-rich layers), iron formation (pyrite-, magnetite-, hematite-, chlorite-, or carbonate-facies), bedded barite, and (or) tourmalinite.

(3) Evidence in the form of premetamorphic alteration that hydrothermal circulation has occurred, shown particularly by the depletion of calcium and sodium and by the introduction of magnesium. The rocks may be highly aluminous and locally highly potassic. In areas of greenschist-facies metamorphism, alteration may be expressed as wide zones of bleached, pyritized, sericitized, and (or) silicified rocks and as chlorite-, chloritoid-, or talc-rich phyllite (±carbonate). At higher metamorphic grades, hydrothermal alteration may be expressed as bleached pyritic quartz-muscovite schist (±phlogopite), cordierite-rich rocks (±anthophyllite and (or) gedrite), chlorite-garnet rocks (±staurolite), quartz-kyanite gneiss (±sillimanite or andalusite), and (or) albitite. An abrupt

change from strongly altered to stratigraphically higher unaltered rocks may mark the upper contact of a nearby stratabound sulfide-bearing zone.

(4) Occurrence of subvolcanic intrusions that may have provided local heat sources for driving hydrothermal circulation systems. Although favorable, such bodies are not critical recognition criteria because adequate circulation may result also from high regional heat flow, transmitted by deep fractures or faults.

(5) Evidence of tectonic extension to maintain open, permeable conditions for fluid transport. Whereas compression tends to reduce the volume of pore space and thereby inhibits circulation, extension tends to increase the volume of pore space and encourage circulation. The occurrences of dike swarms, sheeted intrusions, and listric normal faults parallel to strike, constitute indications of extension normal to strike.

These criteria are listed in approximate order of decreasing importance because, if there are no volcanic rocks (criterion 1), a different ore model is needed to explain known deposits and the subsequent criteria may be increasingly difficult to recognize during mapping in complex metamorphic terranes. It is stressed, furthermore, that younger processes, such as shearing within fault zones or within accretionary mélange (as in the Jim Pond Formation), may dismember originally large deposits and lower the resource favorability of a metavolcanic terrane that otherwise satisfies all five criteria. Application of these criteria to the major metavolcanic belts of figure 6 follows.

Jim Pond Formation

The two colored areas of queried resource favorability that sit astride the Boundary Mountains anticlinorium in northwestern Maine (fig. 6) are within metavolcanic rocks of the Upper Cambrian(?) Jim Pond Formation. Cummings (1988) has described two volcanogenic massive sulfide deposits that occur in these areas. At the Border deposit, in Woburn Township, Quebec, one drill hole showed approximately 9 m of 6 percent zinc within fractured rhyolite; another showed about 6 m of massive sulfides that contain an average of 26.9 percent zinc, 1.3 percent lead, 0.5 percent copper, 12 ounces/ton silver, and 0.035 ounce/ton gold (Cummings, 1988, p. 260). According to Cummings, the sulfides are hosted by gray rhyolite and rhyolite porphyry, and the sulfides apparently "follow stratigraphy" (Cummings, 1988, p. 261). To the best of my knowledge, the metavolcanic host rocks have not been mapped beyond the prospect area. Accordingly, only a small area of probable favorability is shown surrounding the Border prospect on figure 6. At the Alder Stream prospect, in Maine, the best drill-core intercepts showed about 9.5 m containing 0.33 percent copper, 1.5 m containing 1.0 percent copper, and 0.5 m containing 2.6 percent copper (Cummings, 1988, p. 275). The host rocks there are weakly metamorphosed "tuffs, tuffaceous sediments, and mixed flows which contain 2 to 10% pyrite and erratic (low) amounts of Cu sulphides" (Cummings, 1988, p. 275). Probable volcanogenic massive sulfide deposits have been explored at two locations approximately 55 km east of the area of figure 6, along the continuation of the Squirtgun fault (fig. 6, SQF). One is at the Squirtgun flowage dam; the other is about 3 km southwest of the dam (Young, 1968, p. 133). Near the dam, the Squirtgun fault of the prospect area marks the so-called Chase Stream-Churchill Stream lineament and separates the Silurian and Lower Devonian beds of the Moose River synclinorium north of the fault from the more strongly deformed pre-Silurian rocks of the Lobster Mountain anticlinorium south of the fault. The rocks south of the fault are now assigned to the Jim Pond and Hurricane Mountain Formations (see Osberg and others, 1985). Felsic metatuff of the Jim Pond Formation of this prospect area hosts apparently stratabound copper-zinc deposits said to be "quite similar to the ore occurrences in the Bathurst area of nearby New Brunswick" (Young, 1968, p. 133). If these deposits are in fact volcanogenic, their occurrences along the Squirtgun fault are probably coincidental, and the Jim Pond Formation as a whole may be promising for mineral exploration.

The Jim Pond Formation of the colored area that hosts the Alder Stream prospect (fig. 6) is composed of a thick sequence of felsic metavolcanic rocks, thickly stratified epiclastic rocks, and siliceous iron formation that lies above thick, pillowed, basaltic greenstones. The pillowed greenstones are conformably underlain by the Boil Mountain Complex (Boudette, 1982). Almost by definition as ophiolite, the Boil Mountain is extensional in origin. Chert, iron formation, vent-facies fragmental metarhyolite, and contemporary intrusive rocks have been mapped. Although the effects of hydrothermal alteration have not been described (criterion 3 above), all other recognition criteria are satisfied within this area. The Jim Pond Formation is part of an accretionary wedge (Boudette, 1982), however, and some undiscovered major sulfide deposits that may well occur within it might have been tectonically disrupted.

Ammonoosuc Volcanics

Farther south, along the Bronson Hill anticlinorium, the Middle Ordovician Ammonoosuc Volcanics are host to a wide variety of stratabound mineral deposits. The Ammonoosuc is the thickest and most extensive metavolcanic unit in the Sherbrooke-Lewiston region. In the Littleton-Orfordville area, red areas of resource favorability (fig. 6) include the upper and lower members, respectively above and below black sulfidic slate or phyllite of the Partridge Formation. Most of the Partridge Formation is excluded from favorability, but the lower contact of the Partridge is exceptionally favorable for mineral resources because it marks a volcanic hiatus above the thick basaltic and mixed bimodal lower member, and because it is commonly marked by metachert, garnet-rich silicate iron formation, and other exhalative deposits. Although I previously considered the lower member to be generally less than 200 m thick in the Littleton-Orfordville area (and therefore of relatively low resource favorability), I now recognize that it is at least 2 km thick near Littleton, but faulted farther southwest by the Foster Hill sole fault to thicknesses of 200 m or less. The original thickness of the lower member southwest of Littleton may have been considerable, and, therefore, its contact with the overlying Partridge Formation should be considered a prime target.

Near the Milan mine and the Hampshire Hills prospect, the red area of resource favorability includes only the upper member, composed of thickly stratified felsic pyroclastic rocks and volcanic conglomerate, small- to moderate-sized contemporaneous intrusions, iron formations and other exhalites, and wide zones of altered rocks. These rocks lie above thin to thick sequences of massive and pillowed metabasalt of the lower member. Deposits at the Milan mine and Hampshire Hills prospect are in the upper member but near its lower contact. Tectonic extension is indicated by the presence of the intrusion at Chickwolnepy Stream (fig. 6, CS), a metamorphosed body of tonalite and commonly sheeted gabbro and diabase that is inferred to be comagmatic with the Ammonoosuc Volcanics (Moench, 1984) and interpreted to represent an incipient spreading center. Massive sulfide deposits within the Ammonoosuc Volcanics have been mined at the Warren (Ore Hill) mine (Secord and Brown, 1986) and at the Milan mine (Emmons, 1910; Gair and Slack, 1979; R.H. Moench, unpub. field observations). Whereas the Warren deposit is a zinc-lead body that has appreciable silver and only a minor amount of copper, the Milan deposit is a zinc-copper body that has appreciable lead and silver. Total production from the Milan mine, the larger of the two, was probably about 500,000 tons. Other, possibly significant, stratabound sulfide occurrences have been seen in outcrop or rubble. The geologic setting of the Milan deposit, with nearby felsic meta-agglomerate and underlying intensely altered rocks and probable intrusive metafelsite, closely resembles that of the classic Kuroko ores of Japan. All five recognition criteria are thus easily satisfied by large parts of the Ammonoosuc Volcanics.

One chalcopyrite-pyrrhotite deposit, at Hampshire Hills (fig. 6; Pyke, 1985), occurs in silicate-facies iron formation and quartz-kyanite gneiss enclosed in epiclastic metagraywacke of the Ammonoosuc Volcanics. This deposit is several kilometers away from known proximal metavolcanic rocks, so that the recognition criteria for the volcanogenic model may not apply. Pyke (1985) infers that the Hampshire Hills copper deposit accumulated in a hot brine pool with iron-rich, tuffaceous exhalite (now quartz-chlorite-garnet-staurolite-magnetite rock that has feldspathic layers), at some distance from a submarine hydrothermal spring.

Piermont Allochthon

The largest known polymetallic, massive sulfide deposits in the area shown in figure 6 are in the proposed northern extension of the Piermont allochthon, near the international border. These deposits have been explored at the Ledge Ridge prospect (at least 3.7 million tons of 4 percent zinc + copper + lead) and mined in the Clinton River district (reserves about 1.8 million tons of about 2 percent copper, 1.5 percent zinc, and appreciable lead; see Chevé, 1978; Eisenberg, 1983; Stephens and others, 1984; Ebinger, 1985). The Ledge Ridge deposit was found by J.S. Cummings, Incorporated, in an area covered by glacial drift (Cummings, 1988). Regional mapping by several workers and information furnished by the Boise Cascade Corporation (written commun., 1988) indicate, however, that the deposit lies near the structural lower contact of a sequence, possibly more than 500 m thick, of weakly metamorphosed felsic tuff and reworked tuff, fragmental rhyolite, agglomerate, and iron formation apparently conformably above as much as 1,500 m of pillowed, basaltic greenstone. Younging directions determined from pillows in the greenstone sequence (Green, 1968; Eisenberg, 1982; R.H. Moench, field observations) and poorly developed graded bedding in ferruginous metasiltstone and metachert at the greenstone-felsic volcanic contact (R.H. Moench, field observations) indicate that the felsic sequence lies stratigraphically above the greenstone sequence. According to my interpretation, the felsic sequence lies within a tight synclinal crease between two faulted anticlinal belts of metabasalt. The felsic sequence is composed of proximal volcanics related to a large volcanic center. The Silurian East Inlet pluton may be a subvolcanic granitic body within the volcanic center.

Massive sulfide deposits of the Clinton River district are approximately on strike with the Ledge Ridge deposit and occur within the products of the same Silurian volcanic center. The massive sulfide deposits lie within a metamorphosed tuffaceous metachert sequence near the contact between predominantly mafic greenstone flows, tuff, reworked tuff, and coarse pyroclastic rocks and a mixed sequence of feldspathic metagraywacke, graphitic and nongraphitic phyllite, felsic to intermediate metatuff and reworked metatuff, and metamorphosed iron formation (Chevé, 1978; Ebinger, 1985). The mixed sequence is coextensive with the felsic sequence at Ledge Ridge and a bimodal sequence farther south. Although the mixed sequence is considered by workers in Canada to lie stratigraphically below the mafic sequence (Chevé, 1978; Ebinger, 1985; St-Julien and Slivitsky, 1987), the possibility that it lies above the mafic sequence, as proposed in the area of the Ledge Ridge prospect, needs to be examined. The mixed sequence is widely pyritic or pyrrhotitic; local effects of alteration have been described. An extensional tectonic setting is indicated by abundant greenstone dikes and sills that are parallel to the regional structural grain and by the chemical data cited above.

All five recognition criteria thus are satisfied in the area of the Ledge Ridge and Clinton River deposits, and an area (red on fig. 6) can be delineated that is highly favorable for the occurrence of undiscovered massive sulfide deposits. This area coincides with the mapped proximal volcanic host-rock assemblage. The favorable assemblage appears to grade southward into the Dixville quadrangle to a less favorable sequence that contains more distal felsic and mafic metavolcanic rocks and abundant interstratified metasedimentary rocks.

At the Thrasher Peaks prospect (fig. 6), a possibly major stratabound zinc-copper deposit (or deposits) occurs within a proximal volcanic assemblage now assigned to the Lower Devonian Littleton Formation. Because these rocks represent an Early Devonian volcanic center that is only 5 km southeast of the Silurian center that hosts the Ledge Ridge and Clinton River massive sulfide deposits, the rocks at Thrasher Peaks may represent a reactivation of the same center. As described by Young (1968), Fournier (1970), Harwood (1973), Eisenberg (1982, p. 30–34), and Cummings (1988, p. 234–240), pyrite, chalcopyrite, sphalerite, and probably minor galena at Thrasher Peaks occur discontinuously in disseminations and small massive pods in a zone as much as 100 m wide and 5 km long (Eisenberg, 1982, p. 30). Massive accumulations of sulfides occur locally in transposed fold hinges. These deposits are hosted by a deformed and weakly metamorphosed volcanic sequence nearly 1 km wide and 16 km long that is truncated on the southeast by the Thrasher Peaks fault. The sulfide minerals occur mainly in schistose felsic metatuff associated with an extensive layer of rhyolite metatuff-breccia, interstratified with gray phyllite and a thin layer of carbonate-bearing rock. Along the northwestern side of the mineralized belt are lenticular bodies of pillowed basaltic greenstone, basaltic pyroclastic rocks, and probable basaltic volcanic conglomerate and metarhyolite domes or hypabyssal intrusives. On the basis of pillow observations, Eisenberg (1982, p. 34) infers that this sequence youngs to the northwest, but R.A. Cavalero infers an opposite younging direction based on his observations of graded bedding and graded tuff units in drill cores (Boise Cascade Corporation, written commun., 1988). Either way, the host volcanic assemblage has the appropriate thickness, proximal character, and probably related intrusives that, together with the sulfide occurrence, indicate that the entire 6-km-long metavolcanic belt has a high resource favorability for the occurrence of volcanogenic massive sulfide deposits. This favorability may be reduced, however, by tight folding and shearing, related to the Thrasher Peaks fault.

In the southern part of the Piermont allochthon, southwest of the Gore Mountain plutons, many stratabound mineral deposits occur within or near the volcanic-bearing facies of the Silurian Perry Mountain Formation. The only exception is the small copper deposit at Washburn Brook (fig. 6; fig. 4, WB), which occurs in silicate-facies iron formation and associated amphibolite in a magnetite-bearing facies of the Lower Devonian Littleton Formation. The known deposits in the Perry Mountain also are copper dominant and are small; the largest, at the Paddock mine (fig. 6; Hitchcock, 1878, p. 42–43), probably produced less than 50,000 tons of ore (Gair and Slack, 1979). The host rocks there are variably pyritic, interbedded phyllite, tuffaceous metasandstone, felsic ash- to lapilli-metatuff, and basaltic greenstone. Small rhyolitic centers of eruption are indicated by scattered small bodies of fine-grained felsic metaporphyry and granophyre, but no major volcanic center is recognized. Tectonic extension in this area is indicated by swarms of metadiabase dikes and sills and sheeted intrusive bodies and by petrochemical data (Hafner-Douglass, 1986).

The Stevens mine (fig. 6) has been studied in detail by Margeson (1982), who describes the ore as thin, discontinuous laminations composed of pyrite, chalcopyrite, and generally subordinate amounts of arsenopyrite, galena, and sphalerite; locally the laminations are zinc rich. The sulfide laminations are associated with 1-m-thick, iron-rich layers of quartz and pyrite that pass laterally to iron-rich laminations of chlorite, sericite, and quartz. Margeson (1982, p. 124–130) proposes that the base-metal sulfide deposits and associated iron-rich rocks at the Stevens mine precipitated from brine that pooled in a shallow depression on the sea floor, at or near a hydrothermal vent. This model also seems appropriate for most of the other known stratabound sulfide deposits of the southern part of the allochthon. Although recognition criteria 2–5 are satisfied, the presence of abundant interstratified phyllite in the host-rock assemblage would have inhibited hydrothermal circulation. Accordingly, the potential for the occurrence of large massive sulfide deposits in most of this part of the allochthon is probably small.

Possible exceptions are two areas of speculative resource favorability (pink areas on fig. 6). The southwestern area is underlain by thickly stratified, strongly altered pyritic quartz-muscovite schist, commonly having "quartz eyes," quartz metaporphyry, and greenstone, most of which may be intrusive. These rocks, which host two known copper prospects, were mapped previously as Ammonoosuc Volcanics (Billings, 1935; Hall, 1959). On the basis of recent mapping, I consider these rocks to be a strongly altered mixed sedimentary and rhyolitic facies of the Perry Mountain Formation. In the northeastern area, rocks that host the copper-dominant Essex prospect and two small undeveloped chalcopyrite occurrences (fig. 6) contain abundant poorly stratified pyritic "quartz-eye" schist, local felsic lapilli-metatuff and greenstone, as well as metasedimentary rocks also assigned to the Perry Mountain Formation. Also in the northeastern area are small granitic bodies (not shown on fig. 6) mapped by Eric and Dennis (1958). Although Eric and Dennis correlated the granitic bodies with the Ordovician Highlandcroft Plutonic Suite, I have found several other similar bodies elsewhere within rocks of known Perry Mountain Formation. Rocks of both areas satisfy all of the recognition criteria, and to encourage exploration, these areas are shown in pink as having a speculative potential for the occurrence of important massive sulfide deposits.

CONCLUSIONS

The Piermont allochthon of the northern Connecticut valley and its proposed extension to the northeast are a tract of predominantly Silurian subaqueous sedimentary and volcanic rocks that accumulated at moderate water depths (several hundred meters(?)) on an inferred platform just northwest of the western margin of the basin that was ancestral to the Kearsarge-central Maine trough. Sedimentation and bimodal volcanism on the platform were accompanied by the emplacement in an extensional environment of mafic and bimodal dike swarms, sheeted intrusive bodies, and possibly many small granitic plutons, and by hydrothermal sea-floor alteration and sulfide mineralization. Many small, copper-dominant volcanogenic massive sulfide deposits formed where hydrothermal circulation was restricted by low host-rock permeability in volcanic rocks interbedded with abundant shaly sediments. Relatively large, polymetallic massive sulfide deposits formed, however, near a major volcanic center.

This entire assemblage was transported during an early stage of the Acadian orogeny to its present position west of the Bronson Hill and Boundary Mountains anticlinoria and formed the Piermont allochthon. The allochthon is juxtaposed against two other metavolcanic terranes that also contain, or may contain, important stratabound massive sulfide deposits, including the Upper Cambrian(?) Jim Pond Formation of the Boundary Mountains anticlinorium and the Middle Ordovician Ammonoosuc Volcanics of the Bronson Hill anticlinorium.

The potential for occurrence of undiscovered massive sulfide deposits in all three terranes is determined by the same major geologic recognition criteria: (1) adequate thickness of originally permeable volcanic rocks that would host hydrothermal circulation systems, (2) evidence of a volcanic hiatus and chemical sedimentation, (3) evidence of hydrothermal circulation in the form of premetamorphic alteration, (4) evidence of a heat source to drive circulation, and (5) evidence of tectonic extension.

The many small, crosscutting, locally gold-bearing, quartz-carbonate-sulfide veins of the so-called Ammonoosuc gold field and elsewhere within and outside the Piermont allochthon probably formed during brittle fracturing and hydrothermal circulation that accompanied emplacement of the high-level Mesozoic plutons of New Hampshire. The mineral content of these veins may have come from nearby rocks and stratabound sulfide deposits, such as proposed by Fulp and Woodward (1990) for the source of gold in quartz veins in New Mexico.

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Chapter K

Stratabound Zinc-Lead-Copper Deposits in the Cambrian Carbonate-Siliciclastic Shelf Sequence at Lion Hill, West-Central Vermont

By SANDRA H.B. CLARK

U.S. GEOLOGICAL SURVEY BULLETIN 1887

SUMMARY RESULTS OF THE GLENS FALLS CUSMAP PROJECT, NEW YORK, VERMONT,
AND NEW HAMPSHIRE

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By Sandra H.B. Clark[1]

Abstract

Occurrences of sulfide mineralization at Lion Hill, about 4 kilometers north of Brandon, Vermont, have been known at least since the 1870's. Mineralized occurrences consist predominantly of stratabound sphalerite and minor galena and chalcopyrite. The host rock is the Lower Cambrian Monkton Quartzite, a mixed carbonate-siliciclastic sequence that formed part of the Cambrian to Lower Ordovician shallow shelf deposits in the area east of the Adirondack Mountains. Lion Hill is on the overturned eastern limb of the Middlebury synclinorium, and the host rocks are strongly deformed and metamorphosed to greenschist-facies assemblages; primary sedimentary features are only locally preserved. Base-metal sulfides are mostly in metadolostone and occur discontinuously in two main stratabound zones. Sulfide minerals are commonly found in thin stringers elongate in the schistosity planes that formed as metamorphic segregations; these textures indicate that mineralization was pretectonic, or at the latest, syntectonic in origin. Rarely preserved pretectonic textures such as intergranular disseminated sulfides and sulfide fillings of fractures when the sediments were semiconsolidated suggest that one phase of mineralization took place during diagenesis. Fragments of layered magnetite iron formation that contain sphalerite and chalcopyrite are present at one prospect. The coincidence of layered iron formation and stratabound zinc-lead-copper mineralization suggests analogy with some of the Irish carbonate-hosted base-metal deposits and the possibility that the origin of the Lion Hill mineralization was in part syngenetic by submarine exhalative activity. Supporting evidence is in the scarcity of gangue minerals and lack of abundant open-space filling and in the textural features of the sulfides that suggest both diagenetic and epigenetic mineralization. The interpretation of a synsedimentary origin for the Lion Hill deposits suggests potential for commercially important, sediment-hosted lead-zinc-copper-silver deposits in the carbonate-siliciclastic shelf sequence throughout western Vermont, eastern New York, northwestern Massachusetts, and southeastern Quebec.

INTRODUCTION

Occurrences of zinc-lead-copper mineralization in rocks of the Cambrian-Ordovician carbonate-siliciclastic continental shelf are known at many localities in western Vermont, eastern New York, and southeastern Quebec (fig. 1). Although none of the occurrences is known to be economic, mineralization has been drilled by three companies at Lion Hill, north of Brandon, Vt. A prospect at Lion Hill was mentioned by Jacobs (1943[?], 1944[?]), who stated that a 50-ft shaft (the Nicklaw mine?) was sunk in the 1870's, from which an unknown amount of ore was mined. The St. Joseph Lead Company did core drilling during 1941, and although good ore was found, the width did not warrant further work (Jacobs, 1943[?], 1944[?]).

From 1952 through 1954, the New Jersey Zinc Company carried out exploration work at Lion Hill, including mapping of the Fay shaft, soil sampling, and core drilling of 30 holes, some of which were to a depth of 800 ft. Between 1981 and 1982, an exploration drilling program was conducted by Labradex Corporation and eight holes were drilled, the deepest of which was 410 ft. Upon completion of their exploration programs, the New Jersey Zinc Company and the Labradex Corporation gave most of the drill core and the pertinent maps and cross sections to the U.S. Geological Survey. Studies of these materials, along with fieldwork by the author, provide the information upon which the description of the Lion Hill deposit in this report is based.

Additional occurrences of zinc-lead±copper mineralization in the Cambrian-Ordovician shelf sequence (fig. 1) are known near Franklin, Vt. (Clark, 1987); Milton, Vt. (G.C. Gregory, 1982, oral commun., 1987); Dunham, Quebec (Gauthier and others, 1985); Saratoga Springs, N.Y. (Newland, 1919; Lesure and Klemic, 1977); and Hinesburg, Vt. (Butler, 1986). Other localities in Vermont from which sphalerite or galena has been reported in the shelf sequence, but for which exact locations are unknown, include Orwell, Wallingford, and Danby (Morrill and Chaffee, 1964). Although these occurrences are not economic, they are of interest because they indicate potential for significant mineralization in rocks of the shelf sequence.

Manuscript approved for publication April 13, 1990.

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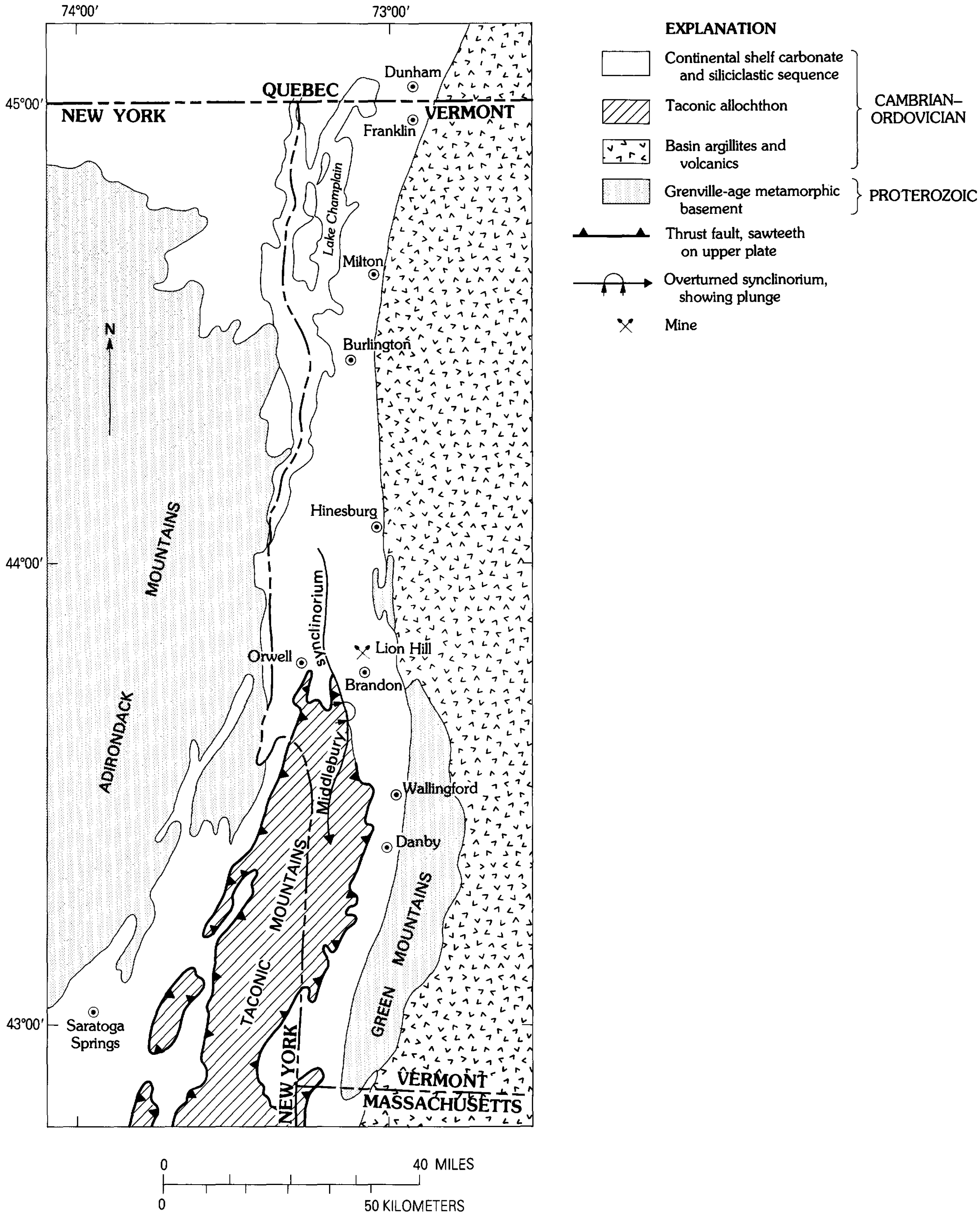

Figure 1. Regional geologic setting of zinc-lead-copper occurrences, western New York and eastern Vermont. Geology from Doll and others (1961) and Fisher and others (1971).

REGIONAL SETTING

Mineralization at Lion Hill is contained in rocks of the Cambrian-Ordovician continental shelf carbonate and siliciclastic sequence (fig. 1) that forms the lowlands of the Champlain and Vermont valleys. The area lies between the Grenville-age metamorphic basement rocks of the Adirondack Mountains to the west and the Green Mountains to the east (see Thompson, this volume). Beginning in Middle Ordovician (Taconic orogeny) time, deep-water sediments

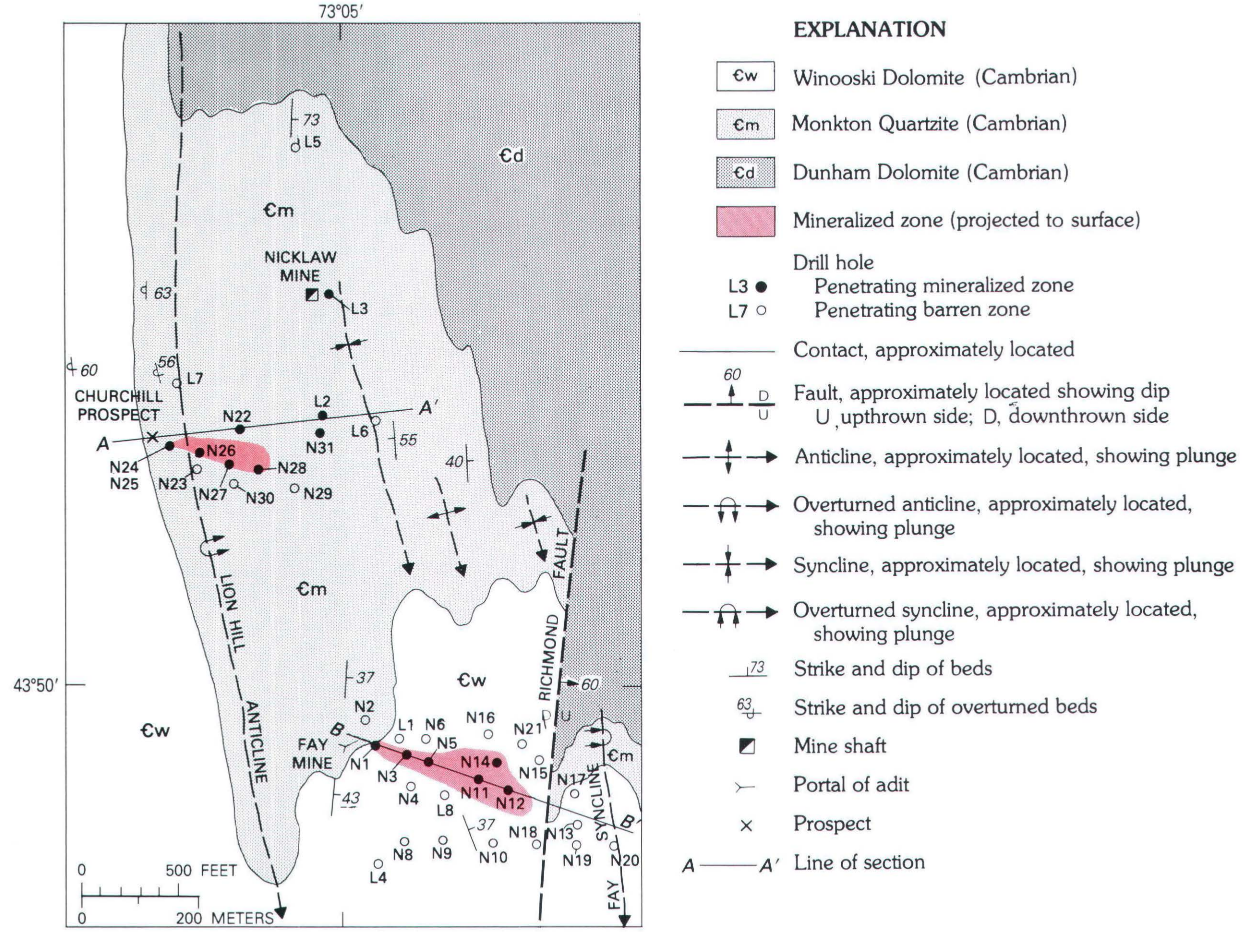

Figure 2. Geologic map of the Lion Hill area showing locations of drill holes and old workings. Cross section *A–A′* is shown in figure 3. Cross section *B–B′* is shown in figure 4.

were thrust over the shallow-water platform sequence as part of the array of events caused by the collision of the continent and an island arc to the east and the collapse of the continental margin (Zen, 1972; Baldwin, 1981). Rocks of the shelf sequence were folded into foreland folds, one of which is the Middlebury synclinorium in which the Lion Hill deposit is located. Lion Hill is about 3 km west of the hinge line where the tectonically stable continental shelf was down faulted into the deep-water basinal environment of the Proto-Atlantic (Iapetus) Ocean to the east (Friedman, 1979) and about 2 km west of the trace of the regional system of north-trending faults that developed during early Mesozoic time (Stanley, 1980) in a zone approximately parallel to the early Paleozoic hinge line (Clark and Slack, 1988).

STRATIGRAPHY AND PALEOENVIRONMENT

The mineralization at Lion Hill is in the unit mapped by Cady (1945) as the Lower Cambrian Monkton Quartzite (fig. 2). The Monkton Quartzite formed as part of the Cambrian to Lower Ordovician platform sequence, which consists of alternating cycles of shallow-water carbonate and siliciclastic sediments that were deposited on a tectonically stable continental crust (Mehrtens, 1985). The Monkton Quartzite is underlain by the Dunham Dolomite, which is predominantly carbonate and records transgressive sedimentation from peritidal through shelf-edge environments on a stable platform (Mehrtens, 1985). The lithologic contact between these two formations is gradational because the environments represented across the contact are similar. The primary difference between the two formations is that the proportion of siliciclastic sand and dolomite differs (Mehrtens, 1985). The Cheshire Quartzite conformably underlies the Dunham Dolomite and is at the base of the sequence.

The Monkton Quartzite has been studied and described in detail in an area north of Burlington, Vt., where it is relatively undeformed and where sedimentary features, such as ripple marks, mud cracks, rain drop impressions, crossbedding, and trace fossils, are well preserved (Keith, 1923; Rahmanian, 1981a,b; Speyer, 1983). The 300-m-thick Monkton Quartzite in this area was formed from shallowing-up sedimentary cycles characterized by repetitive packages of (1) subtidal siliciclastic sand shoal and channels; (2) interbedded siliciclastic sand, silt, and carbonate intertidal sediments; and (3) carbonate muds of the high intertidal and supratidal flat (Mehrtens, 1985; Rahmanian, 1981a,b). These cycles are interpreted as representing prograding tidal-flat deposits. In addition, two siliciclastic facies have been recognized—(1) tidal channels and sand bars and (2) mixed rippled sand and mud drapes of the intertidal zone (Mehrtens, 1985).

In the Brandon, Vt., area, the sedimentary features of the Monkton Quartzite, which are relatively unaffected by deformation in the Burlington area, are only rarely preserved because of more intense deformation and metamorphism. However, some sedimentary features can be seen locally, including medium-scale cross-stratification that has sharp erosional bases and a transition upward into finely laminated material. Also preserved in places are laminar and small fingerlike structures that may be algal fabrics. The similarity of lithology and local preservation of sedimentary fabrics suggest that the tidally dominated environment of deposition of the Monkton Quartzite extended to the Brandon area. Because there is a greater proportion of carbonate and finer terrigenous sediments here than farther north, the Monkton Quartzite may intertongue with and pinch out into open-shelf carbonates (Burt, 1982).

The Monkton Quartzite is conformably overlain by the Winooski Dolomite, a predominantly carbonate unit that may represent a transgressive sequence from peritidal through shelf-edge environments (Mehrtens, 1984, 1985). The Winooski Dolomite is gradationally and conformably overlain by the Danby Quartzite, another mixed carbonate and siliciclastic unit that is interpreted as having been deposited in a complex mosaic of fair-weather- and storm-related sediments on the Cambrian-age carbonate-siliciclastic platform (Mehrtens, 1985).

STRUCTURE

The predominant structural features in the Brandon area are those produced by tectonic deformation. A pervasive schistosity (S_1) has developed commonly parallel to compositional layering in the rocks and is itself folded; this folding of S_1 suggests at least two stages of deformation. A second schistosity (S_2) intersects S_1 and is parallel to the axial planes of the F_2 folds. The structures in the Brandon area reflect their location on the overturned eastern limb of the south-plunging Middlebury synclinorium (fig. 1). The predominant folds are overturned flexural folds that have axes that trend about 10° west of north and plunge about 20° southeast; axial surfaces dip about 50° to 60° east (Burt, 1982). Small-scale folds range from broad, open, and cylindrical similar folds in competent strata to isoclinal folds in calcareous rocks and near zones of shearing and faulting.

Three types of faults have been recognized in the Brandon area: (1) thrusts subparallel to bedding; (2) a smaller break thrust variety (Cady, 1969), which is the rupture of a steeply dipping or overturned limb of an anticline; and (3) bedding-plane shear (Burt, 1982).

The dominant structures at Lion Hill (fig. 2) are the large overturned Lion Hill anticline, which has numerous smaller folds developed on its limbs, and a high-angle reverse fault, identified on the maps and cross sections of the New Jersey Zinc Company as the Richmond fault. It is of the break thrust variety and displaces bedding and contacts. Evidence for this fault consists of brecciated zones and deformed phyllites in drill cores and stratigraphic offset. The Richmond fault may have been reactivated in the general area of an older fault zone that may have served as a conduit for mineralizing fluids.

MINERALIZATION

Two main zones of stratabound mineralization were identified on New Jersey Zinc Company maps and cross sections. One zone intersects the surface at the Fay (also called Oram) mine near the southern end of Lion Hill, and another intersects the surface at the Churchill prospect on the western side of Lion Hill (fig. 2). A third smaller zone of sulfides was identified in core holes (L2 and N31) drilled south of the Nicklaw mine (figs. 2 and 3). Mineralized rock, including pieces of layered magnetite iron formation, were found near the Nicklaw mine at the top of Lion Hill.

The locations of the holes drilled by the New Jersey Zinc Company and Labradex Corporation are shown in figure 2, along with old mines and surface projections of areas identified as mineralized zones on New Jersey Zinc Company maps and cross sections. The western zone of mineralization (fig. 3) is on the overturned limb of a fold about 20 to 30 m from the upper contact of the Monkton Quartzite with the overlying Winooski Dolomite, which, in this area, is below the Monkton. The southern zone of mineralization is on the upright limb of the fold (fig. 4). The drill holes into the southern zone of mineralization begin in the Winooski Dolomite and intersect the Monkton Quartzite at depth; the mineralized zone is about 20 m below the Winooski Dolomite-Monkton Quartzite contact.

The third smaller zone of mineralization was identified during the study of drill cores supplied by Labradex Corporation. This zone is on the upright limb of the fold south of the Nicklaw mine (fig. 3). It may be the continuation of the same layer of mineralization as in the western zone but folded around to the upright limb of the fold. This mineralized zone is also at approximately the same horizon as the Nicklaw mine and the southern zone of mineraliza-

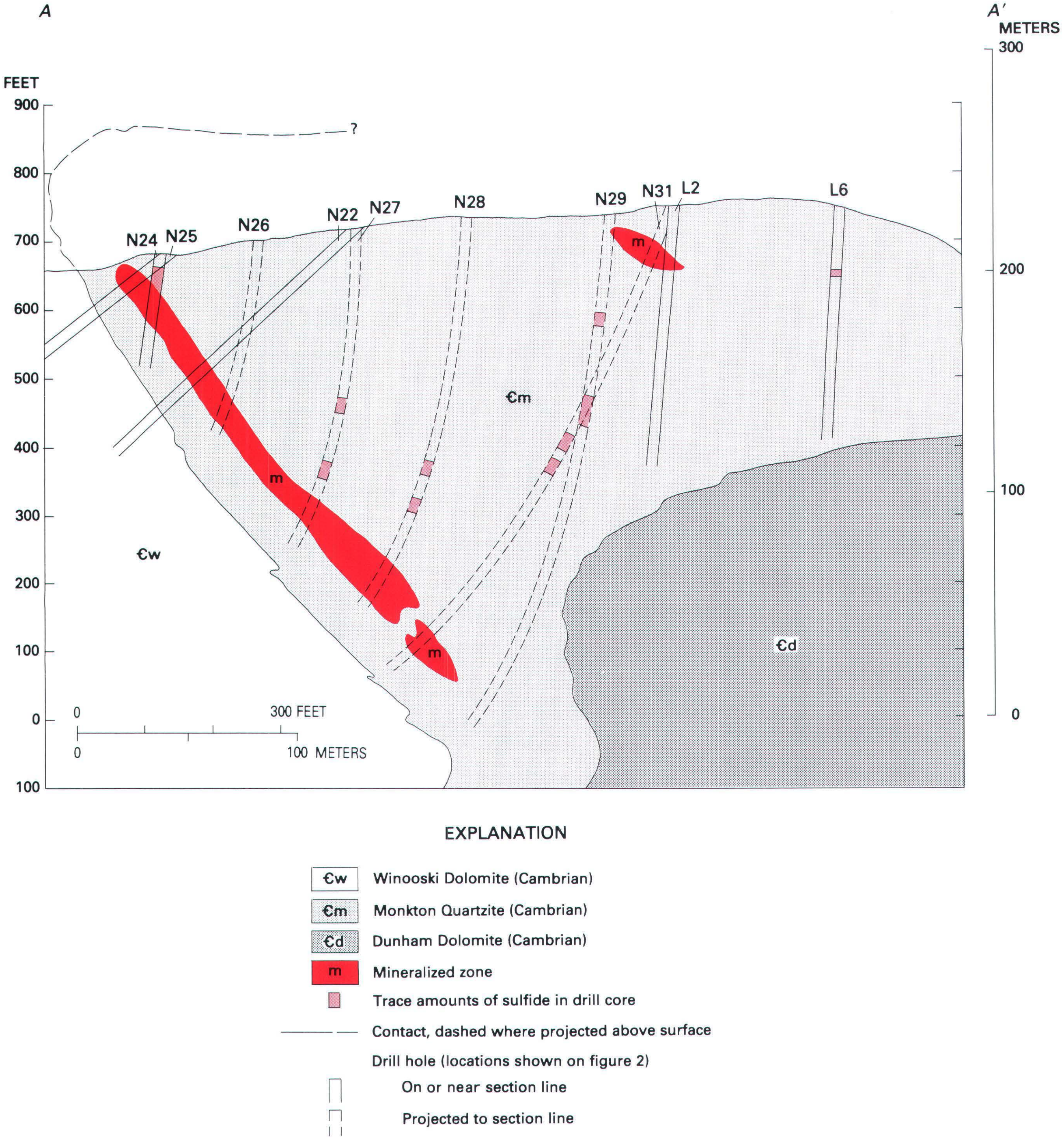

Figure 3. Generalized composite cross section along line *A–A′* (fig. 2), including projected location of the western mineralized zone at Lion Hill.

tion, which are also on the upright limb of the major fold. These relations indicate that the main mineralized zones are stratabound and possibly also stratiform. However, small amounts of sulfides occur both above and below the two main zones of mineralization (figs. 3 and 4), as well as in some drill holes outside these main zones. The mineralized zones are probably lenticular in overall form but in detail are likely to be irregular and discontinuous as the result of complex folding, shearing, and faulting and because they may have been irregular and discontinuous when originally deposited.

The two mineralized zones identified on New Jersey Zinc Company drawings cover approximately 20,000 m^2 and contain approximately 500,000 tons of sulfide-bearing rock. Using the assumptions of a tonnage factor of 12.5 and an average grade of 2 percent combined metal (zinc+

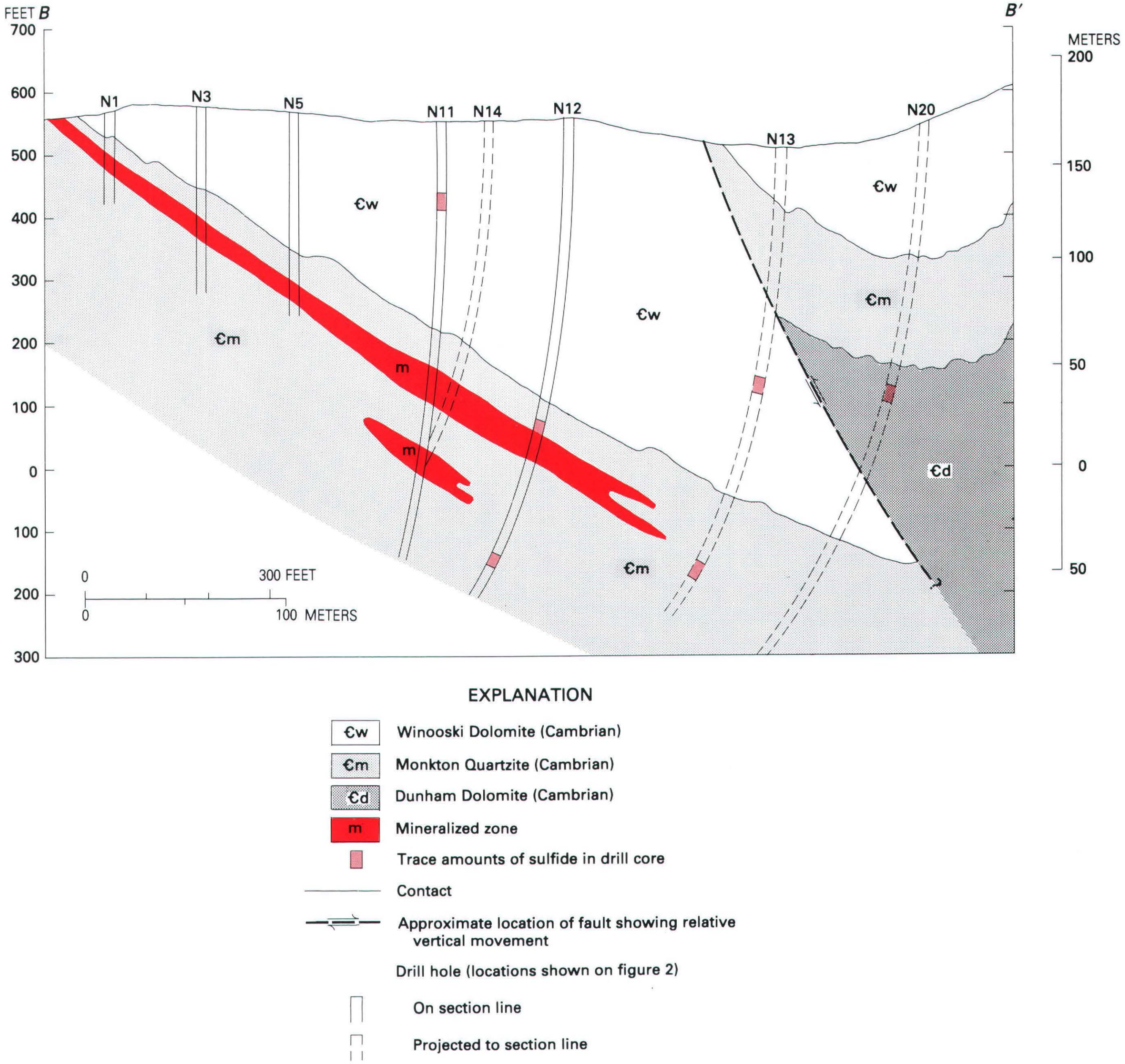

Figure 4. Generalized cross section along line *B–B′* (fig. 2) through the southern mineralized zone at Lion Hill.

lead±copper), the zones thus defined would contain a total of approximately 10,000 tons of sulfide minerals.

The predominant rock types at Lion Hill are metadolostone, phyllite, and metasandstone; gray metadolostone is the most common host rock for mineralization. Mineral assemblages include dolomite, quartz, microcline, biotite, sericite, and chlorite, which indicate greenschist-facies metamorphic conditions. Schistosity and foliation are defined by orientation and segregation of micaceous minerals; elongation of quartz, dolomite, and feldspar; and cataclastic flow textures. This schistosity (S_1) is folded locally, and a second schistosity (S_2) intersects the first and is parallel to axial surfaces of folds (F_2). Accessory minerals include zircon, rutile, sphene, epidote, and leucoxene (some leucoxene specimens have magnetite or ilmenite cores). These minerals are subrounded and appear to be detrital in origin. Tourmaline is commonly present in the rocks in small amounts (<1 percent) and occurs as rounded grains (typically green) that appear to be also of detrital origin. Some tourmaline occurs as euhedral grains (typically brown) that are associated with biotite in schistosity surfaces and appears to have recrystallized during metamorphism. Other tourmaline grains have subrounded green cores and brown rims that may represent metamorphic overgrowths on detrital grains. Pyrite is also abundant locally and forms large (5–6 mm) euhedral crystals in some

Table 1. ICP–AES analyses of drill hole samples from the mineralized areas near the Fay mine, the Churchill prospect, and south of the Nicklaw mine (figs. 2–4)

[Holes drilled through the mineralized zones are identified by an asterisk. Analyses by Elwin Mosier, U.S. Geological Survey]

	Drill hole	Depth interval (ft)	Zn	Pb	Ag	Cu
			(Range, in ppm)			
Fay mine area	*N14	508–574	24–>2,000	0–15,000	0–59	0–1,400
	*N11	369–421	33–24,000	0–1,400	0–1	2–33
	N13	690–710	500–>2,400	0–190	0	3–10
	N10	368–416	13–2,100	0–92	0	0–103
	N10	647–664	19–2,200	0–25	0	0–16
	N4	223–229	85	140	0	21
Churchill prospect area	*N26	211–259	9–21,000	0–7,800	0–9	2–250
	*N28	473–540	41–>3,000	0–1,900	0–1	0–34
Nicklaw mine area	*L2	66–100	57–>2,400	15–100	0	5–63
	*L3	78–95	22,000	140–820	0	7–76
	L6	80–103	13–48	0–17	0	5–38

layers. Much, if not all, of the pyrite is thought to be diagenetic in origin because it is associated with organic-rich layers in phyllites, for example, and in metadolostones that have algal fabrics.

Sphalerite is the most abundant ore mineral. In hand specimen, it is either pale brown or light green, and in transmitted light in thin section, it is colorless or yellow; no growth zoning or color banding is present. Galena is the second most abundant ore mineral. Chalcopyrite generally occurs in minor amounts, and pyrrhotite is rare. All of these sulfide minerals commonly form irregular metamorphic segregations that are elongate within the schistosity and shear planes. They also occur as irregularly shaped, interstitial grains and in discordant veinlets. In the metamorphic segregations, sphalerite and galena are intergrown with quartz, carbonate, microcline, and aligned biotite and chlorite. The predominant textures of the sulfides reflect remobilization and recrystallization that occurred during metamorphism, and primary textures are generally obliterated. However, in some places, the textures of the sulfides appear to reflect original depositional features. In one section of drill hole N28, anhedral, pale-brown sphalerite fills intergranular spaces in metasandstone and is concentrated along bedding surfaces and cross-stratification. Mineralizing fluids apparently completely permeated the rock, and sulfides were deposited interstitially as part of the cementing material. In some thin sections, irregularly shaped disseminated sulfides (most commonly sphalerite) are observed that may fill spaces between dolomite and quartz grains of host rocks but are concentrated along certain beds. Such intergranular disseminated sulfides probably were deposited during diagenesis.

Sulfides locally form the matrix of metadolostone breccias. These sulfides also may have been deposited during diagenesis, but replacement of an earlier cementing material is equally possible. In some metadolostones, thin, irregularly shaped zones are filled with minerals, including sphalerite, that have textures suggesting flowage of material in a plastic state into fractures that cut across bedding and that, in turn, are cut by schistosity. This texture is interpreted as indicating that mineralizing fluids moved through the sediments while they were still uncompacted.

Mineralization also occurs as irregular masses, which discordantly cut the host rock. One such mass cuts both host metadolostone and a calcite-cemented solution breccia zone. This structure indicates that there was either an epigenetic phase of mineralization or a remobilization that was associated with metamorphism, or both.

Gangue minerals, which are sparse, either occur in veinlets or in segregations with sulfides. The principal gangue minerals are quartz, dolomite, and microcline. Cross-cutting veinlets and irregularly shaped blebs of quartz, dolomite, and microcline that lack associated sulfides occur also in the drill core and surface outcrops at Lion Hill. Barite and fluorite are conspicuously absent from the Lion Hill deposit.

Several large pieces of layered magnetite iron formation have been found on the surface near the Nicklaw mine (but not in outcrop or drill core). In addition to magnetite, the iron formation contains abundant sphalerite and chalcopyrite, minor amounts of pyrite and galena, and traces of specular hematite. Grains of magnetite and sulfides are concentrated in alternating layers that are rich in quartz, quartz-carbonate, or biotite-chlorite. The micaceous minerals show a strong schistosity, but euhedral grain borders are commonly preserved in magnetite grains.

Drill core samples from 11 holes were analyzed by using Induction Coupled Plasma-Atomic Emission Spectroscopy (ICP–AES). The results for zinc, lead, silver, and copper from the main mineralized zones in each of the three mineralized areas are summarized in table 1. Zinc values are up to 24,000 parts per million (ppm) (about 2 percent), locally; lead values are up to 15,000 ppm (about 1.5 percent) in the most strongly mineralized holes (N11 and N14). Silver is present but only in the most intensely

Table 2. Analyses of three samples of layered magnetite iron formation collected near the Nicklaw mine

[Analyses by Roosevelt Moore and J.R. Gillison of the U.S. Geological Survey. All elements are in parts per million except for iron, which is in percent]

Sample	1	2	3
Emission spectrographic analyses			
Ag	84	74	19
Cd	620	840	690
Co	23	31	26
Cu	>1,500	>1,500	>1,500
Fe	>24	>24	>24
Ni	160	190	210
Pb	>1,000	>1,000	>1,000
V	>1,000	>1,000	>1,000
Zn	>10,000	>10,000	>10,000
Graphite furnace analyses			
Se	8.1	24	21
Fire assay			
Au	<0.075	<0.075	<0.075

mineralized areas. The highest silver value in the analyzed samples is 59 ppm.

Emission spectrographic analyses of three samples of the iron formation from the Nicklaw mine area (table 2) show high values for silver (19–84 ppm), nickel (160–210 ppm), and vanadium (>1,000 ppm). In addition, the analyses show high contents of iron (>24 percent), copper (>1,500 ppm), lead (>1,000 ppm), and zinc (>10,000 ppm).

ORIGIN

Several features of the mineralization at Lion Hill are analogous to those of the carbonate-hosted, base-metal deposits of Ireland, which are stratabound and locally stratiform. In particular, the association of layered iron formation and stratabound zinc-lead-copper sulfides at Lion Hill is similar to the mineralization at the Tynagh deposit in Ireland (Derry and others, 1965). This association suggests that the Lion Hill deposit may have had a partly syngenetic origin by submarine exhalative activity. The iron formation at Tynagh is thought to have resulted from the local outpouring of hydrothermal solutions that precipitated iron and associated minerals in the nearest undisturbed basin (Clifford and others, 1986). The submarine hot springs at Tynagh are believed to have been the source of fluids for the associated lead-zinc mineralization. At Lion Hill, the disseminated interstitial sulfides, especially where concentrated along bedding, are believed to have formed during diagenesis when the sediments were still uncompacted. The filling of fractures during diagenesis at Lion Hill is similar to the interpreted origin of the dilatant fracture fillings in the Tynagh deposit. In the Tynagh deposit, fracture fillings formed soon after precipitation, when the sedimentary material was in a plastic, semiconsolidated state (Boast and others, 1981). Textures of the Lion Hill sulfides also record epigenetic processes in the irregularly shaped massive bodies. The epigenetic mineralization at Lion Hill may correspond to the postdilatant, fracture-filling stage of discordant mineralization at Tynagh (Boast and others, 1981) but may have been from remobilization that occurred during a later event.

The Lion Hill deposit is noteworthy in that it lacks many of the features characteristic of Mississippi Valley-type lead-zinc deposits. Open-space filling, for example, is uncommon, and gangue minerals are scarce. In addition, there is no evidence of paleokarst or karst fill or other features that would have provided the pathways and open spaces for solution migration and precipitation. Fluid-inclusion filling temperatures and salinities and lead-isotope compositions of galena from Lion Hill are also distinctly different from those of Mississippi Valley-type deposits (Foley, this volume). Therefore, the mineralization at Lion Hill may be analogous to some of the carbonate-hosted base-metal deposits of Ireland, which are a variant of sediment-hosted stratiform (sedimentary-exhalative) deposits (Andrew, 1986; Hitzman and Large, 1986). Because stratiform lead-zinc deposits tend to be higher in grade (and higher in silver) than the large-tonnage, but lower grade Mississippi Valley-type deposits (Gustafson and Williams, 1981), the discovery of stratiform mineralization has greater implications for commercial exploitation. The interpretation of a partly syngenetic (sedimentary-exhalative) origin for some of the lead-zinc-copper mineralization at Lion Hill, although not known to be economic at the present time, suggests potential for commercially important, sediment-hosted, lead-zinc-copper-silver deposits elsewhere in the carbonate-siliciclastic shelf sequence throughout western Vermont and in eastern New York, northwestern Massachusetts, and southeastern Quebec.

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Chapter L

The Lion Hill Zinc-Lead-Copper Deposit, Brandon, Vermont—Source of Lead and Fluid Systematics

By NORA K. FOLEY

U.S. GEOLOGICAL SURVEY BULLETIN 1887

SUMMARY RESULTS OF THE GLENS FALLS CUSMAP PROJECT, NEW YORK, VERMONT, AND NEW HAMPSHIRE

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The Lion Hill Zinc-Lead-Copper Deposit, Brandon, Vermont—Source of Lead and Fluid Systematics

By Nora K. Foley[1]

Abstract

Sphalerite, galena, pyrite, quartz, and minor chalcopyrite and pyrrhotite are found in stratabound veins and shear zones in rocks of the Lion Hill area, near Brandon, Vermont. The sulfide minerals occur predominantly as discordant, epigenetic masses, although some interstitial sulfides may have formed during diagenesis. Homogenization temperatures of primary and secondary fluid inclusions measured in epigenetic sphalerite range from 152 to 196 °C; salinities range from 11.5 to 14 equivalent weight percent NaCl. Measured eutectic temperatures of about −25 °C and a lack of daughter minerals indicate that sodium is the dominant cation in solution in the inclusions.

The lead isotopic compositions of galenas from zinc-lead ore and from iron formation of the Lion Hill area range from 18.253 to 18.632 for $^{206}Pb/^{204}Pb$, from 15.536 to 15.618 for $^{207}Pb/^{204}Pb$, and from 38.126 to 38.496 for $^{208}Pb/^{204}Pb$. These values are similar to those of syngenetic(?) lead-zinc ores at White Creek, New York ($^{206}Pb/^{204}Pb$—18.424; $^{207}Pb/^{204}Pb$—15.563; and $^{208}Pb/^{204}Pb$—38.264). The mineralogy, fluid chemistry, and lead isotopic composition of the Lion Hill zinc-lead-copper deposits are similar to those of the sediment-hosted, stratabound lead-zinc deposits of Ireland.

INTRODUCTION

Minor deposits of zinc and lead are found throughout the northeastern United States in rocks of early Paleozoic age (Clark, this volume). Many of the deposits have features that are typical of some Mississippi Valley-type (MVT) ore deposits, but they are generally of too low a grade and are too limited in extent to be of economic interest. However, a few lead and zinc deposits in the region are, in part, stratiform and syngenetic or diagenetic in origin, and these may have a greater resource potential (Clark, this volume). The deposits at Lion Hill near Brandon, Vt., for example, are thought to be similar to the carbonate-hosted, stratabound base-metal deposits of Ireland (Clark, this volume; Clark and Slack, 1988).

In the Lion Hill area, sphalerite and minor galena, pyrite, quartz, chalcopyrite, and pyrrhotite are found in small quartz-dolomite veins, as disseminations, and in shear zones. The sulfide minerals occur in subeconomic quantities, although small amounts of lead and zinc have been extracted. The geologic setting, mineralogy, and textural characteristics of these veins are described in detail by Clark (this volume). The purpose of this study is to better understand the source(s) of the lead and the composition of the fluids that formed the mineralization at Lion Hill. By characterizing the mineralization in detail, a better assessment may be made of the resource potential of the region.

Common genetic models have been proposed for both MVT and sedimentary exhalative (sedex) -type mineral deposits (for example, Cathles and Smith, 1983; Lydon, 1983); a major difference, the coeval nature of the sedex deposits versus the epigenetic nature of the MVT deposits, is used to distinguish the two types. However, because basinal brines may be responsible for the formation of both sedex and MVT deposits, a geologic environment containing one type also may contain the other (Whelan and others, 1984). Characterizing individual deposits is difficult because sedex deposits can have textural characteristics of both syngenetic and epigenetic origin (for example, LeHuray and others, 1987). Classifying the deposits as one type or the other may actually be much less important, therefore, than identifying a promising sedimentary environment.

ANALYTICAL TECHNIQUES

Samples for lead isotopic study were selected from drill core provided by S.H.B. Clark from part of a collection donated to the U.S. Geological Survey by the New Jersey Zinc Company and the Labradex Corporation. Detailed logs of the drill cores prepared by J.M. Hammarstrom and E-an Zen (unpub. U.S. Geological Survey report, 1972) and by S.H.B. Clark (written commun., 1984) were used to select samples. Galena crystals were separated from samples of drill core that had been cleaned in distilled water in an ultrasonic bath. The crystals were acid-washed in a 7 *N* HNO_3 solution, dissolved in 14 *N* HNO_3, and then precipitated as lead nitrate. The samples were then loaded on rhenium filaments by using the silica gel and phosphoric acid technique (Cameron and others,

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Manuscript approved for publication April 13, 1990.

[1] U.S. Geological Survey.

1969) and were analyzed on a 12-in radius, National Bureau of Standards (NBS) -type solid-source mass spectrometer. Instrumental fractionation was corrected by comparison of repeat analyses of NBS–982 to the absolute ratios given by Catanzaro (1968). All errors are reported as standard error of the mean at the 2-sigma level and are generally less than 0.15 percent of the measured ratios.

Fluid inclusion thermometry measurements were made on primary inclusions in coarse sphalerite taken from drill core. The measurements were made on a gas-flow, heating and cooling stage described by Woods and others (1981). The overall precision of measurements of the homogenization temperature of the liquid+vapor is better than +1 °C, and the precision on the melting point of ice is generally better than +0.2 °C.

GEOLOGIC SETTING

The Lion Hill area, situated on the overturned eastern limb of the Middlebury synclinorium west of the Green Mountain massif, Vermont, was a shallow-water continental shelf environment during Cambrian time (Cady, 1945). The lowermost unit in the sedimentary shelf sequence is the Lower Cambrian Cheshire Quartzite. This unit is overlain by the Dunham Dolomite, which is, in turn, overlain by the Monkton Quartzite and Winooski Dolomite. Two Upper Cambrian units also occur in the area, the Danby Formation, which overlies the Winooski Dolomite, and the Clarendon Springs Formation, the uppermost formation in the Cambrian sequence. The Cheshire, Monkton, and Danby formations constitute terrigenous wedges that thin to the east. The source area for these formations was probably the Middle Proterozoic Grenvillian basement of the Adirondack massif to the west or northwest (Cady, 1945).

The Lion Hill sulfide deposits are hosted by the Lower Cambrian Monkton Quartzite and the Winooski Dolomite. The Monkton Quartzite was described by Cady (1945) to be predominantly interbedded, immature red and blue-gray quartzite that has intercalated layers of reddish and purplish shale and gray to pink dolomite. The overlying Winooski Dolomite is mainly a pink dolomite that has siliceous partings and is gray and mottled near the top of the section. The Monkton grades laterally into the underlying Dunham Dolomite and overlying Winooski Dolomite to the east and south where the two carbonate units are conformable (Cady, 1945).

Base-metal sulfides at Lion Hill are concentrated in elongate, stratabound zones in the upper part of the Monkton Quartzite near its contact with the Winooski Dolomite. The sulfide minerals occur predominantly as metamorphic segregations and as discordant masses that are interpreted to be epigenetic. Some sulfides, though, also occur as relict interstitial minerals that may have formed during diagenesis (Clark, this volume). Recrystallization of the sulfides may have taken place as a result of the regional tectonic deformation during the Taconian orogeny (~440 Ma, Zen, 1967) and possibly during the younger Acadian orogeny (Crosby, 1963). Chlorite-grade regional metamorphism has been documented to the north of Lion Hill by Osberg (1952) and to the south of Brandon by Zen (1960). Cady (1945) described the eastern portion of the Monkton Quartzite as a schistose rock. More recently, strongly deformed rocks that have greenschist-facies metamorphic assemblages have been identified in drill core from Lion Hill (Clark, this volume). Hand samples of sulfide-bearing drill core have phyllitic layers that show a strong schistosity that parallels the fabric of the host quartzite.

The origin of lead-zinc mineralization at Lion Hill is largely obscured by deformational and metamorphic overprinting. Fluid trapped in inclusions in recrystallized sphalerite and calcite suggests that a fluid phase was present during metamorphism of the Lion Hill deposit. Sinha and others (1986) have demonstrated for mylonites that the presence of a fluid phase during deformation may result in increased solution transport, microfracturing, and recrystallization predominantly parallel to the tectonic layering of the rock. Structural deformation accompanying regional metamorphism of both sedex and MVT deposits may result then in coarse, recrystallized sphalerite and galena strung out along shear zones and will localize remobilized sulfides along shear zones, faults, fold axes, and axial surfaces, such as is the case for the Lion Hill mineralization.

The two main mineralized zones at Lion Hill consist of discontinuous stratabound lenses approximately 120 and 180 m in length along strike and 30 and 60 m thick, respectively (Clark, this volume). Both lenses are elongate subparallel to the dip of the axial planes of the major folds and are subparallel to bedding. Scattered occurrences of sphalerite, galena, and chalcopyrite also are found in the Lion Hill anticline to the north of the two main lenses. Fragments of layered magnetite iron formation that contain abundant sphalerite and chalcopyrite and minor pyrite and galena were found on the dump of the Nicklaw mine (Clark, this volume).

FLUID INCLUSION STUDY

Coarse, recrystallized sphalerite was the only mineral that contained fluid inclusions suitable for analysis. The 20- to 50-μm-wide inclusions fall into two groups. The majority of the inclusions are secondary in nature and are aligned along healed microfractures; others of the same general size range are isolated or occur in small groups and may be primary. Both primary and secondary inclusions yielded approximately the same ranges of temperature and salinity values, implying that the two generations of inclusions formed from similar fluids.

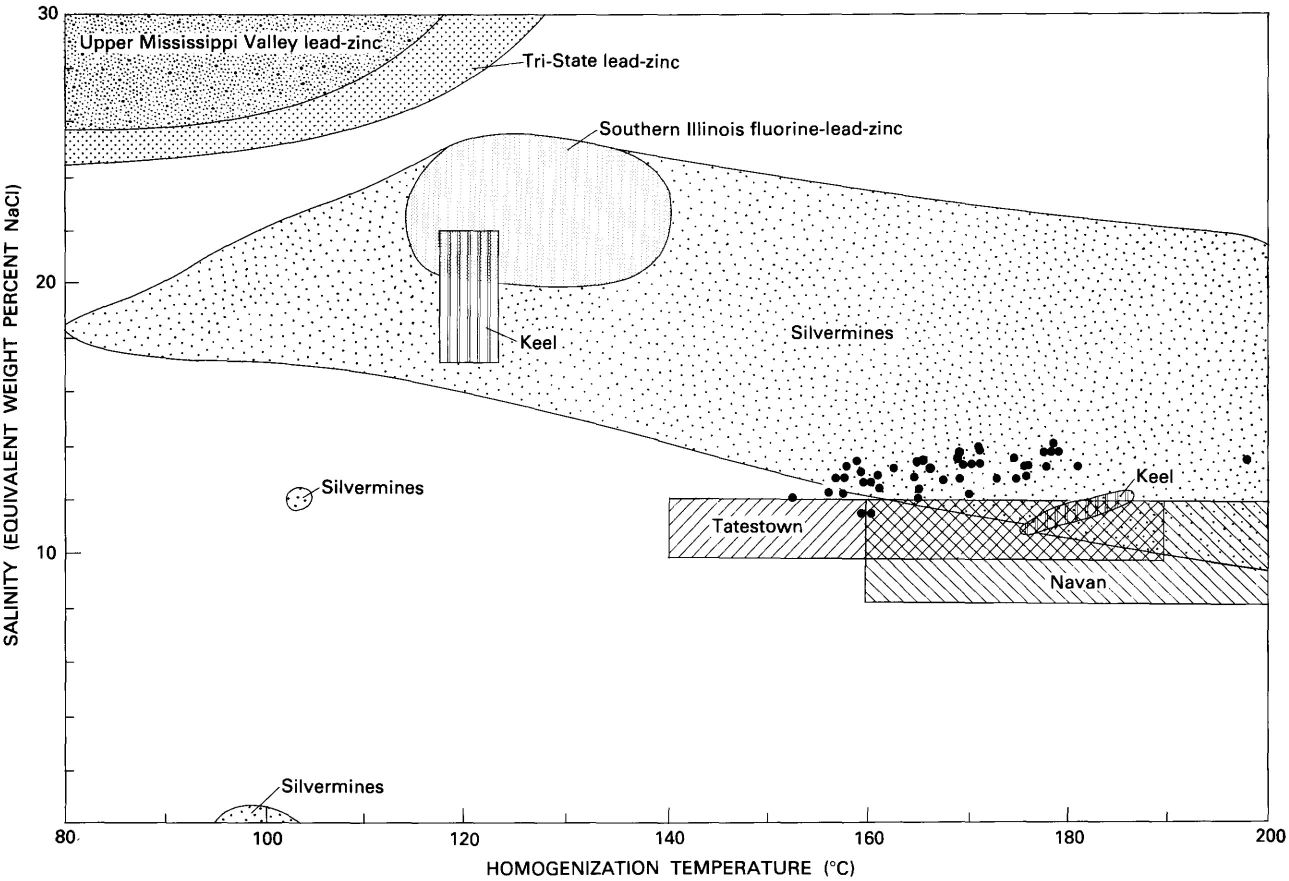

Figure 1. Plot of homogenization temperature versus salinity for fluid inclusions in sphalerite from the Lion Hill deposit (filled circles). Fluid inclusion data for Mississippi Valley-type deposits, including the Upper Mississippi Valley lead-zinc district, the Tri-State lead-zinc district, and the Southern Illinois fluorine-lead-zinc district (Roedder, 1976), and Irish carbonate-hosted deposits, including the Silvermines zinc-lead-copper deposit, Ireland (Sampson and Russell, 1984); the Navan and Tatestown deposits, Ireland (Andrew and Ashton, 1985); and the Keel deposit, Ireland (Roedder, 1976; Caulfield and others, 1986), are plotted for comparison. Data for both primary and secondary inclusions are plotted for the Mississippi Valley-type deposits.

Fluid inclusion thermometry results for discordant sphalerite from Lion Hill are plotted on figure 1. Homogenization temperatures of primary inclusions in sphalerite range from 152 to 196 °C. Measurements of the bulk salinity of the same inclusions, based on temperatures of the melting point of ice, yielded values of 11.5 to 14 equivalent weight percent NaCl. Eutectic temperatures of about −25 to −30 °C suggest that sodium is the dominant cation in solution. The lack of daughter minerals in the inclusions indicates that the solutions are undersaturated with respect to the common salt components even at room temperature.

The measured inclusions occur in sphalerite that was deformed and recrystallized during the Taconic orogeny (Clark, this volume). Fluid trapped in these inclusions most likely represents the fluid present during the peak metamorphic event, because most preexisting inclusions, formed when the ores were originally deposited, would not have survived the deformation and recrystallization. The homogenization temperatures reported here are uncorrected for pressure because the pressure of trapping is unknown, and stratigraphy and fluid systematics place only minimal constraints on the depth of mineralization. During Middle Ordovician time, prior to Taconic thrusting and deformation, at least 1,000 m of sedimentary rock overlay the Monkton Quartzite in the vicinity of Lion Hill (Cady, 1945). However, shallow depth estimates of less than 3 to 4 km indicate geothermal gradients that are in excess of the peak gradients established for the area during either the Taconic or Acadian orogeny. Most simple estimates of pressure correction are based on the effects of hydrostatic versus lithostatic load. In the case of Lion Hill, given the concurrent regional metamorphism, an open system cannot be assumed. The pressure exerted on the growing crystals may have been only a function of local pore fluid pressure because the system, unless completely plastic, was probably isolated from the regional lithostatic load.

The fluid systematics of the Lion Hill mineralization may be best understood by considering the regional metamorphic environment. Rocks in the vicinity of Lion Hill have been metamorphosed to chlorite grade and, locally, to biotite grade (Cady, 1945). Coarser sulfide minerals occur as metamorphic segregations along shears and fractures and in small veinlets; larger pods of mineralization have small halos of recrystallized carbonate. All of these factors indicate that the sulfide minerals were extensively recrystallized during deformation and metamorphism.

At Lion Hill, the low regional metamorphic grade suggests that the fluids present in the analyzed inclusions may have been generated by metamorphic dewatering of the enclosing sedimentary rocks. The fluid compositions obtained for Lion Hill are appropriate for the local rock and mineral assemblages. The lack of visible CO_2 in the fluid inclusions and the lack of graphite in the country rocks are consistent with the regionally low metamorphic grade at Lion Hill and suggest that no additional fluid was generated in situ by decarbonation reactions. The fluid preserved in the inclusions in sphalerite at Lion Hill may represent a combination of intercrystalline fluid, derived by the release of fluids trapped within inclusions in precursor sulfides and gangue, and interstitial pore fluids. The paleofluids preserved during lithification may have been remnants of basinal brines that characterize such continental shelf environments worldwide. Additional fluids may have been added by migration of ground water to the site during metamorphism.

Primary inclusions from unmetamorphosed MVT deposits homogenize at temperatures that range from about 100 to 150 °C (for example, Roedder, 1976) for the main ore-forming stages (fig. 1). The salinity of the ore fluid in such MVT deposits is typically much greater than 15 equivalent weight percent NaCl and commonly exceeds 20 equivalent weight percent NaCl. Daughter crystals of halite are rarely found in MVT deposits; their scarcity suggests that there are appreciable quantities of ions other than sodium in the ore fluids. Low-grade metamorphic overprinting of an MVT deposit may raise the homogenization temperatures of primary inclusions by decrepitation and recrystallization and also result in higher temperature secondary inclusions. The salinities probably will remain high, and the high salinities reflect the original compositions of the ore and pore fluids, unless the flow of metamorphic fluid was adequate to disperse the saline fluids (Roedder, 1984).

Many zinc-lead-silver deposits of Ireland, including Tynagh, Silvermines, and Keel, are thought to be sedimentary-exhalative or diagenetic in origin, although most have associated epigenetic mineralization. Russell and others (1982) have proposed a sedimentary-exhalative origin for the Irish deposits, while others, for example, Hutchinson (1983), feel that the ores are clearly epigenetic. Taylor and Andrew (1978) have suggested a hybrid model for the origin of the Silvermines deposit; this model has stratigraphically lower epigenetic ores deposited in a feeder zone and upper ores deposited syngenetically from exhalations near the sediment-seawater interface. In support of this model, Sampson and Russell (1984) found that fluid inclusions in ore and gangue minerals from the Silvermines deposit homogenized at temperatures ranging from 50 to 260 °C and had salinities primarily from about 0 to 24 equivalent weight percent NaCl (fig. 1). The Silvermines ores are thought to have been deposited when a hot, metal-bearing hydrothermal fluid (250 °C, 12 equivalent weight percent NaCl) encountered a colder, near-surface submarine brine pool (low temperature, 25 equivalent weight percent NaCl). Caulfield and others (1986) extended this hybrid model to explain the epigenetic ores at the Keel deposit; they stressed the dependence of syngenetic, diagenetic, and epigenetic processes of ore deposition on the rate of fluid expulsion and local geologic controls. Their conclusions are based, in part, on the work of Roedder (1976), who measured homogenization temperatures of 175 to 185 °C and salinities of ~11 to 12 equivalent weight percent NaCl for the Keel deposit (fig. 1). Caulfield and others (1986) collected additional data on primary fluid inclusions in sphalerite from the deposit. These data yielded homogenization temperatures of 117 to 123 °C and salinities of 17 to 22 equivalent weight percent NaCl (fig. 1). Andrew and Ashton (1985) also cite values of 8 to 12 equivalent weight percent NaCl and 140 to 200 °C for the synsedimentary-syndiagenetic Navan and Tatestown deposits of Ireland (fig. 1).

Low-grade metamorphism of the Irish sedex deposits may result in higher fluid inclusion temperatures and would probably homogenize the salinities to some intermediate value, unless substantial amounts of ground water were added. The high salinities of some of the Irish deposits are thought to result from mixing between a lower salinity proto-ore fluid and a submarine brine pool (Taylor and Andrew, 1978). Sedex deposits that form in similar environments, but without a brine pool, may not attain such high salinity values, and, if metamorphosed, the values would remain moderately low.

The analyzed fluid in inclusions from Lion Hill may represent a mixture of the precursor ore fluid and a metamorphic intercrystalline fluid. The measured salinities and relative cation proportions of the Lion Hill inclusions are similar to those of the zinc-lead-silver deposits of Ireland, including parts of the Tynagh, Silvermines, and Keel deposits. An MVT precursor, in contrast, would require the addition of a much lower salinity fluid to yield the fluid compositions observed in the Lion Hill inclusions. However, element mobility in the fluid phase, and hence bulk composition of the fluid, may be affected by processes such as compaction, brine expulsion, metamorphism, and so on. The field of data shown on figure 1 for the Lion Hill deposit may have resulted from metamorphic recrystallization and mixing of a low-temperature–high-salinity MVT fluid and a high-temperature–low-salinity fluid (~400 °C, 0

Table 1. Lead isotope compositions of galenas from Lion Hill, Vermont

[Lead isotope uncertainties are given as percent of the ratios at the 95 percent confidence limits. Numbers in locality column refer to drill hole location and depth in feet from the surface]

Sample	Locality	$^{206}Pb/^{204}Pb$	$^{207}Pb/^{204}Pb$	$^{208}Pb/^{204}Pb$	Lithology of wallrock
1	B–31–50′	18.351±0.03	15.546±0.02	38.275±0.04	Monkton Quartzite: Quartzitic dolostone that has micaceous partings and quartz, sphalerite, galena, and pyrite in veins.
2	B–31–560′	18.463±0.01	15.552±0.02	38.368±0.04	Monkton Quartzite: Quartzitic dolostone that has micaceous partings.
3	B–26–240′	18.496±0.01	15.592 ±0.02	38.406±0.02	Monkton Quartzite: Quartzitic dolostone that has sphalerite, galena, and pyrite.
4	B–25–87′	18.483±0.02	15.587±0.02	38.405±0.04	Monkton Quartzite: Calcareous quartzite that has phyllitic layers and partings and biotite, galena, sphalerite, and pyrite in veins.
5	B–17–692′	18.378±0.02	15.551±0.02	38.126±0.04	Monkton Quartzite: Dolostone that has green phyllitic layers and sphalerite, galena, calcite, and quartz in veins.
6	L–3–84.5′	18.362±0.01	15.607±0.01	38.403±0.06	Winooski Dolomite: Dolostone that has phlogopite, biotite, pyrite, and galena in veins.
7	L–2–84′	18.442±0.02	15.583±0.02	38.465±0.04	Monkton Quartzite: Dolostone that has green phyllitic layers and sphalerite and galena in thin, fine-grained zones.
8	B–12–526′	18.374±0.01	15.566±0.01	38.318±0.02	Monkton Quartzite: Phyllitic dolostone that has sphalerite and galena along shears.
9	B–28–531′	18.632±0.01	15.618±0.02	38.496±0.04	Monkton Quartzite: Quartzite that has phyllitic and micaceous layers and dolomite, quartz, sphalerite, and galena associated with the layers.
10	MIF[1]	18.253±0.02	15.536±0.02	38.229±0.04	Magnetite iron formation that has magnetite, galena, sphalerite, chalcopyrite, and minor pyrite in pods that are separated by stringers of quartz, feldspar, and micaceous minerals.

[1] Magnetite iron formation, float sample found near Nicklaw shaft.

equivalent weight percent NaCl). A simpler, and perhaps more likely, explanation, based on relative cation proportions, is an origin by homogenization of a precursor sedex deposit. However, given the inclusion data presented in this study, the author cannot distinguish unequivocally between either a sedex or an MVT origin for the precursor sulfides of the Lion Hill deposit.

LEAD ISOTOPE STUDY

The lead isotopic compositions of galenas from the Lion Hill zinc-lead-copper deposit (table 1, fig. 2) range from 18.253 to 18.632 for $^{206}Pb/^{204}Pb$, from 15.536 to 15.618 for $^{207}Pb/^{204}Pb$, and from 38.126 to 38.496 for $^{208}Pb/^{204}Pb$. The data cannot be used to constrain either the age of the mineralization or the age of the source because the samples do not show a wide enough range in lead isotopic compositions.

Two lines of evidence suggest that some of the spread in $^{207}Pb/^{204}Pb$ values may be controlled locally. Galena from the Winooski Dolomite has a higher value of $^{207}Pb/^{204}Pb$ for a given value of $^{206}Pb/^{204}Pb$ than galena hosted by the Monkton Quartzite. This difference suggests that the Winooski Dolomite may be more radiogenic in $^{207}Pb/^{204}Pb$ than the Monkton Quartzite. The difference in radiogenic nature may be expected if the lead in the Monkton Quartzite was derived originally from relatively unradiogenic feldspars of the Middle Proterozoic Grenvillian basement in the Adirondack or Green Mountains (for example, Zartman and Wasserburg, 1969). Differences in lead isotopic composition of galena from deposits of western Newfoundland were also noted by Swinden and others (1988). They found that lead isotopic values increased with stratigraphic position in a Cambrian-Ordovician sequence and interpreted this trend to indicate mixing of leads from Grenvillian basement and the platformal sedimentary sequence.

The variation in lead isotopic values among samples from a single mineralized lens is less than the variation between lenses, including for the scattered sulfide occurrences north of the main lenses. For example, galenas from the mineralized lens in the Fay syncline are slightly less radiogenic and more restricted in lead isotopic composition than are galenas from the lens in the Churchill anticline. Galenas from scattered sulfide occurrences to the north of the main zones define an isotopic field that encompasses all galena analyses from the main lenses (fig. 2).

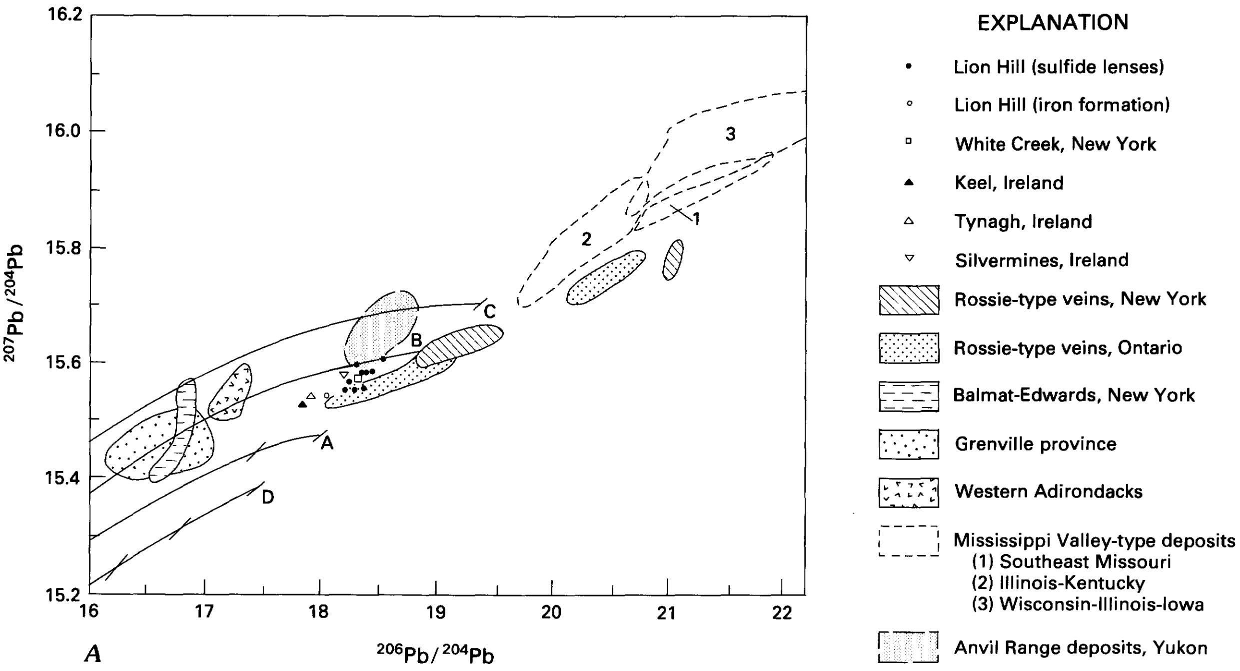

Figure 2. Lead isotope data for galenas from the Lion Hill deposit (stratabound sulfides and iron formation) and the White Creek, N.Y., deposit (stratabound sulfides). (*A*) Plot of $^{207}Pb/^{204}Pb$ versus $^{206}Pb/^{204}Pb$, (*B*) $^{208}Pb/^{204}Pb$ versus $^{206}Pb/^{204}Pb$. For comparison, lead data for galenas are also shown for the Silvermines, Keel, and Tynagh deposits, Ireland (Caulfield and others, 1986); the Rossie-type veins of northwestern New York (Ayuso and others, 1987) and southwestern Ontario (Fletcher and Farquhar, 1982a); the Balmat-Edwards, New York (Doe, 1962; Fletcher and Farquhar, 1982b); the Southeast Missouri lead-zinc district (Doe and Delevaux, 1972); the Illinois-Kentucky and Wisconsin-Illinois-Iowa districts (Heyl and others, 1966); and the Anvil Range deposits, Yukon (Shanks and others, 1987). Rock lead data for the Grenville province and the western Adirondacks are from Zartman and Wasserburg (1969). Curves on both plots are the average lead environments of Doe and Zartman (1979): A, mantle; B, orogene; C, upper crust; and D, lower crust.

Galena from a magnetite iron formation found as float on a mine dump at Lion Hill (the Nicklaw mine) also was analyzed for lead isotopic composition. In addition to galena and magnetite, the iron formation contains abundant sphalerite and chalcopyrite and minor pyrite in pods that are separated by stringers of quartz, feldspar, and micaceous minerals. The galena from the iron formation is isotopically similar to other Lion Hill galenas (table 1) and is only slightly less radiogenic for $^{206}Pb/^{204}Pb$ (18.253), $^{207}Pb/^{204}Pb$ (15.536), and $^{208}Pb/^{204}Pb$ (38.229).

The lead isotopic values for the discordant (epigenetic?) mineralization at Lion Hill and for the iron formation are comparable to those of a small stratabound lead-zinc deposit at White Creek, N.Y. The White Creek deposit (J.F. Slack, oral commun., 1986; Slack and others, 1987) is 80 km to the south of Lion Hill within the Taconian sequence of continental margin, slope-rise facies slate and graywacke that ranges in age from Late Proterozoic to Middle Ordovician (Zen, 1967); the deposit is thought to lie within the Mudd Pond Quartzite Member of the Nassau Formation of Early Cambrian age (see Potter, 1972). Lead isotopic values for a galena from the White Creek deposit are 18.424 for $^{206}Pb/^{204}Pb$, 15.563 for $^{207}Pb/^{204}Pb$, and 38.264 for $^{208}Pb/^{204}Pb$.

Compared to the growth curves of the plumbotectonic model of Doe and Zartman (1979), the lead isotopic data for the Lion Hill galenas overlap the orogene curve (B on fig. 2) and define a field that extends toward the mantle curve (A on fig. 2). The less radiogenic compositions of the White Creek and Lion Hill galenas are similar to those of stratabound sulfide deposits formed in epicontinental basin-and-platform or continental-margin environments (for example, Rammelsberg, West Germany, Doe and Zartman, 1979), although there are some important differences.

Lead isotopic ratios of the Lion Hill galenas are fairly radiogenic relative to the age of the host rocks and display a range in values of $^{206}Pb/^{204}Pb$, $^{207}Pb/^{204}Pb$, and $^{208}Pb/^{204}Pb$. The lead isotopic values of galenas from the Lion Hill sulfide lenses and iron formation and from the White Creek sulfide deposit all yield model lead-lead ages of less than 250 million years. These ages are much younger than the ages of the host Monkton and Winooski Formations (Early Cambrian), the inferred age of the iron formation (Early Cambrian), and the age of the Taconian sequence (Cambrian-Ordovician). However, the calculated lead-lead ages are model dependent, and lead from these deposits may have had a more complex isotopic evolution than that assumed by the model.

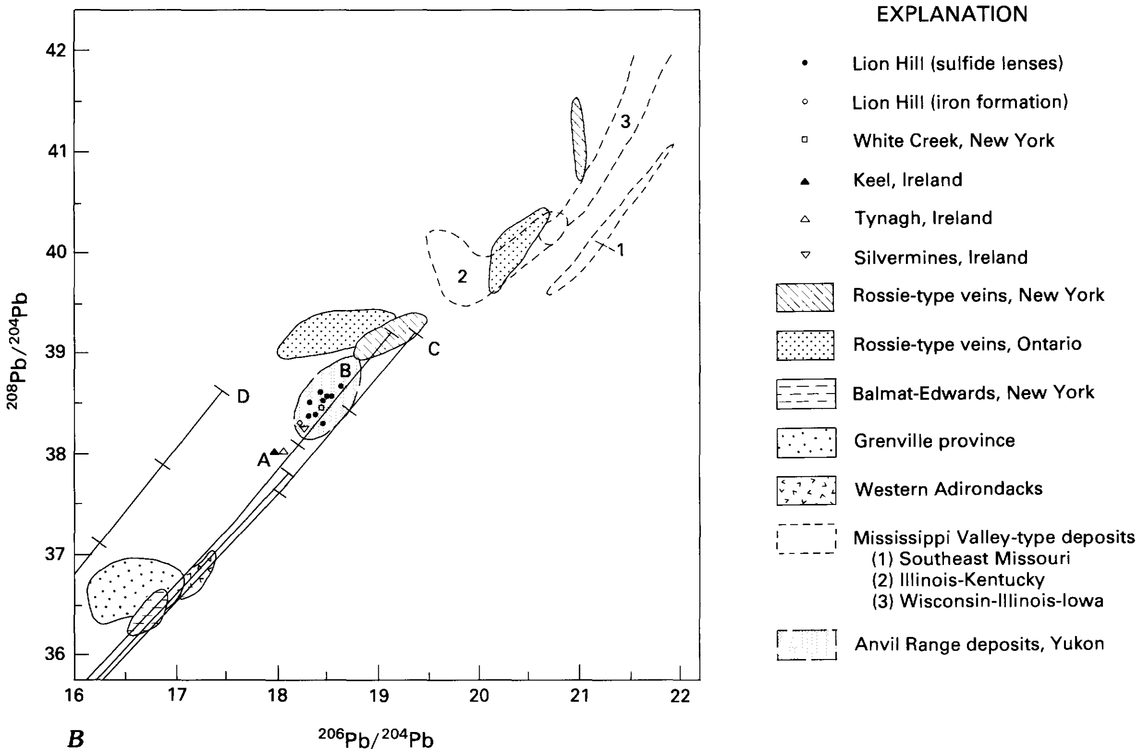

Figure 2. Continued.

Average lead isotope growth curves for sedex-type mineral deposits have been empirically derived for a number of specific, well-defined tectonic environments. For example, Godwin and Sinclair (1982) have constructed an average lead isotope growth curve for shale-hosted, zinc-lead deposits of the Canadian Cordillera. The model has been used to date a number of sedex deposits in the Selwyn basin and in the Ogilvie, Richardson, Wernecke, and Mackenzie Mountains of western Canada. If sufficient lead isotopic data were available for deposits and for host and basement rocks in the northeastern United States, an average lead growth curve could be constructed for the region; this curve might give model ages for the Lion Hill and White Creek deposits that more closely approach the ages of the host rocks.

The discrepancy between the model lead ages and the ages of the host rocks for these deposits may be explained also by contamination of the initial lead isotopic composition during metamorphism. Metamorphosed sedex deposits commonly show evidence of contamination with varying amounts of radiogenic lead (for example, Sullivan mine, Kanasewich, 1968). A more detailed lead isotopic study of the White Creek deposit may show that the lead values measured in this study reflect the addition of radiogenic lead, thus explaining the discrepancy between model lead ages and the age of the host rocks.

Classic MVT ore deposits contain extremely radiogenic and heterogeneous lead isotope values (for example, Doe and Delevaux, 1972), as do some vein deposits in the northeastern United States and southeastern Ontario (for example, Rossie-type veins, Ayuso and others, 1987; Fletcher and Farquhar, 1982a). In contrast, lead isotopic values of galenas from individual carbonate-hosted, base-metal deposits in Ireland exhibit little heterogeneity (although there may be large differences between deposits), and the values are relatively unradiogenic compared to the Carboniferous age of the host rocks (Caulfield and others, 1986). Some shale-hosted, exhalative deposits of Canada, however, show a wider range of lead isotopic values than do the sedex deposits of Ireland (Godwin and Sinclair, 1982). For example, galenas from the Howards Pass deposit have a spread in $^{206}Pb/^{204}Pb$ and $^{208}Pb/^{204}Pb$ values that is similar to those shown for the Lion Hill galenas. Cambrian sedex deposits of the Anvil Range district in the Canadian Yukon (Shanks and others, 1987) also show a similar spread in lead isotopic composition. The marble-hosted Balmat zinc ores in the Grenvillian province of New York State recently have been described as a metamorphosed MVT deposit (Whelan and others, 1984), and galenas from Balmat, like those from Lion Hill, are isotopically heterogeneous. The Lion Hill galenas, like those of the Irish deposits (fig. 2), however, differ from Balmat, Anvil Range, and classic MVT galenas in that they have relatively low $^{207}Pb/^{204}Pb$ values. These low values suggest that any involvement of an upper crustal lead component was limited.

The similarities between the lead isotopic compositions of galenas from the Lion Hill and the White Creek deposits suggest that galena in veins and shear zones at Lion Hill may have formed by local metamorphic remobilization of preexisting lead-zinc-iron sulfides that have lead isotopic compositions similar to that of the unmetamorphosed White Creek galenas. The Lion Hill iron formation also may have been contaminated with radiogenic lead during metamorphism.

SUMMARY

(1) The mineralogy, fluid chemistry, and lead isotopic compositions of the Lion Hill, Vermont, lead-zinc-copper deposits are similar to those of many sediment-hosted, stratabound lead-zinc deposits, especially in Ireland, and bear important similarities to other examples in the general class of sedimentary basin-related mineral deposits.

(2) The discordant mineralization at Lion Hill may have formed by recrystallization of preexisting diagenetic lead and zinc sulfides during deformation of the host rocks.

(3) The most likely age for the formation of the discordant Lion Hill mineralization is syn- or post-Taconian (for example, Middle Ordovician), but precursor lead and zinc sulfides probably existed in the shelf sediments during Early Cambrian time.

(4) The most important result of studying the Lion Hill mineralization may be identifying the environment as favorable for the occurrence of potentially high-grade sedex-type lead-zinc-silver deposits.

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Chapter M

Some Controls on Trace-Element Concentrations, Especially Uranium, in Selected Peat Deposits of Vermont and New Hampshire

By CORNELIA C. CAMERON, PAUL G. SCHRUBEN, and
EUGENE L. BOUDETTE

U.S. GEOLOGICAL SURVEY BULLETIN 1887

SUMMARY RESULTS OF THE GLENS FALLS CUSMAP PROJECT, NEW YORK, VERMONT,
AND NEW HAMPSHIRE

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Some Controls on Trace-Element Concentrations, Especially Uranium, in Selected Peat Deposits of Vermont and New Hampshire

By Cornelia C. Cameron,[1] Paul G. Schruben,[1] *and* Eugene L. Boudette[2]

Abstract

Thirty-eight deposits of Holocene peat were sampled in the Vermont-New Hampshire area of the Glens Falls 1° × 2° quadrangle. The peat deposits were studied to assess the ability of some deposits to accumulate anomalous amounts of certain elements, including Ni, Co, Cu, V, Sr, La, Nd, Eu, Th, and U. The amount of each element in a given deposit depends on the presence of available sources in the underlying bedrock and on the efficiency of ground-water transport from these sources to the peat-forming marsh or swamp. The presence of bedrock fracture zones and solution channels, such as in karst, facilitates element transport.

Uranium is the most geochemically anomalous element and in 79 samples ranges in concentration from 1.0 to 467.0 parts per million and has a mean of 48.28 parts per million. Deposits that have the largest uranium anomalies are located (1) just south of Fern Lake, Vermont, in a swamp that contains an estimated 40,000 tons of commercial quality dry peat and (2) in a marsh east of Sunapee Lake near New London, New Hampshire, where the peat resource is estimated also at 40,000 tons. This report describes where the uranium occurs in the deposits and the relations of the deposits to ancient ecological environments or ecosystems that may influence or control uranium concentrations.

Uranium contents of the peat deposit near Fern Lake are similar to the lowest ore-grade portions of the Flodelle Creek, Washington, deposit, the only surficial uranium mine in the United States. Like the Washington State deposit, the uraniferous peat deposits in Vermont and New Hampshire are relatively nonradioactive. Because these deposits have postglacial ages of less than 10,000 years, daughter products such as radium and thorium have not yet formed in appreciable amounts. Agricultural users of peat should be alerted to the uranium content of these samples.

INTRODUCTION

Thirty-eight deposits of Holocene peat in Vermont and New Hampshire were studied during a mineral resource assessment of the Glens Falls 1° × 2° quadrangle. For the purpose of discussion, core sites of these deposits are divided into five groups, A–E (fig. 1). Group A deposits are located west of the Green Mountains in Vermont and overlie lower Paleozoic sedimentary and metasedimentary rocks, chiefly dolomite and quartzite. The deposits in group B are found east of the Green Mountains and overlie Proterozoic granitic rocks and quartzite and lower and middle Paleozoic metasedimentary and minor metavolcanic rocks. The deposits in group C occur on both sides of the Connecticut River in Vermont and New Hampshire and overlie phyllite, schist, gneiss, and granite of Devonian age and older. The deposits in group D are located in New Hampshire and overlie Devonian two-mica granite and quartz monzonite. Also located in New Hampshire, group E deposits overlie granite, granitic gneiss, and various metasedimentary rocks.

This report examines the relations between ash and trace-element contents and the geological settings of the peat deposits and the ecological environments that existed during deposition. Controls on trace-element concentrations, especially uranium, are also evaluated.

ACKNOWLEDGMENTS

The authors express their gratitude to Joan Esterle, a graduate student from the University of Kentucky, who provided valuable field assistance. Z.A. Brown, C.J. Skeen, W.B. Crandell, and A.F. Dorrzapf analyzed samples by using emission spectrography. D.M. McKown and R.B. Vaughn obtained various chemical data for this report by using the delayed neutron activation method in the U.S. Geological Survey's radiochemistry laboratory in Denver. Their help is gratefully acknowledged.

METHODS AND RESULTS

Samples for laboratory analysis were obtained from cores taken from the surface to the base of each of the 38

Manuscript approved for publication April 13, 1990.

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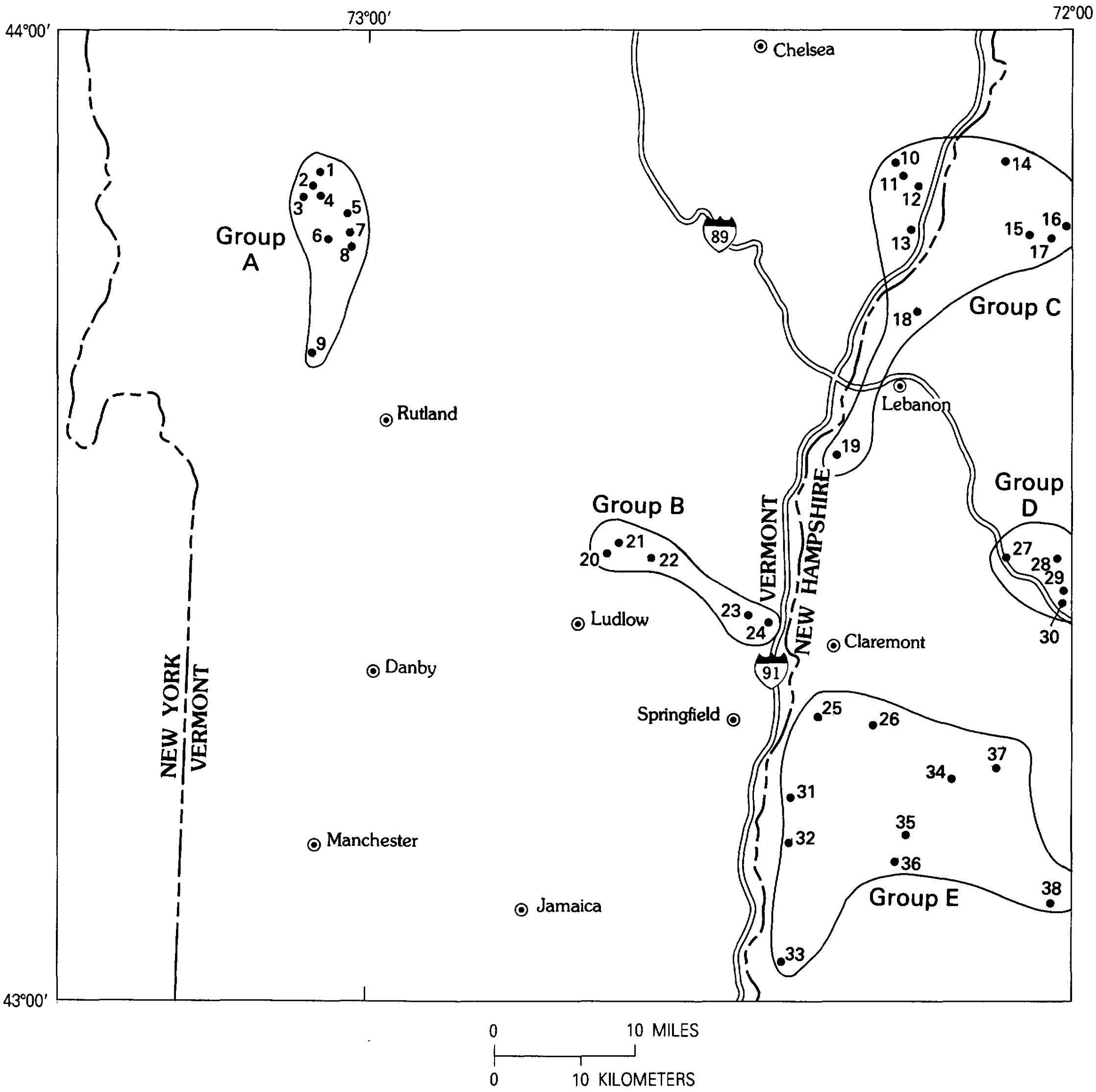

Figure 1. Locations of numbered core sites in Vermont and New Hampshire. Sites have been divided into five groups, A, B, C, D, and E.

deposits by using a Macaulay peat auger. In the laboratory, ash content on a dry basis was determined for each sample. This ash was used later for emission spectrographic analyses. Uranium and thorium data were obtained from the unashed fractions of the samples by using the delayed neutron activation method. Statistics for the entire data set are shown in table 1. In these peat deposits, elemental concentrations are considered to be anomalous at or above a threshold of two standard deviations above the mean. Some of the elements that occur in anomalous amounts are Ni, Co, Cu, V, Sr, La, Nd, Eu, Th, and U. Of these elements, the most anomalous is uranium; in 79 samples, uranium ranges in concentration from 1.0 to 467.0 parts per million (ppm) and has a mean of 48.28 ppm (table 1).

The anomalous elements shown in table 2 are arranged according to the locations of core site groups. Uranium and iron are anomalous only in core groups A and D; calcium and molybdenum are anomalous only in groups A and C; gallium is anomalous only in core site groups D and E; and vanadium is anomalous only in core groups C and D. These trends reflect (1) the types of bedrock, fracture zones, and solution channels through which ground water circulates and (2) the interaction of the organic and inorganic constituents of the depositional environment shown in the core stratigraphy[3] of the deposits (see fig. 2). Groups A and D are discussed in more detail because these are the principal localities where anomalous concentrations of uranium occur in the peat deposits. The focus is on two specific deposits: (1) Messer Pond (core site 30) near New London, N.H., under a marsh covered with heath that rests

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[3] Original measurements in cores were made in feet so this chapter uses English rather than metric units for core descriptions and sections.

Table 1. Ash and element statistics

[Ash is reported on dry basis. Elements are reported in ppm on whole dry sample basis, except for iron, magnesium, calcium, and titanium, which are reported in percent]

Element	Minimum	Maximum	Mean	Deviation	Number of valid samples
Ag	0.006	0.98	0.11	0.12	139
B	.84	86.00	19.34	20.10	139
Ba	18.70	607.62	153.09	126.41	139
Be	.01	4.31	.83	.86	89
Ca	.08	42.72	2.11	5.35	139
Ce	.80	254.94	39.41	42.81	96
Co	.11	51.44	5.01	7.01	139
Cr	.26	82.26	14.00	16.44	139
Cu	.59	131.43	19.17	21.04	139
Eu	.20	10.61	1.49	1.55	68
Fe	.02	5.73	.95	1.03	139
Ga	.03	16.87	3.49	3.78	124
La	.03	206.38	39.06	38.50	139
Mg	.02	2.58	.32	.37	139
Mn	13.39	2,349.16	218.66	308.21	139
Mo	.09	39.47	4.11	4.67	139
Nb	.08	20.57	3.94	3.74	71
Ni	.45	147.13	17.23	21.23	139
Nd	1.30	267.14	57.00	58.25	97
Pb	.29	35.55	7.19	6.82	120
Sc	.13	18.69	4.62	3.90	139
Sr	9.43	405.49	93.18	86.50	137
Th	2.24	99.20	16.51	12.00	15
Ti	.00	.69	.11	.12	139
U	1.00	467.00	48.28	55.24	79
V	.59	112.18	20.61	20.47	139
Y	.28	72.84	16.16	14.59	139
Yb	.02	8.34	1.93	1.69	139
Zn	.40	273.23	55.63	46.97	132
Zr	1.24	233.70	42.73	42.59	137
Ash	1.74	97.92	45.84	27.31	

Table 2. Elements that have anomalous concentrations, arranged according to core site group locations

[The term anomalous refers to those elements that have concentrations at least two standard deviations above the mean]

Element	Core site group				
	A	B	C	D	E
Ba					
Fe	X			X	
Mg	X		X	X	X
Ca	X		X		
Ti				X	
Mn	X			X	
Ag	X			X	X
B		X		X	X
Ni			X		X
Be			X	X	X
Co			X		X
Cr			X	X	X
Cu	X	X	X		
La	X		X	X	X
Mo	X		X		
Nb		X	X	X	X
V			X	X	
Pb	X		X	X	X
Sc			X	X	X
Sr	X		X	X	
Ce			X	X	X
Y	X		X	X	X
Zn			X		X
Zr	X			X	X
Yb	X		X	X	X
Ga				X	X
Nd					X
Eu	X		X	X	X
Th	X			X	X
U	X			X	

on a stratigraphy above glacial drift overlying two-mica granite of Devonian age and (2) along Route 53 south of Fern Lake (core site 1), Brandon Township, Rutland County, Vt., in a predominantly hardwood swamp in karst topography over glaciated Cambrian dolomite. The uranium (U_3O_8) in these deposits is of postglacial age ($<$10,000 yr old) and is not yet in secular equilibrium with daughter products (for example, radium and thorium). The deposits are therefore relatively nonradioactive. However, given time and stability, the parent uranium and associated daughter elements will evolve in the direction of equilibrium and make the deposits more radioactive.

MESSER POND, NEW HAMPSHIRE, AREA

The Messer Pond deposit (fig. 2) contains an estimated 40,000 tons of air-dried moss and reed-sedge peat that have ash contents of less than 25 percent, which is the ash cutoff for commercial quality peat. Core samples from this deposit have a uranium content of 217 ppm in these peats. In the underlying peaty clay, uranium values reach 291 ppm. Figure 2 shows core 30 and the sample locations in relation to core stratigraphy. Sample material can thus be interpreted in terms of environments of deposition and trends in ash and uranium contents.

The Messer Pond area is underlain by the informally named Sunapee granite of Late Devonian age (about 366 Ma; J.N. Aleinikoff, oral commun., 1989). The Sunapee is an evolved, two-mica granite and was emplaced as a north-northeast-trending, postkinematic intrusive sheet less than 1 km thick (fig. 3). To the west, the Sunapee passes into either metamorphosed to upper amphibolite facies semipelite (undivided Silurian and Devonian rocks on fig. 3) or into granite of the Bethlehem Gneiss in the Mount Clough pluton. To the east, the Sunapee granite is in contact with granite of the Kinsman Quartz Monzonite that comprises the Cardigan pluton. The Sunapee granite is younger than all of these enclosing rocks (Chapman, 1952).

A regional study of uranium distribution in the Lower Devonian rocks of New Hampshire was made by Lyons

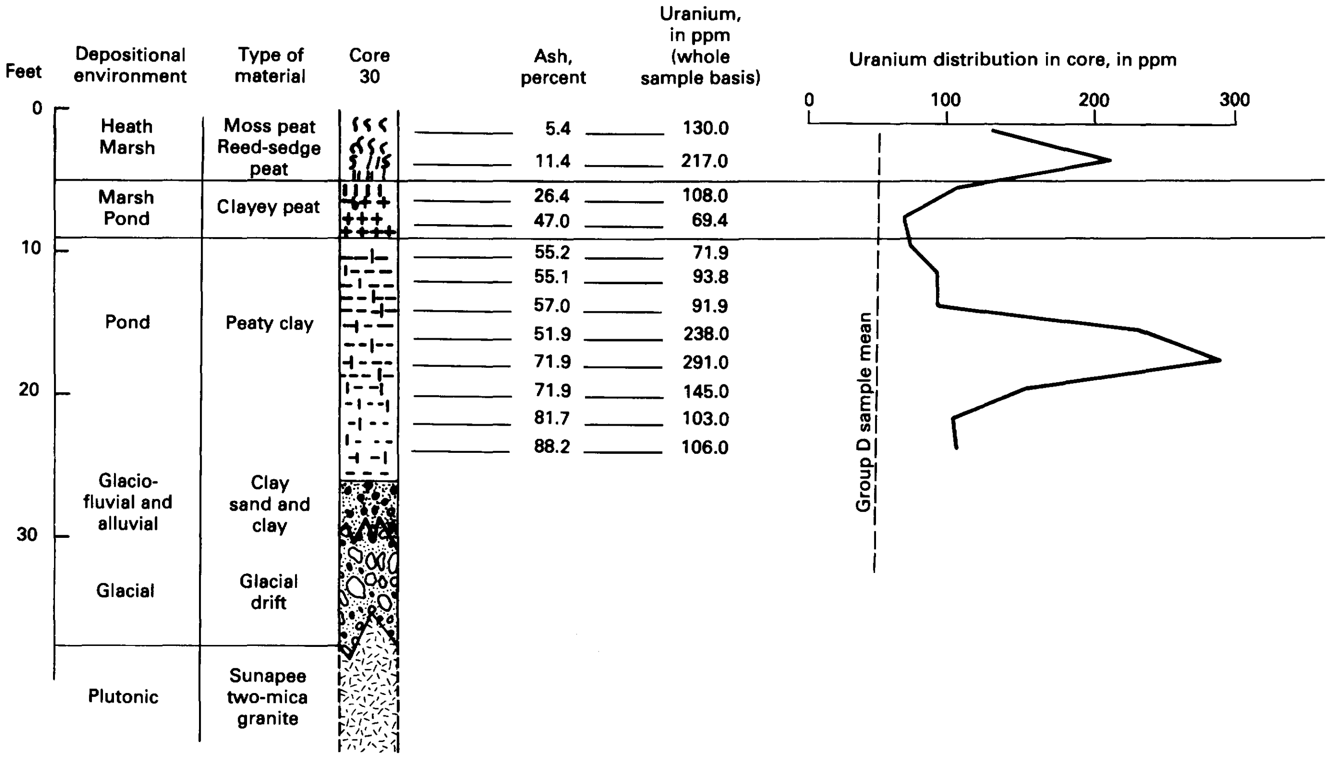

Figure 2. Core 30 (Messer Pond, N.H., deposit) showing the relation of types of materials to depositional environments and to ash and uranium contents. The uranium distribution curve from the surface to the base of the deposit is also shown.

(1964). Secondary uranyl-phosphate uranium minerals occur within the Sunapee granite in an upland roadcut on Interstate Highway I-89, near the town of New London (Boudette, 1977). This occurrence is about 1.61 km from the Messer Pond marsh (fig. 4) and was mapped in detail by Bothner (1978), who established its spatial context within the Sunapee granite. Thorium and uranium distributions in the Sunapee granite are given in table 3.

The U.S. Department of Energy (1982a), as part of the National Uranium Resource Evaluation Program (NURE), supported reconnaissance work to study the uranium content of water and modern sediments in the United States. Ground water at Messer Pond was determined to contain 29 parts per billion (ppb) uranium and 24 ppb uranium at nearby Clark Pond (fig. 4). The concentration of uranium in the pond waters ranges from 0.64 to 1.0 ppb. Another NURE Report (U.S. Department of Energy, 1982b) on the Glens Falls quadrangle includes detailed descriptions of uranium occurrences.

During the 10-yr period after construction of I-89, secondary uranium minerals were mostly leached away by solution and the movement of vadose ground water. This leaching was probably enhanced by acid rain and snow and reinforced by the saline spray from road salting operations. The rapid mobility of uranium in this area led the authors to suspect that Holocene peat deposits below the water table would be ideal sinks for the uranium transported in ground water. This conclusion prompted the study of the peat deposits at the four core sites in group D (fig. 1).

Core 30 at Messer Pond (fig. 2) was studied in the greatest detail and has a total depth of 25 ft. Commercial quality peat, containing less than 25 percent ash, accumulated to a depth of 5 ft in a marsh environment on which a heath developed. This commercial quality peat rests on clayey peat that has ash contents of 25 to 50 percent. This clayey peat represents the base of the marsh and the top of the pond ecological environments. Below the clayey peat is peaty clay that extends to a depth of 24 ft and has ash contents that exceed 50 percent. This peaty clay is the base of the ancient Holocene pond deposit that rests on glaciofluvial sand and clay that overlies glacial drift that rests on the Sunapee two-mica granite.

Uranium content reaches a maximum of 217 ppm in the center of the marsh and again at 291 ppm in the center of the buried pond peaty clay (fig. 2). Trends in the ash and uranium concentration curves at core site 30 are similar to those trends at core site 29 in marshes and heaths that overlie two-mica granite and quartz monzonite.

Nuclear magnetic resonance (NMR) spectra (fig. 5) for ^{13}C obtained from the four samples in the upper 8½ ft of core 30 are correlated with the depositional environments and uranium contents shown on figure 2. Carbohydrates in the organic material decrease with depth, whereas paraffinic contents increase with depth (fig. 5); both are related to the

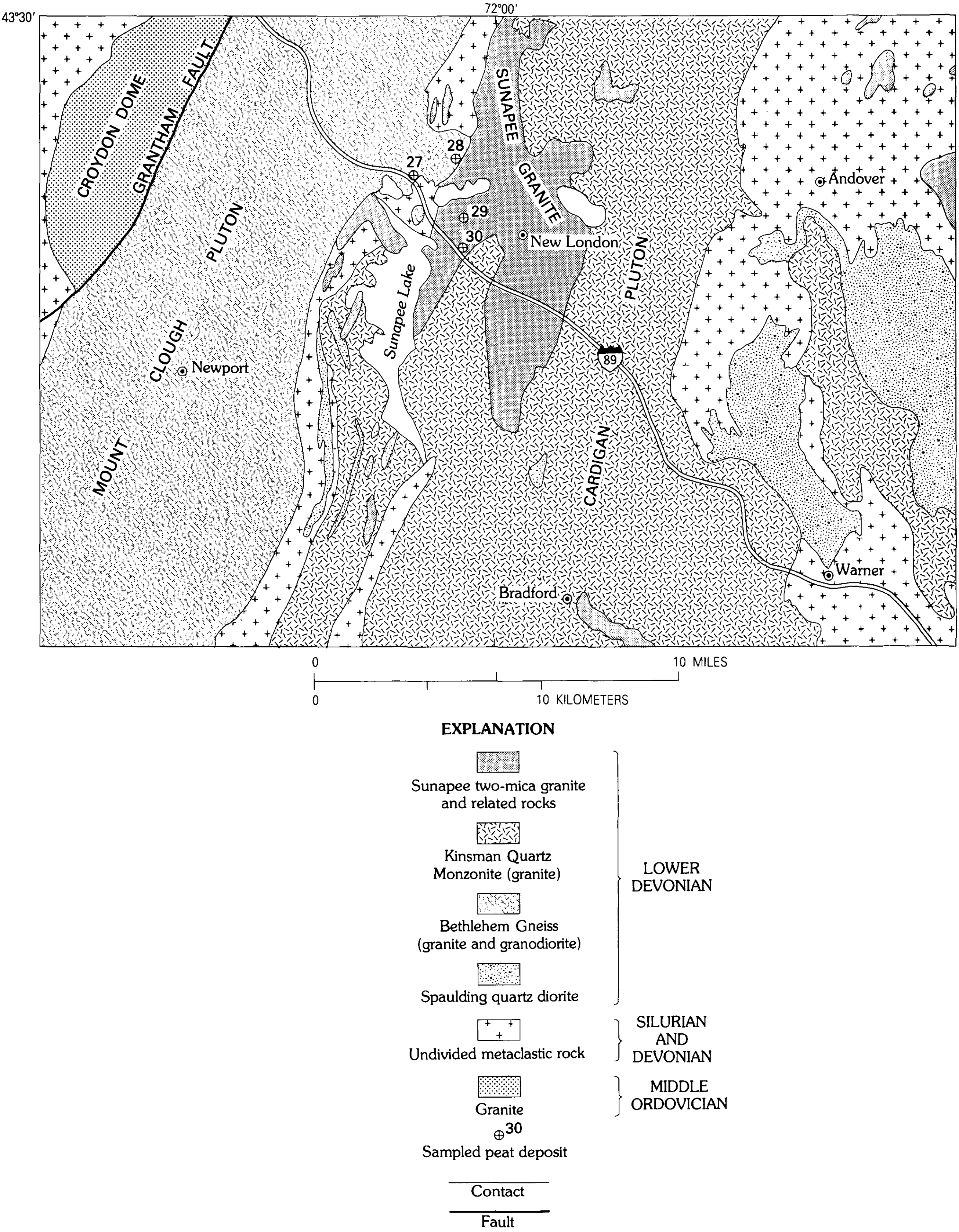

Figure 3. Bedrock substrate of peat deposits at New London, N.H. Geology after Lyons and others, unpublished State map of New Hampshire.

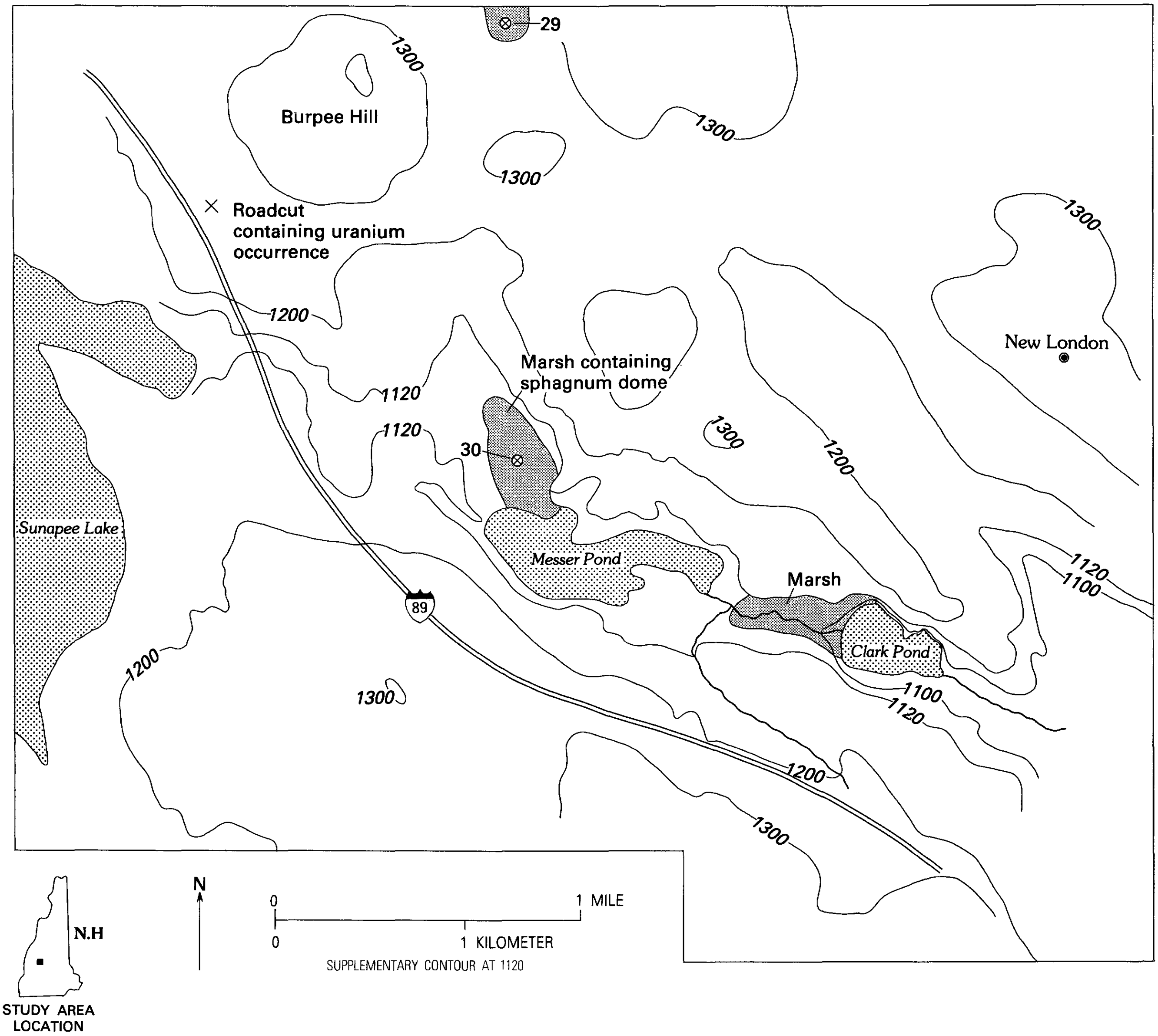

Figure 4. Messer Pond, N.H., area showing location of core 30 in relation to roadcut that contains the uranium occurrence.

aromatic contents as diagenesis progresses. Uranium contents reach a maximum above the base of the marsh (217.0 ppm, figs. 2 and 5) where the paraffinic and carbohydrate peaks are nearly at the same elevation. A relation appears to exist between the ecological environments of organic accumulation and the diagenesis within these environments. An experimental study of algal uptake of U, Ba, V, Co, and Ni from dilute solutions by Mann and Fyfe (1984) reveals uptakes of uranium and barium to be 1,000 to 10,000 ppm by dry weight. The uptake was by two species of freshwater green algae; these plants are common in marsh and pond environments where peat accumulates. Berthelin and Munier-Lamy (1983) also describe microbial mobilization and preconcentration of uranium from various rock materials by fungi. The processes that may be involved in uranium preconcentration and recovery apparently depend on energistic processes of the plant cell, which are influenced by various environmental factors (Berthelin and others, 1985). Their experience indicates that uranium solubilization depends not only on rock or mineral characteristics but also, especially, on the extent of humus formation (induced by vegetation type) and associated microbial activity. Also, uranium adsorption and insolubilization is more effective on ferric oxides and microbial components than on clays and nontransformed plant materials. Gueniot and others (1982) observed uranium fixation on the surface of iron oxihydrox-

Table 3. Thorium and uranium distribution in the Sunapee granite

[Analyses were made by instrumental neutron activation methods]

Sample number	Latitude	Longitude	Th	U	Th:U
S-1[1]	43°24′53″N.	72°01′23″W.	7.2	25.4	0.28
S-2[1]	43°24′52″	72°01′23″	4.9	2.2	2.23
MK-107[1]	43°23′32″	71°58′48″	4.2	3.4	1.24
MK-1[1]	43°21′33″	71°59′48″	8.1	8.1	1.00
MK-108[1]	43°23′51″	71°57′47″	1.5	1.5	1.00
SU-U1a[2]	43°24′50″	72°01′50″	8.6	18.7	.46
Su-U1b[2]	do.	do.	8.2	13.3	.62
Su-U1b[2]	do.	do.	9.4	31.8	.30

[1] Analyses by P.A. Baedecker, U.S. Geological Survey, Reston, Va.

[2] Analysis by D.M. McKown, H.T. Millard, Jr., and M. Scheider, U.S. Geological Survey, Denver, Colo.

ides in placic horizons of hydromorphic soils. Note that anomalous amounts of iron and uranium occur together in core site groups A and D (table 2).

FERNVILLE, VERMONT, AREA

The Fernville, Vt., peat of core site 1 (fig. 6) lies in a swamp in karst topography. This deposit contains an estimated 40,000 tons of air-dried reed-sedge peat. The stratigraphy of the core is as follows: 0–1 ft = muck; 1–3 ft = reed-sedge peat that has an ash content of 22.5 percent and a uranium content of 467 ppm; 3–5 ft = gray clay; 5 ft = sand.

Marsh and swamp vegetation has always predominated at core site 1. The ancient pond environment that existed in the New Hampshire deposit is absent, however. Uranium in ground water enters the Vermont peat deposit and is captured by organic matter as (1) a part of life processes and (2) after plant death during diagenesis. Decay or diagenesis is more severe in the muck than in the reed-sedge peat where plant remains are still recognizable.

The area around Fern Lake (core site 1) is underlain by the lower portions of the lower Paleozoic carbonate bank (Rodgers, 1968). The following descriptions of stratigraphic units are summarized from Cady (1945). The Cheshire Quartzite is largely a pure, massive, white quartzite about 400 ft thick. The upper portion, which is about 50 ft thick, comprises interbedded vitreous orthoquartzite and minor dolomite that grade into the overlying Dunham Dolomite. The lower portion of the unit is somewhat argillaceous. The Dunham Dolomite, which can exceed 1,700 ft in thickness, is primarily a siliceous, buff-weathering, massive dolomite that contains irregularly distributed, well-rounded sand grains. The upper portion of the

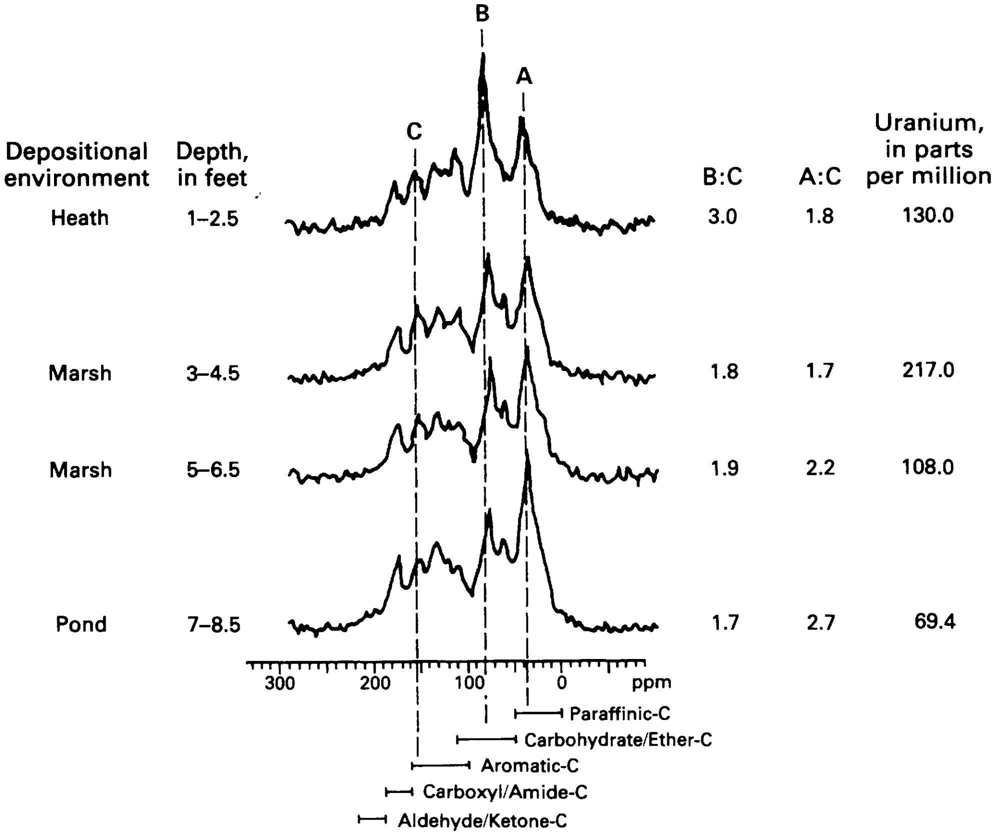

Figure 5. Nuclear magnetic resonance spectra for the upper 8½ ft of core 30, Messer Pond marsh. Analysis by Patrick Hatcher, U.S. Geological Survey, Reston, Va.

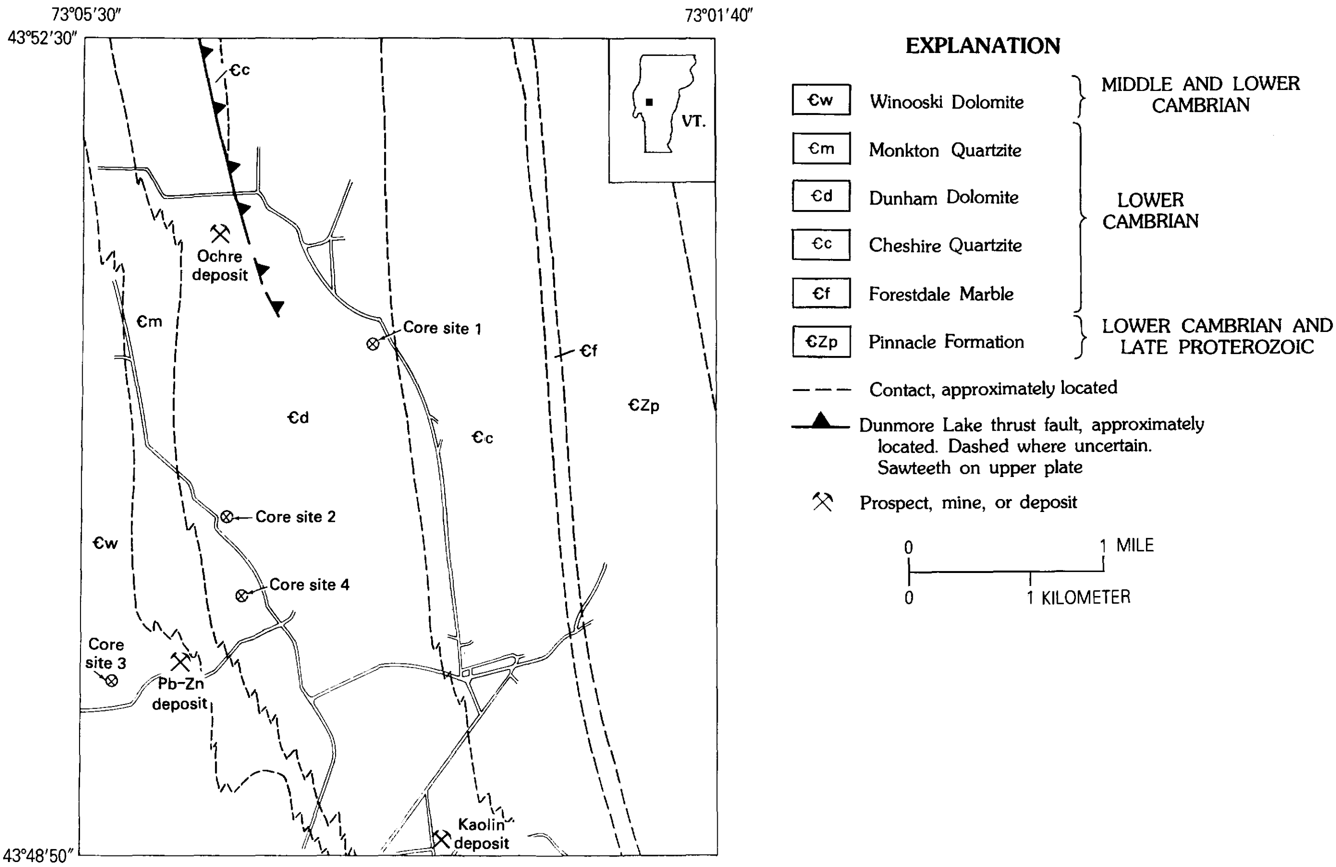

Figure 6. Bedrock geologic map showing locations of core sites 1–4. Geology modified from Cady (1945) and Doll and others (1961). Base adapted from U.S. Geological Survey, Brandon quadrangle, 1988.

Dunham, which contains numerous quartzite beds, grades into the overlying Monkton Quartzite.

The "Brandon residual-formation" unconformably overlies the Dunham Dolomite and Cheshire Quartzite in a zone that extends from the southern end of Lake Dunmore in Vermont to about 1.61 km south of Forestdale, Vt. Clark (1891) indicated that this surficial unit occurs over a large area extending from Vermont to Georgia. The "Brandon residual-formation" was described by Burt (1931[?], p. 115) as "a series of unconsolidated deposits consisting of kaolin, ocher, quartz sand, iron and manganese ores, and lignite." Berry (1919) suggests an Eocene age for the "Brandon," while Barghoorn and Spackman (1949) postulate that it cannot be older than late Miocene. The aggregation of materials suggests that the environment of deposition was a swamp that existed during Tertiary time. Dale (1904) suggests that most of the original source material for the unit was Precambrian detritus, derived from the Green Mountains.

Cady (1945) shows the Lake Dunmore thrust fault near the Ferndale swamp area (core site 1), and Stanley (1980) describes a regional system of high-angle faults in this area. Fracture systems associated with these faults influence circulation of uranium-bearing ground water that feeds the Ferndale swamp. The uranium in the area may have several sources. Proterozoic lithologies of the Green Mountains are known to host disseminated uranium concentrations as well as uraninite-bearing veins (Grauch and Zarinski, 1976; Ayuso and Ratté, this volume). The bedrock uranium occurrences are south of the Fernville area, but the Proterozoic terrane east of Fernville has not been thoroughly evaluated for uranium potential. Uranium deposits near Ludlow (Okemo Mountain) and Jamaica, Vt. (College Hill and Pinnacle Hill), occur in Proterozoic gneiss and schist. Stratabound uranium-bearing zones in quartzofeldspathic schist and gneiss are associated with Proterozoic granite (augen gneiss) and tourmaline quartzite (Ayuso and Ratté, this volume). Elsewhere in Vermont and in other New England States, the Cheshire Quartzite contains anomalous amounts of uranium (R.I. Grauch, written commun., 1985). Lignites and other sediments of the ancient swamps, as implied by the occurrence of the "Brandon residual-formation," may have received uranium in much the same manner that the peat of the Ferndale swamp is receiving uranium today. Thus, the "Brandon," as well as the Proterozoic and Paleozoic rocks, may contribute uranium to the ground water that is moving through and into the karst that contains the Ferndale swamp.

Uranium contents of the peat deposit near Fern Lake (core site 1) are similar to those of the lowest ore-grade portions of the Flodelle Creek, Wash., deposit. The peat deposit at Flodelle Creek (Johnson and others, 1987) is the only surficial uranium mine currently operating in the United States. Future work in the Fernville area may define a systematic distribution of highly uraniferous peat and a possible resource potential for uranium.

DISCUSSION

Uranium is commonly associated with peat, and laboratory studies indicate that adsorption, ion exchange, and reduction all play a role in the mechanism of uranium fixation (Nakashima and others, 1984). Otton and Zielinski (1985) studied a Holocene deposit of clayey peat and peaty clay (ash contents of 40 percent) and clay, silt, and sand that contained a slight amount of organic material in a swamp at Flodelle Creek, Wash. The swamp was fed by uraniferous waters. This deposit was laid down in marshes and swamps that were subject to flooding, which added silt throughout Holocene time. Otton and Zielinski (1985) suggest that the fixation of uranium by organic matter is controlled largely by processes of adsorption and ion exchange, rather than by reduction to relatively insoluble U^{+4}.

The Fernville, Vt., deposit is simpler, ecologically speaking, than the Washington State deposit because it has not been subject to flooding and has developed simply by growth and decay of vegetation. The Vermont deposit more readily lends itself to organic and geochemical studies concerning the role that living and dead plants play in the movement and fixation of uranium within a peat deposit. That ecological environments affect the movement and fixation of uranium is indicated by the peaks on the uranium curves in the Messer Pond, N.H. (fig. 2), deposit. The curve in the ancient marsh is repeated in the ancient pond deposit. A need for further studies of the role plants play in both pond and marsh ecosystems is indicated. Also, the unusual amounts of uranium in deposits that are potential peat resources should be studied, and members of the peat and other agricultural industries should be notified about the trace-element content of this commodity.

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Chapter N

Distribution of Gold, Molybdenum, Tin, and Selected Other Elements in the Vicinity of Mount Ascutney, Windsor County, Vermont

By LESLIE J. COX

U.S. GEOLOGICAL SURVEY BULLETIN 1887

SUMMARY RESULTS OF THE GLENS FALLS CUSMAP PROJECT, NEW YORK, VERMONT, AND NEW HAMPSHIRE

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FIGURES

Distribution of Gold, Molybdenum, Tin, and Selected Other Elements in the Vicinity of Mount Ascutney, Windsor County, Vermont

By Leslie J. Cox[1]

Abstract

A small, composite pluton of Early Cretaceous (122 Ma) age intrudes Middle to Late Proterozoic and early to middle Paleozoic metamorphic terrane at Mount Ascutney in southeastern Vermont. The igneous rocks are a comagmatic series of extrusive flows, tuffs, and breccias and intrusive gabbro, diorite, syenite, and granite. The eastern portion of the pluton consists mostly of hornblende syenite. The syenite stock intrudes middle Paleozoic country rock composed of calcareous quartz-mica schist and lesser amounts of marble, amphibolite, and sulfide-bearing graphitic schist. These rocks are part of the western flank of the Connecticut Valley-Gaspé synclinorium and have been interpreted as having both sedimentary and volcanic protoliths.

Locally anomalous levels of Au, Mo, and Sn, as well as Ba, Bi, B, Co, Cu, La, Pb, Mn, Nb, Ag, W, and Zn are reported in geochemical samples (rock, soil, stream sediment, and heavy-mineral panned concentrate) from the Mount Ascutney area. The distribution patterns of these elements are related to specific lithologies within the study area. A geochemically anomalous area south of the syenite stock is defined by gold-bearing samples that appear related to sulfide-bearing graphitic schists of the Waits River and Gile Mountain Formations. Concentrations of Mo, La, Nb, Sn, W, and Bi are associated with the eastern (mainly syenitic) portion of the Early Cretaceous pluton. The geologic setting and geochemistry of the study area indicate that there is low mineral resource potential for molybdenum in the syenite stock. Tin, tungsten, boron, and zinc appear to be associated with calcareous beds of the Waits River and Gile Mountain Formations within the contact metamorphic aureole of the syenite stock. A low mineral resource potential for tin, tungsten, and zinc in skarn deposits is therefore postulated.

INTRODUCTION

Detailed geochemical sampling within and around the Mount Ascutney pluton in southeastern Vermont was carried out in late July through early August 1984 as part of a multidisciplinary study of the Glens Falls 1° × 2° quadrangle, New York, Vermont, and New Hampshire, under the auspices of the U.S. Geological Survey's (USGS) Conterminous United States Mineral Assessment Program (CUSMAP). The purpose of the sampling was to investigate the source of anomalous values of gold detected by regional reconnaissance work in the Mount Ascutney area (K.C. Watts, Jr., written commun., 1984) and to determine the local distribution of other metal concentrations (including those of W, Sn, B, Pb, and Cu) discovered in the vicinity of Mount Ascutney during a regional geochemical survey of the Glens Falls 1° × 2° quadrangle (Day and others, 1986; Watts, this volume). A detailed geochemical investigation of the Mount Ascutney region also was considered important because of the recent recognition of bedrock occurrences of gold and other metals associated with a similar Mesozoic intrusive stock at Cuttingsville, Vt., located approximately 37 km west of Mount Ascutney (see Robinson, this volume).

GEOLOGIC SETTING

Mount Ascutney is a prominent circular topographic feature in southeastern Vermont, abruptly rising more than 600 m above the surrounding terrain. Little Ascutney Mountain, to the west, is less dramatic in height but is also regionally distinct in relief and arcuate in shape. Both features are underlain by a composite pluton of Early Cretaceous (122.2±1.2 Ma) age (Foland and Faul, 1977; Foland and others, 1985) that intrudes Middle to Late Proterozoic and early to middle Paleozoic gneiss and schist. The Mesozoic rocks (Daly, 1903; Chapman and Chapman, 1940) are a comagmatic series of extrusive flows, tuffs, and breccias and intrusive gabbro, diorite, syenite, and granite. The western stock consists mostly of gabbro and diorite (unit Mzd on fig. 1). The younger eastern stock is elliptical in shape and is composed mostly of hornblende-biotite syenite (unit Mzs on fig. 1). A small circular stock of pink biotite granite, classified as Conway Granite (unit Mzc on fig. 1) on the basis of chemical characteristics, intrudes the

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Manuscript approved for publication April 13, 1990.

[1] U.S. Geological Survey.

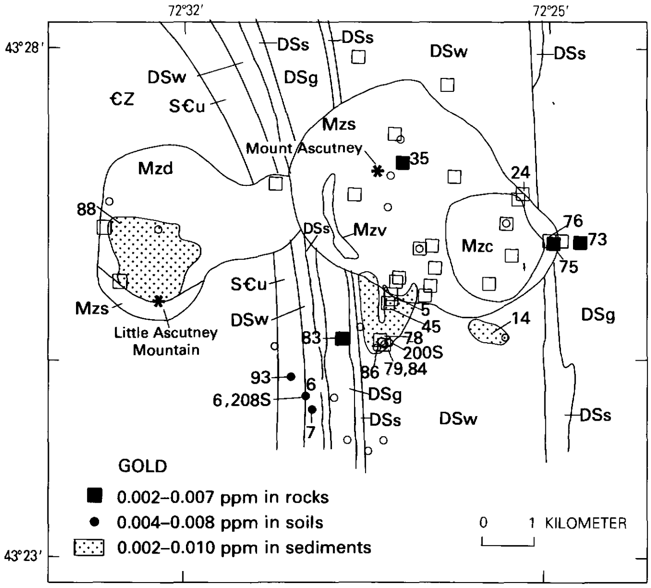

Figure 1. Distribution of gold in rock, soil, and stream sediment samples in the Mount Ascutney area. Geology is after Doll and others (1961) and Chapman and Chapman (1940). Map symbols are Mzc, Conway Granite; Mzs, syenite; Mzd, diorite and gabbro; Mzv, volcanic rocks; DSg, Gile Mountain Formation; DSs, Standing Pond Formation; DSw, Waits River Formation; S€u, Silurian to Cambrian rocks, undivided; Pz, Paleozoic rocks, undivided; €Z, Cambrian and (or) Late Proterozoic rocks. Locations of all rock and soil samples collected are shown by squares and circles, respectively. Filled squares and circles indicate that the rock and soil samples have values within the range given in the legend; open symbols indicate that the samples have values less than the lowest value in the range given in the legend. Locations of drainage basins that have values within the range given in the legend are shown by a stippled pattern. In areas of close sampling, a leader connects the sample number to the sediment sample site on the drainage. The sites of gold-bearing, panned glacial-till samples 200S and 208S (K.C. Watts, Jr., written commun., 1984) also are indicated. ppm=parts per million.

southeastern corner of the syenite. The main syenite stock contains large screens and xenoliths of volcanic rock, as well as small bodies of hornblende granite and numerous dikes (Daly, 1903; Chapman and Chapman, 1940). Recent gravity studies suggest that the syenite-granite stock has the shape of a vertical cylinder extending to a depth of 5.8 km, and that the diorite-gabbro stock may be a 0.3-km-long cylinder underlain by less dense syenite (see Daniels, this volume).

The plutonic igneous rocks (fig. 1) intrude Proterozoic biotite-muscovite gneisses and schists of the Chester dome (west of the diorite-syenite contact) and early to middle Paleozoic quartz-mica and calcareous-mica schists, impure marbles, impure quartzites, and amphibolites (east of the diorite-syenite contact). The Paleozoic schists constitute part of the western flank of the Connecticut River-Gaspé synclinorium, which is interpreted as having both sedimentary and volcanic protoliths (Hepburn and others, 1984). The country rocks were tightly folded and recrystallized by regional middle Paleozoic metamorphism. The carbonate-bearing rocks of the Silurian to Early Devonian Waits River Formation (Hueber and others, 1990) in the thermal aureole surrounding the syenite stock locally display a younger contact metamorphic assemblage superimposed on the earlier regional metamorphic facies (Nielson, 1973).

GEOCHEMICAL SAMPLING

The geochemical survey reported here was designed to replicate the gold anomalies identified in 1983 by K.C. Watts, Jr., and to more precisely define the area(s) of anomalous gold concentration. Four types of samples, as described below, were collected.

Fifty-four stream sediment samples and 21 panned concentrate samples of stream sediment were obtained from active streams in the vicinity of the intrusive complex. The sampled streams are typically between 0.6 to 3.7 km in length and have drainage basins generally less than 2.6 km^2 in area. For several streams, additional samples were collected farther upstream along the same drainage to determine if gold or other elements of interest occur in a higher, smaller part of the basin or if they are restricted to areas underlain by only one rock formation. Panned concentrate samples were collected midstream at sites where heavy minerals may be naturally concentrated due to a sudden drop of water velocity. Stream sediment samples also were collected midstream where mud and silt-sized sediment was deposited.

Thirty-four residual soil samples were collected from 22 locations, shown as open and filled circles in figures 1–7. At 11 of the 22 sites, two soil horizons were sampled. The first soil horizon was sampled within 2.5 cm beneath the organic debris on the surface; in places, this soil sample included fine roots. The second soil horizon was sampled at the first visible change with depth, commonly a color change (such as from brown to a shade of red, yellow, or gray) or, locally, a textural change (such as from loamy to clayey).

Thirty-eight rock samples were collected from outcrops shown as open and filled squares in figures 1–7. The rock samples include both apparently fresh rock and rocks that have visible alteration as indicated by the presence of sulfide phases, rusty-colored or quartz-filled fractures, atypical color, or unusual texture.

METHODS OF STUDY

All samples were analyzed for 31 elements by a six-step, D.C.- (direct current) arc, semiquantitative optical emission spectrographic method (Grimes and Marranzino,

1968) by G.W. Day and B.M. Adrian. In addition, each sample was analyzed by a quantitative atomic absorption technique for gold by E.P. Welsch, D.L. Kelly, and R.M. O'Leary. Methods of sample preparation and complete analyses and histograms of element concentrations (Au, Cu, Pb, Mn, Cr, B, Ag, Co, Ni, Zn, Ba, Sn, Mo, and La) for each of the four sample types are reported by Cox (1987).

For most elements, histogram intervals representing the upper 5 to 15 percent of the sample population closely correspond to the number of samples that have analytical values at or above the mean plus two standard deviations. Therefore, samples that have the highest reported analytical values are classified here as anomalous, the locations of which are shown in element distribution diagrams (figs. 1–7; Cox, 1987, fig. 3). A more strict definition of anomalous value is not attempted because of the extreme variation in chemical composition of the bedrock in the area and the correspondingly wide divergence of values for the average crustal abundance for any given element per rock type. The range of values representing anomalies for rock samples is wider than the 5 to 15 percent range that represents possible anomalies for the other three sample types. In general, those values reported as close to an order of magnitude above average crustal abundance or higher are considered anomalous for a given lithology; for a few rocks, values twice the average crustal abundance or greater are considered anomalous. More than 15 percent of the rock samples contain anomalous levels of some elements.

An exception to the general rules for identifying anomalous samples is made for tin in panned concentrate samples. The locations of panned concentrate samples that contain 30 ppm or greater of tin are shown in figure 2.

The use of more than one sample type enhances detection of element anomalies. For example, areas shown as having anomalous tin (fig. 2) are defined largely by the analyses of panned concentrate samples because anomalous values of tin are not reported for stream sediment samples collected from the same sites. Both tungsten and bismuth would have been missed entirely if panned concentrate samples had not been collected. Yet, analyses of panned concentrate samples failed to identify drainages that have anomalous levels of Au, Mo, Zn, Ag, Cr, or Co. Anomalous concentrations of each of these elements, however, were found in several areas where one or more of the other three sampling media were used.

DISCUSSION—DISTRIBUTION OF GOLD, MOLYBDENUM, TIN, AND OTHER ELEMENTS

Low, yet probably anomalous, levels of Au, Mo, Sn, as well as Ba, Bi, B, Co, Cu, La, Pb, Mn, Nb, Ag, W, and Zn are present in the samples collected from the Mount Ascutney area. Distribution patterns of the geochemically anomalous rock, soil, stream sediment, and panned concentrate samples support the following interpretations made about the relations between the types of bedrock and the element anomalies.

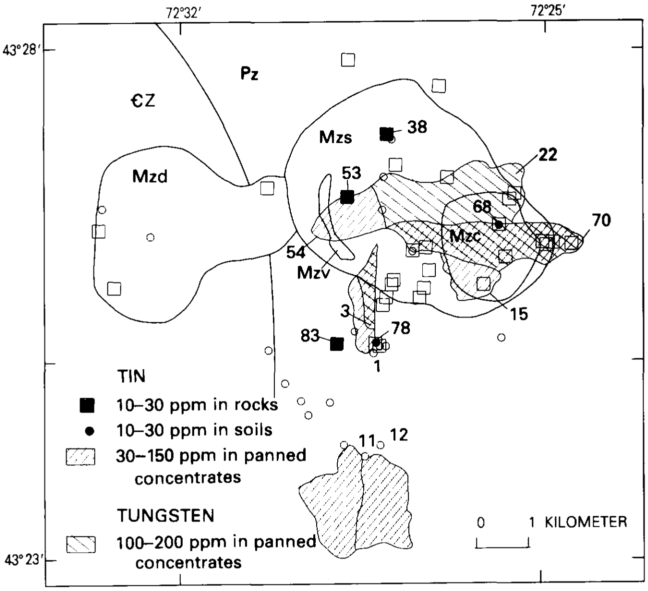

Figure 2. Distribution of tin and tungsten in rock, soil, and panned concentrate samples in the Mount Ascutney area. See figure 1 for explanation of abbreviations and symbols.

Although gold was not detected in any panned concentrate sample, low levels of gold were found in each of the other three sample types. Together these samples crudely define a gold-bearing area south of the syenite pluton at and surrounding the locations of the gold-bearing samples (200S and 208S on fig. 1) identified in a reconnaissance study by Watts (written commun., 1984).

Roughly 2.4 km south of the pluton, several soil samples contain trace amounts of gold. Two samples of residual soil (samples 6 and 7 on fig. 1), containing 8 and 4 parts per billion (ppb) gold, respectively, were collected above pyrite-bearing gray mica schist of the Waits River Formation in the vicinity of reconnaissance sample 208S. Roughly 1 km northwest of this site, 4-ppb gold is reported from bright red, clayey soil overlying the Missisquoi Formation (sample 93 on fig. 1). The Missisquoi Formation consists of carbonaceous schist and amphibolite. Here, the color and texture of the soil suggest that it may be residual after amphibolite. Because levels of 4 to 6 ppb of gold are common for unmineralized mafic igneous rocks (Turekian and Wedepohl, 1961; Gottfried and others, 1972), the amount of gold in this sample probably does not constitute an anomaly. The value of 8 ppb determined for sample 6 is, however, considered anomalous compared to the average of 2 ppb in soils reported by Brooks (1972).

Adjacent basins represented by stream sediment samples 86 and 79 (as well as sample 84, which is a replicate of sample 79) in the vicinity of reconnaissance sample 200S contain between 2 and 10 ppb gold. The absence of gold in stream sediment and panned concentrate samples collected farther upstream in both drainages (Cox, 1987) restricts the gold-bearing source of the drainage basins to the underlying Waits River Formation.

A rusty-black, pyrite-bearing graphitic schist of the Gile Mountain Formation (sample 83 on fig. 1) contains 2 ppb gold and is the only gold-bearing rock sample that is near one of the gold-bearing reconnaissance samples (200S). Lithologically similar schist of the Gile Mountain Formation (sample 73 on fig. 1) from the eastern edge of the Ascutney stock contains 7 ppb gold. A nearby sample of mineralized syenite (sample 75) contains accessory chalcopyrite and 5 ppb gold. All of these rock samples are judged to be anomalous in gold based on average values for sedimentary and igneous rocks (see Rose and others, 1979, p. 557). The isolated occurrence of mineralized syenite suggests that both gold and copper were derived originally from the adjacent Gile Mountain Formation or even from the Waits River Formation. Alternatively, gold may have been introduced into the Paleozoic host rocks adjacent to the syenite intrusion by Mesozoic igneous (or hydrothermal) activity (see Watts, this volume; Robinson, this volume); nevertheless, the area of gold enrichment does not appear to be within the syenite (fig. 1), and the data suggest that the sources of gold in the Ascutney region are sulfide-bearing graphitic schists of the Waits River and Gile Mountain Formations. Closely spaced soil and bedrock sampling and detailed mapping of fracture patterns at the southern edge of the syenite stock may reveal a more distinctive trend for this gold distribution.

South of the pluton, within north-flowing drainage basins that have no contacts with Early Cretaceous plutonic rocks, soils contain trace amounts of Au, as well as anomalous Pb, Cu, Ni, Cr, Co, Ba, and Mn (see Cox, 1987, for distribution patterns). Elements not found in anomalous amounts in any soil or sediment samples within these drainage basins include Zn, Mo, La, Bi, and W. Because concentrations of molybdenum, lanthanum, bismuth, and tungsten are anomalous only in samples that either completely or partially overlie the plutonic rocks, their presence appears to be related to the Early Cretaceous intrusions.

Samples that have anomalously high levels of molybdenum are located within the boundaries of the syenite stock. The molybdenum distribution almost completely excludes the biotite granite (Mzc) as a source (fig. 3). The distribution of anomalous lanthanum (fig. 4) closely parallels that of molybdenum. The distribution of anomalous manganese, where found overlying the syenite stock (fig. 5), is also coincident with samples that have anomalous contents of molybdenum. In this area also, some of the samples that have anomalous contents of lead (fig. 6) and zinc (fig. 7) are coincident with and peripheral to areas of molybdenum concentration (fig. 3). Anomalously high concentrations of lead and zinc, as well as of tin and tungsten, contribute to the regional geochemical signature characteristic of granite molybdenite systems (Ludington, 1986). If molybdenum-mineralized rocks at Mount Ascutney exist, they may be most prevalent in a 0.8- to 1.2-km-wide arcuate band that extends directly west and north from the contact of the biotite granite stock. This band is in an area defined by the coincident anomalies of Mo, La,

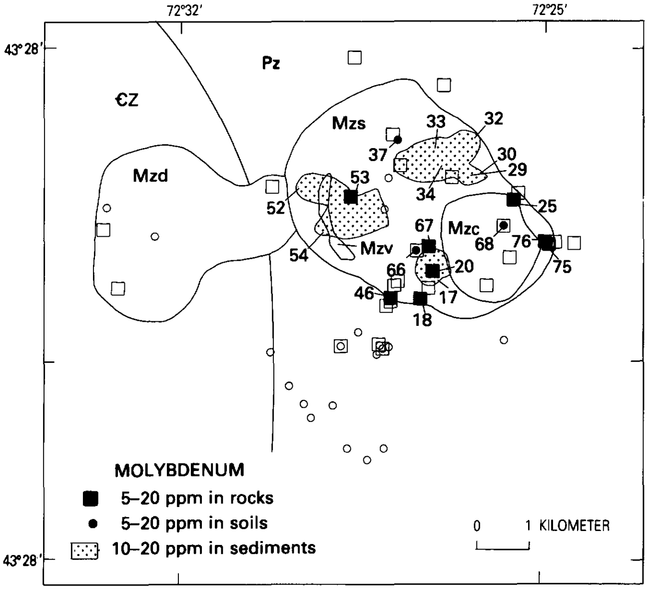

Figure 3. Distribution of molybdenum in rock, soil, and stream sediment samples in the Mount Ascutney area. See figure 1 for explanation of abbreviations and symbols.

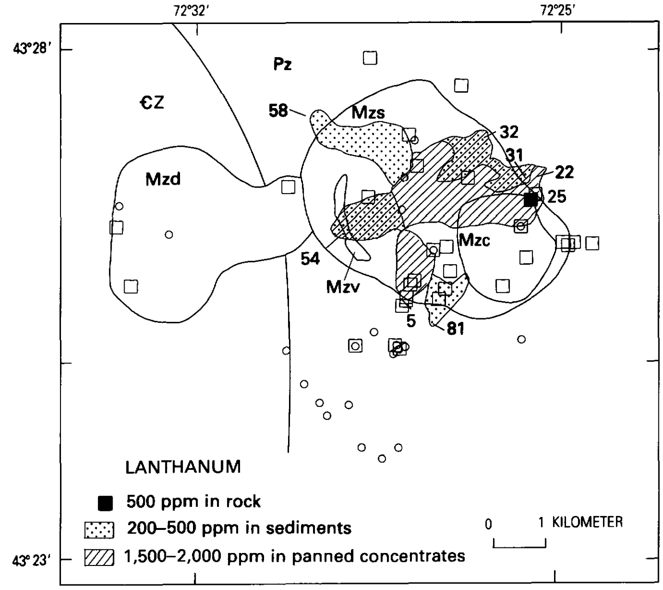

Figure 4. Distribution of lanthanum in rock, stream sediment, and panned concentrate samples in the Mount Ascutney area. See figure 1 for explanation of abbreviations and symbols.

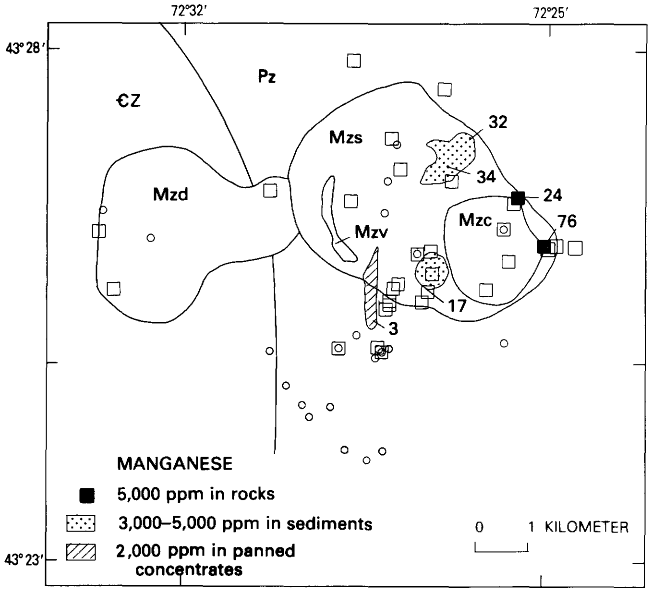

Figure 5. Distribution of manganese in rock, stream sediment, and panned concentrate samples in the Mount Ascutney area. See figure 1 for explanation of abbreviations and symbols.

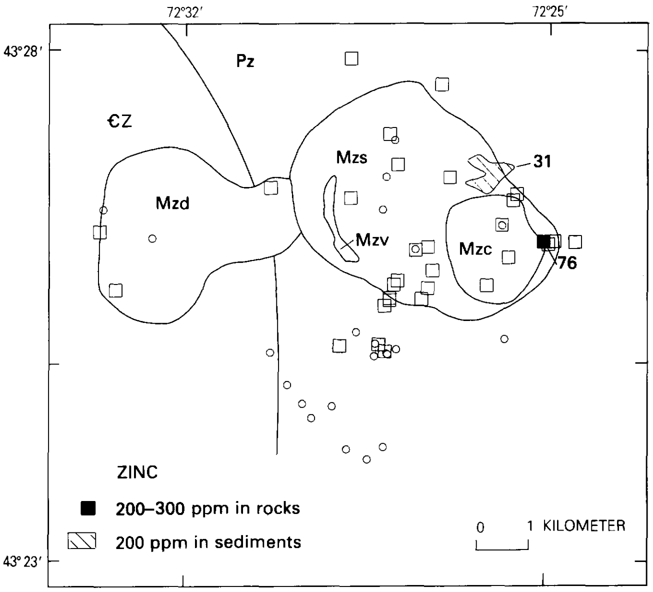

Figure 7. Distribution of zinc in rock and stream sediment samples in the Mount Ascutney area. See figure 1 for explanation of abbreviations and symbols.

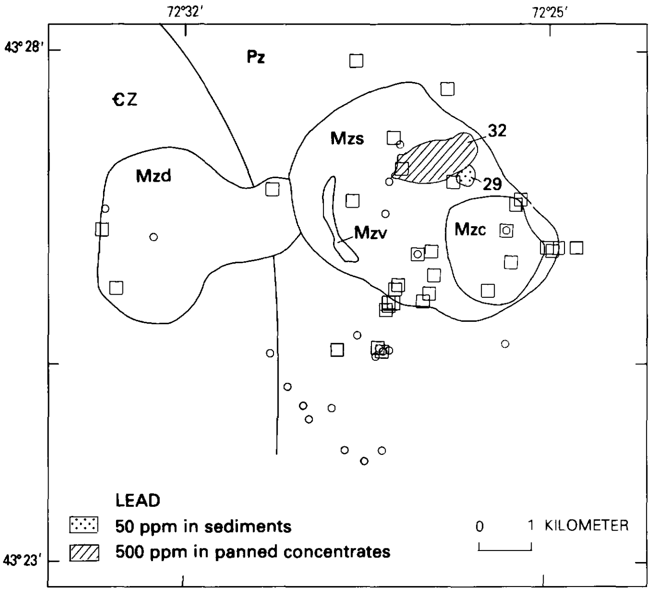

Figure 6. Distribution of lead in stream sediment and panned concentrate samples in the Mount Ascutney area. See figure 1 for explanation of abbreviations and symbols.

Mn, Pb, and Zn (figs. 3–7) and may be associated with high values of Sn and W (fig. 2). In the Mount Ascutney area, the molybdenum may occur as minor disseminations and (or) in veinlets and fractures in the syenite.

The distribution of tin shows a strong association with the biotite granite and with Mesozoic volcanic rocks. The greatest tin concentration (150 ppm) occurs in panned concentrate sample 54 collected from a basin that includes the largest outcrop of volcanic rock (fig. 2). The highest whole-rock tin value (30 ppm) is for an aphanitic volcanic rock (sample 38) obtained from the northern part of the syenite stock (fig. 2). The lack of anomalous concentrations of tin in stream sediment samples draining the small stock of biotite granite and the preferential association of higher tin values with the scattered volcanic bodies suggest that cassiterite-bearing veins may occur within the small areas of remnant volcanic rocks preserved north and west of Mount Ascutney.

Anomalous concentrations of B, Zn, Mn, Pb, Mo, Ag, and Ba (Cox, 1987, table 3) are present in the calcareous rocks of the Waits River and Gile Mountain Formations that lie within approximately 0.4 km of the syenite intrusion. Anomalous levels of boron, tungsten, and tin also are present in panned concentrate and stream sediment samples collected from drainage basins containing the calcareous beds. Nearby soil samples contain anomalous boron, zinc, and tin. The metal-bearing rocks appear to lie within the contact metamorphic aureole identified by Daly (1903) and Nielson (1973).

In general, contact metamorphic aureoles are characteristic sites of skarn deposits produced by evolved magma in anorogenic settings (Einaudi and others, 1981). Orebodies of contact metamorphic association also are characterized by (1) the presence of an igneous source rock, preferably one that evolves a great deal of volatile matter; (2) reactive and soluble host rocks, generally carbonate rocks; and (3) conditions of metal transfer, shown by mineral assemblages and the presence of elements such as phosphorus and boron (Stanton, 1972, p. 616–622).

In the Mount Ascutney area, the first criterion is met, in part, by the presence of the syenite intrusion. The second criterion is satisfied by the description by Nielson (1973) of calc-silicate assemblages indicative of an impure limestone protolith in the Waits River Formation south of the syenite pluton in the vicinity of samples 5 and 45 (fig. 1) and of rock samples containing 5 to 10 percent calcium (reported in this study from samples 24 (pinstriped, calc-silicate-bearing schist), 76 (siliceous marble), and 78 (black carbonaceous schist) (fig. 1)). Occurrences of marble and calcareous rock are otherwise not well documented at Mount Ascutney. The third criterion is partially satisfied by the recognition by Nielson (1973) of silica diffusion in schists within 20 m of the syenite contact and the presence of metamorphic mineral assemblages (diopside-wollastonite-plagioclase-quartz±grossularite±scapolite) that differ from the regional metamorphic assemblage (calcite-ankerite-quartz±muscovite±biotite) described in calcareous schists of the Waits River Formation beyond the contact metamorphic aureole (Nielson, 1973; Hepburn and others, 1984).

The regional strike of the Waits River and Gile Mountain Formations, which is nearly perpendicular to the stock, restricts the location of potential mineralized bodies in carbonate host rocks to the width of the metamorphic aureole. The vertical extent of such bodies, however, is nearly unrestricted. Although the syenite has generated the requisite conditions for solution transfer, it does not otherwise display evidence that a great deal of volatiles evolved during its emplacement. Therefore, the Mount Ascutney area is thought to have low mineral resource potential for undiscovered tungsten, tin, and zinc in skarn deposits associated with the contact metamorphic aureole.

Concentrations of some metals, particularly tin and gold in soil and stream sediment samples collected south of the pluton, may be due to glacial dispersion of metal-bearing bedrock from the north. Larsen (1987) has shown that a boulder train of pebbles and erratics (derived from Mount Ascutney) located south and southeast of the syenite stock represents the direction of ice movement during the Wisconsin glaciation. Watts (unpub. data) shows that samples of stream alluvium and glacial till south of Mount Ascutney reflect a geochemical overprint of glacially transported mineral debris from the plutonic rocks rather than the geochemistry of the lower to middle Paleozoic rocks (B-horizon soil samples lack the overprint). The distribution patterns for tin (fig. 2) indicate that this metal, as discussed above, is derived from the igneous rocks of the Mount Ascutney region; however, the anomalous concentrations of tin in north-draining basins south of the igneous rock exposures are derived either from the lower Paleozoic bedrock or, alternatively, from glacial pebbles and erratics that originated from tin-bearing outcrops directly to the north. Gold similarly may have been derived via glacial transport from bedrock sources to the north because the distribution of gold-bearing samples is almost exclusively south of the pluton (fig. 1). From the available data, no unequivocal conclusions can be drawn regarding the specific source(s) of these metals, and more detailed work will be needed to identify potential bedrock areas that contain concentrations of gold and tin.

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Chapter O

Carbonate-Hosted Gold Mineralization of Hydrothermal-Replacement Origin Adjacent to a Syenitic Stock at Cuttingsville, Vermont, and the Potential for Other Syenite-Related Gold Deposits in New England

By GILPIN R. ROBINSON, JR.

U.S. GEOLOGICAL SURVEY BULLETIN 1887

SUMMARY RESULTS OF THE GLENS FALLS CUSMAP PROJECT, NEW YORK, VERMONT, AND NEW HAMPSHIRE

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Carbonate-Hosted Gold Mineralization of Hydrothermal-Replacement Origin Adjacent to a Syenitic Stock at Cuttingsville, Vermont, and the Potential for Other Syenite-Related Gold Deposits in New England

By Gilpin R. Robinson, Jr.[1]

Abstract

Areas of hydrothermal alteration that have anomalous base- and precious-metal contents are associated with a 100-Ma alkaline stock of the White Mountain Plutonic-Volcanic Suite that intrudes Precambrian metamorphic rocks at Cuttingsville, Vermont. The stock is composed predominantly of two bodies of syenitic rocks in the central and northern parts of the complex and smaller bodies of monzodiorite, hastingsite syenite, and sodalite-nepheline syenite in an arcuate zone around the central syenite body. The youngest intrusive rocks are a series of lamprophyre, trachyte, and xenolith-rich dikes. Geophysical expression of the stock indicates that it is the surface expression of a much larger body at depth and that the exposed portion may represent intrusives emanating from a cupola of a larger subsurface magma chamber.

On the northeastern side of the stock is a massive pyrrhotite deposit that formed as an epigenetic replacement of stratified Precambrian marble. This massive sulfide body, mined during the mid 1800's for the manufacture of copperas (iron sulfate), is rich in arsenic, bismuth, copper, silver, and gold (0.16 to 17 parts per million). Oxide gossan, developed from both the weathered massive sulfide and from roasted ore piles, contains arsenic and gold contents similar to those of the bulk sulfide ore. The replacement sulfide deposit is located adjacent to a large area of quartz-pyrite stockwork veins and associated sericite-pyrite-carbonate alteration in syenitic rocks of the intrusive stock. The quartz-pyrite-sericite-carbonate alteration zone has anomalous arsenic, bismuth, tin, and tungsten; one sample from this locality contains an anomalous gold value of 1.7 parts per million. The alteration zone is poorly exposed and has an unevaluated potential for precious- and base-metal mineralization. Other areas that have anomalous gold or molybdenum mineralization occur in alteration zones near some of the intrusive contacts.

The nature of the intrusive complex at Cuttingsville, the local anomalies of gold, arsenic, bismuth, and molybdenum in areas of hydrothermal alteration, and the type and style of this alteration appear similar to the features characteristic of disseminated gold deposits associated with intrusions of alkalic syenitic rocks in other parts of North America. As such, the mineralization at Cuttingsville appears to represent a class of precious-metal deposits heretofore unrecognized in New England and suggests that a potential for gold may exist with similar intrusive rocks of the White Mountain Plutonic-Volcanic Suite elsewhere in New England and in southeastern Quebec.

INTRODUCTION

A massive pyrrhotite deposit near the town of Cuttingsville in central Vermont was mined from a series of open trenches on the southwestern flank of Copperas Hill during the mid 1800's for the manufacture of copperas (iron sulfate), which was used principally in the tanning industry. The deposit was first described by Hager (1861) in a report to the Vermont State Legislature. A mapping, drilling, and geochemical study of the deposit was initiated in 1948 by Charles Doll, then the State Geologist of Vermont, and anomalously high gold values (up to 0.4 oz/ton) were reported (Doll, 1969). As gold prices increased during the early 1980's, North American Exploration Corporation (Charlottesville, Va.) began a program of exploratory geophysics, drilling, and geochemistry, initiated as a result of Doll's earlier work. Barbara Rudnick, a graduate student at the University of New Brunswick (Canada), recently completed a master's thesis that includes geologic mapping

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Manuscript approved for publication April 13, 1990.

[1] U.S. Geological Survey.

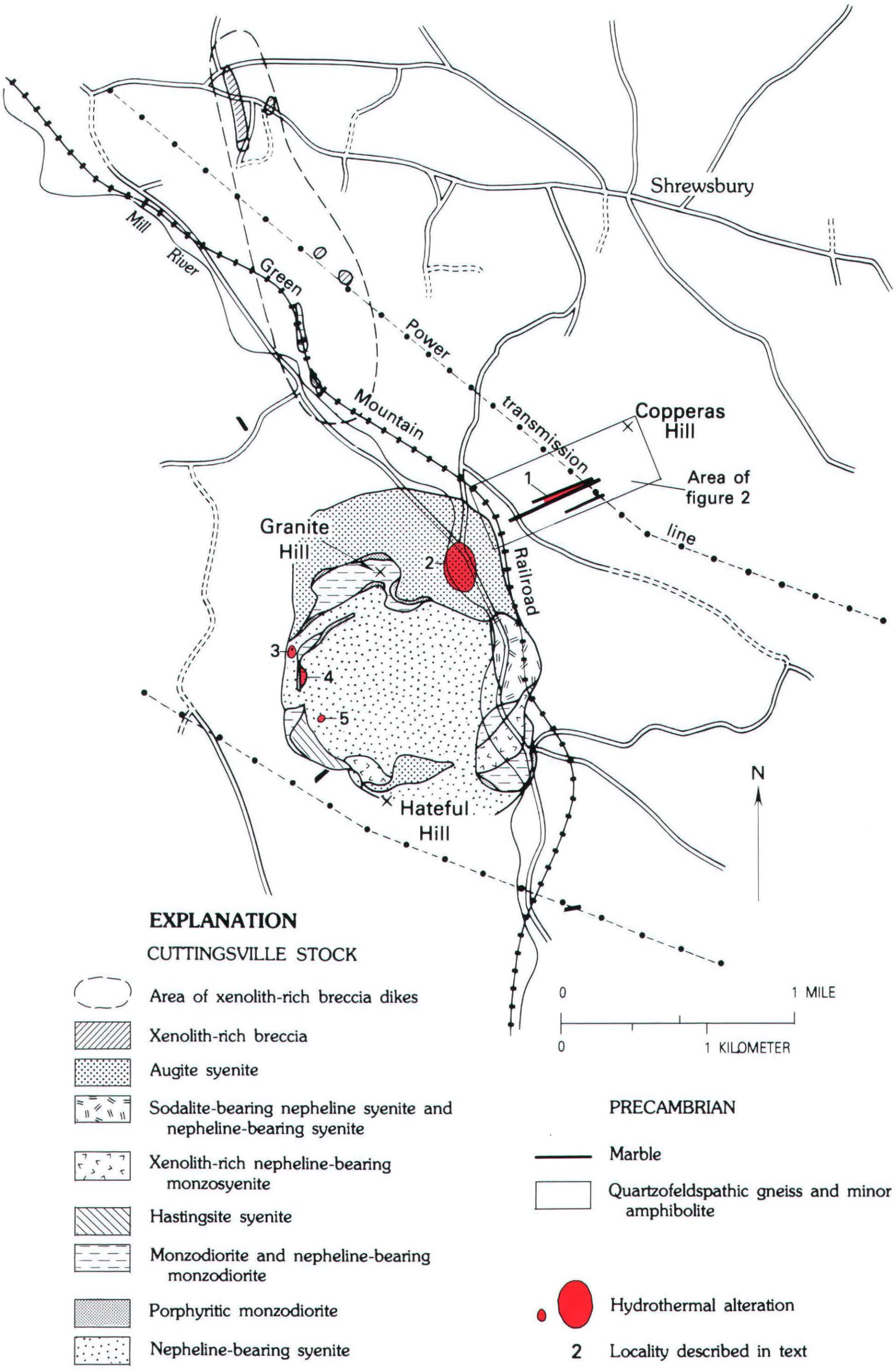


Figure 1. Geology of the Cuttingsville, Vt., area. Modified from Laurent and Pierson (1973). Marble bodies west of the intrusive are from Eggleston (1918). Plutonic rock nomenclature is given in figure 3.

of the area surrounding the pyrrhotite deposit and much of the geophysical information collected by North American Exploration (Rudnick, 1986). The present report is the result of a topical study conducted under the auspices of the Glens Falls 1° × 2° CUSMAP (Conterminous United States Mineral Assessment Program) project at the U.S. Geological Survey. Access to core and core logs held by North American Exploration Corporation and discussions with Barbara Rudnick greatly facilitated this study.

The deposit at Cuttingsville is associated with a 100-Ma alkaline intrusive stock (fig. 1) that intrudes an 1100-Ma Precambrian metamorphic terrane (Eggleston, 1918; Laurent and Pierson, 1973). The deposit forms a series of stratabound lenses of massive pyrrhotite 1 to 6 m

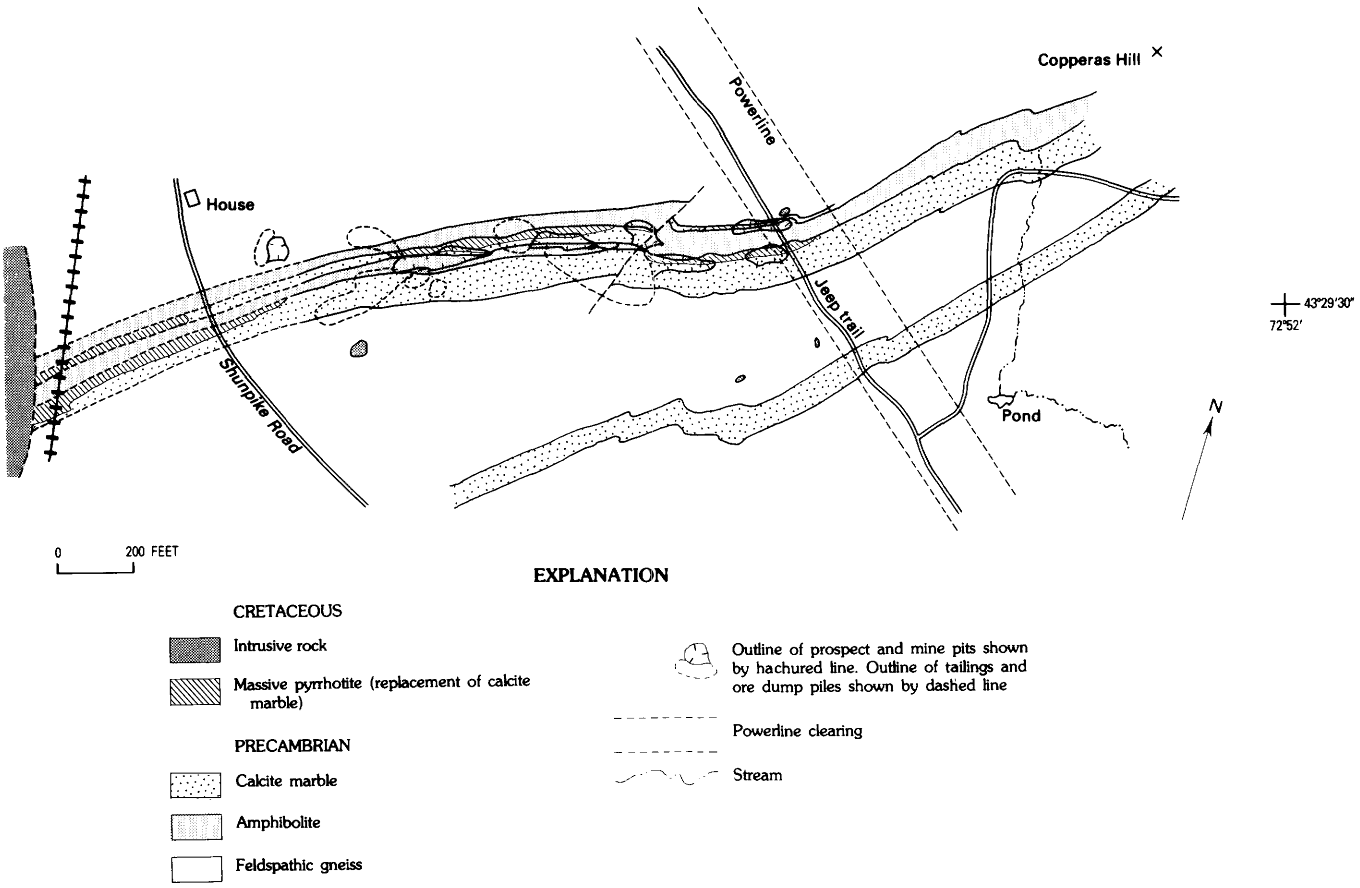


Figure 2. Geology of the Copperas Hill area, Cuttingsville, Vt. Modified from Rudnick (1986).

thick within a body of Precambrian calcite marble enclosed by feldspathic gneiss (fig. 2). Evidence indicates that the massive pyrrhotite formed as an epigenetic replacement of carbonate rocks. This pyrrhotite deposit (fig. 1, locality 1) is located adjacent to a large area of quartz-pyrite stockwork veins and associated sericite-pyrite-carbonate alteration in feldspathic rocks (fig. 1, locality 2) and has an unevaluated potential for additional precious- and base-metal mineralization. The nature of the intrusive complex at Cuttingsville, the local anomalies in gold, arsenic, bismuth, and molybdenum associated with areas of hydrothermal alteration along the margins of the stock, and the type and style of alteration appear to be similar to the features characteristic of other disseminated gold deposits associated with intrusions of alkaline syenitic rocks (Mutschler and others, 1985). As such, the Cuttingsville mineralization may represent a class of precious-metal deposits heretofore unrecognized in New England. Consequently, a potential for gold may exist for other syenitic intrusive rocks of the White Mountain Plutonic-Volcanic Suite (including the Monteregian Hills intrusives) in New England and southeastern Quebec.

CUTTINGSVILLE STOCK

The Cuttingsville intrusion is a small composite stock approximately 2 km in diameter that intrudes 1100-Ma Proterozoic rocks of the Mount Holly Complex in the Green Mountains of Vermont (fig. 1). The intrusion is part of the alkalic to subalkalic White Mountain Plutonic-Volcanic Suite in New England and contains both feldspathoid- and quartz-bearing rocks. The Cuttingsville intrusion is located off the main trend of intrusive centers of the White Mountain Plutonic-Volcanic Suite in New Hampshire but is within the broader province, as defined by McHone (1984), of small intrusive plugs and dikes that have chemical affinities with the White Mountain Plutonic-Volcanic Suite and intrusions in the Monteregian Hills area of southeastern Quebec.

The Cuttingsville stock is composed predominantly of two bodies of syenitic rocks (nepheline-bearing syenite and augite syenite). Smaller bodies of monzodiorite, hastingsite syenite, nepheline syenite, and a nepheline-bearing monzosyenite are present in an arcuate zone around the central nepheline-bearing syenite body (fig. 1; after Laurent and Pierson, 1973). K/Ar ages on biotite from hastingsite syenite (100±2 Ma) and monzodiorite (96.4±2 Ma) (Armstrong and Stump, 1971) make this the youngest dated intrusive body of the White Mountain Plutonic-Volcanic Suite. The age of the Cuttingsville stock is similar to K/Ar biotite ages reported for lamprophyre dikes in the region (McHone, 1984). Field relations and radiometric age determinations provide evidence for the order of intrusive emplacement (Laurent and Pierson, 1973). Petrographic descriptions are from this study, Eggleston (1918), and Laurent and Pierson (1973).

The central nepheline-bearing syenite body is the oldest intrusive phase of the Cuttingsville stock. This syenite is coarse grained and consists chiefly of microperthitic feldspar and oligoclase. Small amounts of nepheline occur in interstitial patches, and augite and biotite are present in some places. Sphene, magnetite, pyrrhotite, apatite, and zircon are accessory phases.

Armstrong and Stump (1971) report a K/Ar age of 100±2 Ma on biotite from the small hastingsite syenite body that intrudes the western margin of the nepheline-bearing syenite body (fig. 1). The hastingsite syenite consists approximately of 80 percent feldspar, two-thirds of which is plagioclase that is intergrown with the remaining one-third of orthoclase. Hastingsite amphibole and biotite occur in roughly equal proportions. Trace amounts of altered nepheline occur in intergranular areas. Accessory phases are apatite, sphene, magnetite, zircon, and pyrite.

Monzodiorite and nepheline-bearing monzodiorite bodies, which, in some places, have porphrytic phases along their outer margins, occur in four small bodies that intrude the nepheline-bearing syenite body near its margin (fig. 1). The monzodiorite is medium to coarse grained and consists of approximately 60 percent labradorite plagioclase, 5 to 10 percent soda orthoclase, and roughly equal amounts of pyroxene, hornblende, and biotite. Both augite and pigeonite are present as pyroxene phases. The hornblende is kaersutite and typically is intergrown with augite. Biotite is commonly intergrown with hornblende. Trace amounts of olivine and nepheline or quartz are present in some areas. Accessory phases include magnetite, sphene, apatite, pyrite, and pyrrhotite. Armstrong and Stump (1971) report a K/Ar age of 96.4±2 Ma on biotite from a gabbroic phase of this body.

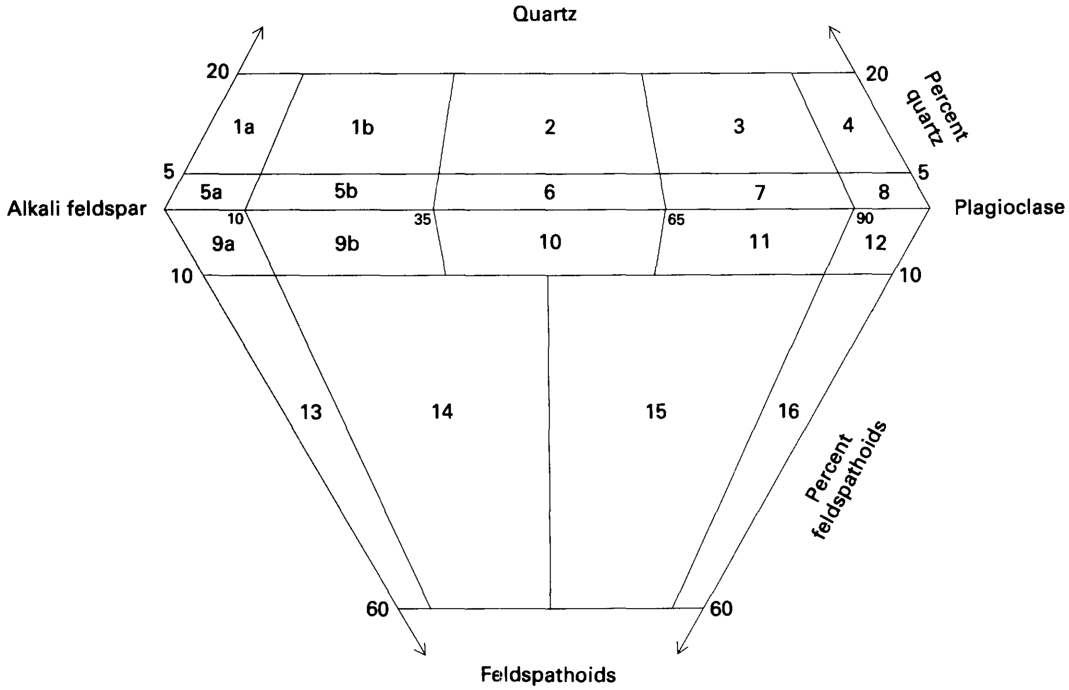
A complex body of feldspathoid-bearing rocks, classified as nepheline-bearing monzosyenite (figs. 1 and 3) and locally containing abundant inclusions and blocks of alkali gabbro in a feldspathic matrix, intrudes monzodiorite and nepheline-bearing syenite along the eastern margin of the syenite body. Outcrops of this unit along the Mill River expose a lithologically variable rock that appears to result from the partial assimilation and impregnation of alkali gabbro by syenitic magma. This hybrid rock is lighter in color than the monzodiorite, and its grain size is highly variable. Approximately 70 percent of the rock is feldspar, of which about two-thirds is plagioclase, and one-third is orthoclase. Small amounts of nepheline and cancrinite (an alteration of nepheline) occur in interstitial patches. Ferromagnesian phases include coarse hornblende, surrounded and intergrown with biotite in some places, and augite, poikilolitically enclosed in hornblende.

A body of sodalite-bearing nepheline syenite is poorly exposed along the eastern margin of the nepheline-bearing syenite body and consists of a coarse-grained aggregate of approximately equal proportions of plagioclase and microperthitic feldspar (some of which may be anorthoclase). Feldspar composes approximately 90 percent of the syenite; the remaining phases are chiefly interstitial pyroxene (aegirine-augite), biotite, nepheline, and sodalite. Apatite, magnetite, sphene, and pyrite are accessory phases.

The small augite syenite body in the southern part of the stock is intrusive into the nepheline-bearing syenite. The larger augite syenite body in the northern part of the stock is reported to be intrusive into the monzodiorite and contains inclusions of porphyritic monzodiorite (Eggleston, 1918, p. 383), although contact relations observed during this study are ambiguous. The augite syenite consists of approximately 75 percent feldspar, chiefly microperthite and oligoclase. Ferromagnesian phases consist of hedenbergitic augite that is rimmed with green hornblende. Quartz is present as a minor phase. Magnetite, zircon, apatite, and sphene are present as accessory phases.

The youngest intrusive rocks are a series of lamprophyre, trachyte, phonolite, and xenolith-rich dikes that cut all units of the stock (Eggleston, 1918), the replacement pyrrhotite body on Copperas Hill (fig. 1, locality 1), and the area of sericite alteration at locality 2 on figure 1. The xenolith-rich dikes are pipelike bodies that contain abundant angular to subrounded fragments of Precambrian and igneous lithologies in a matrix of fine-grained rock fragments and aphanitic mafic dike rock (Doss, 1986). The xenolith-rich dikes apparently represent diatreme vents formed from phreatomagmatic explosions located above a shallow subsurface magma chamber (for example, Lorenz, 1985). This magma chamber was enriched in volatiles, as evidenced by the presence of lamprophyre and xenolith-rich dikes, the saturation of most intrusive units with amphiboles and disseminated sulfides, and the presence of trachyte dikes that have internal pyrite-sericite-carbonate alteration.

Chemical analyses and mineral norms of representative samples of the various intrusive units of the Cuttingsville stock and dike units are given in tables 1 and 2, respectively. Overall, the magma series represented in the Cuttingsville stock is alkaline in character; total alkalis (Na_2O+K_2O) are greater than 6 weight percent in the mafic rocks and greater than 10 weight percent in the felsic differentiates (fig. 4*A*). The series shows strong depletion in ferromagnesian elements and enrichment in alkalis (fig. 4*B*). Most of the rocks are metaluminous (molar



1a. Alkali-feldspar quartz syenite
1b. Quartz syenite
2. Quartz monzonite
3. Quartz monzodiorite
4. Quartz diorite
5a. Alkali-feldspar syenite
5b. Syenite
6. Monzosyenite
7. Monzodiorite/monzogabbro
8. Diorite/gabbro
9a. Foid-bearing alkali-feldspar syenite
9b. Foid-bearing syenite
10. Foid-bearing monzosyenite
11. Foid-bearing monzodiorite
12. Foid-bearing diorite
13. Foid syenite
14. Foid monzosyenite
15. Foid monzodiorite
16. Foid diorite

Figure 3. General classification and nomenclature of plutonic rocks at Cuttingsville, Vt. Classification and nomenclature according to modal mineral content measured in volume percent (after International Union of Geological Sciences Subcommission on the Systematics of Igneous Rocks, 1973); *f*eldspatho*id*=foid.

$Al_2O_3 / (CaO + Na_2O + K_2O - CO_2) < 1 < Al_2O_3 / (Na_2O + K_2O)$), and many of the felsic differentiates border on peralkalinity, although none of the unaltered igneous rocks is peralkaline (molar $Al_2O_3/(Na_2O+K_2O)<1$). Both quartz-normative and nepheline-normative chemistries are present. However, the emplacement order of intrusive bodies and dikes lacks a coherent fractionation trend; mafic and felsic melts, both silica saturated and silica undersaturated, are represented by early and late intrusions. Na_2O/K_2O ratios in unaltered rocks generally remain between 2.5 and 1.5 and show a trend toward lower values at high degrees of fractionation (fig. 4*C*; table 1). Sericite alteration shifts Na_2O/K_2O ratios to values less than or equal to 1 (table 3), largely by sodium loss (fig. 5). Sodic alteration, mainly by sodium-potassium exchange, leads to partial albitization and to Na_2O/K_2O ratios generally greater than 2.5 (table 3).

The magma series at Cuttingsville is characterized by uniform Th/U ratios (approximately 4); high and generally uniform Nb/Y (2.6–14), Nb/Ta (14–21), and La/Yb (29–52) ratios; and high Zr/TiO_2 ratios (0.01–0.4) that increase with increasing Thornton-Tuttle index (tables 1 and 2). The observed ranges of these element ratios are typical of alkalic magma series, and the igneous rocks generally fall in the alkali basalt, basanite, phonolite, and trachyte compositional fields of Winchester and Floyd (1977).

The mafic intrusive rocks at Cuttingsville (table 1: WR84–60 and WR84–95A) have small negative chondrite-normalized europium anomalies ($(Eu/Sm)_{chon}$ near 0.75), Ba/Sr ratios near 1, Na_2O/K_2O ratios near 2, and 100 Rb/Ba ratios of 5.2 and 5.9 (table 1). These values are similar to the Na_2O/K_2O ratios reported for alkali basalt and the range of 100 Rb/Ba ratios of 4.1 to 9.7 reported for primary

Table 1. Geochemistry of fresh and hydrothermally altered igneous rocks, Cuttingsville, Vt.

[Major element values are in weight percent, and trace element values are in ppm. NA, the element content was not determined. The geochemical results reported here were provided by analytical laboratories of the U.S. Geological Survey. Analysts for major elements were J.W. Marienko, M.W. Doughton, P.J. Aruscavage, and Hezekiah Smith. Fluorine and chlorine were determined by specific ion electrode (Norma Rait, analyst). Trace elements were determined by quantitative spectrometric methods (P.J. Aruscavage, M.W. Doughton, and J.S. Kane, analysts) and instrumental neutron activation techniques (C.A. Palmer, J.S. Mee, and L.J. Schwarz, analysts), with the exception of niobium which was determined by spectrophotometric methods (P.J. Aruscavage, analyst)]

	Fresh Rocks								Altered Rocks				
	WR84–60	WR84–84	WR84–86	WR84–53	WR84–95A	WR84–95B	WR84–141C	WR84–93	WR84–54A	CUT6–66	WR84–54C	WR84–54C3	WR84–64
SiO_2	48.6	58.5	67.8	60.0	49.0	66.6	53.5	56.6	64.8	59.5	55.39	57.8	64.96
Al_2O_3	15.0	20.7	16.6	19.8	16.5	17.1	18.1	18.0	17.0	21.8	18.14	26.2	16.45
Fe_2O_3	4.7	1.5	1.7	1.9	4.0	1.4	3.1	3.3	2.3	2.0	4.82	1.3	2.68
FeO	7.3	2.6	1.4	.84	4.4	1.4	5.2	3.0	1.2	.52	6.51	1.8	3.62
MgO	6.3	1.1	.12	.21	5.4	.3	3.0	2.3	.34	.31	.32	.18	.02
CaO	8.9	3.2	.82	.97	7.7	1.2	6.4	4.4	1.1	1.3	.95	.05	.61
Na_2O	4.2	7.4	6.6	8.1	5.6	7.1	5.5	6.4	6.1	6.2	3.63	2.9	8.44
K_2O	1.9	3.8	4.7	5.6	2.6	4.9	2.7	3.5	5.7	6.2	8.98	9.3	2.38
H_2O^+	.61	.72	.20	.73	.93	.20	.84	.49	.11	.91	NA	NA	NA
H_2O^-	.19	.16	.33	.27	.17	.15	.03	.07	.14	.01	NA	NA	NA
TiO_2	2.4	.80	.18	.28	1.8	.40	1.7	1.3	.43	.25	.22	.27	.40
P_2O_5	.87	.31	.04	.05	.55	.10	.79	.41	.15	.05	.05	.03	.05
MnO	.24	.14	.19	.23	.20	.04	.19	.16	.08	.26	.30	.01	.05
CO_2	.10	.22	.50	.70	.58	.39	.29	.32	.74	1.4	NA	NA	NA
Cl_2	.061	.049	.007	.31	.87	.14	.07	.029	.044	.004	NA	NA	NA
F_2	.090	.062	.054	.074	.090	.034	.096	.084	.056	.096	NA	NA	NA
SUM	101.46	101.26	101.24	100.06	100.39	100.45	101.51	100.36	100.29	100.81	99.31	99.84	99.66
0 = f+Cl	−.03	−.02	−.01	−.05	−.12	−.02	−.03	−.02	−.02	−.02			
SUM	99.64	101.24	101.23	100.01	100.27	101.43	101.48	100.34	100.13	100.29	99.31	99.84	99.66
La	96	76	200	126	100	133	100	86	113	146	130	160	130
Yb	2.91	1.90	3.82	4.40	2.32	3.68	3.01	2.74	2.49	3.22	3.8	4.2	6.4
Zr	244	362	573	1,120	256	605	370	453	700	1,220	310	180	900
Hf	5.29	7.13	13.1	22.9	5.59	15.59	7.9	9.02	15.26	19.8	—	—	—
Rb	48	110	130	303	61	150	78	106	137	280	—	—	—
Ba	927	2,440	22	268	1,030	322	923	1,020	377	180	400	510	300
Th	5.76	13.62	36.5	65.5	8.2	53.9	12.5	14.6	25.0	48.8	—	—	—
U	1.41	3.19	7.79	24.1	2.03	13.0	3.25	3.43	3.75	9.27	—	—	—
Nb	81	115	140	380	105	250	115	115	130	210	88	93	89
Ta	5.18	5.51	9.73	19.8	7.03	14.78	7.5	7.19	9.32	6.84	—	—	—
Y	31	14	34	27	15	35	27	24	26	22	19	24	27
Sr	880	1,120	4	57	841	118	964	681	73	28	61	34	47
La/Yb	33	40	52	29	43	36	33	31	45	45	34	38	20
Zr/Hf	46	51	44	49	46	39	47	50	46	62	—	—	—
100Rb/Ba	5.2	4.5	591	113	5.9	47	8.5	10	36	156	—	—	—
Th/U	4.1	4.3	4.7	2.7	4.0	4.1	3.8	4.3	6.7	5.3	—	—	—
Nb/Ta	16	21	14	19	15	17	15	16	14	31	—	—	—
Zr/TiO_2	.010	.045	.32	.40	.014	.15	.022	.035	.16	.49	.14	.067	.23
Nb/Y	2.6	8.2	4.1	14	7	7	4.3	4.8	5.0	9.5	4.6	3.9	3.3
Ba/Sr	1.1	2.2	5.5	4.7	1.2	2.7	1.0	1.5	5.2	6.4	6.6	15	6.4
Na_2O/K_2O	2.21	1.95	1.40	1.45	2.15	1.45	2.04	1.83	1.07	1.00	.40	.31	3.55
$(Eu/Sm)_{chon}$	.76	1.19	.10	.38	.74	.42	.73	.76	.49	.10	—	—	—

basalts (Frey and others, 1978). The felsic differentiates that have highly elevated 100 Rb/Ba ratios greater than 40 (table 1: WR84–86 (591), WR84–53 (113), and WR84–95B (47)) also have low Na_2O/K_2O ratios of approximately 1.5, elevated Ba/Sr ratios of 3 to 6, and strong negative chondrite-normalized europium anomalies ($(Eu/Sm)_{chon}$ <0.5). The strong negative chondrite-normalized europium anomalies, greatly increased Rb/Ba ratio, slightly increased Ba/Sr ratio, and decreased Na_2O/K_2O ratio are consistent with the fractionation of anorthoclase from the melts (Hanson, 1978).

The Zr/Hf ratio of the Cuttingsville magma series remains relatively constant at 39 to 51. This range is somewhat high compared with Zr/Hf ratios of 37 to 45 for primary basalt (Frey and others, 1978) and approaches the average crustal Zr/Hf ratio of 55 (Taylor, 1964). This range is also higher than the range of 30 to 43 reported for most intrusives and dikes in the Monteregian Hills, Quebec, and

Table 1. Geochemistry of fresh and hydrothermally altered igneous rocks, Cuttingsville, Vt.—Continued

Sample number	Type of rock	Latitude	Longitude	Locality number on figure 1	Comments
Intrusive rocks					
WR84–60	nepheline-bearing monzodiorite	43°28′39.4″	72°53′44.5″	—	
WR84–84	hastingsite syenite	43°28′36″	72°53′44.7″	—	
WR84–86	augite syenite	43°29′18.0″	72°53′39.0″	—	
WR84–53	nepheline syenite	43°28′41.9″	72°52′33.2″	—	
Dikes					
WR84–95A	mafic dike	43°28′30.0″	72°52′42.8″	—	
WR84–95B	alkali feldspar trachyte	43°28′30.0″	72°52′42.8″	—	
WR84–141C	nepheline-bearing trachyte	43°29′49.8″	72°53′36.3″	—	
WR84–93	nepheline-bearing trachyte	43°28′26.9″	72°52′39.6″	—	
Altered rocks					
WR84–54A	altered augite syenite	43°29′13.7″	72°52′25.6″	2	slight sericite alteration.
CUT6–66	altered felsic dike	43°29′30″	72°53′00″	5	sericite-pyrite-carbonate alteration.
WR84–54C	altered augite syenite	43°29′13.7″	72°52′25.6″	2	sericite-pyrite alteration.
WR84–54C3	altered augite syenite	43°29′13.7″	72°52′25.6″	2	sericite-pyrite alteration.
WR84–64	altered syenite	43°28′47.2″	72°53′50.2″	3	sodic alteration (albite).

Table 2. Mineral norms and fractionation indices for fresh and hydrothermally altered igneous rocks, Cuttingsville, Vt.
[See table 1 for sample locality information]

	WR84–60	WR84–84	WR84–86	WR84–53	WR84–95A	WR84–95B	WR84–141C	WR84–93	WR84–54A	CUT6–66	WR84–54C	WR84–54C3	WR84–64
Norm (weight percent)													
Q	—	—	10.61	—	—	5.68	—	—	7.08	—	—	4.27	4.40
C	—	.01	.56	.86	—	—	—	—	.74	4.90	.84	11.35	—
or	11.23	22.46	27.77	33.09	15.36	28.96	15.95	20.68	33.68	36.64	53.06	54.66	14.06
ab	24.87	50.27	55.79	50.81	29.54	59.04	42.68	48.85	51.29	51.40	21.81	24.79	71.36
an	16.70	12.21	.28	—	15.62	.87	17.00	10.16	.33	—	4.39	.05	—
ne	5.54	6.49	—	8.36	6.18	—	1.81	2.76	—	.56	4.82	—	—
hl	.10	.08	.01	.51	1.43	.23	.12	.05	.07	.01	—	—	—
ac	—	—	—	—	—	—	—	—	—	—	—	—	.05
di	16.92	—	—	—	11.88	1.48	6.16	5.23	—	—	—	—	2.40
hy	—	—	1.52	—	—	.84	—	—	.85	—	—	2.18	2.64
ol	9.77	3.83	—	.25	7.12	—	6.55	2.98	—	.48	6.86	—	—
mt	6.81	2.17	2.46	2.64	5.80	2.03	4.49	4.78	2.88	1.80	6.99	1.89	3.86
hm	—	—	—	.07	—	—	—	—	.31	.76	—	—	—
il	4.56	1.52	.34	.53	3.42	.76	3.23	2.47	.82	.47	.42	.51	.76
ap	2.02	.72	.09	.12	1.27	.23	1.83	.95	.12	.12	.12	.07	.12
fr	.03	.07	.10	.02	.08	.05	.05	.10	.11	.12	—	—	—
cc	.23	.50	1.14	1.59	1.32	.89	.66	.73	1.59	2.05	—	—	—
sl	—	—	—	.10	—	—	—	—	—	.05	—	—	—
SUM	98.78	100.33	100.67	98.95	99.02	101.06	100.53	99.74	99.87	99.36	99.31	99.77	99.65
A/CNK[1]	.60	.96	1.02	1.01	.67	.94	.79	.83	1.03	1.27	1.04	1.76	.94
A/NK[1]	1.67	1.27	1.04	1.02	1.37	1.01	1.51	1.26	1.05	1.29	1.16	1.77	1.00
Thornton-Tuttle index	41.6	79.2	94.2	92.3	51.1	93.7	60.5	72.3	92.1	88.6	79.7	83.7	89.8

[1] A=molar Al_2O_3; C=molar $CaO-CO_2$; N=molar Na_2O; K=molar K_2O; CNK=C+N+K; NK=N+K.

for the younger intrusions of the White Mountain Plutonic-Volcanic Suite but is similar to the range of Zr/Hf ratios of 49 to 55 reported for the intrusive phases at Mount Megantic, Quebec (Eby, 1985). Eby (1985) interprets the high Zr/Hf ratios to be the result of some degree of crustal source or contamination.

A circular aeromagnetic anomaly approximately 4 km in diameter indicates that the Cuttingsville stock is the surface expression of a much larger body at depth (Daniels, this volume). Two areas of magnetic anomalies exist. The largest is centered on and surrounds the main intrusive stock. The other is an elongate north-south anomaly located

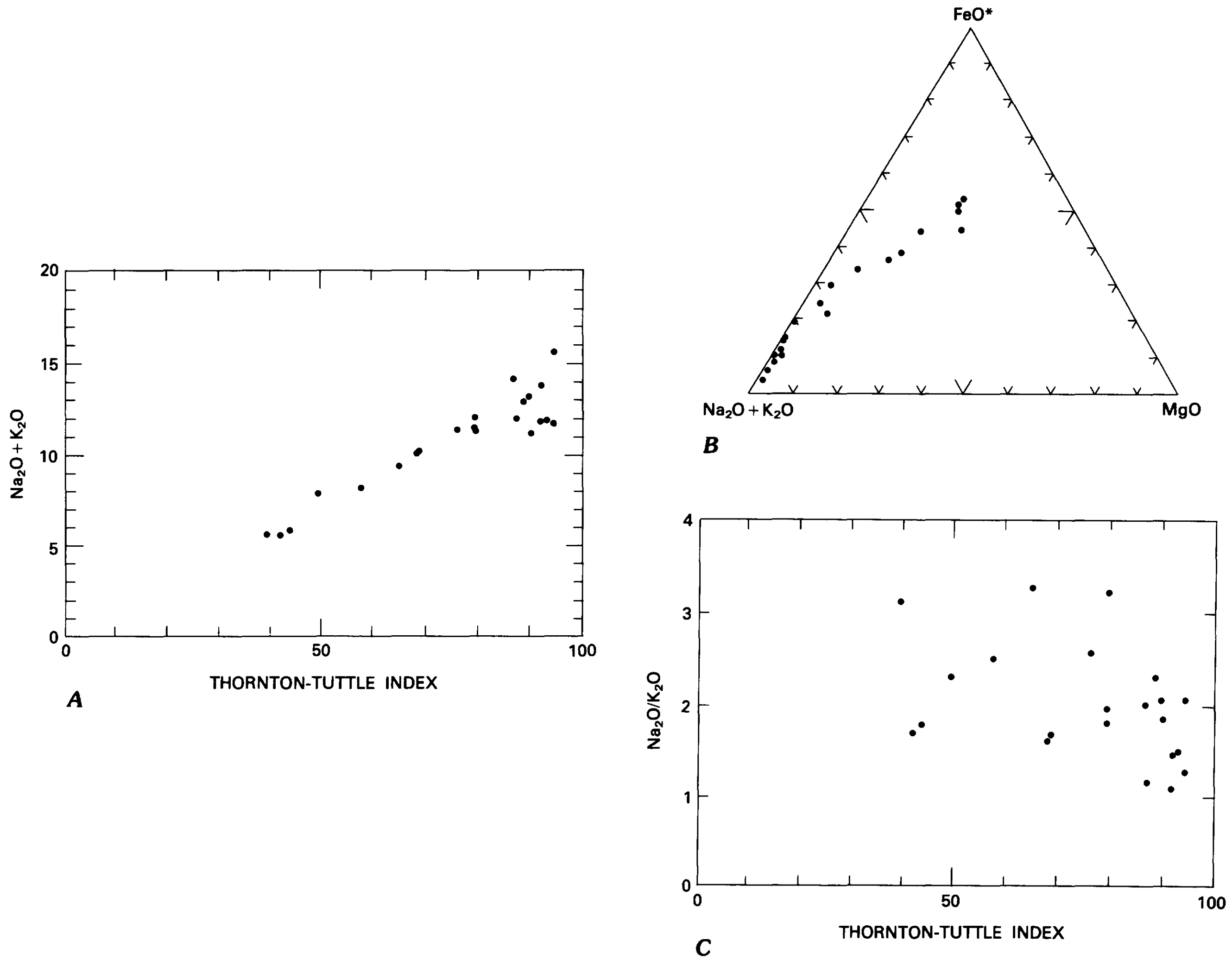


Figure 4. Chemical characteristics of unaltered intrusive rocks in the Cuttingsville stock. FeO* is total iron expressed as FeO. The Thornton-Tuttle index is the percentage sum of quartz and feldspar minerals in the chemical norm and reflects the degree of magma fractionation and differentiation. *A*, Plot of Na_2O+K_2O versus Thornton-Tuttle index. *B*, Ternary plot of MgO, FeO*, and Na_2O+K_2O. *C*, Plot of Na_2O/K_2O versus Thornton-Tuttle index. Chemical data are from table 1, Eggleston (1918), Laurent and Pierson (1973), and G.R. Robinson, Jr. (unpub. data).

north of the intrusive stock in the Shrewsbury area. The Shrewsbury anomaly, centered on a zone of pipelike xenolith-rich dikes that intrude Precambrian gneiss and amphibolite, suggests the presence of a subsurface intrusive plug in this area (fig. 1).

High-pass filtered Bouguer gravity data over the Cuttingsville stock shows a 5 to 6 mGal closed low associated with the stock but centered on the eastern side (Daniels, this volume). The center of this low coincides with the area of hydrothermal alteration at locality 2 (fig. 1). Daniels could not satisfactorily model this observed gravity anomaly using the observed average densities of the igneous and surrounding metamorphic rocks without the addition of an anomalously low-density body to the upper part of the stock. The hydrothermally altered rocks at locality 2 have a low density consistent with this model, and if the altered rocks control the location and existence of this gravity minimum, then a large area of hydrothermal alteration may exist here (Daniels, this volume). However, the main portion of the syenite stock and the zone of xenolith-rich dikes north of the stock do not show a gravity anomaly that would be expected from syenitic rocks intruding Precambrian gneiss (Daniels, this volume). Perhaps the syenite intrusives are underlain at moderate depth by mafic cumulates or gabbro (density 2.93).

The mafic rocks at Cuttingsville have geochemical signatures that are consistent with their derivation from a mantle source, perhaps with some crustal component. Both the mafic and felsic rocks show consistent trends in incompatible element variation. This variation implies that they are all members of a single comagmatic suite. However, the surface exposures of mafic rocks are insufficient to account

Table 3. Trace element geochemistry of hydrothermally altered igneous rocks, Cuttingsville, Vt.

[The localities refer to those shown in figure 1. Trace element values are in ppm. <, the element was below detection limit, as indicated by the following number. ND, the trace element was not determined. The geochemical results reported here were provided by analytical laboratories of the U.S. Geological Survey. Gold was determined by fire assay (Roosevelt Moore, analyst) and atomic absorption (Robert Welsch, analyst) techniques. Arsenic, bismuth, copper, and molybdenum were determined by atomic absorption analysis (J.S. Kane, B.J. Libby, W.M. d'Angelo, and Robert Welsch, analysts). Silver, cobalt, zinc, and tungsten were determined by emission spectrographic methods (C.J. Skeen, Z.A. Brown, and A.F. Dorrzapf, analysts)]

Sample	Au	Ag	Bi	As	Cu	Co	Zn	Pb	Mo	Sn	W	Na_2O/K_2O
Sericite-pyrite alteration of augite syenite, locality 2, figure 1												
WR84–54C	0.004	1.2	4.0	130	19	4.0	430	140	13	9.0	<15	0.4
WR84–54C2	<.005	2.4	12.0	460	48	3.3	330	160	10	8.2	450	.05
WR84–54v	<.005	6.3	16.0	41	22	1.3	98	46	6.7	15	980	.08
Sericite-pyrite alteration, felsic dike, locality 5, figure 1												
CUT6–66	<0.005	0.53	<10	<10	6.3	<1.0	150	11	1.4	5.1	ND	1.0
Altered syenite at intrusive contact, locality 4, figure 1												
WR84–82	1.3	1.5	3.0	<100	130	56	210	20	14	<4.6	ND	0.7
Altered (albitized) syenite at molybdenite prospect, locality 3, figure 1												
WR84–64B2	0.17	0.62	0.039	0.43	33	9.3	120	17	1,100	17	ND	2.5
WR84–64C	<.002	.52	.28	1.5	67	23	95	25	1,800	5.5	ND	1.3
WR84–64P	.002	.22	.068	.81	82	2.1	28	14	140	6.6	ND	3.3

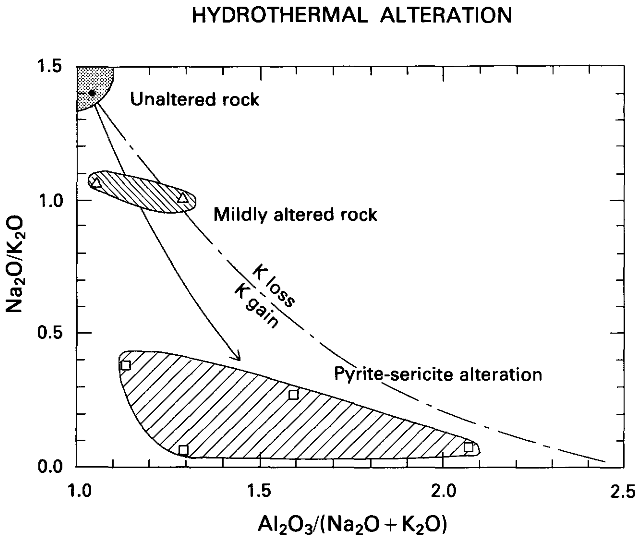


Figure 5. Chemical characteristics of sericite-pyrite alteration at locality 2 (fig. 1), Cuttingsville, Vt. Plot of molar Na_2O/K_2O versus $Al_2O_3/(Na_2O+K_2O)$ for unaltered, mildly altered, and pyrite-sericite altered rock. Sericite-pyrite alteration shows significant sodium loss and only slight potassium gain relative to unaltered syenite on an aluminum-normalized basis. Chemical data are from table 3 and G.R. Robinson, Jr. (unpub. data). Symbols represent analyses of individual samples used to delineate alteration fields.

for the volume of exposed felsic differentiates. In addition, the order of igneous emplacement does not correlate with fractionation trends, and during the youngest phase of intrusive activity, silica-undersaturated mafic and phonolitic magmas and quartz-saturated syenitic magmas all were emplaced as dikes in the same areas within and surrounding the stock. McHone and Corneille (1980, p. 20–21) noted that trachyte and phonolite dikes, especially where abundant, are generally near syenitic intrusive bodies, and the dikes are interpreted as offshoots from these bodies. On the basis of the chemistry, geophysical features, and geologic relations, the Cuttingsville stock is interpreted as a series of intrusions emanating from a cupola associated with a larger magma chamber at depth. This magma chamber appears to have been chemically zoned with a thin layer of silica-undersaturated, halogen-rich phonolitic magma concentrated near the top of the cupola, a middle zone of silica-undersaturated to silica-saturated syenitic magma in the upper part of the chamber, and a lower zone of alkalic mafic magma and crystal cumulates. The central nepheline-bearing syenite body is believed to represent an intrusive plug from the upper part of the chamber, and the younger and smaller satellite intrusive bodies surrounding this syenite body are interpreted as melts tapped from the upper and lower part of the chamber by a ring-fracture system. The halogen- and CO_2-rich trachyte, phonolite, and xenolith-rich dikes represent melts tapped from the upper margin of the cupola. Periodic intrusions of alkali gabbro magma may have entered the chamber from a mantle source.

CARBONATE-HOSTED PYRRHOTITE DEPOSIT

The carbonate-hosted pyrrhotite deposit occurs in a unit of Proterozoic gneiss and amphibolite within the aeromagnetic anomaly associated with the Cuttingsville stock. Small pipelike plugs of trachyte and phonolite intrude the Proterozoic lithologies in this area and presumably are derived from the subsurface igneous chamber that is the source of the magnetic anomaly. Figure 1 shows the position of the sulfide replacement deposit (locality 1) and other areas of alteration relative to the intrusive stock.

The massive pyrrhotite deposit consists of a series of stratabound lenses 1 to 6 m thick within a small body of stratified Proterozoic calcite marble. Field relations and petrographic study clearly show that the massive pyrrhotite replaces the calcite marble. Paragenetic relations of the marble and the massive sulfide deposit are summarized in figure 6; mineral assemblages have been divided into a regional metamorphic (preore stage 0), an ore stage (stages 1 and 2), and a postore alteration stage (postore stage 3). Preore assemblages in the marble (preore stage 0) that developed during Late Proterozoic and (or) Paleozoic regional metamorphism were

(1) calcite+actinolite±phlogopite±quartz+pyrite,

(2) calcite±tremolite±phlogopite±quartz± graphite +disseminated pyrrhotite (monoclinic), and

(3) calcite+chlorite (a retrograde alteration of actinolite).

The oldest period of mineralization (ore stage 1) involved replacement of calcite and actinolite in the marble by ferroan dolomite or ankerite, chlorite, massive pyrrhotite, and minor quartz. This replacement commonly has a border zone of ferroan dolomite or ankerite that has approximately 10 to 20 volume percent pyrrhotite surrounding an interior zone of massive (>80 percent) pyrrhotite. The massive pyrrhotite in places preserves an older metamorphic foliation in the marble as aligned stringers of silicate and carbonate inclusions. The replacement pyrrhotite lenses consist largely of hexagonal pyrrhotite or mixtures of hexagonal pyrrhotite that have some monoclinic pyrrhotite (Rudnick, 1986); locally, veinlets of monoclinic pyrrhotite (paragenetic stage 2A) cut older hexagonal pyrrhotite. This suggests a late development of monoclinic pyrrhotite by either direct precipitation or by alteration of preexisting hexagonal pyrrhotite. A small quantity of magnetite also appears to be associated with some of the late monoclinic pyrrhotite and may be related to the formation of monoclinic pyrrhotite by the oxidation of preexisting hexagonal pyrrhotite, as suggested by Desborough and Carpenter (1965) on the basis of experimental studies. In addition, thin veinlets of arsenopyrite that contain minor chalcopyrite crosscut older hexagonal pyrrhotite in areas of massive pyrrhotite replacement (paragenetic stage 2A). Native gold also has been observed in one sample of massive pyrrhotite as an inclusion in chalcopyrite (Rudnick, 1986). Mineral assemblages of the ore stage (paragenetic stages 1 and 2A) are hexagonal pyrrhotite+ankerite (or ferroan dolomite) + chlorite and trace amounts of arsenopyrite+ chalcopyrite+gold±pyrite+sphalerite±quartz±magnetite. Paragenetically late zones of siderite and siderite+ chlorite+pyrite (paragenetic stage 2B) locally replace both preore calcite and ore-stage hexagonal pyrrhotite+ferroan dolomite (ankerite) along the border of the replacement pyrrhotite bodies.

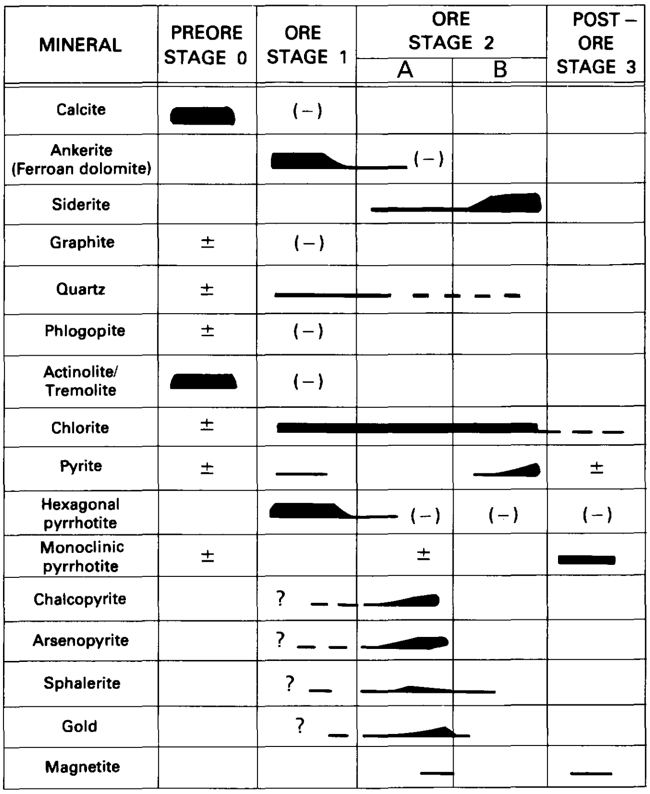


Figure 6. Paragenesis of mineral assemblages in the metamorphic Precambrian marble and epigenetic replacement pyrrhotite body at Cuttingsville, Vt. (-), destruction of a mineral phase.

Late felsic and mafic dikes associated with the intrusive stock cut the replacement pyrrhotite body and locally alter hexagonal pyrrhotite to monoclinic pyrrhotite ±minor pyrite+magnetite (paragenetic postore stage 3). Hexagonal pyrrhotite also is altered in places by supergene processes that produce marcasite along an intricate network of tiny veinlets following fractures and parting planes in the pyrrhotite.

Geochemical analyses of representative samples of massive pyrrhotite and hydrothermally altered rocks, keyed to sample localities in figure 1, are given in tables 3, 4, and 5. The massive pyrrhotite replacement deposit at Cuttingsville is chemically distinct from other stratabound pyrrhotite masses of regional metamorphic origin in Proterozoic marble and calc-silicate rocks elsewhere in the Mount Holly Complex. For instance, the replacement pyrrhotite has significantly lower cobalt and nickel contents than the metamorphic pyrrhotites (fig. 7*A*). The replacement pyrrhotite deposit is enriched in arsenic and bismuth and, in general, is poor in base metals. Concentrations of arsenic in the replacement pyrrhotite deposit are greater than those of copper+lead+zinc in most samples (fig. 7*B*). Variations of

Table 4. Trace element geochemistry of massive sulfide (mostly pyrrhotite) of hydrothermal-replacement and metamorphic origin, Cuttingsville, Vt., and vicinity

[The localities refer to those shown in figure 1. Trace element values are in ppm. <, the element was below detection limit, as indicated by the following number. The geochemical results reported here were provided by analytical laboratories of the U.S. Geological Survey. Gold was determined by fire assay (Roosevelt Moore, analyst) and atomic absorption (Robert Welsch, analyst) techniques. Arsenic, bismuth, and copper were determined by atomic absorption analysis (J.S. Kane, B.J. Libby, W.M. d'Angelo, and Robert Welsch, analysts). Silver, cobalt, and zinc were determined by emission spectrographic methods (C.J. Skeen, Z.A. Brown, and A.F. Dorrzapf, analysts)]

Sample	Au	Ag	Bi	As	Cu	Co	Ni	Zn	Pb
Massive pyrrhotite replacement body, locality 1, figure 1									
CUT6–61	0.22	7.6	17	1,200	1,700	3.1	9.7	180	41
CUT6–69	.38	6.3	110	3,100	1,500	3.8	7.8	160	31
WR83–6A	6.2	31	100	14,000	2,500	12	9.5	210	41
WR83–6B	4.4	19	68	6,400	2,600	5.9	8.2	202	5
WR83–10B	17	81	250	4,600	2,500	9.6	9.2	180	46
WR83–29	.67	5.1	160	10,000	2,000	83	14	150	37
WR83–134	.82	12	160	10,000	2,000	4.7	8.7	150	32
WR83–138	4	21	130	10,000	2,000	20	14	220	5
Gossan, locality 1, figure 1									
WR83–8 (soil)	0.52	3.9	39	1,100	750	2.2	<1.5	200	36
WR83–31 (roasted ore)	.18	20	55	680	900	5.4	<1.5	270	48
Stratabound pyrrhotite of metamorphic origin, Westin, Vt.									
WR83–2	<0.05	4.5	<10	<100	1,550	375	780	215	47

Table 5. Geochemistry of massive pyrrhotite, hydrothermally altered, and other rocks from the Cuttingsville area, Vermont

[The localities refer to those shown in figure 1. The geochemical values reflect the representative range in composition for the analyzed samples in terms of low (L), high (H), and median (M) values. Molybdenum was determined by atomic absorption analysis (J.S. Kane, B.J. Libby, W.M. d'Angelo, and Robert Welsch, analysts). N, tungsten was below the detection limit for emission spectrographic analysis (approximately 15 ppm). The geochemical results reported here are from tables 3 and 4, Doll (1969), and G.R. Robinson, Jr. (unpub. data)]

Rock type locality	Number of samples		Geochemical range, in ppm								
			Au	Ag	Bi	As	Cu	Co	Zn	Mo	W
Massive pyrrhotite	16	L	0.16	5	17	1,000	1,500	4	140		
locality 1		H	17	80	260	18,000	3,500	100	260	<2	N
		M	.75	9.4	105	5,500	2,000	6.5	190		
Gossan	10	L	0.08	1.5	8	130	70	2	150		
locality 1		H	2.8	20	160	4,100	900	10	270	<10	N
		M	1	6	50	1,100	500	5	200		
Pyrite stockwork	5	L	<0.05	1.2	2	20	7.9	1.1	98	6.7	N
locality 2		H	1.7	6.3	16	460	48	4	430	13	980
		M	<.05	2	10	76	20.5	2.2	275	8	250
Molybdenite prospect	5	L	<0.002	0.2	0.038	0.35	33	2.1	28	140	
locality 3		H	.17	.7	.28	1.5	150	33	170	1,800	N
		M	.002	.3	.07	.75	70	10	95	1,000	
Altered syenite	3	L	0.75	1.5	2.6	1.9	70	5	210	14	
locality 4		H	1.3	2	3	2	130	56	210	17	N
		M	.8	2	2.6	1.9	130	17	210	16	
Metamorphic sulfides	3	L		3.7		4.1	640	260	210		
(background)		H	<0.05	5.4	<1	9.7	1,600	380	230	<10	N
		M		4.3		9.5	1,500	370	220		

gold values in samples of the replacement pyrrhotite (0.16–17 ppm) correlate strongly with those of silver (5–80 ppm, fig. 8*A*) and moderately with bismuth (17–260 ppm, fig. 8*B*). High gold values occur with high arsenic (1,000–18,000 ppm, fig. 8*C*) and high copper (1,500–3,500 ppm, fig. 8*D*), although only a poor correlation in variation exists. The altered marble is rich in manganese (1,000 to >5,000 ppm; Doll, 1969; G.R. Robinson, Jr., unpub. data), which is contained mainly in ferroan dolomite and siderite.

Hydrothermal alteration is largely absent in the quartzofeldspathic Proterozoic gneiss bordering the mineralized

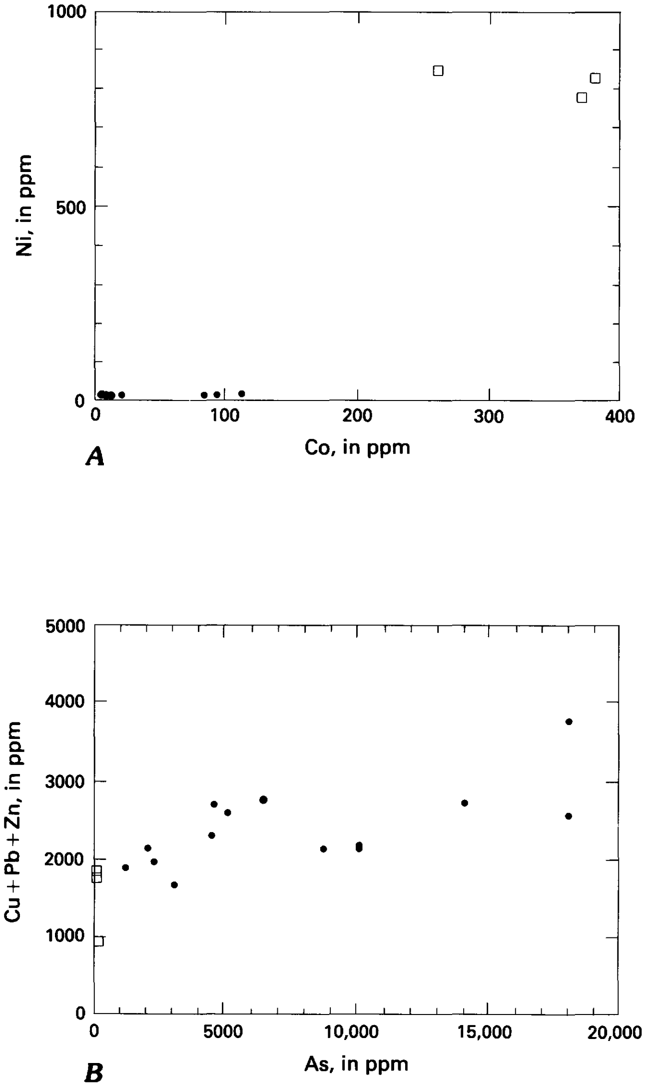


Figure 7. Chemical characteristics of the pyrrhotite replacement body at Cuttingsville, Vt., compared with nearby stratabound pyrrhotite bodies of metamorphic origin in Precambrian marble and calc-silicate rocks. Solid circle, geochemical analyses of the replacement pyrrhotite body; open square, geochemical analyses of samples of metamorphic pyrrhotite bodies. *A*, Plot of Ni versus Co. *B*, Plot of Cu+Pb+Zn content versus As. Chemical data are from table 3 and G.R. Robinson, Jr. (unpub. data).

marble horizon. However, a large area of hydrothermally altered feldspathic rocks occurs to the southwest along strike of the marble horizon in the syenite body near its border (fig. 1, locality 2). In this area, feldspar is partially to completely altered to sericite, chlorite, and carbonate; ferrosilicate minerals are altered to pyrite. The sericite-pyrite alteration, however, shows only slight potassium gain and significant sodium loss relative to unaltered syenite, on an aluminum-normalized basis (fig. 5 and tables 1 and 3). Approximately 2.5 to 10 times more sodium is lost relative to potassium gained on a molar basis. Uranium and strontium are preferentially lost relative to thorium and barium during sericite alteration, leading to increased Th/U and Ba/Sr ratios (table 1). A stockwork of pyrite-quartz-carbonate veins in the zone of sericite alteration (fig. 1, locality 2) contains anomalous As (20–460 ppm), Bi (2–16 ppm), Mn (to 1,300 ppm), Mo (7–13 ppm), Sn (5–30 ppm), and W (to 980 ppm). Trace amounts of molybdenite and wolframite have been identified in these veins. The altered rocks there are only slightly enriched in gold (generally <0.05 ppm) but have up to 1.7 ppm gold in places. Other small areas of hydrothermal alteration that contain pyrite-sericite-carbonate assemblages occur in the syenite stock near its margins (fig. 1, localities 4, 5). A small body of disseminated molybdenite-pyrrhotite-pyrite mineralization, associated with sodic alteration, is also present in the syenite near its western contact with Proterozoic gneiss (fig. 1, locality 3; see also Eggleston, 1918; Jacobs, 1937[?]; Schmidt, 1978).

INTERPRETATION

The mineralogy and geochemistry of the pyrrhotite replacement deposit at Cuttingsville allow constraints to be placed on the conditions of mineralization and the nature of the hydrothermal fluids. The early ore-stage alteration occurred at temperatures less than 325 °C on the basis of the stability of ferroan dolomite + quartz versus actinolite (see also experimental data and metamorphic paragenesis discussed by Winkler, 1974, p. 111–124) and chalcopyrite+hexagonal pyrrhotite versus cubanite+pyrite (Barton, 1979, fig. 7.22). During these stages, temperatures exceeded 250 °C on the basis of the presence of hexagonal pyrrhotite; however, the local occurrence of monoclinic pyrrhotite indicates that alteration at temperatures less than 250 °C took place at least locally. Sulfur fugacities were near the pyrrhotite-pyrite buffer and within the stability field of arsenopyrite. The massive pyrrhotite lenses formed by complete replacement of calcite in the marble. Fringe zones of ferroan dolomite and 10 to 20 volume percent pyrrhotite surrounding the massive pyrrhotite lenses were formed by Fe-Ca cation exchange with the calcite marble. Resultant porosity was filled with coprecipitated pyrrhotite. The late siderite+pyrite assemblages formed at lower temperatures than the other ore-stage assemblages and involved iron enrichment and sulfidation by the reaction of ferroan dolomite+pyrrhotite to siderite+pyrite.

The association of gold with elements such as arsenic and bismuth and anomalously low concentrations of the common base metals such as copper, lead, and zinc in the ores implies that gold was geochemically separated from base metals during the transport and (or) depositional stages of mineralization. Because the probable source rocks supplying metals to the hydrothermal fluids probably had extremely low contents of gold (Gottfried and others, 1972;

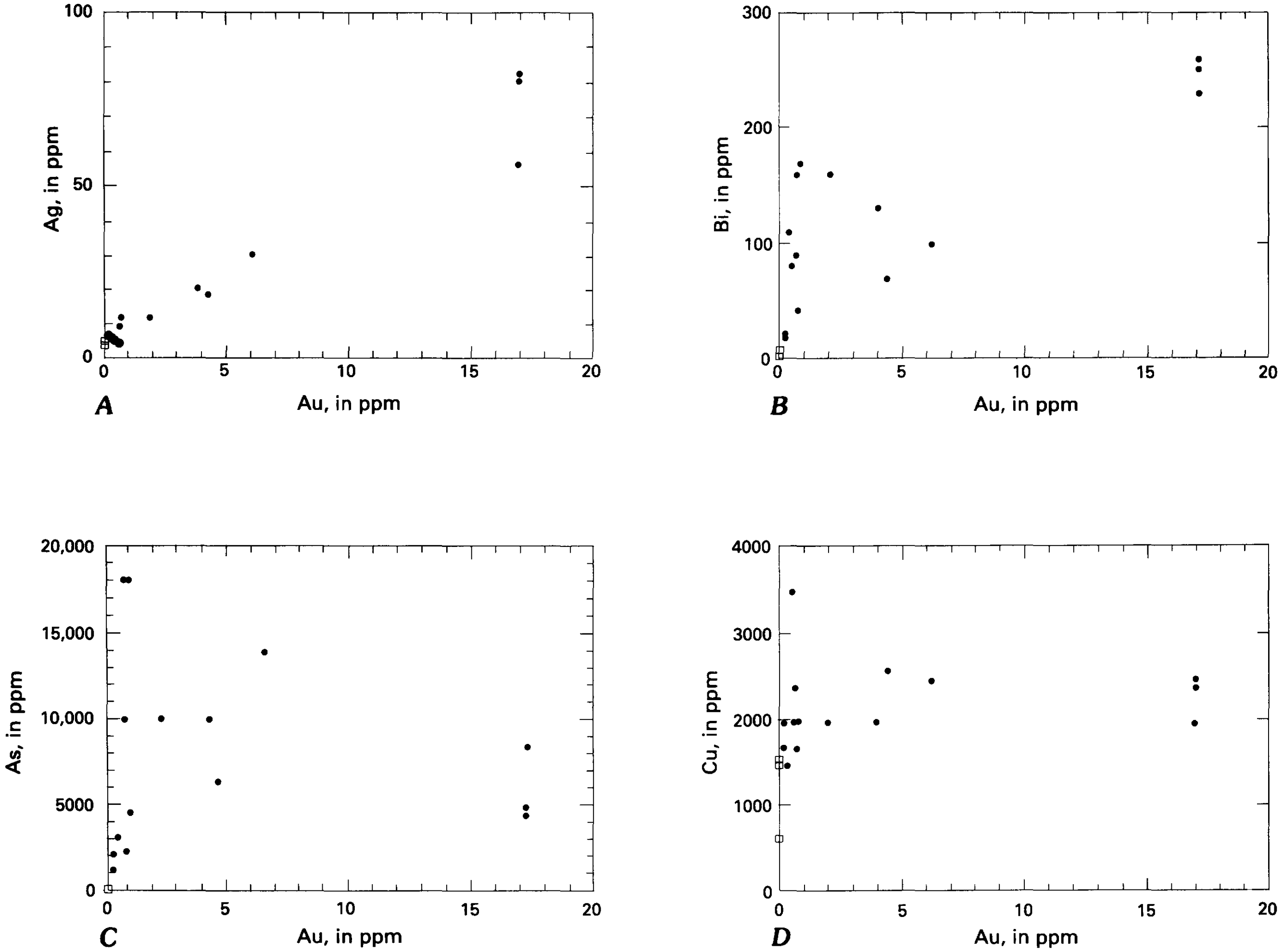


Figure 8. Chemical characteristics of the pyrrhotite replacement body at Cuttingsville, Vt. Solid circle, geochemical analyses of samples from the massive pyrrhotite body; open square, geochemical analyses of samples of stratabound pyrrhotite bodies of metamorphic origin, included here for comparison. *A*, Plot of Ag versus Au. *B*, Plot of Bi versus Au. *C*, Plot of As versus Au. *D*, Plot of Cu versus Au.

Tilling and others, 1973), arsenic, and bismuth and orders of magnitude greater base-metal concentrations, the fluids must have had properties that effectively leached and transported gold, arsenic, and bismuth while not significantly dissolving base metals. At temperatures less than 350 °C, a geochemical separation of gold, arsenic, and bismuth from the common base metals can take place under fluid conditions of (1) moderately low oxygen fugacity, (2) neutral to alkaline pH, and (or) (3) low chloride contents (Seward, 1984). Under these conditions, elements that form soluble complexes with reduced sulfur species, such as Au, As, perhaps Sb, Bi, and to some degree Ag, will be transported instead of elements that form insoluble sulfides, such as Cu, Zn, and Pb. Removal of sulfur from the hydrothermal fluid through the precipitation of iron sulfides leads to the deposition of metals transported as reduced sulfur complexes (for example, gold) while enhancing the solubility of metals transported as soluble chloride complexes (for example, zinc, lead).

Mineral assemblages and geochemical data indicate that the hydrothermal fluids that formed the replacement pyrrhotite deposit were of relatively low ionic strength and generally low f_{S_2} and f_{O_2}. If the pyrite-sericite alteration in the feldspathic rocks is indicative of the character of the fluids entering the marble horizon, the fluids were of near-neutral pH, of moderate to low ionic strength, and rich in CO_2, but undersaturated with respect to calcite. Reaction of the fluids with calcite led to the deposition of pyrrhotite by the reaction $14Fe^{+2} + 16HS^- + 16CaCO_3 + 4H^+ + O_2 = 2Fe_7S_8 + 16Ca^{+2} + 16HCO_3^- + 2H_2O$. Pyrrhotite deposition lowered the reduced sulfur concentration in the fluid, decreasing the stability of bisulfide complexes and leading to the deposition of gold by the reaction $14Fe^{+2} + 8Au(HS)_2^- + 2H_2O = 2Fe_7S_8 + 8Au + 20H^+ + O_2$. In addition, deposition of gold in the sulfide assemblage is enhanced by coupled substitution between +2 cations, such as Fe^{+2} and Cu^{+2}, and +1 and +3 cations, such as Au^+, Ag^+, Bi^{+3}, and As^{+3} (Barton, 1970, p. 193). The correlation of gold variation with that of silver and bismuth and the general occurrence of high gold in arsenic-rich samples from Cuttingsville is consistent with this mechanism for the mineralization.

OTHER CUSMAP STUDIES IN THE CUTTINGSVILLE AREA

Three other USGS studies of the Glens Falls CUSMAP area provide additional data relevant to this investigation. A summary of the pertinent data is given below and the reader should refer to these papers for more detailed information.

An airborne radiometric survey sensitive to the chlorophyll absorption band in spectra of trees was conducted over the Cuttingsville area (Power and Milton, this volume). Spectrally anomalous areas were identified that show a shift in the characteristic long wavelength edge of the chlorophyll absorption band and are inferred to be associated with areas of anomalous soil metals. A number of the anomalous areas correspond with known locations of hydrothermal alteration. The areas identified as spectrally anomalous include

1. the molybdenite prospect (fig. 1, locality 3),
2. the southern portion of the pyritic stockwork (fig. 1, locality 2),
3. two areas flanking Copperas Hill, along the strike belt of the replacement pyrrhotite deposit (fig. 1, locality 1),
4. two areas west of the dashed line enclosing the area of xenolith-rich dikes in the vicinity of Shrewsbury (fig. 1),
5. an area near the crest of Granite Hill, and
6. an area south of Hateful Hill, near the contact of the intrusive stock with Proterozoic rocks.

A regional airborne magnetic and VLF survey was conducted in the Cuttingsville area (Long, this volume). Of particular interest are two areas of high magnetic response coincident with low resistivity that may be indicative of hydrothermally altered rock, a fracture zone, or a mineralized zone (Long, this volume, fig. 9). The western area of coincident geophysical anomalies is approximately along strike of the intrusive contact between monzodiorite and nepheline-bearing syenite and is in the vicinity of a small stockwork vein system controlling hydrothermal alteration and minor gold enrichment at locality 4 (fig. 1). The eastern area of coincident geophysical anomalies occurs at a colluvium-covered area south of the area of hydrothermal alteration at locality 2 (fig. 1).

The results of a regional geochemical survey of the Glens Falls quadrangle utilizing panned heavy-mineral concentrates from stream sediments is described by Watts (this volume). This geochemical survey identified boron, niobium, tungsten, and tin anomalies in the vicinity of Shrewsbury, in the area of xenolith-rich dikes located north of the Cuttingsville stock. Enrichment in this suite of elements is interpreted as a favorable indicator for the presence of metallic mineralization in the Cuttingsville area (Watts, this volume; Slack, this volume).

POTENTIAL FOR PRECIOUS-METAL MINERALIZATION

The mineralization and alkalic igneous suite at Cuttingsville are similar to those of other gold deposits associated with small syenite stocks and dikes that have minor molybdenum mineralization. The nature of the Cuttingsville intrusive complex, the trace-element signatures of both plutonic and hydrothermally altered rocks, and the types of alteration associated with the Cuttingsville pluton, generally fit a model for gold deposits associated with alkaline intrusive rocks recently proposed by Mutschler and others (1985). Examples of other gold mineralization associated with alkaline intrusive rocks include the replacement pyrrhotite deposits at Ketza River, Yukon Territory (Morin, 1982; Toohey, 1986); the replacement and stockwork gold deposits in the northern Black Hills, S.D. (Paterson and others, 1988), and the Judith Mountains area of Montana (Weed and Pirsson, 1898; Giles, 1983); the disseminated and stockwork deposits in the Matachewan area, Ontario (Sinclair, 1982), and possibly similar deposits at Vunda, Fiji (Lawrence, 1978); the Barnat Mines Limited deposit, northwestern Quebec (Issigonis, 1980); and the Golden Sunlight property, Montana (Porter and Ripley, 1985). The replacement pyrrhotite body at Cuttingsville closely resembles the Ketza River gold deposits located in the Pelly Mountains, Yukon Territory (Morin, 1982, p. 99; Toohey, 1986), which appear to be associated with syenitic intrusives (Mortensen, 1982) hosting minor molybdenum mineralization (Wayne Goodfellow, written commun., 1985).

Detailed studies in the Matachewan area have documented an association of gold with syenite. Here gold was produced from low-grade disseminated and structurally controlled mineralization associated with epizonal alkalic syenitic intrusions of Archean age that contain sparse molybdenum and copper (Sinclair, 1982). In this area, syenite-hosted gold and auriferous copper-molybdenum deposits have styles of mineralization and alteration that are similar to those of alkaline porphyry copper-molybdenum systems (Sinclair, 1982). These gold-bearing syenites are typically pink to brick-red and were formed by potassic alteration of gray syenite to potassium feldspar or adularia. The mineralized rocks commonly are highly fractured and are cut by quartz and quartz-carbonate veinlets. Mineralized syenite at Matachewan contains 0.015 to 8 ppm gold (average grade of 0.8 ppm); copper contents are low, typically in the range of 13 to 325 ppm (average=102 ppm). Gold occurs as native gold on disseminated pyrite and in stockwork veinlets. The mineralized syenites generally contain a few percent disseminated pyrite, although pyritic syenite is not uniformly auriferous. Associated accessory minerals include chalcopyrite, molybdenite, and scheelite. Such syenite-hosted gold deposits may represent pyritic alteration zones related to auriferous copper-molybdenum systems (Sinclair, 1982).

The auriferous copper-molybdenum deposits are similarly associated with pink areas of potassic alteration in syenite but have lower precious-metal contents (0.001 to 2.8 ppm gold; average=0.2 ppm) and higher base-metal contents (54 to 3,700 ppm copper; average=835 ppm) than the syenite-hosted gold deposits. The pyrite content also is lower (0.5 volume percent or less) than that of the syenite-hosted gold deposits.

The Cuttingsville stock has potential for disseminated and stockwork gold mineralization similar to that of the Matachewan occurrences. Although there are some chemical differences, the chemistry of the intrusive magma suite comprising monzodiorite, syenite, and nepheline syenite at Cuttingsville resembles that in the Matachewan area (Ploeger and Crocket, 1982; Sinclair, 1982) and the intrusive suite in the Pelly Mountains (Mortensen, 1982). For instance, the Na_2O/K_2O ratio in the intrusive suite at Cuttingsville varies from about 2.5 to 1.5 with differentiation, whereas the suites in the Matachewan and Pelly Mountain areas have Na_2O/K_2O ratios near 1 and locally have $K_2O>Na_2O$. All three of the intrusive suites have high contents of niobium, zirconium, and yttrium and high Nb/Y and Zr/TiO_2 ratios, which are characteristic of alkaline magmatism (Winchester and Floyd, 1977). At Cuttingsville, the pyrite-sericite-quartz-carbonate alteration in feldspathic rocks and the ankerite-siderite alteration in carbonate rocks resemble the alteration associated with felsic- and augite-syenite-hosted vein mineralization at the Macassa mine in the Matachewan area (Kerrich and Watson, 1984) and the CO_2-K-As alteration in mafic rocks associated with gold mineralization elsewhere in the Matachewan area (Davies and others, 1982). Significantly, some analyses of altered syenitic rocks from Cuttingsville that contain a few percent pyrite from localities 2, 3, and 4 (fig. 1) have anomalous contents of gold generally in the range of 0.1 to 0.3 ppm but, in places, as high as 1 ppm. The distribution of these high gold values appears to be erratic, however.

The gold at Cuttingsville is the first documented report of gold mineralization genetically associated with hydrothermally altered alkaline rocks of the White Mountain Plutonic-Volcanic Suite in New England. However, data from the literature suggest that several other reported gold occurrences in New England and southeastern Quebec may be associated with similar Mesozoic intrusives. The Norton mine, on Mount Monadnock in northeastern Vermont, is reported as a gold mine by Morrill and Chaffee (1964). The mineralized zone there is localized within a pink altered syenite of the White Mountain Plutonic-Volcanic Suite, which is cut by an altered bostonite dike (Morrill and Chaffee, 1964, p. 25; Wolff, 1930, p. 146). Dresser (1908) and McGerrigle (1934, p. 91) report a small placer gold locality on the branch of the Chisham River, in the Lac Megantic area of southeastern Quebec, which drains a syenite ring dike on the northeastern side of Mont Megantic. The occurrence of angular gold nuggets, some that have almost needlelike projections (McGerrigle, 1934, p. 94), suggests that the local source may be gold-bearing veins associated with the syenite intrusion. In the same area, Gauthier (1986) reports the presence of a zone of hydrothermal alteration that has anomalous molybdenum, tungsten, and tin. Recently, a silicified and sericitized polymetallic stockwork has been discovered nearby at St. Hermenegilde that contains pyrrhotite, chalcopyrite, molybdenite, native bismuth, bismuth tellurides, and up to 0.18 g/t gold (Gauthier and others, 1989).

The gold metallization at Cuttingsville, Mount Monadnock, and St. Hermenegilde suggests that a potential for porphyry-related gold exists with other syenitic intrusives of the White Mountain Plutonic-Volcanic Suite in New England and southeastern Quebec. Carbonate-hosted replacement-type gold mineralization, such as at Cuttingsville and the Ketza River deposit, Yukon, appears to be preferentially developed in carbonate rocks (principally limestone and calcite marble) above or adjacent to small cupolas of alkali syenite in the upper parts of larger alkalic intrusive complexes. Mineralization involves replacement and dolomitization of the host limestone that results in the development of lenses and irregular zones of massive sulfide that generally have low base-metal contents. Low-grade disseminated and structurally controlled gold deposits associated with epizonal intrusions, such as that in the Matachewan area, characteristically have high sulfide contents (typically >2 percent pyrite), sparse molybdenum and copper mineralization, and significant potassium enrichment (commonly pink). These features imply that gold was transported in alkaline solutions as sulfide complexes, similar to the fluids inferred to have formed the replacement pyrrhotite deposit at Cuttingsville. In both types of deposits, gold was precipitated in association with iron sulfides, probably as a consequence of sulfidation of iron in wall-rock silicates and (or) breakdown of gold-bisulfide complexes. Favorable exploration criteria for locating either syenite-hosted gold-bearing disseminations and stockworks or auriferous carbonate-hosted replacement mineralization have been identified, on the basis of the criteria of Mutschler and others (1985) and by analogy with the mineralized systems in the Pelly Mountains and Matachewan areas of Canada. These exploration criteria are

1. Small epizonal syenite stocks that have
    a. accessory molybdenum and (or) copper mineralization,

b. areas of hydrothermal alteration that have pink potassic alteration and quartz-carbonate-pyrite-sericite stockworks,

c. contents, greater than a few percent, of disseminated sulfide (pyrite) in intrusive rocks,

d. associated trachyte, bostonite, and xenolith-rich dikes, and (or)

e. local geochemical anomalies among one (or more) of the following elements: Au, Ag, As, Bi, Hg, Mo, Te.

2. Areas of xenolith-rich and trachytic syenite dikes that have associated sulfide-bearing alteration, for example:

a. areas of trachyte dikes, especially where abundant, are inferred to indicate proximity to a syenite stock, possibly a cupola; the trachyte dikes are interpreted as offshoots from syenitic intrusions and may not propagate great distances from the parent body (McHone and Corneille, 1980, p. 20–21),

b. local geochemical anomalies among one (or more) of the following elements: Au, Ag, As, Bi, Hg, Mo, Te.

3. Calcite-rich host rocks proximal to favorable igneous bodies described above.

The resource assessment method described by Slack (this volume) for the Glens Falls 1° × 2° quadrangle uses both diagnostic and permissive recognition criteria to rank areas in terms of mineral resource potential. Favorable geology and known mineral occurrences are diagnostic criteria because they are required to be present in the vicinity of nearly all mineral deposits of a given type. Permissive criteria, such as geochemical anomalies, suggest the presence of a particular deposit type but are not required. The absence of a known geochemical anomaly does not preclude the occurrence of a deposit; similarly, the presence of an anomaly does not require the occurrence of a deposit. The presence of only one of these factors alone yields a low resource potential ranking in the scheme of Gair (1989). The combination of a geochemical anomaly with either favorable geology or a known mineral occurrence yields the ranking of a moderate resource potential. The combination of two diagnostic criteria, such as favorable geology and a known mineral occurrence substantiated by a geochemical anomaly, yields a high resource potential. By using these ranking criteria and the favorable geology and geochemical anomaly criteria outlined above, the following target areas that have potential for gold mineralization associated with alkaline intrusives have been identified.

A low resource potential is assigned to the White Mountain Plutonic-Volcanic Suite in general, due to the broad association of this igneous suite with favorable geologic criteria. Diagnostic mineral occurrences are unknown at present, however, numerous geochemical anomalies in arsenic and bismuth are reported for rocks from the White Mountain Plutonic-Volcanic Suite in New Hampshire (Cooley and others, 1984). Specific locations in New England of assigned low potential for both disseminated and stockwork gold mineralization within the White Mountain Plutonic-Volcanic Suite are areas that have favorable syenitic and trachytic host rocks; examples include

1. Outcrop areas of calcite marble on the northwestern, southern, and southeastern sides of the Cuttingsville stock (fig. 1).

2. The syenite stock in the Barber Hill area, Charlotte, Vt.

3. Trachyte dike swarms in the Orchard Point-Shelburne Point areas, Shelburne, Vt.

4. Trachyte dike swarms near Essex, N.Y.

5. Syenite bodies and ring-dikes in the White Mountains of New Hampshire and Maine.

A moderate resource potential for gold mineralization associated with alkaline intrusive rocks is assigned to areas that have favorable geology, local rock geochemical anomalies in gold, arsenic, bismuth, or molybdenum, and (or) known mineral occurrences. Reported gold occurrences that lack geochemical or mineralogical verification and gold occurrences that have an unknown gold source are included in this category. These specific areas are

1. The syenite stock and associated bostonite dikes on Mount Monadnock, Vt.

2. The syenite stock and associated xenolith-rich dikes in the Pollard Hill area, N.H. (see Slack, this volume).

3. East of the town of Ossippee, N.H., in the vicinity of the Pocket Mountain gold prospect (Myers, 1941, locality no. 120; Morrill, 1960).

4. The White Diamond Gold mine in the vicinity of Center Sandwich, N.H. (Myers, 1941, locality no. 156). This is probably the same occurrence as the Diamond Ledge gold mine in the Mount Chocorua quadrangle described by Morrill (1960, p. 33) from which $9,000 in gold was delivered to the Philadelphia mine.

5. The vicinity of composite stocks of the White Mountain Plutonic-Volcanic Suite in the Mount Lafayette and Twin Mountain areas, east-southeast of Franconia, N.H. Placer gold "nuggets" are reported by Morrill (1960) from the north-central part of the Franconia 15-minute quadrangle, presumably from the Gale River that drains the western side of these stocks.

6. An area downstream from the junction of the Carrigan Notch Trail and the Desolation Trail, Crawford Notch 15-minute quadrangle, N.H. Placer gold is reported by Morrill (1960, p. 21).

7. The vicinity of the "long quartz vein," North Conway 15-minute quadrangle. A "gold mine" is reported to be located 400 yd from the vein (Morrill, 1960, p. 7).

8. The vicinity of the composite stock and small intrusive bodies of alkali syenite and granite at Mount Agamenticus and Cape Neddick, Maine. Gold nuggets and gold dust have been described from dredged sediments from the mouth of the Josias River at Perkins Cove, Ogunquit,

Maine (Morrill and others, 1964). D.A. Gust (oral commun., 1989) also reports the occurrence of sulfide-bearing greisens and pyritic rhyolites associated with the Mount Agamenticus complex (see Brooks and others, 1989).

9. The vicinity of the small syenite stock southwest of Denmark, Maine, and the small syenite bodies south of Brownsfield and north of Kezar Falls, Maine (includes the Burnt Meadow Mountain complex). Morrill and others ([?], p. 24) report the Frenchman's gold mine as occurring 1/2 mi southwest of Hiram, Maine, and report a possible gold mine shaft in the Kezar Falls-Parsonsfield, Maine, area. The presence of gold is unverified.

10. The vicinity of Acton, Maine. A series of mines in an arsenic-rich silver-gold vein are reported along the eastern side of the Little River (Rand, 1958; Morrill and others, [?]). The genesis of the veins is unknown, but a small body of alkali gabbro and volcanics of the White Mountain Plutonic-Volcanic Suite occurs south of Acton, Maine.

Areas that have an assigned high resource potential for either disseminated- and (or) replacement-type gold mineralization associated with alkaline intrusive rocks are those localities that have favorable geology, local geochemical anomalies in gold, arsenic, bismuth, and molybdenum, documented hydrothermal alteration, and known mineral occurrences that have documented gold enrichment. Two areas thus defined are at (1) Cuttingsville, Vt., including the syenite stock bodies of nearby calcite marble and the area of xenolith-rich dikes at Shrewsbury, Vt., and (2) Mount Megantic in southeastern Quebec.

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Chapter P

Epigenetic Uranium Mineralization in the Middle Proterozoic Mount Holly Complex Near Ludlow and Jamaica, Vermont

By ROBERT A. AYUSO and CHARLES A. RATTÉ

U.S. GEOLOGICAL SURVEY BULLETIN 1887

SUMMARY RESULTS OF THE GLENS FALLS CUSMAP PROJECT, NEW YORK, VERMONT,
AND NEW HAMPSHIRE

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Epigenetic Uranium Mineralization in the Middle Proterozoic Mount Holly Complex Near Ludlow and Jamaica, Vermont

By Robert A. Ayuso[1] *and* Charles A. Ratté[2]

Abstract

The most important bedrock uranium deposits known in New England are the stratabound veins, pods, and lenses near the Vermont towns of Ludlow and Jamaica, within the Middle Proterozoic Mount Holly Complex in the Green Mountain massif. West of Ludlow, at Grant Brook and Ludlow Mountain, the host rocks to the uranium deposits are sheared, fractured, and cataclastically deformed quartzite. At College Hill and Pinnacle Hill near Jamaica, the host rocks consist of micaceous gneiss interlayered with quartzofeldspathic gneiss, micaceous schist, and pegmatitic zones. Highly mineralized veins at College Hill occur mainly near pegmatite-host rock contacts, and at Pinnacle Hill the mineralization is contained along foliation surfaces in rusty-weathering zones within micaceous schist. Near the top of Pinnacle Hill, however, the mineralization is within a mafic gneiss and forms discontinuous, micaceous bands and (or) lenses. The veins are mineralogically simple and consist of uraninite, quartz, and widely varying amounts of minor and accessory minerals. Near Ludlow, tourmaline is the most important accessory mineral, and near Jamaica, epidote, magnetite, tourmaline, garnet, and chlorite are present in varying amounts.

The uranium mineralization near Ludlow and Jamaica is almost monoelemental in that, with the exception of lead, no significant enrichments occur in thorium, arsenic, or molybdenum or in base, precious, or any other lithophile elements or heavy metals. The bulk composition of the host rocks varies greatly, and strong correlations are lacking between uranium content and alkali element contents or between the concentrations of uranium and those of other major elements. Many samples that contain high uranium abundances at Ludlow Mountain also have high values of Fe_2O_3/FeO. Total rare earth element content of the host rocks is widely variable in mineralized areas near both Ludlow and Jamaica and is not directly related to uranium content.

Uranium values range up to about 1,160 parts per million in rocks closely associated with the mineralization near Ludlow. Uranium and boron are positively correlated, which agrees with the observed association of mineralized pods and veins and tourmaline-rich rocks. Values of U_3O_8 range from 0.2 to 0.6 percent in the mineralized veins and pods; however, values as high as 22 weight percent U_3O_8 have been reported. The relatively low thorium content and low total rare earth element content in the mineralized zones, and the lack of a strong correlation of these elements with uranium, argue against an origin for the mineralization by direct derivation from anatectic melts. We suggest instead that, during the peak prograde metamorphism in the Late Proterozoic, partial melting of the host rocks developed uranium-enriched anatectic pegmatites. Subsequent metamorphic events produced fluids that migrated along shear surfaces and remobilized and concentrated uranium during the Taconian and Acadian orogenies.

INTRODUCTION

Uranium-rich mineralized areas in the Proterozoic basement massif of southern Vermont may be the largest bedrock uranium deposits known in New England. A preliminary evaluation by McHone and Wagener (1982) of uranium potential in these basement rocks indicated a favorable environment for uranium deposits. Epigenetic uranium-bearing veins and small pegmatitic pods and flat lenses occur west of Ludlow and near Jamaica, Vt., on the eastern side of the Green Mountain massif (fig. 1). In both mineralized areas, anomalously high contents of uranium are found in the Mount Holly Complex, a Middle Proterozoic Grenvillian basement complex that forms the core of the Green Mountain massif. The Mount Holly Complex in these areas consists largely of quartzite and quartzofeldspathic gneiss and minor hornblende gneiss and pegmatite (Doll and others, 1961). In the Ludlow area, anomalously high amounts of uranium form a discontinuous stratabound zone for about 1.5 km along strike; in the Jamaica area, the mineralization occurs in several zones of varying thickness for at least 3 km. The full extent of the mineralization in both areas has not been reliably determined, however.

In this paper we provide a general description of the rocks that host the uranium deposits, a geochemical characterization of mineralized and unmineralized rocks of the

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Manuscript approved for publication April 13, 1990.

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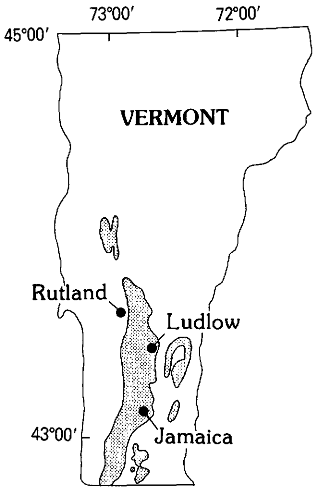


Figure 1. Generalized map showing the locations of the two mineralized areas near Jamaica and Ludlow, Vt. Both of these areas are on the eastern border of the Green Mountain massif (patterned area) near the boundary with metagraywackes, metapelites, and mafic metavolcanic rocks (greenstones) of Proterozoic through Ordovician age (Doll and others, 1961; Karabinos, 1984a). Host rocks that contain the mineralization are units within the Middle Proterozoic Mount Holly Complex. Figure modified after Doll and others (1961).

Mount Holly Complex, and an interpretation of the origin of the uranium deposits. The geologic and geochemical features presented here may aid in the recognition of concentrations of uranium in similar crystalline terranes elsewhere in the Green Mountains and in New England.

HOST ROCKS TO THE URANIUM MINERALIZATION

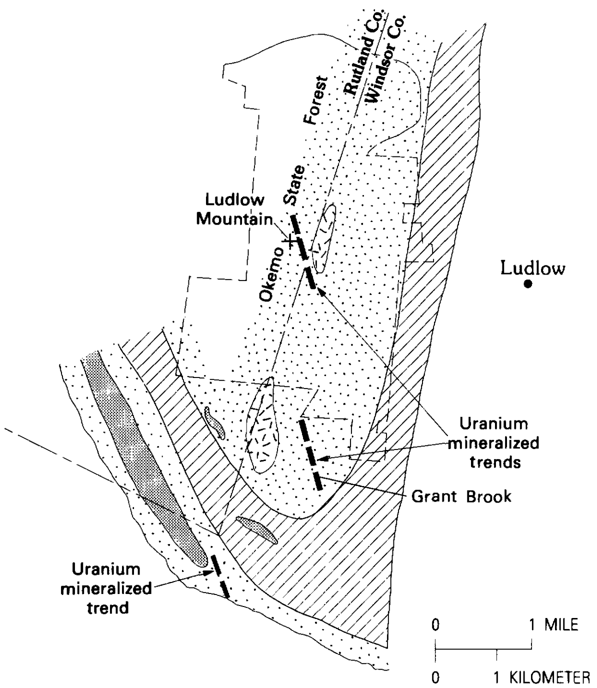
West of the town of Ludlow, at the Grant Brook and the Ludlow Mountain localities, and near the town of Jamaica, at College Hill and Pinnacle Hill, the mineralized zones are confined to the Middle Proterozoic Mount Holly Complex (Doll and others, 1961). In the northern part of the Green Mountain massif, new geologic mapping to the north of Ludlow by Karabinos (1987) documents basement-cover relations, several major thrust faults, and previously unrecognized bodies of augen gneiss (interpreted as metamorphosed granites and (or) rhyolites); U-Pb zircon dates reveal that the augen gneisses of this region crystallized at about 1120 Ma (Karabinos and Aleinikoff, 1988). In the Jamaica area, Ratcliffe and others (1988) show the College Hill-Pinnacle Hill area to be underlain by biotite paragneiss, metavolcanic rocks, and biotite granite gneiss; a few kilometers to the north of the mineralized zones is a large body of metaigneous rocks composed of metatrondhjemite, metatonalite, and metadacite (Ratcliffe and others, 1988, fig. 15). Recent U-Pb zircon dating by Aleinikoff and others (1990) indicates that the massive metadacites near Jamaica are 1351±3 Ma, a coarse-grained metatonalite is 1308±10 Ma, and the granite gneiss on College Hill is 1244±8 Ma.

Near Jamaica, complex changes in P-T conditions and intense deformation occurred as a result of Proterozoic and Paleozoic events (Karabinos, 1984a,b; Chamberlain and Karabinos, 1987). Major Proterozoic regional metamorphism recognized in the area was a prograde event at about 1100 Ma (Ratcliffe and Aleinikoff, 1990), which reached at least upper amphibolite facies conditions and produced anatexis of the country rocks and the formation of local granitic pegmatites (Brace, 1953; Skehan, 1961). Subsequent events involved deformation and metamorphism during the Taconian (Ordovician) and Acadian (Devonian) orogenies (see Burton and others, 1990). According to Karabinos (1984a), garnet grew during both Paleozoic events; Karabinos also indicated that, in the vicinity of Jamaica, two major thrust faults subdivide the area into three basement-cover sequences and thrust faulting transported Proterozoic basement rocks and their cover sequences westward. The age of thrusting near Jamaica is not well constrained, but Karabinos (1984a) tentatively suggested that it is Ordovician on the basis of correlations with Ordovician thrusting in the nearby Taconic Mountains (see, for example, Zen, 1967; Ratcliffe and Hatch, 1979).

GRANT BROOK AND LUDLOW MOUNTAIN

Detailed mapping of the Mount Holly Complex at Okemo State Forest (fig. 2) has outlined a variety of metasedimentary rock units that are modally and texturally discontinuous along strike (C.A. Ratté, unpub. field data). The modal composition of the metasedimentary units is more feldspathic down section, and the units are, in some places, intensely sheared. Such discontinuous shear surfaces locally parallel an earlier foliation and also show protomylonitic and mylonitic textures. Microbreccia is developed in places in the axial (hinge) zones of tight drag folds.

In Okemo State Forest, quartzite (either massive clean or micaceous or feldspar-rich or sheared) is the dominant lithology on the eastern slope of Ludlow Mountain (fig. 2). The massive quartzite consists of beds of clean quartz sand that characteristically contain up to 92 percent quartz and are interlayered with thin beds and (or) laminae of silty or micaceous sand. This unit is primarily of metasedimentary origin. In some places, these massive quartzites are more noticeably gneissic. As the mica content increases, the quartzites become more schistose and result in pelitic gneisses. Feldspar-rich quartzites or arkoses that have a gneissic appearance occur near the base of the massive quartzite unit on the western slopes of Ludlow Mountain. The uranium mineralization occurs in discontin-



EXPLANATION

MIDDLE PROTEROZOIC MOUNT HOLLY COMPLEX

▨ Undivided, mainly biotite-muscovite gneiss, abundant amphibolite, and hornblende gneiss; mica schist, quartzite, and calc-silicate granulite; contains numerous pegmatites and massive granitic and gneissic rocks

⛶ Mainly quartzite and quartz-muscovite schist; contains pegmatites

▩ Calcite and dolomite marbles

▧ Undivided granitic gneisses

Figure 2. Generalized geologic map of the Ludlow area showing the Ludlow Mountain and Grant Brook uranium occurrences. Map modified after Doll and others (1961) and unpublished reports by exploration companies. See figure 1 for location of Ludlow area in the Green Mountain massif.

uous stratabound zones within sheared, mylonitic quartzites, especially where ilmenite, biotite, and tourmaline are present. The sheared quartzites display varying degrees of cataclasis and represent a tectonically derived variety of the massive quartzites exposed at Ludlow Mountain.

Petrographic study of the quartzites shows that recrystallization after shearing resulted in the development of finely textured mosaics of quartz and feldspar parallel to the principal foliation along former microshear surfaces. Phyllosilicate minerals in these samples are concentrated along a primary foliation and along crosscutting crenulate surfaces. Microboudins, microbreccia, aligned clusters of phyllosilicate minerals, and elongated grains of quartz formed during an intermediate and progressive cataclastic event that modified a previously developed fabric. Quartz rodding is common in outcrops of the sheared quartzites. Quartz-feldspar cataclastites that display a marked schistosity and (or) gneissosity developed at an advanced stage of shearing. This schistosity is exemplified by "strung-out" or elongated, rounded, and comminuted quartz and feldspar clusters that contain wispy, parallel streaks of mica and epidote. Ragged plagioclase porphyroblasts that have crenulated borders are partially rotated into the plane of the shear. In addition, aligned clusters of epidote and sericite suggest progressive alteration and shearing of plagioclase. These sheared quartzites, which show gneissic and schistose textures, are noticeably deficient in mafic minerals, except in places where both tourmaline and biotite have been introduced. The growth of epidote, chlorite, and white mica, at the expense of plagioclase, gives these rocks a distinctive green color and reflects a relatively moderate alteration event.

In the Ludlow area, especially on Ludlow Mountain, a significant association exists between highly mineralized pods and lenses and tourmaline-rich rocks, so that the host rocks to the mineralization may be referred to as tourmaline-rich quartzites or tourmalinites. Uranium mineralization is concentrated near shear zones in rusty, micaceous quartzitic schists (epidote-tourmaline-pyrite-hematite-ilmenite-chlorite-biotite-white mica quartz) and, in some instances, within feldspar-rich quartzites (epidote-tourmaline-microcline-perthite quartz). Highly mineralized intervals vary in width from about 10 cm to a few meters and are characterized by a simple mineralogy. A detailed study of the occurrence of the uranium minerals in the Vermont deposits, however, has not been done. Uranium-rich pods and lenses near Ludlow contain, in addition to tourmaline and quartz, small amounts of epidote, ilmenite, chlorite, garnet, and white mica.

COLLEGE HILL AND PINNACLE HILL

Near Jamaica, two uranium occurrences are found at College Hill and Pinnacle Hill (fig. 3). Host rocks that contain uranium mineralization at College Hill consist of complexly interlayered and gradational units of micaceous gneiss, quartzofeldspathic gneiss, and micaceous schist, informally referred to as the banded unit by exploration geologists. The thinly banded quartzofeldspathic gneiss consists of epidote, allanite, microcline, perthite, and quartz; other portions of the gneiss are pegmatitic and contain allanite, epidote, white mica, perthite, and quartz. Uranium mineralization is closely associated with pegmatite dikes, occurs mostly near pegmatite-host rock contacts, and, in some places, occurs within the pegmatites themselves. The pegmatite dikes range widely in size and length from centimeters to 20–30 m and commonly have a leucogranitic mineralogy. Thus, one important observation is that the pegmatites are characteristically associated with miner-

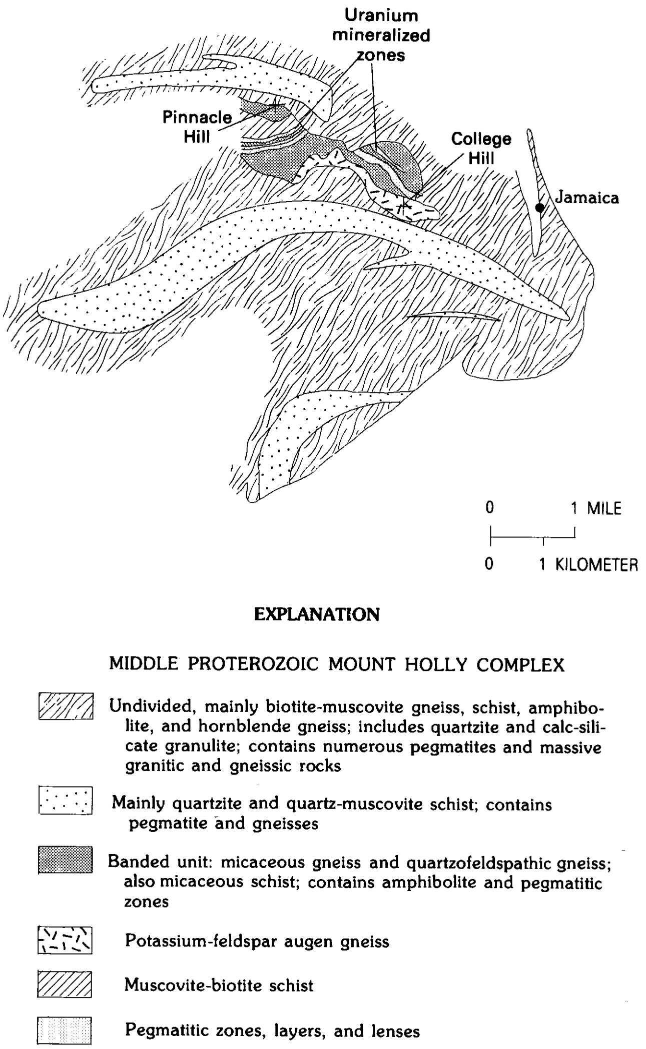


Figure 3. Generalized geologic map of the Jamaica area showing the Pinnacle Hill and the College Hill uranium occurrences. Map modified after Doll and others (1961) and unpublished reports by exploration companies. See figure 1 for location of Jamaica area in the Green Mountain massif.

alized zones. Potassium-feldspar augen gneiss and muscovite-biotite schist also have been identified in the Jamaica area but do not appear to be spatially related to the mineralization. The augen gneiss is primarily a quartzofeldspathic rock that contains prominent porphyroblasts of potassium feldspar. Muscovite-biotite schists interlayered with the banded unit occur mainly in the area near Pinnacle Hill.

As in the Ludlow area, uranium mineralization near College Hill is concentrated in shears, fractures, and joint surfaces that are closely associated with faulted and sheared contacts of pegmatitic bodies and in banded micaceous gneisses that contain abundant garnet, epidote, and magnetite. Uranium mineralization in the Pinnacle Hill area differs from that in College Hill because it is contained along foliation surfaces within micaceous bands in rusty-weathering amphibole and biotite-rich schist and in micaceous quartzite. The uranium mineralization also occurs in a mafic gneiss as dark and narrow, discontinuous bands that consist of varying amounts of tourmaline, garnet, chlorite, epidote, biotite, white mica, and quartz; the more quartzitic samples are generally similar to those on Ludlow Mountain. Some of the micaceous schist bands have been weakly to moderately altered by hydrothermal processes.

Typically, uranium oxide minerals are aligned parallel or subparallel to the dominant foliation and banding of the host rock and form irregular lenses and pods. In general, uraniferous areas of the Mount Holly Complex contain disseminated pitchblende, carnotite, and torbernite (McHone and Wagener, 1982). Near Jamaica, the important uranium-bearing minerals probably also include uraninite, small amounts of brannerite, and uranophane. Mineralized zones contain a similar mineral assemblage as the host rocks, including varying but significant quantities of epidote, magnetite, quartz, chlorite, garnet, and tourmaline. Trace amounts of molybdenite have been found in a few samples. Magnetite-rich samples, which are closely associated with the mineralization, are characteristically brecciated and enclose patches of subhedral epidote, subhedral and fractured garnet, clusters of chlorite, and quartz; these magnetite-rich pods are, in places, intensely altered to hematite.

A possible paragenetic sequence within some of the mineralized pods at College Hill involves the formation of epidote and garnet, followed by magnetite, chlorite, hematite, and quartz. Where tourmaline and biotite are present, tourmaline is embayed and ragged and seems to be replaced by fine-grained, subhedral epidote, chlorite, and quartz. Fractures within tourmaline grains also contain unidentified, zoned, and subrounded amorphous yellow clusters that may be secondary uranium minerals.

At College Hill, the host rocks may have undergone some alteration, including sodium metasomatism (as indicated by albitization), epidotization, hematization of magnetite and mafic minerals, sericitization of feldspar, and chloritization of the mafic micas. This is in contrast to the Ludlow Mountain area, where host rocks that contain uranium pods and veins lack clear evidence for sodium metasomatism, intense sericitization, or high-temperature hydrothermal alteration. At Ludlow Mountain, however, a period of tourmaline and biotite introduction has been recognized.

BEDROCK GEOCHEMISTRY

Bulk rock compositions of host rocks that contain uranium minerals in both the Ludlow and Jamaica areas are highly variable (figs. 4–9). Although the range in silica contents is extensive, from about 46 to 96 weight percent

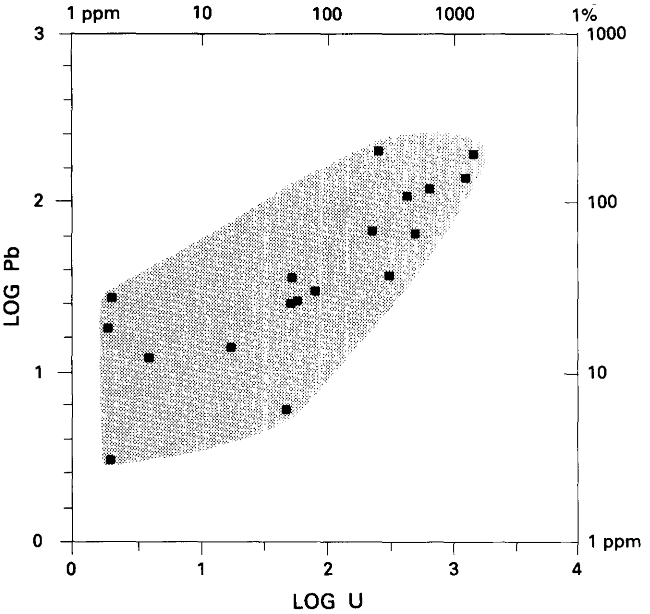


Figure 4. Logarithmic diagram showing the general correlation of lead and uranium in rocks of uranium occurrences near Jamaica, Vt.

SiO_2 (tables 1 and 3), it shows no direct correlation with uranium contents of the mineralized rocks (tables 2 and 4) (Tables follow References Cited section.). Similarly, no strong correlations are evident between uranium and alkali elements or between uranium and other major elements. However, relatively high values of the molar ratio of Fe_2O_3/FeO are found in many of the samples that contain high uranium from the Ludlow Mountain occurrences.

At Ludlow Mountain, the predominant opaque mineral is ilmenite, which occurs in the mineralized pegmatitic pods and flat lenses, as well as in the microbreccias within drag folds and along shear surfaces. Near Jamaica, magnetite and minor pyrite are probably the most common opaque oxides, but magnetite is particularly concentrated in the mineralized pods and lenses and is closely associated with the uranium oxides. Magnetite-rich samples from mineralized zones show relative enrichments in Y, Pb, some of the heavy rare earth elements, Zr, Hf, Cr, Co, and Ni, and Th and U (for example, samples J–16 and J–9, tables 3 and 4).

Near Jamaica, contents of Mo, Sn, W, Cu, Cr, Co, Ni, and Sc are not correlated with those of uranium. Gold,

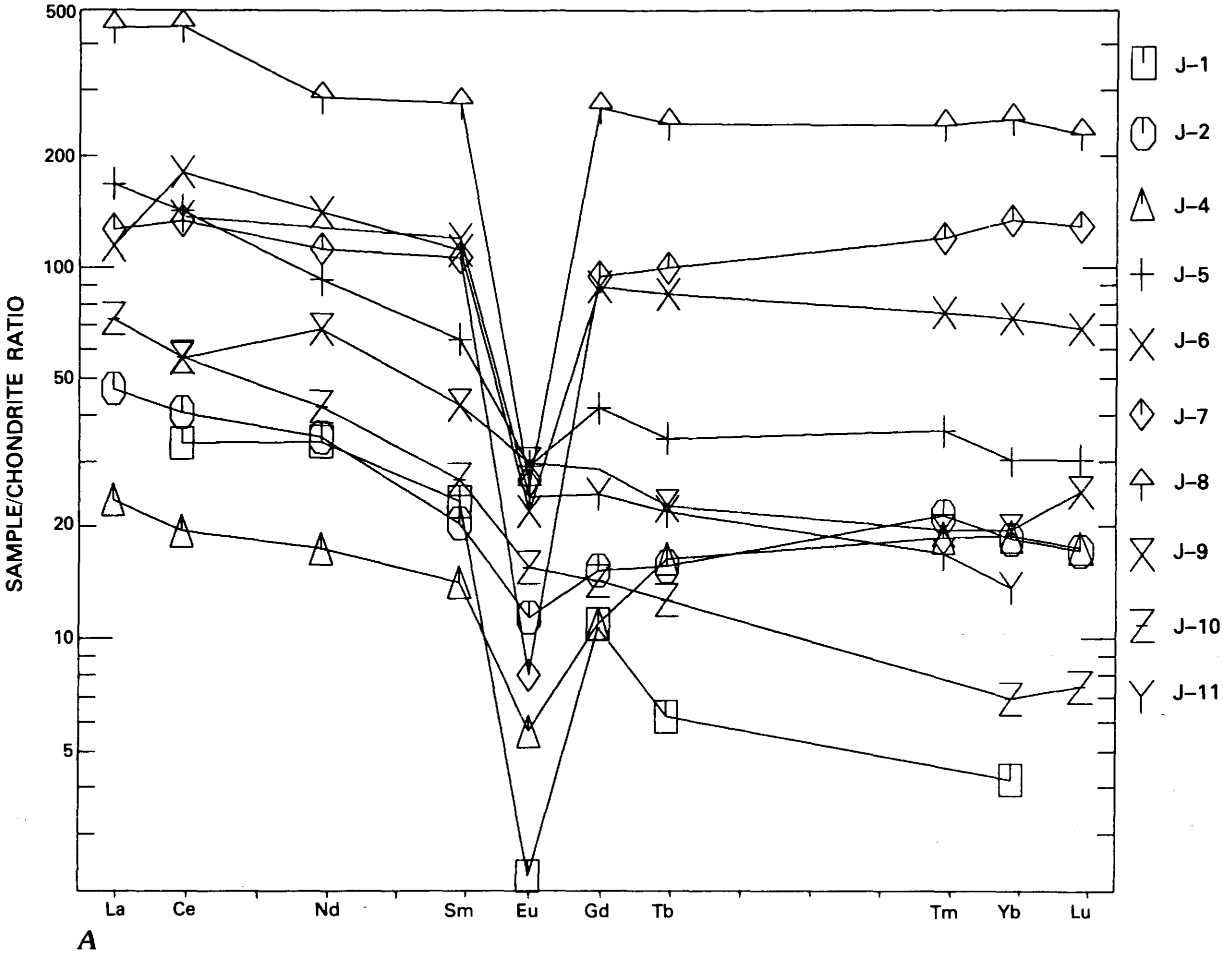


Figure 5. Chondrite-normalized rare earth element patterns for samples of the host rocks and mineralized pods and lenses from Vermont. *A*, Jamaica (samples J–1, J–2, and J–4 to J–11); *B*, Jamaica (samples J–12 and J–14 to J–17); *C*, Ludlow. Normalizing data summarized by Anders and Ebihara (1982).

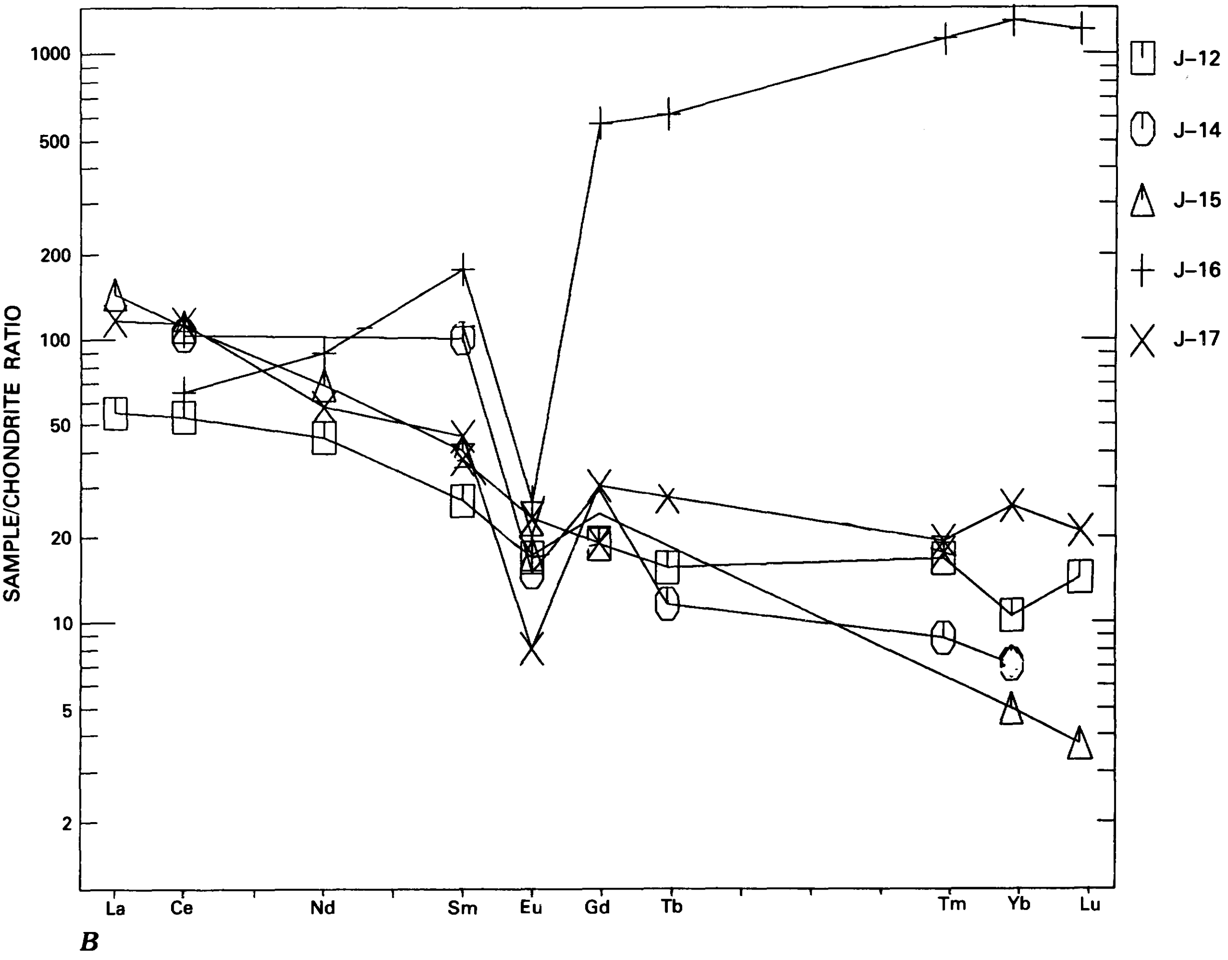


Figure 5. Continued.

silver, bismuth, and cadmium, which were found concentrated in local stream sediments analyzed by semiquantitative methods, are all characteristically below detection limits in rock samples (R.A. Ayuso, unpub. data). Lead contents are typically high (to 200 ppm, table 4) and show a general positive correlation with those of uranium (fig. 4). The uranium mineralization at both areas in Vermont is thus nearly monoelemental, as no significant enrichment in any other lithophile elements, thorium, arsenic, molybdenum, and base, precious, or heavy metals has been detected in these New England deposits.

Total rare earth element (REE) content of the host rocks is widely variable in both mineralized areas (tables 2 and 4) and, more importantly, is not directly related to uranium concentration. In most samples, the total REE content is generally low (<300 ppm), although in some samples it is substantially higher (1,500 ppm; tables 2 and 4). REE patterns on chondrite-normalized plots are characteristically flat to slightly negative in slope as a result of slight enrichment in the light REE compared to the heavy REE; prominent negative europium anomalies are common (fig. 5). Significantly, there is no obvious change in abundance or shape of the REE distributions with proximity to the mineralized zones. One important exception, however, is in a magnetite-rich sample (J–16, table 4, fig. 5*B*). Its chondrite-normalized pattern has a strong negative europium anomaly and a positive slope as a result of the substantial enrichment in heavy REE. Sample J–11 (table 4, fig. 5*A*), which contains high uranium, differs from the magnetite-rich sample in that it has a relatively minor europium anomaly and a pattern that shows moderate enrichment in light REE.

URANIUM AND THORIUM VARIATIONS

In the Ludlow and Grant Brook occurrences, uranium contents range up to about 1,160 ppm in the host rocks intimately associated with the uranium-rich veins, pods, and lenses (table 2). Thorium content in rocks from the area is not uniformly low for all samples but exhibits a wide range in values. Th/U ratios also show a wide range from about 12.4 to 0.008 and display a negative correlation with the uranium content of the rocks (fig. 6).

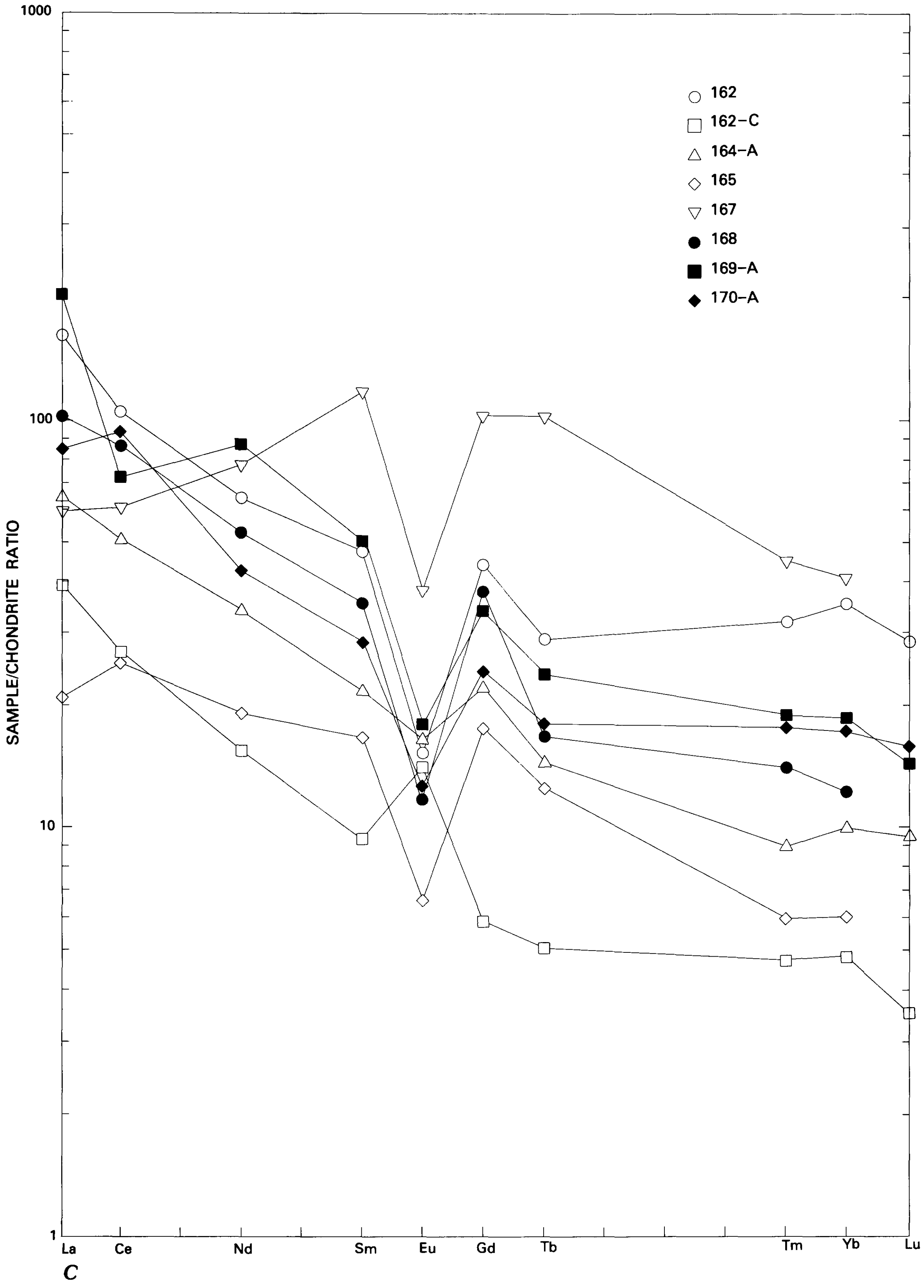


Figure 5. Continued.

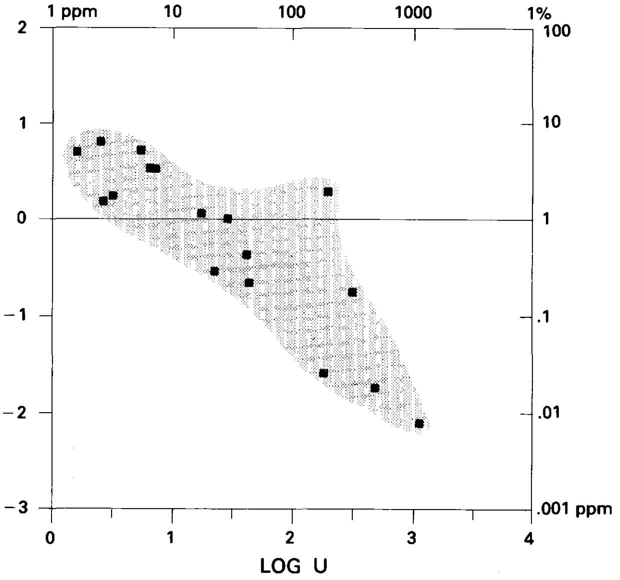


Figure 6. Logarithmic diagram showing the negative correlation between the Th/U ratio and uranium content in rocks of the Mount Holly Complex near Ludlow, Vt.

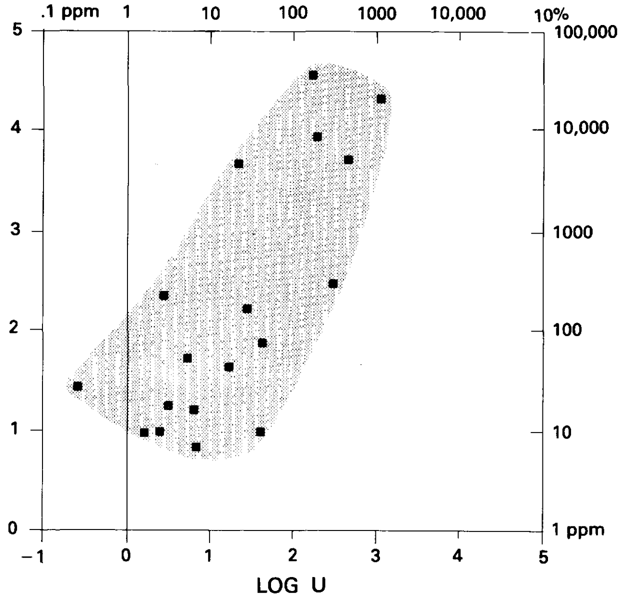


Figure 7. Logarithmic diagram showing boron and uranium contents of rocks from the uranium occurrences near Ludlow, Vt. This diagram shows the direct relation between boron and uranium; this relation agrees with the general modal enrichment in tourmaline near the uranium mineralized zones.

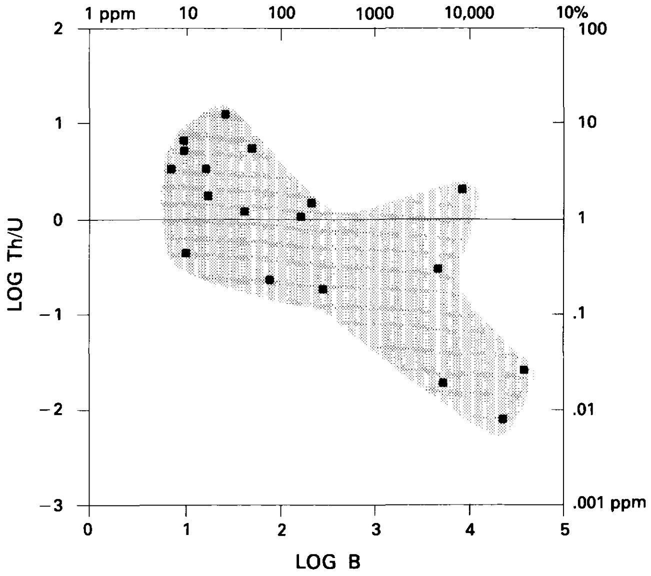


Figure 8. Logarithmic diagram of the Th/U ratio and boron content of rocks from the uranium occurrences near Ludlow, Vt.

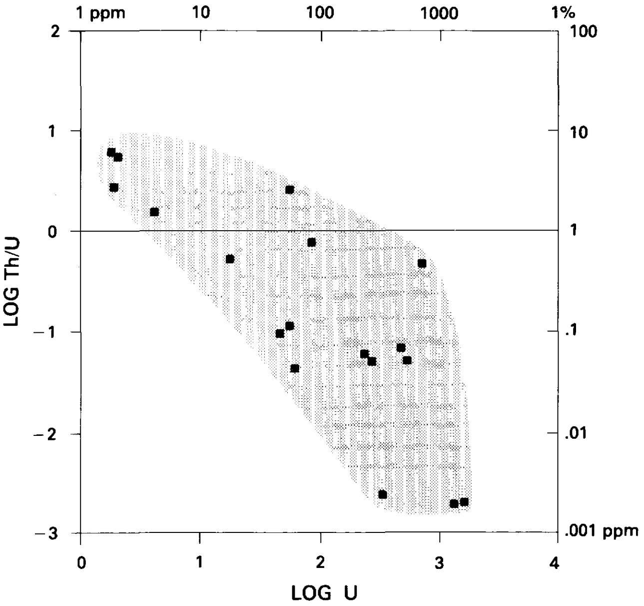


Figure 9. Logarithmic diagram showing the negative correlation between the Th/U ratio and uranium content of the rocks from the uranium occurrences near Jamaica, Vt.

The most significant trace element correlation in the Ludlow area is between uranium and boron in that boron-rich samples have the highest uranium content. Boron ranges up to 3.7 weight percent B_2O_3 in these samples (fig. 7), in agreement with the observed association of mineralized pods and veins with tourmaline-rich rocks. In addition to the general correlation between uranium and boron, Th/U ratios decrease with increasing boron content of the rocks (fig. 8); this correlation between Th/U and B is consistent with the suggestion that uranium and boron gradients may reflect remobilization, transport, and deposition of the uranium oxides during the mineralizing events.

Contents of U_3O_8 range from 0.2 to 0.6 weight percent in the mineralized veins and pods, and anomalous values as high as 22 weight percent U_3O_8 have been recorded in the Jamaica area. Most samples from Jamaica have relatively low thorium contents except for a few samples that have thorium values up to 326 ppm (table 4). Ratios of Th/U vary widely from as low as about 0.001 to 10, and the Th/U ratios decrease with increasing uranium content (fig. 9). Th/U and boron contents near Pinnacle Hill and College Hill are not correlated; this lack of correlation implies that uranium and boron also are not correlated. The lack of correlation also agrees with the observed textural disequilibrium between tourmaline- and magnetite-rich (and uranitite-rich?) mineralized veins. It appears that the mineralizing process responsible for the remobilization of uranium probably did not transport boron, which is in contrast to the situation in the Ludlow area, where the geochemical coherency between uranium and boron may indicate that metamorphic fluids remobilized and transported uranium and boron together.

COMPARISONS WITH THE SWANSON URANIUM DEPOSIT, VIRGINIA

The uranium deposits in Vermont differ in style of mineralization and in their geochemical features from the most important uranium deposit in the Eastern United States, the Swanson deposit near Lynchburg, Va. The Swanson mineralization is structurally controlled and occurs as disseminations, in healed fractures within an augen gneiss unit, and, especially, in cataclastic zones within amphibolite members of this augen gneiss unit (Halladay and others, 1982). At Swanson, the main mineralizing event occurred between 175 and 225 Ma(?). Thus the uranium mineralization here may be significantly younger than that in the Green Mountains. In addition, the mineralizing fluids at Swanson strongly altered the host rocks by sodium metasomatism (albitization), desilication, argillization (hydrogen metasomatism), chloritization, CO_2 metasomatism, and hematization; in the Vermont deposits, however, the effects of alteration are not nearly as intense. Coffinite is the most important uranium mineral at Swanson instead of uraninite, and the coffinite is closely associated with apatite, chlorite, and anatase (Halladay and others, 1982; Halladay, 1983). In contrast, uranium mineralization in Vermont is closely associated with tourmaline and ilmenite(?) at Ludlow and with quartz, garnet, magnetite, and tourmaline near Jamaica. Tourmaline is also a gangue mineral in uranium deposits elsewhere, especially those of the unconformity type in the Athabasca basin, Canada (for example, Rabbit Lake deposit). In contrast to the Ludlow occurrence, however, uranium mineralization at Rabbit Lake contains anomalous abundances of Ni, Co, Cu, Zn, V, Ag, platinum-group elements, As, Bi, and Se (Hoeve and Sibbald, 1978).

Uranium mineralization at Swanson, like that near Ludlow and Jamaica, shows no additional elemental enrichments. At Swanson, the mineralizing fluids were probably high in fluorine, phosphate, and uranium, whereas the fluids at Ludlow were probably high in boron and uranium. The obvious correlation between phosphate and uranium at Swanson agrees with the extensive formation there of high-fluorine apatite, a feature that remains unrecognized at the Vermont deposits (figs. 10 and 11). Fluids associated with the mineralization at Jamaica were probably different from those at Ludlow, however, as suggested by the replacement of tourmaline and by the general abundance of magnetite in the mineralized veins.

DISCUSSION AND CONCLUSIONS

Uranium mineralization in the Ludlow and Jamaica deposits of Vermont is hosted by diverse lithologies within the Middle Proterozoic Mount Holly Complex. Thus, we believe that the composition of the host rocks was not the controlling factor in the localization of the uranium-rich pods and veins. We emphasize that the presence of pegmatitic dikes and veins appears to be a general feature of areas that contain anomalously high contents of uranium. This feature, together with the presence of the strongly sheared and mylonitized rocks of the Mount Holly Complex, may be useful in the search for similar mineralization in other parts of the Proterozoic basement of the Green Mountain massif.

The uranium deposits have not been radiometrically dated, and only broad stratigraphic constraints can be placed on their age. We believe that, because of metamorphism and shearing and fracturing in the host rocks, tectonic events probably produced several stages of uranium mobilization in anatectic pegmatites of the Mount Holly Complex. On the basis of our preliminary results, we suggest that the genesis of the uranium mineralization reflects the interplay of several Late Proterozoic and Paleozoic orogenies. However, because of the complexity of Proterozoic igneous and metamorphic events in the region and because of the intensity of Paleozoic deformation and metamorphism, we cannot preclude the possibility that the mineralization was entirely a result of mobilization of uranium during the Acadian orogeny.

We suggest that, during the peak prograde Grenvillian metamorphic event at about 1100 Ma (Ratcliffe and Aleinikoff, 1990), a first stage of lithophile element concentration occurred through local partial melting of the Mount Holly Complex. This melting produced anatectic pegmatites, or granitic melts that evolved into granitic pegmatites. Our suggestion that pegmatitic segregations generated during these stages of the Grenvillian orogeny are the most significant sources of the uranium mineralization in the Mount Holly Complex agrees with the ideas of

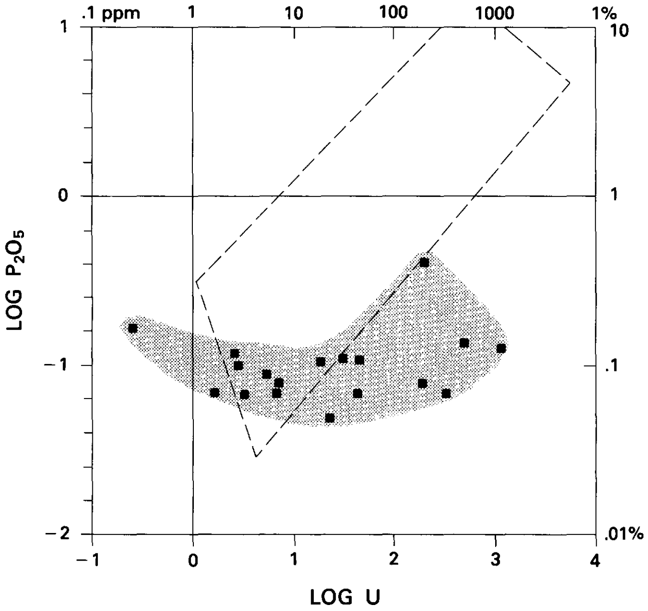


Figure 10. Logarithmic diagram of P_2O_5 and U contents of uranium mineralized rocks from Ludlow, Vt. (shaded area), and from the Swanson uranium deposit, Virginia (area delineated by a dashed line). Note the direct correlation between U and P_2O_5 in the Swanson deposit, which agrees with the extensive formation of apatite in the mineralized zones there. This feature is not observed near Ludlow, where tourmaline is the most important accessory mineral, suggesting that the mineralizing fluids were probably boron rich. Data for the Swanson deposit are from Halladay and others (1982) and from R.A. Ayuso and G.R. Robinson, Jr. (unpub. data).

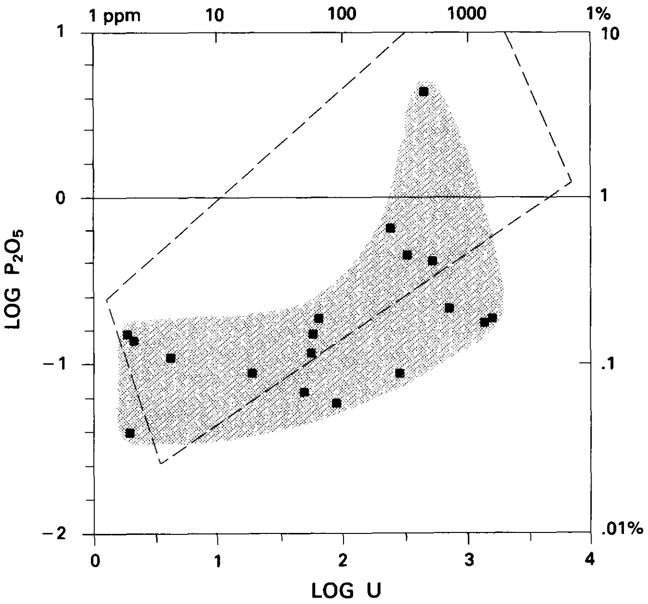


Figure 11. Logarithmic diagram of P_2O_5 and U contents of uranium mineralized rocks from near Jamaica, Vt. (shaded area), and from the Swanson uranium deposit, Virginia (area delineated by a dashed line). Data for Swanson are from Halladay and others (1982) and from R.A. Ayuso and G.R. Robinson, Jr. (unpub. data).

McHone and Wagener (1982) regarding the origin of the uranium-bearing epigenetic veins near Ludlow and Jamaica.

We believe that the relatively low thorium values and the low total REE contents in the mineralized zones, and the lack of direct correlations of these element concentrations with those of uranium, argue against an origin of the mineralization by direct derivation from anatectic melts. It is more likely, in our interpretation, that these anatectic rocks were the precursors to the mineralization and that they represent the first stage of uranium concentration. Subsequent stages necessary for the formation of the uranium mineralization include the westward thrusting of the basement and its lower Paleozoic cover during the Taconian orogeny, which resulted in extensive shearing and fracturing. Studies by Chamberlain and Karabinos (1987) demonstrate that thermal perturbations in thrust-dominated terranes transfer heat from the hotter overriding plate into the cooler lower plate. In the Jamaica area, we believe that Taconian thrusting may have played a critical role and triggered the remobilization of uranium. During a subsequent metamorphic event associated with the Acadian orogeny, fluids released by metamorphic reactions may have migrated along shear surfaces and further remobilized uranium, leading to the formation of the presently exposed mineralized pods and veins.

We prefer to invoke oxidizing metamorphic fluids as the remobilizing agents of uranium and suggest that the mixing of these fluids with more reducing fluids reduced localized uranium mineralization within tourmaline-rich rocks of the Ludlow area. We note, however, that the relation between tourmaline-rich units and uranium mineralization in the Jamaica area is not straightforward and that the style of mineralization there may be quite different compared to that in the Ludlow area. In fact, textural observations of the mineralized rocks near Jamaica suggest that the crystallization of tourmaline probably preceded the introduction of the uranium. In the Jamaica area, tourmaline-rich units locally may have functioned as effective host rocks by promoting mineral reactions and the reduction of U^{+6} to U^{+4} for the precipitation of uranium oxides in the deposits.

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TABLES 1–4

Table 1. Representative major element analyses of rocks from the Mount Holly Complex near Ludlow, Vt.

[All analyses were performed at the U.S. Geological Survey, Reston, Va. Major element values (in weight percent) were determined by rapid-rock methods by F.W. Brown and Herbert Kirschenbaum, analysts. Values for Cl and F (in parts per million) and H_2O were determined by wet chemical methods by Norma Rait and Hezekiah Smith, analysts. The x and y coordinates are Universal Transverse Mercator (UTM) grid, zone 18T. The x coordinate is the easting value in meters; the y coordinate is the northing value in meters. W=Wallingford, L=Ludlow]

Sample	169–A	167	170–A	162	168	162–C	165	164–A	170
SiO_2	57.9	59.5	60.0	68.5	71.8	73.8	74.8	79.0	80.2
Al_2O_3	20.9	18.3	19.6	17.9	14.4	15.1	14.7	12.2	12.6
Fe_2O_3	1.8	3.7	1.3	1.2	3.0	.01	1.3	2.3	.58
FeO	6.6	3.7	6.1	1.7	.36	1.0	.2	.72	.12
MgO	2.0	3.2	2.2	.5	.57	.10	.37	.54	.19
CaO	.25	1.5	1.7	.56	1.3	.80	.31	.06	1.3
Na_2O	.74	.82	2.2	3.3	4.8	3.5	4.3	.38	2.5
K_2O	3.7	5.0	3.1	4.1	1.8	4.8	2.7	2.5	1.8
TiO_2	1.5	1.2	1.0	.11	.41	.06	.12	1.1	.09
P_2O_5	.09	.14	.12	.11	.07	.07	.11	.05	.07
MnO	.10	.11	.10	.12	.07	.05	.06	.04	.02
S	.01	.04	.01	.03	.03	.01	.01	.04	.01
CO_2	.01	.06	.11	.06	.06	.36	.08	.05	.03
H_2O^+	4.5	2.7	2.9	2.2	.96	.89	1.2	1.5	.84
H_2O^-	.09	.14	.04	.08	.03	.01	.03	.02	.02
Total	100.21	100.14	100.51	100.43	99.63	100.61	100.31	100.54	100.41
Cl	35	140	39	21	27	20	21	20	20
F	800	1,800	1,000	800	300	100	400	500	200
Rock type	Pegmatitic, tourmaline-bearing muscovite schist and quartzite	Tourmaline-rich, pegmatitic, micaceous quartzite	Micaceous quartzite and schist; pegmatitic	Sheared, tourmaline-bearing, rusty quartzose gneiss	Pegmatitic, micaceous, tourmaline-bearing rusty quartzose gneiss	Sheared quartzite	Pegmatitic, tourmaline-rich quartzose gneiss	Tourmaline-rich micaceous quartzite pod	Micaceous quartzite
Quadrangle	W	L	W	L	W	L	W	L	W
X coordinate	48017	48083	48018	48109	48022	48109	48078	48091	48018
Y coordinate	6808	6840	6794	6840	4817	6840	6829	6822	6794

Table 2. Representative trace element analyses of rocks from the Mount Holly Complex near Ludlow, Vt.

[All values are in parts per million. Rb, Sr, Ba, Y, Zr, and Nb were determined by Energy Dispersive X-Ray Fluorescence by R.G. Johnson and J. Evans, analysts. Values for rare earth and ferromagnesian elements were determined by neutron activation methods by C.A. Palmer and G.A. Wandless, analysts. Li, Mo, Sn, W, and Zn were determined by wet chemistry methods by A.F. Dorrzapf, P.J. Aruscavage, and W.M. d'Angelo, analysts. Be was determined by induction-coupled-plasma spectroscopy, and B was determined by quantitative spectroscopy by D.W. Golightly, J.S. Kane, and J.D. Fletcher, analysts. U and Th were determined by delayed neutron activation methods by C.A. Palmer, analyst]

Sample	169–A	167	170–A	162	168	162–C	165	164–A	170
Rb	191	400	141	202	128	101	123	125	58
Cs	8	24	4	7	2	2	1	3	1
Sr	65	78	175	89	83	152	48	28	181
Ba	708	1,185	456	722	167	706	273	412	173
Y	39	47	27	46	—	12	13	—	22
La	68	20	28	53	34	13	—	22	—
Ce	69	54	82	93	77	24	24	44	—
Nd	53	47	26	39	32	10	9	—	—
Sm	9	22	5	9	7	2	3	4	—
Eu	1.2	2.7	.8	1.1	.8	1	.5	.7	—
Gd	8.3	26	5.9	11	9	1.4	4.4	5.5	—
Tb	1.1	4.9	.8	1.4	.8	.2	.6	.7	—
Tm	.6	1.3	.5	1.0	.4	.1	.2	.3	—
Yb	3.7	8	3.7	7	2.5	1.0	1.2	2.0	—
Lu	.5	.1	.6	1.0	—	.1	—	.3	—
Zr	298	113	209	115	38	42	57	157	42
Hf	10	5	6	4	2	1	2	4	1
Nb	23	13	12	12	13	3	10	11	5
Ta	1.8	1.7	.9	.9	1.9	.1	1.1	1.9	.4
Th	29	9	17	31	60	6	10	7	19
U	5	471	3	29	316	3	43	22	41
Sc	24	21	23	9	4	1	4	7	1
Cr	120	75	89	20	7	—	3	55	2
Co	16	18	17	4	2	1	4	2	1
Ni	44	47	42	7	5	10	5	12	12
Cu	—	—	—	—	—	—	—	—	—
Zn	120	108	—	35	—	14	32	—	—
Li	57	130	50	28	23	9	31	24	11
Be	3	7	2	5	3	1	4	3	2
B	53	5,300	10	170	290	18	79	4,800	10
Sn	5	9	2	3	1	2	3	10	1
W	2	4	1	2	1	1	2	6	1
Mo	.3	1.3	1.1	.1	21	.7	.2	.7	8
Sb	.4	.7	—	.2	—	—	.7	.2	—

Table 3. Partial chemical analyses of representative rocks from the Mount Holly Complex and of magnetite-rich

[All analyses were performed at the U.S. Geological Survey, Reston, Va. Major element values (in weight percent) were determined by rapid-rock methods determined by wet chemical methods by Norma Rait and Hezekiah Smith, analysts. The x and y coordinates are Universal Transverse Mercator (UTM)

Sample	J–16	J–9	J–6	J–8	J–1	J–12	J–10	J–11
SiO_2	24.52	46.0	59.52	60.21	63.15	64.0	64.62	65.14
Al_2O_3	6.27	15.97	16.9	14.71	19.75	15.7	16.03	16.58
Fe_2O_3	60.67	15.02	6.08	8.54	4.96	5.27	5.43	4.34
MgO	1.05	4.89	2.13	2.48	1.38	2.44	1.49	1.87
CaO	1.09	10.3	5.46	5.33	.6	4.3	3.94	3.40
Na_2O	2.21	2.76	4.46	2.16	1.06	5.08	5.31	5.07
K_2O	.58	1.44	1.87	2.82	5.28	1.13	1.19	.83
TiO_2	1.81	1.41	1.04	1.29	.7	.51	.52	.46
P_2O_5	.22	.45	.42	.15	.09	.19	.15	.19
MnO	.35	.29	.13	.14	.04	.11	.11	.06
S	.01	.03	.01	.09	.11	.01	.01	.04
CO_2	.01	.01	.01	.01	.01	.01	.01	.01
H_2O^+	.53	1.6	1.4	1.5	2.3	.72	.82	.67
H_2O^-	.23	.31	.35	.28	.37	.10	.12	.31
Total	99.55	100.46	99.78	99.71	99.80	99.57	99.75	98.9
Cl	60	250	80	110	50	50	110	40
F	290	810	680	490	550	510	290	380
Rock type	Magnetite-rich, rusty mineralized pods in pegmatite quartzofeldspathic gneiss	Rusty mafic schist in pegmatitic micaceous gneiss	Biotite schist pod in pegmatitic gneiss	Biotite schist pod in pegmatitic gneiss	Red and yellow-stained schistose lenses in biotite schist	Granitic gneiss containing micaceous mineralized pods	Pegmatitic gneiss containing rusty pods	Biotite schist pod in migmatitic gneiss
Area	College Hill	College Hill	College Hill	College Hill	Pinnacle Hill	Pinnacle Hill	College Hill	Pinnacle Hill
X coordinate	47748	47757	47754	47756	47763	47758	47757	47758
Y coordinate	6775	6770	6774	6773	6745	6757	6770	6757

mineralized lenses near Jamaica, Vt.

by F.W. Brown and Herbert Kirschenbaum, analysts. Values for Cl and F (in parts per million) and H_2O were grid, zone 18T. The x coordinate is the easting value in meters; the y coordinate is the northing value in meters]

J–15	J–14	J–2	J–5	J–4	J–7	J–17
65.33	66.0	69.40	69.54	72.92	73.03	75.3
14.97	14.7	13.37	14.9	15.86	16.12	12.5
8.12	6.53	6.17	3.84	.24	1.32	2.5
.63	1.5	1.72	1.1	.11	.39	.81
2.15	2.25	2.83	2.73	.65	1.84	1.74
6.9	5.2	2.68	1.11	5.48	4.54	5.5
.55	1.1	1.44	3.97	4.16	1.62	.79
.46	.48	.81	.33	.03	.08	.2
.11	.18	.14	.12	.04	.06	.09
.07	.1	.13	.06	.02	.02	.07
.01	.91	.01	.01	.01	.01	.01
.01	.01	.01	.01	.01	.01	.01
.45	.53	1.3	1.6	.28	.68	.52
.14	.28	.13	.21	.02	.22	.09
99.90	99.77	100.14	99.53	99.83	99.94	99.93
40	40	60	40	40	40	60
250	570	530	310	50	90	280
Granitic gneiss enclosing biotite schist pods	Biotite schist pod	Feldspathic and biotitic schist containing mineralized pods	Pegmatitic gneiss, migmatitic	Pegmatitic gneiss, migmatitic	Pegmatitic gneiss containing mineralized pods	Pegmatitic gneiss, migmatitic
Pinnacle Hill	Pinnacle Hill	Pinnacle Hill	College Hill	College Hill	College Hill	College Hill
47756	47756	47760	47751	47751	47754	47748
6758	6758	6748	6776	6776	6774	6775

Table 4. Representative trace element analyses of rocks from the Mount Holly Complex and of magnetite-rich mineralized lenses near Jamaica, Vt.

[All values are in parts per million. Rb, Sr, Ba, Y, Zr, and Nb were determined by Energy Dispersive X-Ray Fluorescence by R.G. Johnson and J. Evans, analysts. Values for rare earth and ferromagnesian elements were determined by neutron activation methods by C.A. Palmer and G.A. Wandless, analysts. Li, Mo, Sn, W, and Zn were determined by wet chemistry methods by A.F. Dorrzapf, P.J. Aruscavage, and W.M. d'Angelo, analysts. Be was determined by induction-coupled-plasma spectroscopy, and B was determined by quantitative spectroscopy by D.W. Golightly, J.S. Kane, and J.D. Fletcher, analysts. U and Th were determined by delayed neutron activation methods by C.A. Palmer, analyst]

Sample	J–16	J–9	J–8	J–14	J–6	J–1	J–11	J–12	J–10	J–15	J–7	J–5	J–2	J–17	J–4
Rb	355	139	121	497	241	274	515	51	34	19	65	124	86	42	56
Cs	2	2	3	2	3	4	1	2	1	—	—	1	4	1	—
Sr	41	271	320	199	344	45	268	335	309	324	134	219	213	170	137
Ba	81	179	312	418	259	416	251	238	305	275	335	593	351	110	314
Y	2,310	28	387	—	69	7	—	16	17	12	162	45	29	35	31
Pb	120	37	36	140	66	200	190	26	3	12	30	25	27	14	18
La	35	20	142	44	64	21	47	25	27	44	55	52	15	39	7
Ce	55	48	352	65	129	30	59	56	58	84	109	115	38	98	18
Nd	54	41	172	—	84	20	260	27	25	41	68	56	21	35	11
Sm	34	8	54	20	22	5	24	5	5	8	21	13	4	9	3
Eu	2	2.2	1.9	1.1	1.6	.2	1.8	1.2	1.1	1.3	.6	2.1	.8	.6	.4
Gd	29	—	70	—	23	—	6.3	4.9	3.7	—	24	11	3.9	7.8	2.8
Tb	37	1.1	11.4	.5	3.9	.3	1	.7	.6	—	5	1.6	.7	1.3	.8
Tm	—	.6	7.8	.3	2.4	—	.5	.6	—	—	4	1.2	.7	.6	.6
Yb	273	4.0	52	1.5	15.1	.9	2.8	2.2	1.4	1	28	6.3	3.8	5.4	3.9
Lu	39	.8	7.3	—	2.2	—	—	.5	.2	.1	4.1	1	.6	.7	.6
Zr	944	50	356	37	52	177	16	107	161	173	98	227	249	261	44
Hf	43	2.3	11.6	4.2	2.9	7.7	3.4	3.1	4	5	5	6	8	8	1
Nb	35	8	44	27	15	11	25	11	7	7	10	15	8	15	—
Ta	1	.2	1.5	1.3	.7	1.1	.5	.8	.5	.1	.2	.9	.7	.7	.1
Th	326	.8	142	3	29	15	3	3	5	6	63	6	11	9	11
U	690	323	55	1,340	527	272	1,550	61	2	4	84	54	2	18	2
Sc	15	36	29	13	21	16	12	19	12	11	2	16	14	6	—
Cr	68	239	29	—	—	71	—	28	9	14	—	34	44	5	4
Co	19	40	16	10	12	2	10	12	10	3	1	5	11	4	—
Ni	30	135	9	4	6	4	11	9	5	8	2	15	11	6	2
Cu	10	220	50	56	6	28	110	50	28	30	3	4	12	9	2
Zn	223	135	114	74	65	70	37	69	49	30	13	43	89	61	5
Au	<9	<3	<5	<8	<2	<5	<8	<5	<4	<4	<4	<3	<4	<5	<4
Li	11	27	34	21	27	20	15	18	16	10	10	26	17	9	2
Be	1.4	5.4	1.1	2.6	1.5	1.4	1.2	1.7	1	2	1	2	2	2.1	1.1
B	—	5	4	17	3	16	4	2	2	3	6	8	4	2	5
Sn	1.5	3	5	4	2	<1	<1	1	1	<1	<1	1	1	5	<1

Chapter Q

Geochemistry of Stream Sediments and Heavy-Mineral Concentrates from the Orange County Copper District, East-Central Vermont

By JOHN F. SLACK, PAUL J. ATELSEK, and
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U.S. GEOLOGICAL SURVEY BULLETIN 1887

SUMMARY RESULTS OF THE GLENS FALLS CUSMAP PROJECT, NEW YORK, VERMONT, AND NEW HAMPSHIRE

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Geochemistry of Stream Sediments and Heavy-Mineral Concentrates from the Orange County Copper District, East-Central Vermont

By John F. Slack,[1] Paul J. Atelsek,[1] *and* Jesse W. Whitlow[1]

Abstract

The Orange County copper district, in east-central Vermont, was sampled geochemically by using stream sediments and heavy-mineral concentrates. A total of 152 samples from 76 sites was collected and analyzed by emission spectrographic and atomic absorption methods. This report highlights the distribution of anomalous concentrations of copper, zinc, manganese, and gold in the vicinity of mines and prospects of the district and briefly discusses the occurrence of anomalous cobalt, silver, boron, tin, tungsten, lead, and cadmium. The mined ore deposits consist of pyrrhotite-rich, stratabound massive sulfides within Paleozoic (Silurian and Early Devonian) clastic metasedimentary and minor mafic metavolcanic rocks. Geochemical dispersion of copper, zinc, cadmium, cobalt, silver, manganese, and boron downstream from mine dumps at the sites of major past producers at the Elizabeth, Ely, and Pike Hill mines reflects the mineralogy of the ores and their metamorphic wall rocks. Other anomalies including those for gold, tungsten, tin, and lead constitute a different metal suite that generally is spatially unrelated to the mines or to any known prospects and suggest a different bedrock source. This latter metal suite may be derived from previously unrecognized granite-related mineral deposits, located either in the copper district itself or near postkinematic granite plutons to the north.

INTRODUCTION

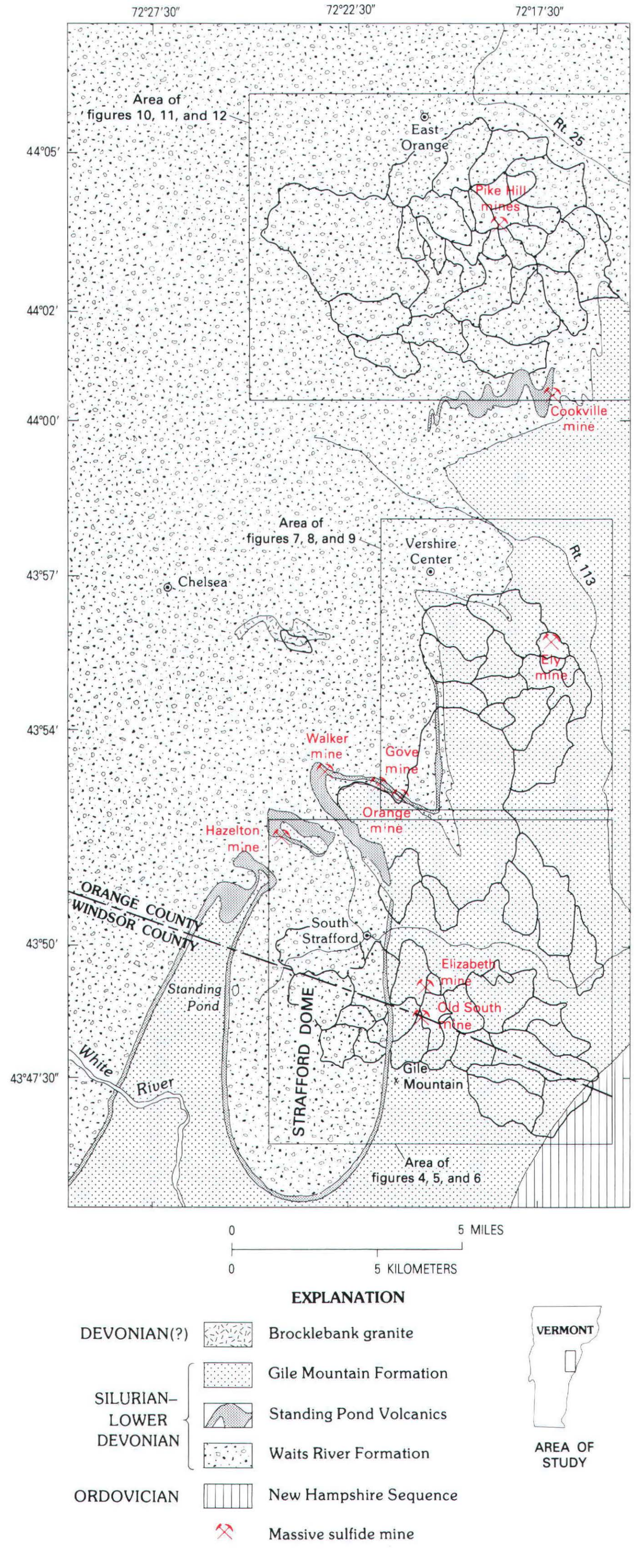
The Orange County copper district, located in east-central Vermont (fig. 1), has a long history of mineral production from cupriferous massive sulfide deposits (White and Eric, 1944). The district has three principal mines, Elizabeth, Ely, and Pike Hill, and many prospects. The Elizabeth deposit was discovered first, in 1793, but early mining attempts for iron failed. Beginning in 1809, the mine produced copperas (iron sulfate), which was used as a dye, in treating timber, in the purification of sewers, and in the manufacture of ink. Copper production in the district started in 1830 with open-pit mining at Elizabeth. During the 1850's, the Ely mine began operation and for a short time reportedly was the largest copper producer in the Nation (Stevens, 1911, p. 1772; Hermance and others, 1949). Production in the district peaked with the reopening of the Elizabeth mine during World War II and ended with its closure in 1958 (see Howard, 1969). Since then there has been no metal mining in the region.

The Elizabeth mine, at the southern end of the district, produced about 3.2 million tons of pyrrhotite ore at average grades of 1.8 percent copper and 0.5 percent zinc, plus minor amounts of silver and gold that were recovered from the copper concentrates (McKinstry and Mikkola, 1954; Howard, 1969; Annis and others, 1983). The centrally located Ely mine (also known as the Copperfield or Vershire mine) produced approximately 0.5 million tons of sulfide ore at comparable copper grades (White and Eric, 1944; Hermance and others, 1949; Kinkel, 1967). The northernmost mines in the district are the closely spaced Union and Eureka properties, about 8 km north of the northern boundary of the Glens Falls 1° × 2° quadrangle. These mines, collectively called the Pike Hill mines, produced less than 0.1 million tons of ore (White and Eric, 1944). The Smith mine, located just to the south of Pike Hill, has no known production. Other mines in the district that have little or no recorded production are Hazelton, Walker, and Orange and Gove north of the town of South Strafford and the Cookville mine between Ely and Pike Hill (fig. 1). Total production in the district is estimated at 4 million tons of massive sulfide ore (Gair and Slack, 1979, 1980).

Chemical analyses of sulfide-rich samples from the mines provide an important foundation for interpreting the drainage geochemical data. Spectrographic and atomic absorption analyses of 15 samples of massive sulfide ore from the Elizabeth, Gove, Ely, and Pike Hill mines show high concentrations of a number of trace elements (Slack and others, 1986; J.F. Slack, unpub. data). In addition to Cu and Zn, the sulfide ores contain elevated values of Ag (to 200 parts per million (ppm)), As (to 200 ppm), Bi (to 44 ppm), Cd (to 1,300 ppm), Co (to 650 ppm), Hg (to 4.9 ppm), Mn (to 8,600 ppm), Mo (to 170 ppm), Ni (to 130

Manuscript approved for publication April 13, 1990.

[1] U.S. Geological Survey.



◀ **Figure 1.** Orange County copper district showing the geology and locations of study areas surrounding the Pike Hill, Ely, and Elizabeth mines. Geology after Doll and others (1961). Waits River and Gile Mountain Formations and Standing Pond Volcanics compose the Vermont Sequence.

ppm), Se (to 160 ppm), and U (to 5.8 ppm). Values for gold are consistently low (<1 ppm). Values for lead are also low (typically <50 ppm); the highest lead concentrations are at the Elizabeth mine (390 ppm). Neither tungsten (detection limit 10 ppm) nor tin (detection limit 5 ppm) was found in any of the analyzed sulfide-rich samples from the mines or from any of the enclosing wall rocks of the deposits.

The geology of the study area is dominated by a thick sequence of highly deformed metasedimentary rocks and minor mafic metavolcanic rocks of Silurian and Early Devonian age (Doll and others, 1961; Hatch, 1988; Hueber and others, 1990). The eastern part of the area is underlain by quartzose mica schist, minor metagraywacke and graphitic mica schist, and sparse quartzite and amphibolite of the Gile Mountain Formation. Lithologies in the western and northern parts of the study area constitute the underlying Waits River Formation that consists of calcareous mica schist and minor calcareous (or ankeritic) metagraywacke and marble. The boundary between these quartzose and calcareous sequences is approximated by the Standing Pond Volcanics, an amphibolite-rich unit that trends roughly north-northeast in the district. This thin unit outlines the Strafford dome and other structural domes to the south; in places it diverges from the contact and is entirely within the Waits River Formation (Doll, 1944[?]; White and Jahns, 1950; Murthy, 1957). In addition to amphibolite, the Standing Pond Volcanics contain minor amounts of metachert, coticule (fine-grained, quartz-spessartine rock), and local epidosite and magnetite iron formation (Annis and others, 1983; Hepburn and others, 1984; J.F. Slack, unpub. field observations). In the Strafford 15-minute quadrangle, the amphibolites are interpreted, on the basis of field relations, mineralogy, and chemistry (Hepburn, 1984; J.F. Slack, unpub. data), as metamorphosed tholeiitic basalts emplaced originally as flows, tuffs, and (or) sills; a small body of metadiabase and fine-grained metagabbro in the Cookville mine area (White and Eric, 1944; J.F. Slack, unpub. field observations) may represent an intrusive feeder to the nearby amphibolites. The surrounding Gile Mountain and Waits River Formations are believed to represent metamorphosed quartz-rich and carbonate-rich turbidites, respectively (Hatch, 1988). All of these rocks have been complexly deformed and regionally metamorphosed to the middle amphibolite facies. Kyanite is common in pelitic schists, and sillimanite is reported from a few localities (see Jacobs, 1944[?]; Annis and others, 1983). The structurally complex setting of the deposits involves two periods of intense folding accompanied by thrust faulting, later doming, and an even later period of local thrust faulting (Offield and Slack, 1990).

The copper deposits occur as stratabound and typically stratiform massive sulfide bodies. Pyrrhotite, the principal sulfide mineral, is associated with minor amounts of chalcopyrite and sphalerite; pyrite is uncommon, and galena is rare. The orebodies apparently are not restricted to any particular lithology or stratigraphic horizon (Annis and others, 1983). The Elizabeth and Ely deposits, for example, are located entirely within the Gile Mountain Formation, whereas the Pike Hill deposits are apparently in the Waits River Formation. The Hazelton, Walker, Orange, Gove, and Cookville deposits are hosted by the Standing Pond Volcanics (fig. 1). Wall rocks associated with the deposits include a variety of unusual lithologies that contain abundant quartz, amphibole, spessartine (manganese-rich) garnet, tourmaline, carbonate, albite, and (or) mica. These lithologies are well developed at the Elizabeth mine, as evidenced by exposures in open cuts and in drill cores (Annis and others, 1983; J.F. Slack, unpub. data) and are believed to represent premetamorphic, hydrothermally altered rocks that formed on or near the sea floor during sulfide mineralization. Limited drill cores available from the Pike Hill mine area show similar unusual rock types, including coticule and magnetite iron formation (J.F. Slack, unpub. data). The Ely massive sulfide deposits have not been studied in enough detail to determine whether they have analogous types of wall rocks.

A geochemical sampling program carried out during 1981–82 focused on drainage basins surrounding the three largest mines of the district. Most of the samples were collected in 1981 and 1982 by J.W. Whitlow, assisted by C.N. Sears, G.L. Cygan, and J.T. Hanley, all of the U.S. Geological Survey. Additional field work was done by P.J. Atelsek in 1982, including in the Pike Hill area where sites that had anomalous metal values were resampled. Selected data that have the most important economic and exploration implications are presented here, including those for Cu, Zn, Co, Au, Sn, W, B, and Mn. The complete geochemical data set of analyses for 31 elements in 301 samples from 82 sites (grab samples, magnetic and nonmagnetic panned concentrates, and stream sediments) is available in Adrian and others (1989).

SAMPLING AND ANALYTICAL METHODS

Sample localities were selected to include drainage basins that contain all of the important mineral deposits in the district. At each site, stream sediment and a panned concentrate were taken. The bulk stream sediment samples consist of the finest loose material available that was dried and sieved to −80 mesh prior to analysis.

The panned concentrates comprise materials that were generally taken from several locations at each sample site. These different locations typically were only a few meters or tens of meters apart, and all were in the same stream course. The concentrates, collected with a shovel, were sieved to −4 mesh (<4.75 mm), and the lighter material was removed by use of a gold pan until a heavy-mineral fraction remained. This fraction was then dried and sieved to −10 mesh (<2.00 mm) and, in the laboratory, was separated by using bromoform (specific gravity=2.85) to further remove the light minerals. The remainder, culled of magnetite by using a hand magnet, was then separated into relatively magnetic and nonmagnetic fractions by use of a Frantz Isodynamic Magnetic Mineral Separator set at 0.6 ampere to reduce interference effects caused by iron. Both of these fractions were chemically analyzed. Splits of the nonmagnetic fractions also were examined for scheelite under an ultraviolet light.

Analyses were made by B.F. Adrian, B.F. Arbogast, R.N. Baker, and D.E. Detra, in U.S. Geological Survey laboratories at Denver, Colo., by using a semiquantitative 6-step emission spectrographic method for 31 elements (Myers and others, 1961; Motooka and Grimes, 1976). The semiquantitative values are reported as 6 steps per order of magnitude (1, 0.7, 0.5, 0.3, 0.2, 0.15, or multiples of ten of these numbers) and are approximate midpoints of geometric brackets whose boundaries are 1.2, 0.83, 0.56, 0.38, 0.26, 0.18, 0.12, and so on. The lower limits of determination in ppm for the elements studied are B, 10; Ba, 20; Co, 5; Cu, 5; Mn, 10; Mo, 5; Ni, 5; Pb, 10; Sn, 10; W, 50; Zn, 200; these limits are doubled for the heavy-mineral concentrates. Gold also was analyzed by quantitative atomic absorption methods by J.E. Gray, A.A. Roemer, and L.J. Sherlock in U.S. Geological Survey laboratories in Denver. The atomic absorption determination limit for gold, which is proportional to the sample size, in this study ranged from 0.05 to 0.71 ppm. The high detection limits were necessary only for a few concentrate samples of minimal size.

DATA PRESENTATION

For ease of examination and delineation of regional background variations, the samples are divided into three groups on the basis of their locations within the major mine areas of the district (Elizabeth, Ely, Pike Hill). The Elizabeth mine area contains 31 sample sites, Ely has 14, and the Pike Hill mine area contains 31 sample sites. Excluding grab samples and magnetic fractions of heavy-mineral concentrates (Adrian and others, 1989), data for 76 sample sites (152 total samples) are reported here.

Histograms

Histograms were constructed to determine anomalous element concentrations relative to those of the regional (background) sample population. Visual inspection of the histograms was used to characterize anomalous samples; samples that have analytical values greater than those of a statistically normal population are considered to be anomalous. A comparison of the data also was made with geochemical analyses of stream sediments and heavy-mineral concentrates from unmineralized clastic metasedimentary sequences elsewhere in the Appalachians (for example, Lesure and others, 1977; Slack and others, 1984a; Slack and others, 1985). One histogram for each anomalous element in each sample type was constructed (figs. 2 and 3); the class intervals represent semiquantitative concentration ranges for the analyzed samples. Quantitative atomic absorption results for gold were grouped into the six-step format to avoid arbitrary interval assignment and to provide semilogarithmic compression of the data.

Anomaly Maps

Identified anomalous samples are portrayed on anomaly maps for each of the three study areas (see figs. 4–12). The anomaly maps are designed to provide information about the areal distribution of the geochemical anomalies relative to the distribution of mines and prospects. Not all of the selected elements show significant anomalous concentrations, however, and therefore only drainage basins that have anomalous values of Au, Co, Cu, Mn, Sn,[2] W,[2] and Zn are plotted. Each of these elements is assigned a different pattern on the maps, except for gold, which is shown by a pink pattern. Assuming local derivation of sediment debris, the source of an anomaly may be anywhere within the drainage basin that contains the sample site. Therefore, the entire basin is shown with the pattern for the anomalous element, even when the location of a probable source (such as an ore deposit) is known. Possible complications of glacial contribution to the anomalies are discussed in more detail below.

RESULTS AND DISCUSSION

Elizabeth Mine Area

Geochemical analyses of heavy-mineral concentrates show a clear derivation of metals from the Elizabeth mine (fig. 4). Panned concentrate 81–1, which was collected from a stream that drains the area around the north pit of the mine, has values of >50,000 ppm copper, 2,000 ppm zinc, 30 ppm silver, and >5,000 ppm boron (table 1). These values are clearly due to drainage from the mine dumps and reflect the mineralogy of the sulfide orebodies that includes chalcopyrite, minor sphalerite, and locally abundant tourmaline (the source of the anomalous boron). In contrast, panned concentrate 81–3, collected from streams that drain the south pit and the Old South mine, lacks base-metal (and boron) anomalies, probably because large dumps are absent in this area. This sample does have a high manganese content, however, as does one sample from a basin to the southeast (81–5). The manganese may reflect the occurrence of coticule rocks in the Elizabeth mine sequence (including the Old South mine), particularly their relatively widespread distribution compared to the stratigraphically restricted sulfide ores (Annis and others, 1983; J.F. Slack, unpub. data; see also Spry, 1990).

The larger drainage basin immediately to the west of the Elizabeth mine contains 1,500 ppm copper in heavy-mineral concentrate 81–16 (fig. 4). No mine dumps or prospects are known in this basin, and the contained anomaly may be due to concealed mineralization, possibly an outcropping of the western limb of the Elizabeth orebody (see Howard, 1969). Alternatively, the presence of this copper may reflect stratabound sulfide deposits associated with amphibolites of the Standing Pond Volcanics and (or) the stratigraphically lower part of the Gile Mountain Formation (see Doll, 1944[?]; White and Eric, 1944; Annis and others, 1983).

Many geochemical anomalies in this southern area lack any spatial association with mines or prospects. Several concentrate samples that contain 1,000 to 1,500 ppm zinc (nos. 81–2, 81–17, 81–25, 81–33, 81–35, and 81–37) are clearly unrelated to the open pits or dumps of the Elizabeth mine, although basin 81–17 is nearly on strike with the orebody to the south. The anomaly in concentrate sample 81–2 (1,500 ppm zinc) probably has a source close to the sample site because the three subbasins slightly upstream show little or no zinc (for example, below analytical detection limits of 500 ppm). Heavy-mineral concentrate 81–19 has a marginal (100 ppm) cobalt anomaly and 2,000 ppm boron.

Basins that contain ≥1,000 ppm zinc in the eastern part of the Elizabeth mine area (concentrates 81–25, 81–33, and 81–35) form a roughly north-south belt (fig.4). Sample 81–31 has anomalous manganese (10,000 ppm). Heavy-mineral concentrate sample 81–26 in this area contains 300 ppm lead (Adrian and others, 1989).

Two basins to the west and southwest of the Elizabeth mine contain very high levels of manganese (>10,000 ppm) in heavy-mineral concentrates. The anomalous samples from these basins (81–11, 81–12) are from the center of the Strafford dome in which the major lithologies are calcareous schists and quartzose marbles of the Waits River Formation. No prospects are known in these two basins,

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[2] In some samples, analytical values for Sn and W are reported as "<20" ppm and "<100" ppm, respectively, which represent metal concentrations close to, but less than, the analytical detection limits for each element. On the anomaly maps, these values are considered close enough to the threshold values to be included in the population of anomalous samples.

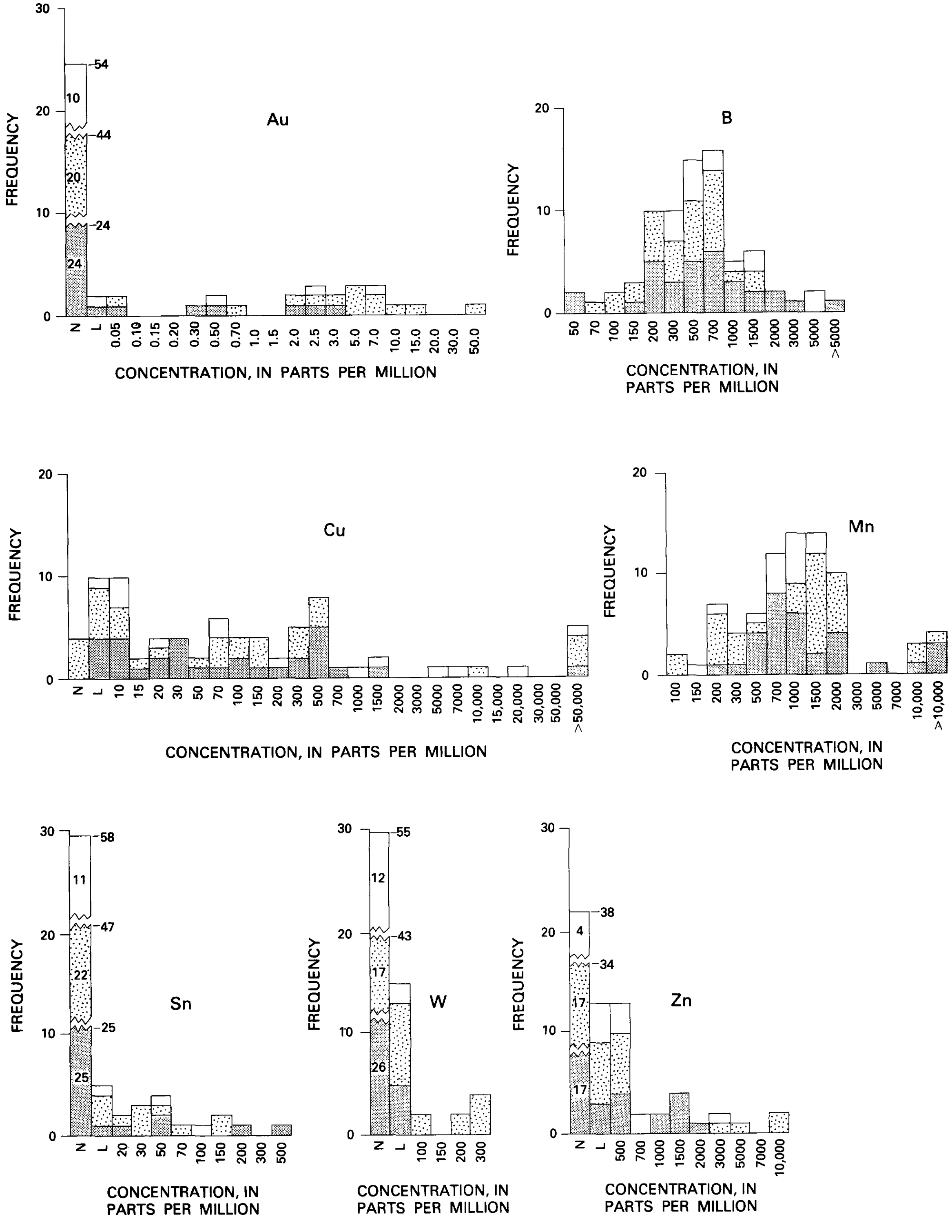

Figure 2. Histograms of selected geochemical data for samples of heavy-mineral concentrates. Shaded (gray) boxes represent data from the Elizabeth mine area, open boxes are from the Ely mine area, and stippled boxes represent data from the Pike Hill mine area. Abbreviations used: N, not detected; L, detected but below limit of determination.

and the Standing Pond Volcanics (a possible source of Mn-garnet) are apparently absent in both drainages. These basins lack anomalous amounts of other metals in heavy-mineral concentrates (except minor gold in 81–12) and seem unrelated to the sulfide deposits of the district. The origin of the high manganese values in basins 81–11 and 81–12 is unknown.

Anomalous gold, tungsten, and tin were detected in several samples from basins in the vicinity of the Elizabeth mine (fig. 5). Gold was found at or below the detection

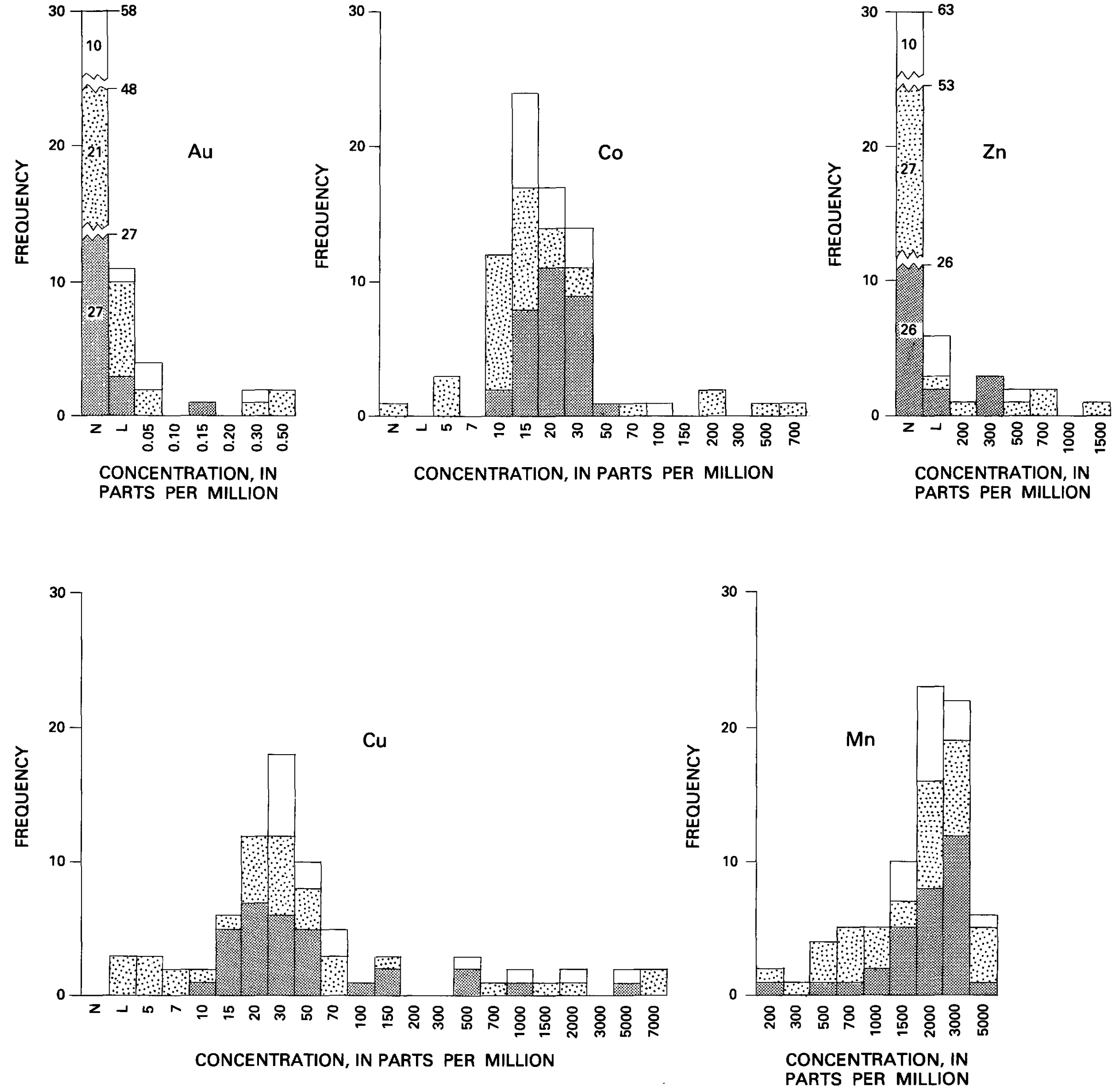

Figure 3. Histograms of selected geochemical data for stream sediment samples. Shaded (gray) boxes represent data from the Elizabeth mine area, open boxes are from the Ely mine area, and stippled boxes represent data from the Pike Hill mine area. Abbreviations used: N, not detected; L, detected but below limit of determination.

limit (0.05–0.71 ppm) in concentrate samples from several basins to the west of the Elizabeth mine area (concentrates 81–12, 81–14, 81–15, 81–16). Within this belt, only sample 81–14, which contains 1.9 ppm gold, has a highly anomalous gold content; no gold was found in basins that drain the open pits or mine dumps. Tin and tungsten anomalies also are spatially unrelated to the mine area, except for one value of detectable (<100 ppm), but not determinable, tungsten in a concentrate sample from basin 81–4 that drains the Old South mine (fig. 5). Similar tungsten values are present in samples from surrounding basins (for example, concentrates 81–10, 81–14, 81–19, 81–22). Tin also occurs in some of these same basins (table 1) although at low levels (≤50 ppm). High values of tin are present in concentrate samples 81–6 (200 ppm) and 81–21 (500 ppm), neither of which is from basins that drain the mine or dumps at Elizabeth.

Geochemical anomalies for stream sediments (table 2) only locally agree with those for heavy-mineral concentrates. The drainage basins that show the most similarity in metal concentrations between these two sample media surround the open pits and dumps of the Elizabeth mine (fig. 6). Sample 81–1 that drains the eastern side of the north pit and the main dumps contains 5,000 ppm copper,

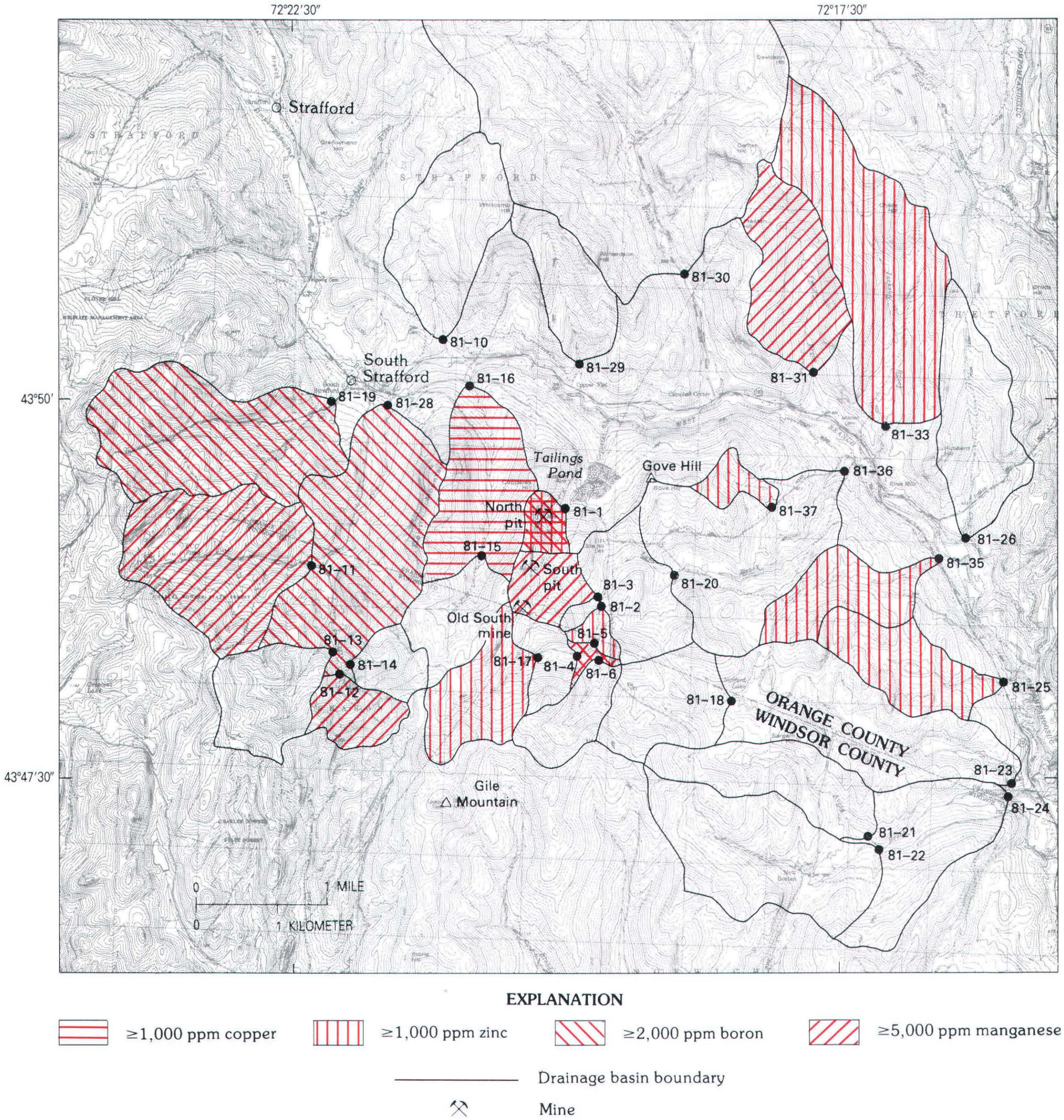

Figure 4. Distribution of geochemically anomalous copper, zinc, boron, and manganese in samples of heavy-mineral concentrates from the Elizabeth mine area. See table 1 for anomalous values. Note that for some drainage basins more than one pattern is shown to indicate the presence of more than one anomalous element value. Base from U.S. Geological Survey, Sharon and South Strafford, 1981. Contour interval 20 feet. National Geodetic Vertical Datum of 1929.

300 ppm zinc, 1.5 ppm silver, 30 ppm molybdenum, and 700 ppm boron (table 2; additional data from D.E. Detra, U.S. Geological Survey, written commun., 1989). The boron value of 700 ppm, although somewhat low for heavy-mineral concentrates, is very high for stream sediments and clearly reflects the occurrence of abundant tourmaline in the sulfide ores. Sample 81–3 also contains high amounts of copper (1,000 ppm), which undoubtedly are derived from mineralized rocks in the south pit and the Old South mine (fig. 6). Along strike to the south, sample 81–6 has anomalous (5,000 ppm) manganese that may reflect clay adsorption of manganese derived from coticules associated with the distal portions of the Elizabeth deposit. Stream sediment sample 81–33 to the east of the mine contains 500 ppm copper (fig. 6); this value is consistent with the copper data for the heavy-mineral concentrate from

Table 1. Selected analyses of heavy-mineral concentrates from the Elizabeth mine area, Vermont

[All values are in parts per million. Analytical detection limits for the elements are as follows: Au, 0.05; B, 20; Cu, 10; Mn, 20; Sn, 20; W, 100; Zn, 500. N=not detected; >=greater than value shown; <=detected but less than value shown]

Sample no.	Au	B	Cu	Mn	Sn	W	Zn
81–1	N	>5,000	>50,000	500	N	N	2,000
81–2	N	1,000	300	2,000	N	N	1,500
81–3	N	150	200	>10,000	N	N	N
81–4	N	700	300	700	N	<100	500
81–5	N	2,000	500	5,000	N	N	N
81–6	N	1,500	70	700	200	N	N
81–10	N	1,500	700	1,000	20	<100	N
81–11	N	50	15	>10,000	N	N	N
81–12	[1]0.45	50	10	>10,000	N	N	N
81–13	<.07	500	30	2,000	N	N	N
81–14	1.90	500	10	1,500	N	<100	N
81–15	.36	200	10	300	N	N	N
81–16	.07	500	1,500	1,500	N	N	<500
81–17	N	700	500	1,000	N	N	1,500
81–18	N	300	30	1,000	N	N	N
81–19	N	2,000	500	700	50	<100	N
81–20	N	300	500	500	N	N	N
81–21	N	1,000	150	700	500	N	<500
81–22	N	300	<10	1,000	N	<100	N
81–23	N	500	50	200	N	N	N
81–24	N	200	30	2,000	N	N	N
81–25	N	500	30	700	N	N	1,500
81–26	N	200	10	1,000	N	N	N
81–28	N	3,000	<10	500	50	N	N
81–29	N	700	<10	700	<20	N	500
81–30	N	700	100	700	N	N	<500
81–31	N	200	20	10,000	N	N	500
81–33	N	700	500	2,000	N	N	1,500
81–35	N	1,000	100	1,000	N	N	1,000
81–36	[1]3.00	200	20	500	N	N	500
81–37	2.40	700	<10	700	N	N	1,000

[1] Gold observed during field and (or) laboratory processing of samples by J.W. Whitlow.

the same site. Similar stream sediment anomalies for copper and zinc are present in sample 81–20, just south of Gove Hill. The only stream sediment sample from this area that contains significant gold is 81–13 (0.15 ppm, table 2).

Ely Mine Area

Most of the anomalies for heavy-mineral concentrates in the Ely mine area (see figs. 7 and 8) seem to be directly related to the mine or to known prospects (see White and Eric, 1944; Hermance and others, 1949). Sample 81–57 that drains the mine has 20,000 ppm Cu, 500 ppm Zn, 3 ppm Ag, and 500 ppm B, whereas sample 81–59 (downstream of the mine) contains >50,000 ppm Cu, 3,000 ppm Zn, 50 ppm Ag, 30 ppm Mo, 6.9 ppm Au, and 5,000 ppm B (table 3). The higher geochemical values in sample 81–59, collected downstream of the mine dumps and tailings, suggest that these materials are a major source of heavy minerals and metals in the streams. Sample 81–63, from a large basin farther downstream from the mine area, contains 7,000 ppm Cu and 100 ppm Sn but lacks anomalous Zn, Ag, Au, or B. Concentrates 81–56, 81–58, and 81–60 contain 1,000 to 5,000 ppm copper; sample 81–58 also has 5,000 ppm boron. The copper values in samples 81–56 and 81–60 may be related to sulfide mineral prospects to the southwest of the Ely mine area (fig. 7). Sample 81–58 is from a basin that drains the subsurface eastern extension of the Ely orebody and apparently is providing copper (from chalcopyrite) and boron (from tourmaline) to the drainage. The higher boron value in this sample (5,000 ppm), relative to that from the immediate mine area (81–57, 500 ppm), suggests that tourmaline-rich rocks or tourmalinites are present in the bedrock to the east of the Ely mine. The low levels of detectable gold (0.47 ppm) and tungsten (<100 ppm) in sample 81–53 to the west of the mine (fig. 8) may reflect undiscovered mineralization near the contact of the Waits River and Gile Mountain Formations or glacially transported material derived from the north.

The stream sediment samples (table 4) display a pattern of anomalies like those of the heavy-mineral concentrates but at much lower levels. In particular, all three samples that drain the mine and dump areas (fig. 9, 81–57, 81–58, and 81–59) contain anomalous (≥500 ppm) copper; sample 81–59 also has 0.25 ppm gold, and sample 81–57 has 100 ppm lead. All of these anomalies presumably are related to drainage from the dumps and tailings in the Ely mine area. The manganese anomaly in sample 81–54 to the southwest of the Ely mine may reflect coticule rocks associated with prospects in this drainage basin (fig. 9) or coticule present in the Standing Pond Volcanics to the west.

Pike Hill Mine Area

The Pike Hill mine area contains three closely spaced mine sites that have extensive dumps. White and Eric (1944) briefly describe the geology and sulfide deposits of the Eureka, Union, and Smith mines. Drainages from these mines yield heavy-mineral concentrates that have the largest Cu, Zn, Co, and Au values in the district, as well as local Sn, W, Pb, Cd, and Mn anomalies (table 5; Adrian and others, 1989).

The largest Pike Hill mine is the Union, which is within the basin of sample 81–16 (fig. 10). A heavy-mineral concentrate from this sample contains 10,000 ppm copper, 3,000 ppm zinc, 150 ppm cobalt, and 100 ppm cadmium. This basin drains into basin 81–15 to the east, where the anomalies persist and partly increase (>50,000 ppm copper, 10,000 ppm zinc, 100 ppm cobalt). The higher copper and zinc values in sample 81–15 reflect the distribution of mine dumps and tailings to the east of the Union mine.

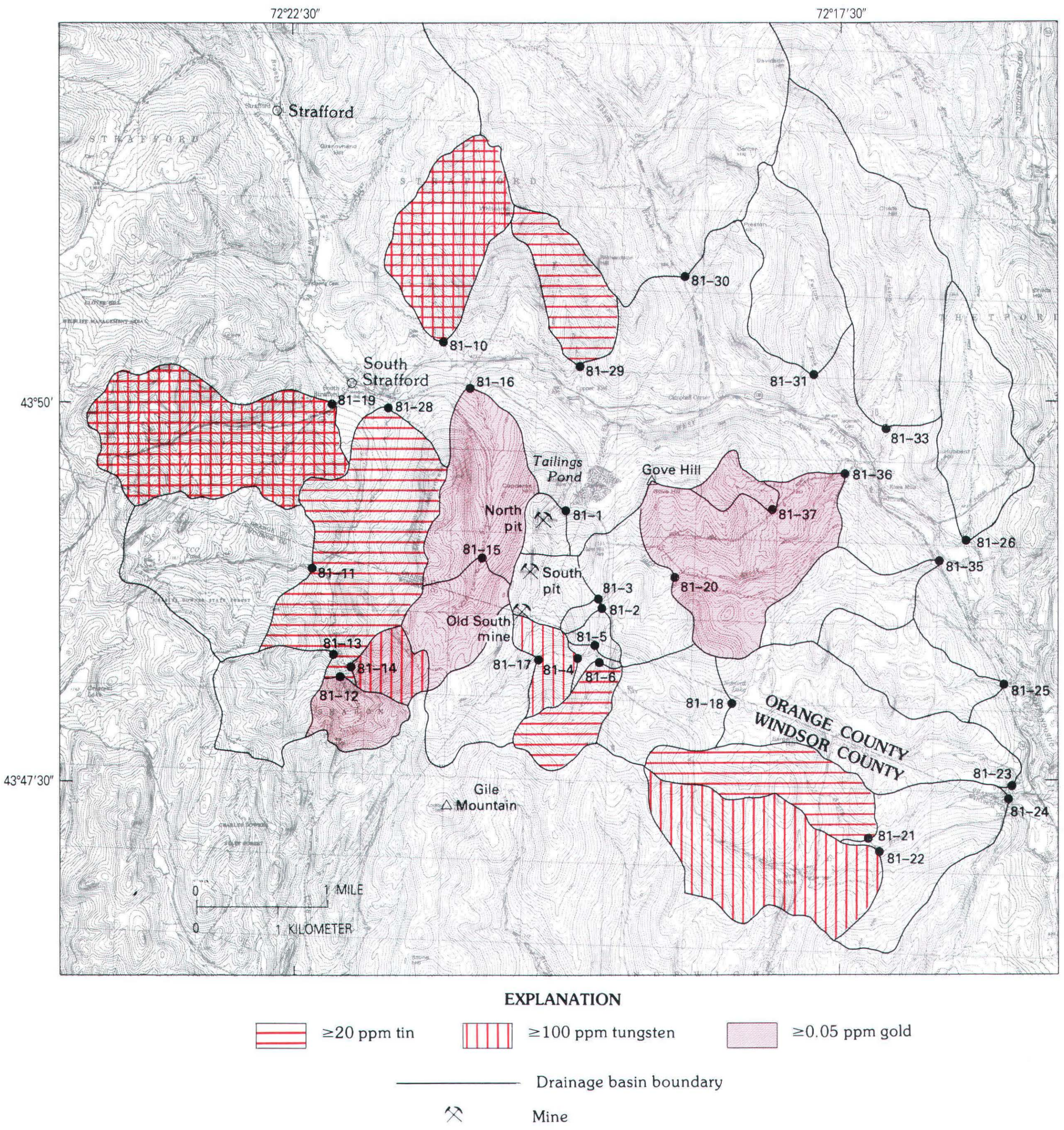

Figure 5. Distribution of geochemically anomalous tin, tungsten, and gold in samples of heavy-mineral concentrates from the Elizabeth mine area. See table 1 for anomalous values. Note that for some drainage basins more than one pattern is shown to indicate the presence of more than one anomalous element value. Base from U.S. Geological Survey, Sharon and South Strafford, 1981. Contour interval 20 feet. National Geodetic Vertical Datum of 1929.

These anomalies continue in concentrate samples farther downstream at sites 81–22 (>50,000 ppm Cu, 5,000 ppm Zn, 50 ppm Ag) and 81–21 (>50,000 ppm Cu, 10,000 ppm Zn, 50 ppm Ag), both of which are from drainages that flow into the Waits River at the eastern edge of the study area. Manganese anomalies are also characteristic of the basins surrounding the mines, including greater than 10,000 ppm in sample 81–16. This high value of manganese probably reflects the presence of spessartine garnet in the concentrates (coticule rocks are present in drill core from here). Elevated manganese values (10,000 ppm) in heavy-mineral concentrates from adjacent drainages including 81–5 and 81–19 (fig. 10) probably have a similar source. The high manganese in sample 81–19 suggests a northern extension of the ore-related sequence of wall rocks associated with the massive sulfide deposits at the Eureka and Union mines.

Table 2. Selected analyses of stream sediments from the Elizabeth mine area, Vermont

[All values are in parts per million. Analytical detection limits for the elements are as follows: Au, 0.05; Co, 5; Cu, 5; Mn, 10; Zn, 200. N=not detected; <=detected but less than value shown]

Sample no.	Au	Co	Cu	Mn	Zn
81–1	N	15	5,000	200	300
81–2	N	15	20	1,000	N
81–3	N	20	1,000	500	N
81–4	N	15	20	1,000	N
81–5	N	15	15	1,500	N
81–6	N	15	15	5,000	N
81–10	N	15	15	1,500	N
81–11	<0.05	20	100	2,000	N
81–12	<.05	20	30	3,000	N
81–13	.15	20	50	3,000	N
81–14	N	20	50	2,000	N
81–15	N	30	50	2,000	N
81–16	N	30	150	3,000	N
81–17	N	30	50	2,000	N
81–18	N	20	30	3,000	N
81–19	N	20	30	2,000	N
81–20	N	50	500	3,000	300
81–21	N	30	30	3,000	N
81–22	N	30	30	3,000	N
81–23	N	30	30	3,000	N
81–24	N	20	20	3,000	N
81–25	N	15	15	3,000	N
81–26	<.05	30	20	700	N
81–28	N	20	20	3,000	N
81–29	N	20	50	1,500	N
81–30	N	15	20	1,500	N
81–31	N	20	20	2,000	N
81–33	N	30	500	2,000	300
81–35	N	10	10	2,000	<200
81–36	N	30	150	3,000	<200
81–37	N	10	15	1,500	N

Table 3. Selected analyses of heavy-mineral concentrates from the Ely mine area, Vermont

[All values are in parts per million. Analytical detection limits for the elements are as follows: Au, 0.05; B, 20; Cu, 10; Mn, 20; Sn, 20; W, 100; Zn, 500. N=not detected; >=greater than value shown; <=detected but less than value shown]

Sample no.	Au	B	Cu	Mn	Sn	W	Zn
81–50	<0.06	300	10	1,000	N	N	<500
81–51	N	500	70	1,000	N	N	<500
81–53	[1].47	500	10	1,000	N	<100	<500
81–54	N	1,000	20	1,000	N	N	700
81–55	N	300	<10	150	N	N	N
81–56	N	500	1,000	700	N	N	N
81–57	N	500	20,000	700	N	N	500
81–58	N	5,000	5,000	1,500	50	N	700
81–59	6.90	5,000	>50,000	500	<20	N	3,000
81–60	N	1,500	1,500	1,500	N	N	500
81–61	N	1,500	200	1,000	N	N	500
81–62	N	700	70	700	N	N	N
81–63	N	700	7,000	700	100	N	<500
82–58[2]	2.30	300	10	200	N	<100	N

[1] Gold observed during field and (or) laboratory processing of samples by J.W. Whitlow.

[2] Sample from the eastern side of the Brocklebank granite (fig. 1).

Basins that contain gold-bearing concentrates trend approximately parallel to strike through the middle of the study area (fig. 11). These anomalies seem to be unrelated to the existing mines and prospects. From north to south, the following samples have anomalous gold values: 82–74, 2.4 ppm; 81–8, 7.5 ppm; 81–10, 7.2 ppm; 82–80, 1.9 ppm; 81–11, 6.0 ppm; 81–12, 2.9 ppm; 81–4, 6.0 ppm; 81–2, 9.0 ppm; 81–14, 41.0 ppm; 82–82, 14.0 ppm; and 82–64, 5.6 ppm. These last two samples, which are from the southern edge of the study area, contain lead anomalies of 300 ppm and 7,000 ppm, respectively (Adrian and others, 1989). Sample 81–14 also has 150 ppm tin and traces of fine-grained scheelite (table 5).

Several of the heavy-mineral concentrates have anomalous tungsten (fig. 11). Samples 81–10, 81–11, 82–68, and 82–82 each contain 300 ppm tungsten, and samples 81–9, 82–64, and 82–74 have 100 to 200 ppm tungsten. The tungsten-bearing drainage basins have no clear relation to known mines or prospects in the Pike Hill area, an interpretation that is supported by the lack of detectable (<15 ppm) tungsten in sulfide ore samples from the district (J.F. Slack, unpub. data).

Data for stream sediments (fig. 12 and table 6) are less sensitive geochemically but generally correlate with those for the heavy-mineral concentrates. This correlation is well documented for copper in basins that drain the Eureka and Union mines and in basins to the east. Surprisingly, gold was found in some stream sediment samples where it was not detected in the heavy-mineral concentrates taken from the same site.

Stream sediments show a well-defined dispersion of geochemical anomalies away from the Union mine dumps. Basin 81–16, which contains the mine, has 7,000 ppm copper, 700 ppm zinc, 500 ppm cobalt, and 100 ppm lead (no lead was present in the concentrate). Downstream, stream sediment sample 81–15 contains 2,000 ppm copper, 500 ppm zinc, 200 ppm cobalt, and 0.35 ppm gold (which was not detected in the concentrate). Sample 81–22, still farther downstream, has 1,500 ppm copper, 200 ppm zinc, and 200 ppm cobalt; the low zinc value is not anomalous. There may be undiscovered mineralization in basin 81–21 because metal values increase here to 7,000 ppm copper, 1,500 ppm zinc, and 700 ppm cobalt (the highest in the district). Alternatively, sample site 81–21 may be unrepresentative of the drainage as a whole, and the high metal values might be due to selected metal concentration, perhaps in organic material in the stream bed.

The gold belt defined by the heavy-mineral concentrates (fig. 11) is not expressed in the stream sediments, except for sample 81–14. This stream sediment sample contains 0.5 ppm gold and 100 ppm lead; these values are

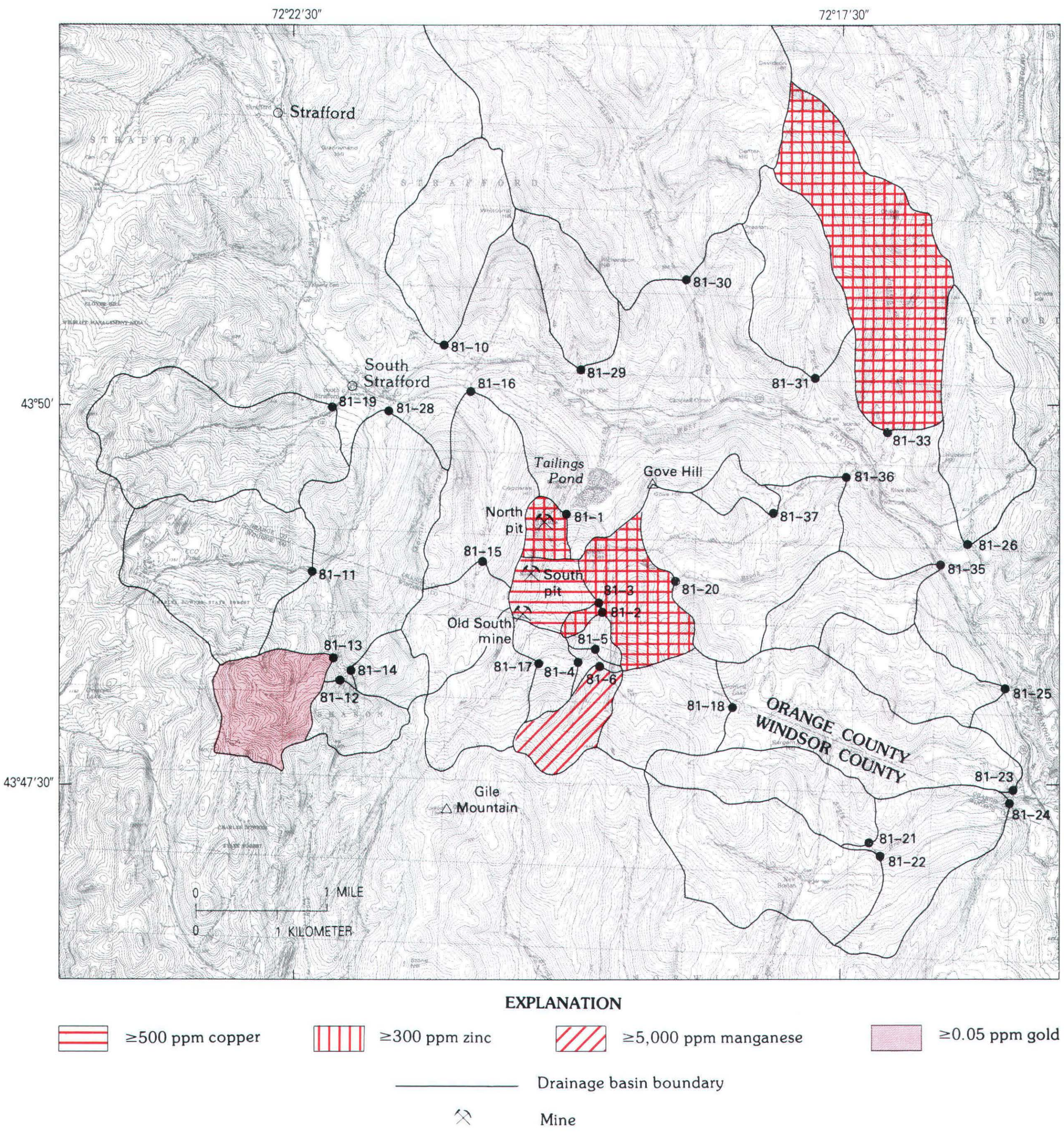

Figure 6. Distribution of geochemically anomalous samples of stream sediment from the Elizabeth mine area. See table 2 for anomalous values. Note that for some drainage basins more than one pattern is shown to indicate the presence of more than one anomalous element value. Base from U.S. Geological Survey, Sharon and South Strafford, 1981. Contour interval 20 feet. National Geodetic Vertical Datum of 1929.

consistent with gold and lead anomalies in the concentrate from the same site. Sample 81–4 has anomalous gold (table 5, 6.0 ppm) in the concentrate, but the stream sediment carries 700 ppm copper, 700 ppm zinc, and no detectable gold (<0.05 ppm). Other than the mine area sample (81–15) and that to the south (81–14), gold was detected in only one isolated sample of stream sediment (82–67, 0.55 ppm) to the west.

CORRELATION WITH OTHER DATA

Parts of the Pike Hill area were studied by geobotanical remote sensing methods by Power and Milton (this volume). Airborne spectral data of the forest canopy were acquired and analyzed for spectral anomalies that may be related to mineralization. Many such anomalies were identified in the Pike Hill area, one of which is directly over the

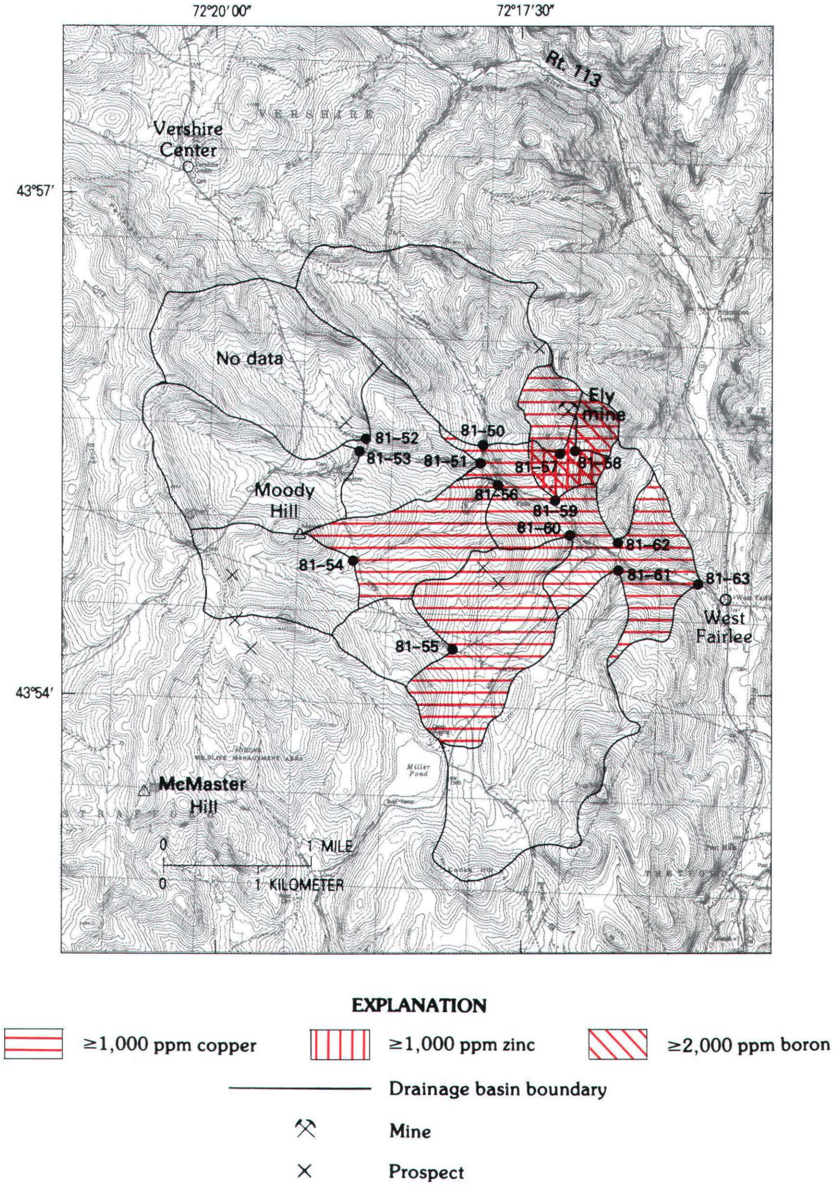

Figure 7. Distribution of geochemically anomalous copper, zinc, and boron in samples of heavy-mineral concentrates from the Ely mine area. See table 3 for anomalous values. Note that for some drainage basins more than one pattern is shown to indicate the presence of more than one anomalous element value. Base from U.S. Geological Survey, Vershire, 1981. Contour interval 20 feet. National Geodetic Vertical Datum of 1929.

Eureka and Union mines. Other spectral anomalies are located in basins that lack known mineralization but which have geochemical anomalies. Strong spectral anomalies occur in basins 81–1, 81–6, 81–15, 81–16, 81–21, 81–22, 82–68, 82–70, 82–72, 82–74, and 82–82. Geochemical anomalies exist in all of these basins. In basin 82–72, where the only anomalous metal is tungsten (<100 ppm), the spectral anomalies are concentrated in a small subbasin to the southeast of Marston Hill (82–67/68), and potential mineralization in 82–72 may be masked by sediment dilution in this large drainage basin (see fig. 11).

REGIONAL CONSIDERATIONS

The geochemical anomalies identified in stream sediments and heavy-mineral concentrates of the district probably reflect multiple sources of metals and other elements. Anomalies for Cu, Zn, Ag, Co, Cd, Mn, and B commonly are directly associated with occurrences of stratabound sulfide deposits in the Elizabeth, Ely, and Pike Hill mine areas. In contrast, drainage basins that contain anomalous gold, tungsten, tin, and lead are, in most cases, separated spatially from the known sulfide deposits. This pattern is

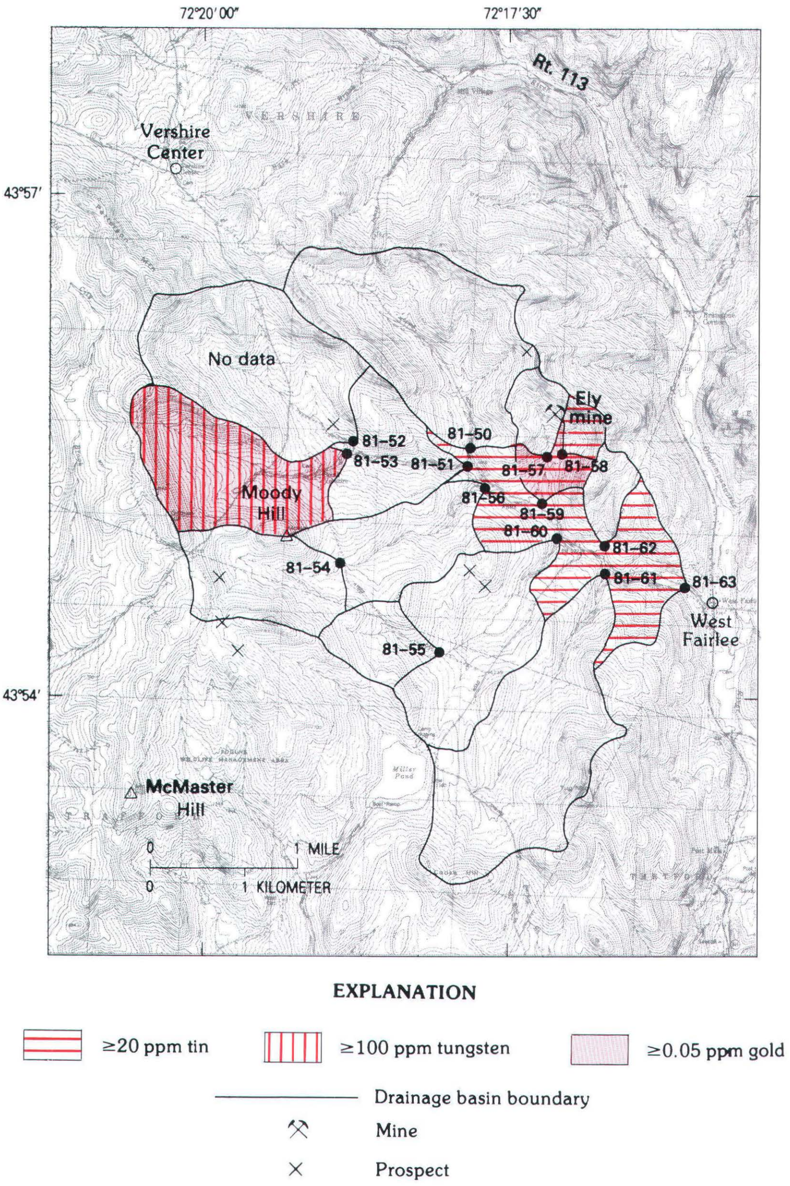

Figure 8. Distribution of geochemically anomalous tin, tungsten, and gold in samples of heavy-mineral concentrates from the Ely mine area. See table 3 for anomalous values. Note that for some drainage basins more than one pattern is shown to indicate the presence of more than one anomalous element value. Base from U.S. Geological Survey, Vershire, 1981. Contour interval 20 feet. National Geodetic Vertical Datum of 1929.

consistent with the mineralogic and chemical composition of the orebodies, which lack significant amounts of gold, tungsten, tin, or lead (Slack and others, 1986; J.F. Slack, unpub. data).

Figure 13 shows the distribution of anomalous copper and gold in the nonmagnetic fraction of heavy-mineral concentrates plotted on a regional geologic base. Drainage basins that contain high copper values are clustered around the largest mines in the district, whereas gold-bearing basins generally are distal. This separation of copper from gold suggests different bedrock sources for these metals; similar separations of gold and base metals are known in other massive sulfide districts of the Appalachians (for example, Craig, 1980; German, 1985). In an earlier report, Slack and others (1984b) suggested that the gold anomalies in the Orange County copper district may be derived from stratabound, exhalative deposits that formed by the same processes as the massive sulfide bodies (that is, coeval with Paleozoic sedimentation and mafic volcanism). Although this is one possible origin, it is equally likely that the gold

Table 4. Selected analyses of stream sediments from the Ely mine area, Vermont

[All values are in parts per million. Analytical detection limits for the elements are as follows: Au, 0.05; Co, 5; Cu, 5; Mn, 10; Zn, 200. N=not detected; <=detected but less than value shown]

Sample no.	Au	Co	Cu	Mn	Zn
81–50	N	20	70	2,000	N
81–51	N	15	30	2,000	N
81–52	N	15	50	1,500	N
81–53	N	15	30	1,500	N
81–54	<0.05	20	30	5,000	N
81–55	N	20	30	1,500	N
81–56	N	15	50	2,000	N
81–57	N	100	2,000	3,000	<200
81–58	.05	30	1,000	2,000	<200
81–59	.25	30	5,000	2,000	500
81–60	.05	15	30	3,000	N
81–61	N	15	30	2,000	N
81–62	N	15	70	3,000	N
81–63	N	30	500	2,000	<200

source is from bedrock occurrences of epigenetic veins. Support for this latter interpretation comes from the presence of anomalous (2.3 ppm) gold in concentrate sample 82–58 (table 3) collected from a small basin that drains the southeastern side of the Brocklebank granite (informal name) about 8 km west of the Ely mine (fig. 13). This granite is unfoliated and probably is Devonian or Carboniferous in age. The gold-bearing concentrate sample taken from the area of this granite lacks anomalous copper (10 ppm) but contains detectable (<100 ppm), but not determinable, tungsten. Anomalous tungsten values, including visible grains of scheelite as large as 3 mm in diameter, also have been obtained by P.J. Atelsek from the nonmagnetic fraction of heavy-mineral concentrates collected near the margins of the Knox Mountain Granite to the north (see Doll and others, 1961, for geologic relations). This tungsten is believed to come from granite-related mineral deposits associated with Devonian or Carboniferous plutons. Some of the gold anomalies in the copper district also may be derived from epigenetic, hydrothermal deposits (including in shear zones and ductile faults), possibly related to metamorphic fluid flow in the Waits River Formation (Ferry, 1990).

Unlike the elements related to the stratabound sulfides, the gold, tungsten, and tin anomalies may not come from the local metamorphic bedrock, as there are no granite plutons exposed near the sulfide deposits. One possible source is the Knox Mountain Granite, from which heavy minerals such as gold, scheelite, cassiterite, and perhaps galena could have been transported to the south in glacial till. This theory is supported by the occurrence of 41 ppm gold and 7,000 ppm lead—the highest values for these metals in the district—in heavy-mineral concentrate 81–14 that drains a north-facing slope at the southern end of the Pike Hill study area; this sample also contains 150 ppm tin and traces of fine-grained scheelite (table 5 and fig. 10; Adrian and others, 1989). The lack of known mineralization in this immediate area and the topographic character of the drainage basin are consistent with derivation of these metals from glacial till that was preferentially deposited against a north-facing topographic barrier.

Table 5. Selected analyses of heavy-mineral concentrates from the Pike Hill mine area, Vermont

[All values are in parts per million. Analytical detection limits for the elements are as follows: Au, 0.05; B, 20; Cu, 10; Mn, 20; Sn, 20; W, 100; Zn, 500. N=not detected; >=greater than value shown; <=detected but less than value shown]

Sample no.	Au	B	Cu	Mn	Sn	W	Zn
81–1	N	700	10	1,500	N	100	<500
81–2	9.00	500	300	1,500	<20	N	N
81–3	N	500	500	1,000	N	N	N
81–4	6.00	700	150	2,000	N	<100	500
81–5	N	700	500	10,000	N	N	500
81–6	N	700	70	2,000	50	N	N
81–7	.75	700	150	2,000	N	N	500
81–8	7.50	1,500	500	1,500	70	<100	<500
81–9	N	1,000	10	1,500	30	200	500
81–10	7.20	1,500	20	1,500	30	300	500
81–11	6.00	700	70	1,000	N	300	<500
81–12	2.90	500	<10	1,500	N	<100	<500
81–13	N	700	10	2,000	30	N	<500
81–14	[1]41.0	700	300	2,000	150	N[2]	500
81–15	.05	500	>50,000	1,500	N	N	10,000
81–16	N	100	10,000	>10,000	N	N	3,000
81–17	N	500	70	2,000	N	N	N
81–18	N	300	100	1,500	N	N	N
81–19	N	500	150	10,000	N	N	<500
81–20	N	300	15	1,500	N	<100	N
81–21	N	200	>50,000	1,000	N	N	10,000
81–22	N	200	>50,000	1,500	<20	N	5,000
82–64	5.60	200	300	200	150	100	N
82–66	N	200	N	100	N	<100	N
82–68	N	300	<10	500	20	300	N
82–70	N	150	100	300	N	<100	N
82–72	N	200	<10	300	N	<100	N
82–74	2.40	100	<10	200	<20	200	N
82–76	N	150	N	200	N	N	N
82–78	N	70	N	100	N	N	N
82–80[3]	1.90	100	N	300	N	N	N
82–82	14.0	300	50	200	N	300	N
82–84[4]	N	300	<10	200	N	<100	N

[1] Gold observed during field and (or) laboratory processing of samples by J.W. Whitlow.

[2] Trace amounts of fine-grained (<1 mm) found during examination of nonmagnetic fraction under ultraviolet light.

[3] Duplicate sample from site no. 81–11.

[4] Duplicate sample from site no. 81–1.

The southern end of the Knox Mountain Granite is approximately 4 km north-northwest of the Pike Hill mines. Surficial geologic studies by Larsen (1972, 1987) have shown that indicator fans of granitic pebbles and cobbles from the Knox Mountain Granite were transported in till as much as 25 km in a southeasterly direction and that those from the small Brocklebank granite (figs. 1 and 13) were

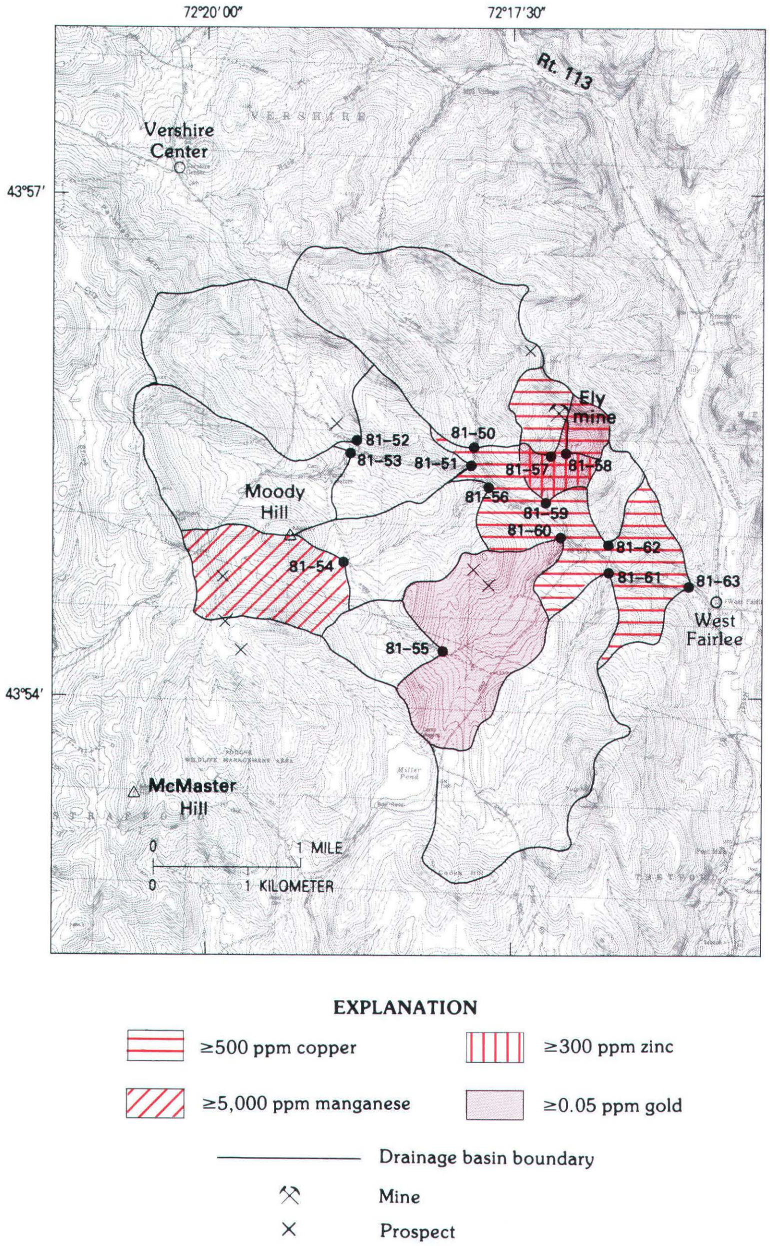

Figure 9. Distribution of geochemically anomalous samples of stream sediment from the Ely mine area. See table 4 for anomalous values. Note that for some drainage basins more than one pattern is shown to indicate the presence of more than one anomalous element value. Base from U.S. Geological Survey, Vershire, 1981. Contour interval 20 feet. National Geodetic Vertical Datum of 1929.

transported as much as 10 km to the southeast. The Pike Hill mine area is southeast of the Knox Mountain Granite and thus is in a favorable location to have received glacial materials from this pluton. The contacts of the Knox Mountain Granite are poorly exposed (Murthy, 1957), and no study has been able to thoroughly evaluate the possible presence (or absence) of contact metasomatism in the calcareous country rocks of the surrounding Waits River Formation. The only recent work on the Knox Mountain Granite is a geochemical and petrologic investigation by Griffin (1982) that documents several intrusive phases and a peraluminous, S-type magma chemistry reflecting an origin by partial melting of a metasedimentary source (see also Griffin and Anderson, 1984). Such peraluminous granites commonly are associated with tungsten, tin, and other metals in hydrothermal mineral deposits throughout the world (for example Kooiman and others, 1986; Dubray and others, 1988), although not all tungsten-related granites are peraluminous (Newberry and Swanson, 1986). Because of the lack of detailed geologic information about the

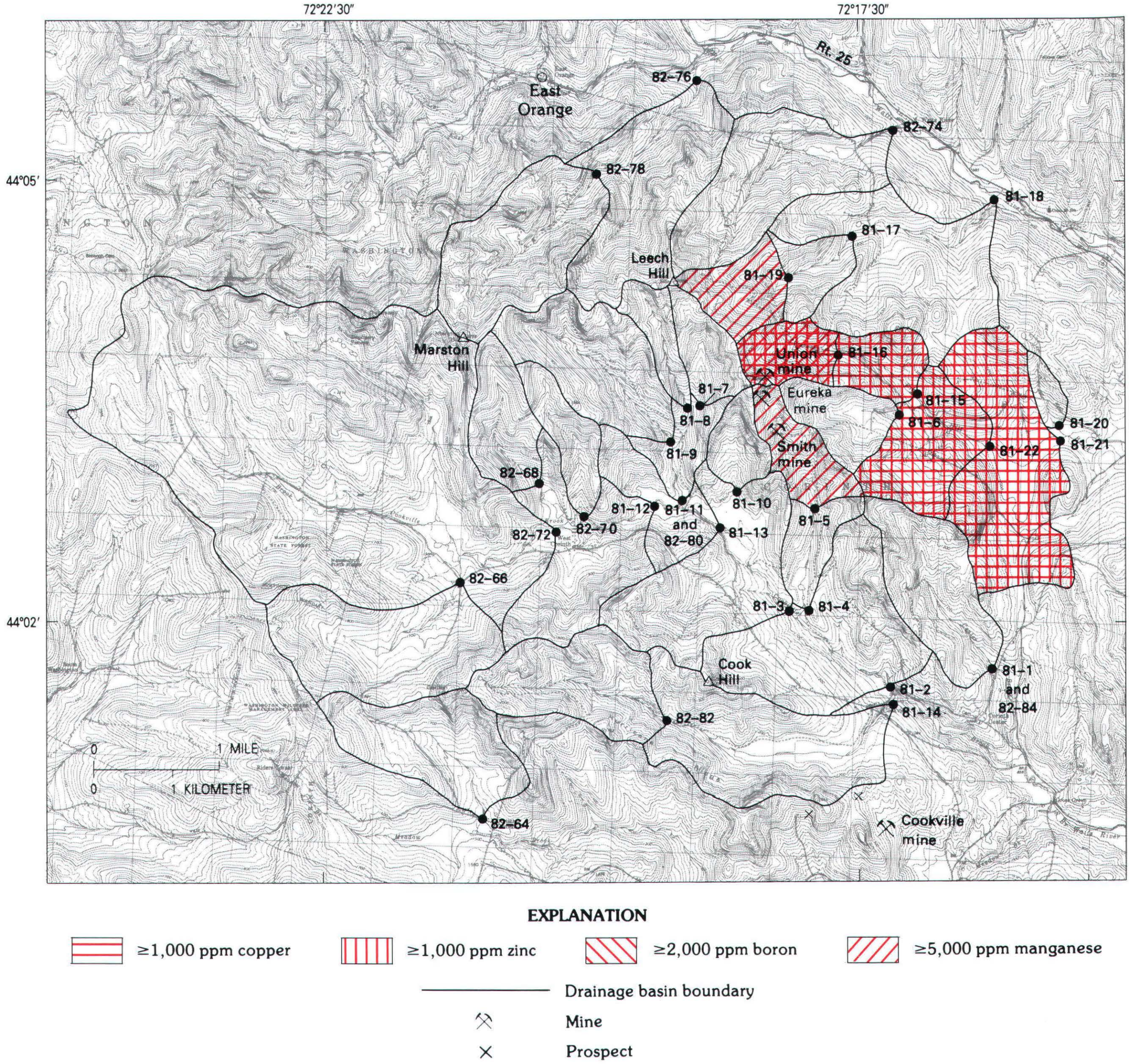

Figure 10. Distribution of geochemically anomalous copper, zinc, boron, and manganese in samples of heavy-mineral concentrates from the Pike Hill mine area. See table 5 for anomalous values. Note that for some drainage basins more than one pattern is shown to indicate the presence of more than one anomalous element value. Base from U.S. Geological Survey, Washington and West Topsham, 1981. Contour interval 20 feet. National Geodetic Vertical Datum of 1929.

contact relations of the pluton, and because of the favorable nature of the igneous rock chemistry, a previously unrecognized potential for skarn and other granite-related hydrothermal mineral deposits is suggested for the Knox Mountain Granite. Additional work may extend this potential to other granitic plutons to the north, including those in the so-called northeast kingdom area of Vermont (Doll and others, 1961; Ayuso and Arth, 1985), where a small scheelite-bearing skarn deposit is developed along the northern contact of the Averill pluton in southeastern Quebec (Gauthier and others, 1985, 1989).

In a reconnaissance survey, Watts (this volume) also found anomalous levels of W and Sn in heavy-mineral concentrates from the Elizabeth and Ely mine areas and high values for Cu, Pb, B, Sr, and Mn in some samples (see Day and others, 1986, for analytical data). Watts (this volume) suggests that mineralized sources for the tungsten and tin anomalies are buried granites beneath the Strafford dome and along the Strafford-Willoughby arch (see Doll and others, 1961, for geologic relations). His model is supported by the presence of granitic and pegmatitic dikes, 2 km to the west of the Pike Hill mines (White and Eric,

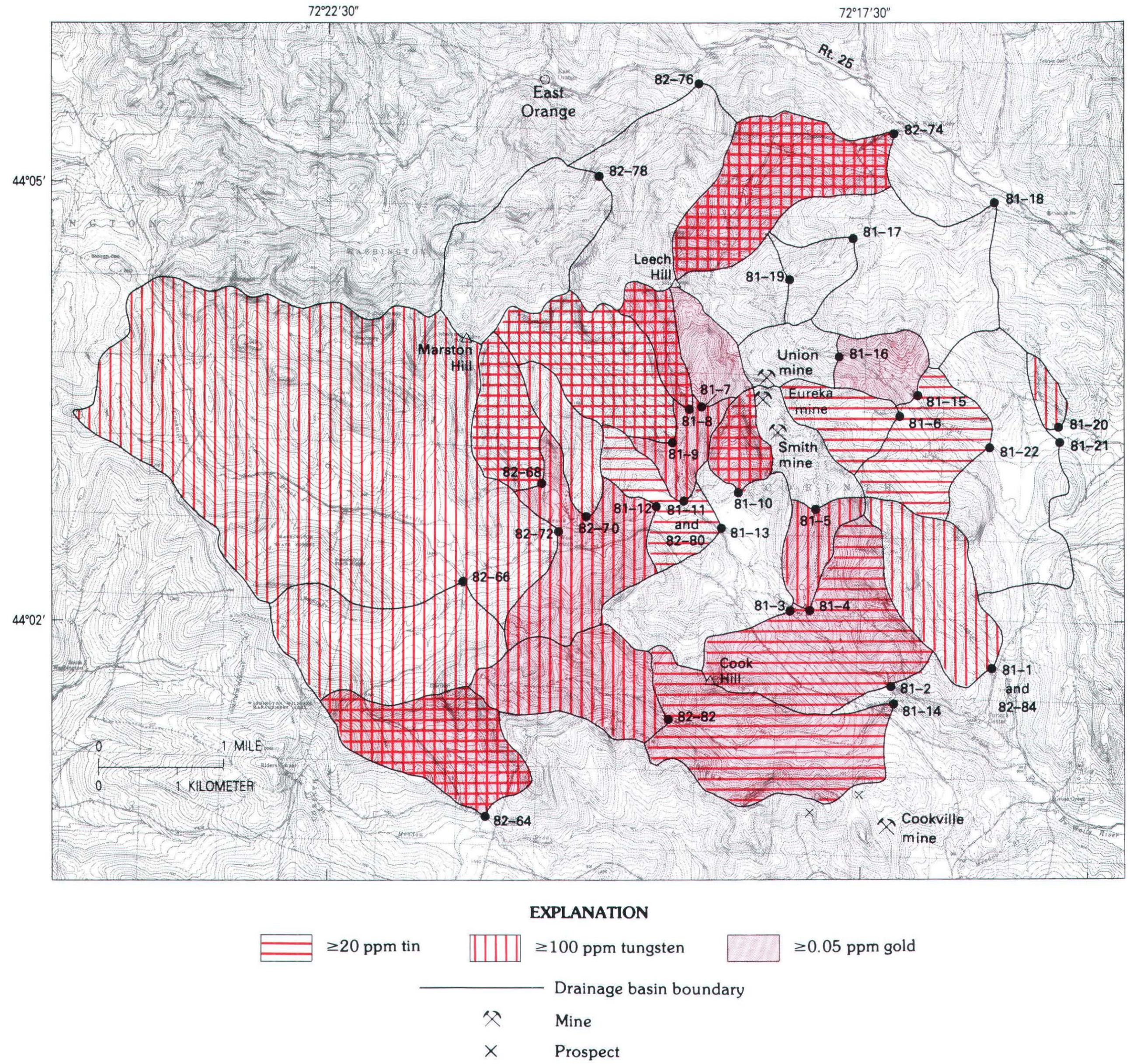

Figure 11. Distribution of geochemically anomalous tin, tungsten, and gold in samples of heavy-mineral concentrates from the Pike Hill mine area. See table 5 for anomalous values. Note that for some drainage basins more than one pattern is shown to indicate the presence of more than one anomalous element value. Base from U.S. Geological Survey, Washington and West Topsham, 1981. Contour interval 20 feet. National Geodetic Vertical Datum of 1929.

1944; Murthy, 1957), and by abundant granitic dikes on Cook Hill, 4 km south of the mines (Murthy, 1957). A gravity survey by Bean (1953) suggests that a low-density mass is present in the core of the Pomfret and Strafford domes, possibly caused by granite at depth beneath the more dense carbonate-rich rocks of the Waits River Formation exposed at the surface. These relations are consistent with a local derivation for the tungsten, tin, and gold anomalies in the district from granite-related, hydrothermal mineral deposits present in areas proximal to the copper deposits (Watts, this volume). However, the lack of known bedrock occurrences of tin, tungsten, gold, and lead in the district and the distribution of anomalies relative to glacial features are equally consistent with a distal source of the metals from outside the district, such as from the area of the Knox Mountain Granite to the north. The more abundant tin, tungsten, and gold anomalies in the Pike Hill area, relative to those near Ely and Elizabeth, support this interpretation. The available geologic and geochemical data do not preclude either model, and more work will be required to determine the bedrock source(s) of tin, tungsten, gold, and lead in the Orange County copper district.

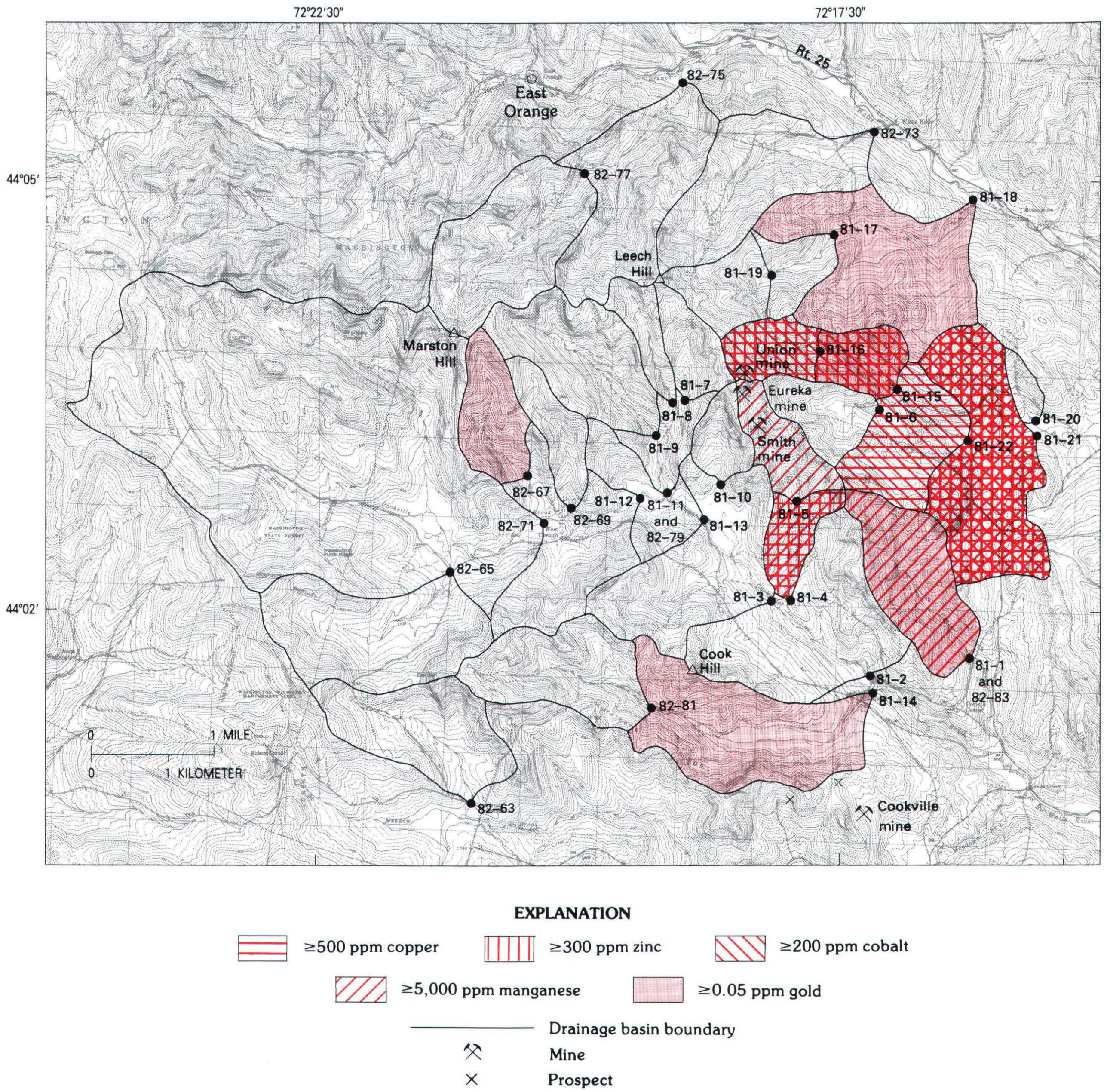

Figure 12. Distribution of geochemically anomalous samples of stream sediment from the Pike Hill mine area. See table 6 for anomalous values. Note that for some drainage basins more than one pattern is shown to indicate the presence of more than one anomalous element value. Base from U.S. Geological Survey, Washington and West Topsham, 1981. Contour interval 20 feet. National Geodetic Vertical Datum of 1929.

Table 6. Selected analyses of stream sediments from the Pike Hill mine area, Vermont

[All values are in parts per million. Analytical detection limits for the elements are as follows: Au, 0.05; Co, 5; Cu, 5; Mn, 10; Zn, 200. N=not detected; <=detected but less than value shown]

Sample no.	Au	Co	Cu	Mn	Zn
81–1	0.05	20	50	5,000	N
81–2	<.05	15	30	3,000	N
81–3	<.05	15	50	2,000	N
81–4	<.05	70	700	5,000	700
81–5	N	15	70	5,000	<200
81–6	N	10	30	3,000	N
81–7	N	15	70	2,000	N
81–8	N	10	30	1,000	N
81–9	N	10	20	2,000	N
81–10	N	20	70	3,000	N
81–11	N	15	30	2,000	N
81–12	N	15	20	2,000	N
81–13	<.05	15	30	1,500	N
81–14	.50	10	30	2,000	N
81–15	.35	200	2,000	3,000	500
81–16	<.05	500	7,000	2,000	700
81–17	N	30	150	3,000	N
81–18	.05	15	20	3,000	N
81–19	<.05	30	50	3,000	N
81–20	N	5	20	1,000	N
81–21	N	700	7,000	5,000	1,500
81–22	<0.05	200	1,500	2,000	200
82–63	N	10	7	300	N
82–65	N	15	10	700	N
82–67	.55	5	5	200	N
82–69	N	10	7	1,000	N
82–71	N	10	5	700	N
82–73	N	N	<5	1,500	N
82–75	N	10	<5	500	N
82–77	N	10	<5	500	N
82–79[1]	N	5	5	500	N
82–81	N	10	20	700	N
82–83[2]	N	20	15	700	N

[1] Duplicate sample from site no. 81–11.

[2] Duplicate sample from site no. 81–1.

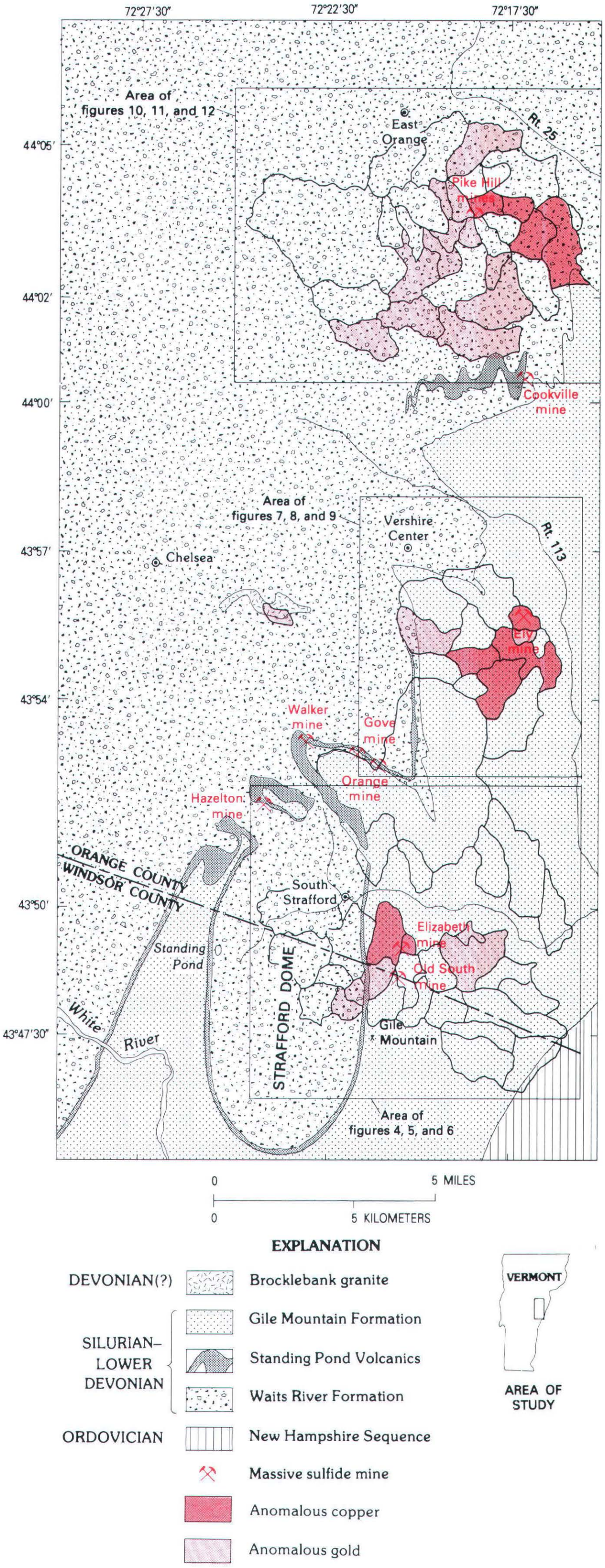

Figure 13. Regional distribution of anomalous values of ▶ copper and gold in heavy-mineral concentrates from drainage basins in the Orange County copper district. Geology after Doll and others (1961). Waits River and Gile Mountain Formations and Standing Pond Volcanics compose the Vermont Sequence.

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Chapter R

Preliminary Assessment of Metallic Mineral Resources in the Glens Falls 1° × 2° Quadrangle, New York, Vermont, and New Hampshire

By JOHN F. SLACK

U.S. GEOLOGICAL SURVEY BULLETIN 1887

SUMMARY RESULTS OF THE GLENS FALLS CUSMAP PROJECT, NEW YORK, VERMONT, AND NEW HAMPSHIRE

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Preliminary Assessment of Metallic Mineral Resources in the Glens Falls 1° × 2° Quadrangle, New York, Vermont, and New Hampshire

By John F. Slack[1]

Abstract

A preliminary assessment of metallic mineral resources is presented for the Glens Falls 1° × 2° quadrangle, New York, Vermont, and New Hampshire. The assessment is based on the use of diagnostic and permissive recognition criteria, including (1) favorable geology; (2) the distribution of mines, prospects, and mineral occurrences; (3) geochemical anomalies (principally in heavy-mineral concentrates of stream sediments); and (4) geophysical anomalies. These recognition criteria are used in conjunction with appropriate ore deposit models to evaluate the potential for different types of metallic mineral deposits; nonmetallic deposits (and fuels) are not considered. Areas that have one recognition criterion are judged to have low resource potential, and areas that have two or more criteria are assigned moderate and high resource potential, respectively.

Areas that have moderate to high resource potential for different deposit types in the quadrangle include (1) volcanogenic massive sulfide deposits of copper and zinc (±lead, silver, gold) in metavolcanic and metasedimentary belts (chiefly early Paleozoic) in western New Hampshire and eastern Vermont; (2) sedimentary exhalative (sedex)-type lead-zinc (±copper, silver, barite) deposits in Proterozoic basement rocks of the eastern Adirondack Mountains, New York, and in Paleozoic rocks constituting both the shelf sequence and the Taconic allochthon of eastern New York and western Vermont; (3) volcanogenic gold (±silver) deposits in early Paleozoic belts of western New Hampshire and eastern Vermont, and possibly also in Proterozoic rocks of the eastern Adirondacks, New York; (4) metamorphogenic gold-bearing quartz veins in the Late Proterozoic to Cambrian eastern cover sequence of the Green Mountain massif in central Vermont; (5) gold associated with altered ultramafic rocks in the eastern cover sequence of Vermont; (6) porphyry-related gold associated with the Mesozoic alkaline intrusive complex at Cuttingsville, Vermont, and possibly also in Proterozoic basement units of augen gneiss in Vermont; (7) epigenetic gold in veins, stockworks, and (or) disseminations within early Paleozoic rocks of the Taconic allochthon of eastern New York and western Vermont; (8) vein-related molybdenum, tin, tungsten, and (or) fluorite in late extensional faults of the eastern Adirondacks, New York, and in western New Hampshire; (9) porphyry-related molybdenum, tin, and (or) tungsten in western New Hampshire associated with Oliverian plutons (Ordovician), with Acadian (Devonian) plutons and pegmatites, and with plutons of Late Devonian (or Carboniferous) age; (10) skarn-type tungsten (and perhaps tin) peripheral to an Acadian(?) granitic pluton in central Vermont; (11) porphyry-related molybdenum, tin, and (or) tungsten associated with the Mesozoic alkaline intrusions at Cuttingsville, Vermont, and Pollard Hill, New Hampshire, and possibly with some Proterozoic felsic gneisses (metamorphosed granites) in the eastern Adirondacks, New York; (12) podiform chromite deposits in ultramafic bodies of central Vermont; (13) stratabound residual iron and (or) manganese deposits in areas of the early Paleozoic shelf sequence of western Vermont; (14) stratabound iron deposits in Proterozoic basement rocks of the eastern Adirondacks, New York, and in Late Proterozoic to Cambrian metasedimentary rocks of central Vermont; (15) uranium in epigenetic veins within Proterozoic basement rocks of the Green Mountains of Vermont; (16) uranium in pegmatites of Proterozoic age in the eastern Adirondacks of New York and the Green Mountains of Vermont and in granites of Devonian age in western New Hampshire; (17) uranium in surficial peat deposits of western New Hampshire and western Vermont; and (18) rare earth elements possibly associated with stratabound magnetite deposits of Proterozoic age in the eastern Adirondacks of New York.

INTRODUCTION

The Glens Falls 1° × 2° quadrangle, covering eastern New York, central Vermont, and western New Hampshire, has a long history of both metallic and nonmetallic mineral production (see Slack and Schruben, this volume). The assessment of mineral potential for the Glens Falls quadrangle presented here is focused exclusively on metallic resources. Economically, however, nonmetallic resources have been and probably will continue to be much more important. For example, pegmatites in New Hampshire (Cameron and others, 1954) and graphite deposits in New York (Alling, 1918) were historically valuable for many

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Manuscript approved for publication April 13, 1990.

[1] U.S. Geological Survey.

Table 1. Principal metallic mineral producers in the Glens Falls 1° × 2° quadrangle

Mine or district	Location (State)	Commodities produced[1]	Production (tons of ore)[2]
Elizabeth	Vermont	Cu, Zn, FeS, Ag, Au	3,200,000
Hammondville	New York	Fe	2,000,000
Ely	Vermont	Cu, FeS(?)	500,000
Fort Ann	New York	Fe	350,000
Crown Point	do.	Fe	100,000
Paradox Lake	do.	Fe	100,000
Skiff Mountain	do.	Fe	<10,000
Orange	Vermont	Cu	<10,000
Gove	do.	Cu	<10,000
Croydon	New Hampshire	Cu, Zn(?)	<10,000
Blood Mountain	Vermont	Cu	<10,000
Copperas Hill	do.	FeS, Cu(?)	<10,000
Oram (Lion Hill)	do.	Zn, Pb	<10,000
Holts Ledge	New Hampshire	W	<10,000
Braintree	Vermont	As	<10,000
Taggart	do.	Au	<10,000
Rooks	do.	Au	<10,000
Tyson Furnace	do.	Fe	<10,000
Brandon	do.	Fe, Mn	<10,000
Plymouth Union	do.	Fe	<10,000

[1] Metals listed in approximate decreasing order of commodities produced; FeS=pyrrhotite.
[2] Estimated.

years. In Vermont, talc, asbestos, and dimension stone (chiefly granite and marble) have been mined extensively and continue to supply a substantial part of the Nation's resources of these commodities (U.S. Bureau of Mines, 1985; Ratté and Ogden, 1989). Slate, garnet, sand and gravel, and stone aggregate also are commercially important locally in the quadrangle. None of these nonmetallic commodities (or fuels) is discussed further in this report.

Table 1 lists the principal metallic mineral producers in the Glens Falls quadrangle, ranked in decreasing order in terms of tons of ore produced. The largest past metal supplier in the quadrangle, and in fact in all of New England, is the Elizabeth mine in east-central Vermont, which produced an estimated 3.2 million tons of ore, mainly for copper and minor byproduct zinc, pyrrhotite, silver, and gold (Howard, 1969; Annis and others, 1983). Next in importance are the Hammondville iron mines in the eastern Adirondack Mountains of New York, which produced about 2 million tons of ore (Newland, 1908); because of the high grades of these iron deposits, they have in actuality yielded the greatest amount of metal. Other mines have had only minor production, chiefly for iron or base metals (table 1). The Oram (Lion Hill) mine near Brandon, Vt., produced some zinc and lead during the 1870's (Grant, 1968). Arsenic was mined from arsenopyrite-rich veins near Braintree, Vt. (Morrill and Chaffee, 1964; Grant, 1968), and minor amounts of tungsten were produced from quartz veins on Holts Ledge east of Lyme, N.H. (J.B. Thompson, Jr., Harvard University, written commun., 1987; Slack and others, 1987). The Taggart mine in the Plymouth-Bridgewater district of east-central Vermont (Perkins, 1904; Perry, 1929[?]), although very small, apparently is the largest mine in the quadrangle known to have produced gold as its principal commodity.

METHODS OF MINERAL-RESOURCE ASSESSMENT

In appraising the mineral resources of the Glens Falls quadrangle, qualitative, rather than quantitative, methods have been employed. Because of a lack of detailed grade and tonnage data for a variety of metal commodities and deposit types in the study area, several recognition criteria are used here to establish degrees of resource potential. Figure 1, modified slightly from Gair (1989), illustrates the methodology. This diagram is the foundation for the mineral-resource assessment of the Glens Falls quadrangle and is based on both diagnostic and permissive recognition criteria. Favorable geology and known mineral occurrences are considered to be diagnostic criteria, because they are required to establish resource potential and are present for essentially all mineral deposits of a given type. A geochemical anomaly, by contrast, is a permissive criterion, as it suggests the presence of a particular deposit but is not required; also, the absence of a geochemical anomaly does not necessarily preclude the possibility of a given deposit type. The presence of favorable geology or known mineral occurrence or any one of the permissive criteria alone yields only a low resource potential, shown in pink on the accompanying maps. A geochemical anomaly combined with either favorable geology or a known mineral occurrence—that is, a permissive and one diagnostic

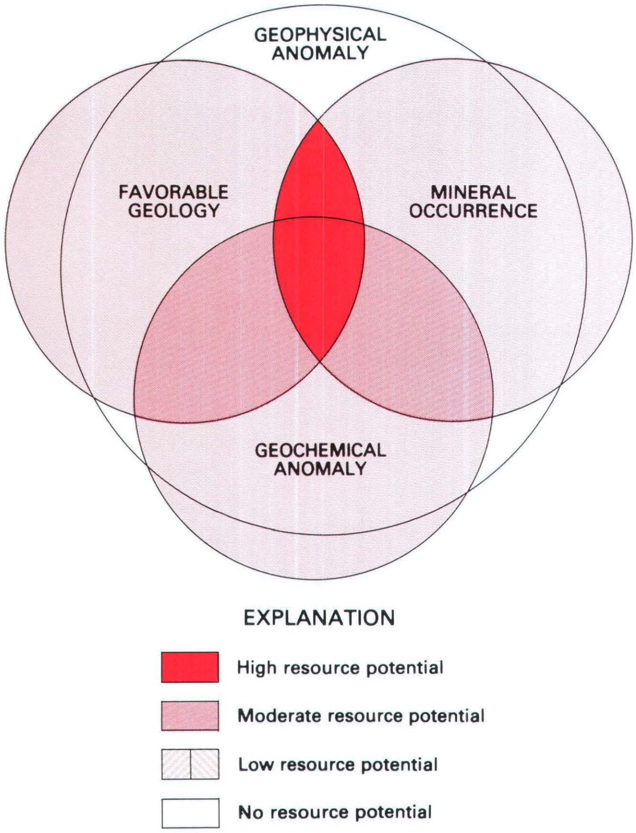

Figure 1. Methodology for estimating mineral-resource potential (from Gair, 1989). See text for description.

criterion—results in a moderate resource potential, shown in medium red (fig. 1). The combination of two diagnostic criteria such as favorable geology and known mineral occurrence yields a high potential, shown in dark red (fig. 1). A geophysical anomaly, although a permissive criterion, has a lower ranking in the evaluation scheme and, by itself, does not define a resource potential. Note especially that, in some figures, more than one type of pink pattern is used, to avoid ambiguity between nearby areas of low resource potential related to different deposit types.

In this study, the author designates areas of mineral potential in the Glens Falls quadrangle largely on the basis of appropriate ore deposit models (for example, Cox and Singer, 1986; Roberts and Sheahan, 1988). Watts (this volume) concludes that many of the geochemical anomalies in heavy-mineral (panned) concentrates of stream sediment in the quadrangle reflect linear belts of mineralization related to epigenetic (probably Mesozoic) hydrothermal processes. This interpretation is in some respects different from that of this author and implies the existence of many undiscovered vein deposits in the region. In this chapter, a resource potential for vein deposits is proposed in areas where vein deposits are known or where mapped faults coincide with panned concentrate anomalies. Where major faults or other field evidence of veins is absent, the bedrock source of the anomalies is considered uncertain and no delineation of vein-related mineral resources is made; this is especially true for base metals (copper, lead, zinc), which commonly occur in small veins that lack resource significance. Many of the regional geochemical anomalies of Watts (this volume) are likely derived from stratabound mineral deposits such as volcanogenic massive sulfides; where the geology is favorable for the existence of such stratabound deposits and (or) where mineral occurrences of this deposit type are known, a resource potential for stratabound deposits is assigned. More detailed work will be needed to document the sources of the panned concentrate geochemical anomalies in the quadrangle and evaluate the differing interpretations of Watts (this volume) and the author.

The resource assessment presented here is preliminary and intended only as a broad summary of the known and potential mineral deposits of the Glens Falls quadrangle. The assessment also is necessarily generalized for illustration at the page scale of this volume. A more thorough assessment is planned for publication in a series of 1:250,000-scale maps that integrate geochemical, geophysical, and remote sensing data with the distribution of mines, prospects, and mineral occurrences and with the results of new mapping studies currently underway in parts of the quadrangle.

In summarizing the mineral-resource assessment of the Glens Falls quadrangle, geochemically related metals are grouped into similar deposit types. The groups of mineral deposits and metals considered here are (1) volcanogenic copper-zinc; (2) sediment-hosted lead-zinc; (3) volcanogenic gold (±silver); (4) vein and porphyry-related gold; (5) tin, tungsten, molybdenum, and (or) fluorite; (6) chromium, nickel, vanadium, and platinum group elements; (7) iron, manganese, and titanium; and (8) uranium, thorium, and rare earth elements. The following discussion of resource assessment for each of the different groups of mineral deposits is linked to separate maps of the Glens Falls 1° × 2° quadrangle (figs. 2–9), each of which has a simplified geologic base (modified from Thompson and others, 1990). Areas of moderate and high mineral potential discussed in the text are keyed to locality numbers listed in table 2.

VOLCANOGENIC COPPER-ZINC (±LEAD, SILVER, GOLD)

Figure 2 shows the mineral-resource assessment for volcanogenic deposits of copper and zinc, with or without accompanying lead, silver, and gold. These types of deposits are chiefly stratabound massive sulfides, which may be hosted in either metasedimentary or metavolcanic rocks (for example, Franklin and others, 1981; Lydon, 1984). In western New Hampshire and easternmost Vermont, small

Table 2. Designated areas of moderate and high mineral-resource potential in the Glens Falls 1° × 2° quadrangle

[Locality numbers are identical to labels on individual maps of mineral-resource potential (figs. 2–9); criteria abbreviations: MO, mineral occurrence; FG, favorable geology; GCA, geochemical anomaly; GPA, geophysical anomaly]

Locality no.	Area	Deposit type	Criteria	Mineral potential
2–1	Blood Mountain mine, Vt. and N.H.	Kuroko-type massive sulfides	MO, FG	High.
2–2	Neal mine, N.H.	do.	do.	Do.
2–3	Croydon mine, N.H.	do.	do.	Do.
2–4	Ammonoosuc Volcanics, N.H. and Vt.	do.	GCA, FG	Moderate.
2–5	Barnard Volcanic Member of Missisquoi Formation,[1] Vt.	do.	do.	Do.
2–6	Ely mine, Vt.	Besshi-type massive sulfides	MO, FG, GCA	High.
2–7	Orange and Gove mines, Vt.	do.	do.	Do.
2–8	Elizabeth mine, Vt.	do.	do.	Do.
2–9	Gile Mountain Formation, Vt.	do.	GCA, FG	Moderate.
2–10	Stowe Formation, Vt.	do.	do.	Do.
2–11	Spathic mine, Vt.	do.	MO, FG	High.
2–12	Pinney Hollow Formation, Vt.	do.	GCA, FG	Moderate.
3–1	Oram (Lion Hill) mine, Vt.	Sedex-type lead-zinc	MO, FG	High.
3–2	Unnamed lead-zinc prospect, Vt.	do.	do.	Do.
3–3	do.	do.	do.	Do.
3–4	Unnamed lead-zinc occurrence, Saratoga Springs, N.Y.	do.	do.	Do.
3–5	Early Paleozoic shelf sequence, Vt.	do.	GCA, FG	Moderate.
3–6	Eastern Adirondack Mountains, N.Y.	do.	do.	Do.
3–7	Taconic allochthon, N.Y. and Vt.	do.	do.	Do.
4–1	Gold occurrence near Bethel, Vt.	Volcanogenic gold	MO, FG	High.
4–2	Gold occurrence near Bellows Falls, Vt.	do.	do.	Do.
4–3	Ammonoosuc Volcanics, N.H. and Vt.	do.	GCA, FG	Moderate.
4–4	Gile Mountain Formation, Vt.	do.	do.	Do.
4–5	Barnard Volcanic Member of Missisquoi Formation,[1] Vt.	do.	do.	Do.
4–6	Stowe Formation, Vt.	do.	do.	Do.
4–7	Adirondack Mountains gold occurrence near Hague, N.Y.	do.	do.	Do.
5–1	Taggart gold mine, Vt.	Metamorphogenic vein gold	MO, FG	High.
5–2	Rooks gold mine, Vt.	do.	do.	Do.
5–3	Unnamed gold prospect, Vt.	do.	do.	Do.
5–4	Early Paleozoic eastern Vermont sequence, Vt.	do.	GCA, FG	Moderate.
5–5	Altered ultramafic body, Chester dome, Vt.	Ultramafic-related gold	do.	Do.
5–6	Cuttingsville intrusive complex, Vt.	Porphyry-related gold[3]	MO, GCA, FG	High.
5–7	Northern extension of Cuttingsville complex,[2] Vt.	do.	GPA, FG	Moderate.
5–8	Sulfidic zones in Proterozoic augen gneiss, Vt.	do.	MO, FG	High.
5–9	Mount Colfax gold prospect, N.Y.	Stockwork vein gold	do.	Do.
5–10	Taconic allochthon, N.Y. and Vt.	do.	GCA, FG	Moderate.
6–1	Extensional faults, N.Y. and N.H.	Fault-related veins	do.	Do.
6–2	Proterozoic granitic gneiss, Adirondack Mountains, N.Y.	Porphyry-related(?) veins	do.	Do.
6–3	Scheelite-bearing veins, Mascoma dome, N.H.	do.	MO, FG	High.
6–4	Molybdenite-bearing veins, Lebanon dome, N.H.	do.	do.	Do.
6–5	Oliverian domes, N.H.	do.	GCA, FG	Moderate.
6–6	Bethlehem Gneiss, N.H.	do.	do.	Do.
6–7	Molybdenite- and cassiterite-bearing pegmatites, N.H.	Granitoid concentration	MO, GCA	Do.
6–8	Acadian(?) pluton, Vt.	Porphyry-related deposits	GCA, GPA, FG	Do.
6–9	Sunapee pluton, N.H.	do.	GCA, FG	Do.
6–10	Cuttingsville intrusive complex, Vt.	do.	MO, FG	High.
6–11	Pollard Hill intrusions, N.H.	do.	do.	Do.

Table 2. Designated areas of moderate and high mineral-resource potential in the Glens Falls 1° × 2° quadrangle—Continued

[Locality numbers are identical to labels on individual maps of mineral-resource potential (figs. 2–9); criteria abbreviations: MO, mineral occurrence; FG, favorable geology; GCA, geochemical anomaly; GPA, geophysical anomaly]

Locality no.	Area	Deposit type	Criteria	Mineral potential
7–1	Rochester verde antique quarry, Vt.	Podiform chromite	MO, FG	High.
7–2	Hinchman prospect, Vt.	Epigenetic vein platinum	GCA, FG	Moderate.
8–1	Brandon mine, Vt.	Residual iron and manganese	MO, FG	High.
8–2	Unnamed iron-manganese mine, Vt.	do.	do.	Do.
8–3	do.	do.	do.	Do.
8–4	do.	do.	do.	Do.
8–5	do.	do.	do.	Do.
8–6	do.	do.	do.	Do.
8–7	do.	do.	do.	Do.
8–8	Tyson mine area, Vt.	Sedimentary iron oxides	do.	Do.
8–9	Plymouth Union iron mine, Vt.	do.	do.	Do.
8–10	Spathic iron mine, Vt.	Volcanogenic siderite	do.	Do.
8–11	Hammondville iron mines, N.Y.	Sedimentary iron oxides	do.	Do.
8–12	Fort Ann iron mines, N.Y.	do.	do.	Do.
8–13	Crown Point iron mines, N.Y.	do.	do.	Do.
8–14	Paradox Lake iron mines, N.Y.	do.	do.	Do.
8–15	Skiff Mountain iron mine, N.Y.	do.	do.	Do.
8–16	Moose Mountain titanium mine, N.Y.	Orthomagmatic ilmenite	do.	Do.
9–1	Ludlow Mountain uranium prospects, Vt.	Epigenetic vein uranium	MO, FG	High.
9–2	East Jamaica uranium prospects, Vt.	do.	do.	Do.
9–3	West Jamaica uranium prospects, Vt.	do.	do.	Do.
9–4	Uranium-rich pegmatite dike, Weston, Vt.	Granite-related uranium	GCA, FG	Moderate.
9–5	I-89 uranium deposit, N.H.	do.	MO,FG	High.
9–6	Rte. 11 uranium occurrence, N.H.	do.	GCA,FG	Moderate.
9–7	Uraniferous peat, near Sunapee, N.H.	Surficial ground-water concentration	MO, FG	High
9–8	Uraniferous peat, near Brandon, Vt.	do.	do.	Do.
9–9	Uraniferous pegmatite, Spar Bed Hill quarry, N.Y.	Granite-related uranium	do.	Do.
9–10	Uraniferous pegmatites, near Brant Lake, N.Y.	do.	do.	Do.
9–11	Uraniferous pegmatite, Fort Ann quarry, N.Y.	do.	do.	Do.
9–12	Uraniferous pegmatite, Overlook quarry, N.Y.	do.	do.	Do.
9–13	Radioactive iron deposits, N.Y.	Sedimentary iron oxides	GCA, FG	Moderate.
9–14	do.	do.	do.	Do.

[1] As used by Doll and others (1961).

[2] Of Chapman (1968).

[3] Includes several gold-bearing deposit types (see Robinson, this volume).

deposits of this type are known in the Ordovician Ammonoosuc Volcanics and equivalent units (for example, Post Pond Volcanic Member of the Orfordville Formation), such as at the Blood Mountain (Waterman) mine, the Neal mine, and the Croydon mine (fig. 2, nos. 2–1 to 2–3). Although different in some respects, these deposits are all predominantly volcanic hosted and contain copper as the major metallic commodity, variable zinc, and only minor amounts of lead, silver, and gold; they may be broadly classified as Kuroko-type, after the well-known Miocene deposits of Japan (Ohmoto and Skinner, 1983). The entire belt of Ammonoosuc Volcanics is outlined in pink because it comprises a favorable geologic setting, and therefore a low resource potential, for Kuroko-type massive sulfide deposits. Areas in the Ammonoosuc volcanic belt that have significant geochemical anomalies in heavy-mineral concentrates of stream sediments (Watts, this volume) are shown in medium red and are designated as having moderate resource potential (fig. 2, no. 2–4). A similar rationale is used for the metavolcanic rocks of the Barnard Volcanic Member of the Missisquoi Formation (as used by Doll and others, 1961) of early Paleozoic age (see Aleinikoff and Karabinos, this volume), also shown in pink to the west of the Ammonoosuc belt in eastern Vermont. This resource potential is based on favorable geology that includes the presence of volcanogenic massive sulfides in probable correlative metavolcanic rocks of the Hawley Formation at the Davis copper-zinc deposit in northwestern Massachusetts (Slack and others, 1983) and at the Leadville zinc-lead-silver deposit along the Quebec-Vermont border

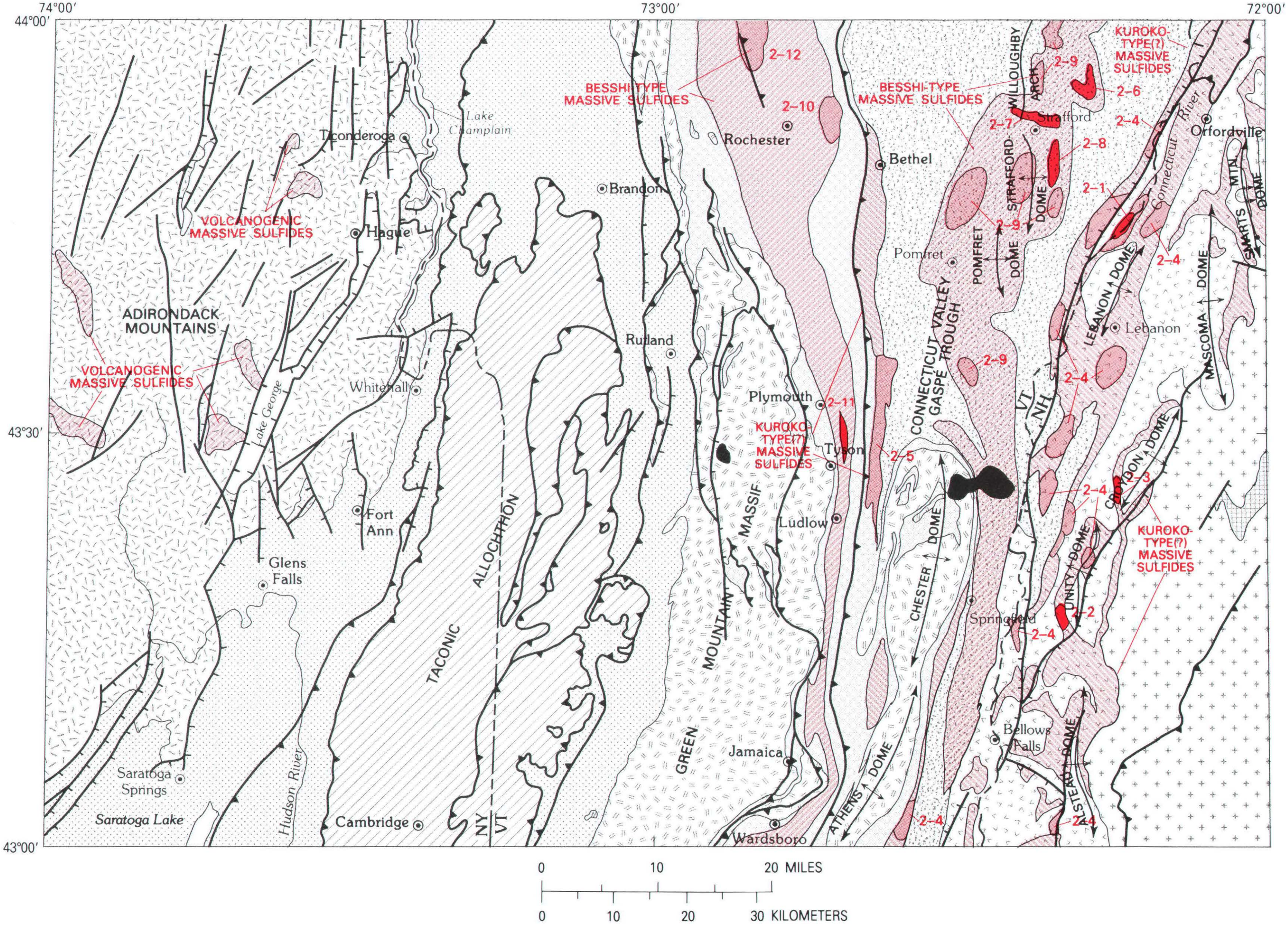

Figure 2. Mineral-resource potential for volcanogenic deposits of copper and zinc (±lead, silver, gold). See table 2 for explanation of locality numbers for high (dark red), moderate (medium red), and low (pink) areas of mineral-resource potential; note that different lined patterns for areas of low resource potential shown in pink refer to different deposit types. See figure 1 for explanation of resource potential patterns.

EXPLANATION

CRETACEOUS AND JURASSIC (?)	▩	White Mountain Plutonic-Volcanic Suite (syenite, granite, and gabbro)
UPPER DEVONIAN	▦	Granite plutons (two-mica granite)
LOWER DEVONIAN AND SILURIAN	░	Merrimack terrane (granitic gneiss, migmatite, minor pelitic schist, and quartzite)
	░	Connecticut Valley-Gaspé terrane (pelitic schist, calcareous schist, and minor amphibolite)
ORDOVICIAN AND CAMBRIAN	░	Parautochthonous shelf sequence (limestone, dolomite, and quartzite)
DEVONIAN, SILURIAN, AND ORDOVICIAN	░	Bronson Hill terrane (metavolcanic rocks, pelitic schist, and granitic gneiss)
ORDOVICIAN, CAMBRIAN, AND PROTEROZOIC Z	▨	Taconic allochthon (slate, graywacke, and minor carbonate)
	□	Eastern Green Mountain terrane (pelitic schist, metagraywacke, and minor amphibolite)
PROTEROZOIC Y	░	Adirondack massif (felsic gneiss, quartzite, anorthosite, amphibolite, calc-silicate rocks, and metagabbro)
	░	Green Mountain massif[1] (felsic gneiss, quartzite, minor amphibolite, and calc-silicate rocks)

High-angle fault—Tick marks on downthrown side where known

Thrust fault—Sawteeth on upper plate

[1] Includes Lincoln Mountain massif and Proterozoic rocks of the Chester and Athens domes.

Figure 2. Continued.

(Gauthier and others, 1985a, 1989). The presence of geochemical anomalies in panned concentrates of stream sediments surrounding the southern portion of the Barnard volcanic belt and the favorable geologic setting there indicate a moderate resource potential for this deposit type (fig. 2, no. 2–5).

Several other volcanosedimentary belts in eastern Vermont (pink areas) are designated as having geologic settings favorable for the occurrence of predominantly sediment-hosted massive sulfide deposits that contain copper and zinc (±silver and gold). Such deposits may be classified as Besshi-type, after the Permian and Mesozoic examples in Japan (for example, Fox, 1984; Slack and Shanks, 1989). The easternmost and largest of the favorable belts is comprised of portions of the Gile Mountain and Waits River Formations and the Standing Pond Volcanics and hosts the stratabound massive sulfide deposits of the Orange County copper district (White and Eric, 1944; Annis and others, 1983), including those at the Ely, Orange and Gove, and Elizabeth mines. In addition to the mineral deposits, this belt contains significant geochemical anomalies in panned concentrates (Slack and others, 1984; Watts, this volume; Slack and others, this volume) and therefore constitutes an area of high resource potential (fig. 2, nos. 2–6 to 2–8, respectively); areas within this belt that have panned concentrate anomalies but no known mineral occurrences are assigned moderate potential for Besshi-type massive sulfide deposits (fig. 2, no. 2–9). To the west, in the north-central part of the quadrangle, is a sequence of Upper Cambrian to Lower Ordovician clastic metasedimentary and minor mafic metavolcanic rocks that compose the Stowe Formation. This sequence is important because of the occurrence in the Stowe Formation of a small cupriferous massive sulfide deposit at the Udall mine east of Morrisville, Vt., in the Lake Champlain 1° × 2° quadrangle (White and others, 1946; J.F. Slack, unpub. data). The presence of a massive sulfide body at the Udall mine suggests that similar mineralization may have taken place elsewhere in the Stowe Formation, thus defining a favorable geologic setting for the occurrence of volcanogenic massive sulfides within this formation in the Glens Falls quadrangle. This favorable setting, and the distribution of geochemical anomalies recognized by Watts (this volume), defines an area of moderate resource potential for this deposit type (fig. 2, no. 2–10).

A potential for volcanogenic massive sulfides has been recognized also in a belt of pelitic metasedimentary rocks and minor greenstones (mafic metavolcanic rocks) to the south, along the eastern margin of the Green Mountain massif. This belt contains the Pinney Hollow Formation and the possibly correlative Mount Abraham Schist of Late Proterozoic(?) or Cambrian(?) age to the north (see Doll and others, 1961; Stanley and others, 1987). Volcanogenic-type mineralization within (or near the contact of) the Pinney Hollow Formation is present at the site of the small Spathic iron mine on Weaver Hill, about 2.7 km northeast of the village of Tyson, Vt. (fig. 2, no. 2–11). The Spathic deposit (Slack and others, 1987; J.F. Slack, unpub. data) consists of stratabound (and apparently stratiform) lenses of massive siderite (iron carbonate) associated with trace to minor (<10 percent) pyrrhotite and chalcopyrite. Some dump samples at the mine contain radiating fibers of pale brown magnesian amphiboles (anthophyllite and (or) gedrite). The presence of these amphibole minerals, which commonly are associated with metamorphosed massive sulfide sequences throughout the world (see Beeson, 1988), indicates that this is not an insignificant iron deposit and represents part of an ancient, submarine hydrothermal system. Accordingly, a high potential for volcanogenic massive sulfide deposits is assigned to the Pinney Hollow Formation at and near the Spathic mine (fig. 2, no. 2–11). In the north-central part of the quadrangle, a moderate resource potential for such deposits is assigned on the basis of panned concentrate anomalies identified by Watts (this volume) in a greenstone-rich portion of the Pinney Hollow Formation (fig. 2, no. 2–12). The remainder of the Pinney Hollow and the Mount Abraham Schist are assigned a low resource potential for massive sulfide deposits (pink areas), on the basis of correlative stratigraphic relations. Lithologic associations and a variety of geochemical anomalies in the Mount Abraham Schist within the Bread Loaf Wilderness, in the Lake Champlain 1° × 2° quadrangle just a few kilometers north of the Glens Falls quadrangle, also suggest a potential for massive sulfide deposits within this belt (Slack and Bitar, 1983).

In the eastern Adirondacks of New York, a speculative low resource potential is assigned to areas that contain abundant layered amphibolite within metamorphic sequences of probable marine sedimentary origin. No regional geochemical anomalies or mineral occurrences are known in these sequences, but the geologic setting is considered favorable for volcanogenic massive sulfide deposits, on the basis of the occurrence of polymetallic massive sulfides in lithologically similar rocks of Grenville age at the Montauban mine in eastern Quebec (Stamatelopoulou-Seymour and MacLean, 1984; Bernier and others, 1987). More work is needed to better evaluate the massive sulfide potential of this region.

SEDIMENT-HOSTED LEAD-ZINC (±COPPER, SILVER, BARITE)

Figure 3 illustrates the mineral-resource assessment for sediment-hosted lead and zinc deposits, which may contain important amounts of associated copper, silver, and (or) barite. Shown in pink as having low resource potential is the Cambrian and Ordovician shelf sequence that consists largely of shallow-water carbonate and siliciclastic rocks (for example, dolomite, quartzite). This sequence contains several lead-zinc deposits and occurrences, the largest of which is the Oram deposit at Lion Hill near Brandon, Vt. (fig. 3, no. 3–1); three smaller occurrences are known to the west, south, and southwest within the same shelf sequence (fig. 3, nos. 3–2 to 3–4). The Oram (Lion Hill) deposit has textural, lithologic, and fluid-inclusion features suggestive of a partly sedimentary-exhalative (sedex) origin (Clark, this volume; Foley, this volume) and is similar to some of the carbonate-hosted, lead-zinc deposits of Ireland that are stratabound and locally stratiform (for example, Hitzman and Large, 1986). Similar analogies have been made by Gauthier and others (1989) for stratabound lead-zinc deposits in the Cambrian Dunham Dolomite of southeastern Quebec. The interpretation of a partly syngenetic origin for the Lion Hill mineralization (Clark and Slack, 1988; Clark, this volume) has major resource implications because the base-metal grades of these types of deposits commonly are very high, relative to those of classic Mississippi Valley-type (MVT) lead-zinc ores, which also are stratabound but not stratiform. In addition, most sediment-hosted, stratiform lead-zinc deposits contain significant amounts of silver, in contrast to the silver-poor MVT ores. Areas within the shelf sequence that have associated geochemical anomalies in panned concentrates of stream sediments (Watts, this volume) are thus considered significant and are designated as having moderate resource potential for this deposit type (fig. 3, no. 3–5).

On the basis of ore deposit models, the Early Cambrian Cheshire Quartzite within the shelf sequence may have a potential for stratabound lead and (or) zinc. Such models are based on the characteristics of sandstone-hosted, lead-zinc deposits throughout the world (Bjorlykke and Sangster, 1981), including those of the large Laisvall lead deposit in northern Sweden (Rickard and others, 1979). The applicability of these ore deposit models to the New England region has been strengthened by the recognition in the Appalachian Mountains of Paleozoic sandstones that are cemented by lead and (or) zinc, including the Silurian Tuscarora Quartzite in the central Appalachians (J.R. Craig, oral commun., 1980) and quartzites of Cambrian and Silurian age in southeastern Quebec and the Gaspé Peninsula (Schrijver and Beaudoin, 1987). Similar lead±zinc deposits may exist within Paleozoic sandstones of the Glens Falls quadrangle, such as the Cambrian Cheshire Quartzite in Vermont and the Silurian Clough Quartzite in New Hampshire. No stratabound disseminations of base metals are known in these quartzite units, however, and geochemical anomalies recognized in associated panned concentrates of stream sediment (Watts, this volume) cannot be unequivocally assigned to these specific stratigraphic units. Following the mineral-resource evaluation of the Cheshire Quartzite in the Bristol Cliffs Wilderness by Slack and Mory (1983), presently available data suggest that no lead or zinc resource can be documented in such rocks within the Glens Falls quadrangle. Accordingly, no areas of resource potential for this deposit type are shown on figure 3. Geologic environments nevertheless may have been favorable for the formation of sandstone-hosted, lead-zinc deposits within the Glens Falls quadrangle, and future exploration in the region directed specifically at this deposit type may be warranted.

In the eastern Adirondacks, a sequence of highly deformed and metamorphosed carbonate and calc-silicate rocks is assigned a low resource potential for Balmat-type

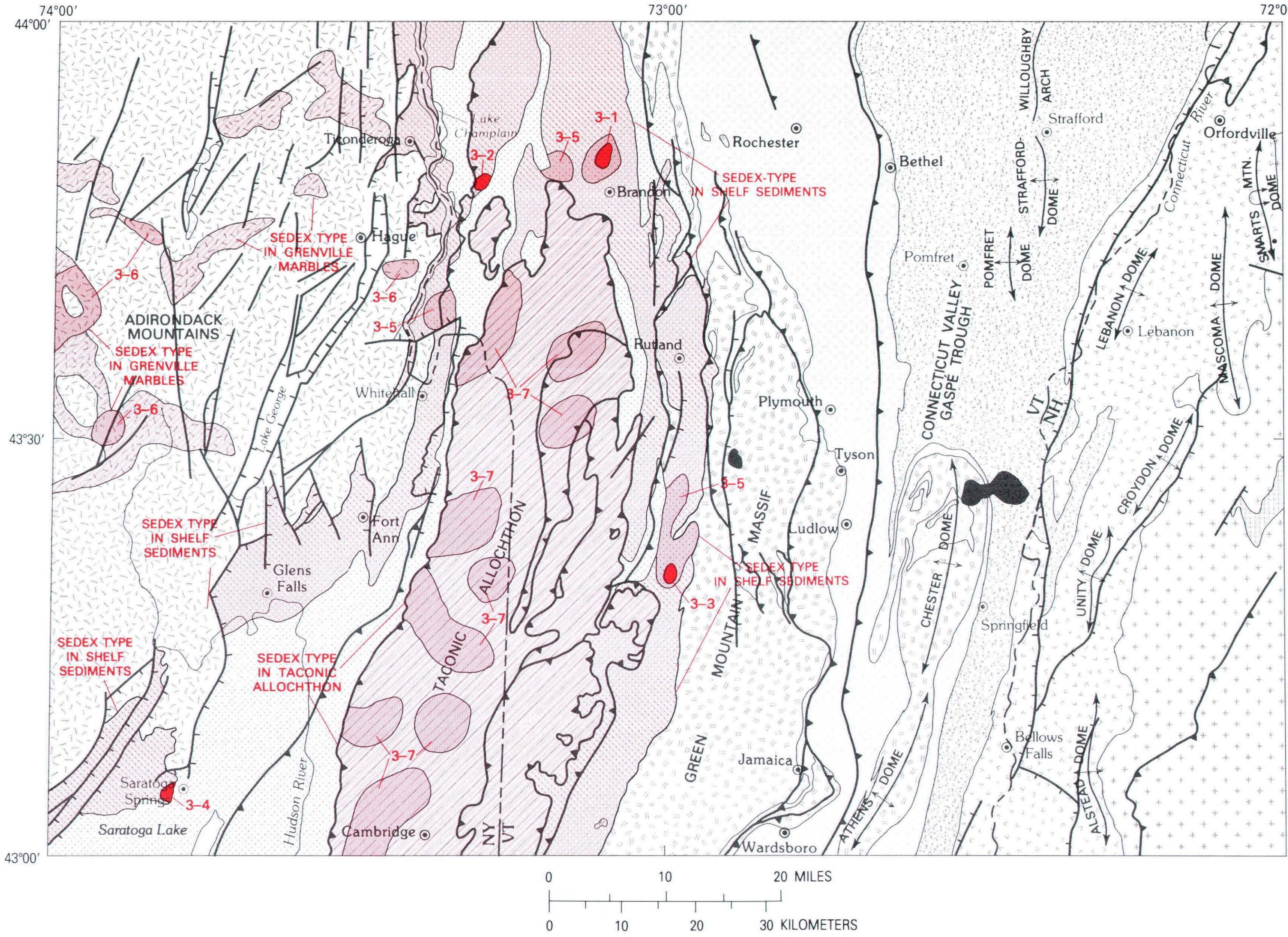

Figure 3. Mineral-resource potential for sediment-hosted deposits of lead and zinc (±copper, silver, barite). See table 2 for explanation of locality numbers for high (dark red), moderate (medium red), and low (pink) areas of mineral-resource potential; note that different lined patterns for areas of low resource potential shown in pink refer to different deposit types. See figure 1 for explanation of resource potential patterns. See figure 2 for explanation of geologic map units and symbols.

massive sulfide deposits. The Balmat ores may be metamorphosed analogs of the stratiform, sediment-hosted class of lead-zinc deposits (deLorraine and Dill, 1982; Whelan and others, 1984), such as those within the Carboniferous carbonate rocks of central Ireland (for example, Hitzman and Large, 1986). The assessment of moderate potential for some areas (fig. 3, no. 3–6) is based on panned concentrate geochemical anomalies, principally for lead, within these belts of marble (see Watts, this volume).

A deposit type that has newly recognized potential in the Glens Falls quadrangle is sedimentary-exhalative or sedex-type polymetallic sulfides in rocks of the Taconic allochthon (Slack and Watts, 1989; Watts, this volume). The geologic setting of the Taconic allochthon is similar to that of other parts of the world that contain major sedex deposits, such as the Selwyn basin in the Canadian Cordillera (Carne and Cathro, 1982; Abbott and others, 1986). Although no bedrock occurrences of this deposit type are known within the quadrangle, one small stratabound lead-zinc deposit is present in Taconic rocks near White Creek, N.Y., just 2 km south of the southern border of the Glens Falls quadrangle (Slack and others, 1987; J.F. Slack, unpub. data). The White Creek deposit consists of abundant sphalerite and galena within a vitreous quartzite unit and local conglomerate lens that compose the Mudd Pond Quartzite Member of the Nassau Formation (see Potter, 1972), which is now considered to be part of the Browns Pond Formation (Rowley and others, 1979). The White Creek sulfides show microscopic features (including deformed sphalerite) that indicate a pre-Taconic age of mineralization. This deposit, although very small, is analogous in many respects to several sediment-hosted, lead-zinc-silver deposits in the Selwyn basin, particularly the Tom and Jason deposits in the MacMillan Pass area of the Yukon (see Bailes and others, 1986; McClay and Bidwell, 1986).

Two stratigraphic sequences appear to be especially favorable hosts for sedex-type lead-zinc-silver deposits in the Taconic allochthon. One of these is the Cambrian Browns Pond Formation, which contains the Mudd Pond Quartzite and the stratabound White Creek deposit, as well as intraformational carbonate conglomerates that, in places, have clasts of massive pyritic sulfide (Slack and Watts, 1989); one of these clasts contains 2.5 percent zinc (K.C. Watts, Jr., oral commun., 1989). The second favorable sequence is the Mount Merino Formation of Middle Ordovician age that comprises dark chert and carbonaceous, locally sulfidic (and silicified), slate (for example, Rowley and others, 1979). This formation also has rare occurrences of pyritic massive sulfide clasts within intraformational conglomerates, such as near Hudson, N.Y., in the Albany 1° × 2° quadrangle (J.F. Slack and K.C. Watts, Jr., unpub. data); equivalent rocks along strike to the northeast in southeastern Quebec (Beauceville Formation) contain a sedex-type zinc-lead-copper deposit (the Champagne prospect) that has significant associated gold and silver (Gauthier and others, 1987, 1989; Godue, 1988). Because of structural complexities, a detailed assessment for sedex deposits in the Taconic allochthon is not possible without more extensive geologic and geochemical data. Nevertheless, some general resource interpretations can be made. Because of the overall geologic similarity between the Taconic allochthon and the Selwyn basin, the entire region of the allochthon is considered to have a low resource potential for this deposit type. Areas that have significant lead, zinc, copper, and barium geochemical anomalies in the Taconics (Day and others, 1986; Watts, this volume) are assigned a moderate potential (fig. 3, no. 3–7).

VOLCANOGENIC GOLD (± SILVER)

Figure 4 shows the mineral-resource assessment for volcanogenic gold deposits. In the eastern part of the map area, in western New Hampshire and eastern Vermont, several belts shown in pink designate areas of low resource potential for deposits of stratabound volcanic-hosted gold ±silver. Such speculative precious-metal deposits may have formed by volcanogenic processes either syngenetically from a submarine hydrothermal system (as exhalative chemical precipitates) or as epigenetic veins and (or) replacements in the shallow parts of volcanic arcs (either in submarine or subaerial settings); review articles on such volcanogenic, typically epithermal, gold deposits include those by Hayba and others (1985) and Bonham (1988). The belts shown on figure 4 in pink outline the distribution of favorable volcanic and volcaniclastic units of early Paleozoic age. Within these belts, dark red areas of high potential are designated on the basis of anomalous gold in rock samples. A pyritic silicified metarhyolite (fig 4, no. 4–1) from the Barnard Volcanic Member south of Bethel, Vt. (J.F. Slack, unpub. data), that contains 0.1 ppm gold, was collected near a known gold prospect and in the vicinity of several gold- and silver-rich panned concentrate localities (Day and others, 1986; Watts, this volume). The dark red area to the south (fig. 4, no. 4–2) reflects the presence of an altered pyritic felsite near Bellows Falls, Vt., that contains 1.0 ppm gold (Watts, this volume). Areas within the volcanic-associated belts shown in medium red (fig. 4, nos. 4–3 to 4–6) contain high gold and (or) arsenic in panned concentrates (Day and others, 1986; Watts, this volume) and are designated as having moderate potential for this deposit type.

In the eastern Adirondack Mountains, two belts of rocks are assigned a low potential for volcanogenic gold (fig. 4). One outlines areas of layered amphibolite associated with marine metasedimentary rocks. The rocks of this belt have broad similarities to the Grenville rocks north of Montreal that host the gold ores of the Montauban mine described by Gauthier and others (1985b) and Jourdain and

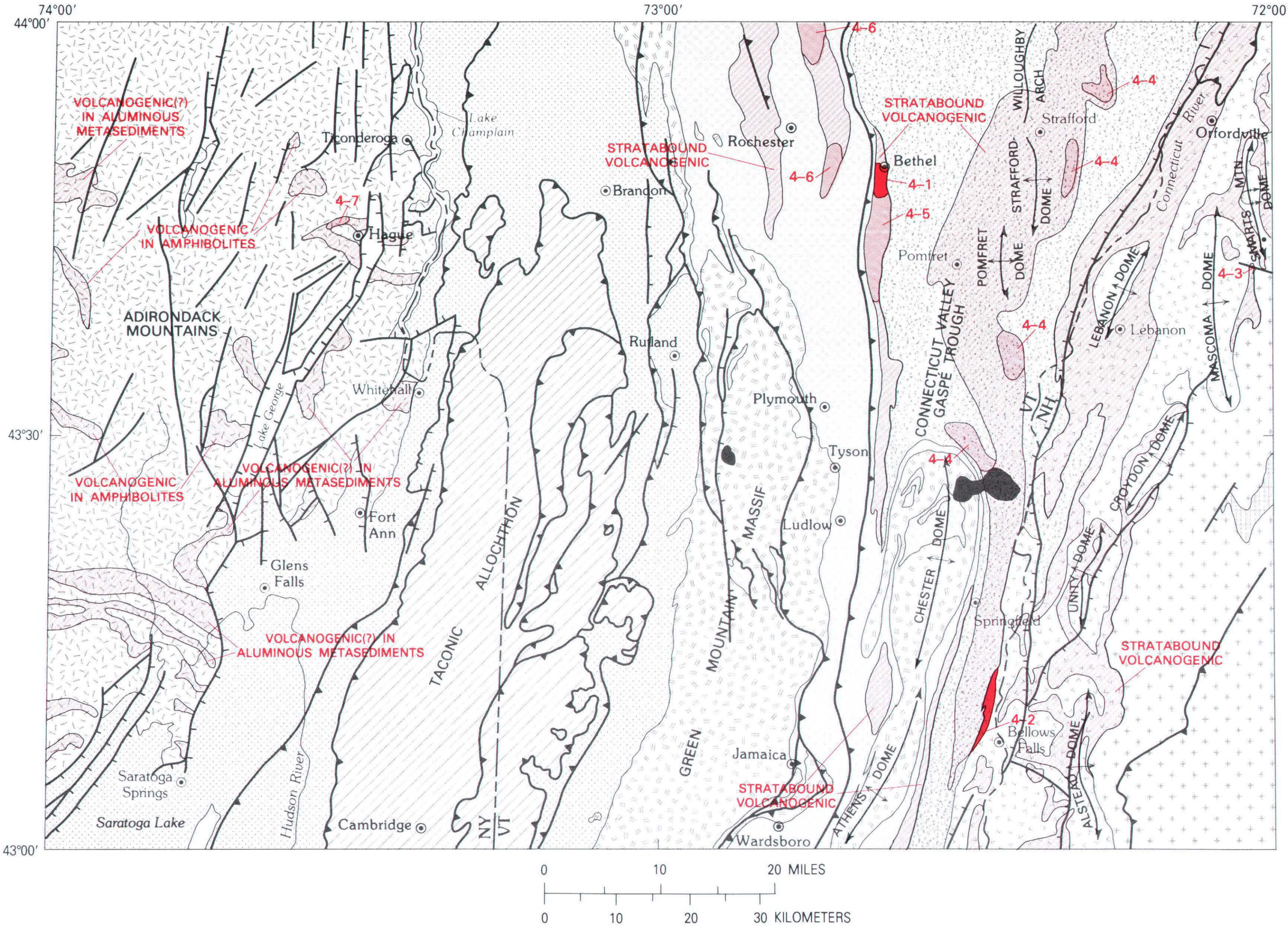

Figure 4. Mineral-resource potential for volcanogenic deposits of gold (±silver). See table 2 for explanation of locality numbers for high (dark red), moderate (medium red), and low (pink) areas of mineral-resource potential; note that different lined patterns for areas of low resource potential shown in pink refer to different deposit types. See figure 1 for explanation of resource potential patterns. See figure 2 for explanation of geologic map units and symbols.

others (1987). The Montauban gold deposit is interpreted to have formed by exhalative processes contemporaneously with local volcanism (Bernier and others, 1987). Although no gold anomalies (in panned concentrates) from this Adirondack belt are known (Day and others, 1986), detailed rock sampling has not been done, and so a resource potential for gold in the amphibolites and surrounding metasedimentary rocks cannot be ruled out.

A second belt of rocks in the Adirondacks comprises a variety of lithologies, including aluminous gneiss, graphitic and sulfidic schist, and minor carbonate or calc-silicate gneiss (see McLelland, this volume). This belt is well known for its stratabound graphite deposits (Alling, 1918), many of which are associated with thin units of very pyritic schist (Dixon Schist of Alling, 1918); sparse amounts of galena, sphalerite, and gahnite (zinc spinel) are present in a few of the deposits (Alling, 1918). Reconnaissance rock sampling of this belt in two areas (Slack and others, 1987; J.M. McLelland and J.F. Slack, unpub. data) reveals minor geochemical anomalies, including 0.17 ppm gold in a graphitic sillimanite-garnet gneiss (khondalite) just west of Hague, N.Y. (fig. 4, no. 4–7). This belt of rocks in the Adirondacks is lithologically similar to that surrounding the Enåsen gold mine in central Sweden, where disseminated gold occurs in a stratabound zone within Proterozoic sillimanite quartzite associated with graphitic and pyritic schist, calc-silicate gneiss, and sillimanite-bearing metapelitic gneiss (Nysten and Annersten, 1985; Willdén, 1986). The anomalous gold value in the rock sample from the Adirondack belt and analogy with the deposit in Sweden suggest that this group of Adirondack rocks also may have a moderate resource potential for stratabound gold, possibly of volcanogenic origin.

VEIN AND PORPHYRY-RELATED GOLD (±SILVER)

On the eastern side of the Green Mountains is a region that contains many gold-bearing quartz veins (Hager, 1861; Perkins, 1904; Perry, 1929[?]; Smith, 1976). The most significant of these veins and their surrounding country rocks are shown on figure 5 as having high resource potential (nos. 5–1 to 5–3). These veins are inferred to be metamorphogenic types of gold deposits, in which metamorphically derived fluids deposited quartz, carbonate, gold, and some base metals in fractures and faults within various metasedimentary (and some metavolcanic) rocks of the eastern Vermont sequence (for example, Pinney Hollow, Ottauquechee, and Missisquoi Formations). From a current resource standpoint, these veins probably have minimal importance because of their small size and apparent low grades (for example, the Taggart mine, see Slack and others, 1987). Many heavy-mineral concentrates of stream sediment in this area contain anomalous amounts of gold and (or) arsenic (Day and others, 1986; Watts, this volume), but the bedrock source of these metals has not yet been established. Accordingly, a broad zone that contains the known gold-bearing veins and the panned concentrate anomalies is designated as having a moderate potential for vein-type metamorphogenic gold (fig. 5, no. 5–4). Placer gold is documented in this general area also (Smith, 1976; Henson, 1982) but like the veins probably has limited economic significance. A resource may exist, though, for gold-bearing paleoplacers in Tertiary terrace gravels, such as those northeast of the town of Tyson, Vt. This possibility has not been thoroughly evaluated during the Glens Falls CUSMAP project.

A potential for gold also may exist in the same general region in association with altered ultramafic bodies. This interpretation is based on the recently discovered gold mineralization related to ultramafic rocks in the Baie Verte peninsula of northwestern Newfoundland (Tuach, 1987; Gower and others, 1988) and on the occurrence models outlined by Buisson and Leblanc (1985, 1986). A reconnaissance study of the gold content of such rocks in the Glens Falls quadrangle and to the north in the Lake Champlain 1° × 2° quadrangle has recently been made (J.F. Slack and B.R. Lipin, unpub. data). This study failed to identify any significant gold concentrations, but a few of the altered ultramafic bodies in the North Troy and Belvidere Mountain, Vt., areas (Lake Champlain quadrangle) contain elevated values (0.1–0.3 ppm) of gold. Some of these bodies in the ultramafic belt of the Lake Champlain quadrangle (see Doll and others, 1961) have relatively high concentrations of arsenic (to 1,200 ppm), antimony (one sample has 92 ppm), cobalt (to 150 ppm), tungsten (one sample has 21 ppm), boron (to 99 ppm), and barium (to 94 ppm). According to Buisson and Leblanc (1985, 1986), this suite of metals is characteristic of gold deposits associated with carbonatized ultramafic rocks. Only one sample analyzed from the Glens Falls quadrangle has elevated metal contents, a talc schist from the core of the Chester dome northwest of Springfield, Vt., that contains 94 ppm barium and 290 ppm copper; this ultramafic body is judged to have moderate resource potential for ultramafic-related gold (fig. 5, no. 5–5). Other areas of ultramafic rock in the Glens Falls quadrangle (Doll and others, 1961; Moench, this volume, fig. 2) are considered to be favorable and are assigned a low resource potential for gold of this deposit type. More detailed field and analytical work will be necessary to fully evaluate the gold potential of these rocks in the region.

In the eastern part of the Glens Falls quadrangle are three areas that have potential for porphyry-related gold (fig. 5). These areas represent the surface outcrop and surrounding country rock of the alkaline intrusive complexes of the Jurassic and Cretaceous White Mountain Plutonic-Volcanic Suite at Cuttingsville and Mount Ascutney, Vt., and at Pollard Hill, N.H. These three areas are assigned a low resource potential on the basis of favorable

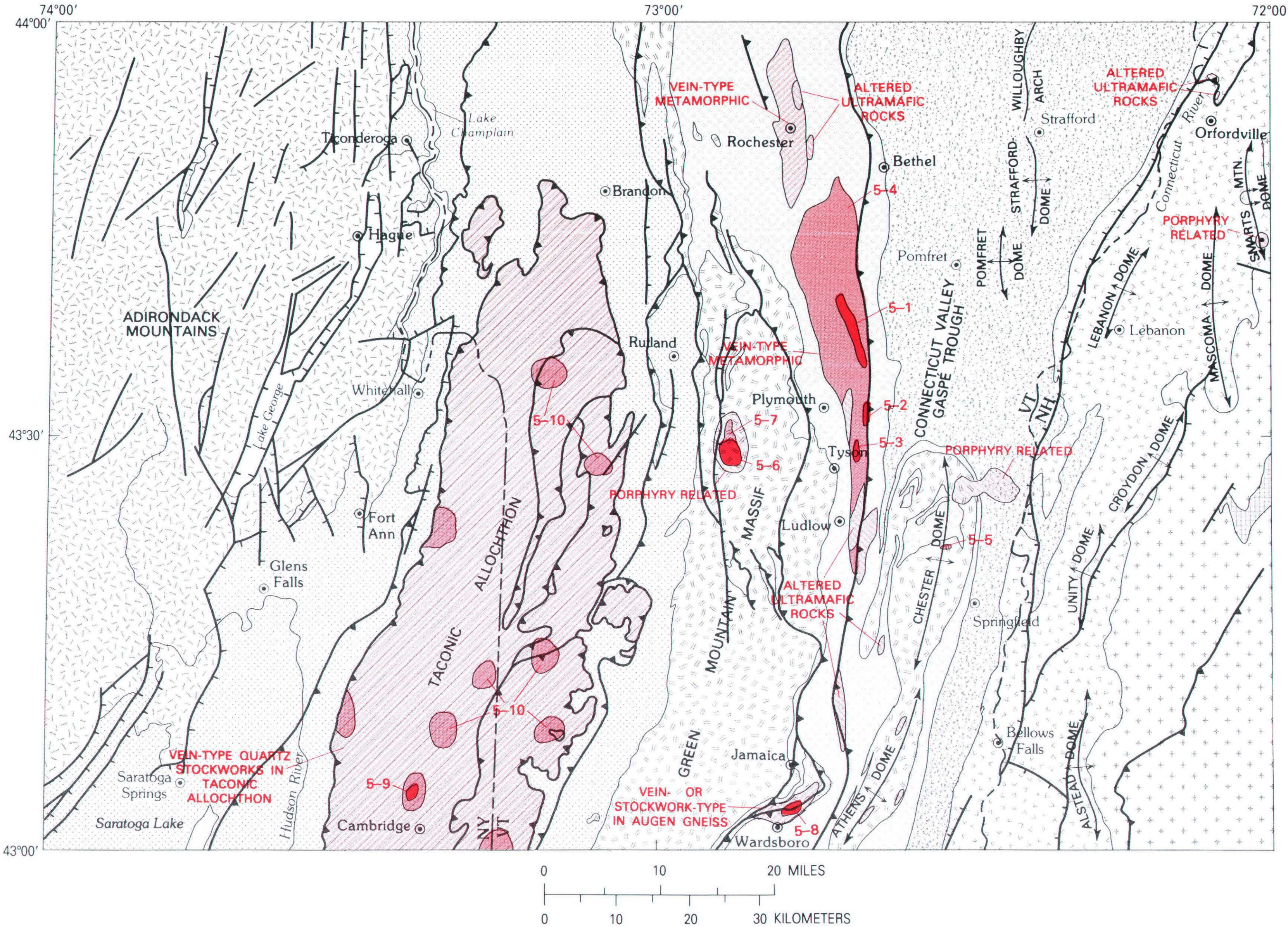

Figure 5. Mineral-resource potential for vein and porphyry-related deposits of gold (±silver). See table 2 for explanation of locality numbers for high (dark red), moderate (medium red), and low (pink) areas of mineral-resource potential; note that different lined patterns for areas of low resource potential shown in pink refer to different deposit types. See figure 1 for explanation of resource potential patterns. See figure 2 for explanation of geologic map units and symbols.

geology (pink), but the Cuttingsville intrusive complex (of Chapman, 1968) additionally has a high potential (no. 5–6 in dark red) because of the presence of bedrock occurrences of gold. These occurrences include a gold-rich pyrrhotite replacement deposit in Precambrian marble, a gold-bearing, quartz-pyrite stockwork in pulaskite syenite, and gold concentrations in altered syenite near a small molybdenum prospect (Robinson, this volume). An area of moderate resource potential for gold (medium red) is shown to the north of the outcrop area of the Cuttingsville complex (fig. 5, no. 5–7) on the basis of geophysical studies that suggest the presence of intrusive rocks there at shallow depth (see Daniels, this volume).

Augen gneisses and adjacent rocks of Proterozoic age within the Green Mountain massif also may have a potential for gold. This assessment is based on recent studies of the White Bay area of northwestern Newfoundland, where concentrations of gold have been found in association with sheared and altered megacrystic granitic rocks of Late Proterozoic age along the eastern edge of the Long Range inlier (Tuach and French, 1986; Tuach, 1987). The occurrences of gold in the White Bay area, some of which are believed to be of Late Proterozoic age (Tuach, 1987), suggest that similar megacrystic augen gneisses in Vermont, which probably are metamorphosed granites, constitute a favorable geologic setting for the occurrence of porphyry-related gold. This hypothesis is strengthened by the discovery of pyrite-rich zones and geochemical anomalies in some of the Vermont augen gneisses (J.F. Slack and Paul Karabinos, unpub. data). One of the most interesting localities is along Wardsboro Brook, 2.2 km northeast of Wardsboro, Vt., where outcrops of sheared augen gneiss locally contain significant amounts of pyrite and minor arsenopyrite in irregular veins and disseminations. The sulfide occurrences are not within late (posttectonic) structures and appear to be deformed by Paleozoic orogenic events; however, the proximity of major thrust faults (see Ratcliffe and others, 1988, p. 135) suggests that the formation of the sulfides may be related, at least in part, to Paleozoic deformation. Geochemical analyses of selected sulfide-rich rocks from this area show elevated values of arsenic (100–200 ppm), molybdenum (5–7 ppm), lead (73 ppm), and zinc (100–160 ppm). Although gold was not found in any of the analyzed samples of gneiss (detection limit 0.05 ppm), the listed metal anomalies (especially for arsenic), and the presence of sulfidic zones, are considered to be promising indicators of mineralization and designate an area of moderate resource potential for gold (fig. 5, no. 5–8). Associated belts of augen gneiss in this part of the Green Mountain massif are shown in figure 5 as having low resource potential for gold. These augen gneisses correspond to the Bull Hill Gneiss of Doll and others (1961) and belong to a group of 965- to 945-Ma metamorphosed granites in the Green Mountain massif and the Chester and Athens domes (Karabinos and Aleinikoff, 1988). More work will be required to evaluate the gold potential of the Bull Hill Gneiss in the Chester and Athens domes and the potential of the older (1120 Ma) augen gneisses in the northern part of the Green Mountain massif (see Karabinos, 1987).

To the west, the Taconic allochthon is shown (fig. 5) as having a low resource potential for gold in quartz veins, quartz stockworks, and (or) irregular disseminations. This type of gold is another newly recognized mineral resource in the quadrangle (Slack and Watts, 1989), a conclusion supported both by a bedrock mineral occurrence and by panned concentrate anomalies for gold and arsenic (Day and others, 1986; Watts, this volume). The bedrock occurrence is at the Mount Colfax prospect north of Cambridge, N.Y. (Dale, 1899, p. 186), where a gold-bearing, quartz-pyrite stockwork is localized within black slate. Such an occurrence probably is tectonically controlled and related to thrust faults formed during emplacement of the Taconic allochthon in Middle Ordovician time and (or) to later Mesozoic extensional faulting. The bedrock occurrence is designated as having high resource potential for gold (fig. 5, no. 5–9). Panned concentrate anomalies within the allochthon are shown as having moderate potential (fig. 5, no. 5–10).

The potential for gold mineralization in the Taconics deserves serious consideration. One possible analog is in the Bohemian massif of Czechoslovakia, where mineralized quartz stockworks in Devonian (Variscan) black slate locally contain elevated values of arsenic, barium, tin, molybdenum, and silver; in places there is as much as 4 to 5 g/t of gold (Aichler and others, 1988; M. Vaněček, oral commun., 1986). This Czechoslovakian gold deposit has precisely the same suite of associated anomalous metals as those in the Taconics (Watts, this volume); the concentrations of tin are especially noteworthy in areas that lack felsic igneous rocks (either plutonic or volcanic) and probably are derived from stannite and (or) tin-bearing sulfosalt minerals (Watts, this volume). General analogies can also be made with the large Muruntau gold deposits of Uzbekistan (Soviet central Asia) that consist of gold-bearing quartz stockworks in deformed black slates (Borodaevskaya and Rozhkov, 1977; Strishkov, 1986); most of the panned concentrate anomalies discovered in the Taconics are represented in the geochemical signature of the Muruntau deposits including elevated values for boron (Watts, this volume, fig. 11) that suggest epigenetic concentrations of tourmaline in the Taconics like those recognized in the Muruntau gold ores (Borodaevskaya and Rozhkov, 1977, p. 35–40). Another possible interpretation is that the gold and some of the other geochemical anomalies in the Taconics reflect the presence of sediment-hosted, Carlin-type gold mineralization, like that in many deposits of the Western United States, especially Nevada (Bagby and Berger, 1985; Percival and others, 1988). The lithologic and tectonic setting of the Taconics (particularly the western and northern low Taconics) is in many respects similar to that of the Nevada deposits, except that the late extensional faulting in the Taconics took place during Mesozoic time, rather than in the late Tertiary as in the Great Basin of Nevada. Such extensional faults, and their intersections with low-angle thrust faults of the region, are considered to be favorable areas for the occurrence of vein or disseminated gold mineralization (Slack and Watts, 1989). The intersection of extensional faults with carbonate beds, like those present in many units of the low Taconics (for example, Hatch Hill and West Castleton Formations), also may have preferentially localized jasperoid formation and the precipitation of fine-grained, Carlin-type gold deposits (see Berger and Bagby, 1990). More detailed field studies are needed to test this hypothesis.

DEPOSITS OF TIN, TUNGSTEN, MOLYBDENUM, AND (OR) FLUORITE

Figure 6 shows the resource assessment for several types of deposits that contain tin, tungsten, molybdenum, and (or) fluorite. Data from mineral occurrences (Slack and Schruben, this volume) and regional geochemistry (Watts, this volume) suggest that many of these elements are spatially associated with late extensional faults, probably of Mesozoic age, both in the eastern Adirondacks and in western New Hampshire. Mineralization in such faults is known in the Westmoreland area, New Hampshire, just south of the Glens Falls quadrangle, where large veins locally contain fluorite, barite, and base-metal sulfides; the fluorite in these veins was mined intermittently from 1911–38 (Bannerman, 1941). Areas surrounding the late extensional faults of the Glens Falls quadrangle are designated as having low resource potential for several metals with or without accompanying fluorite. Other areas surrounding a few of the faults have significant geochemical anomalies for tungsten in heavy-mineral concentrates of stream sediment (Watts, this volume, fig. 8) and are shown as having moderate potential (fig. 6, no. 6–1). The resource importance of these anomalies is unclear, but some of the larger fault zones may host extensive mineralization.

Along the western edge of the quadrangle, Watts (this volume, fig. 9) identified anomalous amounts of tin in heavy-mineral concentrates. The tin-bearing samples come from an area underlain by Proterozoic granitic gneiss as mapped by McLelland (this volume). Unlike the concentrate anomalies for tungsten (Watts, this volume, fig. 8), those for tin are not associated with mapped faults. The tin-bearing samples may be derived from granite-related tin deposits that were deformed and metamorphosed during the Late Proterozoic Grenvillian orogeny, a suggestion consistent with the recent interpretation of the igneous rocks of the Adirondacks as an anorogenic magmatic suite (McLelland, 1986). A moderate resource potential is assigned to areas of granitic gneiss that contain associated tin anomalies (fig. 6, no. 6–2); surrounding granitic gneisses are proposed to have low potential for these types of tin deposits.

In western New Hampshire, diverse geochemical anomalies and mineral occurrences appear to be related to Ordovician, Devonian, and Late Devonian or Carboniferous granitoid plutons. The oldest of these are the Ordovician Oliverian domes, which have associated scheelite- and molybdenite-bearing veins developed along the northern edge of the Mascoma dome (Slack and others, 1987; fig. 6, no. 6–3) and molybdenite-bearing veins along the southern and western contacts of the Lebanon dome (fig. 6, no. 6-4). Major tin and tungsten values in panned concentrates within and surrounding the Smarts Mountain, Mascoma, and Unity domes (Watts, this volume) indicate a moderate resource potential for these metals (fig. 6, no. 6–5). The other Oliverian body within the Glens Falls quadrangle, the Croydon dome, apparently lacks mineral occurrences and geochemical anomalies and on the basis of available information is judged not to have a resource potential.

Geochemical anomalies associated with the syn- or late-kinematic Bethlehem Gneiss (Devonian) in western New Hampshire are very systematic. Although no mineral occurrences are known in the outcrop belt of the Bethlehem Gneiss, mapped areas of this intrusive type have very high associated tin and tungsten anomalies in panned concentrate samples (Watts, this volume). The geochemically anomalous areas are shown as having moderate potential (fig. 6, no. 6–6), while the remaining parts of the Bethlehem Gneiss have low potential. Cassiterite and molybdenite are known to occur in some pegmatites within and peripheral to the Bethlehem Gneiss (Morrill, 1960) and define an area of moderate resource potential for this deposit type (fig. 6, no. 6–7). The presence of these pegmatites may explain selected panned concentrate anomalies, but the apparent lack of tungsten in any of the pegmatites (see Cameron and others, 1954; Morrill, 1960) suggests that the tungsten in the concentrates is derived from a different bedrock source. One possibility is that the tungsten, and perhaps the tin also, is present in porphyry-related veins and (or) greisens, both of which would constitute significant new deposit types in the Glens Falls quadrangle. A greisen tin-tungsten deposit, if present, would likely be limited to areas near the margins of the Bethlehem Gneiss. In contrast, the nearby Kinsman Quartz Monzonite (Devonian) in the extreme southeastern corner of the quadrangle (not separately shown on figure 6), has a near total absence of geochemical anomalies (Watts, this volume) and is judged to lack metallic mineral potential.

At the northern edge of the quadrangle is the southern portion of a small postkinematic Devonian(?) granite body that has major associated tin and tungsten anomalies in panned concentrates (Watts, this volume). This body, shown by Doll and others (1961), also has surrounding airborne magnetic and electromagnetic anomalies (D.L. Daniels, oral commun., 1986; Long, this volume). These

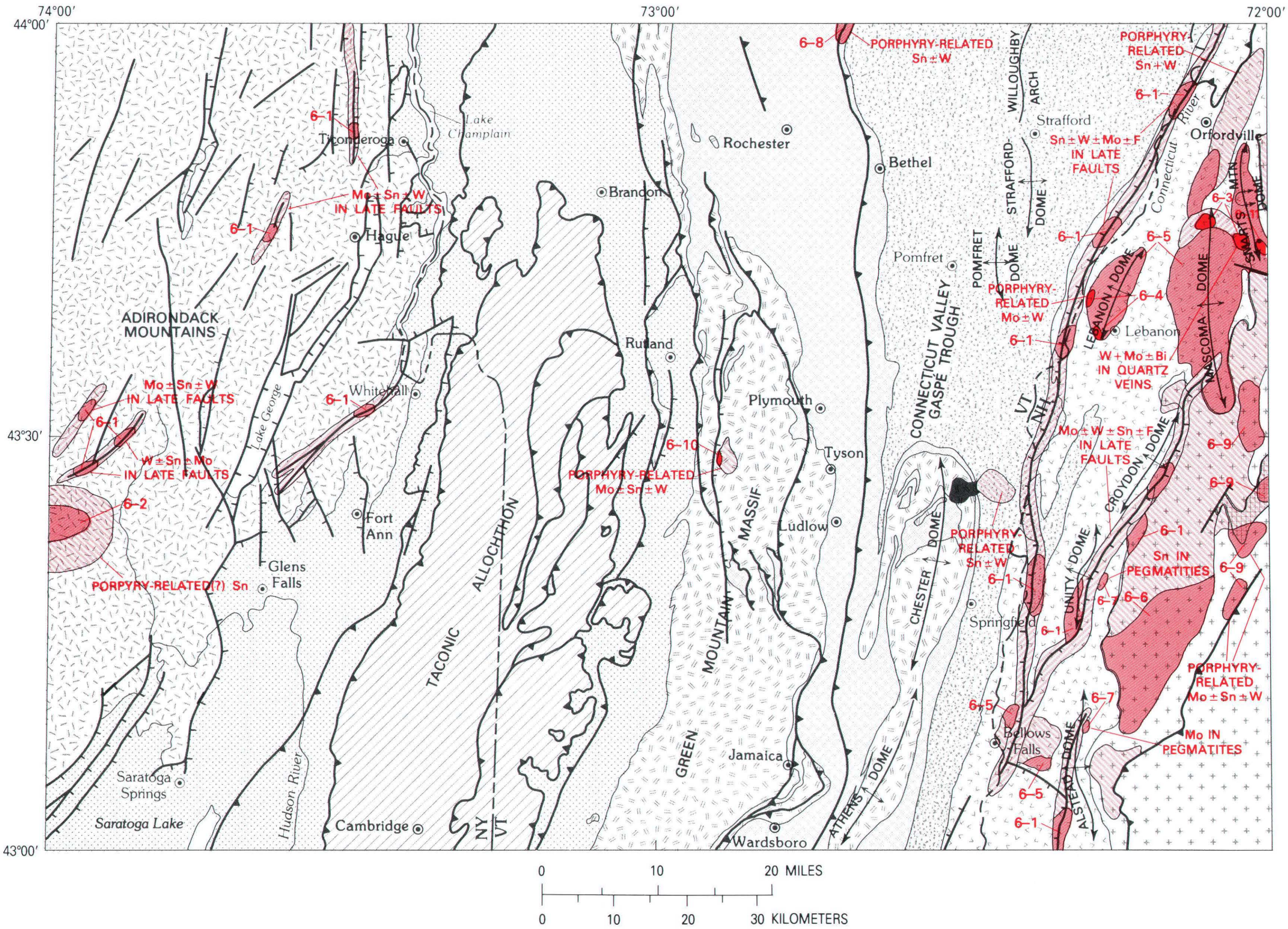

Figure 6. Mineral-resource potential for deposits of tin, tungsten, molybdenum, and (or) fluorite. See table 2 for explanation of locality numbers for high (dark red), moderate (medium red), and low (pink) areas of mineral-resource potential; note that different lined patterns for areas of low resource potential shown in pink refer to different deposit types. See figure 1 for explanation of resource potential patterns. See figure 2 for explanation of geologic map units and symbols.

geochemical and geophysical data suggest that this intrusion may have associated greisen- or skarn-type mineralization and, therefore, a moderate resource potential for these metals (fig. 6, no. 6–8). The presence of very high tungsten values in panned concentrates to the east of this intrusive may reflect skarn-type scheelite associated with the Knox Mountain pluton, just to the north of the Glens Falls quadrangle (Doll and others, 1961), or with buried granitic intrusions postulated to core the Strafford and Pomfret domes (see Watts, this volume). Concentrations of scheelite also have been identified in panned concentrate samples throughout this part of the Connecticut Valley-Gaspé trough (Slack and others, this volume). The occurrence of a small scheelite-bearing, calc-silicate skarn deposit in this same terrane at Lac Lyster, along the northern contact of the Averill pluton near the Vermont border in southeastern Quebec (Gauthier and others, 1985a; 1989), is considered to be an appropriate analogy for the Glens Falls quadrangle.

Along the eastern border of the quadrangle is the western half of the Sunapee pluton, a postkinematic, two-mica granite of Late Devonian age (see Thompson, this volume). The margins of this intrusion also have major associated tin and tungsten anomalies in panned concentrates (Watts, this volume), which suggest possible vein and (or) greisen-type mineralization. Although no mineral occurrences are known here, the areas of the geochemical anomalies and the favorable geologic setting imply a moderate resource potential for tin and tungsten (fig. 6, no. 6–9). Analogies for the Sunapee area are the Mount Pleasant tin-tungsten-molybdenum deposit in New Brunswick and the East Kempville greisen tin deposit in Nova Scotia, both of which are associated with two-mica granites of Late Devonian or Early Mississippian age (Kooiman and others, 1986; Richardson, 1988). Similar granite-related metal deposits are known elsewhere in the northern Appalachians including New Brunswick (Ruitenberg and Fyffe, 1982), Nova Scotia (Chatterjee and Clarke, 1985), and Newfoundland (Whalen, 1980).

The youngest igneous intrusions in the Glens Falls quadrangle, the Mesozoic alkaline bodies of the White Mountain Plutonic-Volcanic Suite, also have associated molybdenum mineralization. Molybdenite is known at a small prospect on the western side of the Cuttingsville stock (Schmidt, 1978; Robinson, this volume) and in the small felsic intrusive body on Pollard Hill in New Hampshire (L.G. Woodruff, U.S. Geological Survey, oral commun., 1986). Minor amounts of fluorite and sphalerite and high contents of copper (to 500 ppm), lead (to 500 ppm), arsenic (to 1,000 ppm), tin (to 500 ppm), and tungsten (to 150 ppm) also are present locally in rocks of the Pollard Hill area (J.F. Slack, unpub. data). These occurrences are indicated on the map as having high resource potential (fig. 6, nos. 6–10 and 6–11); remaining parts of the much larger White Mountain intrusive bodies at Cuttingsville and Ascutney are judged to have low resource potential for porphyry-related molybdenum, possibly also including tin, tungsten, and copper.

DEPOSITS OF CHROMIUM, NICKEL (±COPPER), VANADIUM, AND PLATINUM GROUP ELEMENTS

Figure 7 outlines the potential for deposits of chromium, nickel(±copper), vanadium, and platinum group elements (PGE). With regard to chromium, many podiform chromite deposits are associated with alpine-type ultramafic bodies in the southeastern Quebec Appalachians, along strike to the north of the Glens Falls 1° × 2° quadrangle (Kacira, 1982; Gauthier and others, 1987, 1989). One occurrence of massive podiform chromite is known in a verde antique quarry at Rochester, Vt. (B.R. Lipin, oral commun., 1984; Ratté and Ogden, 1989), and is designated as having high resource potential (fig. 7, no. 7–1). Other areas of exposed ultramafic rock in the quadrangle, to the south near Ludlow, Vt. (Doll and others, 1961), and to the east in the Orfordville, N.H., area (Moench, this volume, fig. 2), are assigned a low resource potential for chromium on the basis of favorable geology but no known occurrences of significant chromite.

On the eastern side of the Green Mountains in Vermont, one occurrence of platinum has been recorded in a discordant gold-bearing quartz vein at the Hinchman prospect (Perkins, 1904). This is the only known platinum locality within the Glens Falls quadrangle (Slack and Schruben, this volume). Because of the small size of such veins, this occurrence probably has limited economic importance and is thus designated as an area of only moderate resource potential (fig. 7, no. 7–2).

Concentrations of platinum group elements may occur in association with ultramafic rocks of the quadrangle. On the basis of this possibility, A.H. Chidester and B.R. Lipin of the U.S. Geological Survey in 1981 and 1982 collected more than 100 samples of ultramafic rocks from Vermont for PGE analyses, including many from within the Glens Falls 1° × 2° quadrangle. The analyses show, however, that PGE concentrations in these rocks are very low, with most samples containing less than 5 parts per billion (ppb) palladium and less than 10 ppb platinum; the highest reported values are 28 ppb palladium and 41 ppb platinum (B.R. Lipin, oral commun., 1986). Associations of PGE with chromitites throughout the world, including those recently discovered in southeastern Quebec (Gauthier and others, 1989; Gauthier and others, 1990), suggest that chromite-rich rocks in Vermont are especially attractive PGE targets, but analysis of a chromitite sample collected by Chidester and Lipin from the Rochester, Vt., verde antique quarry does not show elevated PGE values (<5 ppb each of palladium and platinum). Accordingly, the belt of ultramafic rocks within the Glens Falls quadrangle is judged to lack a resource potential for PGE.

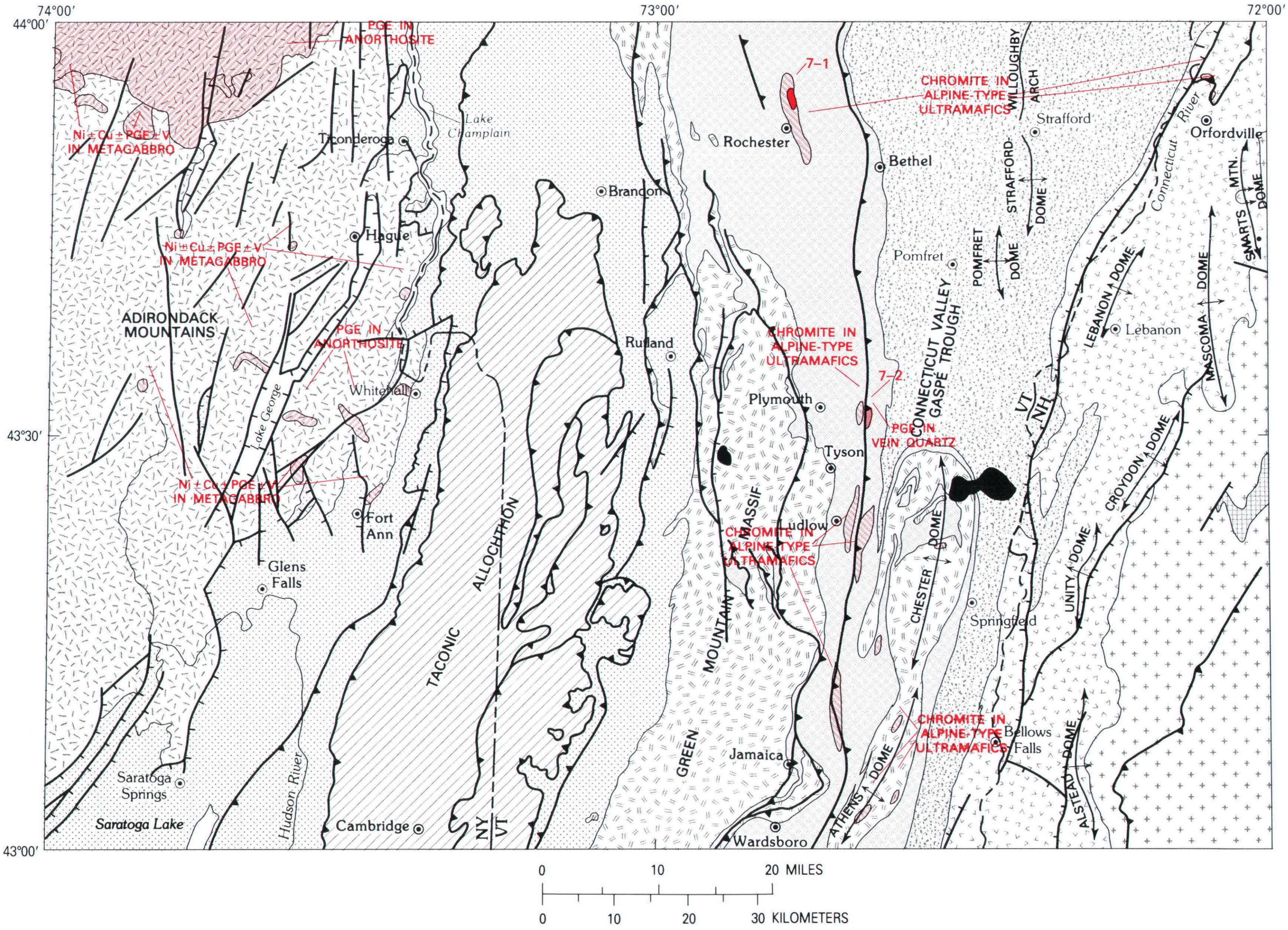

Figure 7. Mineral-resource potential for deposits of chromium, nickel (±copper), vanadium, and platinum group elements (PGE). See table 2 for explanation of locality numbers for high (dark red), moderate (medium red), and low (pink) areas of mineral-resource potential; note that different lined patterns for areas of low resource potential shown in pink refer to different deposit types. See figure 1 for explanation of resource potential patterns. See figure 2 for explanation of geologic map units and symbols.

Anorthositic rocks in the eastern Adirondack Mountains also may contain anomalous amounts of PGE. This speculation is based on the broad association of PGE with anorthosites in many layered mafic intrusions worldwide (for example, Naldrett, 1981; Macdonald, 1987). The anorthosites in the Adirondacks are of the domal or massif type, however, in contrast to those within layered intrusions (for example, Bushveld complex, South Africa; Stillwater complex, Montana). No economic concentrations of PGE are known in domal anorthosites, but the occurrence of as much as 185 micrograms/tonne platinum and 150 micrograms/tonne palladium in the Harp Lake anorthosite in Labrador (Jones, *in* Wardle, 1987) suggests that such a possibility cannot be discounted. Areas of anorthosite and gabbroic anorthosite in the Adirondacks are therefore assigned a low resource potential for PGE (fig. 7).

Also present in the eastern Adirondacks are a series of small metagabbroic intrusive bodies (fig. 7). These mafic bodies are judged to have low resource potential for several metal commodities, including nickel with or without copper, PGE, and vanadium. No mineral occurrences or geochemical anomalies are known to be associated with these bodies, but comparison with similar gabbroic intrusions elsewhere suggests the possibility of mineralization. Analogs may include the Renzy Lake copper-nickel deposit in Quebec and the Macassa nickel-copper-cobalt deposit in Ontario, both of which are associated with Proterozoic metagabbros in the Grenville province (Carter and Colvine, 1985). The Proterozoic metagabbros of central Sweden, in the Baltic Shield, recently have been identified as having possible economic concentrations of vanadium in magnetite-rich zones (Ros and others, 1986), suggesting potential for similar vanadium occurrences in the metagabbroic intrusions of the Adirondacks. PGE are a possibility also, on the basis of deposits of these metals in gabbros in many parts of the world (for example, Macdonald, 1987).

DEPOSITS OF IRON, MANGANESE, AND TITANIUM

Figure 8 shows areas of resource potential for iron, manganese, and titanium. The eastern part of the Cambrian and Ordovician shelf sequence in Vermont contains several small residual iron deposits, some of which have associated manganese (Hager, 1861; Burt, 1931[?]). Areas surrounding these oxide deposits, including the well-known examples near Brandon (fig. 8, nos. 8–1 to 8–3) and those along strike to the south (fig. 8, nos. 8–4 to 8–7) are judged to have high resource potential. The entire portion of this outcrop belt is designated as an area of low resource potential for this deposit type.

To the east are two small belts that contain stratabound iron deposits in Late Proterozoic or Early Cambrian metasedimentary rocks of the eastern Vermont cover sequence, east of the Proterozoic basement of the Green Mountain massif. The small iron deposits near the towns of Tyson and Plymouth, Vt. (fig. 8, nos. 8–8 and 8–9, respectively), consist of lenses of magnetite and (or) hematite along dolomite contacts in the upper part of the Tyson Formation (as used by Doll and others, 1961); these deposits were economically important in the last century (mined 1836–55). A different type of iron deposit occurs on Weaver Hill about 2.7 km northeast of Tyson (fig. 8, no. 8–10), where small stratabound and apparently stratiform lenses of massive siderite in pelitic schist were mined for spathic (carbonate) iron during the 1850's (Hager, 1861).

In the eastern Adirondacks, a suite of basal alaskitic gneisses, outlined in pink, contains several small- to moderate-sized stratabound and stratiform, nontitaniferous magnetite deposits. The areas of the deposits are judged to have high resource potential for iron (fig. 8, nos. 8–11 to 8–15). These deposits include the Hammondville ores and similar iron deposits near Crown Point, Skiff Mountain, and Fort Ann, N.Y. The iron deposits are very similar to the large Mineville-Port Henry low-titanium magnetite ores just to the north of the Glens Falls quadrangle (Newland, 1908) and to the iron ore deposits at Benson mines in the northwestern Adirondacks (Crump and Beutner, 1968).

In the Adirondack highlands, the areas in pink contain and surround gabbroic rocks within the main anorthosite massif. These areas are assigned a low resource potential for the occurrence of titaniferous magnetite deposits of the Sanford Lake type (for example, Gross, 1968). Only one such deposit is known within the Glens Falls 1° × 2° quadrangle, the small Moose Mountain occurrence (fig. 8, no. 8–16). No significant resource potential is apparent for this deposit type beyond the area of the Moose Mountain mine.

DEPOSITS OF URANIUM, THORIUM, AND RARE EARTH ELEMENTS

Figure 9 illustrates the potential for uranium, thorium, and rare earth elements in various deposit types. The high potential areas in the south-central part of the map (fig. 9, nos. 9–1 to 9–3) show the locations of major known uranium deposits in Grenville basement rocks of the quadrangle, at Grant Brook west of Ludlow, Vt., and near Jamaica, Vt. The largest known deposit is in the College Hill area (no. 9–2) west of Jamaica, where much of the uranium is in stratabound veins associated with Grenville-age pegmatite dikes (Ayuso and Ratté, this volume). Moderate potential is assigned to a relatively large pegmatite in the Weston, Vt., area (fig. 9, no. 9–4), which contains at least 200 ppm U_3O_8 in surface outcrops (McHone and Wagener, 1982). Other localized areas of anomalous radioactivity in the Mount Holly Complex (Grauch and Zarinski, 1976; McHone and Wagener, 1982) are judged to have only moderate resource potential for uranium and (or) thorium.

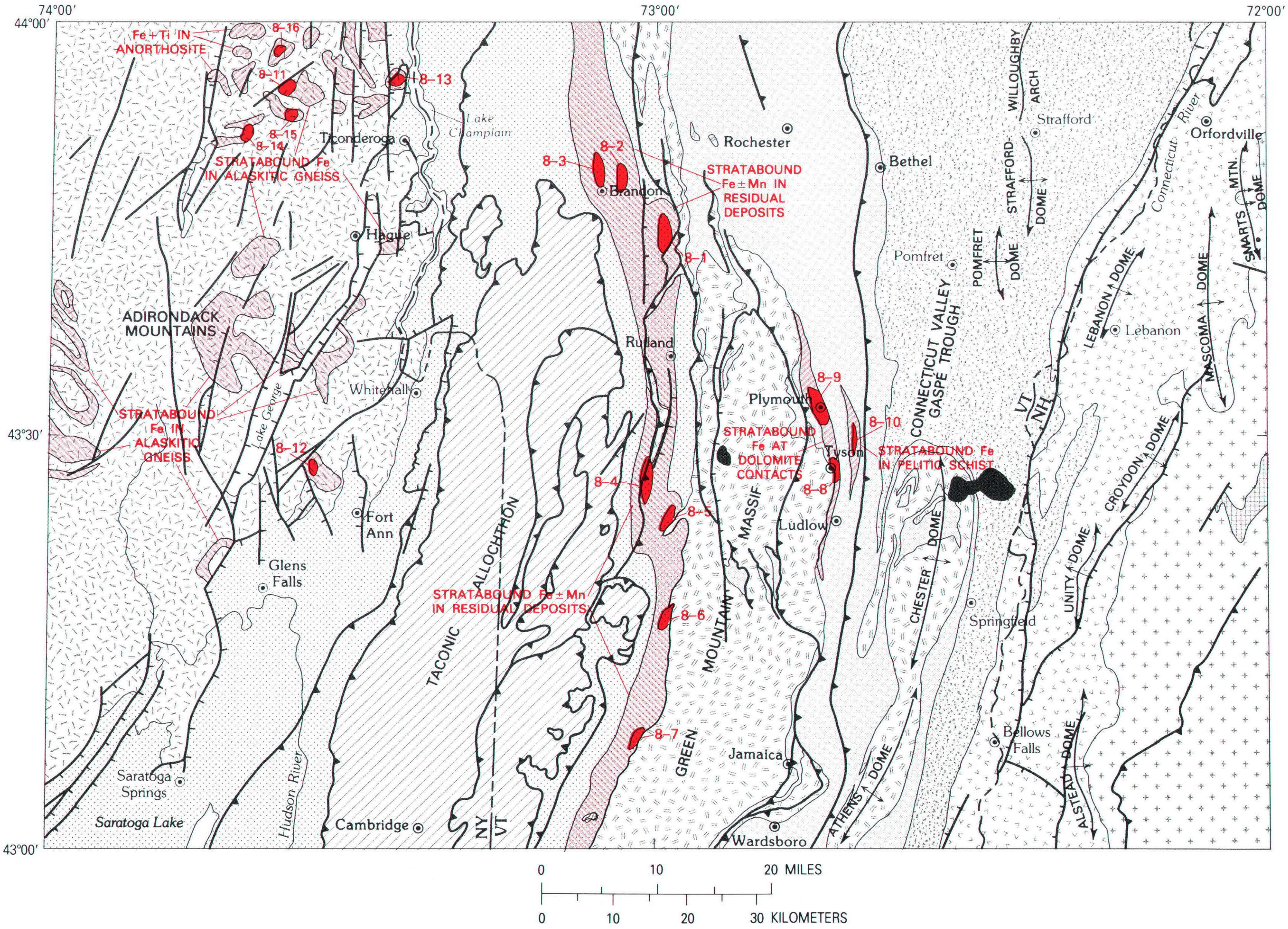

Figure 8. Mineral-resource potential for deposits of iron, manganese, and titanium. See table 2 for explanation of locality numbers for high (dark red), moderate (medium red), and low (pink) areas of mineral-resource potential; note that different lined patterns for areas of low resource potential shown in pink refer to different deposit types. See figure 1 for explanation of resource potential patterns. See figure 2 for explanation of geologic map units and symbols.

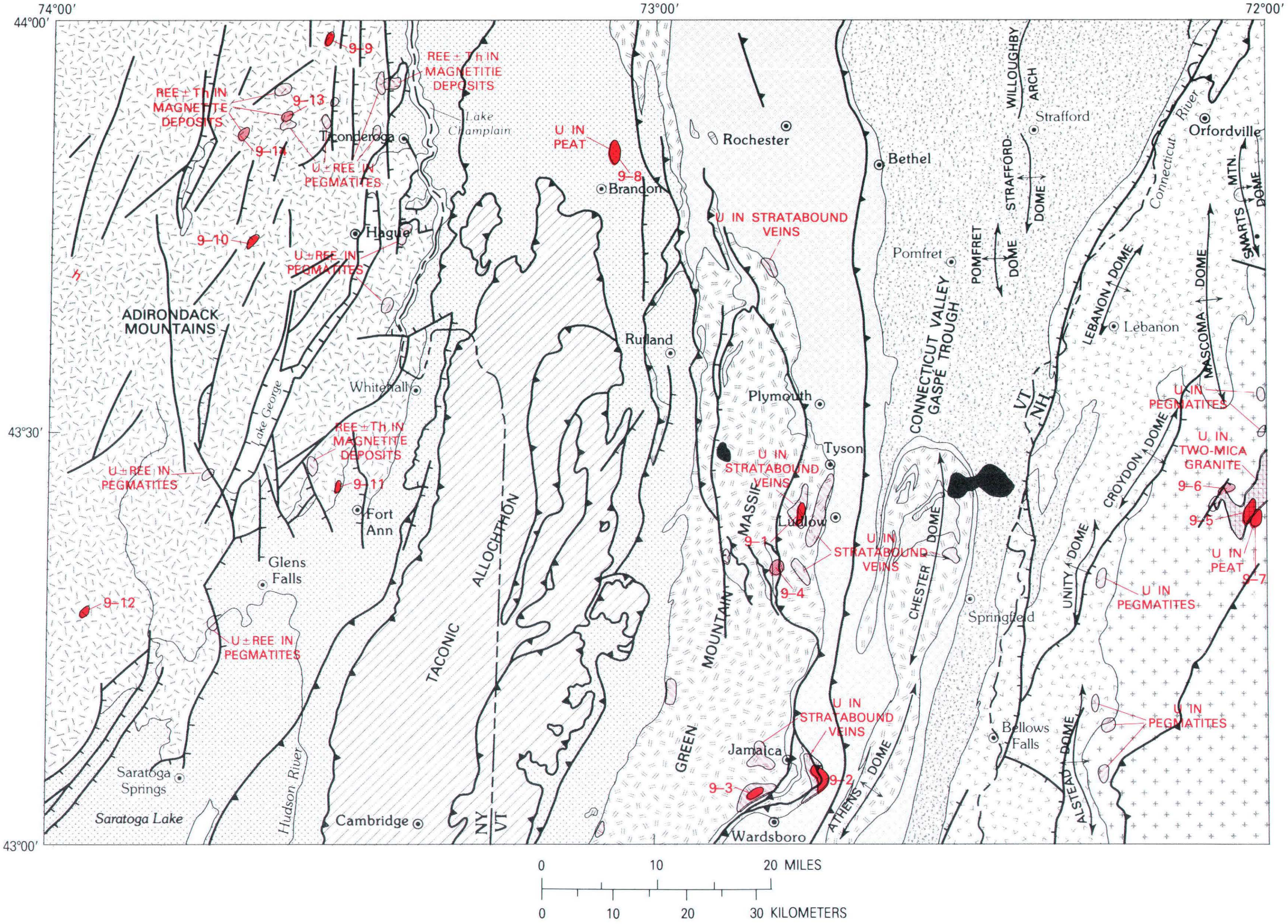

Figure 9. Mineral-resource potential for deposits of uranium, thorium, and rare earth elements (REE). See table 2 for explanation of locality numbers for high (dark red), moderate (medium red), and low (pink) areas of mineral-resource potential; note that different lined patterns for areas of low resource potential shown in pink refer to different deposit types. See figure 1 for explanation of resource potential patterns. See figure 2 for explanation of geologic map units and symbols.

In western New Hampshire, several types of uranium concentrations are recognized. The oldest, apparently, is minor occurrences of uranium minerals in pegmatites associated with the Devonian Bethlehem Gneiss (Morrill, 1960). The Sunapee two-mica granite, of Late Devonian age (see Thompson, this volume), contains secondary uranium minerals along fractures and joints (Bothner, 1978; McHone and Wagener, 1982). The more significant Sunapee mineralization consists of hexavalent uranium minerals (autunite, renardite, torbernite) in the western interior of the pluton, especially in a 5-m-thick zone along interstate highway I-89. Bothner (1978), Lyons (1979), and McHone and Wagener (1982) describe this mineralization in detail; Lyons (1979) estimated a resource of 250 tons of U_3O_8 in the area. On the basis of these studies, a high potential is assigned for granite-related uranium deposits here (fig. 9, no. 9–5). A smaller mineralized zone of secondary uranium enrichment in granite and pegmatite along the northern contact of the pluton (no. 9–6) is judged to have moderate potential. The remaining part of the Sunapee pluton within the Glens Falls quadrangle is considered to have favorable geology and therefore a low potential for uranium of this deposit type.

The youngest known uranium deposits in this area are in Holocene peat (Cameron and others, this volume) in the vicinity of the Sunapee pluton and near Brandon, Vt. (fig. 9, nos. 9–7 and 9–8, respectively). The concentrations of uranium in these surficial peat deposits are very high and are in the range of ore grades at the Flodelle Creek uranium mine in the State of Washington, where uranium also is concentrated in peat (Johnson and others, 1987). Surficial uranium deposits constitute a newly recognized and significant mineral resource within the Glens Falls quadrangle.

Occurrences of radioactive minerals also are known locally along the Precambrian-Cambrian unconformity in eastern New York State. In places, basal quartz-pebble conglomerates within this unconformity display significant response on hand-held scintillometers; the radioactivity reflects minor concentrations of detrital monazite, allanite, and zircon (McHone and Wagener, 1982). This type of mineral occurrence, because of its very restricted distribution, probably is not an important resource and is not shown as such on figure 9.

In the eastern Adirondacks, uranium is concentrated in granitic pegmatites within Grenville basement rocks (Tan, 1966; McHone and Wagener, 1982). Most of these occurrences are in small bodies that have unfavorable geology and thus only moderate potential (fig. 9). However, four of the pegmatites are large ($>2{,}000\ m^2$) and have significant amounts of uranium, like those mined in Grenville basement rocks in the Bancroft, Ontario, area (Robinson, 1960; McHone and Wagener, 1982). These four large pegmatites are therefore judged to have high resource potential for granite-related uranium deposits (fig. 9, nos. 9–9 to 9–12). The reconnaissance work of McHone and Wagener (1982) suggests that these pegmatites also have concentrations of rare earth elements (REE), especially lanthanum (see also Rowley, 1962).

Another mineral occurrence in the Adirondacks consists of anomalous thorium and possibly REE associated with stratabound and stratiform, nontitaniferous magnetite deposits (L.C. Gundersen, U.S. Geological Survey, oral commun., 1985). The similar Mineville-Port Henry magnetite deposits, just to the north of the Glens Falls quadrangle (Newland, 1908), contain significant quantities of REE (McKeown and Klemic, 1956; Beck, 1985), which recently have been evaluated for commercial extraction (Jenkins, 1985). Reconnaissance radiometric surveys of the smaller, nontitaniferous magnetite deposits in the Glens Falls quadrangle by L.C. Gundersen (U.S. Geological Survey, oral commun., 1986) suggest that a few of the deposits, particularly those at Skiff Mountain and near Paradox Lake, N.Y. (fig. 9, nos. 9–13 and 9–14), may have anomalous concentrations of thorium and REE within the magnetite ores (McKeown and Klemic, 1956) and (or) in surrounding country rocks (see Gundersen, 1984). More work will be required to fully evaluate the potential of these magnetite deposits for REE.

Concentrations of uranium are also known in carbonate rocks of the early Paleozoic shelf sequence along the Vermont-Quebec border (Gauthier and others, 1985a, 1989; Prud'homme, 1986). Brecciated dolomite of the Cambrian Milton Dolomite in this area (Clarendon Springs Dolomite in Vermont) locally contains black phosphatic fragments that have elevated values of uranium, zinc (as sphalerite), and lead (as galena). A positive correlation exists between the concentrations of uranium and zinc in these fragments, although the contents of both metals are generally low, in the range of hundreds of ppm (Michel Gauthier, oral commun., 1985). Some of the richest mineralization appears to be on the Vermont side of the border in what is known as the Highgate uranium deposit; quantitative data are unavailable on this occurrence, however. There is no indication of similar uranium-zinc mineralization within coeval rocks of the Glens Falls 1° × 2° quadrangle, and so no resource potential is assigned at this time. More detailed work in the future is necessary to thoroughly evaluate the possible occurrence of this deposit type in the quadrangle.

CONCLUSIONS

The Glens Falls CUSMAP project has identified many types of metallic mineral resources within the quadrangle, some of which were previously unrecognized. Major known and potential metallic resources are (1) volcanogenic massive sulfide deposits (copper+zinc±lead ±silver±gold) in early Paleozoic volcanosedimentary belts; (2) stratiform (sedex-type) lead+zinc(±copper±silver ±barite) deposits in the early Paleozoic shelf sequence and in the Taconic allochthon; (3) volcanogenic gold(±silver) deposits in early Paleozoic volcanosedimentary belts; (4) vein-type (metamorphogenic) gold deposits in the eastern cover sequence of central Vermont; (5) gold associated with altered ultramafic rocks; (6) disseminated or stockwork gold deposits in black slates and limestones of the Taconic allochthon; (7) porphyry-related gold deposits associated with the Mesozoic alkaline intrusions of the White Mountain Plutonic-Volcanic Suite; (8) fault-related tin-tungsten-molybdenum-fluorite deposits; (9) contact-metasomatic (skarn) tungsten deposits associated with Paleozoic plutons; (10) granitoid-associated tin-tungsten-molybdenum-copper deposits related to Paleozoic and Mesozoic intrusions; (11) stratabound uranium-bearing veins and lenses in Grenville-age basement rocks; (12) uranium deposits in two-mica granites and granitic pegmatites; and (13) uranium in surficial peat deposits.

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Water-Supply Papers are comprehensive reports that present significant interpretive results of hydrologic investigations of wide interest to professional geologists, hydrologists, and engineers. The series covers investigations in all phases of hydrology, including hydrogeology, availability of water, quality of water, and use of water.

Circulars present administrative information or important scientific information of wide popular interest in a format designed for distribution at no cost to the public. Information is usually of short-term interest.

Water-Resources Investigations Reports are papers of an interpretive nature made available to the public outside the formal USGS publications series. Copies are reproduced on request unlike formal USGS publications, and they are also available for public inspection at depositories indicated in USGS catalogs.

Open-File Reports include unpublished manuscript reports, maps, and other material that are made available for public consultation at depositories. They are a nonpermanent form of publication that may be cited in other publications as sources of information.

Maps

Geologic Quadrangle Maps are multicolor geologic maps on topographic bases in 7.5- or 15-minute quadrangle formats (scales mainly 1:24,000 or 1:62,500) showing bedrock, surficial, or engineering geology. Maps generally include brief texts; some maps include structure and columnar sections only.

Geophysical Investigations Maps are on topographic or planimetric bases at various scales; they show results of surveys using geophysical techniques, such as gravity, magnetic, seismic, or radioactivity, which reflect subsurface structures that are of economic or geologic significance. Many maps include correlations with the geology.

Miscellaneous Investigations Series Maps are on planimetric or topographic bases of regular and irregular areas at various scales; they present a wide variety of format and subject matter. The series also includes 7.5-minute quadrangle photogeologic maps on planimetric bases that show geology as interpreted from aerial photographs. Series also includes maps of Mars and the Moon.

Coal Investigations Maps are geologic maps on topographic or planimetric bases at various scales showing bedrock or surficial geology, stratigraphy, and structural relations in certain coal-resource areas.

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Miscellaneous Field Studies Maps are multicolor or black-and-white maps on topographic or planimetric bases on quadrangle or irregular areas at various scales. Pre-1971 maps show bedrock geology in relation to specific mining or mineral-deposit problems; post-1971 maps are primarily black-and-white maps on various subjects such as environmental studies or wilderness mineral investigations.

Hydrologic Investigations Atlases are multicolored or black-and-white maps on topographic or planimetric bases presenting a wide range of geohydrologic data of both regular and irregular areas; principal scale is 1:24,000, and regional studies are at 1:250,000 scale or smaller.

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