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W. Harrison, M. Flower, M. Sood, M. Tisue, and D. Edgar

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CRYSTALLINE ROCKS OF THE NORTHEASTERN UNITED STATES

by

W. Harrison, M. Flower,* M. Sood,** M. Tisue, and D. Edgar

Graphics by M. Stahl and B. Termaat

Energy and Environmental Systems Division
Geoscience and Engineering Group

October 1983

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through
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Project Management Division

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CONTENTS

FOREWORD xiii

CONTRIBUTORS xiv

REVIEWERS xviii

ACKNOWLEDGMENTS xix

GEOCHRONOLOGY xx

ABSTRACT 1

EXECUTIVE SUMMARY 2

1 INTRODUCTION 4

1.1 Purpose and Scope 4
1.2 Sponsorship 4
1.3 Plates 4
1.4 References for Section 1 5

2 SYNOPSIS OF CRYSTALLINE ROCKS 6

2.1 Rock Nomenclature 6
2.2 Genetic Classification and Environment of Crystalline Rocks 6
2.3 Chronologic and Tectonic Classification of Crystalline Rocks in the Northeastern United States 10
2.3.1 Grenville and Avalonian 12
2.3.2 Taconic 13
2.3.3 Acadian and Post-Acadian 14
2.3.4 Alleghenian and Post-Alleghenian 14
2.4 Caveats 14
2.5 Presentation of Individual Pluton Reports 15
2.6 References for Section 2 15

3 CRYSTALLINE ROCKS OF MAINE 18

3.1 Grenville and Avalonian 18
3.1.1 Chain Lakes Massif 18
3.2 Taconic 21
3.2.1 Attean Quartz Monzonite 21
3.3 Acadian and Post-Acadian 23
3.3.1 Biddeford Pluton 23
3.3.2 Bottle Lake Quartz Monzonite 24

CONTENTS (Cont'd)

3.3.3 Cadillac Mountain and Somesville Plutons..... 27
3.3.4 Calais Gabbro-Diorite Complex..... 28
3.3.5 Center Pond Quartz Monzonite..... 31
3.3.6 Chiputneticook Quartz Monzonite..... 34
3.3.7 Concord and Related Plutons..... 36
3.3.8 Cranberry Lake Pluton..... 36
3.3.9 Flagstaff Lake Complex 39
3.3.10 Gardner Lake Gabbro-Diorite Complex..... 41
3.3.11 Gouldsboro Pluton..... 42
3.3.12 Granodiorite of Whitneyville..... 44
3.3.13 Hallowell and Togus Plutons..... 45
3.3.14 Hartland Pluton..... 48
3.3.15 Katahdin Batholith..... 50
3.3.16 Lead Mountain Pluton..... 52
3.3.17 Lexington Pluton..... 55
3.3.18 Lucerne Pluton..... 58
3.3.19 Lyman Pluton..... 60
3.3.20 Meddybemps and Related Granite..... 62
3.3.21 Mooselookmeguntic Pluton..... 66
3.3.22 Mt. Waldo Pluton..... 70
3.3.23 Moxie Pluton..... 72
3.3.24 Peaked Mountain Pond Gabbro-Diorite Complex..... 77
3.3.25 Phillips Pluton..... 78
3.3.26 Pierce Pond Pluton..... 81
3.3.27 Pocamoonshine Gabbro-Diorite..... 83
3.3.28 Quaker Hill Pluton..... 85
3.3.29 Redington Pluton..... 87
3.3.30 Rome Pluton..... 89
3.3.31 Sebago Lake Batholith..... 91
3.3.32 Seboeis Complex..... 95
3.3.33 Sedgwick Granite..... 97
3.3.34 Seven Ponds Pluton..... 98
3.3.35 Sugarloaf Pluton..... 101
3.3.36 Tunk Lake Pluton..... 102
3.3.37 Umbagog Granodiorite..... 105
3.3.38 Waldoboro Pluton..... 107
3.3.39 Webhannet Pluton..... 108
3.4 Alleghenian and Post-Alleghenian..... 111
3.4.1 Agamenticus Complex..... 111
3.5 References for Section 3..... 113

4 CRYSTALLINE ROCKS OF NEW HAMPSHIRE..... 124

4.1 Grenville and Avalonian..... 124
4.1.1 Massabesic Gneiss..... 124

4.2 Taconic 126
4.2.1 Ayer Intrusive Complex 126
4.2.2 Jefferson Dome 126
4.2.3 Lost Nation Group 128
4.2.4 Mascoma and Lebanon Domes 129
4.2.5 Newburyport Granodiorite 133
4.2.6 Westmoreland-Swanzey and Related Domes 133
4.3 Acadian and Post-Acadian 137
4.3.1 Ashuelot Pluton 137
4.3.2 Cardigan Pluton 139
4.3.3 Concord and Related Plutons 140
4.3.4 Exeter Pluton 142
4.3.5 Fitchburg Intrusive Complex 146
4.3.6 Fitzwilliam Granite 146
4.3.7 Granite Pluton near Barrington 147
4.3.8 Lincoln Pluton 148
4.3.9 Long Mountain Granite 150
4.3.10 Mt. Clough Pluton 152
4.3.11 Sebago Lake Batholith 154
4.3.12 Umbagog Granodiorite 154
4.3.13 Weare and Spaulding Plutons 154
4.3.14 Winnipesaukee Pluton 156
4.4 Alleghenian and Post-Alleghenian 158
4.4.1 Ossipee Mountains Pluton 158
4.4.2 Pilot-Pliny Complex 161
4.4.3 White Mountain Batholith 163
4.5 References for Section 4 169

5 CRYSTALLINE ROCKS OF VERMONT 174

5.1 Grenville and Avalonian 174
5.1.1 Chester Dome 174
5.1.2 Mt. Holly Complex, Northern Exposure 176
5.1.3 Mt. Holly Complex, Southern Exposure 178
5.1.4 Sadawga and Ray Pond Domes 181
5.2 Taconic 183
5.2.1 Barnard Gneiss 183
5.2.2 Lost Nation Group 185
5.3 Acadian and Post-Acadian 187
5.3.1 Averill Pluton 187
5.3.2 Echo Pond Pluton 188
5.3.3 Knox Mountain Pluton 190
5.3.4 Maidstone Pluton 193
5.3.5 Nulhegan Pluton 195
5.3.6 Victory Pluton 198

CONTENTS (Cont'd)

5.3.7 Willoughby Pluton..... 200
5.4 References for Section 5..... 202

6 CRYSTALLINE ROCKS OF NORTHEASTERN NEW YORK..... 205

6.1 Grenville and Avalonian..... 205
6.1.1 Alexandria Granite Gneisses..... 205
6.1.2 Diana Gneiss and Tupper/Saranac Metasyenite Complexes..... 211
6.1.3 Granite Gneiss..... 218
6.1.4 Hornblende Granite and Hermon Granite Gneisses..... 221
6.1.5 Main Metanorthosite Massif and the Snowy Mountain, Thirteenth Lake, and Oregon Anorthosite Domes..... 225
6.1.6 Olivine Metagabbros..... 231
6.1.7 Stark Metasyenite Complex..... 235
6.2 References for Section 6..... 239

7 CRYSTALLINE ROCKS OF MASSACHUSETTS..... 244

7.1 Grenville and Avalonian..... 244
7.1.1 Berkshire Massif..... 244
7.1.2 Dedham Granodiorite..... 247
7.1.3 Massabesic Gneiss..... 251
7.1.4 Milford Granite..... 252
7.1.5 Mt. Holly Complex, Southern Exposure..... 253
7.1.6 Pelham Dome..... 253
7.1.7 Salem Gabbro-Diorite..... 257
7.1.8 Sterling and Ponaganset Gneisses..... 259
7.2 Taconic..... 259
7.2.1 Andover Granite..... 259
7.2.2 Ayer Intrusive Complex..... 261
7.2.3 Barnard Gneiss..... 266
7.2.4 Cape Ann Intrusive Complex..... 266
7.2.5 Glastonbury Gneiss..... 269
7.2.6 Monson Gneiss..... 270
7.2.7 Newburyport Granodiorite..... 273
7.2.8 Sharpners Pond Diorite..... 275
7.3 Acadian and Post-Acadian..... 277
7.3.1 Belchertown Intrusive Complex..... 277
7.3.2 Fitchburg Intrusive Complex..... 281
7.3.3 Hardwick-Coys Hill Granite..... 283
7.3.4 Williamsburg Granodiorite..... 285
7.4 References for Section 7..... 287

CONTENTS (Cont'd)

8 CRYSTALLINE ROCKS OF RHODE ISLAND..... 296

8.1 Grenville and Avalonian..... 296
8.1.1 Bulgarmarsh Granite..... 296
8.1.2 Esmond Granite..... 298
8.1.3 Sterling Pluton Group..... 301
8.1.4 Ten Rod Granite Gneiss..... 305
8.2 Taconic..... 308
8.3 Acadian and Post-Acadian..... 308
8.3.1 Scituate Granite..... 308
8.4 Alleghenian and Post-Alleghenian..... 311
8.4.1 Narragansett Pier Granite..... 311
8.5 References for Section 8..... 312

9 CRYSTALLINE ROCKS OF CONNECTICUT..... 315

9.1 Grenville and Avalonian..... 315
9.1.1 Bear Hill Massif..... 315
9.1.2 Berkshire Massif..... 320
9.1.3 Housatonic Massif..... 322
9.1.4 Manhattan Prong..... 326
9.1.5 Sterling Plutonic Group..... 326
9.1.6 Waterford Gneiss Group..... 330
9.2 Taconic..... 332
9.2.1 Ayer Intrusive Complex..... 332
9.2.2 Brookfield Gneiss..... 332
9.2.3 Glastonbury Gneiss..... 334
9.2.4 Granitic Gneisses..... 339
9.2.5 Harrison Gneiss..... 342
9.2.6 Monson Gneiss..... 345
9.2.7 Norwalk Area Gneisses..... 349
9.3 Acadian and Post-Acadian..... 354
9.3.1 Canterbury Gneiss..... 354
9.3.2 Nonewaug Granite..... 358
9.4 Alleghenian and Post-Alleghenian..... 360
9.4.1 Narragansett Pier Granite..... 360
9.5 References for Section 9..... 360

10 CRYSTALLINE ROCKS OF SOUTHEASTERN NEW YORK, NORTHERN NEW JERSEY, AND SOUTHEASTERN PENNSYLVANIA..... 365

10.1 Grenville and Avalonian..... 365
10.1.1 Baltimore Gneiss Complex..... 365
10.1.2 Honey Brook Complex..... 368
10.1.3 Housatonic Massif..... 373

CONTENTS (Cont'd)

10.1.4 Manhattan Prong..... 373
10.1.5 Reading Prong, Northwestern Massif..... 376
10.1.6 Reading Prong, Southeastern Massif..... 380
10.1.7 Reading Prong, Western Extension..... 383
10.2 Taconic..... 384
10.2.1 Bedford Gneiss..... 384
10.2.2 Granitic Gneiss..... 384
10.2.3 Harrison Gneiss..... 385
10.2.4 Norwalk Area Gneiss..... 385
10.3 References for Section 10..... 385

APPENDIX: AGES OF CRYSTALLINE ROCK BODIES DESCRIBED IN THIS REPORT..... 389

FIGURES

2.1 Gradational Boundaries of Regional Metamorphic Facies..... 7

2.2 Classification of Plutonic Igneous Rocks..... 8

2.3 Composition of Metamorphic Rocks of Higher Temperature Ranges..... 9

2.4 Schematic Representation of Three Orogenic Environments of Granitic Rocks..... 12

3.1 Major Tectonic Features and Plutons of the Rangeley Lakes-Dead River Basin..... 20

3.2 Mount Desert Island and Lithologic Relations of the Cadillac Mountain Pluton..... 29

3.3 Sketch Map of Geology along St. Croix River near Calais, Maine, Showing the Relationship of Calais Diorite to the Other Granitic Bodies..... 30

3.4 Geologic Relations of the Chiputneticook Pluton to the Pokiok Batholith of New Brunswick..... 35

3.5 Bouguer Gravity Anomaly Map over the Mt. Katahdin Batholith..... 53

3.6 Generalized Geologic Relations of the Lucerne Pluton..... 59

3.7 Bouguer Anomaly Map over the Eastern Coast of Maine..... 61

3.8 Residual Bouguer Gravity Map for Part of the Eastern Coast of Maine..... 65

3.9 Geologic Map of the Mooselookmeguntic and Other Plutons of the Rangeley Lake Area..... 68

3.10 Isometric Block Diagram of Central Parts of Rangeley and Phillips Quadrangles Showing Early Folds and Faults, Preferred and Alternative Interpretations of Plutonic Structure, Metamorphic Zoning, and Superposed Late Recumbent Folds..... 70

3.11 Generalized Geologic Map of Southeastern Spencer, Southern Pierce Pond, Northeastern Stratton, and Northern Little Bigelow Mountain Quadrangles..... 82

3.12 Bouguer Anomaly Map over Waldoboro and Sebago Lake Plutons in Southwestern Maine..... 95

3.13 Geologic Map of Tunk Lake Pluton..... 103

3.14 Residual Gravity Anomaly Map of Waldoboro Pluton..... 109

4.1 Geologic Relations of Massabesic Gneiss in the Milford 15-min Quadrangle, New Hampshire..... 125

4.2 Geologic Map of the Northern Part of the Mascoma Dome..... 130

4.3 Major Oliverian Domes in the Keene-Brattleboro Area and Associated Structural Elements..... 134

4.4 Bouguer Anomaly Map Covering Cardigan, Weare, Spaulding, and Concord Plutons..... 143

4.5 Structure of the Seacoast Region of New Hampshire..... 145

4.6 Distribution of Lithological Types in the Exeter Pluton..... 146

4.7 Simplified Geologic Map of the Dixville Quadrangle..... 151

4.8 Contour Diagrams of Poles of Joints in the Bethlehem Gneiss in the Bellows Falls Quadrangle..... 154

FIGURES (Cont'd)

5.1 Geologic Relations in the East Barre Area..... 191

5.2 Geologic Relationships of the Nulhegan Pluton..... 196

6.1 Geologic Quadrangles of the Adirondack Area..... 206

6.2 Map of Major Rock Types in the Alexandria, Tupper/Saranac, and Hermon Complexes of the Northwest Adirondacks..... 207

6.3 Generalized Trends of Foliation and Linear Structure of the Alexandria, Tupper/Saranac, and Hermon Complexes..... 210

6.4 Geological Map of the Diana Syenite Gneiss Complex between Natural Bridge and Harrisville..... 213

6.5 Geology of the Snowy Mountain Dome Area..... 226

6.6 Bedrock Geology of Area between Elizabethtown and Mineville, Eastern Adirondacks..... 232

7.1 Location of the Berkshire Massif..... 245

7.2 Geologic Map of the Western Edge of the Berkshire Massif Showing Associated Major Thrust Faulting..... 248

7.3 Geologic Map of the Pelham Dome..... 255

7.4 Geologic Map of the Harvard-Ayer Area..... 264

7.5 Geologic Map of the Clinton Area..... 265

7.6 Alkalic Granitic Plutons in Boston Bay Area..... 267

7.7 Generalized Geologic Map Showing Distribution of Major Rock Units in Part of Northeastern Massachusetts..... 276

7.8 Geologic Map of Williamsburg Granodiorite Area..... 286

9.1 Geologic Map of Western Connecticut and Adjacent Eastern New York..... 316

9.2 Geologic and Tectonic Maps of the Lake Waramaug Area..... 319

9.3 Modal Analyses of the Housatonic Gneisses..... 324

FIGURES (Cont'd)

9.4 Geologic Map of West-Central Connecticut..... 325

9.5 Geologic Relations of Glastonbury Gneiss in Massachusetts and Connecticut..... 335

9.6 Proposed Correlation of Stratified Units in Western Massachusetts, Western Connecticut, and Southeastern New York..... 343

9.7 Geologic Map of the White Plains-Glenville Area in Connecticut and New York..... 344

9.8 Lithologic Map for the Norwalk North and Norwalk South Quadrangles..... 350

9.9 Orientation Diagram of 133 Sets of Slickensides in the Norwalk North and Norwalk South Quadrangles..... 354

9.10 Geologic Map of Canterbury Gneiss Area..... 355

10.1 Geographical Relationship between the Honey Brook Upland, Chester Valley Region, and Glenarm Terrane of the Southeastern Pennsylvania Piedmont..... 369

10.2 Geologic Map of the Piedmont, Southeastern Pennsylvania..... 371

10.3 Generalized Geologic Map of Manhattan Prong..... 375

TABLES

2.1 Contrasting Characters of Orogens on a Structural Basis..... 11

A.1 Ages of Crystalline Rock Bodies of Maine..... 391

A.2 Ages of Crystalline Rock Bodies of New Hampshire..... 396

A.3 Ages of Crystalline Rock Bodies of Vermont..... 399

A.4 Ages of Crystalline Rock Bodies of Northeastern New York..... 401

A.5 Ages of Crystalline Rock Bodies of Massachusetts..... 402

A.6 Ages of Crystalline Rock Bodies of Rhode Island..... 405

TABLES (Cont'd)

A.7 Ages of Crystalline Rock Bodies of Connecticut..... 406

A.8 Ages of Crystalline Rock Bodies of Southeastern New York, Northern New Jersey, and Southeastern Pennsylvania..... 409

PLATES*

I Major Crystalline Rock Associations and Faults of the Northeasten United States

II Distribution of Described Crystalline Rock Bodies of Maine

III Distribution of Described Crystalline Rock Bodies of New Hampshire

IV Distribution of Described Crystalline Rock Bodies of Vermont

V Distribution of Described Crystalline Rock Bodies of Northeastern New York

VI Distribution of Described Crystalline Rock Bodies of Massachusetts

VII Distribution of Described Crystalline Rock Bodies of Rhode Island

VIII Distribution of Described Crystalline Rock Bodies of Connecticut

IX Distribution of Described Crystalline Rock Bodies of New York, New Jersey, and Pennsylvania

*The plates are available under separate cover in ANL/ES-137, Part 2.

FOREWORD

This topical report was used as a data base for the "Northeastern Regional Geologic Characterization Report," Public Draft, May 1983, prepared by Battelle Memorial Institute, Columbus, Ohio. The U.S. Department of Energy (DOE) sponsored the present work, first through Battelle's Office of Nuclear Waste Isolation (ONWI) and subsequently through the Office of Crystalline Repository Development (OCRD). Project monitors for this Argonne National Laboratory effort were Ronald R. Nicks, Dale E. Markley, and Stephen J. Kowall of ONWI and OCRD. Wyman Harrison of Argonne's Energy and Environmental Systems Division (EES) was the project manager and is responsible for the materials presented by each contributor to the report; he was assisted in project administration by Dorland E. Edgar, also of EES. Profs. Martin F.J. Flower, University of Illinois at Chicago, and Mohan K. Sood, on leave from Northeastern Illinois University, were in charge of the literature survey of crystalline rock complexes.

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Appendix	Ages of Crystalline Rock Bodies Described in This Report	M. Sood
Plate I	Major Crystalline Rock Associations and Faults of the Northeastern United States	P. Barosh, M. Tisue, W. Harrison, M. Sood, and M. Flower

Plate II	Distribution of Described Crystalline Rock Bodies of Maine	M. Flower, M. Sood, G. Kuecher, W. Harrison, and M. Tisue (assisted by B. Bach, H. Levine, S. Lovett, S. Swihart, R. Warner, and C. Witt)
Plate III	Distribution of Described Crystalline Rock Bodies of New Hampshire	M. Flower, M. Sood, G. Kuecher, and M. Tisue (assisted by C. Leitch, H. Levine, and C. Witt)
Plate IV	Distribution of Described Crystalline Rock Bodies of Vermont	M. Flower, M. Sood, G. Kuecher, and M. Tisue (assisted by B. Bach, K. Borgelt, and C. Witt)
Plate V	Distribution of Described Crystalline Rock Bodies of Northeastern New York	M. Flower, G. Kuecher, and M. Tisue (assisted by S. Lovett)
Plate VI	Distribution of Described Crystalline Rock Bodies of Massachusetts	M. Sood, M. Flower, G. Kuecher, W. Harrison, and M. Tisue (assisted by F. Trent and C. Witt)
Plate VII	Distribution of Described Crystalline Rock Bodies of Rhode Island	M. Flower, G. Kuecher, M. Sood, and M. Tisue (assisted by B. Bach and S. Lovett)
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Plate IX	Distribution of Described Crystalline Rock Bodies of New York, New Jersey, and Pennsylvania	M. Flower, G. Kuecher, M. Sood, W. Harrison, and M. Tisue (assisted by B. Bach, H. Levine, F. Trent, and R. Wogelius)

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In addition to the above reviewers, the entire report was reviewed in draft form by personnel of EG&G, Idaho, Inc., and the U.S. Geological Survey, Reston, Virginia. Although these latter reviews were conducted mainly to aid OCRD in its composition of a pre-screening document based on the present report, many technical criticisms and suggestions from these reviewers were incorporated into this report.

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Era	Period	Million Yr Ago*		Northeastern Region Orogenic Association
Ceno-zoic	Tertiary	1.6		
		66		
Mesozoic	Cretaceous		100	
		144		
	Jurassic			
		208	200	
	Triassic			Post-Alleghenian
		245		
Paleozoic	Permian			
		286		Alleghenian
	Pennsylvanian		300	
		320		
	Mississippian			Post-Acadian
		360		
	Devonian		400	Acadian
		408		
	Silurian			
		438		
	Ordovician			Taconic
		505	500	
	Cambrian			
		570		
Precambrian			600	
	Late Proterozoic			Avalonian
			700	
		1000	1000	
	Middle Proterozoic		1100	Grenville
		1600	1600	

*Based on *Decade of North American Geology 1983 Geologic Time Scale*, Geol. Soc. America (1983).

CRYSTALLINE ROCKS OF THE NORTHEASTERN UNITED STATES

by

W. Harrison, M. Flower, M. Sood, M. Tisue, and D. Edgar

Graphics by M. Stahl and B. Termaat

ABSTRACT

This report, which includes a series of maps, is a compendium of the available information on exposed crystalline rocks in Maine, New Hampshire, Vermont, northeastern New York (Adirondack Mountains), Massachusetts, Rhode Island, Connecticut, southeastern New York, northern New Jersey, and southeastern Pennsylvania. Crystalline rocks are defined in this report as bodies of medium- and coarse-grained igneous and high-grade metamorphic rocks. The study was undertaken to provide background information to assist in evaluating the geologic suitability of such rocks for isolating high-level radioactive waste. Available information describing the size, shape, and location; relative and absolute age; origin; petrography; secondary intrusions; country rock; structure; and geophysical characteristics is presented for each crystalline rock body in the study region that has an area of at least 78 km^2 (30 mi^2).

EXECUTIVE SUMMARY

Because of their high strength and relatively low hydraulic conductivity, large bodies of crystalline rocks of igneous and high-grade metamorphic origin are potentially suitable sites for underground repositories for high-level radioactive waste. Sponsored by the U.S. Department of Energy (DOE), this report surveys the available information on the characteristics of crystalline rock bodies of the northeastern United States whose mapped areas, within which outcrops occur, equal or exceed about 78 km^2 (30 mi^2).

Battelle Memorial Institute's Office of Crystalline Repository Development (OCRD) has defined crystalline rocks as medium- and coarse-grained igneous and high-grade metamorphic rocks, and has specified that the region to be discussed in this report comprises Maine, New Hampshire, Vermont, northeastern and southeastern New York, Massachusetts, Rhode Island, Connecticut, northern New Jersey, and southeastern Pennsylvania.

All available information was reviewed to obtain the following specific information on individual crystalline rock bodies: size, shape, and location; relative and absolute age; origin; petrography, including mineralogy, textural relations, and alteration; secondary intrusions; characteristics of the country rock; rock structure; and geophysics. The results of the review are mapped in Plates I-IX and described, state by state, in the body of this report.

Each crystalline body is considered in both a chronologic and tectonic context. Four broad chronological groupings of crystalline rock bodies were established for this report: Middle and Late Proterozoic (about 1600-570 million years [m.y.] ago), Cambrian to Silurian (about 570-408 m.y. ago), Devonian to Pennsylvanian (about 408-286 m.y. ago), and Permian to Cretaceous (about 286-66 m.y. ago). Because the several tectonic and igneous events generally took place over long periods of time, this report discusses the crystalline rock bodies in terms of their relationships to the main tectonic episodes, rather than in terms of their absolute ages. All individual or related plutons and crystalline rock bodies meeting the minimum 78-km^2 (30-mi^2) criterion are grouped by state and then by tectonic association; that is, Grenville and Avalonian, Taconic, Acadian and post-Acadian, or Alleghenian and post-Alleghenian (see Geochronology, p. xx).

Grenvillian and Avalonian crystalline rock bodies, which are the most complex in the northeastern region, crop out in all the states of the region. They are highly folded and faulted, and extremely foliated and fractured, and they have been subjected to high-grade metamorphism. These Precambrian rock bodies consist of complex associations of orthogneiss and paragneiss, granite, tonalite, anorthosite, granulite, syenite, gabbro, and minor amounts of amphibolite. They are often gneissic and migmatitic.

Crystalline rock bodies associated with the Taconic and Acadian-post-Acadian tectonic episodes are distinguishable primarily by composition, fabric, and structure, all of which reflect differences in parental magma composition, fractionation history, emplacement mode, temperature of country rock, and thermotectonic modifications.

The nonstratified Taconic rock bodies exhibit features suggestive of forcible injection and in situ fractionation, and were altered by Acadian metamorphism. Characteristic rock types may be classified as granitic, syenitic, dioritic, gabbroic, and minor ultramafic and alkalic intrusives that show weak to strong tectonic fabric. Many of the Taconic crystalline rock bodies are stratified and composite, containing mafic and felsic fractions.

Acadian crystalline rock bodies are represented by generally uniform gabbro, diorite, and granite. Many gabbros have associated granitic phases. Granitic rocks have igneous flow structure, some of them with tectonic fabrics. The older crystalline rock bodies are gneissic and more elongated than the younger ones, and tend to be oriented parallel to Acadian folds. They often enclose domed masses of the Taconic Oliverian series.

Post-Acadian intrusive rocks are less distinctive than those of the Taconic and Acadian intrusive episodes. The posttectonic plutons are near the earlier syntectonic plutons and are associated with abundant pegmatite. In general, these intrusives are mainly of Late Devonian and Mississippian age and form a sequence of ultramafic, gabbro, norite, granodiorite, monzonite, and granite bodies.

The Alleghenian and post-Alleghenian granitic rocks occupy much of east-central and northern New Hampshire, the western fringes of Maine, southeastern Vermont, and central Rhode Island. Granite is the dominant rock type. Syenite and quartz syenite are common in the form of ring dikes; plug-like bodies of gabbro and diorite occur but are rare.

In addition to differences in composition based on tectonic and chronological associations, syntectonic intrusives generally have moderately to strongly foliated internal structures, while posttectonic intrusives usually have a more isotropic rock fabric.

1 INTRODUCTION

1.1 PURPOSE AND SCOPE

This topical report reviews and analyzes available geologic information relating to crystalline rocks of the northeastern United States. The study was undertaken to help evaluate the suitability of such rock bodies for isolating high-level radioactive waste.

Crystalline rocks are defined herein to include medium- and coarse-grained igneous and high-grade metamorphic rocks. The nine states of the study region (see Plate I) have numerous outcrops of crystalline rocks. The present study summarizes available information on each crystalline rock body whose mapped area equals or exceeds about 78 km^2 (30 mi^2). Subjects treated in this report include the geometrical aspects of the size and shape of the rock bodies; temporal aspects of their relative and absolute age; descriptive aspects of petrography, rock structure, and geophysical signature; and the genetic aspect of rock origin. A companion report (Harrison et al., 1983) describes the geology, hydrology, and mineral resources of the crystalline rock areas of the region. Drafts of these reports were used as sources for a separate report, a review draft of which was published by the Office of Crystalline Repository Development (1983a).

In this report, only those crystalline rock bodies whose mapped areas as shown on Plates II-IX equal or exceed the above-mentioned minimum area are considered. Preparation of the report involved no field work and relied mainly on published literature. Some supplemental information gathered from the public domain and from knowledgeable experts was incorporated to increase the comprehensiveness and utility of the report.

1.2 SPONSORSHIP

This work was sponsored by the U.S. Department of Energy (DOE) to provide background geological information for its Crystalline Rock Project (Office of Crystalline Repository Development, 1983b). The present report is being published on behalf of DOE as a multiple-use resource document of potential benefit to many governmental, educational, industrial, and personal users. As such, it represents a transfer to the public of knowledge gained through expenditure of public funds.

1.3 PLATES

Accompanying this report under separate cover are nine plates. Plate I shows the distribution of major crystalline rock bodies and faults in the northeastern United States. Plates II-IX, which were used to compile Plate I, show the distribution of crystalline rocks in Maine; New Hampshire; Vermont;

northeastern New York (Adirondack Mountains); Massachusetts; Rhode Island; Connecticut; and southeastern New York, northwestern New Jersey, and southeastern Pennsylvania, respectively. In addition, Plates II-IX show the locations of major faults, active and inactive mines, and quarries.

All plates were prepared by the authors of this report from available published information, including the most recent versions of state geologic maps. Boundaries of crystalline rocks shown on state maps were modified, wherever possible, to reflect information obtained from more recent publications. Thus, crystalline-rock boundaries shown on the accompanying plates may not always correspond with those on previously published state geologic maps. Some of these boundaries remain controversial because of local overburden, inadequate field data, or conflicting geologic interpretations. Bedrock maps change over time as new information becomes available, as different geologists assume responsibility for compilations, and as guidelines for the compilation of maps change.

1.4 REFERENCES FOR SECTION 1

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Office of Crystalline Repository Development, *Northeastern Regional Geologic Characterization Report* (review draft), Battelle Memorial Institute, Columbus, Ohio (1983a).

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2 SYNOPSIS OF CRYSTALLINE ROCKS

This synopsis summarizes the chronologic, spatial, and petrogenetic aspects of crystalline rock bodies within the context of tectonic evolution. Crystalline rocks are defined for the purpose of this report to include all medium- and coarse-grained igneous and high-grade regional metamorphic rocks. The pressure-temperature regimes that have affected the crystalline rocks of the region have been predominantly those that produce metamorphism in the range from greenschist to granulite (see Fig. 2.1).

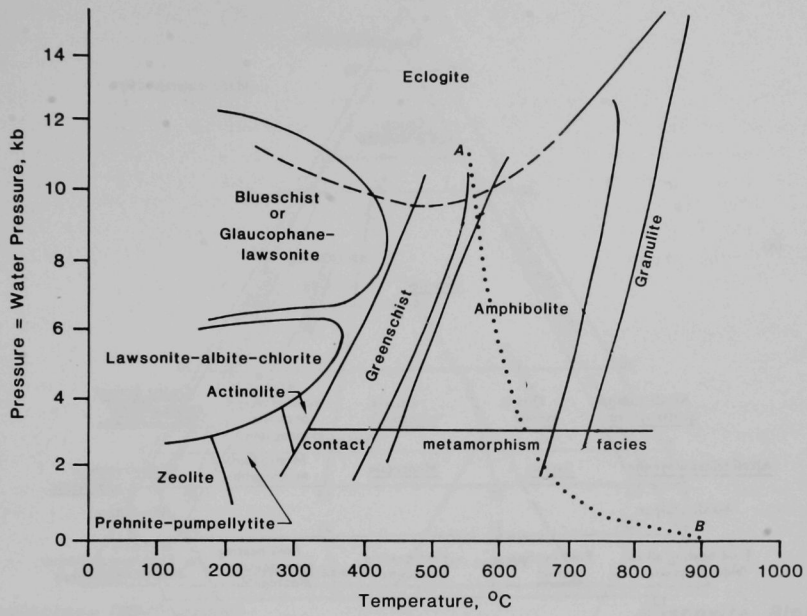
2.1 ROCK NOMENCLATURE

For the detailed geologic discussions in Secs. 3-10, the mineralogic classification of Streckeisen (1976) has been adopted for plutonic igneous rocks (see Fig. 2.2). As an example of how this type of diagram is used, note that a foid syenite (e.g., a nepheline syenite) is a plutonic rock with 10-60% feldspathoid minerals and a plagioclase-to-total-feldspar ratio of less than 10. The classification of Winkler (1976) has been followed for metamorphic rocks. Figure 2.3 defines the compositional domains of high-grade quartzo-feldspathic crystalline rocks. The contrasting chemical and mineralogic characteristics of granitic rocks are probably related to their contrasting tectonic environments.

2.2 GENETIC CLASSIFICATION AND ENVIRONMENT OF CRYSTALLINE ROCKS

Plutonic igneous rocks reflect long and complex emplacement histories (Pitcher, 1978 and 1979), which can be interpreted from petrologic and geochemical information, spatial relationships, and chronologic studies. A spatial and temporal connection is recognized between regional deformation and the generation and emplacement of granitic magma (Read, 1957; Stephansson, 1975). This relationship is usually referred to in terms of epizonal, mesozonal, and catazonal associations, and reflects the ductility contrast between an intrusive body and country rock (Buddington, 1959; Hamilton and Myers, 1967). This ductility contrast involves three major variables: lithologic contrast, relative crustal level of emplacement, and the time gap between intrusion and regional metamorphism.

Granite formed through "ultrametamorphism" is a possibility, but the connection may not be direct because of the often extensive period between metamorphism and intrusion. Experimental studies (Wyllie, 1977) suggest that water content and temperature are not high enough in anorogenic settings to permit crustal remelting without supplemental heat from mantle-derived magma (Brown and Hennessy, 1978; Brown, 1979). However, in the orogenic environment, breakdown of mica during prograde metamorphism of a sedimentary wedge may provide a substantial amount of water, which depresses the



Typical Mineralogical Characteristics of Facies

Zeolite	Quartz + laumontite (zeolite) + chlorite
Blueschist	Glaucophane + lawsonite
Greenschist	Albite + epidote + actinolite ± chlorite
Amphibolite	Hornblende + plagioclase (An > 20) in mafic rocks and kyanite in pelitic rocks
Granulite	Augite - orthopyroxene + Mg-Fe garnet
Eclogite	Feldspar-free assemblage (omphacitic pyroxene ± (Mg,Fe)Al garnet in mafic rocks)

Fig. 2.1 Gradational Boundaries of Regional Metamorphic Facies (P_{H_2O} = P, except in the low-temperature range of the eclogite facies; dashed boundary indicates limits of the eclogite facies under conditions of low water pressure [P_{H_2O} < P]; the dotted curve [AB] shows the beginning of melting in the simplified granite system quartz-orthoclase-albite in the presence of excess water) (Source: Adapted from Turner, 1981)

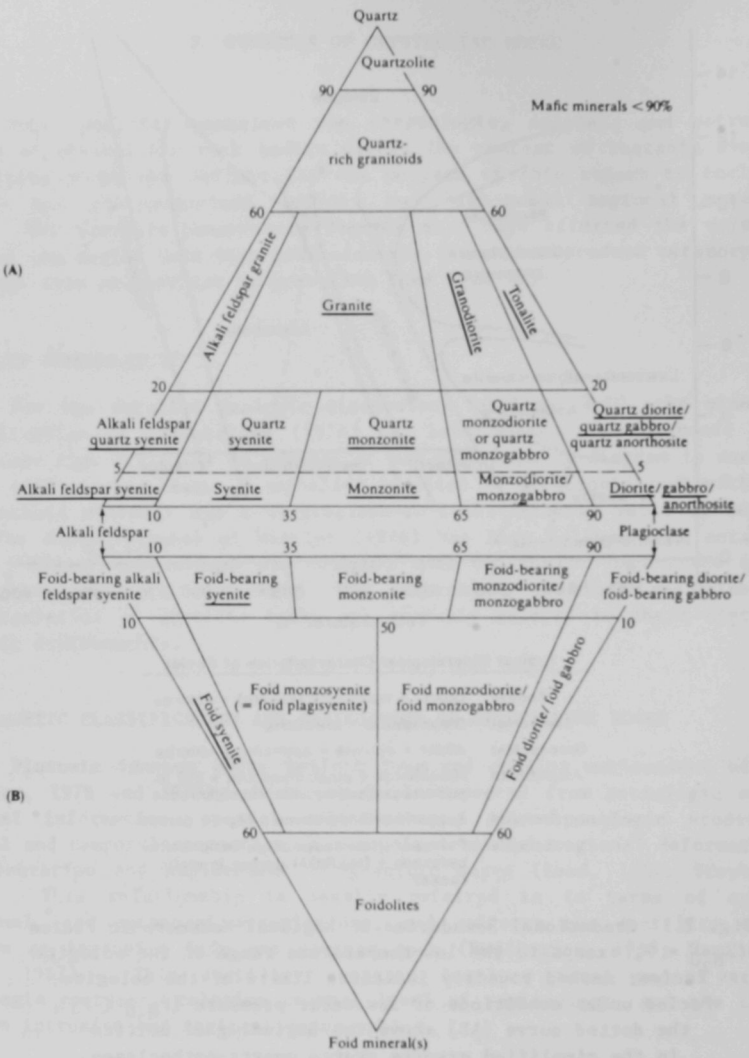
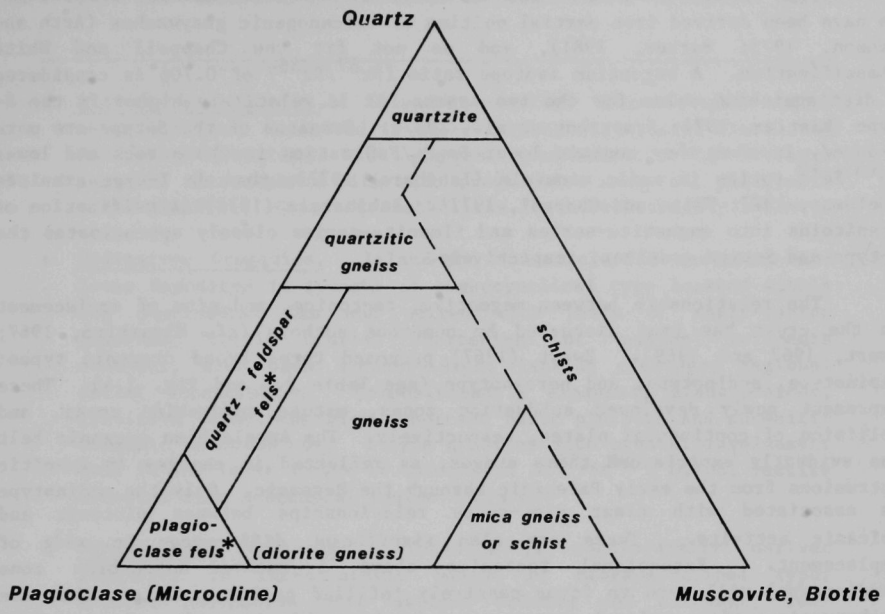


Fig. 2.2 Classification of Plutonic Igneous Rocks (to be included in these triangles, an igneous rock must be macroscopically crystalline and must contain at least 10% plagioclase, alkali feldspar, and either quartz [triangle A] or feldspathoid minerals [triangle B]; rock names commonly referred to in this report are underlined) (Source: Streckeisen, 1976)



*Fels implies a massive (nonfoliated) rock.

Fig. 2.3 Composition of Metamorphic Rocks of Higher Temperature Ranges (Source: Adapted from Winkler, 1976)

melting point of the rock, bringing it within the range of the prevailing geothermal conditions. In general, granitic magma is generated in harmony, in both time and space, with its metamorphic environment (Pitcher, 1979).

Chemical and mineralogic studies define two main granite series, each of which is characteristic of a type of orogenic belt. Both series are calc-alkaline, but each is represented by a distinctive compositional range. The approximate ratio of gabbro-diorite to tonalite-granodiorite to granite is typically 15:50:35 for the first type and 2:18:80 for the second (Pitcher, 1979).

Chapell and White (1974) differentiated granite types into S-type and I-type, by source material. The S-type is peraluminous (Shand, 1951), biotite-muscovite granite and is probably produced through prograde metamorphism of pelitic metasedimentary material. In comparison, the I-type is metaluminous, biotite-hornblende granite derived from basic igneous material or melting of deep-seated igneous material. However, voluminous

Precambrian shield granites, both metaluminous and peraluminous, are thought to have been derived from partial melting of volcanogenic graywackes (Arth and Hanson, 1975; Barker, 1981), and do not fit the Chappell and White classification. A strontium isotope ratio (Sr^{87}/Sr^{86}) of 0.706 is considered a distinguishing value for the two types: it is relatively higher in the S-type (Kistler, 1974; Armstrong et al., 1977). Granites of the S-type are more reduced, in that they contain lower Fe_2O_3/FeO ratios in whole rock and lower Fe^{+3}/Fe^{+2} ratios in mafic minerals (Ishihara, 1977), than do I-type granites (Coleman, 1977; White and Chappel, 1977). Ishihara's (1977) classification of granitoids into magnetite-series and ilmenite-series closely approximates the I-type and S-type granites, respectively.

The relationship between magmatism, tectonics, and site of emplacement in the crust has been discussed by numerous authors (cf. Miyashiro, 1967; Zwart, 1967 and 1969). Zwart (1967) proposed three broad orogenic types: alpinotype, andinotype, and hercynotype (see Table 2.1 and Fig. 2.4). These represent newly developed subduction zones, mature subduction zones, and collision of continental plates, respectively. The Appalachian orogenic belt has evidently experienced these stages, as reflected in changes in granitic intrusions from the early Paleozoic through the Mesozoic. Only the andinotype is associated with clear time-space relationships between plutonic and volcanic activity. There are also significant differences in mode of emplacement. Extensional tectonics above a mature subduction zone (andinotype) will tend to favor passively infilled cauldron plutons that are disharmonious with regional structures and metamorphic isograds. Compressive, ductile tectonic regimes of intracontinental orogens (hercynotype) will favor emplacement as diapirs that are harmonious with tectonic lineation and metamorphic isograds. Mode of emplacement, in addition to the subsequent tectonic and metamorphic overprint, is an important determinant of the rock fabric as the magma cools.

2.3 CHRONOLOGIC AND TECTONIC CLASSIFICATION OF CRYSTALLINE ROCKS IN THE NORTHEASTERN UNITED STATES

Each crystalline rock body in this study is considered in both a chronologic and tectonic context. Four broad chronologic groupings of crystalline rock bodies were defined for this report: Middle and Late Proterozoic (about 1600-570 million years [m.y.] ago), Cambrian to Silurian (about 570-408 m.y. ago), Devonian to Pennsylvanian (about 408-286 m.y. ago), and Permian to Cretaceous (about 286-66 m.y. ago). Plate I shows the locations of crystalline rock bodies of these groupings. As both tectonic and igneous events generally occurred over considerable periods of time, this report, for consistency, discusses the crystalline rock bodies in terms of their relationships to the main tectonic episodes rather than in terms of their absolute ages. All individual or related plutons meeting the minimum 78-km^2 (30-mi^2) criterion are grouped by state and then by tectonic association; that is, Grenville and Avalonian, Taconic, Acadian and

Table 2.1 Contrasting Characters of Orogens on a Structural Basis

- Alpinotype Orogenies. Island-arc-derived volcaniclastic sediments and lavas deposited in oceanic trenches; crustal shortening involving thrusting with nappes predominant; high-pressure regional metamorphism with wide progressive zonation; ultrabasic rocks abundant. Granitoid batholiths characteristically absent.

- Andinotype Orogenies. Island-arc volcaniclastic sediments and lavas deposited in troughs of eugeosynclinal type located within the continental lip and paired with belts of shelf-facies clastics; little crustal shortening but vertical movements dominant, with open drape-folding, lacking cleavage; regional burial metamorphism. Compositionally expanded, disharmonious, granitoid batholiths with important basic plutonic and andesitic volcanic associations. I-type granites with crustal involvement only in the later stages of evolution; $Sr^{87}/Sr^{86} < 0.706$; restite material from subcrustal source included.

- Hercynotype Orogenies. Nonvolcanic, continentally derived sediments in intracratonic basins of miogeosynclinal type; crustal shortening with upright folding and cleavage; low-pressure metamorphism with prograde zonation; ultrabasic rocks rare. Compositionally contracted, harmonious, granitoid batholiths with only minor basic association and generally lacking contemporaneous volcanics. S-type or mixed S- and I-type granites; Sr^{87}/Sr^{86} ratio > 0.706; inherited xenocrysts derived from recycled crustal rocks.

Source: Pitcher, 1979, partly after Zwart, 1969.

post-Acadian, or Alleghenian and post-Alleghenian (see Geochronology, p. xx). Lyons et al. (1982) proposed a continuous magmatism from Taconic to Acadian, and also a relative prominence of Alleghenian events in eastern New England.

The plutons associated with each of these tectonic episodes may be distinguished by mineralogic and chemical composition, fabric, and structure, all of which can be attributed to differences in parent magma composition, fractionation history, emplacement mode, wall-rock temperature, and especially tectonic environment (Page, 1968).

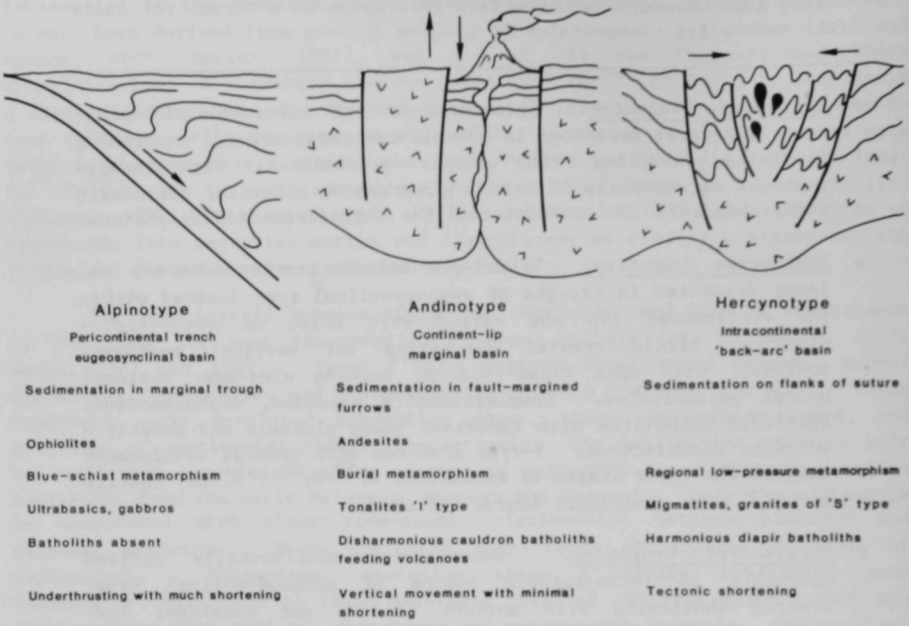


Fig. 2.4 Schematic Representation of Three Orogenic Environments of Granitic Rocks (Source: Adapted from Pitcher, 1979)

2.3.1 Grenville and Avalonian

In the northeastern region, Precambrian intrusives of Grenville and Avalonian age are almost always metamorphosed. They are faulted, extremely fractured, and invariably foliated.

Grenville rocks form the metamorphosed core of western New England and northeastern New York; in contrast, Avalonian rocks are the basement in eastern New England. The degree of structural and lithologic complexity is relatively greater in the Grenville than in the Avalonian rocks. In Maine, the Chain Lakes massif is a portion of 1600-m.y.-old crust (Naylor, 1975).

In the Adirondack region of northeastern New York, rocks 1050-1300 m.y. old are well exposed in a dome-shaped body. Anorthosite-gabbro, granite-granitic gneiss, and syenite-quartz syenite (mangerite-charnockite) gneiss lithologies are well represented. All of these lithologies were modified by the high-grade metamorphism of the Grenville orogeny, which produced complex and heterogeneous lithologic characteristics and strong tectonic fabrics. The rocks display a well-developed, northeast-trending structural grain and east-

northeast-trending high-angle faults. Lithologically similar bodies have discontinuous surface expressions, with some having collective areas of 2000-4300 km^3 (700-1700 mi^3). Gravity contrast data suggest that the anorthosite bodies extend to depths as great as 32 km (20 mi), while the Alexandria granite gneiss extends to 5 km (3 mi). Anorthosites also exhibit heat flows of 0.80×10^{-6} $cal/cm^2 \cdot s$, which is lower than the average for shield rocks of about 1×10^{-6} $cal/cm^2 \cdot s$.

The Precambrian crystalline rock bodies underlying the Hudson, Housatonic, and Berkshire highlands are mostly 950-1300 m.y. old. They form the metamorphosed core of the Appalachian orogen, the western margin of which marks a prominent gravity high consistent with the northeast-trending tectonic fabric. In addition to the common intrusive lithologies, there are complex and heterogeneous gneisses, amphibolites, and migmatites. The Ramapo-Canopus fault zone is an important structural feature in the Hudson Highlands, which separates it into eastern and western lithic zones (Hall, 1980). The charnockitic and alaskitic gneisses in the eastern Hudson Highlands demonstrate their petrologic affinity to those of the Adirondacks. The Housatonic and Berkshire lithologies are somewhat similar, but the latter highland is structurally more complex.

Crystalline rocks of Avalonian age (550-600 m.y.) occur to a limited extent in the highland areas, but they form a significant portion of the basement in eastern New England. Although they are largely interlayered tonalites and amphibolites, they are also represented by feldspathic and granitic gneisses and amphibolites in coastal Maine, and by gabbro, diorite, granodiorite, and granite in coastal Massachusetts and Rhode Island.

2.3.2 Taconic

Crystalline rock bodies associated with the Taconic orogeny may be pretectonic, syntectonic, or posttectonic; they were intruded into unmetamorphosed Cambrian and Ordovician sediments under unique conditions. Many of these plutons belong to the Highlandcroft and Oliverian magma series that show internal structures (e.g., foliation or lineation that conforms to a dome).

In general, Taconic magmas seem to have been forcibly injected, fractionated in situ, and metamorphosed during the subsequent Acadian orogeny. Early foliation is marked by grains of biotite, feldspar, and quartz; later foliation is marked by fine-grained aggregates. The textures of the plutons are a composite of relict igneous and superimposed metamorphic fabrics. Recrystallized secondary minerals are much finer grained than primary minerals. Gneissose wall rock is often volcaniclastic in origin, but may not show cataclastic structures because of complete metamorphic recrystallization. In the vicinity of faults of Permian-Triassic age, Taconic bodies may show a third local foliation that is at variance with the other two. In general, these bodies are compositionally diverse and contain both mafic and felsic fractions. They may be porphyritic and equigranular, and are

distinguishable from other pre-Devonian intrusives by their retention of primary igneous structures. Aplite and pegmatite veining is present.

2.3.3 Acadian and Post-Acadian

The Acadian to post-Acadian rock bodies of the New Hampshire plutonic series, in New Hampshire and Maine, form sheetlike or blocklike intrusives. Individually, they tend to be elongate, and the group forms a flattened oval elongated to the northeast. In general, Acadian rock bodies tend to have gneissose margins with coarser, uniform central portions. Contact metamorphic effects due to high temperatures are present.

Posttectonic members of the series occur side-by-side with syntectonic members. Emplacement by stoping and intrusion along fractures and Acadian weak zones is typical, as are faint foliation and medium- to fine-grained textures. Abundant pegmatite also is characteristic. Although these members are similar to the Acadian plutons, they show little evidence of forceful injection. Also, the contact metamorphism that they produce in the host rock tends to be retrograde instead of prograde metamorphism.

In New Hampshire, crystalline rocks of this age and tectonic association are the early- to late-tectonic tonalite-granite of the Kinsman and Bethlehem intrusives; the gabbro-monzonite of the Spaulding series; and the highly aluminous, two-mica, Concord-type post-Acadian granites. In Maine, such rocks are gabbroic but have associated granite phases. The easternmost intrusions of this age are alkalic. In overall characteristics, the Acadian and post-Acadian crystalline rock bodies are of relatively uniform lithologies and somewhat isotropic. The older bodies sometimes exhibit prominent flow features, but the younger ones lack foliation.

2.3.4 Alleghenian and Post-Alleghenian

The late Paleozoic to Mesozoic bodies of Alleghenian and post-Alleghenian association are the least common in the region. They occur most frequently as cylindrical intrusions, as indicated by gravity data; the most prominent of them are members of the White Mountain magma series, which are mainly alkalic granite, syenite, and monzonite bodies. The intrusions are relatively homogeneous and isotropic.

2.4 CAVEATS

Problems abound in interpreting stratigraphic relations and also in comparing radiometric age data derived over two decades at different laboratories by different methods. Such problems are compounded by the various tectonic episodes and the isotopic reequilibration that may have occurred. In addition, distinguishing between primary igneous foliation and

tectonic foliation is often difficult, and intrusive contact relationships are frequently ambiguous. In this report, syntectonic and posttectonic pluton associations based on significant differences in plutonic rock fabric have been emphasized, but with the understanding that some readers may disagree with some of the designations.

2.5 PRESENTATION OF INDIVIDUAL PLUTON REPORTS

Sections 3-10 present the available information on crystalline rocks in Maine, New Hampshire, Vermont, the Adirondack area of northern New York, Massachusetts, Rhode Island, Connecticut, and the area contained by southeastern New York, southeastern Pennsylvania, and northern New Jersey. Within each state or area, those crystalline rock bodies or complexes that met the minimum-area criterion of 78 km^2 (30 mi^2) were numbered, beginning in the southwest corner of the state or area and proceeding clockwise; these are the numbers by which the rock bodies are identified on the plates accompanying this report, and the numbers have no other significance. In the text section for each state, the plutons are discussed alphabetically within the tectonic categories specified above in Secs. 2.3.1-2.3.4. The appendix provides a chronological list of the crystalline rock bodies in each state, according to radiometric or lithostratigraphic age.

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3 CRYSTALLINE ROCKS OF MAINE

3.1 GRENVILLE AND AVALONIAN

3.1.1 Chain Lakes Massif (Plate II, Pluton 21)

3.1.1.1 Size, Shape, and Location

The Chain Lakes massif is a roughly oval composite body located on Maine's northwestern Canadian border. Boudette and Boone (1976) report that the massif forms the core of the Boundary Mountains anticlinorium. The massif is a layered sequence of gneissic granofels, gneiss, schist, gneissic quartzite, and thinly layered amphibolite. This body is about 3100 m (10,170 ft) thick, and outcrops are found over an area of about 860 km^2 (332 mi^2).

3.1.1.2 Relative and Absolute Age

On the basis of stratigraphic position, these rocks have been considered to be pre-Silurian. However, Naylor et al. (1973) reported a Pb^{206}-Pb^{207} radiometric age of 1500-1600 m.y. for the Chain Lakes massif. An ophiolite body of Cambrian or Ordovician age is tectonically attached to its southern margin and is referred to as the Boil Mountain complex by Boudette (1982).

3.1.1.3 Origin

The mode of emplacement has been obscured by subsequent tectonic events. The rocks were derived from a volcanic-sedimentary protolith subjected to amphibolite-granulite facies metamorphism, perhaps during the Precambrian (Boudette, 1978). Following metamorphism, the massif was probably rafted to its present position during the Taconian-Caledonian event as a detached fragment of Europe or Africa (Boudette, 1982).

3.1.1.4 Petrography

Mineralogy and Textural Relations. The Chain Lakes massif may be considered to be composed of two units. The lower unit is overlain by gneiss. It is predominantly massive, light to medium gray, medium to coarse grained, and chloritic and sericitic. The rock might be called a granofels, although it contains some thin layers and laminations; it is a quartz-feldspar-sillimanite-muscovite-biotite gneiss, with inclusions of fragments of older rock a few centimeters across, and quartz nodules (Boudette and Boone, 1976).

The lower unit grades upward into a more systematically bedded, lithologically heterogeneous upper unit. However, gneiss that is mineralogically comparable to the lower unit, but evenly layered, is abundant in the upper unit. The layers vary from 1 cm (0.4 in.) to 10 cm (4 in.) thick, and have internal laminations. In addition to the gneiss, the upper unit also contains chloritic and sericitic quartzofeldspathic schist, thick-bedded feldspathic quartzite, felsic metavolcanic rocks, thick-bedded and laminated quartzite, black schist, and finely laminated amphibolite. The gneiss, schist, and quartzite commonly are intergradational (Boudette and Boone, 1976).

The Boil Mountain complex rocks, though not part of the Chain Lakes massif, include serpentinite, pyroxenite, gabbro, and epidiorite, which are generally highly altered; however, unaltered examples occur in the northeastern part.

Alteration. The entire massif was metamorphosed to sillimanite grade during Precambrian time. The rocks have been further altered by later hydrothermal and metamorphic events.

3.1.1.5 Secondary Intrusions

An ophiolite sequence, the Attean quartz monzonite, and the Seven Ponds pluton are the major secondary intrusive phases in the massif.

3.1.1.6 Country Rocks

The Chain Lakes massif is in contact with the Ordovician Attean quartz monzonite to the north and northwest, the Cambrian-Ordovician Boil Mountain complex to the south, and Devonian intrusives to the west (see Fig. 3.1).

3.1.1.7 Rock Structure

Where rocks of the massif are gneissic or schistose, foliation is parallel to the compositional layering. Relatively abundant, small fragments of metamorphic rocks within these layers have a foliation that varies in angular divergence from the regional foliation, which postdates that in the fragments (Boudette and Boone, 1976). Fracturing is extensive (Albee and Boudette, 1972).

The Chain Lakes massif forms the core of the Boundary Mountains anticlinorium. The regional structural trend in this area is consistently northeast to southwest.

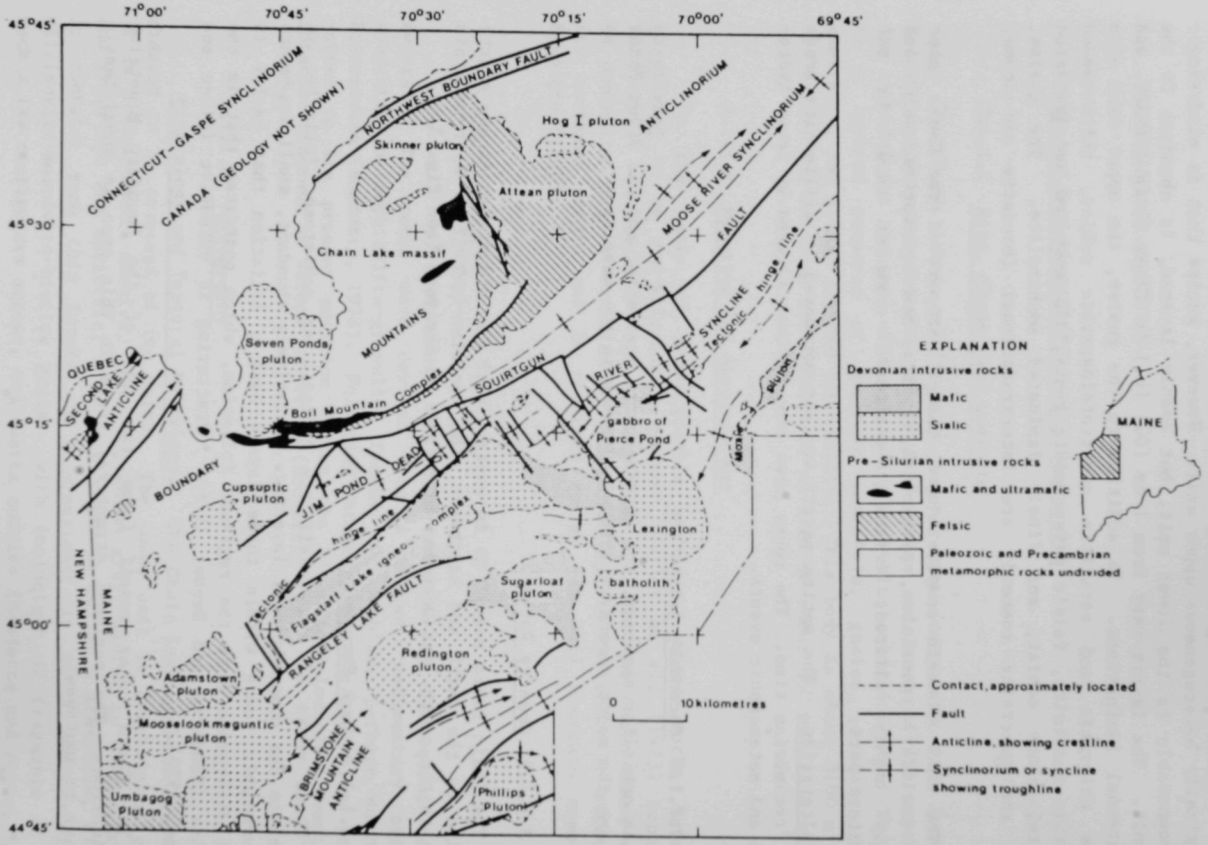

Fig. 3.1 Major Tectonic Features and Plutons of the Rangeley Lakes-Dead River Basin (Source: Boudette, 1982. Copyright 1982, Geol. Assoc. Canada; reprinted with permission)

3.2 TACONIC

3.2.1 Attean Quartz Monzonite (Plate II, Pluton 22)

3.2.1.1 Size, Shape, and Location

The Attean quartz monzonite crops out in the Attean, Spencer, Long Pond, Pierce Pond, and Skinner quadrangles. Outcroppings are found throughout an irregularly shaped area about 32 km (20 mi) long and 29 km (18 mi) wide, covering about 435 km^2 (168 mi^2). It is located on the crest and southeastern limb of the Boundary Mountain anticlinorium, adjacent to the Chain Lakes massif.

3.2.1.2 Relative and Absolute Age

An Ordovician age is postulated for the Attean quartz monzonite (Albee and Boudette, 1972). It has been dated at 356 ± 15 m.y., on the basis of K-Ar ratios in biotite, and at 395-500 m.y. on the basis of the Pb-α method (Faul et al., 1963). Boone et al. (1970) and Boudette and Boone (1976) quote a U-Pb zircon age of about 470 m.y. for the pluton, thus making it equivalent to the Highlandcroft plutonic series (Billings, 1956) of New Hampshire.

3.2.1.3 Origin

Attean quartz monzonite is close to, and may be co-intrusive with, the mafic Alpine complex. Together, they form an elongate intrusive prism. The Attean may be petrogenetically related to the Alpine complex, representing the late lithic phase of the Ordovican igneous activity that produced the Alpine.

3.2.1.4 Petrography

Mineralogy and Textural Relations. The Attean quartz monzonite is a mottled pink-and-green, medium- to coarse-grained porphyritic rock. Phenocrysts are tabular potassium feldspar grains 1-5 cm (0.4-2 in.) in length. The groundmass has a uniform grain size and consists of quartz and plagioclase in equant grains 3-5 mm (0.1-0.2 in.) across. The texture is massive, with little alignment of feldspar grains. Major minerals include quartz (25-30%), potassium feldspar (30-35%), plagioclase (35-40%), and mafics (5-10%). Accessory minerals include hornblende, biotite, chlorite, apatite, ilmenite, sphene, magnetite, calcite, apatite, and zircon. A highly altered version of the rock is found in the north and west of the outcrop area. In this facies the rock has a cataclastic schistosity due to alignment of quartz and mica grains. Quartz grains are granulated and give the rock a hobnailed appearance. The overall rock is green and has a mortar structure (Albee and Boudette, 1972).

Alteration. The degree of alteration of the Attean quartz monzonite varies from place to place, being greater to the north and west in the Attean quadrangle. Alteration of plagioclase has produced white mica, clay, and clinozoisite-epidote. Micaceous minerals and quartz may be aligned, producing a micaceous schistosity. Indications of cataclasis include mortar structure, undulatory extinction, and granulation in quartz (Albee and Boudette, 1972).

3.2.1.5 Secondary Intrusions

Dark-gray mafic dikes as wide as a few meters (several feet) cut the Attean quartz monzonite. The dikes are very fine grained, and consist of aggregates of hornblende, chlorite, clinozoisite-epidote, plagioclase, quartz, white mica, ilmenite, and sphene (Albee and Boudette, 1972).

The Attean quartz monzonite is intruded into by the Hog Island granodiorite of Devonian age. This is a light-gray, medium-grained rock containing plagioclase (40-55%), potassium feldspar (microperthite) (15-20%), quartz (20-35%), and biotite (5%). The plagioclase and quartz are finer grained (1-2 mm, or 0.04-0.08 in.) than the potassium feldspar (4-6 mm, or 0.16-0.24 in.). The rock is massive and slightly jointed (Albee and Boudette, 1972).

3.2.1.6 Country Rocks

The Attean quartz monzonite was intruded into pre-Upper Silurian granofelsic and dioritic rock masses of the Chain Lakes massif.

3.2.1.7 Rock Structure

The Attean quartz monzonite contains abundant joints with variable spacing. Few of these joints contain slickensides, mineral filling, or alteration. Several schistose shear zones are parallel to and appear to be associated with the joint systems. The dip of most joints is vertical but some are horizontal. The strike of jointing is variable, but a common set appears to strike about north-south. A fault striking north-northwest occurs in the southern half of the Attean quadrangle. The dip of the fault plane is nearly vertical, and the east side is downthrown (Albee and Boudette, 1972).

A secondary schistosity is present in the northwestern part of the outcrop (Albee and Boudette, 1972). No folding has been observed in the Attean quartz monzonite. Because planar features are very rare in this rock, observations of folding are difficult (Albee and Boudette, 1972).

3.2.1.8 Geophysics

The Attean quartz monzonite has a magnetic susceptibility that is lower than the values for the older granofels and diorite. The magnetic pattern

over the monzonite is generally uniform, although higher values appear to be associated with the granofels roof pendants. In this region, topographically low areas tend to correspond with magnetically low areas underlain by the Attean quartz monzonite (Albee and Boudette, 1972).

3.3 ACADIAN AND POST-ACADIAN

3.3.1 Biddeford Pluton (Plate II, Pluton 3)

3.3.1.1 Size, Shape, and Location

The Biddeford pluton lies on the coast of Maine not far from the New Hampshire border, in the Biddeford 7.5-min quadrangle and the northeastern corner of the Kennebunk 15-min quadrangle. The pluton is elongated about 14 km (9 mi) in an approximately northwesterly direction inland; it is about 10 km (6 mi) wide in its dry-land exposure. Outcrops appear in a roughly oval area of more than 102 km^2 (39 mi^2). The Biddeford pluton is just southeast of the Lyman pluton.

3.3.1.2 Relative and Absolute Age

The Biddeford pluton does not appear to belong to the New Hampshire plutonic series; it is probably younger (Doyle et al., 1967). Lyons et al. (1957) established an age of 303 m.y. for the granite of the Biddeford pluton by use of Pb-α analysis, and Gaudette et al. (1975) determined an age of 345 $\pm$ 12 m.y. by Rb-Sr analysis in whole rock. Page (1968) also suggested that the Biddeford pluton resulted from an igneous event more recent than the activity that created the New Hampshire series.

3.3.1.3 Origin

The Biddeford pluton, with others assigned to a Late Devonian(?) plutonic series by Page (1968), is thought to have been emplaced by forceful injection and stoping, taking the form of cylindrical stocks and sill-like structures. In the territory north and south of the Biddeford pluton, numerous stringers of granitic rock have intruded into the country rock and produced isolated roof pendants, as evidenced by the attitude of the bedding in these blocks.

3.3.1.4 Petrography

Mineralogy and Textural Relations. Hussey (1962) described the Biddeford pluton as considerably more homogeneous in color, texture, and

composition than either the Lyman or Webhannet plutons. It consists of a light-gray to pinkish biotite granite, grading into a quartz monzonite. The Biddeford pluton is medium to coarse grained and evenly textured.

Alteration. Alterations of this pluton are not mentioned in the literature.

3.3.1.5 Secondary Intrusions

Information on secondary intrusions is lacking.

3.3.1.6 Country Rocks

The Biddeford pluton contacts the metamorphosed calcareous rocks of the Berwick formation; the metasediments of the Kittery formation, which contain calcareous sandstone, siltstone, and graywacke; and the metasediments of the Eliot formation (Doyle et al., 1967).

3.3.1.7 Rock Structure

The Biddeford pluton shows no significant foliation toward its northwestern end, while its southeastern extension, especially near the coast, shows moderate to strong foliation.

Three major faults in Southern Maine run parallel to the regional structural fabric. Hussey and Newburg (1978) interpreted the Nonesuch River fault as a right lateral fault. The topographic lineament in the Lyman pluton may be related to the reactivation of the fault after emplacement.

3.3.2 Bottle Lake Quartz Monzonite (Plate II, Pluton 27)

3.3.2.1 Size, Shape, and Location

The Bottle Lake complex is located in the Grand Lake area of east-central Maine about 22 km (14 mi) from the New Brunswick border. It extends for about 45 km in the east-west direction (Ayuso, 1979) with its center at the junction of Penobscot, Hancock, and Washington counties.

The Bottle Lake complex has a mapped area of about 1100 km^2 (425 mi^2). Its shape is roughly that of two overlapping circular bodies, with a narrow tail protruding about 15 km (9 mi) to the northeast. Folded, low-grade metasediments of the Merrimack synclinorium surround the complex (Ayuso and Wones, 1980). The Norumbega fault zone is adjacent to the complex on its southern side, truncating the lower edge of the eastern lobe.

3.3.2.2 Relative and Absolute Age

The Bottle Lake complex consists of two related intrusions. The western lobe of the complex, called the Passadumkeag River granite, yielded a Rb-Sr whole-rock age of 360 ± 16 m.y. (Ayuso and Wones, 1980). The eastern lobe of the complex, the Whitney Cove granite, gave a K-Ar date, in biotite, of 370 m.y., and a Pb-α age, in zircon, of 380 m.y. (Faul et al., 1963). The northeast extension of the Bottle Lake complex is thought to be part of the Whitney Cove granite, which was dated by Ayuso and Arth (1983) at 380 ± 5 m.y. by the Rb-Sr method applied to whole rock, and at 393 ± 11 m.y. by the Pb-Pb method on zircon. Local structural relationships indicate a relative age for the Bottle Lake complex of a bit later than Early Devonian.

3.3.2.3 Origin

Larrabee et al. (1965) mentioned the occurrence of country rock inclusions in the western lobe of the complex near the southern margin. On the basis of field relations, Ayuso (1979) proposed that the eastern body was emplaced before the western body. Ayuso and Wones (1980) mentioned numerous inclusions in the core of the Passadumkeag River granite (the western lobe). A well-developed contact aureole also indicates fluid emplacement. This intrusion is clearly posttectonic (Loiselle, 1981).

3.3.2.4 Petrography

Mineralogy and Textural Relations. The Bottle Lake complex consists of two major phases -- the granite of Whitney Cove and the granite of the Passadumkeag River. The Passadumkeag River granite has two distinct facies with different textures — core and rim. The core facies is a gray, coarse-grained, porphyritic hornblende-biotite quartz monzonite with rapakivi texture. The rim facies is a pink hornblende-biotite granite that is seriate to equigranular. The Whitney Cove granite also has core and rim facies. The core is a pink, medium- to coarse-grained, anhedral, seriate biotite granite with equidimensional texture. The Topsfield facies, which is related to the Whitney Cove rim facies, is very similar both mineralogically and texturally to the rim facies (Ayuso and Wones, 1980).

For the Bottle Lake complex as a whole, alkali feldspar is about Or-85-95 in composition, while plagioclase crystals, which are zoned normally, range in composition from An-0 to An-50 (Ayuso and Wones, 1980). Small amounts of magnetite, ilmenite, allanite, sphene, zircon, apatite, and sulfides are present (Ayuso and Wones, 1980).

Alteration. Secondary alteration of the Bottle Lake complex is not prominent, although some secondary metamorphic recrystallization is evident, and could be related to the intrusion of the Passadumkeag River magma (Ayuso, 1979; Ayuso and Wones, 1980).

3.3.2.5 Secondary Intrusions

The Whitney Cove granite contains aplite and pegmatite dikes, generally less than 6 mm (0.25 in.) in width. These dikes are more abundant near the granite contacts.

3.3.2.6 Country Rocks

The Bottle Lake complex is bounded on all sides except the northeast by what is thought to be the Vassalboro formation, a calcareous pelite of early Silurian age (Osberg, 1979). On the northeast are undifferentiated volcanics and pelitic siltstones thought to be Late Ordovician to Early Silurian, and a unit of graywackes, siltstones, and slates of Late Cambrian to Early Ordovician age (Ayuso and Wones, 1980). The metamorphic aureole of the complex extends away from the body generally less than 1.2 km (0.75 mi) (Larrabee et al., 1965).

3.3.2.7 Rock Structure

Although joints and fractures are abundant, they seem to have no preferred orientation. In the Grand Lake area, the Whitney Cove granite is cut by the left lateral Norumbega fault system and a northeasterly cataclastic zone. An east-to-west fault, southeast of East Musquash Lake, bisects the two bodies of the pluton (Ludman, 1978; Ayuso, 1979). The northeasterly Norumbega fault system is a major regional structure, 3-4 km (2-2.5 mi) wide, that contains subparallel faults (Wones and Thomson, 1979) The associated fabric suggests that the principal comprehensive stress was from east to west.

Feldspars form limited foliation zones with northeasterly orientation in the core of the Passadumkeag River granite. Randomly oriented foliation zones, probably flow structures, also occur in the Whitney Cove granite. No folding in or associated with the Bottle Lake complex has been reported.

3.3.2.8 Geophysics

A magnetic survey was completed in the Bottle Lake area and was used to trace covered contacts (Larrabee et al., 1965). Studies by Hodge et al. (1982) suggest that the Lead Mountain and Bottle Lake batholiths connect underground, as they show a continuous Bouguer anomaly that indicates a body 8-10 km (5-6 mi) thick.

3.3.3 Cadillac Mountain and Somesville Plutons (Plate II, Pluton 39)

3.3.3.1 Size, Shape, and Location

Together, Cadillac Mountain and the Somesville pluton make up a roughly circular, stock-like unit of rocks extending over an area of about 118 km^2 (45 mi^2). They constitute the center, and most of the area, of Mount Desert Island, between Blue Hill Bay and Frenchman Bay on the coast of Maine.

3.3.3.2 Relative and Absolute Age

Chapman (1969) proposed that the Somesville pluton intruded into the Cadillac Mountain granite and is therefore younger, but both Cadillac Mountain and the Somesville pluton are Devonian in age (also see Doyle et al., 1967). Faul et al. (1963) have given a K-Ar biotite age of 389 m.y. for the Somesville pluton in Hall's Quarry, and an age of 400 m.y. for granite of Cadillac Mountain (Great Head).

3.3.3.3 Origin

Chapman (1969) considered the Cadillac pluton to have formed by crystallization of a granitic magma. He based this conclusion on the decrease of grain size toward the contact with the country rock, the fact that the granite cuts across the country rock and fills fractures in it, the angular, diversely oriented inclusions of country rock, and the perthitic feldspar that appears inverted from a high-temperature form. The magma would have been emplaced in a large chamber by ring fracturing and cauldron subsidence, producing a saucer-like structure. This mode is suggested by the subcircular outline of the plutons, the smoothly curved but sharp contacts, the paucity of granitic apophyses, the presence of ring dikes that intrude into the shatter zone along the contact, the existence of the shatter zone, the web-vein and web-breccia pattern in the country rock, the external centripetal dip pattern, and the chilled contacts representing rapid ascent of magma.

Light-colored inclusions, rich in quartz and alkali feldspar, were derived from stratified rocks of the Bar Harbor series and felsic volcanic material. Dark-colored inclusions, rich in plagioclase and hornblende, represent fragments of mafic volcanic material and intrusive doleritic rock (Chapman, 1969).

3.3.3.4 Petrography

Mineralogy and Textural Relations. The Somesville pluton is a biotite-muscovite granite that intrudes into the northwest side of the Cadillac Mountain pluton.

The Cadillac Mountain pluton has a zonal structure more than 1 km (0.6 mi) wide extending inward from the country rock; the pluton becomes progressively coarser-textured toward the core (see Fig. 3.2). It is composed of a pale pink to greenish-gray, medium- to coarse-grained, hypidiomorphic granular hornblende granite. The chief minerals are microperthite (60%), quartz (30%), and hornblende (5%). Apatite, magnetite, and zircon are accessory minerals. Locally, biotite and sphene are found. Quartz is intergrown with microperthite, but is commonly interstitial with feldspar.

Hollister et al. (1974) related the weak and scattered molybdenum mineralization associated with Cadillac pluton to the brecciated contact zone.

Alteration. Alteration is prominent near the contact zone, and is expressed in the development of sericite from feldspar and of chlorite from the mafic mineral components.

3.3.3.5 Secondary Intrusions

Late-stage dike phases are present near the contact.

3.3.3.6 Country Rocks

The Cadillac and Somesville plutons intrude into the mafic and felsic phases of the Early Devonian Bays-of-Maine igneous complex and the older stratified volcanic and sedimentary units. A continuous shatter zone surrounds the Cadillac in the country rock but not in the Somesville pluton (see Fig. 3.2). The breccia developed from the tuffaceous beds and well-stratified siltstones of the Bar Harbor series. Locally, the breccia represents disrupted mafic sills and dikes. Undisturbed beds of the Bar Harbor series are found in and beyond the outer edge of the shatter zone.

3.3.3.7 Rock Structure

A brecciated and shattered zone is developed in the country rock at the contact. Jointing, both tectonic and due to cooling, is well developed, and in places is mineralized or filled with quartz veins (Hollister et al., 1974).

3.3.4 Calais Gabbro-Diorite Complex (Plate II, Pluton 29)

3.3.4.1 Size, Shape, and Location

The Calais gabbro-diorite complex is located south of Calais on the New Brunswick border (see Fig. 3.3); it extends south about 10 km from near the

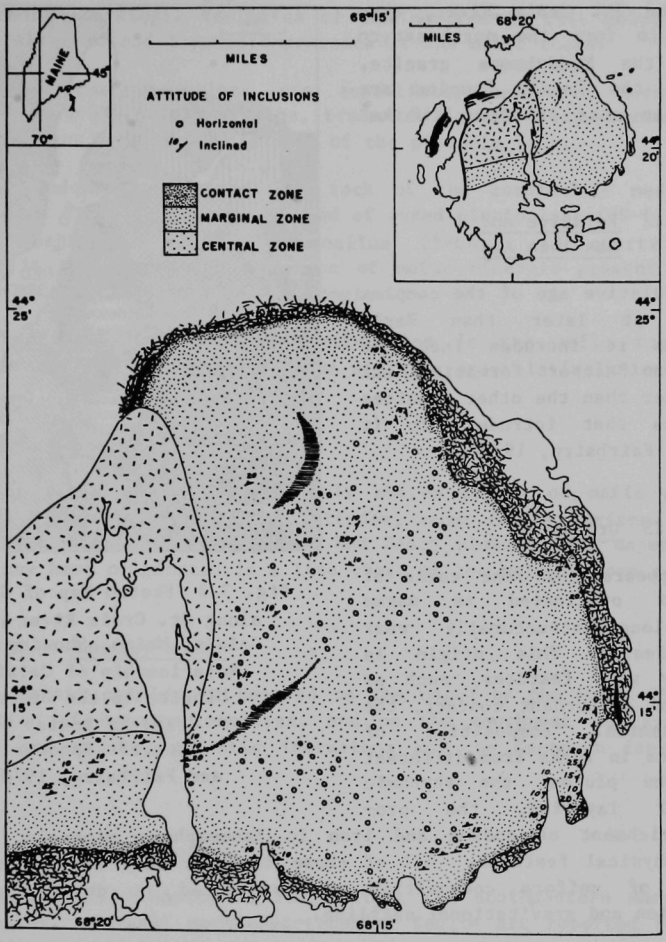

Fig. 3.2 Mount Desert Island and Lithologic Relations of the Cadillac Mountain Pluton (stipple: Cadillac Mountain hornblende granite; dashes: Somesville biotite granite) (Source: Chapman, 1969)

mouth of the St. Croix River, then turns west to form the northeastern border of the Meddybemps granite. Outcrops of the Calais Complex are found over an area of about 118 km^2 (45 mi^2).

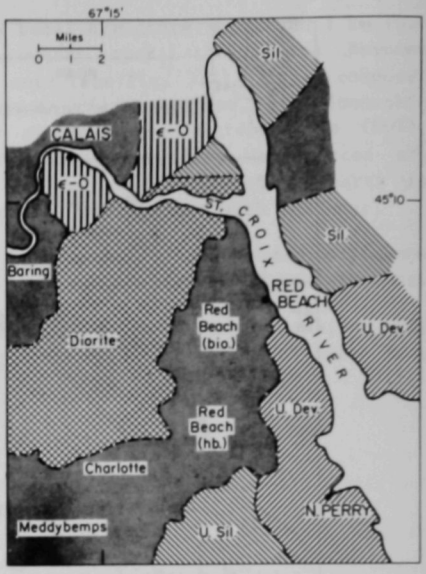

Fig. 3.3 Sketch Map of Geology along St. Croix River near Calais, Maine, Showing the Relationship of Calais Diorite to the Other Granitic Bodies (Source: Spooner and Fairbairn, 1970)

3.3.4.2 Relative and Absolute Age

The relative age of the complex is placed at later than Early Devonian, as it intrudes into the Lower Devonian Eastport formation, but somewhat older than the other granites of the area that intrude into it (Spooner and Fairbairn, 1970).

3.3.4.3 Origin

It appears that the rocks of the complex originated as magma intruded along structurally weak discontinuities. This origin is indicated by many features, such as the sharp and discordant contact, the uniform chemical composition of mineral phases in rocks from different areas in the pluton, the rhythmic compositional layering, the progressive enrichment of sodium and iron in isomorphous minerals, and the associated physical features. All of these features lead to the conclusion that magma of uniform composition was subjected to fractionation by crystallization and gravitational settling.

3.3.4.4 Petrography

Mineralogy and Textural Relations. Mafic rocks in the Calais complex are represented by layered norites and nonlayered gabbros, with norites being prominent in the northwest border of the plutonic belt. They have been well described by Amos (1963). The norite contains bronzite (orthopyroxene) (15-17%); calcic plagioclase (50-54%); and augitic clinopyroxene (12-13%), with minor amounts of olivine (2-3%), hornblende (up to 7%), and biotite (1-3%). Variations in the plagioclase, augite, and olivine proportions characterize

the different mineralogic varieties of the norites in this complex and also define the streaked and layered appearance of the mafic rocks.

The gabbros themselves are essentially unlayered normal gabbros (plagioclase, 48-58%; orthopyroxene, trace to 2%; and clinopyroxene, 10-38%); they are abundant in the central part of the plutonic belt.

The predominant intermediate rock of the complex is medium-grained biotite-quartz diorite. It is composed of zoned plagioclase (39-51%), biotite (25-28%), hornblende (1-5%), microcline (5-6%), and quartz (13-23%). Variations in the proportion and type of mafic minerals present define the other less abundant diorites in the complex.

Alteration. The prominent alteration in the mafic rocks is the formation of chlorite and epidote. The diorites are relatively less altered.

3.3.4.5 Secondary Intrusions

Small dikes and less definitely shaped masses of mafic pegmatites, together with pods and veinlets of sulfides, occur in the norites. Granites, e.g., the Meddybemps, Charlotte, and Red Beach plutons, are major intrusive rocks and may have originated in a late phase of the main igneous activity.

3.3.4.6 Country Rocks

The Calais complex is bounded on the north by the Lower Silurian Oak Bay formation of metaconglomerates, and on the southwest and south by the Meddybemps and related granitic intrusions and the Eastport formation (see Fig. 3.3).

3.3.4.7 Rock Structure

Ludman (1977) reported a fault along the southwestern margin of the Calais complex. Several northeast-trending faults are reported in the belt (Amos, 1963). Large-scale and small-scale petrologic layering is a well-developed feature of the mafic rocks.

3.3.5 Center Pond Quartz Monzonite (Plate II, Pluton 26)

3.3.5.1 Size, Shape, and Location

The Center Pond quartz monzonite, also known as the Lincoln granite (Larrabee et al., 1965), is a roughly elliptical intrusive body that trends northeast to southwest. It occupies a large part of the Winn quadrangle and

parts of the Lincoln, Passadumkeag, and Saponac quadrangles in Penobscot County in east-central Maine. Outcrops of the pluton are found over an area approximately 24 km (15 mi) long and 10 km (6 mi) wide, covering approximately 154 km^2 (59 mi^2) (Scambos and Loiselle, 1979 and 1980). Cold Stream Pond, Eskutassis Pond, Upper Cold Stream Pond, Upper Pond, Mattanawcook Pond, and Caribou Pond are glacial lakes that lie within the outlines of this pluton.

3.3.5.2 Relative and Absolute Age

The Center Pond pluton intrudes into Silurian and older metasedimentary units, and is itself cut by a fault associated with the post-Acadian tectonic activity of the Norumbega fault zone (Ayuso and Wones, 1980; Scambos and Loiselle, 1980). These relationships indicate a relative age of Devonian. Larrabee et al. (1965) cited several written communications and published values of absolute age of the Center Pond quartz monzonite: (1) 340 m.y. and 370 m.y., Pb-α; (2) 349 m.y. and 359 m.y., K-Ar; and (3) 350 ± 100 m.y., Rb-Sr. The high uncertainty factor for the Rb-Sr date is due to the presence of a large amount of normal strontium relative to radiogenic strontium (Larrabee et al., 1965). Loiselle et al. (1983b) found a Rb-Sr whole-rock age of 372 ± 10 m.y. and a Pb-Pb zircon age of 380 ± 5 m.y. for the emplacement of the pluton.

3.3.5.3 Origin

Emplacement as a fluid magma is indicated by the weak primary foliation, by evidence of lit-par-lit intrusion, by metasedimentary inclusions near the contacts; and by a wide, thermally altered aureole around the pluton (Loiselle and Ayuso, 1980). Geochemical anaylsis shows that the rocks are similar to I-type granitoids, thus precluding their derivation from preexisting metasedimentary units (Loiselle and Ayuso, 1980).

3.3.5.4 Petrography

Mineralogy and Textural Relations. The Center Pond pluton consists of calc-alkaline quartz monzonite, monzonite, and granite (Scambos and Loiselle, 1979). Typical rock specimens are composed of abundant hornblende, nearly equal amounts of quartz, plagioclase, and microcline, and lesser amounts of biotite and muscovite. Accessory minerals are primary sphene, zircon, allanite, and opaques; hydrothermal minerals are chlorite, epidote, and sericite (Loiselle and Ayuso, 1980). The amphibole/biotite ratio in the Center Pond granite varies from 0 to 1.2. The texture is predominantly medium grained and equigranular, although both medium-grained aplitic and coarse-grained porphyritic facies are present. Local porphyritic zones include tabular phenocrysts of microcline that are 2.5 cm (1 in.) long and 0.6 cm (0.25 in.) wide. Microcline phenocrysts, biotite books, and lenses measuring

up to 30 cm x 10 cm (12 in. x 4 in.) exhibit subparallel orientation, which produces foliation. Near the margins of the pluton are abundant inclusions of metasedimentary country rock. In addition, partially resorbed mafic inclusions are randomly distributed throughout the granite body (Loiselle and Ayuso, 1980).

Alteration. Hydrothermal alteration has changed some biotite and feldspar components to chlorite and sericite.

3.3.5.5 Secondary Intrusions

The Center Pond pluton is cut by a northeast-to-southwest-trending mylonite belt that is approximately 1.5 km (0.9 mi) wide (Loiselle and Ayuso, 1980).

3.3.5.6 Country Rocks

The Bottle Lake pluton lies near the Center Pond quartz monzonite, and the contact-metamorphic aureoles of the two merge. The adjacent bedrock formations include metamorphosed gray and green siltstone, sandstone, shale, black shale, chert, and a conglomerate of quartzose metagreywacke. The Center Pond quartz monzonite intrudes into strata of the Merrimack synclinorium that have been regionally metamorphosed to chlorite grade. The metasediments immediately surrounding the pluton are metasiltstone and slate of Silurian age and a range of metamorphosed sandstone, chert, conglomerate, and greywacke equivalent to the Vassalboro formation. The metasiltstone and slate exposed along the main road eastward from Lincoln contain iron and calcium carbonates, interlaminated metasandstone and metasiltstone, and quartzose metaconglomerate (Larrabee et al., 1965). Inclusions of the country rock are present near the contacts.

3.3.5.7 Rock Structure

A major strike-slip fault trending to the northeast cuts diagonally across the map plan of the Center Pond pluton (Ayuso and Wones, 1980). Along the 18-km (11-mi) fault, the northern half of the pluton is offset 2.6 km (1.6 mi) to the northeast. The fault is associated with the Norumbega fault zone of post-Acadian age.

The foliation expressed by the alignment of microcline phenocrysts and mafic minerals may represent a primary flow feature. The orientation of the grains may have occurred under a confining pressure while the rock was still partially molten and mobile. The foliation is variably developed through the intrusive rock body.

The Center Pond quartz monzonite pluton has not been deformed by folding events.

3.3.6 Chiputneticook Quartz Monzonite (Plate II, Pluton 28)

3.3.6.1 Size, Shape, and Location

The Chiputneticook quartz monzonite is also known as the Skiff Lakes granite (Lutes, 1980). It is a small part of the Pokiok batholith; it is exposed in an arcuate band along the islands and shores of the Chiputneticook lakes in eastern Maine on the U.S.-Canadian border (see Fig. 3.4). The major part of the batholith extends northeast at least 64 km (40 mi) into New Brunswick. The small sliver of the batholith that occurs in Maine is a strip about 29 km (18 mi) long and 2-6 km (1-4 mi) wide on the western and southwestern shores of the lakes, with intermittent outcrops in the Vanceboro, Forest City, Danforth, and Amity 7.5-min quadrangles.

3.3.6.2 Relative and Absolute Age

The Chiputneticook quartz monzonite intrudes into Silurian and older metasedimentary rocks. An absolute age of 400 m.y. determined by the K-Ar method was reported by Faul et al. (1963) for a sample from the Danforth quadrangle; Tupper and Hart (1961) dated a sample from Fosterville, New Brunswick, at 380 m.y. McCutcheon et al. (1981) provided a definitive Rb-Sr age of 389 $\pm$ 20 m.y. for the Pokiok batholith.

3.3.6.3 Origin

The mode of emplacement of the Chiputneticook quartz monzonite is thought to be the same as that of other Devonian intrusives in Maine, i.e., fluid magma injection. Evidence of flow foliation within a meter or so of the border of the pluton supports this point of view.

3.3.6.4 Petrography

Mineralogy and Textural Relations. The rocks range in composition from a light-gray to medium-gray biotite quartz monzonite to a light-gray granite. The grains average larger than 6 mm (0.25 in.) in length. The rock is commonly porphyritic, with very large, euhedral potassium-feldspar phenocrysts that sometimes exhibit rapakivi texture, but the porphyritic texture does not occur within a meter or so of the pluton margins. Lutes (1980) recognized an extension of the Skiff Lake granite body in New Brunswick as a zoned biotite granite that grades from porphyritic to somewhat massive from the rim to the core of the pluton (see Fig. 3.4).

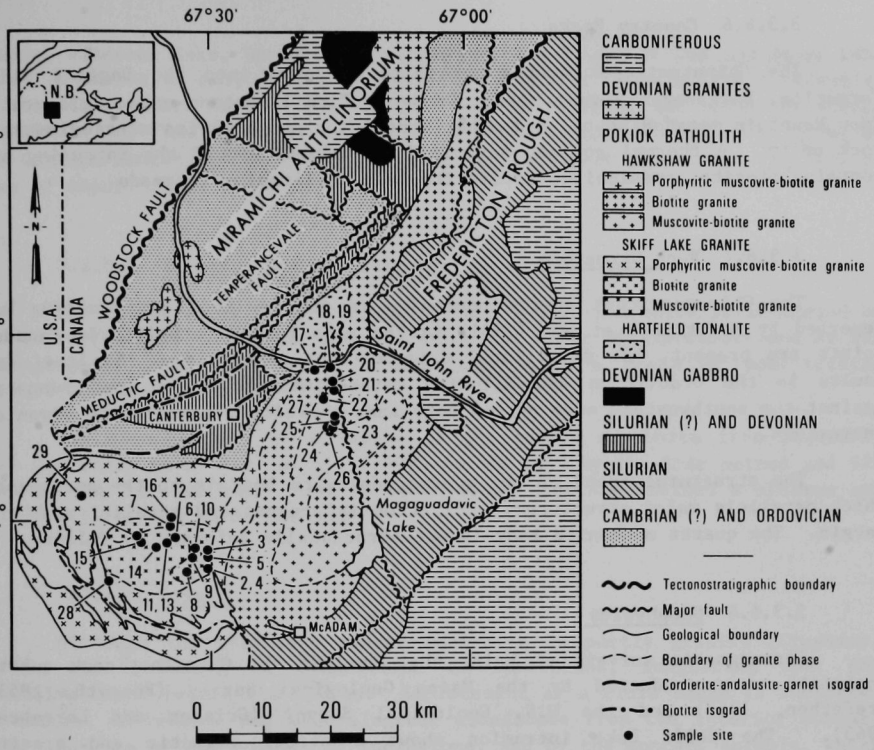

Fig. 3.4 Geologic Relations of the Chiputneticook Pluton to the Pokiok Batholith of New Brunswick (Source: McCutcheon et al., 1981)

3.3.6.5 Secondary Intrusions

In addition to the mutually intrusive relations of the various phases of the pluton (see Fig. 3.4), a west-southwest-trending diabase dike of post-Devonian age cuts across the Chiputneticook quartz monzonite at approximately 45° 44' N. latitude. Aeromagnetic data indicate that this steeply dipping dike extends 160 km (100 mi). Its length in the Chiputneticook quartz monzonite body is approximately 5 km (3 mi); it is 60 m (200 ft) wide at Dark Cove in the Danforth quadrangle (Larrabee et al., 1965). The rock is medium grained and contains 3-5% magnetite.

3.3.6.6 Country Rocks

The Chiputneticook quartz monzonite intrudes into the Daggett Ridge formation, Baskahegan Stream slate, Croperly Turn limestone metaconglomerate, Snow Mountain metasediments, and other Ordovician and Silurian metasedimentary rock units. A thermal contact aureole is developed around the intrusive and contains biotite, andalusite, cordierite, and sillimanite isograds.

3.3.6.7 Rock Structure

The Chiputneticook quartz monzonite is not faulted, but faulting was reported by McCutcheon et al. (1981) in the main Pokiok batholith in Canada. Joints are present, but directional data are not available. At least two faults in the Ordovician and Silurian metasedimentary units are truncated against the southwestern edge of the body and, therefore, predate the Devonian intrusions.

The structural trend of the quartz monzonite body is to the northeast, which parallels major structures. Foliation is present and parallel near the margin. The quartz monzonite body is not structurally deformed by folding.

3.3.6.8 Geophysics

Aeromagnetic surveys of Maine that include the Chiputneticook quartz monzonite were conducted by the Maine Geological Survey (Forsyth, 1953; Trefethen, 1953) and the U.S. Geological Survey (Griscom and Larrabee, 1963). The Skiff Lake intrusion shows a strong magnetic and gravity expression indicating it to be separate from other phases of the batholith (McGrath, 1970).

3.3.7 Concord and Related Plutons (Plate II, Pluton 5)

See Sec. 4.3.12 and Plate III, pluton 7. This rock type crops out in Maine over an area of only 30 km^2 (19 mi^2); the major part is located in New Hampshire, and it is therefore described with the plutons of that state.

3.3.8 Cranberry (Wabassus) Lake Pluton (Plate II, Pluton 35)

3.3.8.1 Size, Shape, and Location

The Cranberry Lake pluton is a north-to-south elongated body located in eastern Maine just south of the Bottle Lake complex. The pluton is best exposed in the Wabassus Lake quadrangle along the eastern shore of Grand Lake, between Grand Lake Stream and Dyer Cove Point (Larrabee, 1964b), and along the

road to Wabassus Lake, between the village of Grand Lake Stream and Grand Lake Brook (Larrabee et al., 1965). It also crops out in the Big Lake quadrangle, in the southeastern part of the Nicatous Lake quadrangle, and along the shores of Machias Lake and Wabassus Lake. It is approximately 45 km (28 mi) long at its maximum length and 23 km (14 mi) wide at its widest point, and occupies an area of about 435 km^2 (168 mi^2).

3.3.8.2 Relative and Absolute Age

The relative age of the Cranberry Lake quartz monzonite is indicated by its intrusion into the surrounding Silurian and older metasediments and by its petrographical and geochemical similarities with a group of post-Acadian intrusive bodies in Maine.

The absolute age of a sample of the quartz monzonite from Cranberry Lake was reported by Faul et al. (1963) as 356 m.y. by the K-Ar method and 420 m.y. by the Pb-α method. Samples from Peaked Mountain yielded a minimum age of 332 m.y. by the Rb-Sr method.

3.3.8.3 Origin

Although the Cranberry quartz monzonite is partly granite (Stuesser, 1966), the variation in composition is continuous, suggesting that one intrusive event formed the pluton. Emplacement as a fluid magma is indicated by changes in texture and mineralogical abundances from the interior outward to the borders of the pluton, by inclusions, and by the contact aureole, which is characteristically well developed at the contacts with associated Devonian intrusions and metasediments.

3.3.8.4 Petrography

Mineralogy and Textural Relations. The body consists of a pink or gray, biotitic or hornblendic quartz monzonite. The rock is generally equigranular and coarse grained, with particle sizes ranging from 3-10 mm (0.125-0.375 in.). In some places the rock contains phenocrysts of microcline as large as 25 mm (1 in.) long and up to 6 mm (0.25 in.) wide. On Amazon Mountain the monzonite is medium gray and leucocratic (Larrabee et al., 1965). Biotite and hornblende increase in abundance near inclusions and at the borders of the pluton.

The rock can be described as a porphyritic perthite-oligoclase-quartz-biotite granite (Stoeser, 1966). The perthite is gray to pink, occurring as anhedral to subhedral crystals up to 25 mm (1 in.) long. The oligoclase grains are anhedral and less than 13 mm (0.5 in.) long. Anhedral gray quartz grains less than 13 mm (0.5 in.) in diameter are abundant, and smaller, black, anhedral to subhedral biotite grains are moderately abundant. The groundmass

matrix consists of feldspar, quartz, and biotite grains less than 2.5 mm (0.1 in.) in diameter. Phenocrysts of orthoclase, quartz, and minor plagioclase occur in a phenocryst-to-groundmass ratio of approximately 1 to 3. Rapakivi structure is variably developed throughout the Cranberry quartz monzonite.

Alteration. Alteration is not clearly described in the literature.

3.3.8.5 Secondary Intrusions

Small aplite and pegmatite dikes and quartz veins occur in the quartz monzonite bodies of the area, but they are less than 15 cm to 0.6 m (6 in. to 2 ft) thick.

3.3.8.6 Country Rocks

The magma intruded into rocks of Cambrian or Ordovician age and the Silurian Kellyland formation, which is sometimes called the Pale argillite. All of the country rock was metamorphosed to chlorite grade -- the Cambrian-Ordovician rocks to quartzite, sandy metasiltstone, a minor thin metaconglomerate, and slate, and the Kellyland formation to predominantly metasiltstone and metasandstone. The Kellyland formation is silicified to varying degrees. Metasedimentary rocks of Cambrian or Ordovician age may have been faulted upwards into younger rocks prior to intrusion of the quartz monzonite magmas (Larrabee et al., 1965).

The contact metamorphic aureole, which extends less than 1.2 km (0.75 mi) away from the pluton, contains magnetite and biotite in great abundance. The aureoles are more extensive in surrounding slates and metasiltstones than in the quartzites (Larrabee et al., 1965).

3.3.8.7 Rock Structure

The outcrop area of the Cranberry Lake quartz monzonite may have been subjected to two periods of fault activity that postdate the intrusion of the monzonite. The earlier discontinuous faults trend northeasterly to east-northeasterly and are parallel to the strike of the major Springhill fault near Fredericton, New Brunswick, and the Fundy and Lubec faults to the southeast (Larrabee, 1964). There is some displacement within the Cranberry Lake quartz monzonite along the Norumbega fault, which bounds the body at its north end. A second fault system in the Wabassus Lake quadrangle trends west-northwest.

Minor fractures exist in the pluton, but they do not form a major joint system. A zone of sheared quartz monzonite occurs near Nicatous Lake. The Cranberry quartz monzonite does not exhibit foliation, flow structures, or

preferred orientation of mineral grains. Fold structures in the pluton have not been described in the available literature.

3.3.9 Flagstaff Lake Complex (Plate II, Pluton 19)

3.3.9.1 Size, Shape, and Location

The Flagstaff Lake complex (Boone, 1973) is an elongate belt of rocks ranging from gabbro to granite in west-central Maine. The belt, occupying an area of about 282 km^2 (109 mi^2), strikes southwestward from the Little Bigelow Mountain across the Stratton and Kennebago Lake quadrangles to the northern part of the Rangely quadrangle. The complex is batholithic in size, with outcrops occurring over a length of about 40 km (25 mi) and a width of 5 to 8 km (3-5 mi).

3.3.9.2 Relative and Absolute Age

Boone (1973) assigned a Devonian age to the Flagstaff Lake complex, as it intrudes into the Cambrian(?) or Ordovician(?) Dead River formation and folded structures in the Lower Devonian Seboomook formation on the northeast; therefore the intrusion is probably late Acadian.

3.3.9.3 Origin

The Flagstaff Lake complex is a southeast-dipping felsic sheet that intruded along pre-existing structural weaknesses or discontinuities (Boudette, 1978). The garnet granofels contact zone is believed to represent the residue from the Quimby and Greenvale Cove formations assimilated by the intrusive through anatexis (Moench et al., 1982).

3.3.9.4 Petrography

Mineralogy and Textural Relations. The gabbroic rocks (gabbro, troctolite, norite, etc.) in the complex are massive, equigranular, and medium to coarse grained. Pyroxene gabbros are predominant, with hypersthene diopsidic augite. The modal composition of the pyroxene hornblende gabbro is plagioclase, 50%; hornblende, 10-45%; diopside, 4%; magnetite, 2%; biotite with minor chlorite, 1.6%; quartz, 0.3%; hypersthene, 0.1%; and sphene, 0.1% (Boone, 1973). Eastward, from the Flagstaff Lake basin to the Round Top Mountain lobe, the rock becomes finer grained with a doleritic texture.

The dioritic rocks in the complex are inequigranular, fine to medium grained, and weakly porphyritic. Polygranular aggregates of quartz mantled by pyroxene or hornblende are common. Plagioclase phenocrysts are formed. The

modal composition of the coarse-grained hornblende quartz diorite is plagioclase, 52%; quartz, 20%; hornblende, 18%; biotite with minor chlorite, 8.5%; hypersthene, 0.3%; diopside, 0.3%; and perthite, 0.2%. Where the quartz content decreases, the rock becomes a quartz-bearing hornblende-hypersthene diorite (Boone, 1973).

Quartz monzonite in the complex is light gray, medium to coarse grained, massive, and porphyritic with a hypidiomorphic groundmass. Phenocrysts of microcline-perthite and plagioclase are subequant, blocky, 1.5 to 3 cm (0.6-1.2 in.) long, and constitute about 5% to 15% of the rock. The modal composition is plagioclase, 38%; quartz, 27%; perthite, 23%; biotite with minor chlorite, 10%; chlorite, 1.5%; muscovite, 0.5%; and trace amounts of magnetite, apatite, and garnet (Boone, 1973).

Garnet granofels in the complex is medium to coarse grained and massive. The major minerals are almandine garnet (5-60%) and andesine (up to 50%) (Moench et al., 1982). Biotite and quartz are accessory minerals. Hypersthene and amphibole or cordierite are common.

Alteration. In the central portion of the larger masses, there is no alteration of the gabbroic rocks, but elsewhere, they are commonly retrograded to hornblende-chlorite-biotite-bearing assemblages (Boone, 1973). Saussuritic (chlorite-epidote) alteration is pervasive in the Kennebago Lake and Biglow Mountain area.

3.3.9.5 Secondary Intrusions

In the complex, the Pierce Pond pluton, Sugarloaf massif, and the Lexington batholith are secondary intrusions. In the diorite and gabbro facies, fine-grained, inequigranular, porphyritic, garnetiferous granite occurs as dikes and apophyses. Albite-rich pegmatite dikes are abundant (Boone, 1973).

3.3.9.6 Country Rocks

To the east of the complex, the Pierce Pond Pluton lies in fault contact; to the southeast is the Lexington batholith. Sugarloaf massif is towards the south, and the Huston diorite-granite composite stock intrudes at the southern end of Flagstaff Lake. Metasedimentary rocks of the Seboomook, Carrabasset, and Dead River formations surround the complex. To the north, an aureole is present at the boundary. The contact zone is marked by prominent garnet granofels.

The complex conforms to the regional trend of the folded metasedimentary rocks. Locally, small-scale gradational contacts are found. Spessartine-almandine garnet is found at the contact between quartz monzonite and the country rock.

3.3.9.7 Rock Structure

The most extensive faults (parallel to the major fault traces) cut the quartz monzonite and pegmatite dikes below the Long Falls Dam. Cataclastic zones 0.5 m (20 in.) wide bound the faults where the quartz monzonite was hydrothermally altered. The longest fault trace, trending northwest from Hill 1900 to the Pierce Pond quadrangle, is in en-echelon relationship with the fault offsetting the batholith contact north of West Carry Pond (Boone, 1973).

3.3.9.8 Geophysics

The available data indicate that the fault sets in the Flagstaff Lake Complex are not seismically active (Boone, 1973).

3.3.10 Gardner Lake Gabbro-Diorite Complex (Plate II, Pluton 31)

3.3.10.1 Size, Shape, and Location

The Gardner Lake complex is an irregularly indented, somewhat rounded body of gabbros and diorites in southeastern Maine (Westerman, 1979). The complex extends over an area of 241 km^2 (93 mi^2) in the Gardner Lake quadrangle. It is about 25 km (15 mi) in its longest axis from east to west, and about 11 km (7 mi) in its shortest axis from north to south.

3.3.10.2 Relative and Absolute Age

Gates (1978) and Doyle et al. (1967) identified Gardner Lake as part of the Devonian igneous intrusive rocks known as the Bays-of-Maine complex defined by Chapman (1962).

3.3.10.3 Origin

The Gardner Lake complex is an intrusive pluton with an overall alkalic composition.

3.3.10.4 Petrography

Mineralogy and Textural Relations. Gardner Lake rocks are massive, equigranular, fine- to medium-grained gabbroic rocks composed mainly of plagioclase and hornblende, with varying amounts of augite (Young, 1963). They include norites, gabbros, and diorites, as do the other alkalic rocks of the area (Amos, 1963; Bickford, 1963). Sulfides are accessory minerals.

Alteration. Alterations are pervasive near local mineralized zones. Young (1963) stated that there may have been a minor late-stage sulfide remobilization.

3.3.10.5 Secondary Intrusions

Sulfide stringers and veinlets are common in the basic rocks (Young, 1963).

3.3.10.6 Country Rocks

The Meddybemps granite is in discordant contact to the north. To the northwest lies the Pocamoonshine gabbro-diorite. The granodiorite at Whitneyville is in contact with the Gardner Lake complex to the southwest. Silurian rhyolitic to basaltic flows, tuffs, and sedimentary rocks are host rocks to the Gardner Lake complex (Doyle et al., 1967). Mineralized zones grade into the diabase (Young, 1963).

3.3.10.7 Rock Structure

A well-developed foliation is present, formed from biotite.

3.3.10.8 Geophysics

Young (1963) reported magnetic anomalies associated with the sulfide mineralization on various prospects in the Gardner Lake 15-min quadrangle. On the basis of gravity contrast data, Biggi and Hodge (1983) postulate that the Gardner Lake pluton is a thin body underlain by granitic rocks.

3.3.11 Gouldsboro Pluton (Plate II, Pluton 38)

3.3.11.1 Size, Shape, and Location

The Gouldsboro is an irregular pluton that belongs to the granitic-granophyric phase of the Bays-of-Maine igneous complex. It extends over an area of about 105 km^2 (41 mi^2) on the east side of Frenchman Bay on the coast of Maine. The minimum and maximum dimensions of the area within which outcroppings are found are approximately 8 and 17 km (5 and 11 mi), respectively.

3.3.11.2 Relative and Absolute Age

The Gouldsboro pluton is part of the late phase of the Bays-of-Maine emplacement. It intrudes into the Ordovician Ellsworth schist, and thus is probably of Devonian age (Chapman, 1962; Doyle et al., 1967).

3.3.11.3 Origin

Chapman (1962) stated that the Bays-of-Maine complex is a mafic-felsic assemblage that was emplaced within and beneath Silurian and Devonian volcanic and sedimentary rocks. The Gouldsboro pluton and other, associated, granites may represent differentiated portions of the gabbros (Chapman, 1962), which reached the surface along deep crustal fractures (Biggi and Hodge, 1983).

3.3.11.4 Petrography

Mineralogy and Textural Relations. The granite is normal and fine to medium grained. It is composed of microcline-perthite, albite, biotite, and quartz. Small amounts of hornblende are also present (Young, 1962).

Alteration. Hydrothermal mineralization and alteration have affected the rocks (Young, 1962).

3.3.11.5 Secondary Intrusions

No significant intrusions are described except for mineralized stringers. Chapman (1962) stated that the swarms of xenoliths represent remnants of a contact zone below a flat, undulating, granite-granophyric-phase roof.

3.3.11.6 Country Rocks

To the east of the Gouldsboro pluton lies a unit of granodiorite, up to Narraguagus Bay, that Doyle et al. (1967) and Chapman (1962) identified as a similar rock type; it may be an extension of the pluton, but no detailed information is available to verify this identification. Towards the south lies the Atlantic Ocean; the Bar Harbor series lies to the west in discordant contact with the Gouldsboro. Near Jones Cove, at the northwest side of the pluton, greenstone is in contact with the Gouldsboro rocks. Towards the north lie the Ellsworth schist and the Tunk Lake pluton.

3.3.11.7 Rock Structure

Sheets of granitic rocks intrude the layered and laminated gabbroic phase of the Bays-of-Maine complex (Chapman, 1962). Foliation is pronounced in the diorite. Local shear zones are present at the Gouldsboro prospect, in Gouldsboro Township (Young, 1962).

3.3.11.8 Geophysics

Young (1962) has described magnetic anomalies associated with mineralized zones.

3.3.12 Granodiorite of Whitneyville (Plate II, Pluton 32)

3.3.12.1 Size, Shape, and Location

In eastern Maine, near the town of Whitneyville, lies an arcuate body of granodiorite that occupies an area of 100 km^2 (39 mi^2) in the Columbia Falls, Whitneyville, and Wesley quadrangles. Its length and width are 20 and 5 km (12 and 3 mi), respectively.

3.3.12.2 Relative and Absolute Age

The granodiorite of Whitneyville is part of the Jonesport intrusive series (Terzaghi, 1946). The series is younger than the Jonesboro granite (Terzaghi, 1946), which has been assigned a Pb-α zircon age of 430 m.y. by Faul et al. (1963). The granodiorite is also classified as part of the Devonian granitic-granophyric phase of the Bays-of-Maine complex (Chapman, 1962; Doyle et al., 1967). Pajari et al. (1974) and Westerman (1973) reported ages of 408 ± 14 to 424 ± 24 m.y. for the granites and gabbros of the Bays-of-Maine complex, suggesting that part of it may be Late Ordovician (also see Westerman, 1979).

3.3.12.3 Origin

The granodiorite of Whitneyville resulted from the late-stage, comagmatic granitic-granophyric phase of the igneous activity that developed the Bays-of-Maine complex from an alkalic magma (Chapman, 1962).

3.3.12.4 Petrography

Mineralogy and Textural Relations. The granodiorite is coarse grained and contains angular aggregates of pyroxene. It is composed of oligoclase, 80%; quartz, 10%; perthite, 4%; and biotite (Terzaghi, 1946).

Alteration. Terzaghi (1946) reported that pyroxene is partially altered to biotite, amphibole, and/or chlorite.

3.3.12.5 Secondary Intrusions

Small dikes and veins of granites, microgranites, and pegmatites can be found (Terzaghi, 1946).

3.3.12.6 Country Rocks

Contacts with the country rocks are largely inferred; in some places contact is marked by fine-grained albite porphyry (Terzaghi, 1946). The Peaked Mountain Pond gabbro-diorite complex lies to the east; the Jonesboro granite is the wall rock to the south. All other surrounding areas are primarily Silurian rhyolitic and basaltic flows and tuffs (Doyle et al., 1967).

3.3.12.7 Rock Structure

Jointing and small-scale shearing is evident in the granodiorite (Chapman, 1962).

3.3.12.8 Geophysics

On the basis of gravity models, Biggi and Hodge (1983) suggested a thickness of 0.8 km (0.5 mi) for the Whitneyville body.

3.3.13 Hallowell and Togus Plutons (Plate II, Pluton 9)

3.3.13.1 Size, Shape, and Location

The Hallowell and Togus plutons are about 2 km (1.2 mi) apart in south-central Maine, bracketing the city of Augusta. The plutons occupy an outcrop area of approximately 123 km^2 (47 mi^2), with overall conformity to an oval shape. Two protuberances extend from the main granitic mass in a southeasterly direction both north and south of Augusta. Several outlying satellite bodies occur in the surrounding metasediments.

3.3.13.2 Relative and Absolute Age

The Hallowell granite and its satellite bodies intrude into metasedimentary rocks of the Silurian Waterville formation and the Lower

Silurian(?) Vassalboro formation, which indicates a maximum relative age of post-Silurian (Osberg, 1968 and 1979; Ferry, 1978). Two conflicting absolute age dates occur in the literature. Hurley et al. (1958) reported a K-Ar age of 260 m.y., obtained from a biotite sample from the Hallowell granites. Osberg (1968) reported primarily Devonian Pb-α ages for several nearby stocks of the same mineralogy.

Dallmeyer and Van Breeman (1978) reported a radiometric age of 387 ± 11 m.y. for the Hallowell and 394 ± 8 m.y. for the Togus pluton, on the basis of Rb-Sr whole-rock data.

3.3.13.3 Origin

The Hallowell pluton is a discordant, coarse-grained, igneous body emplaced to a depth of approximately 11 km (7 mi) below the surface (Ferry, 1978). The main mass consists of multiple magmatic intrusions of contrasting structural relations and petrology. Initially, concordant lenses of granodiorite and tonalite were emplaced. The dominant rock type was then emplaced along channelways, accompanied by stoping of large masses of both the previously intruded granodiorite and tonalite, and the surrounding metasedimentary country rocks (Barker, 1964). Tongues of the main granitic mass intruded into the country rocks along cracks, fractures, and bedding planes. Irregular pegmatite veins about 2.5 cm (1 in.) thick were subsequently intruded, crosscutting the foliation in the granites and the schistosity in the country rock (Ferry, 1978). Injection of a few thin, vesicular, alkalic olivine basalt dikes of Late Jurassic to Early Triassic age is the last recorded igneous event.

3.3.13.4 Petrography

Mineralogy and Textural Relations. The Hallowell pluton is basically a muscovite-biotite-garnet quartz monzonite. The composition of the main granitic mass falls within the following ranges: 26-36% (by volume) quartz, 19-43% potassium feldspar, 17-41% plagioclase feldspar, 0-6% biotite, 0-10% muscovite, and minor accessory minerals (Barker, 1964). This range in composition reflects the multiple-intrusion nature of the pluton: up to four different magmatic pulses have been recognized in single outcrops. The more mafic minerals crystallized first, followed by the more felsic minerals from the residue. Quartz grains are anhedral, equidimensional, and average 8 mm (0.3 in.) in diameter. Potassium feldspar is present as microcline and microperthite megacrysts up to 4 cm (1.5 in.) long. Biotite, the only mafic mineral present, occurs as flakes up to 5 cm (2 in.) in diameter, and is frequently cut by even larger muscovite grains. Garnet occurs as red anhedral or dodecahedral grains 7-40 mm (0.3-1.5 in.) in diameter. The plagioclase is anhedral, equidimensional to slightly elongate, and has an average composition

of Ab-90, An-8, Or-2. Minor and trace quantities of apatite, zircon, monazite, ilmenite, pyrrhotite, hematite, and magnetite occur.

Alteration. All of the granites near Augusta, including the Hallowell, contain plagioclase and microcline that have been altered from 1-10% by volume to the mineral assemblages muscovite, calcite, and quartz; muscovite, epidote, and quartz; or muscovite, calcite, epidote, and quartz. The biotites have been metasomatically altered from 1-90% by volume to the mineral chlorite. Small but significant amounts of carbon dioxide, water, hydrogen, and potassium have been added to the intrusives during this metasomatic event. This alteration process occurred soon after the emplacement of the granites; the magmatic rock acted as a heat source, circulating metasomatic fluids through both the granite and country rock (Ferry, 1978 and 1979).

3.3.13.5 Secondary Intrusions

The two contrasting types of secondary intrusions that occur in the pluton are the pegmatitic and basaltic (Mesozoic) dikes.

3.3.13.6 Country Rocks

The country rocks of the Silurian Waterville formation and the Lower Silurian(?) Vassalboro formation consist of metamorphosed interbedded sandstones, pelites, and argillaceous carbonates. The carbonate rocks constitute only a small percentage of the rock surrounding the intrusion (Osberg, 1968 and 1979; Ferry, 1976). Nonetheless, the carbonates played an important role as a source of carbon dioxide in the metasomatic reactions between sedimentary and igneous rock (Ferry, 1978). The Waterville formation at its lowest metamorphic grade consists of quartz, plagioclase, muscovite, and chlorite. The metamorphic grade increases as one approaches the pluton, passing through the biotite, garnet, andalusite-staurolite, and cordierite assemblage into a sillimanite mineral assemblage adjacent to the pluton (Osberg, 1971; Ferry, 1980). These isograds have not been mapped on the east side of the pluton, where igneous rock is in contact with the Vassalboro formation. Porphyroblasts associated with the igneous-induced metamorphic event cut across the schistosity, indicating that thermal metamorphism postdates deformation. The zoisite metamorphic isograd has been mapped for both formations and exhibits a concentric geometry around the pluton (Ferry, 1978).

3.3.13.7 Rock Structure

The Hallowell pluton is cut by vertical joint sets whose attitudes are N. 60° W. and N. 30° E., meeting at 90° (Barker, 1964). The strike of the one set is parallel to the long axis of the pluton. No documented faults cut across the intrusion.

The granite body has what appears to be a flow texture. Biotite flakes are vertical, with a strike of N. 20° E. This strike is parallel to both the long axis of the intrusion and one of the prominent joint sets. The long axes of stoped blocks and large xenoliths also are parallel to this foliation. Smaller xenoliths exhibit random orientation.

The pluton belongs to the late orogenic suite of granites of the Devonian Acadian orogeny. There is no evidence of later deformation.

3.3.14 Hartland Pluton (Plate II, Pluton 10)

3.3.14.1 Size, Shape, and Location

The Hartland pluton is a discordant intrusive in south-central Maine. The major axis of this roughly elliptical pluton trends in a northeasterly direction, roughly parallel to regional structures in the area. Outcroppings of the pluton occur predominantly in the Skowhegan and Pittsfield 15-min quadrangles, but the pluton also extends a little southward into the Waterville quadrangle; it is partially covered by Great Moose Lake. The Hartland pluton is about 30 km (19 mi) long and 10 km (6 mi) wide, and occupies an area of 195 km^2 (75 mi^2).

3.3.14.2 Relative and Absolute Age

Potassium-argon age determinations by Hurley (1964) indicate an age of 375 m.y. for the mica of the Hartland granite. Faul et al. (1963) also used the K-Ar dating method, and arrived at an age of 356 m.y. for the granite. However, their Pb-α age determinations on zircon indicated an age of 450 m.y. Faul et al. (1963) suggested that a recrystallization of mild intensity affected the K-Ar mica dates but not the Pb-α zircon dates (Toys, 1965). On the other hand, Borns (1959) stated that no recrystallization has taken place in the Hartland pluton (Toys, 1965). Dallmeyer et al. (1982) gave a 360 ± 8 m.y. Rb-Sr whole-rock age.

3.3.14.3 Origin

The Hartland pluton is chemically and mineralogically similar to S-type granites described from other tectonic provinces, which suggests that they were derived by partial melting of sedimentary material (Chappell and White, 1974; Dallmeyer et al., 1982). The depth of intrusion is estimated to be less than 10 km (6 mi), due to the absence of muscovite (Ferry, 1978) as a magmatic phase. Such a depth consideration is supported by the Acadian low-grade regional metamorphism in the vicinity of the intrusion. Lithological differences may be a reflection of initial magma composition.

3.3.14.4 Petrography

Mineralogy and Textural Relations. The Hartland pluton is a heterogeneous, equigranular, medium-grained granodiorite intrusion that locally grades into quartz monzonite and biotite granite (Dallmeyer et al., 1982). Toys (1965) gave a modal composition of microcline-perthite, 45%; quartz, 35%; oligoclase, 10%; and biotite, 7%; with lesser amounts of hornblende (2%) and chlorite (1%). Accessory minerals are sphene, zircon, and magnetite (Toys, 1965).

Quartz is anhedral and usually exhibits undulatory extinction. Oligoclase occurs as subhedral to euhedral grains and commonly forms intergrowths with the quartz to produce myrmekite. Biotite is the more common ferromagnesian mineral and is often associated with chlorite. The anhedral grains of hornblende are usually bent and exhibit very little chloritization (Toys, 1965).

Alteration. Sericite is common in thin section and probably results from the weathering of feldspar. Most of the chlorite observed appears to be after biotite, with which it is associated; some chlorite is associated with hornblende grains.

3.3.14.5 Secondary Intrusions

There is no mention of secondary intrusions in the literature.

3.3.14.6 Country Rocks

The Hartland plutonic body intrudes into the polydeformed and metamorphosed sequence of Silurian-Devonian sedimentary rocks (Pankiwskyj et al., 1976; Osberg, 1979) with Buchan-type metamorphic zones increasing to the south (Osberg, 1971). Hartland in general is surrounded by a well-developed contact aureole. Xenoliths of the country rocks occur in the pluton.

3.3.14.7 Rock Structure

Gently plunging, nearly isoclinal folds underlie the region around the southern tip of the pluton. The axial traces of these folds are oriented N. 40° E. and help form the regional structural trend. The pluton is discordant to the bedding and foliations in the country rocks, even though the general shape is elongated with the regional structural grain.

3.3.15 Katahdin Batholith (Plate II, Pluton 24)

3.3.15.1 Size, Shape, and Location

The Katahdin batholith is located in north-central Maine in the Katahdin quadrangle, the northern portion of the Jo-Mary Mountain quadrangle, the western portion of the Stacyville quadangle, the eastern portion of the Harrington Lake quadrangle, the northwestern portion of the Norcross quadrangle, and the southern portion of the South Traveler Mountain quadrangle (these are all 15-min quadrangles). The entire southern half of Baxter State Park, which includes Mt. Katahdin, is underlain by the pluton. The batholith has a roughly elliptical shape, with a 56-km (35-mi) long axis that strikes approximately northeast; it is about 32 km (20 mi) wide. Almost three-fourths of the 1283-km^2 (495-mi^2) area in which surface exposures of the granite mass are found lies outside the park's southern and eastern boundaries.

3.3.15.2 Relative and Absolute Age

The absolute ages obtained by Faul et al. (1963) using both Rb-Sr and K-Ar methods fall in the range 355-390 m.y., which is Devonian. This age is roughly the same as the stratigraphically determined relative age of post-Early Devonian (Caldwell, 1960). The manner in which the pluton cuts across structures of the Acadian orogeny also indicates that it is post-Early Devonian in age. Loiselle et al. (1983a) reported a Rb-Sr whole-rock age of 388 $\pm$ 5 m.y., but a Pb-Pb zircon age of 414 $\pm$ 4 m.y., which is older than the early Devonian value in use.

3.3.15.3 Origin

The Katahdin batholith is a discordant igneous body emplaced by diapirism, or passive stoping with associated assimilation and digestion of overlying roof rocks (Neuman, 1967). The contact zone on its southern, eastern, and western sides consists of breccias and partially assimilated, contact-metamorphic hornfelsic sedimentary and volcanic units. Extensive shouldering of the surrounding rocks occurred during the intrusive event, causing faulting and modifying the pre-intrusion structural relationships. Numerous stope blocks of the now eroded roof rock occur in the granite as xenoliths. The depth of emplacement is estimated at 1 km (0.6 mi) (Loiselle et al., 1983b).

3.3.15.4 Petrography

Mineralogy and Textural Relations. The Katahdin pluton is composed of quartz monzonite (Griscom and Larrabee, 1963), consisting of potassium feldspar (30% by volume), plagioclase (30%), quartz (30%), and biotite

(5-10%). The plagioclase feldspar has an average composition of An-22. The feldspar crystals are anhedral and can be as large as 10 mm (0.4 in.) across. Quartz grains are euhedral and average 5 mm (0.2 in.) in diameter. The biotite grains are anhedral and average about 2-3 mm (0.1-0.15 in.) in diameter. The quartz monzonite is by far the most abundant rock type. Two minor facies, one containing hornblende and the other with a granophyric texture, also occur in the Katahdin and Harrington Lake quadrangles but not in the Stacyville quadrangle (Neuman, 1967). The contact breccia zone on the eastern flank can be up to a mile wide and has a slightly different mineralogy. It varies from equal amounts of xenoliths and granitic matrix to predominantly xenoliths. The granitic matrix in this zone is highly variable as a result of different degrees of assimilation, and can range from fine-grained aplites and pegmatites to areas devoid of quartz.

Alteration. The Katahdin quartz monzonite has been subjected to synintrusive interactions between rock and water. The feldspars have been sericitized, causing a change in color from gray to pink in the main mass of the quartz monzonite (Neuman, 1967). This alteration of the feldspars appears to have been restricted to the upper portions of the intrusion, as evidenced by the pink quartz monzonite being the most common rock type at the higher elevations on Mt. Katahdin (Caldwell, 1960). This color change does not occur in the matrix granitic rocks of the surrounding breccia zone. There has also been a certain amount of alteration of biotite to chlorite.

3.3.15.5 Secondary Intrusions

The only major secondary intrusion is a large Devonian diorite that cuts across the extreme southwestern tip of the Katahdin batholith.

3.3.15.6 Country Rocks

The Lower Devonian Seboomook formation surrounds the southern third of the batholith. This formation consists of approximately 1500 m (5000 ft) of interbedded and intercalated fine-grained sandstone, dark-gray siltstone, and slate. The southwestern tip consists of a large diorite intrusion.

The oldest rock units intruded into by the granite at the eastern boundary are metavolcanics, which are collectively referred to as greenstones. They span a compositional spectrum from basalt to andesite to dacite. Relative ages indicate an age of Ordovician, which associates them with the earlier Taconic orogeny. The next youngest unit stratigraphically, which is also Ordovician, is the Wassataquick chert. This unit consists of thinly bedded cherts interbedded with minor tufaceous units. The next unit, found only in fault blocks adjacent to the intrusion, is an Upper Silurian conglomerate. It is a gray, polymictic pebble and cobble conglomerate, whose clasts are composed of quartz, quartzite, and shale fragments. The matrix is a micaceous, quartzose sandstone.

The mineralogy and petrology of the rocks on the western flank are virtually unknown. The only statement that can be made is that they consist of Ordovician metasediments and metavolcanics.

The country rocks of the northwestern and northern flanks of the pluton are composed of Lower Devonian Traveler rhyolite, which consists of many separate flows and ash falls (Rankin, 1962 and 1968). This formation is regarded as the ash-flow eruptive counterpart of the Katahdin magmas (Hon et al., 1981).

3.3.15.7 Rock Structure

In the Mt. Katahdin area, the quartz monzonite is cut by very closely spaced vertical joints. Slopes in the area consist primarily of rubble. A small zone of faulting occurs in the west central portion of the batholith, just west of the southwestern corner of Baxter State Park.

The main body of the batholith is devoid of foliation. The breccia zone contains a concentration of country rock xenoliths with preferred orientation. The zone also exhibits preferred orientation of biotite flakes. Such orientation indicates some synintrusive shearing.

No postintrusive folding of this pluton body has been reported.

3.3.15.8 Geophysics

Geophysical surveys in the area have been limited to the country-rock margins of the batholith. Electromagnetic surveys have traced the greenstone metavolcanics.

Studies indicate a near-circular residual gravity low of -32 mgal (see Fig. 3.5) over the Katahdin (Hodge et al., 1982). Gravity contrast profiles indicate that the pluton continues to a depth of about 3 km (1.9 mi).

3.3.16 Lead Mountain Pluton (Plate II, Pluton 36)

3.3.16.1 Size, Shape, and Location

Outcrops of the Lead Mountain pluton are found in the Lead Mountain, Great Pond, Ellsworth, Tunk Lake, and Cherryfield 15-min quadrangles in eastern Maine. Its map plan is nearly triangular, with irregularly shaped extensions at the northern end (in the Nicatous Lake quadrangle) and at the southeast corner. The pluton is approximately 58 km (36 mi) long and 35 km (22 mi) wide at its maximum extent; outcrops occur over an area of about 1048 km^2 (405 mi^2).

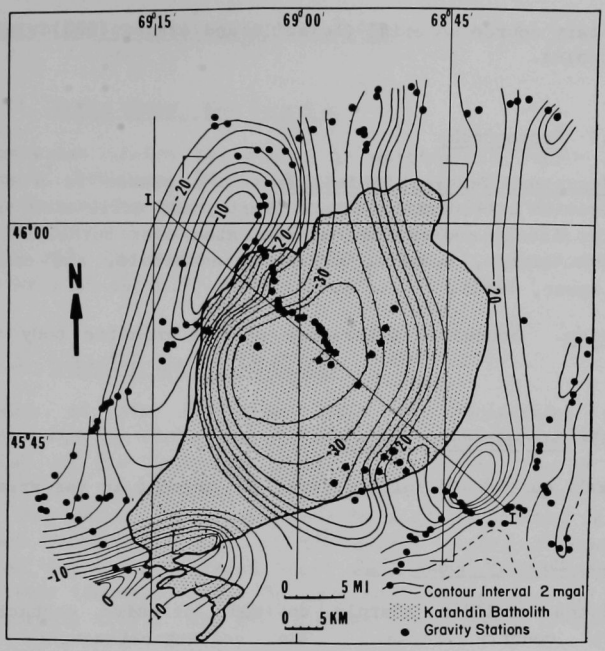

Fig. 3.5 Bouguer Gravity Anomaly Map over the Mt. Katahdin Batholith (Source: Hodge et al., 1982)

3.3.16.2 Relative and Absolute Age

The Lead Mountain pluton intrudes into the lower Paleozoic metasedimentary rocks of the greenschist facies, and is associated with post-Acadian activity in eastern Maine (Page, 1968; Loiselle and Ayuso, 1980), suggesting that its relative age is Late Devonian. Radiometric analyses of rocks with mineralogies equivalent to that of the pluton yielded a Rb-Sr whole-rock age of 392 ± 17 m.y. (Loiselle et al., 1983b).

3.3.16.3 Origin

The pluton exhibits a chemical composition characteristic of granitic magma emplacement in nonorogenic settings. The rock has a higher absolute K_2O content, a higher K_2O/Na_2O ratio, a higher absolute Rb content, a lower CaO content, and a lower K/Rb ratio than other Maine intrusives interpreted as I-type granitoids (Loiselle and Ayuso, 1980). The magma does not appear to be derived from the melting of metamorphosed pelitic sedimentary strata or from

mature sedimentary source material (Loiselle and Ayuso, 1980); it was probably emplaced by stoping.

3.3.16.4 Petrography

Mineralogy and Textural Relations. The pluton is a coarse-grained alkalic granite with subhedral feldspar, quartz, and mafic mineral grains. It has a low color index, and biotite predominates over hornblende. Accessory minerals include zircon, apatite, allanite, amphibole, and opaque minerals (Loiselle and Ayuso, 1980).

Alteration. Secondary alteration of the granitic body has not been described.

3.3.16.5 Secondary Intrusions

Very small quartz and sulfide veins are present in the granitic body.

3.3.16.6 Country Rocks

The pluton exhibits sharply defined intrusive contacts with the metasedimentary country rocks. The contact-metamorphic aureole is approximately 1 km (0.6 mi) wide (Loiselle and Ayuso, 1980) and extends into rocks of Cambrian to Silurian age.

3.3.16.7 Rock Structure

Faulting in the Norumbega fault zone has cut and deformed the north end of the pluton. The post-Acadian displacement is a right lateral strike-slip motion parallel to the trend of the Merrimack synclinorium. Although the pluton is fractured by local jointing, there is no major joint system.

No major flow structures are present on a large scale; local gneissoid banding occurs in isolated outcrops or samples. No fold structures in the granite body have been described.

3.3.16.8 Geophysics

Gravity data of Hodge et al. (1982) suggest that this pluton most likely connects with the Bottle Lake complex in the subsurface, and may extend to a depth of 10 km (6 mi).

3.3.17 Lexington Pluton (Plate II, Pluton 17)

3.3.17.1 Size, Shape, and Location

The Lexington pluton is located in west-central Maine in the Little Bigelow Mountain, Stratton, Pierce Pond, and Kingfield quadrangles. Small portions occur in eastern Bingham and Anson quadrangles. It is elongated from north to south, with a smooth east side and a knobby west side. The length is about 35 km (22 mi); the width is about 10 km (6 mi). Outcroppings are found throughout an area of about 307 km^2 (118 mi^2).

3.3.17.2 Relative and Absolute Age

Intrusion of the Lexington into the Seboomook, the youngest metasedimentary unit in the area, places a maximum relative age of Early to Middle Devonian on the Lexington pluton. The truncation of Acadian fold structures in the metasediments indicates post-Acadian, or at least posttectonic, intrusion. The metamorphic halo of the pluton is superimposed on the contact metamorphic halo of the nearby Sugarloaf gabbroic massif, indicating that intrusion occurred after emplacement and crystallization of the Sugarloaf body (Boone, 1973). The only radiometric study of the Lexington pluton was by Naylor (1973), who gave a Rb-Sr age of 360 ± 12 m.y.

3.3.17.3 Origin

Forcible emplacement of the Lexington pluton occurred in three phases, which are reflected in the contrasting mineralogies and petrologies that occur in three lobes of the pluton. Emplacement of the pluton marked the final stage of the structural evolution of the area. Pre-intrusive Acadian fold structures to the east of the intrusion are relatively undisturbed and, at map scale, the pluton is strongly discordant to those structures. On the western side of the pluton, the surrounding country rock exhibits features directly related to emplacement of the intrusion. The close proximity of several gabbroic massifs to the west and north of the Lexington pluton also has caused superimposition of both structural and metamorphic effects on the intervening metasediments. All indications are that the granitic bodies post-date the gabbroic intrusives.

Forced emplacement of the magma was accompanied by stoping of wall and roof rocks. Large roof pendants of Carrabassett and Seboomook pelitic rocks are present, and have been metamorphosed to a granofels. Both sharp and diffuse boundaries occur at igneous rock contacts, with tongues, dikes, and sills intruding into the metasedimentary country rocks. On the basis of gravity data and internal fabric, Koller (1978) proposed that the pluton evolved by lateral spreading of pulses of granitic magma from a feeder from the north.

3.3.17.4 Petrography

Mineralogy and Textural Relations. The main mass of the Lexington pluton consists of three different mineralogies and textures (Boone, 1973). As contact relations between these three phases have not been determined, there is no reason to consider any of them as secondary.

The north lobe of the intrusion is an equigranular biotite-magnetite quartz monzonite that is coarse grained, with some minor porphyritic variants. It is massive, with no evidence of zoning due to flow or fractionation. The average bulk composition of feldspar is An-20, with a zoning from more calcic cores to albitic rims. The minor mineral, biotite, is an iron-rich phase, while the alteration product, chlorite, is of the greenalite variety. Octahedrons of magnetite occur everywhere, but in very small proportions.

The central part of the pluton consists of very coarse-grained quartz monzonites and quartz syenite porphyries, in which a concentration of granofelsic roof pendants occurs. The phenocrysts in the syenitic phase are almost entirely alkalic feldspar. Continuous gradations from quartz monzonite porphyries to quartz syenite porphyries occur. The average modal composition of the quartz monzonite porphyries is quartz, 16-30%; alkalic feldspar, 20-55%; and albitic plagioclase, 17-45%; with minor volumes of biotite, magnetite, apatite, muscovite, and tourmaline. Mineral segregation and alignment of phenocrysts are pervasive in this phase of the intrusion.

The southern lobe of the intrusion, which occurs in the northeastern quarter of the Kingfield 15-min quadrangle and the northwestern quarter of the Anson 15-min quadrangle, is a fine- to medium-grained biotite-muscovite quartz monzonite. A single petrographic mode analysis yielded the following composition: quartz, 31%; alkali feldspar, 28%; plagioclase, 28%; biotite, 5%; and muscovite, 7%; with minor to trace quantities of magnetite, apatite, and tourmaline.

Alteration. Plagioclase feldspar phenocrysts in the porphyritic phase of the intrusion have been largely altered to polygranular aggregates of muscovite, biotite, and epidote. Roughly 30% of all biotite is partially altered to chlorite and magnetite. Large crystals of muscovite occur occasionally as inclusions in plagioclase. These are probably replacement crystals, being cogenetic with chloritic alteration of biotites.

3.3.17.5 Secondary Intrusions

Although there is no field evidence for determining the order of intrusion of the three lobes, possible sequences of emplacement have been inferred on the basis of indirect petrologic evidence. If the percentage of mafic minerals and the calcic nature of the plagioclase is taken as a simple fractionation index, the crystallization sequence would be from the moderately

mafic quartz monzonite and quartz syenite porphyry to the less mafic, equigranular quartz monzonites, which have more quartz.

3.3.17.6 Country Rocks

The country rock consists of a sequence of Middle Silurian to Lower Devonian metasedimentary rocks. The Middle Silurian Perry Mountain formation consists of potassic metashales and cyclically interbedded quartz-rich metasandstones. The Middle to Upper Silurian Small Falls formation consists of black sulfidic metashales interbedded with thin, quartzose metasandstones and sulfidic calc-silicate rocks. The Upper Silurian to Lower Devonian Madrid and Carrabassett formations consist of quartzites, metagraywackes, and metaslates. The Lower Devonian Seboomook formation consists of interbedded and intercalated fine-grained sandstones, dark-gray siltstones, and slates. Coarse, pelitic granofelsic roof pendants of the Seboomook and Carrabassett formations occur in the syenitic porphyry phase of the intrusion. Complexes of igneous dike swarms also occur in these pendants.

3.3.17.7 Rock Structure

No postintrusion faulting of the igneous body has occurred. Data on joint sets could not be found in the literature.

The preferred orientation of feldspar grains and biotite flakes in the porphyritic phases of the intrusion indicates extensive flow. The absence of preferred orientation in the equigranular mineralogic phases indicates that this is a primary texture and not tectonically induced. No folding of either the intrusion or the surrounding country rock occurred in post-Acadian time.

3.3.17.8 Geophysics

On the basis of a regional gravity study by Kane and Bromery (1968), the Lexington pluton was thought to be one of three cupolas of a large granitic mass. The Phillips and Redington intrusives were proposed as the other two cogenetic bodies. Although the quartz monzonite phases of all three intrusions are similar, the syenitic phase occurs only in the Lexington pluton. The gravity field has a maximum negative anomaly over the foliated porphyritic syenite phase of the intrusion, and indicates that the intrusion is at least 2700 m (9000 ft) thick at that point. A comprehensive gravity study by Koller (1978) suggested that the Lexington pluton is roughly horizontal and tabular at depth, with a maximum vertical thickness of 3500 m (11,500 ft). It has vertical contact with the rocks in the north and dips at low angles in the other directions.

3.3.18 Lucerne Pluton (Plate II, Pluton 41)

3.3.18.1 Size, Shape, and Location

The Lucerne pluton is located in Hancock and Penobscot counties. It is exposed about 5 km (3.1 mi) from the ocean and extends inland about 66 km (41 mi) in a northeasterly direction. The extrapolated surface area is about 670 km^2 (259 mi^2). The mapped shape of the pluton is that of an elongated triangle. The Lucerne pluton intrudes into three to four fault-bounded blocks that are elongated parallel to the coast (see Fig. 3.6). The pluton is adjacent to the Norumbega fault zone on its northwestern side. Local metamorphic rocks are high-grade pelites, which have been moderately to severely tilted and folded.

3.3.18.2 Relative and Absolute Age

The Lucerne body has been dated by several means. Faul et al. (1963) arrived at a K-Ar age on mica of 365 m.y. Loiselle et al. (1983b) found an age of 371 ± 21 m.y. by the Rb-Sr whole-rock method, and 380 ± 5 m.y. by Pb-Pb analysis on zircon. The pluton cuts the northeasterly Acadian foliation.

3.3.18.3 Origin

The Lucerne granite was emplaced by forceful intrusion. The evidence includes drag folding of surrounding metasediments, inclusions of country rocks in the body of the intrusion, dikes of the intrusion in surrounding rocks, and the spreading of the magma along preexisting planes of weakness in country rocks (Wones, 1980).

3.3.18.4 Petrography

Mineralogy and Textural Relations. The subhedral biotite-microperthite granite of the Lucerne pluton has two major facies -- seriate (north and south of Sunny Side fault) and porphyritic (Wones, 1980). The pluton is abundantly coarse grained, hypidiomorphic, and sparingly kapakivi textured (only 5% of the alkali feldspar crystals are mantled by plagioclase). The principal minerals are quartz (21-36%), alkali feldspar (32-53%), plagioclase (19-30%), and biotite (2-9%). Accessory minerals include fluorite, tourmaline, ilmenite, and apatite.

Alteration. Secondary alteration of the Lucerne pluton is minor, and is represented by uralitized pyroxene, chloritized biotite, and secondary magnetite (Wones, 1980).

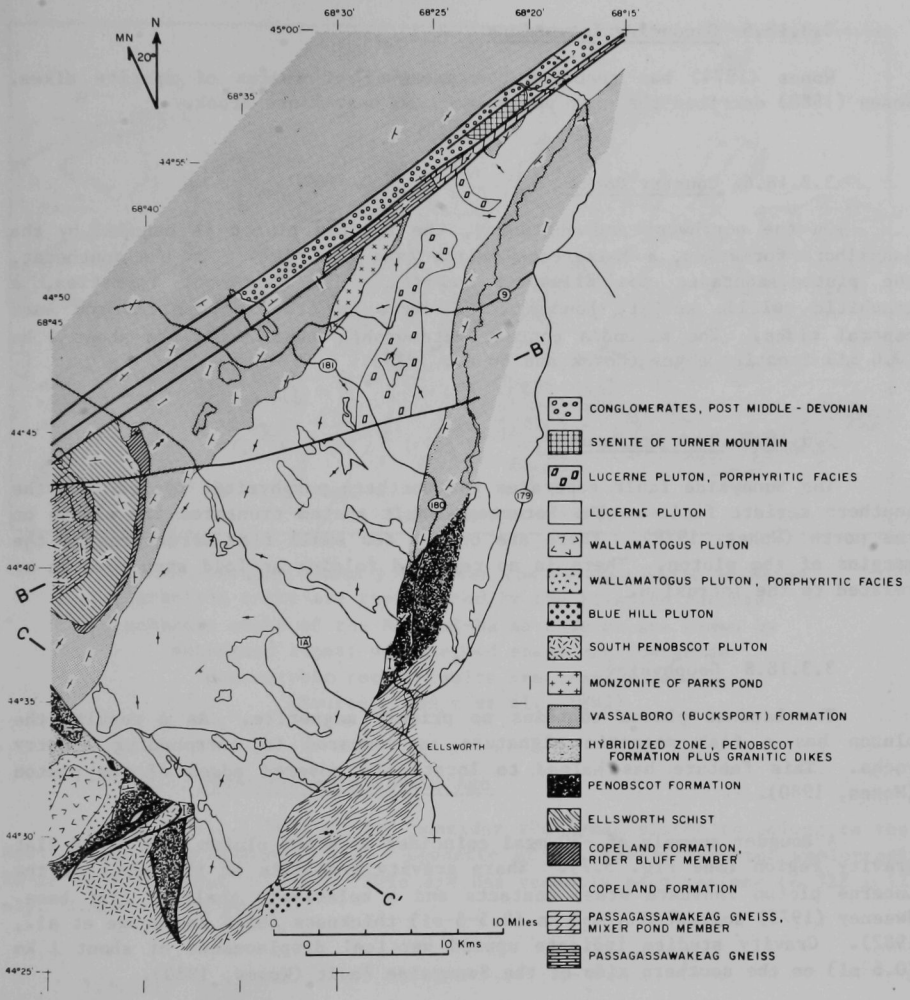

Fig. 3.6 Generalized Geologic Relations of the Lucerne Pluton (Source: Wones, 1980. Copyright 1980, Am. Mineralogist; reprinted with permission)

3.3.18.5 Secondary Intrusions

Wones (1974) has reported the presence of swarms of apatite dikes. Wones (1980) decribed the only pegmatite from near Moose Brook.

3.3.18.6 Country Rocks

On the northwest and northeast, the Lucerne pluton is bounded by the Vassalboro formation, a calcareous pelite (see Fig. 3.6). On the southeast, the pluton contacts the Ellsworth schist. The Penobscot formation, a graphitic sulfide schist, bounds the Lucerne on its southwestern and east central sides. The pluton's contact metamorphic aureole extends about 1 km (0.6 mi) from its edges (Novak and Wones, 1979).

3.3.18.7 Rock Structure

The Sunnyside fault separates the northern porphyritic facies from the southern seriate facies. The Norumbega fault system truncates the pluton on the north (Wones, 1979). There are only a few small flow structures at the margins of the pluton. There is no reported folding or fold structure in or related to the intrusion.

3.3.18.8 Geophysics

The Lucerne pluton contains no primary magnetite. As a result, the pluton has a flat magnetic signature as compared to surrounding country rocks. This feature has helped to locate the covered edges of the pluton (Wones, 1980).

A Bouguer anomaly of -30 mgal coincides with the pluton in a broad flat gravity region (see Fig. 3.7). Sharp gravity gradients at the edges of the Lucerne pluton indicate steep contacts and a relatively shallow, flat base. Sweeney (1975) suggested a 4-8-km (2.5-5-mi) thickness (also see Hodge et al., 1982). Gravity studies indicate upward vertical displacement of about 1 km (0.6 mi) on the southern side of the Sunnyside fault (Wones, 1980).

3.3.19 Lyman Pluton (Plate II, Pluton 4)

3.3.19.1 Size, Shape, and Location

The Lyman pluton is approximately 32 km (20 mi) long and 8 km (5 mi) wide, and is roughly rectangular with somewhat rounded ends. Outcrops occur within an area of about 205 km^2 (89 mi^2), primarily in the Kennebunk and Buxton 15-min quadrangles. The pluton parallels the regional structure, extending in a northeasterly direction.

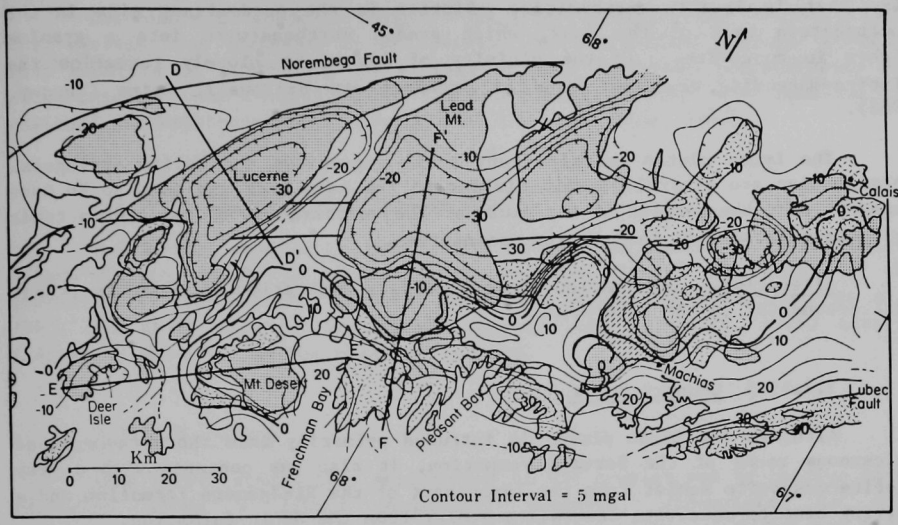

Fig. 3.7 Bouguer Anomaly Map over the Eastern Coast of Maine (granitic rocks are represented by the stippled pattern; gabbroic rocks of the Bays-of-Maine complex are shown by patterned lines; unpatterned areas are low-grade metamorphic rocks; faults are heavy lines) (Source: Hodge et al., 1982)

3.3.19.2 Relative and Absolute Age

Doyle et al. (1967) do not consider the Lyman pluton to belong to the New Hampshire plutonic series. Gaudette et al. (1975) reported whole-rock Rb-Sr ages of about 440 m.y. This age has recently been revised to 322 ± 12 m.y. (Gaudette, in press).

3.3.19.3 Origin

Page (1968) interpreted the Lyman pluton as belonging to a Late Devonian plutonic series, and as being emplaced by forceful injection and stoping.

3.3.19.4 Petrography

Mineralogy and Textural Relations. The Lyman pluton is a light-gray to pinkish-gray biotite-muscovite granite (Hussey, 1962) of fine to medium grain

size that is locally porphyritic. Biotite is the predominant mica in the southwestern half of the mass, which grades northeastward into a granite richer in muscovite. In the vicinity of Lyman, it closely resembles the biotite-muscovite granites found at Freeport and Hallowell, Maine (Hussey, 1962).

The Lyman pluton contains many small pegmatites and aplite stringers. These bodies are mineralogically similar to the main body and tend to be more abundant near the margins of the pluton. The northern end of the pluton tends to be associated with more massive pegmatites.

Alteration. Information about secondary alteration of minerals is not well documented in the literature.

3.3.19.5 Country Rocks

Although the Lyman pluton is intruded primarily into the metamorphosed calcareous rocks of the Berwick formation, it also has contacts with a gray biotite-muscovite schist from the lower part of the Rindgemere formation and a silvery biotite-muscovite-staurolite schist from the Gonic formation.

3.3.19.6 Rock Structure

The Lyman pluton shows slight foliation at its southern end and becomes increasingly foliated northward. The foliation generally follows the inferred contacts of the rock body. In the center of the pluton, the foliation trends northeastward, following the regional structural trend of the surrounding country rock. At the pluton's northern and southern ends, the foliation becomes nearly east-west in orientation (Hussey, 1962).

Hussey and Newberg (1978) interpret the topographic lineament expressed by the Lyman pluton as an indication of the reactivation of the Nonesuch River fault in the area.

3.3.20 Meddybemps and Related Granite (Plate II, Pluton 30)

3.3.20.1 Size, Shape, and Location

The Meddybemps granites and several adjacent intrusions (the Baring, Charlotte, and Cooper plutons) are small, related plutons in southeastern Maine. They together cover about 147 km^2 (57 mi^2); they are treated as one complex in this report. This belt of intrusions extends from near Calais, at the mouth of the St. Croix River, southwestward into the Bays-of-Maine igneous complex (see Fig. 3.3). The Baring granite is located southwest of the Calais gabbro-diorite; the Meddybemps granite is southeast of the Baring granite; and to the east of the Meddybemps granite is the Charlotte granite.

The composite shape is roughly that of a rectangle with rounded corners that extends from northwest to southeast. The surrounding geologic environment consists of moderately to steeply dipping low-grade metasediments and metavolcanics. The intrusives and the surrounding country rocks parallel the regional northeast-southwest structure of the area (Amos, 1963).

3.3.20.2 Relative and Absolute Age

The Baring, Charlotte, and Meddybemps granites intrude into the Calais gabbro-diorite, which indicates a Middle Devonian time of emplacement of the intrusions. Radiometeric K-Ar and Rb-Sr age data on biotite (Faul et al., 1963) and Rb-Sr data on whole rock (Spooner and Fairbairn, 1970) clearly indicate a 400-m.y. age for these intrusions.

3.3.20.3 Origin

Amos (1963) states that emplacement of the complex was by passive magma intrusion along a linear structural weakness, accompanied by some assimilation. This is supported by the gravity studies of Biggi and Hodge (1983). No mention of brecciation, xenoliths of country rock in the intrusions, or folding of country rocks due to intrusion was found in the literature.

3.3.20.4 Petrography

Mineralogy and Textural Relations. The following description is excerpted from Amos (1963). The Meddybemps granite consists of a light-salmon, medium-grained biotite granite. The alkali feldspar is not perthitic, and the plagioclase is normally zoned. The Baring granite is a homogeneous, white, medium-grained, subporphyritic biotite-perthite granite. The potassium feldspar is microcline, and the nonperthitic plagioclase ranges from An-24 to An-38. Traces of magnetite and chlorite are present. The Charlotte granite is a light-tan to light-brick-red, medium-grained biotite granite. The alkali feldspar is nonperthitic, and the plagioclase is normally zoned.

Other rock types include fine-grained biotite granite (An-23-26), fine-grained granophyre, and quartz-rich to quartz-poor quartz monzonite. Micrographic intergrowths of quartz and potassium feldspar are common, as are myrmekite, cuneiform intergrowths, and other irregular intergrowths. Another common characteristic of these intrusions is an abundance of miarolitic cavities (Abbott, 1978).

Alteration. There is little mention in the literature of secondary alteration, except for weathering.

3.3.20.5 Secondary Intrusions

Small pegmatite and aplite dikes are sparsely scattered throughout the complex (Amos, 1963).

3.3.20.6 Country Rocks

The Baring granite is bounded on the northwest by the pre-Silurian(?) Charlotte group, which consists of dark argillite, slate, mica schist, quartzite, phyllite, mica gneiss, and metamorphosed mafic volcanic flows. The Charlotte group also occurs on the southwestern side of the Meddybemps granite. To the south of he Charlotte granite occur the Pembroke formation, the Quoddy shale, and the Eastport formation. The Pembroke formation is a gray shale, which dips gently north. The Lower Silurian Quoddy shale is an easterly dipping (20°-30°) shale with intercalated rhyolites and tuffs. The Lower Devonian Eastport formation, to the south of the Red Beach granite (Spooner and Fairbairn, 1970) is a moderately northeast-dipping series of basalt, rhyolite, and andesite flows mixed in with several thin red shales and argillites. The Upper Devonian Perry formation lies to the east of the Red Beach granite, and is a gently southeasterly dipping unit of conglomerate, arkose, siltstone, shale, and interbedded basalt flows. The contact aureoles of the intrusions extend less than 460 m (1500 ft) into the country rocks (Amos, 1963).

3.3.20.7 Rock Structure

A number of northwest-trending faults cut the intrusive complex. Two faults that have each been traced for 3-5 km (2-3 mi) cut the Red Beach granite but have lateral displacements of less than 0.8 km (0.5 mi). A fault at least 8 km (5 mi) long occurs between the Charlotte granite and the Meddybemps granite. Amos (1963) shows two extensive faults along the southwestern edge of the Baring granite. The only indication of flow in the complex occurs in the Baring granite, i.e., as scattered areas of aligned microcline crystals and areas of aligned biotite crystals near contacts (Amos, 1963). Late Carboniferous folding and faulting are well known in southwestern New Brunswick not many miles to the east (Rast and Stringer, 1975).

3.3.20.8 Geophysics

Hodge et al. (1982) reported an increase in Bouguer values in this area. They also suggested that the Jonesboro and Northfield plutons may be connected at depth. The plutonic bodies are typified by circular anomalies on a residual gravity map (see Fig. 3.8) (Biggi and Hodge, 1983).

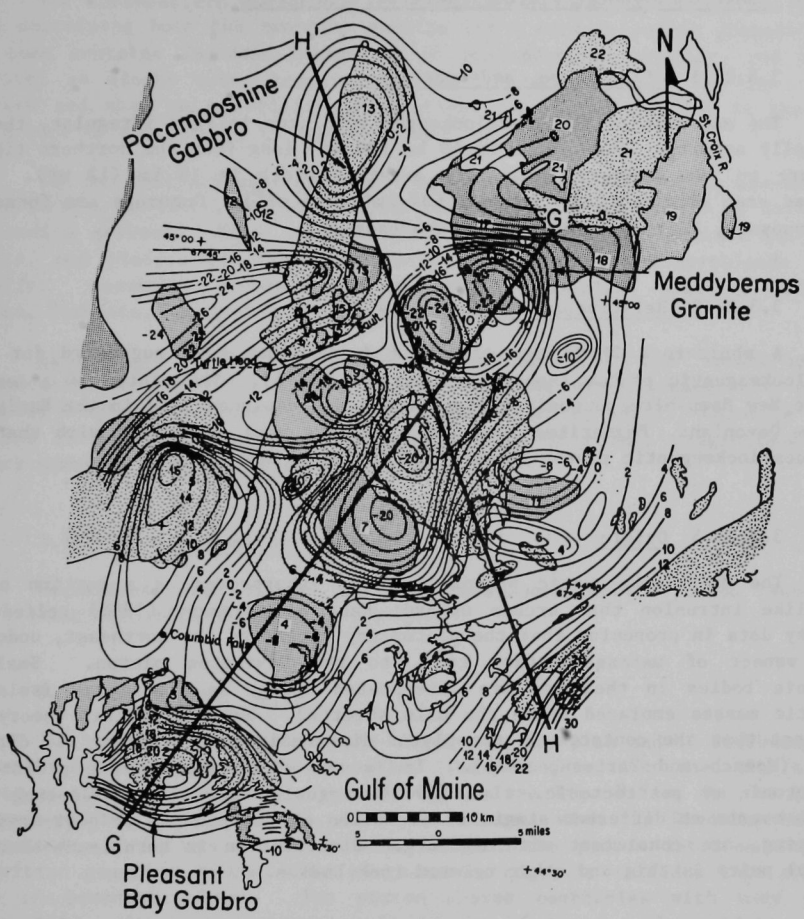

Fig. 3.8 Residual Bouguer Gravity Map for Part of the Eastern Coast of Maine (relevant mafic and felsic plutons: 5 - Peaked Mountain Pond gabbro-diorite complex; 6 - granodiorite of Whitneyville; 10 - Gardner Lake gabbro-diorite complex; 13 - Pocamoonshine gabbro-diorite; 17 - Meddybemps granite; 18 - Charlotte granite; 19 - Red Beach granite; 20 - Baring granite; 21 - Calais gabbro-diorite complex) (Source: Hodge et al., 1982)

3.3.21 Mooselookmeguntic Pluton (Plate II, Pluton 13)

3.3.21.1 Size, Shape, and Location

The map plan of the Mooselookmeguntic pluton is very irregular, though basically arcuate. The pluton is 48 km (30 mi) long from the northern tip of the arc to the southern tip. Its greatest width is 19 km (12 mi). The exposed area of the pluton is over 634 km^2 (245 mi^2). Outcrops are found in the Oquossoc, Rangeley, and Rumford quadrangles.

3.3.21.2 Relative and Absolute Age

A whole-rock Rb-Sr age of 379 ± 6 m.y. has been suggested for the Mooselookmeguntic pluton (Moench and Zartman, 1976). The pluton is a member of the New Hampshire plutonic series, which was emplaced late in the Early or Middle Devonian. Pegmatites from the area have ages consistent with that of the Mooselookmeguntic pluton (Moench and Zartman, 1976).

3.3.21.3 Origin

The Mooselookmeguntic pluton has been interpreted as a portion of a sheetlike intrusion that arches upward; Kane and Bromery (1968) relied on gravity data in proposing that the pluton may extend to the northeast, under a thin veneer of metasedimentary rock, to the Lexington pluton. Smaller plutonic bodies in the area are interpreted either as cupolas or isolated granitic masses emplaced above the main sheet(s). Supporting this theory is the fact that the contacts of the pluton dip shallowly outward in all directions (Moench and Zartman, 1976). It is not clear whether the pluton is syntectonic or posttectonic. Lux (1983) suggested that either cooling and uplift events at different stages of evolution of the pluton, or post-Acadian reheating, are consistent with the argon distribution in hornblende-biotite mineral pairs in this and other related rock bodies.

3.3.21.4 Petrography

Mineralogy and Textural Relations. The northeastern part of the pluton is composed mostly of a binary quartz monzonite. Biotite in this rock is rare or absent. Pegmatite dikes are locally abundant and form many of the higher elevations in the region because of their resistance to weathering.

The south-central portion of the pluton is composed of a two-mica granite that is light gray in color and fine to medium grained. The minerals microcline and oligoclase occur in similar percentages. Quartz accounts for nearly one third of the rock, while biotite accounts for only 1-2%. Accessory minerals are apatite and zircon.

The southeastern portion of the pluton is composed of a transitional zone containing both the two-mica granite and a sphene-bearing granodiorite. The zone contains abundant inclusions of granodiorite, tonalite, and quartz diorite, as blocky cubes roughly 1 m (3 ft) on a side. They are coarse grained and show no particular orientation. Primary foliation in the more mafic rock is cut sharply by the granite.

The far southeastern corner of the pluton is composed of sphene-bearing granodiorite, tonalite, and quartz diorite. The rocks are medium to dark gray and medium grained. Major constituent minerals include oligoclase-andesine, quartz, and biotite (13-25%). Microcline, muscovite, and hornblende occur locally. Accessory minerals include sphene, allanite, apatite, epidote, zircon, ilmenite, and pyrite (Moench and Hildreth, 1976).

Alteration. The finer-grained varieties of the sphene-bearing granodiorite in the southeastern corner of the Mooselookmeguntic pluton commonly weather to a rust color (Moench and Hildreth, 1976). Chlorite after biotite is extremely common. Saussuritization of calcic plagioclase also occurs commonly (Moench and Zartman, 1976).

3.3.21.5 Secondary Intrusions

Pegmatites are common in the Mooselookmeguntic pluton and the surrounding metasedimentary country rock.

3.3.21.6 Country Rocks

The contact zone is extremely complex in map plan and is surrounded in places by a contact metamorphic aureole. The pluton is in contact with a wide ariety of igneous and metasedimentary rocks (see Fig. 3.9). In the inner, concave portion of the pluton, it is in contact with the Devonian Umbagog granodiorite. On its outer northern margin, the pluton intrudes into an Ordovician granite; on the outer northeastern edge, it intrudes into a small body of Devonian gabbro. The pluton shares boundaries with many metasedimentary units and contains small blocks of these rocks, which may be remnants of roof pendants (Moench and Hildreth, 1976).

Along the outer northern edge of the pluton, the wall rock consists of a Cambrian-Ordovician sulfidic shale, a slate and metasiltstone of the Aziscohos formation, and other unnamed units. Also present are undifferentiated sedimentary rocks, including graywacke, sandstone, siltstone, and shale; a metamorphosed shale; and a sandstone having subordinate quartz-rich polymictic conglomerate.

The southern outer margin of the pluton is in contact with metamorphosed sandstone and shale; dark-gray sulfidic phyllite, schist, and gneiss,

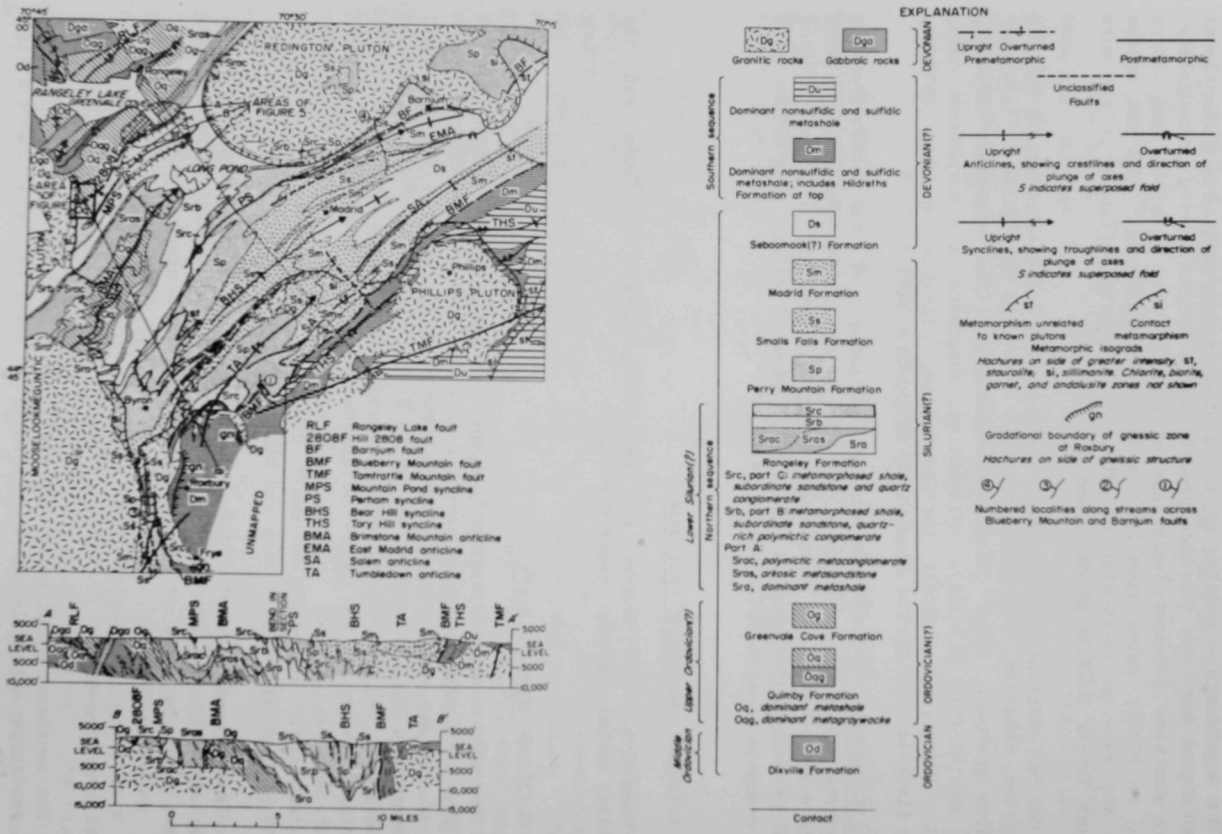

Fig. 3.9 Geologic Map of the Mooselookmeguntic and Other Plutons of the Rangeley Lake Area (Source: Moench, 1970b)

calc-silicate gneiss and granulite of the Boot and Howard Pond members of the Littleton formation; and rhythmically bedded shale and sandstone. The far southern margin is in contact with dark-gray slate, schist, gneiss (including parts of the Littleton formation above the Boot member and some of the Upper Rindgemere formation), sulfidic schist, and minor amounts of metaconglomerate and metasandstone. The inner northern margin of the pluton is in contact with a phyllite that is relatively rich in chlorite and a slate with variable amounts of interbedded quartzite. The granodiorite of the Umbagog pluton shares a boundary with the Mooselookmeguntic pluton at approximately the midpoint of the arc of its concave side.

The southern inner margin of the pluton is in contact with the Ordovician metamorphosed felsic to mafic volcanic rocks, interbedded metavolcanics of the Ammonoosuc volcanics, and minor amounts of granite gneiss, metaconglomerate, metasandstone, sulfidic metasandstone, and slate (Moench, 1970a; Moench and Hildreth, 1976).

3.3.21.7 Rock Structure

The area surrounding the eastern margin of the Mooselookmeguntic pluton abounds in apparently inactive faults that are oriented in a northeastly or northwesterly direction. Many of these are truncated by the pluton.

The primary foliation in the two-mica granite suggests syntectonic emplacement; it consists of subparallel orientation of micas and feldspars that is not immediately noticeable. The sphene-bearing granodiorite usually is well foliated.

The area surrounding the pluton has undergone at least three episodes of metamorphism. The younger metamorphic features can be attributed to a late deformation that resulted in the last two metamorphic events. The most recent deformation manifests itself in small recumbent and large flexural folds; complex slip cleavage and a related schistosity; domes; and conspicuous flexural cross folds that trend in a northwesterly direction. The fold and fault interrelationships in the Phillips and Rangley quadrangles are displayed in Fig. 3.10.

Kane and Bromery (1968) suggested that the Mooselookmeguntic pluton and several other plutons in the area are related to a subhorizontal sheetlike intrusion originally emplaced at a depth of 11-14 km (7-9 mi). Moench and Zartman (1976) noted that the eastern contact of the pluton, which dips gently to the east, could extend at depth to the Lexington pluton. They described the Lexington pluton as composed of a cupolalike mass wider than it is thick and a subhorizontal sheet believed to occur at a depth of 1.5-3 km (1-2 mi). The thickness of the sheet is believed to be about 1.5-2 km (1-1.2 mi) (Moench and Zartman, 1976).

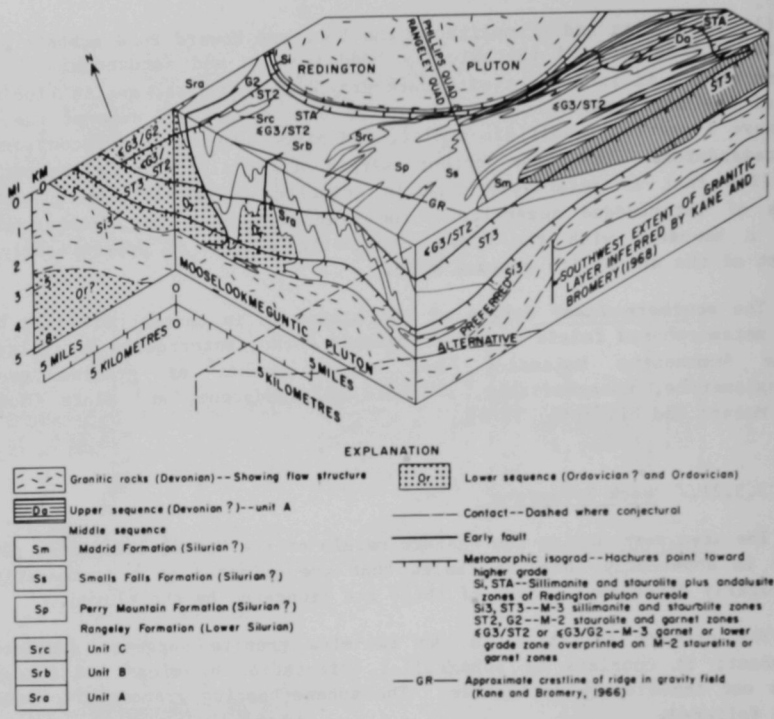

Fig. 3.10 Isometric Block Diagram of Central Parts of Rangeley and Phillips Quadrangles Showing Early Folds and Faults, Preferred and Alternative Interpretations of Plutonic Structure, Metamorphic Zoning, and Superposed Late Recumbent Folds (Source: Moench and Zartman, 1976)

3.3.22 Mt. Waldo Pluton (Plate II, Pluton 42)

3.3.22.1 Size, Shape, and Location

The Mt. Waldo pluton is located in eastern Hancock County, Maine, west of the Lucerne pluton and northwest of the Sedgwick granite. It is exposed on the west bank of the Penobscot River about 1.3 km (0.8 mi) from the ocean and extends inland for about 10 km (6 mi). Outcrops occur within a rough triangle with rounded corners, whose surface area is about 123 km^2 (47 mi^2).

3.3.22.2 Relative and Absolute Age

Zartman et al. (1970) arrived at a K-Ar age on biotite for the Mt. Waldo granite of 325 m.y. Sweeney (1975) quoted a personal communication from D.G. Brookins in which Brookins reported a Rb-Sr whole-rock age of 390 ± 10 m.y. for Mt. Waldo rocks. The Mt. Waldo pluton intruded into metamorphosed rocks older than Middle Devonian.

3.3.22.3 Origin

The mode of emplacement is disputed. Trefethen (1944) stated that flow structures within the granite and possible drag effects in surrounding country rocks indicate forceful intrusion along faults or fractures. King (1977) said that numerous small inclusions of country rock in the Mt. Waldo granite and the map pattern indicate stoping as a more likely means of emplacement.

3.3.22.4 Petrography

Mineralogy and Textural Relations. The Mt. Waldo granite is a light- to medium-gray, medium- to coarse-grained, porphyritic biotite granite (Trefethen, 1944). Perthitic microcline is the predominant potassium feldspar. Nonperthitic plagioclase is oligoclase (An-12-15) (King, 1977), and hornblende is occasionally abundant. Trace amounts of apatite, magnetite, tourmaline, zircon, zoisite, muscovite, and secondary chlorite are present.

Alteration. Secondary alteration of the Mt. Waldo granite occurs to a minor extent.

3.3.22.5 Secondary Intrusions

Local occurrences of an irregular hornblende diorite may be an early phase of the Mt. Waldo pluton. Aplite dikes are relatively common, as are simple pegmatite lenses (Trefethen, 1944).

3.3.22.6 Country Rocks

The northern half of the Mt. Waldo granite is bounded by the Passagassawakeag gneiss. The Vassalboro formation (Kellyland-Bucksport formation), a calcareous pelite, occurs to the east and southwest. The southeastern corner of the pluton is bounded by the Penobscot formation, a graphitic sulfide schist. The pluton's metamorphic aureole extends out less than 2 km (1.25 mi) and is a hornblende hornfels facies (Winkler, 1967).

3.3.22.7 Rock Structure

Map-scale joints are infrequent, but strike roughly west-northwest. A similar set, possibly related to the set within the pluton, occurs in adjacent metamorphic rocks. Many of the fractures show minor offsets of dikes and slickensides (Trefethen, 1944).

The Mt. Waldo granite is famous for its rock bursts, which are sudden and often violent breakings of rock caused by the rapid or instantaneous release of accumulated strain energy in highly stressed rock masses (Lee et al., 1979).

Much of the Mt. Waldo granite contains a distinct flow pattern, parallel to the contacts, that is formed by subparallel feldspar phenocrysts (Trefethen, 1944). There are no major folds in or related to the Mt. Waldo pluton.

3.3.22.8 Geophysics

Gravity studies (Sweeney, 1976) and modeling indicate an average thickness for the Mt. Waldo granite of less than 7 km (4.4 mi); much of the body is probably about 5 km (3 mi) thick. The southern contact is near vertical (Sweeney, 1975); the western side dips at an angle of less than 60° west; and the eastern side dips about 75° west (Sweeney, 1976). The gravity low is outlined by a -20 mgal contour on the Bouguer map (see Fig. 3.7).

3.3.23 Moxie Pluton (Plate II, Pluton 23)

3.3.23.1 Size, Shape, and Location

The Moxie pluton is a mafic intrusive body located in west-central Maine, at the southwestern end of the Mt. Katahdin pluton. The pluton is about 70 km (43 mi) long and, on average, about 5 km (3 mi) wide. Three major protuberances of greater width occur in the south and central portions of the intrusion. Outcrops are found within a total area of approximately 371 km^2 (143 mi^2). Gravity survey data (Kane, 1961) suggest that the northeastern part of the pluton may have the form of a 3-km (1.8-mi) thick, steeply dipping dike, and that the wide irregular areas farther northeast and southwest reflect undulations in a large sheetlike body that is several miles thick (Espenshade, 1972). The long axis of the intrusion has an approximate strike of N. 55° E., which is parallel to the strikes of regional structures. Outcrops are found in the southern half of First Roach Pond quadrangle, the southeastern quarter of Moosehead Lake quadrangle, the northwestern quarter of the Greenville quadrangle, the southeastern quarter of the Forks quadrangle, and the northern quarter of the Bingham quadrangle.

Metamorphosed country rocks form topographically high landforms, and unmetamorphosed sedimentary and igneous rocks have weathered into lowlands and valleys. Morainal and outwash deposits from Pleistocene glaciers cover most of the bedrock in the area.

3.3.23.2 Relative and Absolute Age

The mafic rocks of the pluton intrude into Upper Silurian and Lower Devonian metasedimentary rocks (Espenshade and Boudette, 1967; Espenshade, 1972). Adjacent to the pluton these rocks have been metamorphosed to hornfels, indicating that emplacement and metamorphism were contemporaneous. This places a maximum relative age of Early to Middle Devonian on the intrusive event. The absence of younger rocks precludes estimating a minimum relative age. Faul et al. (1963) report a K-Ar date of 393 m.y. from biotite contained in a norite sample.

3.3.23.3 Origin

Field investigations of textural and mineralogic variations (Espenshade and Boudette, 1967; Espenshade, 1972), coupled with magnetic data (Kane and Bromery, 1968), indicate that the intrusive is thick, steeply dipping, and sheetlike. As indicated by magnetic data, the body has a 60° dip southeasterly, with the southeastern contact being the top of the intrusive and the northwestern contact the floor. Texture is produced by discordant flow orientation of phenocrysts with respect to the wall rock. Both the textural and geophysical data suggest that the large oval protuberances are conduits. The width of the contact aureole is much larger (up to 3 km [2 mi]) than those associated with the more felsic intrusives in Maine. This can be attributed to mafic magmas being hotter.

The divergence of fold axes in the country rocks as they approach the intrusion indicates a degree of forceful injection (Espenshade and Boudette, 1964). There are few reported occurrences of xenoliths in the pluton, but this could be due to complete melting of inclusions by the high-temperature magma. The inclusions that do occur are xenoliths of hornfels, and they occur only adjacent to the contacts. Very tight minor folds occur in the hornfels adjacent to the pluton but not in the nonmetamorphosed sedimentary rocks, which also indicates forceful injection (Espenshade and Boudette, 1967). Schulman and Hon (1983) proposed, from mineralogic and geochemical considerations and the distribution of high-magnesium and high-iron rocks, that the central part of the pluton resulted from fractionation of a high-aluminum gabbroic magma.

3.3.23.4 Petrography

Mineralogy and Textural Relations. The Moxie pluton consists of several related mafic rock types. Troctolite and norite are the most abundant, with minor occurrences of gabbro, dunite, and diorite. The troctolite is a dark, medium-grained rock composed of plagioclase (50-80%), olivine (15-40%), orthopyroxene (4-12%), clinopyroxene (0-3%), and minor amounts of hornblende, biotite, opaque minerals (probably magnetite), and apatite (Espenshade, 1972). Plagioclase composition is An-55-75, while that of olivine is Fo-26-75. Troctolite dominates in the northern and western outcrops, indicating that it may form the dominant rock on the floor of the intrusion.

Norites intergrade with the troctolite; they are brown, medium- to fine-grained rocks with a modal composition of plagioclase, 50-85%; olivine, 0-6%; orthopyroxene, 13-40%; clinopyroxene, 0-10%; and hornblende, 0-10%; with minor amounts of biotite, magnetite, cummingtonite, and apatite (Espenshade, 1972). Plagioclase composition is An-25-50; olivine is of unknown composition. Norite is the dominant rock type along the southeastern contact of the intrusion. In general, magnesium-rich mineralogies dominate the southern portions, and iron-rich mineralogies dominate the central and northern portions of the intrusion. Acheson and Hon (1983) reported that cumulate and noncumulate portions define the iron-enriched and calc-alkaline phases, respectively.

Alteration. The mafic rocks of the Moxie pluton characteristically are very fresh and unmetamorphosed (Espenshade, 1972). Minor alteration does occur in the dioritic phases, which are a less common rock type in the complex. The absence of weathered or altered zones has been attributed to differential erosion during Pleistocene glaciation. The altered and weathered rocks were stripped off because they were less resistant.

3.3.23.5 Secondary Intrusions

The Moxie pluton contains a diverse suite of mafic rock types ranging from gabbros to diorites. When compared to other major stratiform intrusives containing the same suite of rocks, the Moxie intrusive mass is conspicuous in that it lacks regular cyclic layering (Espenshade, 1972). Poor bedrock exposure complicates attempts to relate individual rocks to differentiation schemes. Although classification of the various mineralogies as either primary or secondary is not supported by field evidence, the complex interrelationships and distribution of known rocks indicate multiple intrusions.

Although rock types other than the norite or troctolite occur, no secondary dikes or sills are recognized to be the result of separate magmatic events. The Moxie pluton is cut by fine-grained felsic dikes at Rum Cove and lower Wilson Pond. Additional exposures of small dikelike bodies of felsic

rock at Trout Pond and north of Moore Bog do not appear to intrude into the mafic rock (Espenshade and Boudette, 1967). In the northeastern to central part of the pluton small, fine-grained granitic bodies of potassium feldspar, quartz, and biotite are exposed along the eastern side of Moosehead Lake and east of Beaver Cove. Espenshade (1972) mapped these as small dikes or sills of quartz monzonite and granodiorite. He also observed quartz-feldspar pegmatites a few inches thick in that part of the Moxie pluton. However, none of these occurrences can be proven to be differentiated secondary intrusions.

3.3.23.6 Country Rocks

The Devonian sedimentary country rocks have been subjected to four separate events leading to four characteristic mineralogies: (1) primary sedimentation, (2) secondary regional metamorphism, (3) contact metamorphism, and (4) retrograde metamorphism resulting from diagenesis. The primary sedimentary mineralogy occurs in a lower 760-m (2500-ft) section of limey sandstone capped by 1500-1800 m (5000-6000 ft) of pelitic rocks. The pelitic sequence consists of a 150-m (500-ft) shale unit interbedded with fine-grained sandstone and feldspathic sandstone, a middle 900-1050-m (3000-3500-ft) shale unit with very localized thin beds of sandstone, and an upper 600-750-m (2000-2500-ft) sequence of dark gray shales commonly interbedded with graded sandstone beds 13-30 cm (5-12 in.) thick. Prior to intrusion, the entire sequence experienced a regional metamorphism to chlorite grade, which transformed the shales to slates.

The slates, siltstones, and sandstones intruded into by the Moxie and Katahdin plutons have been thermally altered to hornfels-grade metamorphic rocks. These contact aureoles are more resistant to weathering and erosion than the crystalline igneous rocks, and stand out as prominent ridges around the intrusives. In many circumstances, these ridges serve as the sole indicator of the boundaries of the intrusives. Directly adjacent to the Moxie pluton, the sillimanite-cordierite metamorphic zone has a gneissic to granoblastic texture, with some vestiges of original bedding. Accessory minerals are quartz, perthitic potassium feldspar, rare zoned plagioclase, biotite, muscovite, and andalusite. This zone is occasionally absent but may extend up to 3 m (10 ft) from the intrusion, at which point it is replaced by the andalusite-amphibole metamorphic belt. This zone, which varies from less than 300 m (1000 ft) to more than a kilometer (several thousand feet) wide, consists of a light-gray schist. The grain size varies from coarse to fine across the zone. The inner coarse-grained portion contains the accessory minerals quartz, biotite, feldspars, and muscovite. Actinolite and tremolite are common in the more calcareous sequences. After the disappearance of andalusite, biotite persists as the lowest-grade contact metamorphic index mineral for up to 8 km (5 mi) from the intrusion.

The mineral isograds for the higher-grade zones are parallel to the contacts of the intrusion. The biotite isograd differs in that it intersects

and joins biotite isograds for the adjacent Onawa dioritic pluton, Bald Mountain stock, and Shirley Blanchard stock. All of these isograds have wider zones south of the intrusive. This widening can be attributed to two effects -- the southerly dip of the intrusive and the greater heat flux from the top of the intrusion.

3.3.23.7 Rock Structure

Structural deformation of the region apparently took place before intrusion. Although the country rock is folded and fractured, shear zones and associated cataclastic structures in the Moxie pluton are rare (Espenshade and Boudette, 1967). Joints and shear zones trend northwest and are sometimes incompletely lined by the alteration mineral cummingtonite. South of Big Indian Pond, on the southeastern contact of the pluton, are two fault exposures (Espenshade and Boudette, 1967).

Plagioclase crystals arranged in planar orientation parallel to compositional layering produce a laminated flow structure dipping 20°-60° N. Two interpretations of this flow structure are discussed by Espenshade and Boudette (1967). One is that the linear orientation of plagioclase crystals was created by significant convection currents during fractional crystallization of the magma. The second interpretation, which does not preclude the first, is that the lamination formed principally by gravitational settling but may have been accompanied by convection currents causing linear orientation. The flow structures at the margins of the pluton are nearly conformable with the southeast-dipping contacts. Within the intrusive body, however, flow structures are highly discordant to the dip of the contacts (Pfann, 1979).

The Moxie pluton and associated granitic stocks have undergone deformation. Since the surrounding country rocks are tightly folded, the pluton apparently intruded and solidified after the events that deformed the country rock. There are small-scale, tight folds within a limited region of the contact metamorphic aureole that have been attributed to forceful intrusion of the magma (Espenshade and Boudette, 1967).

3.3.23.8 Geophysics

An aeromagnetic survey by Bromery et al. (1963) showed strong magnetic anomalies related to the Moxie pluton, which indicate the structural orientation of the contacts and the occurrence of associated buried gabbro bodies. High magnetic anomalies occur on Little Squaw Mountain (27 units) and west of Blue Ridge between Rum Mountain and Cranberry Pond (25 units) on the southern side of the Moxie pluton. The magnetic data on the north side of the pluton do not exhibit a pattern as closely related to the shape of the mafic body (Espenshade, 1972).

On the basis of paleomagnetic studies, Pfann (1979) concluded that the Moxie pluton is a concordantly intruded, 500-to-2500-m (1640-to-8200-ft) thick, sill-like body that has been folded into steeply plunging folds in a post-emplacement deformation of moderate to severe intensity.

Data from a gravity survey by Kane and Peterson (1961) show that the Moxie pluton is near a major, conspicuously linear, positive gravity anomaly (Espenshade and Boudette, 1967).

3.3.24 Peaked Mountain Pond Gabbro-Diorite Complex (Plate II, Pluton 33)

3.3.24.1 Size, Shape, and Location

Peaked Mountain Pond is an oval complex of gabbros and diorites in eastern Maine between the Lead Mountain pluton and the Whitneyville granodiorite (Westerman, 1979). The complex underlies an area of about 112 km^2 (43 mi^2); it is an oval about 7.5 km (4.7 mi) long and 3.8 km (2.3 mi) wide.

3.3.24.2 Relative and Absolute Age

Doyle et al. (1967) assigned a Devonian age to the complex.

3.3.24.3 Origin

The Peaked Mountain Pond pluton is part of the mafic igneous intrusion represented by gabbro-diorite complexes, e.g., Gardner Lake and Calais (Amos, 1963) along the east coast of Maine.

3.3.24.4 Petrography

Mineralogy and Textural Relations. The gabbro-diorite is medium to coarse grained. It is primarily composed of plagioclase, hornblende, and pyroxene.

Alteration. Evidence of alteration is not reported in the literature.

3.3.24.5 Secondary Intrusions

Late-stage granitic differentiates form small intrusions (Chapman, 1962).

3.3.24.6 Country Rock

The Lead Mountain pluton is in contact with the west side of the Peaked Mountain Pond pluton. To the east lies the granodiorite at Whitneyville. Peaked Mountain Pond is bordered by slate, schists, gneisses, and quartzite to the south and west.

3.3.24.7 Rock Structure

No major structural features are described for the body.

3.3.24.8 Geophysics

On the basis of gravity data, Biggi and Hodge (1983) suggested that the Peaked Mountain body is probably a sheetlike intrusion 1.5-3 km (1-2 mi) thick.

3.3.25 Phillips Pluton (Plate II, Pluton 12)

3.3.25.1 Size, Shape, and Location

The Phillips pluton is located in west-central Maine in the southern half of the Phillips 15-min quadrangle and in the western half of the Dixfield 15-min quadrangle. The pluton has an irregular, elongate shape, with two lobes connected by a narrow neck. The northeastern lobe is approximately 10 km (6 mi) long from the northeast to the southwest and 6-8 km (4-5 mi) wide. The larger lobe lies to the southwest and is 24 km (15 mi) long and 8-10 km (5-6 mi) wide. Outcrops occur in an area of about 218 km^2 (84 mi^2).

3.3.25.2 Relative and Absolute Age

Geophysical evidence indicates that the Phillips and nearby Redington and Lexington plutons are cupolas of a continuous subsurface mass (see Secs. 3.3.17 and 3.3.29). Therefore, the age of the Phillips pluton is probably the same as that of the other two plutons, i.e., Early to Middle Devonian. The Mooselookmeguntic pluton, a short distance to the west and possibly continuous with the Phillips pluton at depth, has a Rb-Sr age of 379 ± 6 m.y. (Moench and Zartman, 1976). This is approximately the boundary between Early and Middle Devonian.

3.3.25.3 Origin

Plutons of the New Hampshire plutonic series have two contrasting modes of emplacement. They were intruded either by forced injection or by passive

digestion of country rock along margins and roofs. The Phillips pluton was forcibly intruded. Its contacts with country rocks are sharp and steep, and exhibit both concordance and discordance on a local scale (Moench and Zartman, 1976). The bedding, foliations, lineations, and fold axes of the metasedimentary country rocks wrap around the pluton. Megascopic foliation in the core of the pluton is absent, implying post-tectonic emplacement. A faintly preferred orientation of micas, localized near the contacts, has been interpreted as a slip cleavage formed during emplacement (Pankiwskyj, 1964).

3.3.25.4 Petrography

Mineralogy and Textural Relations. The rocks of the Phillips pluton are fine to medium grained and typically equigranular. Minerals exhibit random orientations except near contacts with country rocks. The pluton consists of granodiorite, with minor amounts of quartz monzonite and granite. The granodiorite phase is leucocratic, with mafics comprising less than 7%. The modal composition is 15-20% quartz and 75-80% feldspar, with a 7:1 ratio of plagioclase to microcline. Minor mineral phases (less than 1%) are muscovite, biotite, epidote, and sphene; trace quanities of apatite, zircon, and magnetite occur. The only volumetrically significant rock of different character, which was intruded synchronously with the granodiorite, is a quartz diorite, which occurs at or near contacts with country rocks. This phase is slightly more mafic than the coarser-grained granodiorite, with biotite composing up to 15% (Pankiwskyj, 1964; Moench and Zartman, 1976).

Alteration. Saussuritization (replacement of a mineral by the assemblage zoisite, epidote, albite, calcite, sericite, and zeolites) of the feldspars has been noted by Moench (1954), and probably reflects late deuteric alteration. Replacement of the biotites by chlorite also has been observed (Pankiwskyj, 1964). The K-Ar dating of biotite, which indicated a metamorphic re-equilibration of the rock during the Pennsylvanian, is further evidence of extensive alteration at the onset of the Alleghenian orogeny (Pankiwskyj, 1964).

3.3.25.5 Secondary Intrusions

Two minor secondary intrusions occur in the pluton. The oldest is a light-colored quartz monzonite, which contains inclusions of the preexisting granodiorite. The monzonite and granodiorite are, in turn, pervasively cut by late-stage aplitic and pegmatitic dikes. The pegmatites, which have a greater density in the country rock than in the pluton, consist of quartz, clevelandite, potassium feldspar, biotite, black tourmaline, garnet, apatite, and some beryl. The quartz monzonite differs from the main mass of the intrusion in having a higher percentage of muscovite and a lower percentage of biotite (Moench, 1954; Pankiwskyj, 1964).

3.3.25.6 Country Rocks

The country rock surrounding the pluton consists of approximately 6000 m (20,000 ft) of metasedimentary rocks. This thickness has been subdivided into three stratigraphic units on the basis of lithology. The basal sequence is about 1200 m (4000 ft) of metamorphosed Silurian sandstones, calcareous sandstones, shales, calc-silicate rocks, and noncalcareous shales. The middle sequence is 1500 m (5000 ft) of Lower Devonian sulfidic to nonsulfidic shale, sandstone, calcareous sandstone, shale, and calc-silicate rock. This unit is capped by the Lower Devonian Hildreth formation, which consists of calcareous metashale. The upper sequence is 2100 m (7000 ft) of interbedded sandstones and shales (Osberg, 1968). These rocks have been metamorphosed above the regional metamorphic grade to hornfels around the pluton for a distance of 1 km (0.6 mi) from the contact. The hornfels facies forms a resistant ridge around the easily eroded granodiorite (Moench, 1954).

3.3.25.7 Rock Structure

Two phases of development of joint sets are recognized in the Phillips pluton. The first phase consists of joints and fractures filled with aplite and pegmatite. Poor outcrop exposure of the easily eroded pluton prevents determination of the regional attitudes of this set. The second set, which is devoid of filling, cuts across both the pluton and the country rock. These joints have three different structural attitudes. The strongest set is vertical and strikes east to west, perpendicular to the regional structure. The second most prominent set is near vertical and parallels the regional structure, with a north-to-south strike. A minor northwest-to-southeast-striking set is present, which is also vertical (Moench, 1954).

A large fault, the Tamtrattle Mountain fault, cuts across both the intrusion and the country rock. The fault parallels the contact of the northern lobe of the intrusion in the Phillips quadrangle and cuts across the narrow neck connecting the two lobes of the intrusion. The fault is associated with a heavily silicified zone and areas of slip cleavage in adjacent rocks.

The Phillips pluton is a mainly equigranular, nonfoliate body. The only noticeable flow texture is at or near the margins of the pluton, where a slip cleavage has oriented micas parallel to the regional structure (Pankiwskyj, 1964). The pluton shows no evidence of postintrusive folding.

3.3.25.8 Geophysics

A model for the three-dimensional shape of the Phillips pluton has been developed by Kane and Bromery (1968) using gravity data. The results of their analysis indicate that the pluton is a cupola of a granitic sheet that includes both the Reddington and Lexington plutons. In the area between

outcrops of the three cupolas, the granite is estimated to be 1500-3050 m (5,000-10,000 ft) below the surface, and to have an average thickness of 1500 m (5000 ft). These interpretations are tenative at best, as the gravity data were never terrain-corrected (Moench, 1976).

3.3.26 Pierce Pond Pluton (Plate II, Pluton 18)

3.3.26.1 Size, Shape, and Location

The Pierce Pond pluton of northwestern Maine crops out in the Pierce Pond, Little Bigelow Mountain, and Stratton 15-min quadrangles. The pluton extends over 141 km^2 (54 mi^2); its long axis trends in a northeasterly direction (Boone, 1970) for about 23 km (14 mi). Much of the western half of the pluton is covered by Flagstaff Lake, while Pierce Pond lies predominantly in the northeastern part of the pluton. A small lobe of the pluton extends away from the main body in a southeasterly direction into the Lexington pluton. The Pierce Pond pluton is intimately associated with the Flagstaff pluton and may be continuous with the Kennebago pluton, the area between them being occupied by the Flagstaff pluton (see Fig. 3.11).

3.3.26.2 Relative and Absolute Age

The Pierce Pond pluton is a member of the New Hampshire plutonic series (Boone, 1970 and 1973) which is, in Maine, a northeasterly trending series of granodioritic to gabbroic rocks, from the Umbagog pluton on the Maine-New Hampshire border to the Moxie pluton in north-central Maine. Radioisotopic data indicate that these bodies were emplaced in the Early to Late Devonian, an age that is supported by the stratigraphic relations of the plutons.

3.3.26.3 Petrography

Mineralogy and Textural Relations. The Pierce Pond pluton is composed chiefly of normal gabbro but also contains lesser amounts of noritic gabbro (Boone, 1970 and 1973). The gabbro is massive, nonfoliated, and coarse grained. Major minerals include plagioclase, clinopyroxene, orthopyroxene, and olivine. Lesser amounts of the following minerals, including accessory minerals, are found: ilmenite, magnetite, chalcopyrite, pentlandite, pyrrhotite, hornblende, apatite, and biotite. The presence of these minerals from one locality to another is variable. Along its southwestern edge and at the northwestern tip of the body, part of the pluton abruptly becomes granitic in composition. Major minerals of this phase include plagioclase, biotite, quartz, and potassium feldspar. Lesser amounts of apatite and magnetite are found, while ilmenite occurs as an accessory mineral.

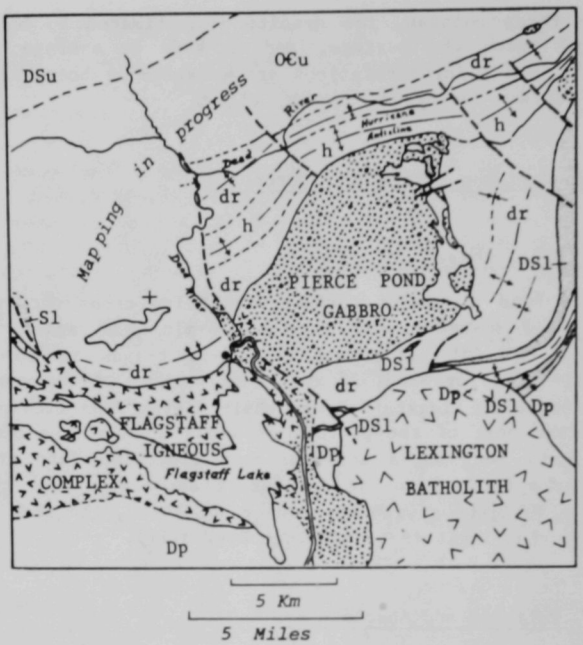

Fig. 3.11 Generalized Geologic Map of Southeastern Spencer, Southern Pierce Pond, Northeastern Stratton, and Northern Little Bigelow Mountain Quadrangles (Source: Boone et al., 1970)

Alteration. Limited alteration of mineral constituents is present.

3.3.26.4 Secondary Intrusions

Veins and dikes of a biotite-tourmaline pegmatite cut portions of the southern part of the pluton (Boone, 1973).

3.3.26.5 Country Rocks

The Pierce Pond pluton is surrounded by a contact-metamorphic aureole. The pluton has metamorphosed portions of the slate and metasiltstone of the Aziscohos formation, which is exposed around all except the southern edge of the pluton. This formation also is in contact with the younger

Flagstaff and Lexington plutons and the dark-gray slate and metasandstone of the Seboomook formation (Doyle et al., 1967).

3.3.26.6 Rock Structure

Minor evidence of flowage in the Pierce Pond pluton is present on the east side of Pierce Pond Mountain, Hedgehog Mountain, and on an island in Otter Pond Cove. The banding is entirely local and displays some crystal sorting and settling. Little compositional variation occurs among the layers.

3.3.26.7 Geophysics

The Pierce Pond pluton is situated in a zone of predominantly northeast-trending inactive faults. There are also faults that strike in a northwesterly direction.

3.3.27 Pocamoonshine Gabbro-Diorite (Plate II, Pluton 34)

3.3.27.1 Size, Shape, and Location

The Pocamoonshine gabbro-diorite is a north-to-south-trending intrusive body in eastern Maine. It is approximately 29 km (18 mi) long and 10 km (6 mi) wide; outcrops occur over approximately 141 km^2 (54 mi^2) in the Big Lake and Wesley 15-min quadrangles.

3.3.27.2 Relative and Absolute Age

The relative age of the Pocamoonshine gabbro-diorite is Devonian (Doyle et al., 1967) because it intrudes into Silurian and older metasedimentary units and is crosscut by granitic dikes of Devonian age. No published absolute age is available.

3.3.27.3 Origin

The intrusive magmatic origin of the gabbro-diorite body is indicated by the inclined contacts, rock textures, the contact-metamorphic aureole, and inclusions of contact-metamorphosed metasiltstone and slate. Smaller inclusions have a lit-par-lit structure, the planes of which strike N. 60°-80° E. and dip vertically. Page (1968) states that the mode of emplacement of a group of Late Devonian plutons was by forceful injection and stoping.

3.3.27.4 Petrography

Mineralogy and Textural Relations. The Pocamoonshine gabbro-diorite is a medium-grey-green to black gabbro-diorite ranging from a fine-grained, salt-and-pepper hornblende diorite to a coarse-grained olivine gabbro (Larrabee et al., 1965). The olivine gabbros contain mineral grains greater than 8 mm (0.3 in.) in diameter. The rock texture varies from gabbroic to diabasic. Minerals in the gabbro-diorite include clinopyroxene, olivine (altered to brown hornblende), poikilitic primary hornblende, magnetite, apatite, and reddish-brown biotite.

North of the main gabbro-diorite body, near the mouth of Berry Brook Flowage in Waite quadrangle, a coarsely crystalline rock resembling the Pocamoonshine gabbro-diorite crops out on three very small, closely spaced islands (Larrabee, 1964a). The rock is a coarse-grained granodiorite consisting of dark-green to black hornblende crystals averaging 3-5 mm in length, white plagioclase, orthoclase, and green-black biotite that contains zircon. Megascopic petrographic similarities with the Pocamoonshine gabbro-diorite suggest that it may connect beneath the surface to the main body of gabbro-diorite (Larrabee et al., 1965).

Alteration. Most of the original olivine and clinopyroxene grains have altered to brown hornblende. Schillerization is apparent in the clinopyroxene grains, and relict schiller structure persists in the altered grains. The granodiorite occurrence directly north of the Pocamoonshine gabbro-diorite pluton exhibits alteration of biotite to chlorite and plagioclase to sericite.

3.3.27.5 Secondary Intrusions

The gabbro-diorite is intruded into by dikes of medium-grained biotite granite 15-20 cm (6-8 in.) wide.

A pegmatitic facies containing graphic intergrowths of feldspar and hornblende aggregates up to 15 x 10 cm x 8 cm (6 x 4 x 3 in.) and crystals as large as 1.2 x 2.5 cm (0.5 x 1 in.) is exposed along Dog Brook.

3.3.27.6 Country Rocks

The country rocks into which the Pocamoonshine gabbro-diorite was intruded are early Paleozoic metasedimentary units, i.e., the dark argillite division of the Charlotte Group of Ordovician age (phyllite, schist, slate, and quartzite) and the argillaceous-siliceous Kellyland formation of Silurian age (mainly metasiltstone). The contacts between the intrusive and the country rock dip outward, the eastern side more steeply than the western side (Larrabee, 1964a; Larrabee et al., 1965). The eastern contact is faulted, and decreases in inclination beneath the surface to the east (Larrabee, 1964a). Tongues of intrusive gabbro alternate with hornfels along the western and

southern contacts, where the pluton interfingers with the contact-metamorphosed Kellyland formation (Larrabee, 1964a; Larrabee et al., 1965).

The aureole is well developed and extends approximately 0.8-4 km (0.5-2.5 mi) away from the pluton in map plan. Contact aureoles are marked by increased amounts of magnetite, biotite, sillimanite, graphite, and crystals of chiastolite up to 1 x 16 cm (0.5 x 6.5 in.) in size. The Kellyland formation also contains cordierite, andalusite, pyrite, and pyrrhotite at its thermally retrograde-metamorphosed contact with the Pocamoonshine gabbro-diorite.

3.3.27.7 Rock Structure

The pluton is traversed by joints, but the joints are not organized in a major system. The body is faulted along its eastern contact by a northeast-trending fault that dips steeply to the east. The northern quarter of the body is offset by an east-northeast-trending fault that extends across a large area, including the entire width of the Pocamoonshine gabbro-diorite in the Big Lake quadrangle.

The structural lineations of the gabbro-diorite have a northerly trend and may reflect intrusive flow patterns. The Pocamoonshine gabbro-diorite is not folded. It does contain rare local layering in the contact zone. The layering dips steeply to vertically, which precludes a relationship to the dip of the contact (Larrabee, 1964a).

3.3.27.8 Geophysics

The pluton is marked, in part, by a circular gravity anomaly enclosed by +2- to -6-mgal contours (see Fig. 3.8).

3.3.28 Quaker Hill Pluton (Plate II, Pluton 7)

3.3.28.1 Size, Shape, and Location

The Quaker Hill pluton is a two-mica granite that lies west of Augusta on the east bank of the Androscoggin River. The pluton is arc-shaped, about 20 km (12 mi) long by 5 km (3 mi) wide, and underlies an area of about 87 km^2 (33 mi^2) in the Livermore Falls and Lewiston quadrangles.

3.3.28.2 Relative and Absolute Age

No radiometric age is available for the Quaker Hill pluton, but lithostratigraphic correlation indicates that it is part of the New Hampshire

plutonic series (Doyle et al., 1967). Moench et al. (1982) indicated that Quaker Hill is lithologically similar to the Sebago batholiths, and also the Mooselookmeguntic pluton. However, Gaudette et al. (1975) and Lyons (1979) proposed that the two-mica granites may be Carboniferous or Permian in age.

3.3.28.3 Origin

Quaker Hill lies in the high-grade area of Acadian metamorphism. As it has similarities to the S-type granites, it may be derived from partial melting of metasedimentary source rocks. Its intrusive character is reflected in its discordant relations to the bedding and foliation in the Silurian-Devonian(?) sequence. Xenoliths of polydeformed rocks occur in the intrusion. The smaller plutonic bodies in the area of the Mooselookmeguntic pluton (Moench and Zartman, 1976) and the Sebago Lake pluton (Hodge et al., 1982) may be cupolas or isolated granitic masses emplaced above a main, sheet-like batholithic intrusion.

3.3.28.4 Petrography

Mineralogy and Textural Relations. The Quaker Hill pluton is composed of a pale gray, fine- to medium-grained, equigranular two-mica granite that is massive or weakly foliated. Locally, muscovite granite and two-mica granodiorite are found. Fisher (1941) mapped the entire pluton as a granodiorite. Major minerals are microcline, oligoclase, quartz, and biotite. Garnet, tourmaline, sillimanite, zircon, and apatite are accessory minerals (Moench et al., 1982).

Alteration. Quaker Hill lies in a sillimanite isograd (Moench and Zartman, 1976).

3.3.28.5 Secondary Intrusions

Pegmatites and aplites are common in the Quaker Hill pluton. The northeast corner of the pluton was intruded into by a Mesozoic alkalic gabbro and syenite in the Androscoggin Lake area (Moench et al., 1982).

3.3.28.6 Country Rocks

The Quaker Hill pluton is almost totally surrounded by Silurian-Devonian(?) metamorphosed calcareous sandstone, siltstone, and graywacke. The pluton also has metamorphosed quartz wacke, conglomerate, siltstone, and shale as wall rock to the north, and the Waterville formation as wall rock to the west (Doyle et al., 1967). All contact relations are discordant.

3.3.28.7 Rock Structure

Even though the pluton has a general orientation consistent with the regional fabric, it has discordant relations with foliation, etc., in the country rocks.

3.3.28.8 Geophysics

Kane and Bromery (1968) and Hodge et al. (1982) provided detailed gravity studies for Maine, and their data correlate well with the observed geology. No specific data are available for the Quaker Hill pluton, but it is located in an area of structural and metamorphic complexity and near the junction of regional gravity gradients found by Kane and Bromery (1968). The Sebago Lake batholith to the southwest and the Waldboro pluton to the southeast are two prominent gravity anomalies.

3.3.29 Redington Pluton (Plate II, Pluton 15)

3.3.29.1 Size, Shape, and Location

The Redington pluton is a batholith located in west-central Maine in the southwestern quarter of the Stratton, the northwestern quarter of the Phillips, and the northeastern quarter of the Rangely 15-min quadrangles. The intrusion has a roughly elliptical shape, with a long axis of approximately 18 km (11 mi) that strikes N. 55° E., and a short axis of about 8 km (5 mi). A small protuberance projects to the southeast. In addition, a number of small satellite granitic bodies occur in the intrusion's contact metamorphic halo. The mapped area shown in Plate II is about 136 km^2 (52 mi^2).

3.3.29.2 Relative and Absolute Age

The pluton intrudes into mainly Silurian sedimentary rocks, roof pendants of which partially cover portions of the intrusion. The sedimentary rocks include the Lower Silurian Rangeley formation, the Middle Silurian Perry Mountain formation, and the Middle to Upper Silurian Small Falls formation. This indicates a relative age for the intrusion of Late Silurian to Early Devonian. No stratigraphic indication of the minimum age of the intrusion is available, as younger Paleozoic rocks are absent. Structural interpretations have assigned the intrusion to the New Hampshire plutonic series (Moench, 1970).

Radiometric dates for the Redington pluton are lacking. Limited but inconclusive data exist for the Phillips pluton to the south, which is considered to be a cupola associated with a large intrusion of which the Redington pluton is also a part (Kane and Bromery, 1968).

3.3.29.3 Origin

The Redington pluton shows evidence of forced injection into the metasedimentary country rocks, as do its cogenetic sisters, the Phillips and Lexington plutons. Regional structures terminate against and wrap around the intrusion. Discordant relations are found on both large and small scales. In contrast to the Phillips pluton, where sedimentary rocks dip away from the intrusion, the sedimentary rocks surrounding the Redington pluton dip towards the intrusion (Moench and Zartman, 1976). Large roof pendants occur that are oriented opposite to the regional structure (Moench, 1954). Contacts exhibit intimate intercalation of the metasedimentary country rock and the igneous intrusion, with granitic tongues intruded along bedding planes. Moench and Zartman (1976) infer this pluton to be essentially sheetlike.

3.3.29.4 Petrography

Mineralogy and Textural Relations. The occurrence of large roof pendants indicates that erosion has exposed only the very top of the pluton. Mineralogy and petrology are therefore restricted to the suite of rocks at the top of the intrusion, and may not be representative of the whole body. The cogenetic relationship of the Redington and Phillips plutons suggested by Kane and Bromery (1968) is in part supported on a mineralogic basis. The two intrusions have very similar compositions but contrasting textures, reflecting different degrees of erosion into their respective magma chambers (Moench, 1954). The Redington pluton, which partially retains its erosion-resistant roof rock, has excellent outcrop exposures. In contrast to the equigranular Phillips pluton, it is coarsely porphyritic, containing abundant small inclusions of country rock. The elongate inclusions, feldspar phenocrysts, and biotite flakes have a planar alignment. The xenoliths are rounded to angular (Moench, 1954).

Alteration. Alteration in the Redington granodiorite is the same as in the Phillips stock.

3.3.29.5 Secondary Intrusions

The lack of information in the literature on secondary intrusions may indicate that they are not significant in the area.

3.3.29.6 Country Rocks

The surrounding country rock consists of sedimentary units displaying mineralogies of both a regional metamorphic grade and a hornfelsic grade, the latter from contact metamorphism. Away from the contact aureole, the Silurian metasedimentary rocks consist of (1) interbedded and intertonguing metashale, (2) feldspathic metasandstone, (3) polymictic metaconglomerate, (4) quartz

metaconglomerate of the Rangeley formation, (5) potassic metashale and cyclically interbedded metasandstone of the Perry Mountain formation, (6) black sulfidic metashales interbedded with thin quartzose metasandstone, and (7) sulfidic calc-silicate rocks of the Small Falls formation (Moench and Zartman, 1976). Adjacent to the intrusion and atop it in the roof pendants, these contact rocks have been subjected to a secondary (contact) metamorphism to hornfels grade, which caused decreases in silica, magnesium oxide, and calcium oxide, and increases in aluminum oxide and potassium oxide (Moench, 1954). Texturally, the hornfels exhibits no preferred orientation in its petrofabric.

3.3.29.7 Rock Structure

No recorded faults occur in the Redington pluton, and a systematic analysis of joint sets does not exist in the literature.

The preferred orientation of feldspar phenocrysts, xenoliths, inclusions, and biotite flakes indicates a flow texture. This interpretation is supported by the lack of stress-induced foliation in the adjacent hornfels metamorphics. The turning and twisting of the flow textures in what appears to be eddies (turbulent flow) further supports the interpretation of this foliation as a primary flow texture. The turbulent pattern precludes determining structural attitudes on this foliation (Moench, 1954).

By the nature of the primary flow texture and lack of foliation in the hornfels country rock, the pluton appears not to have been subjected to post-intrusive plastic deformation.

3.3.29.8 Geophysics

As discussed in Secs. 3.3.17 and 3.3.25, the Redington pluton is considered to be one of three cupolas of a large granitic sheet (Kane and Bromery, 1968). Gravity interpretations indicate an average thickness of 1520 m (5000 ft).

3.3.30 Rome Pluton (Plate II, Pluton 11)

3.3.30.1 Size, Shape, and Location

The Rome pluton is a roughly triangular body, about 25 km (15 mi) on each side, that underlies an area of about 461 km^2 (178 mi^2). It lies predominantly in the Norridgewock 15-min quadrangle, but extends into the Waterville, Augusta, and Farmington 15-min quadrangles. A significant part of the pluton is covered by moderately large lakes, of which Great Pond is the largest. This pluton is one of the larger bodies in a concentration of intrusives found in the western part of Maine.

3.3.30.2 Relative and Absolute Age

The Rome pluton is a member of the New Hampshire plutonic series. Dallmeyer (1979) dated the pluton at 314 m.y. and 281 m.y., by incremental-release analysis of the Ar^{40}-Ar^{39} content. Dallmeyer states that these ages are actually much younger than the emplacement of the pluton and probably reflect a later metamorphic episode. Such metamorphic reheating would have affected the argon retention of the rocks.

3.3.30.3 Origin

Furlong (1960) noted the presence of included blocks and roof pendants, indicating that stoping probably played a role in the emplacement of the Rome pluton. Foliated margins suggest probable syntectonic emplacement (Osberg, 1979).

3.3.30.4 Petrography

Mineralogy and Textural Relations. The Rome pluton consists of a medium-grained, equigranular biotite granodiorite. Major minerals are plagioclase, microcline, quartz, and biotite. Lesser amounts of hornblende and garnet occur locally in limited amounts (Furlong, 1960).

Alteration. Mafic minerals show alteration to chlorite.

3.3.30.5 Secondary Intrusions

Pegmatites in the Rome granodiorite are generally elongated and associated with fractures in the rock (Furlong, 1960; Osberg, 1979).

3.3.30.6 Country Rocks

The regional structural trend in the area of the Rome pluton is north-northeast. At the northwestern edge of the pluton, near the town of New Sharon, the pluton is in contact with calcareous metamorphosed rocks of Silurian-Devonian age, including parts of the Vassalboro and Berwick formations. Proceeding clockwise around the perimeter of the pluton, the granodiorite is in contact with a belt of undifferentiated metasedimentary rocks of Ordovician-Silurian age. Interbedded in this belt are thin beds of metamorphosed calcareous siltstones and limestones. At its northeastern tip near Larone, the pluton is in contact with an Ordovician-Silurian calcareous quartzite that contains minor amounts of sulfidic quartzite and phyllite.

From near the southernmost tip of Great Pond to past North Fairfield, the pluton is in contact with an Ordovician-Silurian metasedimentary quartz

wacke, conglomerate, siltstone, and shale. Interbedded within these units are minor amounts of limestone and marble and thin beds of slate or schist. The southern tip of the pluton is in contact with metamorphosed Silurian calcareous sandstone, siltstone, and graywacke, a section of which contains parts of the Kittery formation.

The south-to-northwest margin of the pluton is slightly more complex than the others. From the southern tip of Long Pond to a little past the town of Vienna, the country rock is Ordovician-Silurian metamorphosed quartz wacke, conglomerate, siltstone and shale, units that are identical to those found along the northeast-trending edge of the pluton near North Fairfield. Between this and another belt of the same rock type is an Ordovician-Silurian calcareous quartzite and minor sulfidic quartzite and phyllite identical to the units found near the town of Larone on the northeastern tip of the pluton. Interbedded in this rock are beds of limestone or marble and thin beds of slate or schist. Extending north from near Vienna to near New Sharon is more of the previously mentioned Silurian calcareous rocks found near Belgrade, and Ordovician-Devonian metamorphosed conglomerate, siltstone, and shale.

3.3.30.7 Rock Structure

Steeply dipping joints and horizontal sheeting are common in the Rome pluton (Furlong, 1960). The Rome granodiorite contains noticeable foliation, which is evident near the margins of the body (Osberg, 1979), suggesting syntectonic emplacement.

3.3.31 Sebago Lake Batholith (Includes Songo Granodiorite) (Plate II, Pluton 6)

3.3.31.1 Size, Shape, and Location

The irregularly shaped Sebago Lake batholith extends over roughly 2714 km^2 (1048 mi^2) in southwestern Maine, forming one of its largest intrusives, and the largest outcrop of any crystalline rock in Maine. Outcrops occur in the Bethel, Bryant Pond, Fryeburg, Norway, Poland, Lewiston, Kezar Falls, Sebago Lake, Gray, Freeport, and Portland 15-min quadrangles. A portion of the pluton extends westward into the state of New Hampshire. Much of the batholith's area is covered by numerous lakes. Exfoliation domes, some of which are over 0.8 km (0.5 mi) in diameter, are fairly common in the Bethel region.

3.3.31.2 Relative and Absolute Age

The Sebago Lake batholith is regarded as a member of the New Hampshire plutonic series. Guidotti (1965) reported various mid-Paleozoic ages for the

pluton based on work by Hurley in 1965. The values range from 252 m.y. to 340 m.y. for six K-Ar determinations on mica. Dallmeyer (1979) reported an age of 227 m.y. based on Ar^{40}-Ar^{39} analysis of biotite (Guidotti, 1965); this is most likely a reset age, as the country rocks also give the same age.

3.3.31.3 Origin

Fisher (1962) indicated that the Songo granodiorite member of the Sebago Lake batholith should not be attributed to metasomatic activity. As evidence, he cited an apparent roof pendant in this area of the pluton and the inclusion of large blocks of metasedimentary country rock. Also, a slightly domal structure is suggested by the gentle dipping of primary foliation in the granodiorite toward the margins of the body. The foliation in the granodiorite does not dip as steeply as that of the wall rock, which locally dips toward the margins of the pluton.

The numerous inclusions of metasedimentary country rock typically are slabby and oriented parallel to the foliation of the granodiorite. Although the contacts between the inclusions and the granodiorite usually are sharp, some of them are gradational. This tends to indicate that forced injection played an important role in emplacement (Fisher, 1962). Guidotti (1965) agreed that the Sebago Lake batholith was forcibly injected, and added that it may have migrated in a northeasterly direction as it ascended through the country rock.

Page (1968) suggested that the Sebago Lake batholith belongs to the New Hampshire plutonic series, the members of which are elongate bodies that together form a flattened oval, elongate to the northeast, with its ends in Connecticut and Maine. The Sebago Lake batholith, along with a large plutonic mass in east-central New Hampshire, lies roughly in the center of this oval. The plutons of the New Hampshire plutonic series generally were intruded and crystallized under stresses induced by the Acadian orogeny. The Sebago Lake batholith and others of the series tend to parallel Acadian folds and other linear features (Page, 1968).

3.3.31.4 Petrography

Mineralogy and Textural Relations. The rock of the northern extension of the Sebago Lake batholith is a gray, medium-grained, subhedral biotite granodiorite. The grain size usually is 1-4 mm in length. Major constituent minerals include quartz, plagioclase (An-28-30), microcline, orthoclase, and green to brown biotite. Modal analysis of the Songo granodiorite yields quartz, 23%; microcline, 7%; orthoclase, 2%; plagioclase (An-33 average), 51%; and biotite, 16% (Fisher, 1962). The rock is locally porphyritic, having orthoclase phenocrysts approximately 4 cm (1.6 in.) long, which are much smaller and less oriented than the phenocrysts in the Kinsman quartz monzonite.

Minor amounts of amphibole, chlorite, sphene, and sericite occur in the granodiorite. Accessory minerals include muscovite, pyroxene, magnetite, pyrite, apatite, zircon, allanite, and leucoxene (Fisher, 1962). Brown and green biotite occur in similar concentrations in the granodiorite. Fisher (1962) indicates that the green biotite tends to be deficient in magnesium and titanium.

The far eastern portion of the Sebago Lake batholith consists of medium- to fine-grained, pinkish muscovite granite. Major minerals include quartz, potassium feldspar (mostly microcline), plagioclase, muscovite, and biotite, with accessory minerals of zircon and chlorite. Modal analysis of the granite of the Woodbury area yields quartz, 30%; potassium feldspar, 35%; plagioclase (An-10), 25%; muscovite, 5%; biotite, 2%; and accessory minerals, 3% (Fisher, 1941).

Alteration. Secondary alteration of chlorite after biotite and sericite after plagioclase is present. The green biotite found in the pluton could be the result of an early stage of hydrothermal alteration (Fisher, 1962).

3.3.31.5 Secondary Intrusions

The roughly ovoid Mount Pleasant stock of Mesozoic age intrudes into the interior of the Sebago pluton at 44° 2' N. latitude and 70° 50' W. longitude.

The Songo granodiorite region is cut by numerous pegmatites, which are particularly abundant near the Songo granodiorite itself. The pegmatites range from a fraction of a meter to many kilometers (several inches to miles) across, taking the form of sills, dikes, discordant lenses, and irregularly shaped bodies. They tend to be simple quartz-feldspar-mica rocks, although less common minerals like garnet, tourmaline, beryl, lepidolite, and apatite are occasionally present (Guidotti, 1965).

3.3.31.6 Country Rocks

The Sebago Lake batholith, owing to its large size, is in contact with various rock types. The batholith as a whole tends to parallel the fold axes formed in the Acadian orogeny. At different points along its Maine perimeter, it shares boundaries with parts of the Shagg Pond, Rindgemere, Littleton, Thompson Mountain, Billings Hill, Eliot, Waterville, Noyes Mountain, Moody Brook, Patch Mountain, Vassalboro, and Berwick formations. The pluton is unconformable on a regional scale, even though it is locally conformable (Guidotti, 1965).

The Shagg Pond formation, the Concord member of the Littleton formation, and a section of the lower Rindgemere formation are gray

biotite-muscovite gneiss and schist. This rock type is in contact with the northwestern and southwestern borders of the pluton, extending into New Hampshire. Sulfide gneiss and schist, which are characteristic of parts of the Thompson Mountain and Billings Hill formations and the Wilbur Mountain member of the Littleton formation, are in contact with the Songo granodiorite at the northern end of the pluton, immediately north of the Moll Ockett fault (Doyle et al., 1967). Across the fault to the south, the Sebago Lake batholith is in contact with metamorphosed shale, siltstone, and sandstone of the Eliot, Waterville, Noyes Mountain, and Moody Brook formations. Along this northeastern margin, the Sebago Lake batholith also cuts metamorphosed limestone and calcareous siltstone of the Patch Mountain and Berry Ledge formations, in addition to unnamed members of the Eliot and Waterville formations (Doyle et al., 1967). On its southeastern border, the pluton cuts calcareous metasedimentary rock of the Vassalboro and Berwick formations. Almost adjacent to the Maine-New Hampshire border, the pluton is in contact with a granodiorite of Devonian age (Doyle et al., 1967).

3.3.31.7 Rock Structure

Five sets of joints -- four vertical and one horizontal -- are found in the Sebago Lake batholith. The horizontal set may be the result of expansion due to regional unloading (Fisher, 1962). The vertical joints that strike parallel to the regional fold structure are probably release fractures formed when the northwest-to-southeast compressional forces lessened.

The Moll Ockett fault extends eastward from the Songo granodiorite. Elongated zones of silicified material having a cataclastic texture are reported (Fisher, 1962). However, no displacement along these zones was observed.

Various phenomena in the Sebago Lake batholith have been interpreted as evidence of flow. Primary(?) foliation and slabby inclusions of the wall rock lie in a domal orientation and are particularly evident near the margins of the body. If the foliation was caused by magmatic flow, drag on now eroded roof pendants could have caused the foliation to lose its gently dipping attitude in the interior of the pluton (Fisher, 1962). Local elongated shapes in conformity with the regional structural trend indicates that foliation, in part, may be secondary.

3.3.31.8 Geophysics

On the basis of previous gravity studies (Kane and Bromery, 1968; Kane et al., 1972), Moench and Zartman (1976) depicted the pluton as a relatively thin and shallow subhorizontal sheet. This view explains many of the complex metamorphic isograds studied by Moench and Zartman in the extreme northern end of the pluton. Gravity lows near the southeast corner of the Sebago batholith (see Fig. 3.12) contrast with the general northwesterly regional trend (Hodge

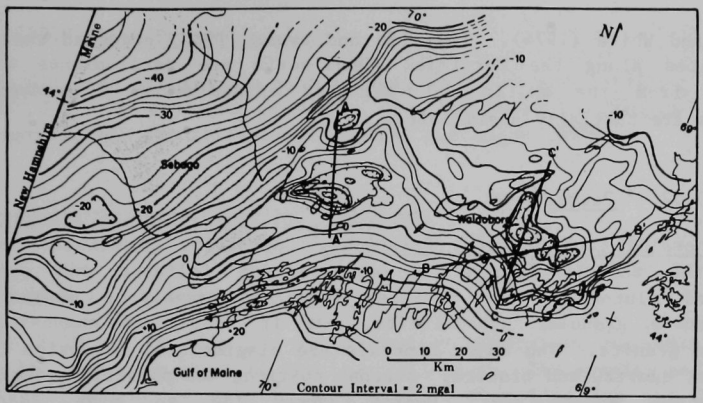

Fig. 3.12 Bouguer Anomaly Map over Waldoboro and Sebago Lake Plutons in Southwestern Maine (granitic rocks are shown in stippled pattern (Source: Hodge et al., 1982)

et al., 1982). Some calculations have indicated that the Sebago may be only 1 km (0.6 mi) thick, but somewhat different assumptions about the value of the density contrast most appropriate for the calculations suggest that the granite may instead be continuous to great depth (Hodge et al., 1982).

3.3.32 Seboeis Complex (or Schoodic Lake Granite) (Plate II, Pluton 25)

3.3.32.1 Size, Shape, and Location

The Seboeis complex (also known as the Schoodic Lake granite) is an elongated body north of the Nozumbega fault system in north-central Maine, containing both nonfoliated and foliated granitoids. The complex is about 28 km (17 mi) long and 3 to 5 km (2 to 3 mi) wide; outcrops occur over an area of about 123 km^2 (47 mi^2).

3.3.32.2 Relative and Absolute Age

Doyle et al. (1967) suggested a Devonian age for the plutonic rocks.

3.3.32.3 Origin

Foliated granites, the dominant facies in the complex, have chemical characteristics similar to, though more potassic than, the I-type granite of

Chappell and White (1974). Loiselle and Ayuso (1980) proposed that none of the granites along the Merrimack synclinorim are peraluminous enough to originate from the melting of pelitic metasediments. Metasedimentary inclusions are partially assimilated.

3.3.32.4 Petrography

Mineralogy and Textural Relations. According to Loiselle and Ayuso (1980), the Seboeis Complex is composed of foliated and subordinate nonfoliated plutons. The foliated unit is largely a medium-grained, hypidiomorphic granular granodiorite, but it ranges in composition from diorite to granite. The major minerals are plagioclase, potassium feldspar, hornblende, quartz, and biotite. Zircon, apatite, and possibly primary opaque minerals are accessory. Quartz shows low to high degrees of recrystallization. The non-foliated unit is a medium- to fine-grained, hypidiomorphic granular granite with low color indices. The major minerals are potassium feldspar, quartz, biotite, muscovite, and sphene. Zircon, allanite, and opaque minerals are accessory. A trace amount of amphibole is present. The unit contains unoriented metasedimentary inclusions.

The foliated (Mattamiscontis) granite has aplitic facies. There is a very high Rb-Sr ratio in the complex.

Alteration. Loiselle and Ayuso (1980) attribute the formation of sericite, epidote, chlorite, sphene, and opaque minerals to the hydrothermal stage.

3.3.32.5 Secondary Intrusions

There are no data on the presence of secondary intrusions in the complex.

3.3.32.6 Country Rock

The Seboeis complex is primarily surrounded by the lower Devonian Seboomook formation. Silurian quartzite and meta-graywacke is found at the southern end of Schoodic Lake. On the eastern side of the complex lies Silurian metamorphosed gray and green siltstone, sandstone, shale, graywacke, conglomerate, and metasedimentary iron-manganese deposits (Doyle et al., 1967).

A metamorphic aureole 1-2 km (0.6-1.2 mi) wide, locally irregular, is present around the entire complex (Doyle et al., 1967; Loiselle and Ayuso, 1980). The contact relations of the nonfoliated granite are sharp and discordant, whereas the foliated granitoid has locally sharp, but possibly generally gradational, contacts (Loiselle and Ayuso, 1980).

3.3.32.7 Rock Structure

The foliated granite has weak to moderate foliation trending N. 30° E. to N. 45° E. It contains abundant metasedimentary inclusions that are oriented parallel to the foliation (Loiselle and Ayuso, 1980).

3.3.33 Sedgwick Granite (Plate II, Pluton 40)

3.3.33.1 Size, Shape, and Location

The Sedgwick granite is a subcircular pluton on the east side of Penobscot Bay, directly south of the Lucerne pluton. The granite underlies an area of approximately 95 km^2 (37 mi^2) between Blue Hill Bay and Eggemoggin Reach, in the Blue Hill quadrangle. Its minimum and maximum map dimensions are about 10 and 15 km (6 and 9 mi), respectively.

3.3.33.2 Relative and Absolute Age

Brookins (1967) reported a Rb-Sr whole-rock age of 395 ± 15 m.y., which agrees with the Devonian age assigned to the Bays-of-Maine region by Faul et al. (1963) and Doyle et al. (1967)

3.3.33.3 Origin

Sedgwick granite was emplaced as part of the late Acadian magmatism in Maine.

3.3.33.4 Petrography

Mineralogy and Textural Relations. Sedgwick is a light-gray to buff, medium- to coarse-grained, equigranular biotite granite. Major minerals are potassium feldspar (microcline and orthoclase), soda-lime feldspar (oligoclase), dark smoky quartz, and muscovite (Rand, 1958). Zircon and magnetite are accessory minerals. Feldspars exhibit a bluish-white color and are up to 2.5 cm (1 in.) long.

Alteration. Forsyth (1953) attributed the presence of tourmaline, coarse muscovite, and sericitized albite in the Ellsworth schist, at its contacts with the granite, to hydrothermal activity soon after the intrusion.

3.3.33.5 Secondary Intrusions

No significant intrusions are reported.

3.3.33.6 Country Rocks

On the north and southeast, the Sedgwick granite is bounded by the Ellsworth schist in the Blue Hill and Blue Hill Neck areas. The Sedgwick grades into quartz monzonite, and may have minor amounts of molybdenite at its contacts with Ellsworth. The Castine volcanics are in contact with the granite in the southwest and west; to the northwest lies the Penobscot granite, surrounded by gabbro and diorite.

3.3.33.7 Rock Structure

Young (1962) reported white quartz veins 2.5-18 cm (1-6 in.) wide filling a fracture system at the Leigh Limeburner ("J.S. Douglas") mineral prospect in the Sedgwick granite. Jointing is not well developed in the granite; a few joints strike from N. 47° W. to N. 35° E., with dips of 64° to 82° (Young, 1962).

3.3.34 Seven Ponds Pluton (Plate II, Pluton 20)

3.3.34.1 Size, Shape, and Location

The Maine portion of the Seven Ponds pluton covers roughly 115 km^2 (44 mi^2) in northwestern Maine. It is about 19 km (12 mi) long and 6-8 km (4-5 mi) wide, resembling a rounded body bounded by the international boundary to the north and joined to a smaller rounded body to the south. The northern mass has a diameter of about 8 km (5 mi), which is about 1.6 km (1 mi) more than the diameter of the southern portion of the pluton. Doyle et al. (1967) did not show any continuity between the two lobes. Westerman (1980) describes them as separate plutons: the larger, northern body is called the Chain of Ponds pluton, while the smaller body to the south is called the Big Island Pond pluton. The name used in this report, "Seven Ponds Pluton," is taken from Harwood (1973). The pluton is dotted by many small to moderately sized ponds and lies in the Arnold, Cupsuptic, Chain Lakes, and Kennebago Lake 15-min quadrangles. It is continuous with the Spider Lake granite of Quebec.

3.3.34.2 Relative and Absolute Age

On the basis of its similarity to the two-mica granitic rocks of the New Hampshire plutonic series, Harwood (1973) assigned a Middle to Late Devonian age to the Seven Ponds pluton. Because the rocks are relatively

unaltered and undeformed, they postdate the regional folding and metamorphism of the Early Devonian, thereby suggesting a Late Devonian age.

3.3.34.3 Origin

The Seven Ponds pluton contains narrow roof pendants in its outer quartz monzonite phase, indicating that stoping played a role in its emplacement. The pluton could belong to the Late Devonian plutonic series of Page (1968), or it could be an early posttectonic phase of the New Hampshire plutonic series.

3.3.34.4 Petrography

Mineralogy and Textural Relations. The Seven Ponds pluton consists of an outer belt or zone of biotite-muscovite quartz monzonite that grades inward to a hornblende-biotite granodiorite. The quartz monzonite is light gray, medium to coarse grained, and equigranular to subporphyritic. The quartz monzonite generally contains biotite and pink alkali feldspar near its contacts with metasedimentary country rocks. Major minerals include quartz, microcline, plagioclase, biotite, and muscovite. Accessory minerals include garnet, hematite, hornblende, clinozoisite, chlorite, zircon, apatite, magnetite, and myrmekite. The quartz grains usually contain rutile needles.

Hornblende in the quartz monzonite gradually increases toward the center of the pluton, where the predominant rock type is a biotite-hornblende granodiorite. A transition zone several hundred feet wide exists between the quartz monzonite and granodiorite. The rock type is intermediate, usually containing about 2% hornblende. The many ponds and swamps covering the granodiorite phase result from its being less resistant to weathering.

The granodiorite is gray, fine to medium grained, and locally porphyritic, containing phenocrysts of hornblende and plagioclase that range in size up to 1.3 cm (0.5 in.). The phenocrysts tend to be oriented parallel to the primary foliation of parallel biotite grains. Major minerals include quartz, microcline, plagioclase, biotite, and hornblende. Lesser amounts of clinozoisite, apatite, sphene, and myrmekite are found. Accessory minerals include chlorite, zircon, and magnetite. Rutile is sometimes present as needles in the quartz grains (Harwood, 1973).

Alteration. Biotite is commonly replaced by chlorite and ilmenite. Chlorite and epidote after hornblende are found. Sericite is observed locally (Harwood, 1973). Much alteration occurs along the joints.

3.3.34.5 Secondary Intrusions

Pegmatites and one lamprophyre dike occur in the area. Serpentinite and serpentinized ultramafics are found in the pluton, as are basalt, felsic porphyry, porphyritic basalt, medium-grained granite, and metamorphosed basaltic dikes.

3.3.34.6 Country Rocks

The contact of the pluton dips steeply away from the mass. Harwood (1973) shows the pluton to be a strongly discordant, stocklike intrusion.

The Seven Ponds pluton intrudes into the Chain Lakes massif. The southern edge of the lower lobe of the pluton also is in contact with metamorphosed felsic to mafic volcanic rocks and interbedded metasediments of the Ammonoosuc volcanics, unnamed greenstone, and metamorphosed dark-gray and black shale, chert and sandstone, all of Ordovician age. Included are parts of the Partridge and Dixville formations. The Seven Ponds outcrop is exposed across part of the area that is assumed to be underlain by the Chain Lakes pluton (see Sec. 3.1.1).

3.3.34.7 Rock Structure

Westerman (1980) compiled a lithologic and brittle-fracture map of the Seven Ponds pluton and vicinity. Harwood (1973) also notes the presence of joints in the body. The joints in the granodiorite phase form rectangular or rhombohedral networks. The joints form distinct raised areas because their greater quartz content makes them more resistant to weathering (Harwood, 1973).

Primary foliation occurs in part of the northern portion of the northern lobe of the Seven Ponds pluton, associated with apparent roof pendants in the pluton. The granodiorite phase shows more primary foliation than the quartz monzonite, because of the parallelism of its biotite flakes. This foliation is vertical and conforms to the contact with the quartz monzonite through the transition zone. Hornblende phenocrysts locally define a lineation that plunges 35° N.

The Seven Ponds pluton is bordered to the southeast by the northeast-trending Boundary Mountain anticline. On the northwestern side, it is bounded by the Frontiac syncline. The regional structural trend in this area is consistently northeast to southwest. No folding is present in the pluton.

3.3.35 Sugarloaf Pluton (Plate II, Pluton 16)

3.3.35.1 Size, Shape, and Location

The Sugarloaf pluton (called Sugarloaf gabbroic massif by Boone [1973]), is an irregular kidney-shaped crystalline rock body lying in the Stratton, the Little Bigelow Mountain, and the Kingfield quadrangles (Boone et al., 1970). Outcrops occur over an area of about 113 km^2 (41 mi^2). The pluton is almost completely surrounded by four neighboring plutons: the Redington, which it almost contacts, and the Lexington, Pierce Pond, and Flagstaff bodies, all within 5-10 km (3-6 mi) of the Sugarloaf.

3.3.35.2 Relative and Absolute Age

The Sugarloaf pluton is inferred to be of Acadian age, as it truncates the folds in the Lower Devonian Seboomook formation. Zartman et al. (1970) reported a K-Ar age of 397 ± 12 m.y. on biotite from a mica granite dike in the pluton.

3.3.35.3 Origin

The pluton is a magmatic intrusion that was forcibly emplaced at a shallow depth, probably a a sheet. Xenoliths of high-grade metamorphic rocks represent roof pendants in the pluton.

3.3.35.4 Petrography

Mineralogy and Textural Relations. The pluton is composed largely of gabbroic mafic rocks that are massive, layered, and have well-preserved cumulate textures in parts of the body. They are medium to coarse grained, with subordinate pegmatitic textural phases.

Mineralogic variants are present, but all the rock are generally composed of plagioclase and augite-hypersthene. Differences in olivine and hornblende proportions define different rock types. Secondary alteration minerals are abundant. Tourmaline, ilmenite, and pyrrhotite are the important accessory minerals.

Alteration. Retrogressive modification and alteration are pervasive in the pluton. Chlorite, epidote, serpentine, and actinolite are retrograde products of mafic minerals and the plagioclase. Biotite may be an alteration mineral.

3.3.35.5 Secondary Intrusions

No significant secondary intrusions are described.

3.3.35.6 Country Rocks

The pluton is surrounded by a contact aureole about 1.6 km (1 mi) wide.

3.3.35.7 Rock Structure

Fractures are inferred from sheared zones trending to the northeast. Mineral layering present is primarily due to gravity setting. No folding has occurred since the intrusion.

3.3.35.8 Geophysics

Bromery et al. (1963) interpreted the lack of dominant aeromagnetic anomalies as indicating that the pluton is a thin intrusive sheet.

3.3.36 Tunk Lake Pluton (Plate II, Pluton 37)

3.3.36.1 Size, Shape, and Location

The Tunk Lake pluton is in southeastern Maine in Hancock and Washington counties. It is about 6.5 km (4 mi) inland, northeast of Frenchman's Bay. Its exposed and inferred surface area is about 179 km^2 (69 mi^2); its surface shape is that of a slightly flattened circle with the long axis running east-west (see Fig. 3.13). It is almost completely surrounded by rocks of the Bays-of- Maine igneous complex, which includes gabbros and granites.

3.3.36.2 Relative and Absolute Age

Although the absolute age of the Tunk Lake pluton is not available, adjacent intrusions truncated by the Tunk Lake pluton have been dated. The North Sullivan quartz monzonite, west of the Tunk Lake pluton, has a K-Ar biotite age of 369 m.y., and the granite north of the Tunk Lake pluton has a K-Ar biotite age of 367 m.y. (Faul et al., 1963). Therefore, the Tunk Lake pluton has a minimum relative age younger than Middle Devonian.

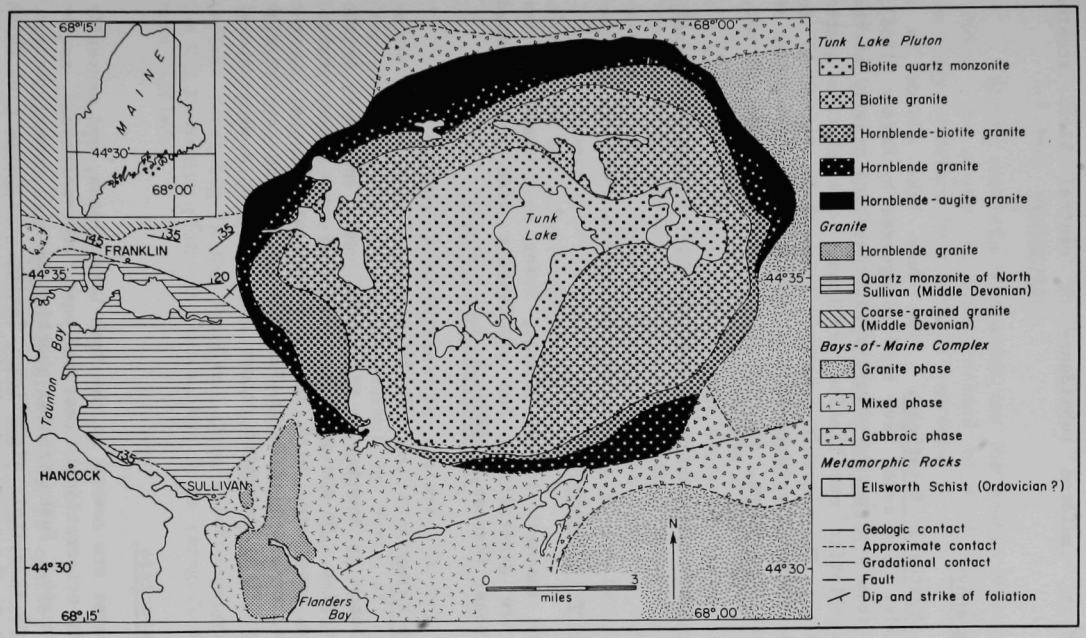

Fig. 3.13 Geologic Map of Tunk Lake Pluton (Source: Karner, 1968)

3.3.36.3 Origin

The circular shape, zoned character, and marginal xenoliths of the pluton indicate magma emplacement by piecemeal stoping along zones of ring fractures, accompanied by cauldron subsidence or central block stoping (Karner, 1968).

3.3.36.4 Petrography

Mineralogy and Textural Relations. The Tunk Lake pluton is a concentrically zoned body. Six zones have been recognized by Karner (1968), ranging in composition from a magnetite-aegerine-microperthite granite at the rim, through hornblende-biotite and biotite granites, to a biotite quartz monzonite at the core. The perthite ranges from Ab-21 to Ab-45 in composition. The nonperthitic plagioclase is oligoclase. Trace amounts of magnetite, hematite, zircon, sphene, and allanite also are present (Karner, 1968).

Alteration. Secondary alteration of the Tunk Lake pluton occurs only to a minor extent.

3.3.36.5 Secondary Intrusions

Aplite dikes are abundant in the pluton, being most common in the central area. They are generally a few centimeters wide and oriented northwest or north-northeast (Karner, 1968).

3.3.36.6 Country Rocks

The Tunk Lake pluton is bounded on the north, east, and south by the Bays-of-Maine igneous complex. On the north and northwest is a coarse-grained Middle Devonian granite associated with the Lead Mountain pluton; on the southwest is the quartz monzonite of North Sullivan. A limited contact with the possibly Ordovician Ellsworth schist occurs on the pluton's west-central side (Karner, 1968).

3.3.36.7 Rock Structure

Joints and fractures are common in the Tunk Lake pluton, and generally have one of the following orientations: east and west, west-northwest, north-northeast, or east-northeast. Subhorizontal sheeting is also common (Karner, 1968). There are several relatively significant faults, the largest of which is a fault zone adjacent to the south side of the pluton.

Possible flow structures include scattered occurrences of oriented mafic inclusions and perthite crystals (Karner, 1968). There are no reported folds in or directly related to the Tunk Lake pluton.

3.3.36.8 Geophysics

Abrupt magnetic boundaries for the pluton suggest steep contacts (Wing, 1958). The pluton is characterized by a circular gravity low whose outer boundary is marked by a -5-mgal contour.

3.3.37 Umbagog Granodiorite (Includes Adamstown Pluton) (Plate II, Pluton 14)

3.3.37.1 Size, Shape, and Location

The Maine portion of the Umbagog granodiorite underlies about 235 km^2 (91 mi^2). A smaller portion of the intrusive extends westward into the state of New Hampshire. The pluton lies in the Oquossoc, Errol, Milan, and Old Speck Mountain 15-min quadrangles. A small arm of the Umbagog granodiorite extends eastward into the Mooselookmeguntic pluton (pluton 13). The western edge of the Maine portion of the Umbagog pluton is bordered by Umbagog Lake. In the southern part of the Umbagog pluton, the land tends to be marshy and dotted with many small ponds.

3.3.37.2 Relative and Absolute Age

The Umbagog granodiorite is assigned to the Highlandcroft plutonic series (Late Ordovician) by Hussey and Pankiwskyj (1980), but to the New Hampshire plutonic series (Devonian) by Doyle et al. (1967). Green and Guidotti (1968) described it as belonging to the Highlandcroft plutonic series, but noted the possibility that it was emplaced during the Devonian Acadian orogeny. Milton (1961) also noted the possibility that the pluton belongs to the New Hampshire plutonic series. Green (1960) obtained a K-Ar age of 323 m.y. for the biotite of the Umbagog pluton, but this age may reflect resetting in the Permian. This report assigns the Umbagog pluton to the New Hampshire plutonic series, but it is recognized that the issue has not been settled.

3.3.37.3 Origin

Where the Umbagog granodiorite is in contact with older rocks, the contact is smooth and sharp; where the pluton is intruded into by the younger Mooselookmeguntic pluton, the contact is a zone of intrusive breccias (Milton, 1961; Doyle et al., 1967). Field and aeromagnetic evidence indicates that the contact is steep, possibly dipping under the intrusive. Dikes of the Umbagog

granodiorite in the country rock and inclusions of older metasedimentary rocks in the pluton are rare. Milton (1961) notes the occurrence of a section of flat-lying strata consisting of highly metamorphosed sedimentary rocks that are surrounded by the granodiorite of the Umbagog pluton. He infers these strata to be part of the roof of the pluton. To summarize, the pluton is thought to be a cylindrical plug with a flat-lying roof. Concordance of the pluton with the country rock indicates forceful syntectonic injection as a possible mechanism of emplacement (Milton, 1961).

3.3.37.4 Petrography

Mineralogy and Textural Relations. The Umbagog intrusive is composed predominantly of a medium-grained, dark granodiorite that is massive to weakly foliated. The darker minerals clump together to form conspicuous clots. Major minerals include plagioclase, quartz, potassium feldspar, hornblende, and biotite. Lesser amounts of augite also occur. Accessory minerals include epidote, allanite, sphene, apatite, and opaque minerals.

Some of the hornblende may be metamorphically reworked hypersthene and plagioclase. One locality near the contact with older rocks contains biotite, augite, and hypersthene but no hornblende. Milton (1961) believed that the intrusive rock was originally a norite that was subsequently metamorphosed. Some of the biotite in the central, unaltered portion of the pluton may have a metamorphic origin.

Alteration. There is evidence of metamorphic reworking of hypersthene and plagioclase to hornblende, and of hornblende to biotite.

3.3.37.5 Secondary Intrusions

No pegmatites are associated with the Umbagog granodiorite (Milton, 1961). The Mooselookmegmuntic pluton is the major secondary intrusion into the Umbagog.

3.3.37.6 Country Rocks

The Umbagog pluton intrudes into the chloritic phyllite and slate of the Albee formation along its northern contact. Variable amounts of laminated quartzite are interbedded with these rocks. To the south, the pluton intrudes the metamorphosed felsic to mafic volcanic rocks of the Ammonoosuc volcanics. These metavolcanics are interbedded with other metasedimentary rocks.

3.3.37.7 Rock Structure

The Umbagog granodiorite generally is massive to weakly foliated. Near its contact with the Mooselookmeguntic pluton, strong foliation and lineation are developed.

3.3.37.8 Geophysics

An aeromagnetic survey over the Umbagog intrusive indicates that the contact of the pluton with the country rock is sharp and steep and that it may dip inward under the granodiorite.

3.3.38 Waldoboro Pluton (Plate II, Pluton 8)

3.3.38.1 Size, Shape, and Location

The Waldoboro pluton of south-central Maine lies north of Muscongus Bay on the Atlantic coast. The pluton lies predominantly in the Waldoboro 15-min quadrangle, but it also extends north into the Liberty, west into the Wiscasset, and south into the Monhegan 15-min quadrangles (Doyle et al., 1967). Outcrops occur over roughly 376 km^2 (145 mi^2), in an area about 35 km (22 mi) long and no more than 15 km (9 mi) wide. The shape of the pluton in map view is highly irregular. Musconus Bay covers part of the southern portion of the Waldoboro pluton.

3.3.38.2 Relative and Absolute Age

Zartman et al. (1970) published a K-Ar age of 295 ± 9 m.y. for a granite specimen from near the town of Waldoboro, but this young age probably reflects thermal reequilibration during the Permian. The Waldoboro pluton belongs to the New Hampshire plutonic series and has a Devonian stratigraphic age (Doyle et al., 1967).

3.3.38.3 Origin

No conclusions concerning the emplacement of the Waldoboro pluton can be drawn from the available information; it is a composite whose main body is a two-mica granite, along with hornblende granite (Osberg and Guidotti, 1974; Newberg, 1975). The pluton was probably forcibly emplaced, as indicated by localized deformation at the margins (Newburg, 1975).

3.3.38.4 Petrography

Mineralogy and Textural Relations. The two-mica granite of the Waldoboro pluton is light to medium gray, unevenly textured, and fine to medium grained. Major minerals include orthoclase, plagioclase, quartz, biotite, and muscovite.

Alteration. Alteration is assumed to be minimal, as it is not much reported.

3.3.38.5 Secondary Intrusions

Pegmatites are present in the pluton.

3.3.38.6 Country Rocks

The extreme northwestern margin of the pluton is in contact with a thinly interbedded quartzite and phyllite. Proceeding in a southeasterly direction across the regional structural trend, the pluton is in contact with tuffaceous metasandstone and metavolcanic rock, which are predominantly felsic. Also present are subordinate amphibolite, calc-silicate gneiss, and biotite-sillimanite gneiss. The pluton intrudes the calcareous metasedimentary rocks. On its northeastern side, the Waldoboro pluton cuts a bed of metamorphosed shale, siltstone, and minor sandstone. The country rock to the southeast and southwest are referred to as undifferentiated metasediments.

3.3.38.7 Rock Structure

Although fractures occur in the pluton, no information concerning directional trends was available.

3.3.38.8 Geophysics

The Waldboro pluton is marked by a small negative Bouguer anomaly (Hodge et al., 1982). The residual gravity map correlates well with the geology (see Fig. 3.14). Gravity profiles show that granite extends beneath the metamorphic rocks. It is thickest at the center and tapers to the west.

3.3.39 Webhannet Pluton (Plate II, Pluton 2)

3.3.39.1 Size, Shape, and Location

The Webhannet pluton is structurally and mineralogically more complex than the nearby Biddeford pluton. It consists of a rock body formed from

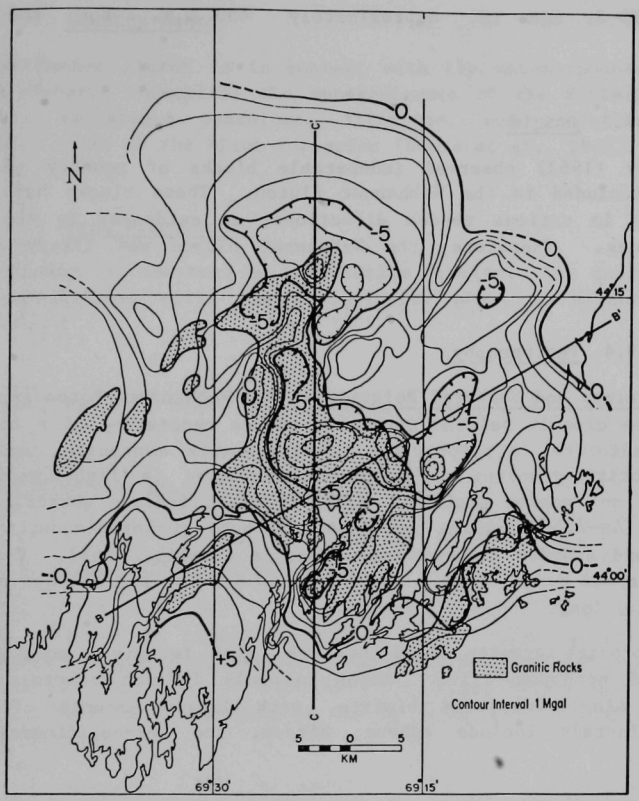

Fig. 3.14 Residual Gravity Anomaly Map of Waldoboro Pluton (Source: Hodge et al., 1982)

three closely related intrusive phases. The Webhannet pluton, like the Lyman pluton to the north, is aligned in a northeasterly direction. It is more than 26 km (16 mi) long, and at its widest point is about 8 km (5 mi) wide. The Webhannet pluton lies predominantly in the Kennebunk 15-min quadrangle, and extends slightly in a southwesterly direction into the York Harbor and the Dover East 7.5-min quadrangles. The Webhannet pluton crops out in an area of about 128 km^2 (50 mi^2).

3.3.39.2 Relative and Absolute Age

Although preliminary findings suggested that the Webhannet pluton predated the New Hampshire plutonic series, Gaudette et al. (1975) reported

whole-rock Rb-Sr ages of approximately 400 m.y. for the Webhannet pluton.

3.3.39.3 Origin

Hussey (1962) observed innumerable blocks of country rock (Kittery formation) included in the Webhannet pluton. These blocks have apparently been rotated in various random directions, as evidenced by their observed bedding planes. Therefore, the Webhannet pluton was likely emplaced by stoping.

3.3.39.4 Petrography

Mineralogy and Textural Relations. The Webhannet pluton is a composite mass of three closely related intrusive phases consisting of a biotite-hornblende granodiorite, a porphyritic biotite-quartz monzonite, and a nonporphyritic biotite granite. The quartz monzonite is light gray, strongly porphyritic, and coarse grained. Major minerals include quartz, microcline, plagioclase (An-25), biotite, and lesser amounts of hornblende, with sphene, magnetite, and apatite as accessory minerals (Hussey, 1962). Zoned plagioclase phenocrysts range up to 5 cm (2 in.) in length, but average about 1-2 cm (0.5-0.75 in.) long.

The biotite granite phase is pink, even textured, medium to coarse grained, and nonporphyritic. Major minerals include quartz, microcline, plagioclase (An-15-20), and biotite, with lesser amounts of muscovite. Accessory minerals include sphene, zircon, and opaque minerals (Hussey, 1962).

The granodiorite phase of the Webhannet pluton consists of several bodies ranging in composition from basic quartz monzonite through quartz diorite. The average composition of these bodies is that of a granodiorite, and they occur in the previously described biotite granite and quartz monzonite phases. The rocks are typically medium gray and fine grained, with plagioclase phenocrysts. Quartz, plagioclase (An-35-40), microcline, biotite, and hornblende are present, with accessory minerals being sphene, apatite, and unnamed opaque minerals (Hussey, 1962).

Alteration. The lack of information in the literature on the presence of secondary minerals is assumed to indicate that the rocks are not significantly altered.

3.3.39.5 Secondary Intrusions

No secondary intrusions into this body are mentioned in the literature.

3.3.39.6 Country Rocks

The Webhannet pluton is in contact with the metamorphosed calcareous rocks of the Berwick formation; the metasediments of the Kittery formation, which contain calcareous sandstone, siltstone, and graywacke; and the metasedimentary rocks of the Eliot formation (Doyle et al., 1967).

3.3.39.7 Rock Structure

No foliation is observed in the biotite granite. The quartz monzonite shows local foliation, while the granodiorite phase is slightly foliated (Hussey, 1962).

3.4 ALLEGHENIAN AND POST-ALLEGHENIAN

3.4.1 Agamenticus Complex (Plate II, Pluton 1)

3.4.1.1 Size, Shape, and Location

Hussey (1961) described the Agamenticus complex as a strongly discordant body composed of alkalic intrusive rocks. The complex is roughly circular, having a diameter of approximately 10 km (6.5 mi), with an area of a little over 78 km^2 (30 mi^2). It lies almost entirely within the northwestern quarter of the York 15-min quadrangle and is assigned to the White Mountain series (Hussey, 1961).

3.4.1.2 Relative and Absolute Age

Foland et al. (1971) established a K-Ar biotite age of 216 ± 4 m.y. for the Agamenticus complex. Christopher (1969) arrived at a fission-track age of 172 m.y. Foland and Faul (1977) reported a K-Ar age of 228 ± 5 m.y. The complex intrudes into the Webhannet pluton, clearly establishing it as the younger body.

3.4.1.3 Origin

Stoping and passive intrusion is the mode of emplacement inferred for the Agamenticus complex, as for the rest of the plutons of the White Mountain magma series.

3.4.1.4 Petrography

Mineralogy and Textural Relations. The Agamenticus complex is composed of four major rock types: an alkalic granite, an alkalic syenite, a contaminated alkalic granite, and a porphyritic biotite granite (Hussey, 1962). The alkalic granite is exposed in a ring-shaped mass, which crops out on all sides of the pluton except on the east and along York beach. It is usually light buff to gray in color, and is medium grained, uniform in texture, and nonporphyritic (Hussey, 1962). Hussey (1962) described the alkalic syenite as a medium- to dark-greenish-gray rock that is medium grained, evenly textured, and nonporphyritic. The syenite contains such sodium-rich minerals as arfvedsonite, riebeckite, aegirine, aegirine-augite, and sometimes hastingsite. Some quartz is present, and accessory minerals include fayalite and apatite. The feldspar ranges from microperthite to antiperthite (Hussey, 1962).

The contaminated alkalic granite varies considerably in color, composition, and texture. It occupies an irregular ring-shaped zone, which is usually located between the alkalic granite and the alkalic syenite. The contaminated granite probably resulted from partial assimilation during emplacement of the older alkalic syenite by the younger alkalic granite.

A stock of biotite granite, approximately 5 km (3 mi) long and 4 km (2.5 mi) wide, occupies the central part of the Agamenticus complex. This granite is compositionally similar to the general Conway-type granite of Billings (1956), being composed of light-pink, fine-grained, porphyritic biotite granite. Major minerals present are orthoclase, plagioclase (An-15-20), quartz with undulatory extinction, and biotite. Accessory minerals include zircon, apatite, and biotite (Hussey, 1962).

3.4.1.5 Secondary Intrusions

The alkalic syenite is cut by numerous dikes and stringers, which are composed of fine-grained, buff-colored aplitic material and a coarse-textured alkalic pegmatite having a greenish color (Hussey, 1962).

3.4.1.6 Country Rocks

The Agamenticus complex contacts the Webhannet pluton on its northwestern side and the calcareous metasediments of the Kittery formation for the rest of its periphery. The complex is surrounded by a contact-metamorphic aureole (Doyle et al., 1967).

3.4.1.7 Rock Structure

No data are available.

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4 CRYSTALLINE ROCKS OF NEW HAMPSHIRE

4.1 GRENVILLE AND AVALONIAN

4.1.1 Massabesic Gneiss (Plate III, Pluton 20)

4.1.1.1 Size, Shape, and Location

The Massabesic gneiss is an elongated northeast-trending crystalline rock body that extends for 120 km (75 mi) across New Hampshire, from Brookline and Mason near the Massachusetts border to the vicinity of Rochester near the Maine border. The mapped area of the pluton as shown on Plate III is about 470 km^2 (181 mi^2).

4.1.1.2 Relative and Absolute Age

The Massabesic gneiss, first described by Sriramadas (1966), is a complexly folded migmatite consisting of dark paragneiss crosscut by light-colored orthogneiss. Sriramadas (1966) and Greene (1970) interpreted the Massabesic as the high-grade metamorphic equivalent of the Berwick formation (see Fig. 4.1), thereby assigning it a Devonian age. Aleinikoff et al. (1979) published an age of 646 m.y. (minimum) for the Massabesic (both paragneiss and orthogneiss). The Massabesic also contains a 475-m.y.-old intrusive orthogneiss unit (Besancon et al., 1977).

4.1.1.3 Origin

The paragneiss unit of the Massabesic gneiss contains amphibolite, schists, calc-silicate, and quartzite layers, and may represent metamorphosed and granitized volcaniclastic rocks of the Avalon chain (Rankin, 1975; Aleinikoff et al., 1979). A pink microcline gneiss unit consists of metamorphosed intrusive sheets.

4.1.1.4 Petrography

Mineralogy and Textural Relations. The Massabesic is lithologically somewhat complex. Sriramadas (1966) described the gneiss unit as composed of microcline gneiss, 70%; white oligoclase gneiss, 25%; and amphibolite, 5%. The pink microcline (cf. the microcline granite unit of Freedman, 1950) is massive to well-foliated, medium- to coarse-grained rock composed of microcline, 25-44%; quartz, 20-25%; oligoclase, about 10%; biotite, 15-30%; and muscovite, 1-5%. Magnetite is conspicuous and may account for up to 5% of the rock. Sillimanite is also present (up to 10%), and is common along slip

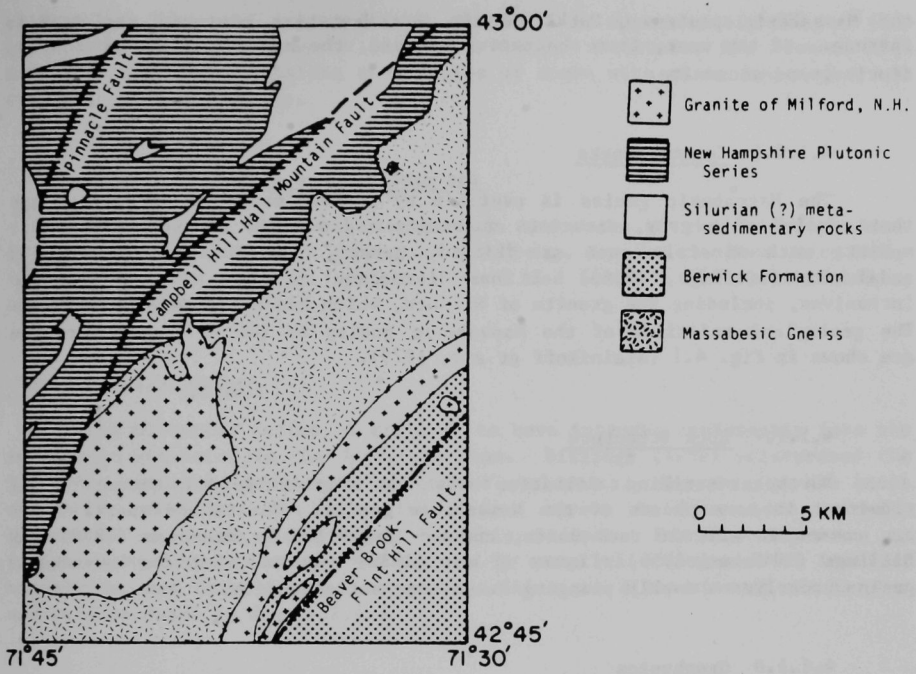


Fig. 4.1 Geologic Relations of Massabesic Gneiss in the Milford 15-min Quadrangle, New Hampshire (Source: Aleinikoff, 1978)

planes. Oligoclase gneiss (paragneiss) occurs in both porphyritic and nonporphyritic forms, and has foliation and banding similar to the pink gneiss. Both the gneisses are differentiated on the basis of absence or presence of pink microcline. White gneiss is composed of oligoclase, 20-40%; quartz, 20-40%; biotite, 20-35%; and muscovite, 0-10%. Almandine and other garnet may make up to 10% of the rock. Fine-grained amphibolite occurs in bands, and less commonly in lenses. Major minerals are hornblende, 25-65%; diopside, up to 15%; quartz, 15-25%; and sphene, about 2%.

Alteration. The feldspars are sericitized, whereas biotite shows a change to chlorite.

4.1.1.5 Secondary Intrusions

Pegmatites ranging from small dikes to bodies 140 m (450 ft) long and 15 m (50 ft) wide are abundant. The Pawtuckaway complex of the White Mountain series, a small ring structure about 1.5 km (1 mi) in diameter, intrudes into

the Massabesic pluton. Rocks of the New Hampshire plutonic series also intrude: to the west, from the center outward, the rock types are monzonite, diorite, and monzonite.

4.1.1.6 Country Rocks

The Massabesic gneiss is overlain by metasediments of Ordovician age that consist of a gray, micaceous quartzite and a gray, coarse-grained mica schist, with minerals such as biotite, garnet, sillimanite, and locally andalusite (Billings, 1956). It has discordant contacts with the younger intrusives, including the granite of Milford, which has a U-Pb age of 275 m.y. The geological relations of the Massabesic gneiss in the Milford quadrangle are shown in Fig. 4.1 (Aleinikoff et al., 1979).

4.1.1.7 Rock Structure

Northeast-striking foliation with a complex folded structure is prominent in some phases of the Massabesic gneiss. Joints and quartz veins are common in all the rock units, and are described by Freedman (1950) and Billings (1942 and 1956). Lyons et al. (1982) reported that the Massabesic gneiss underlies a doubly plunging Massabesic anticlinorium.

4.1.1.8 Geophysics

Detailed gravity surveys of the Massabesic gneiss suggest a platelike shape for the complex (Anderson, 1978).

4.2 TACONIC

4.2.1 Ayer Intrusive Complex (Plate III, Pluton 23)

See Sec. 7.2.2 and Plate VI, pluton 12. Only about 21 km^2 (8 mi^2) of this complex is in New Hampshire; the largest part of its area lies in Massachusetts, and it is discussed with the plutons of that state.

4.2.2 Jefferson Dome (Plate III, Pluton 10)

4.2.2.1 Size, Shape, and Location

The Jefferson dome is part of the north-south trending en echelon pattern of domes in the west-central part of the state. It is about 64 km (40 mi) long and up to 20 km (12 mi) wide, and extends from Franconia to northeast

of Berlin. The total mapped area in Plate III is approximately 674 km^2 (260 mi^2). The Jefferson dome is part of a larger structure, the Bronson Hill anticlinorium, which consists of a series of domes with cores composed of the Oliverian plutonic series.

4.2.2.2 Relative and Absolute Age

Various ages have been assigned to the rocks of the Jefferson dome. Zartman and Leo (1981) have determined a 455-m.y. age for the dome using U-Th-Pb isotopes. Using Rb-Sr isotopes, Foland and Loiselle (1981) published an age of 441 ± 5 m.y.

4.2.2.3 Origin

The Oliverian series is believed to have intruded conformably into the Ammonoosuc volcanics in the Jefferson dome. Billings (1956) interpreted the Oliverian series to represent "granitized" volcanics. However, on the basis of field relations, the Oliverian magmas appear to have been intruded separately as large, lens-shaped magmatic bodies. Forceful intrusion is indicated by flow structures now defined by parallel arrangement of biotite, hornblende, and large feldspar crystals. (Also see discussion of the Mascoma and Lebanon domes in Sec. 4.2.4.)

4.2.2.4 Petrography

Mineralogy and Textural Relations. The rocks of the Oliverian series show considerable range in composition and texture. Major rock types and textures include hornblende quartz monzonite, fine-grained gray quartz monzonite, coarse syenite, coarse granite, biotite quartz monzonite, and porphyritic biotite quartz monzonite. Most of the facies are medium or coarse grained, granoblastic, and foliated. Compositional variations are expressed in large variations in the modal mineralogy: quartz, a trace to 40%; potassium feldspar (usually microcline), 17-76%; plagioclase, 10-50%; biotite, a trace to 10%; hornblende, 0-18%; and magnetite, 0-10%; with local accessory amounts of muscovite, chlorite, sphene, apatite, zircon, allanite, epidote, pyroxene, garnet, rutile, tourmaline, andalusite, and sillimanite. Plagioclase composition varies from An-13 to An-42. Grain size varies from 0.1 to 12 mm (0.003-0.47 in). Pegmatite and aplite are associated with some of the rock units (Chapman et al., 1944).

Alteration. Chlorite is the principal alteration mineral. Metasomatic changes are present at the contacts with the country rocks. Devonian meta morphism affected the rocks of the area (Osberg, 1978).

4.2.2.5 Secondary Intrusions

There is no information in the literature concerning secondary intrusions, which suggests that they may be absent or minimal.

4.2.2.6 Country Rocks

In the depressions between domes are younger rocks. The upper contact of the Oliverian series is with the Ammonoosuc volcanics, which reach a thickness of several thousand meters on the south flank of the Jefferson dome (Billings, 1956). The Ammonoosuc volcanics consist mainly of potassium-feldspar-poor quartz-plagioclase-biotite gneiss, with small amounts of amphibolite, quartz-muscovite-pyrite schist, and lime-silicate granulite. The biotite gneiss is gray and fine grained (Chapman et al., 1944). Before they were metamorphosed to biotite gneisses, the Ammonoosuc volcanics were probably pyroclastics ranging in composition from rhyolite to dacite (Chapman et al., 1944; Billings et al., 1946).

4.2.2.7 Rock Structure

The southeastern flank of the Jefferson dome is broken by the Pine Mountain fault, which runs 21 km (13 mi) northeast and dips steeply. The southeastern side is downthrown (Chapman et al., 1944).

Foliation is prominent near the margins of the dome, outlining the structure formed during the Devonian. The dip of the foliation varies from place to place, indicating minor folding in addition to that of the major anticline (Chapman et al., 1944; Billings et al., 1946). The Jefferson dome is part of the larger Bronson Hill anticlinorium (Billings, 1956). The major anticline trends northeast and is asymmetric, with the northwestern limb dipping about 45° NW and the southeast limb dipping about 75° SE (Chapman et al., 1944).

4.2.2.8 Geophysics

Jaupart et al. (1982) gave a heat flow value of 0.96×10^{-6} cal/cm^2·s for the area.

4.2.3 Lost Nation Group (Plate III, Pluton 9)

See Sec. 5.2.2 and Plate IV, pluton 8. Although outcrops of this group are found over an area of about 282 km^2 (109 mi^2) in New Hampshire, it has been studied mostly from the Vermont side, and it is therefore discussed with the plutons of Vermont.

4.2.4 Mascoma and Lebanon Domes (Plate III, Pluton 6)

4.2.4.1 Size, Shape, and Location

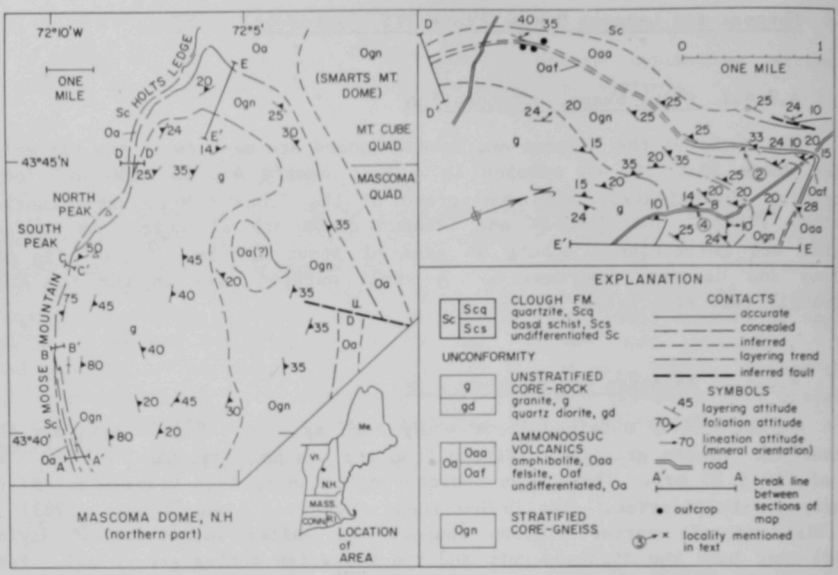
The rocks of the Mascoma and Lebanon domes are members of the Oliverian series domes that lie en echelon in a belt nearly 450 km (250 mi) long, coinciding approximately with the crest of the Bronson Hill anticlinorium (Naylor, 1968). The Mascoma and Lebanon domes are elliptical in surface outline and collectively occupy an area of about 210 km^2 (81 mi^2) in the Mascoma and Hanover quadrangles. A small portion lies in the Mt. Cube quadrangle (see Fig. 4.2).

4.2.4.2 Relative and Absolute Age

Naylor (1969) obtained Rb-Sr whole rock ages of 440 ± 20 m.y. for the Mascoma and Lebanon granites. Zircon from the Mascoma dome has a $^{207}Pb/^{206}Pb$ age of 450 ± 25 m.y. These data indicate a Middle to Late Ordovician age of primary crystallization. The U-Pb-zircon studies of Lyons et al. (1983) on the Highlandcroft series plutons support the ealier suggestion of Naylor (1968) that both the Highlandcroft and the Oliverian series are coeval. Part of the Highlandcroft suite, however, may have been derived from the 1000-m.y.-old Precambrian crust, as variable contamination by xenocrystic zircons is observed in the rocks of the series (Lyons et al., 1983).

4.2.4.3 Origin

Hadley (1942) believed that the Mascoma dome and the adjacent Smarts Mountain dome are phacolithic intrusions that were subsequently folded into domes. Chapman et al. (1944) considered the composite igneous core to have resulted from the forceful intrusion of several successive lenticular magmatic bodies. Bean (1951) interpreted the gravity data to mean that the dome is either a thick laccolith or a thick sheetlike intrusion that has been domed upward. Such mechanisms do not explain the uniform and consistent contact between the Oliverian core rocks and the Ammonoosuc volcanics. Naylor (1968 and 1969) studied the Mascoma dome as an example of the Oliverian domes. He divided the dome into stratified and unstratifed core rocks. He further proposed that the stratified rocks are probably metamorphosed Ammonoosuc volcanics, whereas the granitic core is the result of intrusive igneous activity, and is not remobilized basement. The Lebanon granite represents a syntectonic stock that has domed its roof (Chapman, 1939; Lyons, 1955; and Naylor, 1969).



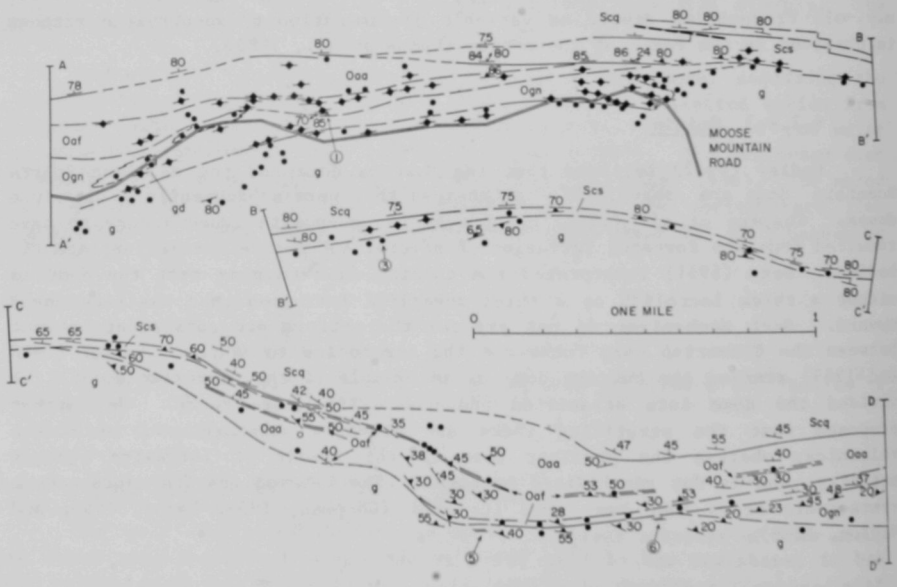


Fig. 4.2 Geologic Map of the Northern Part of the Mascoma Dome (A-A', B-B', etc., show locations of detail maps shown below) (Source: Naylor, 1969)

4.2.4.4 Petrography

Mineralogy and Textural Relations. In the Mascoma dome, stratified core rocks are typified by massive layers of medium-grained plagioclase-quartz-biotite-epidote gneiss of intermediate igneous composition whose upper parts are more heterogeneous than the lower parts. Core rocks are unstratified and homogeneous. Chapman (1939) described four types of igneous rocks, representing unstratified core, that grade into one another. These four types have been identified as granite, quartz monzonite, sodic granodiorite, and quartz diorite. Granite occurs in a small, nearly circular area in the western part of the main body. It is porphyritic, medium grained, granoblastic, and composed of quartz, 40%; oligoclase, 42%; potassium feldspar, 12%; and biotite, 3%. The mineral composition suggests that the rock is a granodiorite. Minor amounts of muscovite, magnetite, apatite, and sphene also are present. Moderate foliation is produced by parallel arrangement of biotite grains and flattened quartz aggregates.

Quartz monzonite is most abundant and occupies the central part of the main mass. It is medium-grained, gneissic, and composed of quartz, 30%; microcline, 24%; plagioclase, 33%; and biotite, 11%; with minor epidote, muscovite, magnetite, apatite, and sphene. The quartz shows strain shadows and granulation. With increasing microcline, monzonite grades into granite.

Sodic granodiorite occupies a broad, arcuate band around the north and east sides of the granite and quartz monzonite. It is medium grained, strongly foliated, and composed of a friable aggregate of quartz, 48%; plagioclase, 41%; potassium feldspar, 7%; and biotite, 4%; with minor muscovite, epidote, magnetite, apatite, and sphene components. Thin, parallel lamellae of biotite are distributed throughout the granodiorite, giving it considerable schistosity. The granodiorite generally is uniform, but the outer part contains inclusions of feldspathic hornblende-biotite schist and layers of gneiss, which may represent parts of the Ammonoosuc volcanics.

Quartz diorite is confined to a complex zone between the three other bodies and the Ammonoosuc volcanics. The zone contains four rock types: (1) biotite-quartz diorite (most abundant) composed of quartz, 29%; plagioclase, 64%; and biotite, 6%; with minor epidote, muscovite, chlorite, magnetite, apatite, sphene, and zircon; (2) siliceous biotite-quartz diorite (rare) composed of quartz, 51%; plagioclase, 42%; and biotite, 6%; with minor muscovite, magnetite, apatite, zircon, and garnet; (3) hornblende-quartz diorite (which is locally intruded into biotite-quartz diorite), composed of quartz, 25%; plagioclase, 53%; biotite, 10%; and amphibole, 12%; with minor potash feldspar and muscovite; and (4) granodiorite, which is similar to that of the main body, but is finer grained and occurs in sills. Many of these rocks are well foliated, and all are distinctly granular with characteristics similar to the stratified core rocks.

The relationships between the granodiorite and the granite and quartz monzonite indicate that the former is older than the other two. Hadley (1942)

concluded that the granite is the result of hydrothermal replacement of granodiorite or quartz monzonite by microcline.

Naylor (1968 and 1969) disputed Chapman's (1939) gradational subdivision of the rocks; he interpreted much of the dome to be metamorphased felsic volcanic rocks of the Ammonoosuc formation, and only the granites in the core as intrusives.

In the Lebanon dome, there are two distinct igneous rock types — a coarse-grained granitic core and a finer-grained, more mafic border phase that surrounds the core. The two are gradational near their contacts.

The border phase is a medium- to fine-grained rock, ranging in composition from gabbro to quartz monzonite. Although it possesses good gneissic structure in most places, only a pencil structure is present in others. The rock is highly sheared; quartz grains are granulated and recrystallized. Locally, quartz constitutes 25% of the rock, and plagioclase constitutes 30-65%. Epidote, calcite, and sericite are abundant and probably represent alteration products. Muscovite is locally abundant and thought to result from hydrothermal alteration of biotite. Other minor constituents include apatite, zircon, ilmenite, magnetite, and pyrite.

The Lebanon granite is a medium- to coarse-grained rock with poorly developed gneissic structure. Microcline is the most abundant mineral, occurring in amounts up to 55%. Secondary plagioclase (usually sodic oligoclase) is present in minor amounts. Quartz constitutes roughly 30% of the rock. Biotite is scarce. Muscovite and epidote are present and believed to be secondary.

Alteration. Mascoma dome rocks show sparse alteration, as seen in the presence of minor epidote. Both the Lebanon granite and the bordering gneiss are altered. The border gneiss locally contains up to 5% secondary calcite. Epidote and sericite also are present and are thought to be secondary. Muscovite is locally abundant, resulting from hydrothermal alteration of biotite. The Lebanon granite is less altered, but contains plagioclase, muscovite, and epidote as alteration products.

4.2.4.5 Secondary Intrusions

The oligoclase granodiorite and the quartz diorite of the Mascoma dome are cut, in the southeastern part of the area, by irregular pegmatite dikes, which are probably associated with the Mt. Clough pluton (Hadley, 1938).

4.2.4.6 Country Rocks

The Mascoma and Lebanon domes are surrounded by the Ammonoosuc volcanics, which are variably metamorphosed felsic rocks.

4.2.4.7 Rock Structure

Oliverian domes are generally well jointed. The Lebanon dome has several different joint sets (Lyons, 1981). The most prominent joints in the Mascoma dome dip steeply toward the center of the dome. Many of those in the granodiorite have been filled with quartz-tourmaline veins. These are bordered by bleached zones 2 cm to a few meters (an inch to several feet) wide, from which the biotite has been entirely removed.

Strong foliation parallel to the bedding is observed in the domes. This foliation is secondary and dips at the margin outward from 20° to 30°. Low-angle dips were recorded in the interior, but foliation is virtually absent in the core of Lebanon dome. The dome is overturned toward the south or southeast (Bean, 1953).

4.2.4.8 Geophysics

A large gravity low -- a Bouguer anomaly of approximately -10.5 mgal -- is associated with the Mascoma and Lebanon domes. Calculations indicate that dips of the contact increase at depth, and that the Mascoma dome extends to a depth of approximately 5000 m (16,000 ft) and the Lebanon dome to a depth of 3650 km (12,000 ft) (Bean, 1953).

Jaupart et al. (1982) reported that the heat flow is higher for these plutons than for the other rocks of the area. For the Lebanon area, heat flow determined for Oxfordsville quartzite is 0.76×10^{-6} cal/cm^2·s.

4.2.5 Newburyport Granodiorite (Plate III, Pluton 22)

See Sec. 7.2.7 and Plate VI, pluton 13. Outcrops of this body are found in an area of only about 25 km^2 (9.6 mi^2) in New Hampshire; the largest portion is located in Massachusetts and therefore it is described with the plutons of that state.

4.2.6 Westmoreland-Swanzey and Related Domes (Plate III, Pluton 3)

4.2.6.1 Size, Shape, and Location

Most of the Westmoreland-Swanzey dome is located in the Keene-Brattleboro area of New Hampshire (see Fig. 4.3). It is nearly divided into two lobes by tongues of metamorphic rock immediately west of the city of Keene. The northwestern lobe is about 16 km (10 mi) wide and 11 km (7 mi) long; the eastern lobe is about 11 km (7 mi) wide and 26 km (16 mi) long. Two prongs of the northwestern lobe extend northward into the Bellows Falls quadrangle. Strips of similar rock extend across Massachusetts (see pluton 8 on Plate VI) and into Connecticut (see pluton 11 on Plate VIII).

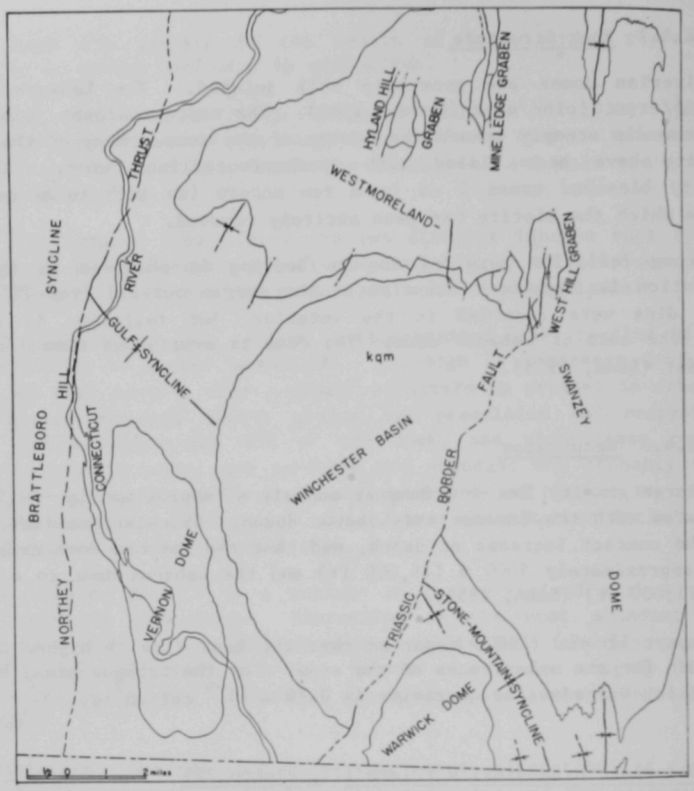


Fig. 4.3 Major Oliverian Domes in the Keene-Brattleboro Area and Associated Structural Elements (Source: Moore, 1949)

Warwick dome continues into the adjoining Northfield and Mt. Grace quadrangles in Massachusetts; Vernon dome, an oval body, occupies the lowland between the southern end of Wantastiquet Mountain and the New Hamphsire-Massachusetts state line. Westmoreland, Swanzey, Vernon, Warwick, and Surrey Mountain domes are grouped together due to similarities of age and lithology. They collectively occupy an area of about 259 km^2 (100 mi^2).

4.2.6.2 Relative and Absolute Age

The domes are considered to be part of the Oliverian series, to which Zartman and Leo (1981) gave an age of 455 m.y.

4.2.6.3 Origin

Westmoreland-Swanzey and associated domes were emplaced by forceful injection of thick composite sheets. The event occurred in the early stages of the Taconic orogeny, while overlying rocks were essentially flat. Subsequent doming by diapiric uprise of the rocks occurred during the Acadian orogeny of Early to Middle Devonian time to give the present structural forms (Lyons, 1981).

4.2.6.4 Petrography

Mineralogy and Textural Relations. All these domes are lithologically similar and consist predominantly of granodiorite gneiss. The core gneisses are trondhjemitic, and are probably a plutonic phase of the Ammonoosuc (Leo, 1980). The description that follows focuses on the Westmoreland-Swanzey dome. The gneiss of the northwestern lobe of the dome is predominantly granodiorite; however, the northwestern part of this lobe is quartz monzonite, and local areas of granite are present. Moore (1949) reported modal analyses for all three types of rock. For the granodiorite, he found plagioclase, 44%; quartz, 37%; and potassium feldspar, 11%; with minor biotite, muscovite, epidote, and chlorite; for the quartz monzonite, he reported quartz, 33%; plagioclase, 31%; potassium feldspar, 28%; and biotite, 5%; with minor muscovite, chlorite, and magnetite; and for the granite, he found potassium feldspar, 45%; quartz, 30%; plagioclase, 18%; biotite, 3%; and muscovite, 3%, with minor chlorite and magnetite. The granodiorite of the southeastern part of this lobe contains significant hornblende in addition to the more abundant biotite. Gneiss in the eastern lobe of the pluton is quartz diorite of the following composition: plagioclase, 53%; quartz, 32%; biotite, 5%; and hornblende, 5%; with minor potassium feldspar, chlorite, and epidote. Local areas are enriched in potassium feldspar. Accessory amounts of magnetite, garnet, epidote, muscovite, apatite, zircon, sphene, and rutile are present in most rocks.

Contacts between the varieties of gneiss are gradational, and the different rock types are presumed to be contemporaneous. Thin schlieren of fine-grained, biotite-rich gneiss are abundant near the border of the intrusive. These schlieren are probably reworked inclusions of amphibolite from the Ammonoosuc volcanics.

Alteration. Chlorite and sericite are present in the gneiss as alteration products.

4.2.6.5 Secondary Intrusions

Small pegmatites occur either in shear zones or in dikelike or sill-like masses.

4.2.6.6 Country Rocks

The western half of the Westmoreland-Swanzey dome is overlain by interbedded medium-grade metamorphic rocks of the Ammonoosuc volcanics. The predominant rock types are biotite-quartz-feldspar gneiss and amphibolite; some actinolite-quartz-feldspar gneiss is present.

The eastern half of the eastern lobe is overlain by high-grade metavolcanics. The most abundant rocks are biotite-quartz-feldspar gneiss and amphibolite similar to those of the middle-grade zone, but with a coarser grain size. Among the less common rocks are sillimanite-quartzite, talc-muscovite schist, and hornblende-chlorite-talc schist. Thin beds of coarse anthophyllite schist also occur.

Bedding is well defined throughout most of the metavolcanics; beds range in thickness from about a centimeter to a few meters (a fraction of an inch to several feet). Foliation is well developed and parallel to the bedding.

On the western side of the eastern lobe, stratified rocks are absent, and gneiss of the Westmoreland-Swanzey pluton is in contact with the Kinsman quartz monzonite along a Triassic-Jurassic border fault.

4.2.6.7 Rock Structure

A Triassic-Jurassic border fault dated at 160 m.y. (Lyons, 1981) runs along the western side of the eastern lobe of the dome.

Foliation of the gneiss is believed to be primary, resulting from intrusion of a partially crystallized magma. It is well developed around the border of the dome, but decreases in intensity toward the interior. The foliation and bedding of the Ammonoosuc volcanics, the foliation and bedding of the gneiss, and the contact between the two are all parallel.

The average outward dip of the foliation along the border of the northwestern lobe of the Westmoreland-Swanzey dome is 25-40°. At the extreme western end of the northwestern lobe, foliation in the gneiss dips outward from the center. Toward the west, the dips steepen and become overturned.

Dips in the foliation around the eastern lobe are generally steeper (65°) than those around the northwestern lobe (about 30°). At the northwestern extremity of the eastern lobe, the structure is overturned.

4.2.6.8 Geophysics

Jaupart et al. (1982) reported a heat flow value of 1.42×10^{-6} $cal/cm^2 \cdot s$ for the Amonoosuc volcanic rocks in the Westmoreland area. This is

higher than the values determined for the Lebanon area, and higher than the mean heat flow for the metasedimentary rocks.

4.3 ACADIAN AND POST-ACADIAN

4.3.1 Ashuelot Pluton (Plate III, Pluton 2)

4.3.1.1 Size, Shape, and Location

The Ashuelot pluton occurs in the Keene-Brattleboro area in the southwestern corner of the state. The pluton occupies an area of about 116 km^2 (45 mi^2) and is roughly elliptical, about 18 km (11 mi) long and 10 km (6 mi) wide. The pluton is bisected by the Ashuelot River and is marked by a hilly region and lowland swamps.

4.3.1.2 Relative and Absolute Age

The Ashuelot pluton is composed of Kinsman quartz monzonite, which intrudes into the Lower Devonian Littleton formation (Moore, 1949). Lyons and Livingston (1977) reported an age of 411 ± 19 m.y. from a Rb-Sr whole-rock isochron. This age presents a problem, as it lies outside the commonly accepted time span (400-350 m.y.) for the Devonian. This problem is common to all bodies of the New Hampshire plutonic series.

4.3.1.3 Origin

The Ashuelot pluton was forcefully injected as a thick sheet after earlier stratified deposits had been intensely folded. The pluton contains inclusions of folded country rock, and the schistosity of some of the inclusions is not parallel to the foliation of the Ashuelot. Because the phenocrysts in the quartz monzonite show very little granulation, the Ashuelot must have been intruded into the Littleton formation near the end of the deformation (Moore, 1949).

4.3.1.4 Petrography

Mineralogy and Textural Relations. The Kinsman quartz monzonite is a major lithologic unit of the New Hampshire plutonic series and has quartz diorite and granodiorite as compositional variants. The Kinsman quartz monzonite of the Ashuelot pluton is a medium- to coarse-grained, light-gray porphyritic rock containing large tabular phenocrysts of white feldspar. In most places the phenocrysts are microcline or microperthite; elsewhere they are plagioclase. The largest phenocrysts are about 4 x 2.5 x 1.5 cm (1.6 x

1.0 x 0.6 in.). The percentage of phenocrysts is highly variable, averaging about 10% of the rock but ranging to a maximum of about 25%. Owing to this variation in the amount and composition of the phenocrysts, portions of the Ashuelot pluton are granodiorite, while other parts are quartz diorite. The groundmass is chiefly feldspar, quartz, biotite, and muscovite. In the Ashuelot pluton, the Kinsman quartz monzonite has a hypidiomorphic granular texture. Essential minerals are oligoclase or andesine, quartz, potassium feldspar, biotite, and muscovite. The plagioclase (An-23-37) occurs as homogeneous grains. Garnet, sillimanite, zircon, apatite, pyrite, magnetite, sphene, and rutile are present as accessory minerals (Moore, 1949).

Alteration. Schists found in the Ashuelot pluton are inclusions, rather than alteration products. A large schistose body of the Littleton formation about 5 km (3 mi) long in a north-south direction and 1.6 km (1 mi) wide is present in the east-central part of the pluton.

4.3.1.5 Secondary Intrusions

Minor pegmatitic intrusions are present.

4.3.1.6 Country Rocks

The Ashuelot pluton intruded into the Lower Devonian Littleton formation, which consists chiefly of interbedded schist and numerous subordinate thin quartzites. It also contains calc-silicate rocks, volcanics, and quartz conglomerate. Bedding is well defined throughout most of the formation.

4.3.1.7 Rock Structure

The Ashuelot pluton is bordered on its eastern margin by a Triassic border fault that trends N. 25° E. and dies out in the vicinity of Keene. The western side of the fault is downthrown, the fault plane dipping steeply northwest. Although the Kinsman quartz monzonite is well foliated near its contact with the Littleton formation, foliation is weak throughout most of the pluton's interior.

4.3.1.8 Geophysics

Gravity profiles across the Ashuelot pluton indicate a maximum thickness of 0.8 km (0.5 mi) (Lyons, 1982). Jaupart et al. (1982) reported a heat flow value of 1.50×10^{-6} cal/cm^2·s for Kinsman tonalite, which is almost twice the value of the lowest heat flow determined in New Hampshire.

4.3.2 Cardigan Pluton (Plate III, Pluton 5)

4.3.2.1 Size, Shape, and Location

The Cardigan pluton is an elongated body about 97 km (60 mi) long north and south, and from 10 to 25 km (6 to 15 mi) wide. A narrow spur about 2 km (1.2 mi) wide juts south-southwest from the center, extending for about 35 km (21 mi). The southern end is at West Peterborough, and the northern end is 3 km (2 mi) north of Groton in west-central New Hampshire. The mapped area of the pluton, as shown on Plate III, is approximately 1194 km^2 (461 mi^2).

4.3.2.2 Relative and Absolute Age

On the basis of field relations, Fowler-Lunn and Kingsley (1938) interpreted the Cardigan pluton as having been intruded into older rock during the Late Devonian. A Rb-Sr age of 402 ± 19 m.y. has been reported by Lyons and Livingston (1977).

4.3.2.3 Origin

The Cardigan pluton is a synkinematic, elongate, basin-shaped pluton intruded during the Acadian orogeny. It was intruded as a sheet concurrently with the development of a series of west-facing nappes in Devonian and older metasediments (Lyons, 1982).

4.3.2.4 Petrography

Mineralogy and Textural Relations. The Cardigan pluton is composed of the Kinsman quartz monzonite, which belongs to the New Hampshire plutonic series. The rock is dark gray, medium to coarse grained, and is massive to strongly foliated. The principal minerals are oligoclase-andesine, quartz, potassium feldspar, biotite, and muscovite. An average modal composition is quartz, 26%; potassium feldspar, 45%; plagioclase, 13%; biotite, 10%; and muscovite, 5%. The unit is sometimes porphyritic, with phenocrysts of potassium feldspar averaging 3.5-5 cm (1.4-2.0 in.) in length. Locally, these phenocrysts constitute 10% of the rock but may account for as much as 20%. The amount of oligoclase-andesine decreases as the number of these phenocrysts of potassium feldspar increases. The average composition of the non-porphyritic phases is granodiorite, whereas the average composition of the porphyritic phases is quartz monzonite. The rock exhibits hypidiomorphic granular texture (Billings, 1956).

Alteration. A granoblastic and cataclastic texture is found in the northern end of the pluton (Fowler-Lunn and Kingsley, 1937).

4.3.2.5 Secondary Intrusions

There are no secondary intrusions in the Cardigan pluton. There are slablike inclusions of mica schist that have been reworked to a rock resembling the Kinsman quartz monzonite.

4.3.2.6 Country Rocks

The Cardigan pluton intruded into the Littleton formation of Lower Devonian age (Lyons and Livingston, 1977). The Littleton formation consists of schists and quartzites. The quartzose members vary from fine- to medium-grained quartzites and garnetiferous quartz-mica schists to rocks that appear much like fine-grained granites. All gradations of schist are found, i.e., from rocks in which only quartz and mica show in hand specimens through garnet-sillimanite schists (Fowler-Lunn and Kingsley, 1937).

4.3.2.7 Rock Structure

The Kinsman quartz monzonite of the Cardigan pluton exhibits a coarse foliation, and phenocrysts of potassium feldspar and granulated aggregates of other minerals show a lineation (Fowler-Lunn and Kingsley, 1937). Foliation in the Kinsman quartz monzonite is parallel to its contacts with schist inclusions. Bullen (1977) reported a folding event after emplacement of the Cardigan pluton that produced an axial-plane foliation in the associated rocks.

4.3.2.8 Geophysics

Gravity data collected for members of the New Hampshire plutonic series indicate that the Cardigan pluton was emplaced as a thin sheet with a maximum depth of 2500 m (8200 ft) (Nielson et al., 1976). This figure agrees well with the 2000-m (6550-ft) depth estimated by Billings (1956). The pluton has an elongate basin shape, with a gentle easterly dip on the western side and a steep to overturned dip on the easterly side. Heat-flow values reported by Jaupart et al. (1982) for the Kinsman quartz monzonite were high.

4.3.3 Concord and Related Plutons (Plate III, Pluton 7)

4.3.3.1 Size, Shape, and Location

The Concord pluton forms an elongate body in the Concord, Rumney, Cardigan, and Plymouth quadrangles, occupying an area of 272 km^2 (105 mi^2). Fitzwilliam granite, around Fitzwilliam, and Bickford granite, near Franconia (Williams and Billings, 1938), are other names for similar granite in the

respective geographic areas (Emerson, 1917). Large bodies of Concord-type granite have been mapped in the Suncook, Winnipesaukee, and Dixville quadrangles. In all, nine disconnected bodies of Concord-type granite are shown on Plate III, disposed around the Concord pluton in a ring nearly 100 km (60 mi) in diameter. Collectively, these bodies of Concord-type rock crop out over an area of about 727 km^2 (280 mi^2). The description that follows pertains primarily to the Concord pluton itself.

4.3.3.2 Relative and Absolute Age

Lyons and Livingston (1977) obtained Rb-Sr whole-rock ages of 359 ± 11 m.y. and 330 ± 3 m.y., which are similar to the age found for the Concord granite from the type area, but younger than the coeval Barre granite in Vermont (Naylor, 1971).

4.3.3.3 Origin

The Concord granite is the youngest member of the New Hampshire plutonic series. Nielson et al. (1976) interpreted the parallelism of foliation to the intrusive contacts to indicate a posttectonic intrusion of the Concord. The magma was probably emplaced by a process of brittle fracturing and cauldron stoping.

4.3.3.4 Petrography

Mineralogy and Textural Relations. The term "binary granite" has been used to describe the Concord and related plutons. Green (1970) reported that quartz monzonite is the most representative rock of such plutons, but the composition varies from granite to granodiorite. Concord-type granites are fine- to medium-grained hypidiomorphic granular rocks. They are locally porpryritic and often slightly foliated.

The Concord pluton is medium to coarse grained and equigranular. It is fine grained near Concord. The earlier phase is somewhat foliated. The principal minerals are quartz, 31%; potassium feldspar, 28%; oligoclase, 29%; biotite, 5%; and muscovite, 7% (Green, 1970).

Alteration. There is no information in the literature on secondary minerals in the pluton, which may indicate that alteration is not significant.

4.3.3.5 Secondary Intrusions

Pegmatite intrusions from a few meters to a few kilometers long and hundreds of meters thick are particularly associated with the Concord granites and may represent a late-stage phenomenon (Bannerman, 1943; Olson, 1950; and Cameron et al., 1954).

4.3.3.6 Country Rocks

The Concord pluton is intrusive into the Littleton formation, and its apophyses intrude into the Kinsman quartz monzonite (Billings, 1956).

4.3.3.7 Rock Structure

No major faults or folds are described for the plutons. Fowler-Billings (1949) reported northeast-trending, silicified fracture zones with moderate northwesterly dips in the northern and northeastern portions. These zones trend north and south with vertical or steep westerly dips in the southern portions. The silicified zones average 46 m (150 ft), wide and are roughly parallel to the strike. Foliation in the Concord granite varies from pluton to pluton, but is commonly weak. The foliation roughly outlines the shape of the pluton. The Concord pluton is not deformed, except for a small foliated zone near the contact and minor straining of quartz grains (Billings, 1956).

4.3.3.8 Geophysics

Figure 4.4 shows that the Concord pluton occupies a gravity low. Some of it lies below the Littleton formation, to a depth of about 2 km (1.2 mi) (Nielson et al., 1976). Jaupart et al. (1982) gave a heat flow of approximately 1.0×10^{-6} cal/cm^2·s for the area.

4.3.4 Exeter Pluton (Plate III, Pluton 21)

4.3.4.1 Size, Shape, and Location

The Exeter pluton is an elongate body about 7 km (4 mi) wide and 32 km (20 mi) long that trends northeast and occupies a mapped area of approximately 115 km^2 (44 mi^2). It is located in the seaboard lowland province of New Hampshire; topographic relief is provided by hills that range in height from 24 m (80 ft) to 152 m (500 ft), capped with glacial material (Sundeen, 1971).

4.3.4.2 Relative and Absolute Age

The Exeter pluton is a member of the Hillsboro plutonic series. Preliminary Rb-Sr dating and structural relations suggest an age of approximately 390 m.y. (Gaudette et al., 1975). The pluton is cut by dikes belonging to the White Mountain series of Late Triassic age (Novotny, 1969).

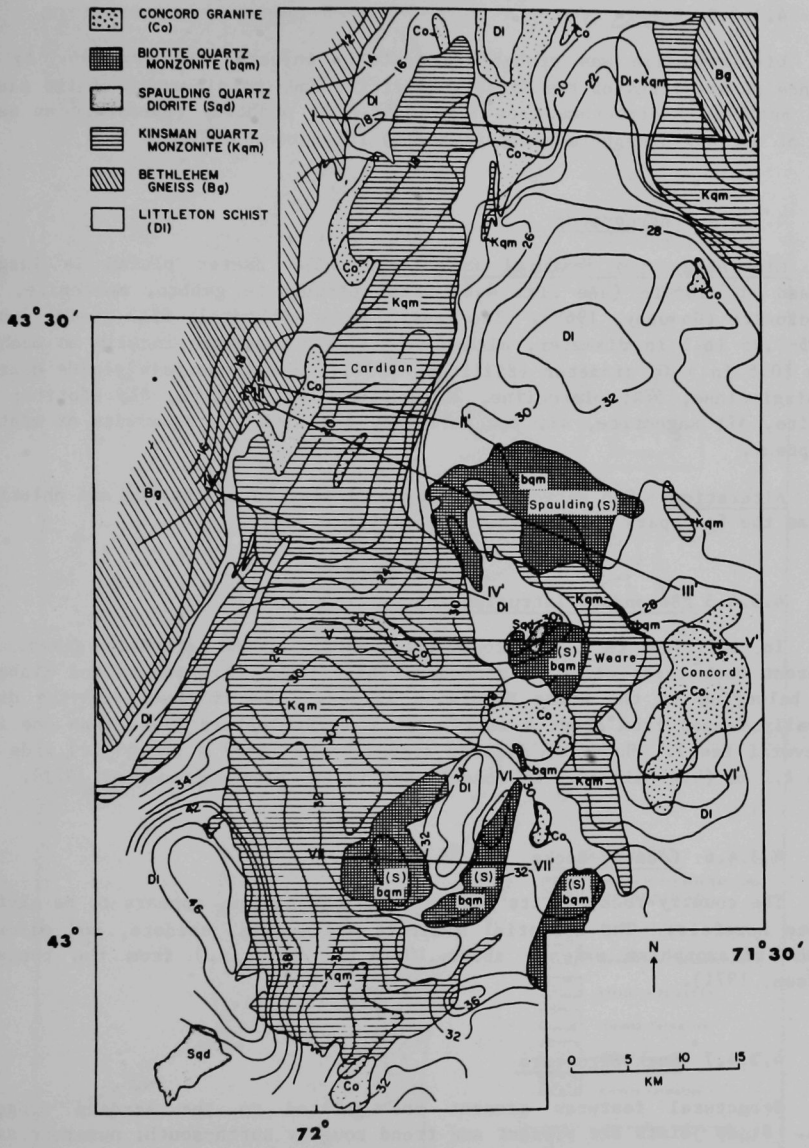


Fig. 4.4 Bouguer Anomaly Map Covering Cardigan, Weare, Spaulding, and Concord Plutons (contour interval is 2 mgal; roman numerals refer to lines of gravity profiles) (Source: Adapted from Nielson et al., 1976)

4.3.4.3 Origin

Diorite magma was probably forcefully injected, as evidenced by the presence of xenoliths of the Kittery quartzite and the situation of the pluton in an anticlinal structure (see Fig. 4.5). It probably represents an early phase of the New Hampshire plutonic series (Freedman, 1950).

4.3.4.4 Petrography

Mineralogy and Textural Relations. The Exeter pluton is largely composed of diorite (see Fig. 4.6) with subordinate gabbro, monzonite, and granodiorite (Novotny, 1969). The crystals are subhedral, with grains 3-6 mm (0.125-0.25 in.) in diameter, although biotite flakes are locally as much as 13 mm (0.5 in.) in diameter (Billings, 1956). Modal analysis yields quartz, 6%; plagioclase, 57%; microcline, 2%; augite, 14%; uralite, 8%; biotite, 3%; chlorite, 5%; magnetite, 4%; and sericite, 1%; with trace amounts of apatite and sphene.

Alteration. The mafic rocks show modification to uralite and chlorite, whereas the feldspars are somewhat sericitized.

4.3.4.5 Secondary Intrusions

In addition to many crosscutting aplitic and granitic dikes, two different rock types occur as secondary intrusions: camptonite and diabase, both belonging to the White Mountains series. The thickness of the dikes generally ranges from about a centimeter to a few meters (less than one inch to several feet). However, a large diabase dike 15-18 m (50-60 ft) wide and up to 1.2 km (0.75 mi) long is conspicuous in the pluton (Sundeen, 1971).

4.3.4.6 Country Rocks

The country rock, at its contact with the pluton, appears to be albite-epidote hornfels. The essential minerals are quartz, epidote, and calcite. Contact metamorphism extends about 10-15 cm (4-6 in.) from the contacts (Sundeen, 1971).

4.3.4.7 Rock Structure

Structural features present are related to the Acadian orogenic phase. Steep joints are present and trend roughly north-south; numerous small faults are present in all the rock units, but the Portsmouth fault is the only prominent fault in the area. Foliation is present and defines the overturned anticlinal structure coincident with the pluton (see Fig. 4.6).

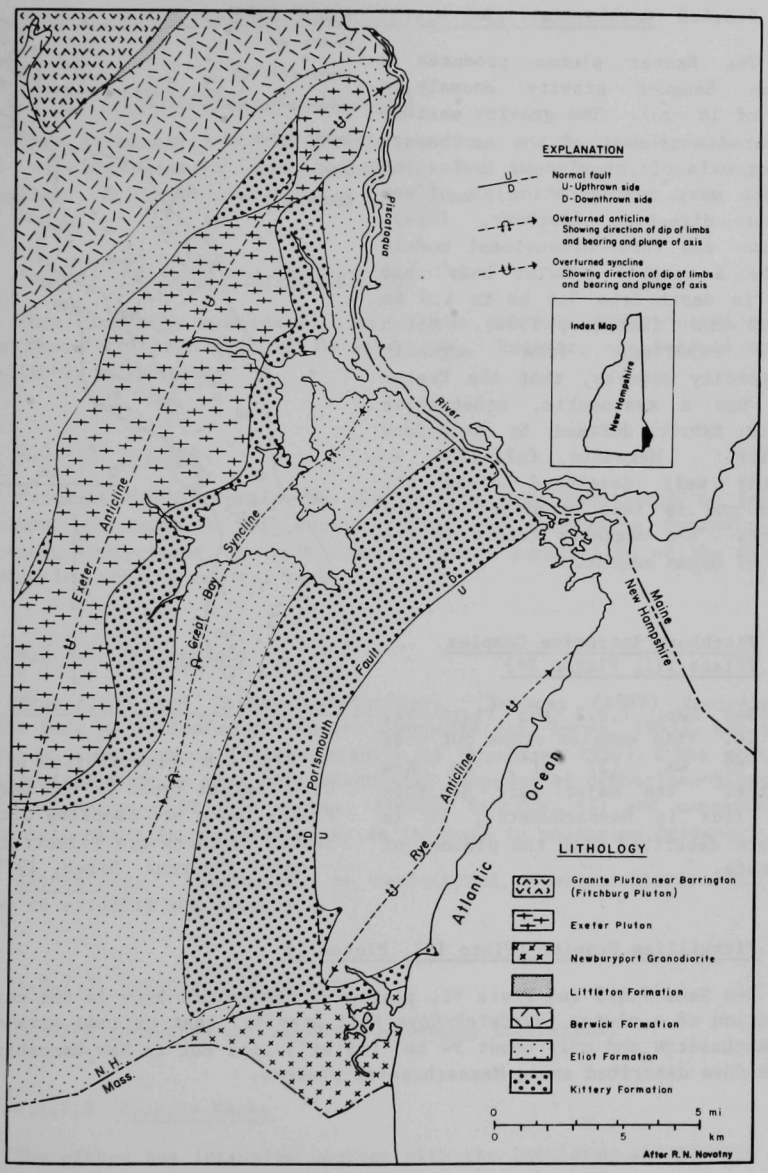


Fig. 4.5 Structure of the Seacoast Region of New Hampshire (Source: Novotny, 1969)

4.3.4.8 Geophysics

The Exeter pluton produces a positive Bouguer gravity anomaly in excess of 16 mgal. The gravity maximum is located northwest of the northeast-trending axis of the pluton, indicating that the mass concentration is offset in that direction (Bothner, 1974). Both two- and three-dimensional models describe a roughly tabular body that ranges in depth from 0.6 km to 1.9 km (0.4-1.2 mi) (Bothner, 1974). Birch (1979) reported, from magnetic suceptibility studies, that the Exeter pluton has a systematic, penetrative magnetic fabric defined by secondary magnetite. Magnetic foliation is extremely well developed along the margins and is less prominent in the interior. The magnetic fabric is the result of magma ascent.

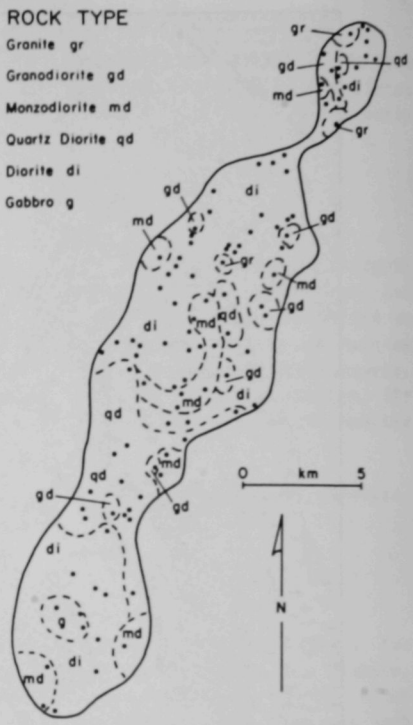


Fig. 4.6 Distribution of Lithological Types in the Exeter Pluton (dots are sampling sites; dashed contacts are illustrative only) (Source: Birch, 1979)

4.3.5 Fitchburg Intrusive Complex (Plate III, Pluton 24)

See Sec. 7.3.2 and Plate VI, pluton 10. This complex crops out over about 69 km^2 (27 mi^2) in New Hampshire. The major part of this pluton lies in Massachusetts; it is therefore described with the plutons of that state.

4.3.6 Fitzwilliam Granite (Plate III, Pluton 1)

See Sec. 7.3.3 and Plate VI, pluton 9. This rock body is the northern termination of a pluton (Hardwick-Coys Hill granite) that extends across most of Massachusetts and only about 94 km^2 (36 mi^2) lies out in New Hampshire; it is therefore described among Massachusetts plutons.

4.3.7 Granite Pluton near Barrington (Plate III, Pluton 19)

4.3.7.1 Size, Shape, and Location

This irregular body of crystalline rock is located in the seacoast region of southeastern New Hampshire, in the Pawtuckaway and Dover quadrangles. It occupies an area of approximately 179 km^2 (69 mi^2).

4.3.7.2 Relative and Absolute Age

This pluton is considered part of the New Hampshire plutonic series, and is therefore considered of Devonian age. However, no absolute age data are available at present.

4.3.7.3 Origin

Freedman (1950) considered the granite near Barrington to represent emplacement of magma by piecemeal stoping and permissive intrusion in a syntectonic environment, as is reflected in the parallelism of the fabric in the intrusion and the country rock.

4.3.7.4 Petrography

Mineralogy and Textural Relations. Novotny (1969) described the plutonic body as predominantly quartz monzonite (cf. Lyons, 1983), but biotite granite is also present (binary granite of Freedman, 1950). The medium- to coarse-grained rock is gray and essentially composed of orthoclase-microcline, 28-51%; oligoclase, 16-35%; quartz, 24-28%; biotite, 4%; and muscovite, 4%. The granitic phase is represented by an increase in potassium feldspar.

Alteration. Plagioclase is sericitized, whereas biotite is partly altered to chlorite and magnetite.

4.3.7.5 Secondary Intrusions

Granite pematites are abundant secondary intrusions.

4.3.7.6 Country Rocks

The pluton has intrusive contact with the Littleton and Berwick formations (see Fig. 4.5).

4.3.7.7 Rock Structure

Faults are marked by silicified zones that trend northeast and northwest. Flint Hill is one such silicified zone, which reaches a length of almost 1 km (0.6 mi). The faults commonly range to 100 m (327 ft) long and 3 to 30 m (10 to 100 ft) wide. Joints strike northeast and dip steeply to northwest or southeast, characterizing a gently dipping sheetlike intrusion. Foliation is parallel to the fabric of the country rock, which trends north-northeast and dips steeply (Novotny, 1969).

4.3.7.8 Geophysics

Jaupart et al. (1982) gave a heat-flow value in the Barrington area of $0.99\text{-}1.87 \times 10^{-6}$ cal/cm^2·s.

4.3.8 Lincoln Pluton (Plate III, Pluton 8)

4.3.8.1 Size, Shape, and Location

The Lincoln pluton includes a large, lenticular body 40 km (25 mi) long and up to 10 km (6 mi) wide, which extends from 8 km (5 mi) southeast of Woodstock on the south to 3 km (2 mi) south of Twin Mountain on the north. Another part to the south is an irregularly angled strip of rock 2-5 km (1.2-3 mi) wide and about 30 km (18 mi) in its total length; together these bodies occupy about 221 km^2 (85 mi^2).

4.3.8.2 Relative and Absolute Age

The Lincoln pluton is composed of Kinsman quartz monzonite, which belongs to the New Hampshire plutonic series. Billings (1956) assigned this intrusive to the Late Devonian. However, recent Rb-Sr data indicate an age of 402 ± 19 m.y. (Lyons and Livingston, 1977).

4.3.8.3 Origin

Contact relations have been interpreted by Billings (1956) to indicate forceful injection.

4.3.8.4 Petrography

Mineralogy and Textural Relations. The Lincoln pluton is composed of Kinsman quartz monzonite; it is dark gray, medium to coarse grained, and massive to foliated. An average modal analysis shows the principal minerals

to be quartz, 26%; potassium feldspar, 45%; plagioclase, 13%; biotite, 10%; and muscovite, 5%. The intrusive is locally porphyritic, with phenocrysts of potassium feldspar averaging 3-5 cm (1.1-2.0 in.) long. These phenocrysts constitute 10% of the rock in places, and occasionally account for up to 20%. The amount of oligoclase-andesine decreases with an increase in the number of phenocrysts of potassium feldspar. The nonporphyritic phases are, on average, granodiorite, whereas the porphyritic phases are mostly quartz monzonite. The rock generally exhibits hypidiomorphic granular texture (Billings, 1956).

Alteration. The only modification of the Lincoln pluton reported in the literature is the development of granoblastic texture.

4.3.8.5 Secondary Intrusions

The Lincoln pluton is cut by dikes and stringers of the Mt. Garfield porphyritic syenite of the White Mountain batholith. This syenite is a fine-grained, light-gray porphyry, rich in dark minerals. The phenocrysts are chiefly microperthite. A portion of the Lincoln pluton is isolated by intrusions of the Mesozoic White Mountain magma series.

4.3.8.6 Country Rocks

The Lincoln pluton intrudes into the Littleton formation of Lower Devonian age. It consists of schists and quartzites. The quartzose members vary from fine- or medium-grained quartzitic and garnetiferous quartz-mica schists to rocks that appear much like fine-grained granites. Among the schists all gradations may be found, from rocks in which only quartz and mica show in hand specimens to garnet-sillimanite schists (Fowler-Lunn and Kingsley, 1938).

4.3.8.7 Rock Structure

The Kinsman quartz monzonite shows north-south foliation with a vertical dip near contacts. Elsewhere, the foliation is parallel to the contacts. The body is shaped like a large sill or lens that dips vertically or with a very steep northwestern component (Williams and Billings, 1938). The Lincoln pluton intruded into sedimentary rocks during the final stages of their folding, and is itself not folded.

4.3.8.8 Geophysics

The Kinsman quartz monzonite is reported to have a heat-flow value of 1.47 to 1.50 x 10^{-6} cal/cm^2·s (Jaupart et al., 1982).

4.3.9 Long Mountain Granite (Plate III, Pluton 12)

4.3.9.1 Size, Shape, and Location

The Long Mountain granite is an irregular, narrow-waisted body about 28 km (17 mi) long from southwest to northeast, with widths from west to east that vary from 2 to 20 km (1.2 to 12 mi); it occupies an area of about 191 km^2 (74 mi^2). It crops out in Coos County in the southern part of the Dixville quadrangle and the northern part of the Percy quadrangle. Hatch (1963) reported that Nash Bog Pond stock is a continuation of the Long Mountain pluton (see Fig. 4.7). Green (1964) mapped its extension (or similar rock) east into the Errol quadrangle.

4.3.9.2 Relative and Absolute Age

The pluton is composed of Concord granite, and thus belongs to the New Hampshire plutonic series. A radiogenic age of 323 ± 11 m.y. has been reported by Lyons and Livingston (1977) for Concord granite from west-central New Hampshire. The various bodies of Concord granite have frequently been interpreted as being the same age (Billings, 1956).

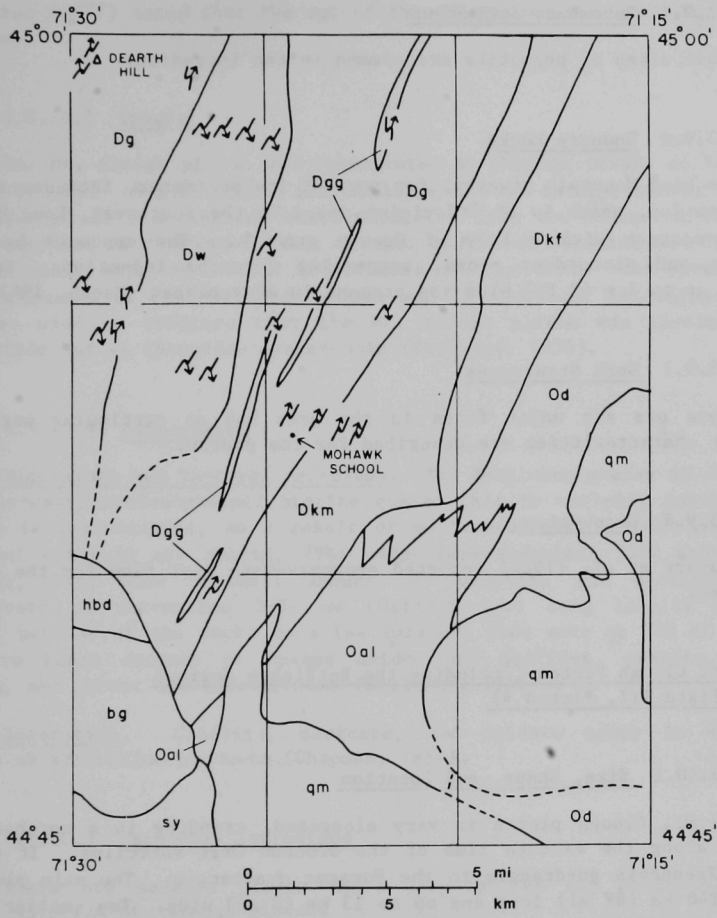
4.3.9.3 Origin

Because the Concord granite intruded into the region after the Acadian orogeny, Billings (1956) suggested that the country rock was deformed in a brittle manner. If it had been emplaced during the Acadian orogeny, the country rock would have deformed in a more plastic manner. Billings (1956) interpreted shattered country rock as indicative of forceful injection.

4.3.9.4 Petrography

Mineralogy and Textural Relations. The Long Mountain granite pluton has two monzonite phases. One ranges from fine to medium grained but is locally coarse; it is a muscovite-biotite quartz monzonite. The other phase, generally equigranular but locally porphyritic, is a gray biotite quartz monzonite. The main mineral assemblages are quartz, 25-35%; potassium feldspar (as orthoclase and microcline), 23-32%; plagioclase, 30-35%; biotite, 5-6%; and muscovite, 2-5% (Hatch, 1963).

Alteration. Modification of feldspars is shown by the presence of sericite; modification of biotite is shown by the presence of chlorite (Hatch, 1963). Alteration, however, is minimal.



Explanation

bg	Biotite granite
sy	Syenite
hbd	Hornblende-biotite diorite
qm	Quartz monzonite (Long Mountain Pluton)
Dg	Gile Mountain formation
Dgg	Greenstone and amphibolite
	Kidderville formation
Dkm	Mafic volcanic member
Dkf	Felsic volcanic member
Dw	Waits River formation
Od	Dixville formation
Oal	Albee formation
	Drag fold

Fig. 4.7 Simplified Geologic Map of the Dixville Quadrangle (directions of plunge of drag folds associated with second deformation are shown) (Source: Adapted from Hatch, 1963)

4.3.9.5 Secondary Intrusions

Small dikes of pegmatite are common in the intrusion.

4.3.9.6 Country Rocks

The Long Mountain granite, for most of its perimeter, intrudes into the Albee formation, which is of Ordovician age. To the southwest, Long Mountain shares a contact with a body of Conway granite. The contacts have both concordant and discordant zones, suggesting forceful injection. Zones of migmatite up to 1 m (3 ft) wide are present in many places (Hatch, 1963).

4.3.9.7 Rock Structures

There are six major folds in the area but no particular structural control or characteristics are described for the pluton.

4.3.9.8 Geophysics

Jaupart et al. (1982) reported above-average heat flow for the plutons of the area.

4.3.10 Mt. Clough Pluton (Including the Bethlehem Gneiss) (Plate III, Pluton 4)

4.3.10.1 Size, Shape, and Location

The Mt. Clough pluton is very elongated, trending in a northeasterly direction along the eastern limb of the Bronson Hill anticline. It extends from the Franconia quadrangle to the Sunapee quadrangle. The main pluton is at least 140 km (87 mi) long and up to 13 km (8 mi) wide. Two smaller bodies nearby are parallel to the main pluton; these are both 2-3 km (1.2-1.8 mi) wide, and are respectively 15 and 20 km (9 and 12 mi) long. The various outcrops of this rock occupy an area of 796 km^2 (307 mi^2). The pluton varies in depth from 0.8 km (0.5 mi) to 5 km (3 mi) (Billings, 1956).

4.3.10.2 Relative and Absolute Age

The Mt. Clough pluton is composed of Bethlehem gneiss. It intrudes into the Early Devonian Littleton formation. Lyons and Livingston's (1977) whole-rock Rb-Sr isochron gave an age of 405 $\pm$ 78 m.y. Because the Littleton formation may be as old as Middle Silurian (Billings and Fowler-Billings, 1975), and because of the wide age range obtained experimentally, Lyons and

Livingston (1977) noted that the age of Mt. Clough could be from Early to Late Devonian.

4.3.10.3 Origin

The Mt. Clough pluton was interpreted by Chapman (1939) to be a large, sheetlike mass. Heald (1950) and Billings (1956) interpreted it as having been injected as a horizontal sheet separated from the Oliverian series by a septum composed of the Ammonoosuc, Clough, and Fitch formations. This septum and horizontal sheet were then folded into a large arch. The parallelism of linear elements in the Bethlehem gneiss and the adjacent Littleton formation has been used as evidence that the Mt. Clough pluton was involved in the deformation and is therefore syntectonic (Billings, 1956).

4.3.10.4 Petrography

Mineralogy and Textural Relations. The Bethlehem gneiss is a dark-gray to light-gray, medium-grained biotite gneiss that is variably foliated. The texture is granoblastic, as a result of postconsolidation deformation. The principal minerals are quartz, 29%; oligoclase-andesine, 33%; and potassium feldspar, 19%; with a small amount of muscovite and biotite. White phenocrysts of microcline 3-20 mm (0.1-0.8 in.) long locally constitute several percent of the rock; in a few locales, they make up 10% of the rock. Minor to trace amounts of opaque oxides and sulfides, apatite, epidote, calcite, and garnet are also present (Billings, 1956).

Alteration. Chlorite, sericite, and epidote occur in very minor amounts as alteration products (Chapman, 1937).

4.3.10.5 Secondary Intrusions

There are several places where lamprophyres and other mafic rocks intrude into the Mt. Clough pluton.

4.3.10.6 Country Rocks

The Mt. Clough pluton intrudes into the Littleton formation of Early Devonian (Middle Silurian?) age. This formation consists of a gray, micaceous quartzite and a gray, coarse-grained mica schist, with such minerals as biotite, garnet, sillimanite, and locally andalusite (Billings, 1956).

4.3.10.7 Rock Structure

The Bethlehem gneiss is strongly foliated, more so near the margins of the pluton. The contacts are generally concordant with the bedding and foliation in the adjacent metasediments. The Mt. Clough pluton and the surrounding metasediments were folded into a large arch (Billings, 1956). The pluton is sandwiched between the Skitchewang and Fall Mountain nappes (Thompson et al., 1968).

Kruger (1946) reported that strike and cross joints are well developed in the Bethlehem gneiss of the Bellows Falls quadrangle. A plot of 200 joints showed that N. 25° E. joints are more common than N. 80° W. joints (see Fig. 4.8). Both types of joints are also found in the wall rocks.

4.3.10.8 Geophysics

Heat flow values of 1.17-1.36 x 10^{-6} cal/cm^2·s are reported for the Bethlehem gneiss (Jaupart et al., 1982).

4.3.11 Sebago Lake Batholith (Plate III, Pluton 15)

See Sec. 3.3.31 and Plate II, pluton 6. Although about 223 km^2 (86 mi^2) of this body lies in New Hampshire, a tenfold larger area appears in Maine, and it is therefore discussed with the Maine plutons.

4.3.12 Umbagog Granodiorite (Plate III, Pluton 13)

See Sec. 3.3.37 and Plate II, pluton 14. Most of this pluton lies in Maine; only about 35 km^2 (14 mi^2) occurs in New Hampshire, and it is therefore described with the plutons of Maine.

4.3.13 Weare and Spaulding Plutons (Plate III, Pluton 18)

4.3.13.1 Size, Shape, and Location

The Weare pluton and the plutons of the Spaulding series (Lyons, 1983) lie in an irregular north-to-south band

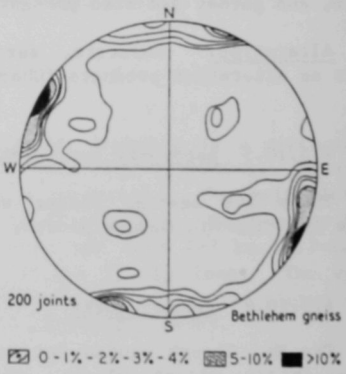


Fig. 4.8 Contour Diagrams of Poles of Joints in the Bethlehem Gneiss in the Bellows Falls Quadrangle (plotted on equal area net lower hemisphere) (Source: Kruger, 1946)

about 65 km (40 mi) long and from 3 to 20 km (1.9 to 12 mi) wide, with outcrops mainly in the Concord, Hillsboro, and Milford 15-min quadrangles; collectively, these plutons occupy a mapped area of about 641 km^2 (247 mi^2).

4.3.13.2 Relative and Absolute Age

These plutons belong to the New Hampshire plutonic series of Devonian age. Lyons and Livingston (1977) have dated the Kinsman quartz monzonite (which composes the fabric of the Weare pluton as well as the Cardigan and Lincoln plutons) at 411 ± 19 m.y. on the basis of Rb-Sr whole-rock data.

4.3.13.3 Origin

Acadian igneous activity in New Hampshire probably occurred in three distinct phases, from a source near the crust-mantle interface (Nielson et al., 1976). As described in Secs. 4.3.2 and 4.3.7 (for the Cardigan and Lincoln plutons, respectively), the Kinsman quartz monzonite represents an early intrusion of magma that was brought upward by density contrast and tectonic squeezing. The Weare pluton represents forceful magma injection during a cycle of recumbent nappe development. This event was followed by forcible emplacement of the bodies of the Spaulding series during the period of horizontal compression as the second phase of igneous intrusion. Magmas probably followed openings along earlier members of the series or kneaded their way through the infrastructure (Nielson et al., 1976). Both the Weare and the Spaulding plutons were probably injected as sheets under tectonic control.

4.3.13.4 Petrography

Mineralogy and Textural Relations. The Weare pluton is represented by a gray, medium- to coarse-grained phase of the Kinsman quartz monzonite. The pluton is porphyritic, with phenocrysts of feldspar constituting about 10% of the rock. Granoblastic and cataclastic features are present. The modal analysis indicates oligoclase-andesine, 23%; potassium feldspar, 30%; quartz, 29%; biotite, 14%; and muscovite, 4%. Quartz diorite and granodiorite are also present as compositional variants (Billings, 1956).

The plutons of the Spaulding series range in composition between quartz diorite and biotite quartz monzonite. The rocks are dark to light gray with well-developed, foliated, medium-grained texture. The principal minerals are oligoclase-andesine, 23-57%; orthoclase, 6-30%; quartz, 16%; biotite, 17-29%; and hornblende, 1%.

Alteration. The rocks show limited development of secondary minerals.

4.3.13.5 Secondary Intrusions

As with other plutons of the New Hampshire series, late-stage pegmatites are present in the Weare and Spaulding plutons (Billings, 1956).

4.3.13.6 Country Rocks

The plutons intruded into the gray micaceous quartzite member of the Littleton formation.

4.3.13.7 Rock Structure

According to Englund (1971), Acadian folding between the Cardigan and Winnipesaukee plutons occurred in three stages: the F_1 folding is recumbent and is mappable into the Hillsboro quadrangle, F_2 folding is oriented about northwesterly axes; and F_3 folding trends about northeasterly axes and is the dominant type. The Kinsman quartz monzonite, of which the Cardigan, Lincoln, and Weare plutons are composed, displays S_1 and S_3 foliation, implying initiation of deformation with recumbent folding, which was superimposed upon a northeasterly fabric (Nielson et al., 1976). As the Spaulding series is probably the late stage of the same igneous activity, the tectonic control may have imprinted similar syntectonic to late tectonic structural features, but of more variable degree, as in the adjacent plutons.

4.3.13.8 Geophysics

Nielson et al. (1976) have provided detailed gravity data for the area (see Fig. 4.4). On the basis of gravity and geologic data, they proposed that the Weare pluton was once coextensive with other Kinsman monzonite bodies, and that its present configuration is due to erosion. The Weare pluton has a maximum thickness of 1 km (0.6 mi).

Jaupart et al. (1982) reported high heat-flow values for the plutonic rocks in south-central New Hampshire.

4.3.14 Winnipesaukee Pluton (Plate III, Pluton 17)

4.3.14.1 Size, Shape, and Location

The Winnipesaukee pluton is an irregularly elliptical body located in east-central New Hampshire south of the White Mountain batholith. Its major and minor axes are approximately 39 km (24 mi) and 35 km (22 mi), respectively; its mapped area, measured by planimeter, is about 783 km^2 (302 mi^2).

4.3.14.2 Relative and Absolute Age

The Winnipesaukee pluton intrudes into the Littleton formation, making it younger than Early Devonian. Quinn (1944) believed this pluton was emplaced during Late Devonian time. Rubidium-strontium dating by Lyons and Livingston (1977) yielded an age of 402 ± 19 m.y. for the Kinsman quartz monzonite, a prominent component of the pluton.

4.3.14.3 Origin

A magmatic origin is assumed for the Kinsman quartz monzonite of the Winnipesaukee pluton (Billings, 1956; Englund, 1976), as it has clear intrusive contacts and xenoliths of the Littleton formation elongated parallel to its foliation and margin. Magma probably was emplaced by a combination of forceful injection and plastic stoping (Nielson et al., 1976). The magmas were channeled into or between country rocks being deformed by recumbent folding.

4.3.14.4 Petrography

Mineralogy and Textural Relations. The Winnipesaukee pluton is a composite body, consisting of the Meredith phase of the Kinsman quartz monzonite and the Winnipesaukee quartz diorite. The Meredith phase of the Kinsman quartz monzonite is rich in phenocrysts of potassium feldspar. The rock is dark gray, medium to coarse grained, and massive to foliated. The feldspar grains are mostly microcline, 2-4 cm (0.8-1.6 in.) long and commonly parallel. Many potassium feldspar grains have Carlsbad twins. The groundmass is quartz, oligoclase, biotite, and muscovite; accessory minerals are garnet and sillimanite. Billings (1956) classified this rock as a granite because potassium feldspar is more abundant than plagioclase.

The Winnipesaukee quartz diorite is dark to light gray, medium grained, and massive to well foliated. The principal dark mineral is biotite; the light-colored minerals, which predominate, are quartz and sodium plagioclase, either andesine or oligoclase (Billings, 1956). Quinn (1944) reported a grain size of 2-4 mm (0.08-0.16 in.), although phenocrysts as large as 1 cm (0.4 in.) in diameter are common. Although quartz diorite is the most abundant rock type, granodiorite, quartz monzonite, and even granite (where orthoclase is present) also occur.

4.3.14.5 Secondary Intrusions

The Winnipesaukee pluton was intruded into by the Ossippee Mountain, Red Hill, and Belknap Mountain complexes, of Mesozoic age.

4.3.14.6 Country Rocks

The Winnipesaukee pluton intrudes into and is underlain by the Littleton formation, which consists of schist and quartzite. Its texture, structure, and mineral composition led Quinn (1944) to interpret the Littleton as formed by high-grade metamorphism. These rocks were of sedimentary origin, having been composed mainly of shale, sandstone, shaly sandstone, and sandy shale.

4.3.14.7 Rock Structure

The Winnipesaukee quartz diorite and the Meredith phase of the Kinsman quartz monzonite are strongly foliated parallel to the bedding planes. Open to isoclinal folds with axes oriented northeasterly are present. Lyons and Clark (1971) proposed that Kinsman quartz monzonite of the Cardigan pluton was emplaced synchronously with the first-stage recumbent folding. However, the plots of foliation by Englund (1977) suggest that the pluton may have been emplaced in the latest deformational stage.

4.3.14.8 Geophysics

The gravity survey of New Hampshire by Nielson et al. (1976) includes a profile across the Winnepesaukee pluton that defines a maximum depth of 3500 m (11,450 ft). The gravity data of Englund (1976) for the Holderness quadrangle further confirm the shallowness of the plutonic rocks.

Jaupart et al. (1982) reported heat-flow values of 1.08-1.42 x 10^{-6} cal/cm^2·s for the complex.

4.4 ALLEGHENIAN AND POST-ALLEGHENIAN

4.4.1 Ossipee Mountains Pluton (Plate III, Pluton 16)

4.4.1.1 Size, Shape, and Location

The Ossipee Mountains pluton occupies a circular area roughly centered on longitude 71°15' W. and latitude 43°45' N. It is about 14 km (9 mi) in diameter and consists of three lithologies: (1) moat volcanics, (2) porphyritic quartz syenite, and (3) Conway granite. Two other lithologically related, roughly circular bodies, lying to the northwest and the southeast of the main Ossipee body, are included. The three exposures together occupy an area of about 234 km^2 (90 mi^2).

4.4.1.2 Relative and Absolute Age

The igneous rocks were intruded into Middle Devonian rocks of the New Hampshire plutonic series. The Conway granite has been dated by K-Ar methods at 121 m.y. (Foland and Faul, 1977).

4.4.1.3 Origin

First, a few thousand meters (feet) of volcanic flows and tuffs were extruded and deposited on an erosional surface consisting of mica schist, pegmatite, and granitic rocks of the New Hampshire plutonic series. Volcanic features and fault zones probably developed that were similar to those found at many modern calderas. Eruptions may have taken place along rifts as well. Faults then broke the volcanics into large blocks, which weakened the crust over the area. The volcanics began to drop down at an even faster rate when an arcuate fracture zone surrounding the mountains developed. This zone was produced by the intrusion from below of Albany porphyritic quartz syenite and Conway granite, and served as a path for further shallow volcanic intrusions and surface eruptions. Later, the Albany quartz syenite was intruded into, along this arcuate fracture zone, to form a ring dike. Finally, the central stock of Conway granite was emplaced by stoping (Wilson, 1969).

4.4.1.4 Petrography

Mineralogy and Textural Relations. The Albany porphyritic quartz syenite forms a ring-dike structure that completely circumscribes the intrusion. Hand specimens commonly contain a few scattered phenocrysts of anorthoclase that are mantled by a thin rim of potassium feldspar. Nonmantled phenocrysts of microperthite are abundant and average 7-9 mm (0.27-0.35 in.) in length. The rimmed phenocrysts of anorthoclase are a different color from the nonrimmed phenocrysts of microperthite. Where the microperthites are pale olive green, pink, or light gray, the mantled phenocrysts of anorthoclase are gray with cream colored rims, gray with white to pink rims, and dark gray with cream-colored rims. The rims of potassium feldspar are about one-tenth as thick as the grains of anorthoclase, which are commonly about 5 mm (0.2 in.) long. Quartz also is present as small equant phenocrysts that average 4 mm (0.16 in.) in diameter. These grains show sharp extinction and are free of strain shadows. Many minerals are present as small grains in the groundmass of the syenite. The light-colored minerals are apatite and microscopic quartz, which is equant and has grains 0.2 mm (0.08 in.) in diameter. Several dark-colored minerals occur in the groundmass: hastingsite, biotite (rare), augite, fayalite, and magnetite. Modal analysis yields quartz, 15-19%; microperthite, 59-66%; anorthoclase, 6-14%; and hastingsite, 7-10%; with trace amounts of augite, biotite, magnetite, zircon, apatite, and fayalite.

Alteration. Though the granites show ordinary surface weathering, no detailed alteration effects have been discussed in the literature.

4.4.1.5 Secondary Intrusions

The Conway granite was emplaced as a stock following intrusion of the Albany porhyritic quartz syenite. It is typically a medium- to coarse-grained, light-pink- to buff-colored, equigranular biotite granite. The grains of quartz and feldspar are 7-12 mm (0.27-0.47 in.) in diameter. Accessory minerals are hastingsite, fayalite, apatite, zircon, rutile, fluorite, allanite, and molybdenite. One-half to three-fourths of the alkali feldspar is microcline perthite, but the perthite contains coarse stringers and irregular patches of the sodic phase. About 5-20% of the feldspar is albite, which is present in individual laths twinned by the albite law. Some feldspar grains are faintly zoned. The quartz is generally a dark smoky color and occupies one-fourth to one-third of the rock. Modal analysis yields quartz, 31%; microperthite, 43%; albite, 20%; and biotite, 6%; with trace amounts of accessory minerals.

4.4.1.6 Country Rocks

The Ossipee Mountains pluton was intruded into igneous rocks of the New Hampshire plutonic series.

4.4.1.7 Rock Structure

The absence of prominent structural features, even of foliation or banding, is attributable to posttectonic intrusion of the pluton. Localized fracturing is present.

4.4.1.8 Geophysics

Sharp and Simmons (1978) presented detailed gravity and magnetic survey data for the Ossipee Mountains. The complex as a whole has an irregularly shaped bottom, extending to a depth of at least 4 km (2.5 mi). The eastern stock of Conway granite is, in places, only a few meters (feet) thick, and is probably underlain by an unexposed but much thicker stock of either diorite or gabbro.

4.4.2 Pilot-Pliny Complex (Including Percy Peaks and Gore Mountain) (Plate III, Pluton 11)

4.4.2.1 Size, Shape, and Location

The four roughly circular stocks of this complex have a linear northwesterly disposition. From north to south they are the Gore Mountain, Percy Peak, Pilot, and Pliny bodies. Collectively, they occupy an area of about 235 km^2 (91 mi^2). Because all four bodies are lithologically similar, only the Pilot-Pliny bodies are described in detail.

The Pilot-Pliny complex forms a subcircular shape, in plan view, that is about 21 km (13 mi) long and 16 km (10 mi) wide. Billings (1956) described the complex as consisting of two parts — the northern Pilot complex and the southern Pliny complex. The ring dikes and stocks of the Pilot complex are related to several centers. The ring dikes of the Pliny complex are concentric about a center 3 km (2 mi) northeast of the highest part of the range. The Cresent range ring dike is concentric about this center or one further to the east (Billings, 1956).

4.4.2.2 Relative and Absolute Age

The Pilot complex intrudes into the Albee formation of Late Ordovician age and the Whitefield gneiss of Ordovician age. The Pliny complex intrudes into the Whitefield gneiss. The Conway granite, which is part of the complex, has been dated at 181 m.y. (Foland and Faul, 1977).

4.4.2.3 Origin

The complex exhibits typical ring-dike structure, except that there are no subsided volcanics. Emplacement occurred along arcuate fractures, with syenitic rocks forming arcuate bodies on the edges of cores of granitic stocks (Chapman, 1976).

4.4.2.4 Petrography

Mineralogy and Textural Relations. Two bodies of riebeckite granite border the Pilot complex, one northeast and one southwest. The riebeckite granite is a white, medium-grained rock speckled with black, stubby crystals of riebeckite. Many specimens have a small amount of elongate sheaves (1-4 mm, or 0.04-0.16 in.) of golden yellow astrophyllite. The quartz is clear and vitreous. The texture is typically hypiodiomorphic granular. Microperthite is the principal feldspar. Modal analysis yields quartz, 32%; microperthite, 50%; plagioclase, 11%; riebeckite, 6%; and biotite, 1% (Billings, 1956).

One large body on the western edge of the Pilot complex and two small bodies to the north are composed of syenite. The large body is medium to coarse grained. Modal analysis yields quartz, 2%; microperthite, 73%; plagioclase, 15%; pyroxene, 2%; hastingsite, 7%; biotite, 1%; and small amounts of apatite (Billings, 1956). Of the two bodies in the north, one is 5 km (3 mi) long and the other is 1.6 km (1 mi) long: both are syenite porphyry. Phenocrysts 1-8 mm (0.04-0.31 in.) long are mostly orthoclase, but there is also some augite. Near the contacts the groundmass is very fine grained and black; elsewhere it is dark green and medium grained. The principal minerals are orthoclase, hornblende, and hedenbergite (Billings, 1956).

The Conway granite forms the largest body in the complex, which is about 16 km (10 mi) long and 13 km (8 mi) wide. It is a pink biotite granite, medium to coarse grained (Billings, 1956). In the coarse-grained phase, the pink crystals of microperthite range from 2 mm to 6 mm (0.08-0.24 in.) in length. The quartz grains are somewhat smaller. The biotite flakes average 2 mm (0.08 in.) in diameter (Billings, 1956). In the medium-grained phase, the feldspars average 2-3 mm (0.08-0.12 in.) in diameter, whereas the quartz averages about 2 mm (0.08 in.) in diameter (Billings, 1956). Conway granite also constitutes the Percy Peak stocks.

The Pliny complex is composed of quartz monzonite, hastingsite granite, and quartz syenite with a small plug of Conway granite. The quartz monzonite occurs on the north side of the Pliny complex. The rock is dark gray, equigranular, and hypidiomorphic. Modal analysis yields orthoclase, 28%; plagioclase, 36%; quartz, 9%; hornblende, 15%; and biotite, 12%; with trace amounts of magnetite, apatite, and sphene (R.W. Chapman, 1942).

The hastingsite granite forms a small circular body to the east. It is pink, fine grained to medium grained, and equigranular to subporphyritic (Billings, 1956). Modal analysis yields quartz, 24%; microperthite, 61%; plagioclase, 9%; hastingsite, 3%; and biotite, 3%; with trace amounts of opaque oxides and apatite (Billings, 1956).

The quartz syenite forms a ring-dike structure that encompasses the southern, eastern, and western portions of the complex. The rock is medium grained and usually equigranular; near contacts, however, it is locally subporphyritic or porphyritic (Billings, 1956). Modal analysis yields quartz, 7%; microperthite, 73%; plagioclase, 11%; pyroxene, 1%; hedenbergite, 7%; and biotite, 1% (Billings, 1956).

In Gore Mountain, on the western portion of the complex, two very small bodies to the south and southwest of the main body of granite are syenitic in composition. Modal analysis yields quartz, 2%; microperthite, 73%; plagioclase, 15%; pyroxene, 2%; and biotite, 1%; with trace amounts of apatite (Billings, 1956). The remainder of the stock is medium- to coarse-grained Conway granite.

Alteration. Secondary alteration is minimal.

4.4.2.5 Secondary Intrusions

There are mutually intrusive relations between the different phases of the rock bodies, along with late development of dikes.

4.4.2.6 Country Rocks

The Pilot complex intrudes into the Albee formation of Late Ordovician age. This formation comprises light-gray to white micaceous quartzite, gray quartz-mica schist, and gray mica schist that may contain porphyroblasts of biotite, garnet, or staurolite (Billings, 1956). Both the Pilot and the Pliny complexes intrude into the Whitefield gneiss, which is a dark-gray, medium-grained, foliated biotite granodiorite gneiss of Middle or Late Devonian age (Billings, 1956).

4.4.2.7 Rock Structure

The complex represents a group of ring-dike structures formed by cone fracturing or subsidence. Although each one is likely to be largely isotropic in structure, jointing is to be expected in such bodies, and joint fractures may be reactivated at intervals (Chapman, 1976). Hatch (1963) and Swift (1966) mapped crosscutting joints in the Gore Mountain stock.

4.4.2.8 Geophysics

Jaupart et al. (1982) reported a heat-flow value of 0.96×10^{-6} $cal/cm^2 \cdot s$ for the rocks of the area, but it should be higher for the intrusive granitic bodies.

4.4.3 White Mountain Batholith (and Belknap Mountain Ring-Dike Complex) (Plate III, Pluton 14)

4.4.3.1 Size, Shape, and Location

The White Mountain batholith, which is part of the White Mountain plutonic-volcanic series, is a composite batholith. Other than the large pluton of Conway granite, the batholith consists of several ring-dike complexes and stocks. Also included is the circular Belknap Mountain ring-dike complex lying to the south of the Winnipesaukee pluton. The total mapped area of the complex is about 1293 km^2 (499 mi^2).

4.4.3.2 Relative and Absolute Age

The White Mountain batholith intrudes into Early Devonian metasediments and Devonian intrusives of the New Hampshire plutonic series (Billings, 1934 and 1956). Radiometric dates of 170-195 m.y. (Creasy, 1974) indicate a Late Triassic to Early Jurassic age for the Mt. Osceola granite and Conway granite (Creasy, 1974). As indicated by Rb-Sr measurements on whole rock, igneous activity occurred in the western half of the batholith about 186-196 m.y. ago, whereas in the eastern half, it occurred about 175 m.y. (Eby and Creasy, 1983).

4.4.3.3 Origin

The batholith developed along deep-seated fracture zones (Billings and Williams, 1935; Billings, 1956; and Chapman, 1976). Magma collected at nodal points in a system of intersecting fracture zones, either in the upper mantle or at the base of the crust. Magma pockets were transformed into diapiric shapes, and ellipsoidal masses became detached from the reservoir rock. These discrete chambers of relatively light, superheated basaltic magma moved upward through the crust, aided by plastic flowage, melting, and stoping of adjacent rock (Chapman, 1976). Eby and Creasy (1983) proposed a lower crustal origin for the Mt. Osceola, Mt. Lafayette, and Conway granite bodies on the basis of low initial Sr and high Th-U isotopic values.

4.4.3.4 Petrography

Mineralogy and Textural Relations. The White Mountain batholith is a composite body composed of 13 lithologies. Recently Lyons (1983) expanded the list to 27 identified lithologic units. They range from syenites through granites. What follows is a petrographic description of ten major units from oldest to youngest, based on Creasy's (1974) study.

Franconia Notch Syenite. Two syenites belong to this designation -- a medium-grained syenite and a mafic syenite. The medium-grained syenite is light gray and composed of gray-white feldspar, biotite, and amphibole. A few tabular, anhedral crystals of feldspar create a subporphyritic texture. A modal analysis yields quartz, 5%; microperthite, 66%; plagioclase, 20%; ferrohastingsite, 5%; biotite, 3%; and opaque minerals, 1%.

The equigranular mafic syenite is composed of dark-gray and light-gray feldspar, abundant biotite, and some amphibole. A modal analysis yields microperthite, 64%; plagioclase, 20%; hastingsite, 4%; biotite, 7%; and chlorite, 5%.

Carrigan Porphyritic Syenite and Trachytic Porphyry. The Carrigan syenite porphyry is a medium-grained, gray rock composed mainly of gray-white feldspar and a subordinate amount of black amphibole. Set in this groundmass

are phenocrysts (5-10%) of chalky white to pink feldspar that are about 10 mm (0.4 in.) across. The rock contains 70% anhedral grains of orthoclase microperthite (of which 5% occurs as phenocrysts), 20% anhedral grains of common hornblende or hastingsite, and 8% quartz, as anhedral grains filling interstices between grains of feldspar and as granophyric intergrowths forming discrete grains and rims on the orthoclase microperthite. This pluton was emplaced about 196 m.y. ago, as indicated by Rb-Sr-whole rock data (Eby and Creasy, 1983).

The trachytic porphyry of Mt. Carrigan contains 25% phenocrysts (3-10 mm or 0.12-0.4 in.) of anhedral pink feldspar set in an aphanitic black groundmass. Phenocrysts of orthoclase microperthite ranging from 0.5 to 1.5 mm (0.02-0.06 in.) in size account for 25% of the rock and are set in a groundmass composed of "patchy," interlocking anhedral grains of feldspar.

Passaconaway Syenite. This rock is a medium- to coarse-grained, green, equigranular syenite composed mainly of microperthite. Ferrohastingsite is the chief mafic mineral, but some fayalite and ferrohedenbergite are present.

Pyroxene Syenite. This rock is massive, dense, black where fresh, and composed of fine- to medium-grained feldspar. Phenocrysts are greenish-gray, rounded feldspars grains 5-8 mm (0.2-0.31 in.) in diameter, which account for approximately 10% of the rock. Two different pyroxene syenites are described by Creasy. One is from Stony Brook in the North Conway quadrangle, with a modal analysis of quartz, 7.3%; orthoclase, 80.4%; fayalite, 0.1%; ferroaugite, 0.4%; ferropigeonite, 3.5%; ferrohastingsite, 6.5%; and iron-titanium oxides, 1.8%. The other is from Loon Mountain in the Franconia quadrangle, with a modal analysis of microperthite, 83.0%; fayalite, 2%; ferrohedenbergite, 10%; ferrohastingsite, 3%; and iron-titanium oxides, 2%.

Mt. Lafayette Granite Porphyry. This porphyry is a massive, fine-grained rock that is dark green to gray where fresh and light brown to gray where weathered. Quartz and feldspar phenocrysts occur in variable amounts. The quartz phenocrysts are smaller than 5 mm (0.2 in.) in diameter and are irregularly rounded and fractured. The feldspar phenocrysts are 1-4 mm (0.04-0.16 in.) in diameter and occur as subrectangular or fragmented crystals, which are green to gray in color and exhibit iridescence. Phenocrysts of fayalite and ferrohedenbergite are present in minor amounts (1-2%), and most are enclosed as reaction rims of ferrohastingsite. A modal analysis yields quartz phenocrysts, 5.1%; perthite phenocrysts, 47.2%; groundmass quartz, 12.1%, groundmass feldspar, 27.7%; plagioclase, 0.7%; fayalite, 0.3%; ferrohedenbergite, 0.4%; ferrohastingsite, 6.2%; and opaque minerals, 0.3%. Eby and Creasy (1983) put the time of emplacement for this phase at 193 m.y. on the basis of Rb-Sr whole-rock dating.

Mt. Garfield Porphyritic Quartz Syenite. Fresh specimens of the Mt. Garfield porphyritic quartz syenite are dark green, but the rock is weathered gray or pink in outcrop. Phenocrysts are subhedral to euhedral perthite, which are subordinate to angular to rounded amber quartz. Phenocrysts of

fayalite and ferrohedenbergite occur in minor amounts, surrounded by ferrohastingsite. The groundmass is composed of greenish feldspar, clear quartz, and discrete grains of amphibole. Creasy reports three groundmass textures: (1) allotriomorphic granular; (2) panallotriomorphic with granular quartz; and (3) panallotriomorphic with interstitial quartz and granophyre. Quartz phenocrysts are 2-4 mm (0.08-0.16 in.) long and are well rounded and fractured. Perthite and microperthite phenocrysts range from 1 mm to 15 mm (0.04-0.6 in.); cumulocrysts 10-20 mm (0.4-0.8 in.) across are common. Quartz is found in the groundmass as grains 0.1-0.2 mm (0.004-0.008 in.) across. The feldspar is anhedral and intergrown with the ferrohastingsite to form an interlocking panallotriomorphic texture.

Albany Porphyritic Quartz Syenite. In hand specimen, the Albany porphyritic quartz syenite consists of subhedral to anhedral rounded or rounded tabular phenocrysts of feldspar, and sparse, rounded phenocrysts of amber quartz set in a fine-grained groundmass. The groundmass is pale brown to gray (green where fresh) and speckled with black grains of ferrohastingsite. The percentage of feldspar phenocrysts is 25-50% and averages about 33% of the whole rock. A modal analysis yields quartz phenocrysts, 4%; groundmass quartz, 12%; microperthite phenocrysts, 35%; groundmass perthite, 29%; microcline, 1%; oligoclase, 5%; ferrohastingsite, 10%; annite, 3%; and iron-titanium oxides, 1%. The rock is holocrystalline with phenocrysts of feldspar and quartz set in a panallotriomorphic granular groundmass composed of microperthite, orthoclase, quartz, plagioclase, ferrohastingsite, and annite. Identical syenite forms an arcuate mass in the Belknap Complex.

Mt. Osceola Granite. This rock is a medium- to coarse-grained granite that is green where fresh but commonly weathers chalky white. The subhedral to anhedral grains of greenish feldspar form an interlocking network that encloses rounded grains of smoky quartz and anhedral grains of quartz and mafic minerals. The granular quartz and mafic minerals -- chiefly ferrohastingsite, amphibole, and subordinate annite -- are strongly concentrated and not randomly distributed throughout the hand specimen. Microperthite, quartz, and ferrohastingsite form a hypidiomorphic granular texture. Fayalite and ferrohedenbergite may occur as rounded grains, as inclusions within microperthite, and as cores inside grains of ferrohastingsite. A range of ten modal analyses yields quartz, 18-30%; microperthite, 50-70%; plagioclase, 0-10%; fayalite, trace-3%; ferrohedenbergite, trace-2%; amphibole, 2-3%; annite, 1-2%; iron-titanium oxides, 1%; and accessory minerals, 1%.

Hart Ledge and Mt. Tremont Syenites. The Hart Ledge syenite is coarse, blue-gray or green where fresh, though usually weathered buff or brown. Blocky feldspar and smaller black grains of amphibole account for 95% of the hand specimen. Quartz is sparse, except in miarolitic cavities. The Hart Ledge syenite has a panallotriomorphic, inequigranular texture of intergrown microperthite and amphibole. A modal analysis yields quartz, 2%; microperthite, 65%; plagioclase, 5%; anorthoclase, 15%; ferrohedenbergite, 1%;

ferrorichterite, 10%; and annite, 2%. The Mt. Tremont quartz syenite is medium grained, buff to white where fresh, though commonly weathered to limonite brown. The texture is made of interlocking, tabular, euhedral or subhedral crystals of microperthite averaging 4 mm (0.16 in.) in length, intergrown in a panhypidiomorphic inequigranular texture, with quartz and amphibole filling the interstices. A modal analysis yields quartz, 8-18%; microperthite, 73-84%; aegerine-augite, 0.5-1%; ferrorichterite, 5.5-7%; opaque minerals, 1%; and accessory minerals, 0.5%.

Riebeckite Granite. This granite is creamy-white in color and composed of anhedral or subhedral tabular crystals of white feldspar and rounded grains of clear quartz. Interstitial grains of black or dark blue-black amphibole and honey-brown astrophyllite are peppered throughout the rock. A modal analysis yields quartz, 20%; microperthite, 72%; riebeckite, 7%; astrophyllite, 0.5%; and accessory minerals, 0.5%. In thin section, a hypidiomorphic texture is exhibited by intergrown microperthite, quartz, and riebeckite. The anhedral to subhedral crystals of microperthite are 1-4 mm (0.04-0.16 in.) in size. Quartz forms equant grains with angular margins; it does not form irregular interstitial grains. Riebeckite forms angular interstitial grains, with some part of every grain in contact with quartz. The riebeckite contains numerous inclusions of fluorite, astrophyllite, sphene, zircon, and quartz.

Conway Granite. The Conway granite is a pink, inequigranular biotite granite consisting of tabular, euhedral crystals of pink alkali feldspar, anhedral interstitial grains of white plagioclase, and rounded grains of clear quartz. Books of biotite form interstitial anhedral crystals. Miarolitic cavities are common and are bounded by euhedral crystals of quartz and feldspar. The porphyritic variety contains large rounded phenocrysts of white feldspar 7-15 mm (0.28-0.6 in.) long and subhedral, subangular to rounded grains of clear quartz up to 5 mm (0.2 in.) long, set in a pale-brown and fine-grained, pasty groundmass of quartz and feldspar. Biotite is lightly peppered throughout this groundmass as discrete black grains. A modal analysis yields quartz, 27.4%; microperthite, 50.4%; orthoclase, 18.8%; and annite, 3.4%, with trace amounts of accessory minerals. The nonporphyritic Conway granite is composed of microperthite, oligoclase, quartz, and biotite intergrown in an inequigranular, panallotriomorphic texture. Allanite, sphene, zircon, and fluorite are accessory minerals. This granite is part of a central stock in the Belknap Complex that also includes several other phases. At the northwestern apex is the Gilford gabbro, which is a small intrusive mass (Modell, 1936). The groundmass is composed of an ophitic intergrowth of labradorite and augite crystals that are 1-3 mm (0.04-0.12 in.) long. Large ovoid crystals of hornblende, about 3 cm (1.2 in.) in diameter, enclose crystals of labradorite (Billings, 1956). The normative composition of this rock is orthoclase, 8.9%; albite, 22.0%; anorthite, 26.7%; nepheline, 1.4%; diopside, 13.45%; olivine, 10.5%; magnetite, 4.8%; ilmenite, 9.9%; and apatite, 1.7%. Eby and Creasy (1983) give a time of emplacement of 175 m.y.

The area mapped as diorite in the northern part of the complex is actually a breccia composed of angular fragments of diorite intruded into by biotite granite and syenite (Modell, 1936). The most common rock is a dark, fine-grained diorite with grains about 1 mm (0.04 in.) in diameter. There are also coarse-grained varieties of diorite where the hornblende crystals are sometimes over 2 cm (0.8 in.) long (Billings, 1956). The normative composition is quartz, 9.9%; orthoclase, 9.5%; albite, 42%; anorthite, 13.9%; diopside, 12.3%; hypersthene, 8.11%; ilmenite, 7.5%; magnetite, 3.5%; and apatite, 1.7%; with 0.7% accessory minerals (Modell, 1936).

Two separate bodies of monzonite exist in the complex -- the Ames and Gilmanton bodies. The Ames monzonite is a small body in the northwestern corner of the complex that impinges on Lake Winnipesaukee. The Gilmanton monzonite forms a large arcuate mass on the extreme southwestern area of the complex. The mineralogy of these bodies is so similar that one description suffices. Though texture varies (Billings, 1956), the most common variety is a medium-grained, equigranular to subporphyritic, light-gray to gray rock composed of quartz, 2%; microperthite, 17%; plagioclase, 65%; pyroxene, 4%; amphibole, 5%; biotite, 5%; opaque oxides, 1%; and trace amounts of opaque sulfides, plus 1% accessory minerals and alteration minerals (Billings, 1956).

The two bodies of syenite have been called the Belknap syenite by Modell (1936). The rock is typically coarse grained and green to greenish gray on fresh surfaces (Modell, 1936; Billings, 1956). The rock consists of microperthite and hornblende, with smaller amounts of oligoclase, sodium orthoclase, biotite, magnetite, goethite, apatite, zircon, quartz, and sphene. They are similar to the syenites of the main White Mountain body.

Alteration. Secondary alteration appears to be minimal, but is evidenced by the presence of chlorite and sericite.

4.4.3.5 Secondary Intrusions

The many lithological phases have mutually intrusive relations. Dikes are also present.

4.4.3.6 Country Rocks

The White Mountain batholith intrudes into both the Littleton formation of Early Devonian age and two units of the Middle Devonian New Hampshire plutonic series -- the Kinsman quartz monzonite and a binary granite. The Littleton formation consists of a gray, micaceous quartzite and a gray, coarse-grained mica schist (Billings, 1956).

4.4.3.7 Rock Structure

Ring dikes produced by subsidence and cone fracturing are the major structures that dominate the complex. Jointing is present and seemingly has been reactivated from time to time. Structural discontinuities are not very well described.

4.4.3.8 Geophysics

Analysis of gravity data shows that the density contrast between the granite of this batholith and the surrounding metamorphic rocks extends downward at least 1520 m (5000 ft) (Joyner, 1963). On the basis of gravity data, Sharp and Simmons (1978) interpreted the White Mountain batholith as a vertical cylinder of granitic rock.

Roy and Decker (1965) and Birch et al. (1968) reported high heat-flow values (up to 2.27×10^{-6} cal/cm^2·s) from the White Mountain batholith, which is consistent with the above-average level of radioactivity in these rocks. Also see Jaupart et al. (1982) for updated data.

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5 CRYSTALLINE ROCKS OF VERMONT

5.1 GRENVILLE AND AVALONIAN

5.1.1 Chester Dome (Plate IV, Pluton 3)

5.1.1.1 Size, Shape, and Location

The core of the Chester dome of southeastern Vermont is elliptical, almost 32 km (20 mi) long by 11 km (7 mi) wide, and covers approximately 317 km^2 (122 mi^2) as determined by planimeter. It consists predominantly of the Wilmington gneiss of the Mt. Holly complex. Also exposed in the Chester dome is the Readsboro member of the Cavendish formation and the Bull Hill gneiss.

5.1.1.2 Relative and Absolute Age

A Precambrian age is indicated (Rosenfeld, 1954 and 1968; Faul, 1963) for the gneisses of the Mt. Holly complex, which are unconformably overlain by members of the Cambrian(?) Cavendish formation.

5.1.1.3 Origin

Rosenfeld's (1968) description of the emplacement of the Chester dome is based on a detailed study of rotated garnets. An early deformational event in the sedimentary rocks of the Waits River formation took the form of a major subhorizontal flow in a west-southwest direction. Major recumbent folds are associated with this first event. Dome and cleavage-arch formation followed. The gneissic rocks of the dome represent metamorphosed sediments remobilized during the thermotectonic events affecting the area.

5.1.1.4 Petrography

Mineralogy and Textural Relations. The Mt. Holly complex of the Chester dome is described by Doll et al. (1961) as consisting primarily of a fine- to medium-grained biotite gneiss. The gneiss locally contains abundant muscovite, with chlorite being more common in the western areas. The rock varies widely: it can be locally massive and granitoid, schistose, fine grained, or compositionally layered. Marbles and quartzites are interbedded with gneisses. Also common and locally abundant are amphibolite and hornblende gneisses (Doll et al., 1961).

The quartzites are locally very massive, reaching thicknesses of 9 m (30 ft) or more. Micaceous quartzite and quartz-mica schist commonly contain

garnet or chlorite after garnet pseudomorphs. The schists are sometimes graphitic. The calcite and dolomite marbles are interbedded with medium- to coarse-grained calc-silicate granulite. Other minerals present in major proportion are phlogopite, actinolite, and diopside (Doll et al., 1961).

Alteration. The presence of chlorite, saussurite, and epidote indicates that secondary alteration affected the rocks of the area during subsequent metamorphic episodes.

5.1.1.5 Secondary Intrusions

The Chester dome was intruded into, on its northeastern side, by the 120-m.y.-old Mt. Ascutney complex. This complex is a composite feature, comprising granite, gabbro-diorite, and syenite stocks. It is roughly 9 km (6 mi) long and almost 5 km (3 mi) wide at its maximum width, with an area of about 23 km^2 (9 mi^2). Its map plan is roughly in the shape of a dumbbell. Occurring in the northwestern corner of the Claremont 15-min quadrangle, it also extends slightly westward into the northeastern corner of the Ludlow 15-min quadrangle. These stocks were emplaced in the Cretaceous period and are correlated with the White Mountain series.

There is evidence of Precambrian granite and pegmatitic dikes in the Mt. Holly Complex (Rosenfeld, 1954). They exhibit retrograde metamorphism, indicating emplacement before Paleozoic metamorphism. Although little information is available on these intrusives, they have been described as diklets several centimeters (inches) across to large bodies measured in square kilometers (square miles) (Rosenfeld, 1954).

5.1.1.6 Country Rocks

The Mt. Holly complex of the Chester dome is in contact with the metasediments of the Cavendish and Hoosac formations and the igneous rocks of the previously described Mt. Ascutney complex. The Wilmington gneiss of the Mt. Holly complex underlies the Bull Hill gneiss of the Cavendish formation, which is thought to be of Cambrian age. The gneiss contains a quartz-plagioclase-microcline-biotite assemblage characterized by locally abundant microcline augen up to several centimeters (inches) in length. Also present is a fine- to medium-grained quartz-plagioclase-biotite or biotite-muscovite gneiss (Doll et al., 1961).

5.1.1.7 Rock Structure

The Chester dome is the product of two very different deformational forces. Related to the earlier deformation are two large sigmoidal folds, the Ascutney and Star Hill sigmoids. Features previously thought to be anticlines and synclines have been reinterpreted as expressions of this type of folding (Rosenfeld, 1968).

5.1.1.8 Geophysics

With geological corrections, the Precambrian rocks of the Mt. Holly complex exhibit a heat flow of 1.13×10^{-6} cal/cm^2·s (Birch et al., 1968).

Gravity data define a northeasterly regional trend of gravity lows with Bouguer-anomaly contour values of -25 mgal to -35 mgal and unsymmetrical closure (Hildreth, 1979).

5.1.2 Mt. Holly Complex, Northern Exposure (Plate IV, Pluton 5)

5.1.2.1 Size, Shape, and Location

The northern part of the Mt. Holly complex is exposed in west-central Vermont in the Green Mountain anticlinorium. Outcrops of the complex are found in an elongated strip 35 km (22 mi) long by 1-3 km (0.5-2 mi) wide, joined near the waist to another parallel strip about 13 km (8 mi) long by 2-4 km (1-3 mi) wide. These strips cover an area of about 82 km^2 (32 mi^2). The rock is primarily gneiss (75%), with a large portion of schist and quartzite (20%). Minor amounts of amphibolite, greenstone, marble, and lime silicate rocks (4%) also are present. Only about 1% of the rocks in the complex are igneous (Brace, 1953). Many compositional and textural variations exist in the complex due to its large area, and it underlies many of the major structural features found in southern Vermont.

5.1.2.2 Relative and Absolute Age

No fossils have been found in the highly recrystallized rocks of the Mt. Holly complex. Because the complex is unconformably overlain by strata that are known to be Cambrian in age, it has been assigned to the Precambrian. Pegmatites cutting the Precambrian of Vermont have been dated at 1000 m.y. (Faul et al., 1963).

5.1.2.3 Origin

An igneous and sedimentary origin is suggested for the Mt. Holly complex. Interbedded marbles and quartzites attest to the sedimentary origin, while felsic and basic intrusions indicate that igneous processes also played an important role. The complex has undergone extreme recrystallization as a result of at least two metamorphic events (Osberg, 1942; Brace, 1953). MacFadyen (1956) stated that the complex underwent a Precambrian orogeny, during which the rocks were subjected to medium- to high-grade metamorphism. These rocks then underwent a second orogeny in the Ordovician, during which retrograde metamorphism altered them to a grade lower than that produced by the earlier event.

5.1.2.4 Petrography

Mineralogy and Textural Relations. Because of the diverse origins of the complex, its rock types and mineral assemblages tend to vary. One of the most common rock types in the complex is quartz-biotite-microcline gneiss (Osberg, 1952; MacFadyen, 1956). The dominant rock type in the Lincoln Mountain 15-min quadrangle is a quartz-microcline-albite-sericite gneiss. Other important assemblages include quartz-microcline-sericite gneiss, quartz-albite-sericite gneiss, and quartz-perthite-sericite gneiss. These rocks are usually a mottled gray, although some phases are pink (Cady et al., 1962).

To the southwest of the Lincoln Mountain quadrangle, near the town of Ripton, a quartz-feldspar-biotite granulite crops out. When fresh this rock is buff in color, but it weathers to a gray to green. Approximately two-thirds of the feldspar is microcline, the rest being plagioclase. Also commonly found in the area is a gray, finely divided quartz-feldspar-biotite-muscovite gneiss with layers 1-2 mm thick that contain rounded to augen-shaped quartz and feldspar grains. The thicker layers are separated from each other by thin layers of muscovite, biotite, and chlorite, with muscovite being more common (Osberg, 1952).

Alteration. Associations of sericite and epidote with feldspars, and of chlorite with mafic minerals (e.g., garnet) are evidence of retrogression of the primary minerals (Osberg, 1952).

5.1.2.5 Secondary Intrusions

The only bodies of igneous origin in the Lincoln Mountain quadrangle are pegmatite dikes, which occur predominantly in the gneiss and granulite phases of the Mt. Holly complex. The pegmatites commonly contain tourmaline (Cady et al., 1962). Osberg (1952) described the presence of albite porphyry dikes and ultramafic pods in the Green Mountain anticlinorium, with diabase dikes being the youngest phase.

5.1.2.6 Country Rocks

The northern exposure of the Mt. Holly complex is in contact with the Cambrian(?) Pinnacle formation to the west, the Cambrian(?) Hoosac formation to the east, and a small portion of the Cambrian(?) Underhill formation on the extreme northeastern tip of the exposure. The metasediments of the Pinnacle formation include a schistose graywacke that is gray to buff in color. The metasediments of the Hoosac formation contain a quartz-sericite-albite-biotite-chlorite schist, which is characterized by albite porphyroblasts. The predominant rock type of the Underhill formation is a silvery or gray-green quartz-sericite-albite-chlorite-biotite schist that contains many lenticular bodies of white quartz (Doll et al., 1961).

5.1.2.7 Rock Structure

Osberg (1952) and Brace (1953) described slip and fracture cleavage in the metamorphic rocks of the anticlinorium, along with rotational features displayed in garnets. Although Cady et al. (1962) indicated that granular rocks show fracture cleavage, they made no direct statement concerning the Mt. Holly complex.

The northern portion of the Mt. Holly complex contains two major anticlines. The Ripton anticline is a doubly plunging anticline that culminates near Ripton (Osberg, 1952). A longer, doubly plunging anticline occurs east of the Ripton anticline; this was described as the Bread Loaf anticline by Osberg (1952), and as the Lincoln anticline by Doll et al. (1961). Between the two anticlines is the Sugar Hill syncline (Osberg, 1952).

Strike joints are steeply dipping and trend N. 10° E. to N. 30° E.; their attitude is roughly parallel to the axes of the minor folds. Cross joints strike N. 70° W. to N. 85° W. and dip steeply north or south (Osberg, 1952). Steeply dipping foliations and plunging lineations are oriented with the major structures of the area (Osberg, 1952; Brace, 1953).

Schists of the Mt. Holly complex are exposed along the west limb of the Ripton anticline. Here, a series of sharp folds is overturned to the west. The structure of the eastern limb, which is composed of less-foliated, more-granular rocks, is more difficult to determine (Osberg, 1952). The Bread Loaf (Lincoln) anticline exposes its core of gneissic material for a greater distance as it plunges to the north and south. The Sugar Hill syncline is sharply overturned to the west and is made more complex by smaller folds that also are overturned to the west (Osberg, 1952).

5.1.2.8 Geophysics

Bean (1953) reported that a gravity high corresponds to the axis of the Green Mountains. This high is thought to be a northward continuation of the gravity high found east of the Blue Ridge gravity low in Appalachian areas further south.

5.1.3 Mt. Holly Complex, Southern Exposure (Plate IV, Pluton 2)

5.1.3.1 Size, Shape, and Location

The core of the Green Mountain anticlinorium extends approximately 115 km (71 mi) from southwestern to west-central Vermont in a banana-shaped band up to 20 km (12 mi) wide in some parts, with outcrops found over an area of about 1545 km^2 (596 mi^2). It is primarily composed of the metasediments of the Mt. Holly complex. In the type locality, these rocks consist of

fine-grained biotite gneiss, augen gneiss, garnet-quartz-muscovite schist, rusty-colored muscovite schist, amphibolite, quartzite, and saccharoidal marble (Osberg, 1952). These rocks are similar to the Grenville lithologies in New York and are probably of an igneous and sedimentary origin (Brace, 1953; MacFadyen, 1956).

5.1.3.2 Relative and Absolute Age

No fossils have been found in the highly recrystallized rocks of the Mt. Holly complex. However, it lies unconformably beneath less-metamorphosed rocks containing Early Cambrian fossils. On the basis of this relationship, the complex was assigned to the Precambrian by Doll (1961).

5.1.3.3 Origin

The rocks of the Mt. Holly complex consist of metamorphosed sedimentary and igneous rocks (MacFadyen, 1956). A sedimentary origin for the nonigneous rocks is indicated by the many interbedded quartzites and marbles.

5.1.3.4 Petrography

Mineralogy and Textural Relations. One of the most common rock types found in the Mt. Holly complex in the area around Rutland is a biotite-microcline gneiss that often occurs as large homogeneous masses but also as beds a meter or so (several feet) thick interbedded with garnet gneiss, quartzite, and limestone. The gneiss is gray and has bands or augen of pink feldspar. These bands are 3 to 13 mm (0.125 to 0.5 in.) thick and contain abundant clear quartz, some broken microcline grains, and sodic plagioclase. The bands are separated by thin layers of dark schist, which are composed of sericite, biotite, chlorite, and some epidote (Brace, 1953).

In the vicinity of the Equinox 15-min quadrangle, a unit composed of a quartz-biotite-microcline gneiss that is very similar to the common rock type in the Rutland area was observed. The unit in the Equinox area contained quartz and bands of pink microcline with biotite. A very similar unit was found in the Bennington quadrangle area (MacFadyen, 1956; Hewitt, 1961). MacFadyen noted that the gneiss in the Bennington area contained abundant blue quartz. At the southern end of the Green Mountain anticlinorium near the Massachusetts-Vermont border lies the Stamford granite gneiss. It is usually gray, coarse grained, and porphyritic. It contains large rectangular or rounded perthitic microcline phenocrysts, blue quartz, albite, microcline, biotite, and epidote (MacFadyen, 1956).

Alteration. At least two metamorphic events are recorded in the Mt. Holly complex. The first produced metamorphic minerals up to garnet in the affected argillaceous rocks. The second metamorphism, which probably was a

part of the regional metamorphism of Cambrian-Ordovician rocks in the area, retrograded those higher-grade minerals to a lower grade, producing chlorite, epidote, and sericite.

5.1.3.5 Secondary Intrusions

Pegmatitic material, commonly containing tourmaline and blue quartz, intrudes into the Mt. Holly complex. The Cuttingsville stock, which is composed primarily of nepheline syenite, pulaskite, biotite syenite, and essexite (Doll et al., 1961), was intruded into the Mt. Holly complex. On the basis of K-Ar analysis, this event was dated approximately 100 ± 2 m.y. ago (Armstrong and Stump, 1971).

5.1.3.6 Country Rocks

The Mt. Holly complex underlies the Mendon formation, a Lower Cambrian stratigraphic unit, over most of the west side of the Green Mountain anticlinorium. MacFadyen (1956) suggested that the Mendon formation is the stratigraphic equivalent of the Dalton formation of western Massachusetts. The metasediments include a schistose quartzite, which contains pebbles of blue quartz and feldspar (Doll et al., 1961). The complex also is in contact with a quartz-sericite-albite-biotite-chlorite schist of the Hoosac formation on the east side of the anticlinorium in the Rutland area.

5.1.3.7 Rock Structure

The Green Mountain anticlinorium is the major structural feature of the southern portion of the Mt. Holly complex. Several small anticlines also are present throughout the exposure. The anticlinorium is generally convex to the east and is overturned and locally overthrust to the west. (See the discussion in Sec. 5.1.2.7 of rock structures associated with the northern exposure.)

5.1.3.8 Geophysics

After making geological corrections, Birch et al. (1968) reported the heat flow in the central region of the southern exposure of the Mt. Holly complex to be approximately 1.22×10^{-6} cal/cm^2·s.

5.1.4 Sadawga and Ray Pond Domes (Plate IV, Pluton 1)

5.1.4.1 Size, Shape, and Location

The Sadawga and Ray Pond domes crop out in the Wilmington 15-min quadrangle in southern Vermont; the two domes compose an irregular, mulilobed shape roughly 24 km (15 mi) long by 12 km (7.5 mi) wide. Exposed in the core of these domes is the Wilmington gneiss. This gneiss is exposed within an area of roughly 317 km^2 (122 mi^2) in southern Vermont. It is a member of the Precambrian Mt. Holly complex, which has its best exposures to the west in the core of the Green Mountain anticlinorium (Skehan, 1961).

5.1.4.2 Relative and Absolute Age

The Wilmington gneiss is interpreted as being Precambrian in age for several reasons. A profound angular unconformity exists between the Mt. Holly complex and younger Precambrian metasediments (Skehan, 1961). These metasediments are in turn separated from overlying metasediments of known Cambrian age by another but less distinctive unconformity. Each of these breaks in the geologic record is interpreted as representing a significant hiatus.

5.1.4.3 Origin

The Sadawga and Ray Pond domes apparently were formed by the upward movement of indigenous rocks some time after the formation of the Green Mountain anticlinorium. Skehan (1961) states that the domes are the result of cascade or spruce-tree folding, as opposed to simple anticlinal or synclinal folding. The necessary forces probably resulted from the vertical movement of the core gneiss relative to the overlying mantle of less-metamorphosed sedimentary rocks. The shear sense in the dome indicates that it was formed by the movement of these upper beds relative to the lower beds toward the axis of an adjacent syncline. The higher-grade metamorphism in the core of the dome may be the result of the high thermal conductivity of the gneiss in the dome (Skehan, 1961).

5.1.4.4 Petrography

Mineralogy and Textural Relations. The medium- to coarse-grained Wilmington gneiss of the Mt. Holly complex is found in the Sadawga and Ray Pond domes. The majority of the rock is well-banded, slightly foliated microcline augen gneiss. Small amounts of schistose gneiss and medium-grained microcline gneiss also are present (Skehan, 1961).

The Wilmington gneiss is identical to the Stamford granite gneiss, except that it does not contain abundant blue quartz. Albite, microcline,

oligoclase, and biotite are relatively abundant in the porphyroblasts; quartz, muscovite, and magnetite are found in the groundmass. Accessory minerals include zircon, sphene, tourmaline, pyrite, and apatite (Skehan, 1961).

Alteration. Epidote-group minerals present in the groundmass of the rock are likely alteration minerals.

5.1.4.5 Country Rocks

The Wilmington gneiss is in contact with the metasedimentary rocks of the Hoosac formation, the Readsboro schist, and the Heartwellville schist. The Hoosac formation is Cambrian in age and consists of medium- to coarse-grained muscovite-chlorite-biotite-garnet-quartz schist and muscovite schist. The muscovite schists are divided into thin graphitic albite and nonalbite types.

The Precambrian Readsboro schist is a garnet-chlorite-biotite-quartz-albite augen schist and gneiss. It is rusty to gray in color and graphitic near its base. The Heartwellville schist, also of Precambrian age, consists of a graphite-muscovite-(chlorite)-garnet-quartz lensed schist that is black to gray. Also present is a dark-green chlorite-muscovite-(garnet)-quartz lensed schist (Skehan, 1961).

5.1.4.6 Rock Structure

Jointing is particularly well developed in the Wilmington gneiss. Quartz veins directly related to the joints are common in the gneiss, but also are found in the surrounding metasediments. These veins range in thickness from several centimeters to a few meters (a few inches to several feet), usually extending for no more than 9-15 m (30-50 ft). The veins tend to strike parallel to the axial traces of the nearby Wilmington anticline and dip 60-90° to the east (Skehan, 1961).

The Sadawga dome is one of the major structural features in southern Vermont, occurring on the east flank of the Green Mountain anticlinorium, which is locally overturned and overthrust to the west. Both the Ray Pond and Sadawga domes are close to the anticlinorium and complicate the structure of the area considerably (Skehan, 1961).

5.1.4.7 Geophysics

A gravity high is associated with the major axis of the Green Mountain anticlinorium over most of its length. At its southern end, the axis of this high drifts to the east of the anticlinorium axis (Diment, 1968).

5.2 TACONIC

5.2.1 Barnard Gneiss (Plate IV, Pluton 4)

5.2.1.1 Size, Shape, and Location

The Barnard gneiss, a part of the Barnard volcanic member, is a narrow belt about 375 km (214 mi) long and 2-4 km (1-2.5 mi) wide, stretching from central Vermont almost to the southern border of Massachusetts. In Vermont, its 250-km (155-mi) length surfaces primarily in the Randolph, Woodstock, Ludlow, Saxtons River, and Brattleboro quadrangles, within an area of about 210 km^2 (81 mi^2). It lies on the east limb of the Green Mountain anticlinorium and continues into an area of dome-and-arch structures in eastern Vermont, bending around the north end of the Chester Dome before continuing northward. Chang et al. (1965) estimated the Barnard gneiss as being 600-900 m (2000-3000 ft) thick. However, Ern (1963) estimated a thickness of almost 2000 m (6000 ft) in the Randolph quadrangle, and a thickening southward to 3000 m (8400 ft).

5.2.1.2 Relative and Absolute Age

The Barnard gneiss is stratigraphically equivalent to the Cram Hill formation and is included in the Missisquoi formation (Doll et al., 1961). On the basis of Cram Hill correlations with graptolitic Magog slates of southern Quebec, an age of Middle Ordovician is given to the formation (Ern, 1963).

5.2.1.3 Origin

The Barnard gneiss represents metamorphosed interbedded felsic and mafic volcanics. The included minor schist and phyllite represent metamorphosed sediments.

5.2.1.4 Petrography

Mineralogy and Textural Relations. According to Ern (1963), the Barnard gneiss unit contains biotite gneiss, hornblende gneiss (locally garnetiferous or calcareous), and amphibolite units with intercalated schists and greenstones.

Barnard gneiss is light gray to dark green, medium to coarse grained, well foliated, and locally sheared. The chief minerals are quartz, albite-oligoclase, biotite, and hornblende. Epidote, chlorite, muscovite, and calcite are accessory minerals. In some places the amount of calcite is sufficient to make it a major component. The modal composition of this

biotite gneiss is quartz, 17%; plagioclase, 31%; biotite, 29%; epidote, 9%; calcite, 6%; and opaque minerals, 2%. The Barnard unit lies in a garnet isograd, and almandine garnet is present in some places.

Alteration. Altered phenocrysts of albite and quartz are common relics in the Barnard gneiss (Ern, 1963; Chang et al., 1965). Possible amygdules of epidote are found in the amphibolites.

5.2.1.5 Secondary Intrusions

In the gaps between the boudins in the amphibolite bed, quartz and feldspar are the vein-filling materials, but no major secondary intrusive phases are reported.

5.2.1.6 Country Rocks

From west to east, the following stratigraphic units are in the vicinity of the Barnard gneiss: the Stowe formation, the Moretown member and in some places the Whetstone Hill member of the Missisquoi formation, the Barnard volcanic member and in some places the Shaw Mountain formation, the Northfield formation, and the Waits River formation (Ern, 1963; Chang et al., 1965). In the central section of the formation, the Barnard unit wraps around the Chester dome.

There is a transitional zone between the Barnard gneiss and the underlying Moretown formation (Ern, 1963). At the southern end, the transition zone is a few meters (feet) wide and at the northern end it is up to 30 m (100 ft) wide. In the chlorite-sericite-quartz schist unit of the Moretown member, there is an upward increase in the thin beds of ankerite-bearing, dark green chloritic amphibolite.

5.2.1.7 Rock Structure

Jointing is well developed, striking N. 30° E. and cutting across the lithologic boundaries. The lineation of prismatic crystals is well developed in the hornblendic rocks of the Barnard member. The lineation plunges north-northeast, parallel to the direction of extension and minor fold axes.

In Curtis Hollow, southeast of Bridgewater, a large anticlinal fold with minor tight folding is found in the Barnard member. The Barnard amphibolite beds are folded with both shears and boudinage locally developed. Boudins appear prominently in the thin amphibolite beds that lie between thicker, light-colored gneiss beds. Chang et al. (1965) found no evidence of crosscutting relations along the contacts between the light-colored amphibolites and the dark gneisses. Within the greenstones, augen of calcite are frequently found. Ern (1963) concluded that the development of

two generations of minor folds and of cleavage shows that there are two stages of deformation in the Randolph quadrangle.

5.2.1.8 Geophysics

Geologic and gravity contrasts (Diment et al., 1980) distinguish the Precambrian domes (the Berkshire-Green Mountain massif) in western Vermont from the Paleozoic metamorphic rocks of the Connecticut Valley-Gaspe synclinorium in the eastern part of Vermont. The Taconian Line separates the Middle Ordovician Missisquoi formation (including the Barnard gneiss) from the Silurian Shaw Mountain formation (Hatch, 1982) and may represent a major stratigraphic break. The Missisquoi formation continues into Massachusetts and Connecticut. The Barnard member may therefore represent an island arc on a continental fragment.

5.2.2 Lost Nation Group (Plate IV, Pluton 8)

5.2.2.1 Size, Shape, and Location

Very irregular in shape, the Lost Nation group is found in an area approximately 26 km (16 mi) long and 16 km (10 mi) wide at its widest point (measured perpendicular to its major axis). It occurs over roughly 179 km^2 (69 mi^2) of northern Vermont and New Hampshire. It crops out in the Guildhall and Whitefield 15-min quadrangles of Vermont and New Hampshire, and the Percy and Mount Washington 15-min quadrangles of New Hampshire.

5.2.2.2 Relative and Absolute Age

The Lost Nation group is in contact with the Albee formation of Ordovician age. The group is a member of the Highlandcroft series, which is associated with the Taconic orogeny (Johansson, 1963). Recent U-Th-Pb isotopic dating of the Highlandcroft series (Lyons, 1981) demonstrates that it has an age of 444-451 m.y., which is Middle to Upper Ordovician.

5.2.2.3 Petrography

Mineralogy and Textural Relations. The rocks of the Lost Nation group are massive and subporphyritic to porphyritic, and range in composition from diorite to quartz monzonite. They are typically subhedral and coarse to fine grained, and are composed principally of andesine. Other major minerals include quartz, biotite, chlorite, microcline, hornblende, and epidote. Accessory minerals include chlorite, sericite, sphene, pyrite, apatite, magnetite, rutile, and zircon. Phenocrysts of microcline, averaging about 1 cm (0.4 in.) in length, occur in the quartz monzonite (Johansson, 1963).

Alteration. Feldspar has been altered to sericite and saussurite; chlorite after biotite is another common alteration in this group, as is the filling of fractures with epidote. Sillimanite-grade metamorphism has affected the area, with its intensity increasing to the northwest (Johansson, 1963).

5.2.2.4 Secondary Intrusions

Many tongues, dikes, and sills are injected into the disturbed zone surrounding the Lost Nation group. These intrusions are very similar in composition to the main body. Aplite dikes are locally common in the Highlandcroft granodiorite; they are described as light-colored, anhedral, granular aplite dikes, usually 2.5-15 cm (1-6 in.) thick. These dikes seem to be similar in age to the pluton. Microfaults in andesine and numerous quartz veins, which average 10 cm (4 in.) thick, cut the granodiorite and aplite dikes. Also present are postdeformational mafic dikes, which cut all of the previously described rocks (Johansson, 1963).

5.2.2.5 Country Rocks

The contact between the Lost Nation group and the country rock takes the form of an intensely fractured and crushed zone; at many localities, it is metamorphosed to hornfels or is schistose (Johansson, 1963). The contact between the Albee formation and the Lost Nation group is not sharp and can best be described as a zone of contact metamorphorphism. Doll (1961) described the Albee formation as consisting of quartzite, phyllite, and schist. Among the various quartzites present in the Albee formation are a gray, massive variety, which is white when weathered, and a feldspathic type interbedded with slate. Feldspathic and quartzose argillaceous phyllites are interbedded with the quartzites.

The Lost Nation group is in contact with the Mesozoic Pilot-Pliny complex on its northwestern side, the metasediments of the Albee formation, and a major fault surface along its southeastern side. Across the fault lies the Ordovician Whitefield gneiss, which is a dark-gray, commonly foliated, medium-grained granodiorite gneiss (Billings, 1956; Doll et al., 1961).

5.2.2.6 Rock Structure

Although joints strike in all directions in the crystalline rocks of the Lost Nation group, two sets are especially prominent. These sets strike from N. 20° E. to N. 60° E. and from N. 25° W. to N. 60° W. They dip almost vertically and do not exhibit unusual mineralization. Limonite staining is common along the joint surfaces. Sericite and chlorite are common in the proximity of the joints. A few major faults are inferred from changes in metamorphic grade and stratigraphic continuity.

5.3 ACADIAN AND POST-ACADIAN

5.3.1 Averill Pluton (Plate IV, Pluton 13)

5.3.1.1 Size, Shape, and Location

The Averill pluton is roughly elliptical, with a major axis of about 23 km (14 mi) that trends roughly east-west on the United States-Canada border in northeastern Vermont. The pluton crops out in the northwestern corner of the Averill and the northeastern corner of the Island Pond 15-min quadrangles, as well as in a small area in Quebec. The granite is sporadically exposed over an area of 192 km^2 (74 mi^2). The area is covered by glacial overburden and is heavily forested.

5.3.1.2 Relative and Absolute Age

No radiometric age is available for this crystalline rock body. However, Billings (1956) correlated it to the White Mountain magma series and assigned it a Mississipian(?) age.

5.3.1.3 Origin

Myers (1964) proposed that the pluton represents a posttectonic intrusion, as evidenced by lithologic homogeneity and discordant contact relations. It may be related to the White Mountain magma igneous activity, due to its similarity to Conway granite and other posttectonic intrusives of similar age.

5.3.1.4 Petrography

Mineralogy and Textural Relations. The Averill pluton has no internal structure and tends to be rather homogeneous. It is a two-mica granite that is gray to pink, medium to coarse grained, and composed of quartz, 31-43%; biotite, 2-7%; muscovite, 1-5%; potassium feldspar, 44-48%; and oligoclase, 0.5-15%. Accessory minerals include apatite, pyrite, muscovite, and zircon. Myers (1964) described the granite's texture as hypidiomorphic granular and locally subporphyritic because of large potassium feldspar grains. Microcline phenocrysts occur as Carlsbad twins.

Alteration. Sericitization of potassium feldspars is common. Chlorite after biotite is also observed (Myers, 1964).

5.3.1.5 Secondary Intrusions

Myers (1964) noted that pegmatites are common in the Averill granite, generally striking N. 20° E., which is parallel to one of the joint sets. These pegmatites have approximately the same composition as the granite, but contain more muscovite. Two generations of pegmatites appear to be present.

5.3.1.6 Country Rocks

The country rock is a gray, muscovite-quartz phyllite or schist. Interbedded and gradational with this rock, on the eastern edge of the pluton, is a gray, micaceous quartzite and graywacke.

5.3.1.7 Rock Structure

Although one set of joints in the Averill granite was observed to strike N. 20° E. (Myers, 1964), most of the joints in the vicinity of the Averill lakes were observed to trend N. 40° W. Joints in the Island Pond quadrangle tend to strike in all directions; the majority of the joints has a dip greater than 75° (Goodwin, 1963). There is a fault near Averill Mountain, where a down-dropped block of metasediments has been preserved in the pluton. Slickensides are often observed in the Averill granite along the fault plane (Myers, 1964).

5.3.2 Echo Pond Pluton (Plate IV, Pluton 12)

5.3.2.1 Size, Shape, and Location

The Echo Pond complex lies predominantly in the Island Pond 15-min quadrangle, but extends slightly in a westward direction into the Memphremagog 15-min quadrangle. It is irregular in shape. Its longest dimension is oriented in a southeasterly direction; this axis is roughly 18 km (11 mi) long. The complex has a maximum width of about 10 km (6 mi), and an area of about 78 km^2 (30 mi^2). The southern end of the body joins with the westernmost part of the Nulhegan pluton.

5.3.2.2 Relative and Absolute Age

No absolute age data are available on the Echo Pond complex. Doll et al. (1961) assigned it to the New Hampshire plutonic series, which makes it Devonian in age. The complex intrudes into the metasediments of the Lower Devonian Gile Mountain formation.

5.3.2.3 Origin

Page (1968) included the Echo Pond pluton in a group of plutons thought to have been emplaced by forceful injection and magmatic stoping. These intrusives are described as sills or cylindrical stocks.

5.3.2.4 Petrography

Mineralogy and Textural Relations. The Echo Pond complex contains a variety of igneous rocks. Although true granite is the predominant rock type, lesser amounts of monzonite, quartz monzonite, diorite, and gabbro also are present. The granite phase is dark to light gray, medium to coarse grained, and usually porphyritic. Large euhedral phenocrysts of potassium feldspar occur in the southern half of the complex. Major minerals include quartz, potassium feldspar, oligoclase, and biotite; lesser amounts of muscovite also occur. The quartz occurs interstitially and as small dispersed grains. Accessory minerals include apatite, zircon, sulfide minerals, and sphene (Goodwin, 1963).

The monzonite phases of the complex occur only near the metasediments along the southern edge of the complex. Major minerals include plagioclase (andesine), hornblende, and some quartz. Accessory minerals include magnetite, sphene, apatite, epidote, and zircon. Phenocrysts of euhedral to subhedral plagioclase can be found in the monzonite. Quartz occurs interstitially as small patches when viewed in thin section (Goodwin, 1963).

The fine-grained diorite contains many euhedral to subhedral plagioclase phenocrysts. Other major minerals are plagioclase feldspar and hornblende. Quartz is a minor constituent. Sphene, apatite, and magnetite are the most common accessory minerals (Goodwin, 1963).

The gabbro is coarse grained and dark gray to greenish black. It is massive and characterized by deep weathering. The relatively abundant minerals are labradorite, andesine, potassium feldspar, hornblende, and iron minerals, while apatite is a common accessory mineral. Very little quartz is present (Goodwin, 1963).

Alteration. Much of the hornblende in the complex has been altered to biotite and then altered to chlorite. The feldspars commonly show high degrees of sericitization (Goodwin, 1963).

5.3.2.5 Secondary Intrusions

Goodwin (1963) stated that dikes are not as numerous in the Echo Pond complex as they are in other plutons in the area, including the Averill granite. More than half of the dikes lie in the southwestern half of the complex and consist of aplite, fine-grained granite, or pegmatite.

5.3.2.6 Country Rocks

The Echo Pond pluton intrudes into the metasediments of the Gile Mountain formation, which is a gray, quartz-muscovite schist. A gray, micaceous quartzite and a calcareous, micaceous schist are interbedded and intergradational with the quartz-muscovite schist. The quartz-muscovite schist commonly contains porphyroblasts of biotite, garnet, and staurolite and, in the immediate vicinity of the pluton, sillimanite (Doll et al., 1961; Goodwin, 1963; Myers, 1964; and Albee, 1968).

5.3.2.7 Rock Structure

Joints are very common in the Echo Pond complex. Longitudinal joints strike N. 45° W. and dip steeply. A set of cross joints strikes N. 35° E. and has more gentle dips (Goodwin, 1963).

The monzonite and diorite commonly exhibit banding, which was interpreted by Goodwin (1963) as flow planes. This interpretation was based on similar dimensional orientation of inclusions, orientation of the bands, and strike of the contact (Goodwin, 1963).

5.3.3 Knox Mountain Pluton (Plate IV, Pluton 6)

5.3.3.1 Size, Shape, and Location

The Knox Mountain pluton is located in the East Barre, Plainfield, and St. Johnsbury 15-min quadrangles in east-central Vermont. It is elongated, with a roughly north-south orientation (see Fig. 5.1). Outcrops occur within an area of approximately 223 km^2 (86 mi^2). Murthy (1957) raised the question whether the Knox Mountain and Barre plutons are separate exposures of the same mass. He performed a qualitative spectrographic analysis of the two granites and found them to be very similar in trace element composition, but found no direct evidence of a relationship between the two. Although the Knox Mountain granite closely resembles the Barre granite, it is on the average lighter in color. The Knox Mountain granite is not as homogeneous in color and composition as the Barre granite.

5.3.3.2 Relative and Absolute Age

Murthy (1957), Hall (1959a and b), and Konig (1961) each found the Knox Mountain pluton to be Devonian in age. Murthy (1957) published a Rb-Sr age of 330 ± 25 m.y. on a biotite from the Barre granite, which was believed by Murthy (1957) and Konig (1961) to be contemporaneous with the Knox Mountain granite. However, the 380 m.y. Rb-Sr mineral isochron age established by

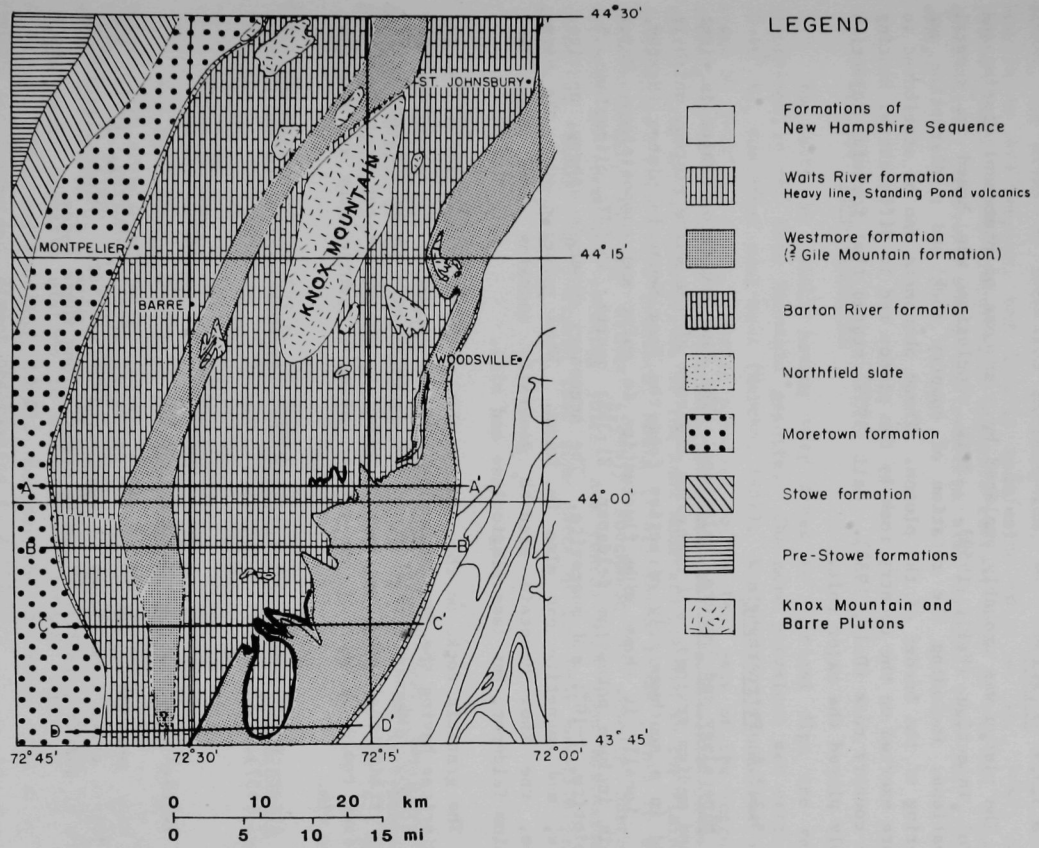

Fig. 5.1 Geologic Relations in the East Barre Area (Source: Murthy, 1957)

Naylor (1971) is regarded as a much more accurate indicator of the true age of the Barre (and Knox?) granites.

5.3.3.3 Origin

The pluton was probably emplaced by a process of piecemeal stoping and forceful injection (Hall, 1959). This conclusion was based on several observations, including the rotation of country rock and a disruption and shattering of the border of the pluton. These phenomena can be attributed to pressure exerted on the country rock by the pluton and to differential heating of the country rock (Daly, 1933). Hall (1959) stated that forceful injection probably played the major role.

5.3.3.4 Petrography

Mineralogy and Textural Relations. The Knox Mountain granite is light gray and medium grained, and resembles the rock of the Barre pluton, which is exposed in a southwesterly direction from the Knox Mountain pluton (Murthy, 1957). Locally it shows wide fluctuation in color and composition. Major minerals include potassium feldspar, 21-35%; quartz, 45-53%; oligoclase, 5-17%; biotite, 2-10%; and muscovite. The accessory minerals sphene, apatite, epidote, and magnetite can also be found. When compared with the Barre granite, the Knox Mountain granite generally contains more quartz and potassium feldspar, but less plagioclase and mica.

The granitic rocks on the eastern side of the Knox Mountain pluton are granodiorites having the same mineralogical composition as the Barre "granite" (Hall, 1959). (Maynard [1934] found the Barre granite to be a granodiorite, but the traditional name has been retained.) The Barre and Knox Mountain intrusives could be part of one pluton having a granite core and a granodiorite rim.

Alteration. Secondary alteration minerals include chlorite and epidote (Murthy, 1957).

5.3.3.5 Secondary Intrusions

Many pegmatites and quartz veins occur locally (Murthy, 1957), and cross joints often form planes of weakness (Konig, 1961). Mineralization in the form of quartz pegmatite dikes is common along these zones.

5.3.3.6 Country Rocks

The Knox Mountain granite lies primarily in a gray, quartzose, and micaceous crystalline limestone, which is interbedded and gradational with

gray quartz-muscovite phyllite or schist. The limestone weathers to form a distinctive brown, earthy crust. Closer to the pluton and in areas of increased metamorphism, the limestone contains minerals such as actinolite, hornblende, zoisite, diopside, wollastonite, and garnet. The phyllite and schist, in areas of increased metamorphism, may contain garnet, biotite, kyanite or sillimanite, and locally andalusite (Doll et al., 1961; Albee, 1968). Hall (1959b) reported a migmatic contact zone in the St. Jonesbury quadrangle. However, where the contacts are visible, they are generally sharp.

5.3.3.7 Rock Structure

Two sets of joints having very steep to vertical dips are readily recognized in the Knox Mountain granite. The more prevalent set strikes N. 20-40° E., the other east-west (Murthy, 1957). Konig (1961) found that joints trend N. 5-25° E. (i.e., roughly parallel to the long axis of the intrusion) and dip 60-90° either to the northwest or southeast. Cross joints are readily visible from aerial photographs of the Knox Mountain granite in the Plainfield quadrangle. From field evidence, Konig (1961) found these joints to strike west-northwest, dipping either steeply southwest or steeply northeast. Surfaces of northeast striking joints are coated with iron hydroxide and chlorite. Konig (1961) stated that some parts of the Knox Mountain granite exhibit primary foliation.

5.3.3.8 Geophysics

Diment (1968) reported a gravity high of 20 mgal, representing a density contrast of 0.1 g/cm^3 with the surrounding rocks. He interpreted the pluton to conform to a truncated cone extending to a depth of 7 km (4.5 mi).

5.3.4 Maidstone Pluton (Plate IV, Pluton 9)

5.3.4.1. Size, Shape, and Location

The Maidstone pluton lies predominantly in the Guildhall 15-min quadrangle, but extends slightly into the Burke 15-min quadrangle. It is highly irregular in shape; outcrops occur over an area of approximately 109 km^2 (42 mi^2). The pluton is a member of the New Hampshire plutonic series and is one of the large granitic plutons in northeastern Vermont (Doll et al., 1961; Johansson, 1963; and Woodland, 1965). Because the pluton is covered by heavy forests and overburden, direct observation is difficult.

5.3.4.2 Relative and Absolute Age

No absolute age data are available. Doll et al. (1961) assigned the pluton to the New Hampshire plutonic series, making it Devonian in age. It intrudes into the metasediments of the Lower Devonian Gile Mountain formation and the Ordovician Albee formation.

5.3.4.3 Origin

Page (1968) included the Maidstone pluton in a group of plutons thought to have been emplaced by forceful injection and stoping. These intrusions are described as sills or as cylindrical stocks.

5.3.4.4 Petrography

Mineralogy and Textural Relations. The Maidstone pluton contains several hypidiomorphic granular granites. A large percentage of the rock falls between a true granite and a quartz monzonite. Although most of the granitic rocks in the Maidstone pluton are medium to coarse grained and massive, they are locally subporphyritic. Their colors range from near white to dark gray and are locally pink to salmon. The majority of the rock is binary granite, but there are a few areas where muscovite is altogether lacking (Goodwin, 1963). Modal analysis yields quartz, 28.8%; microcline, 25.1%; plagioclase, 25.1%; orthoclase, 17.8%; chlorite, 2.9%; muscovite, 3.6%; and biotite, 1.7%. Accessory minerals include apatite, sphene, pyrite, and magnetite (1.2%).

Alteration. Chlorite after biotite is a common alteration, as is sericite after feldspar. This latter example of alteration can be observed along cleavages in the feldspar (Johannson, 1963).

5.3.4.5 Secondary Intrusions

Pegmatitic dikes are rare in the body of the pluton, but some occur near the margins of the body. Microcline is the most abundant feldspar, usually occurring with plagioclase, albite, and smoky quartz. Muscovite is present as small books, and an opaque beryl can be found locally as small, yellow crystals. Black tourmaline also is present (Johansson, 1963).

5.3.4.6 Country Rocks

The Maidstone pluton intrudes into the metasediments of the Gile Mountain formation, which is a gray, quartz-muscovite schist. A gray, micaceous quartzite and a calcareous, micaceous schist are interbedded and intergradational with the quartz-muscovite schist. The quartz-muscovite

schist commonly contains porphyroblasts of biotite, garnet, and staurolite and, in the immediate vicinity of the pluton, sillimanite (Doll et al., 1961; Johannson, 1963).

The Maidstone pluton also contacts, at its southeastern side, the metasediments of the Ordovician Albee formation. The Albee formation consists of a massive gray quartzite and a feldspathic quartzite, which weathers white. The quartzites are interbedded with greenish-gray slate, phyllite, feldspathic phyllite, and quartzose argillaceous phyllite. The Albee formation is metamorphosed in the immediate vicinity of the pluton to micaceous quartzite, quartz-mica schist, mica schist, and hornfels containing sillimanite porphyroblasts (Doll et al., 1961; Johannson, 1963).

5.3.4.7 Rock Structure

Two sets of joints are especially conspicuous in the granite of the Maidstone pluton. These sets trend from N. 20° E. to N. 60° E. and from N. 25° W. to N. 60° W. Another set that is almost horizontal also occurs in the area (Johannson, 1963). The rocks of the Gile Mountain and Albee formations are tightly folded (Johannson, 1963). The Monroe Line, a major structural feature, is intruded by the Maidstone and Victory plutons (Lyons, 1983).

5.3.5 Nulhegan Pluton (Plate IV, Pluton 11)

5.3.5.1 Size, Shape, and Location

The Nulhegan pluton is roughly circular in shape at the surface and has an average diameter of about 15 km (9 mi) (see Fig. 5.2). It crops out over an area of approximately 154 km^2 (59 mi^2), about three-fifths of which is in the Island Pond 15-min quadrangle. Most of the remainder of the pluton is exposed in the Averill 15-min quadrangle to the east; a very small portion of the pluton is exposed in the Burke 15-min quadrangle. The pluton is expressed topographically as a structural basin, the rim of which consists of the metasediments of the Gile Mountain formation (Goodwin, 1963).

5.3.5.2 Relative and Absolute Age

No absolute age data are available on this pluton. Doll et al. (1961) assigned it to the New Hampshire plutonic series, making it Devonian in age. The pluton intrudes into the metasediments of the Lower Devonian Gile Mountain formation.

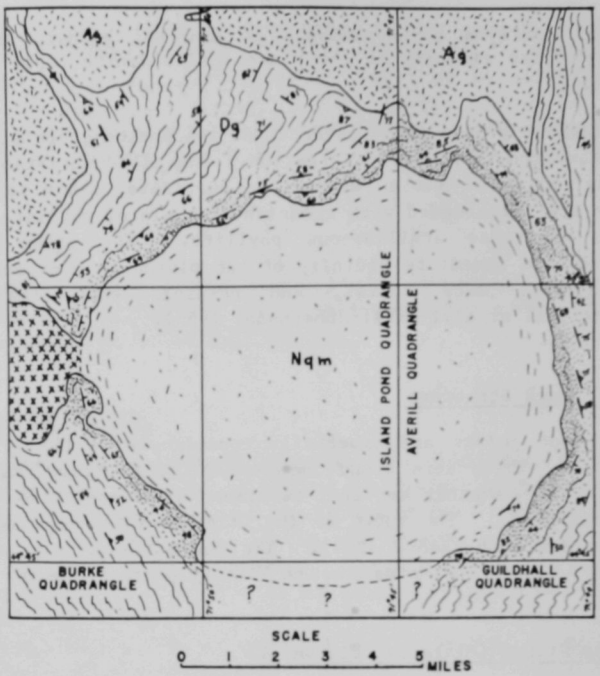

Fig. 5.2 Geologic Relationships of the Nulhegan Pluton (Source: Myers, 1964)

5.3.5.3 Origin

Page (1968) included the Nulhegan pluton in a group of plutons thought to have been emplaced by forceful injection and stoping. These intrusives are described as sills or as cylindrical stocks.

5.3.5.4 Petrography

Mineralogy and Textural Relations. The Nulhegan pluton is composed of coarse- and fine-grained quartz monzonite. The coarse-grained variety is similar to granodiorite in mineral composition, while the medium- to fine-grained quartz monzonite is more granitic in composition. Both varieties are light to dark gray, depending on the amount of mafic constituents.

In thin section, the Nulhegan quartz monzonite has a hypidiomorphic granular texture. Major minerals include euhedral to subhedral grains of brown and green biotite (13-27%), plagioclase (oligoclase and andesine), and potassium feldspar. Quartz grains (15%) are interstitial and do not have a

crystalline form. Hornblende occurs as subhedral grains and is usually replaced by chlorite. Accessory minerals include sphene, apatite, zircon, magnetite, and pyrite (Goodwin, 1963; Myers, 1964).

Alteration. Chlorite has replaced some of the hornblende in the quartz monzonite (Myers, 1964). An altered zone surrounds many of the xenoliths, which are presumed to be from the Gile Mountain formation. This zone takes the form of a band in the quartz monzonite, which usually is brown (Goodwin, 1963).

5.3.5.5 Secondary Intrusions

The Nulhegan pluton is cut by many small granitic dikes oriented parallel to a major set of joints that trend in a northwesterly direction. Pegmatitic dikes are common near the edge of the intrusion. They are uncommon more than 0.8 km (0.5 mi) from the contact. Major minerals in the pegmatites include potassium feldspar, quartz, muscovite, black tourmaline, and almandite. Several mafic dikes also are present; these are thought to be the youngest rocks in the pluton (Goodwin, 1963; Myers, 1964).

5.3.5.6 Country Rocks

The Nulhegan pluton intrudes into the metasediments of the Gile Mountain formation, which is a gray, quartz-muscovite schist. A gray, micaceous quartzite and a calcareous, micaceous schist are interbedded and intergradational with the quartz-muscovite schist. The quartz muscovite schist commonly contains porphyroblasts of biotite, garnet, and staurolite; in the immediate vicinity of the pluton, it contains sillimanite (Doll et al., 1961; Goodwin, 1963; Myers, 1964; and Albee, 1968).

5.3.5.7 Rock Structure

A major set of joints strikes N. 56° E. and dips approximately 90°. Many of the dikes in the pluton have strikes parallel to this direction. A second set of dikes appears to strike N. 27° W. These dikes are cut and displaced by the former set. The second set of dikes could represent another set of joints in the pluton.

The quartz monzonite exhibits faint banding, which was interpreted by Goodwin (1963) as flow banding. It trends about N. 45° W., which is parallel to the long axis of a majority of the xenoliths near the contact. The banding, which also tends to have a parallel to subparallel relationship with the contact, is absent in the interior of the pluton (Goodwin, 1963).

5.3.6 Victory Pluton (Plate IV, Pluton 7)

5.3.6.1 Size, Shape, and Location

The Victory pluton crops out over an area of roughly 78 km^2 (30 mi^2), lying almost entirely in the southeastern corner of the Burke 15-min quadrangle. It is very irregular in shape but can be described as a disturbed linear body trending northeast, with a roughly circular mass joined to its eastern side. Much of the southern part of the pluton is covered by glacial debris and forests (Woodland, 1965).

5.3.6.2 Relative and Absolute Age

This pluton belongs to the New Hampshire plutonic series. It intrudes into the metasediments of the Lower Devonian Gile Mountain formation and the Ordovician Albee formation (Doll et al., 1961).

5.3.6.3 Origin

Page (1968) included the Victory pluton in a group of plutons thought to have been emplaced by forceful injection and stoping. These plutons are described as taking the form of sills or of cylindrical stocks.

5.3.6.4 Petrography

Mineralogy and Textural Relations. The western lobe of the Victory pluton consists of a fine- to (rarely) medium-grained, hypidiomorphic granular rock that is occasionally porphyritic (Woodland, 1965). Major minerals of the biotite or biotite-muscovite quartz monzonite include oligoclase, orthoclase, microcline, plagioclase, quartz, biotite, and muscovite. Accessory minerals include apatite, zircon, tourmaline, and sphene.

A medium-grained melanocratic phase of oligoclase or oligoclase-andesine occurs locally. This rock contains lesser amounts of apatite, microcline, amphibole, orthoclase, biotite, hornblende, and quartz. Accessory minerals include tourmaline, sphene, apatite, and zircon. This phase may represent a separate intrusion rather than an altered or contaminated zone (Woodland, 1965).

In the northern part of the western lobe of the Victory pluton (East Haven Mountain area), a medium-grained biotite granite is exposed. It contains microcline, orthoclase, antiperthite, oligoclase, biotite, quartz, and myrmekite; apatite occurs as an accessory mineral.

Alteration. Oligoclase in the granite of the Kirby-East Haven Mountain area is kaolinized and sericitized. Biotite in the East Haven Mountain area

and in the Miles Mountain area is partially altered to chlorite. Oligoclase in the Miles Mountain area is partially altered to kaolin.

5.3.6.5 Secondary Intrusions

Many granitoid dikes are observed on Miles Mountain, which is located near the southern part of the eastern lobe. These dikes range in composition from a light-gray granite to a dark-gray tonalite (Woodland, 1965).

5.3.6.6 Country Rocks

On the western side of the Victory pluton is an extensive granite-hornfels complex that extends northward from the area around Kirby Mountain, following the margin of the pluton, to the Burke and Umpire Mountains area. The contact between these two phases is usually sharp, even though locally intricate. The rocks of this complex are extremely durable and tend to form many of the topographic highs in the area. Although the hornfels is often of sillimanite grade, it has retained much of its original bedding, schistosity, and cleavage. It commonly has numerous granite stringers.

Pods of a dark, biotite-rich rock lacking internal structure can be found locally. This rock is distinct from the granite of the Victory pluton, even though it is equigranular, fine grained, and somewhat granitic in appearance. The contact between this rock and the granite is less sharp than that between the hornfels and granite phases. A more prevalent migmatite, which also occurs in the hornfels-granite complex, is an intimate mixture of granitic stringers, highly metamorphosed hornfels, and the dark, structureless, biotite-rich rock previously mentioned.

The Victory pluton intrudes into the metasediments of the Gile Mountain formation of Early Devonian age. The formation has been described as a gray quartz-muscovite schist. A gray, micaceous quartzite and a calcareous, micaceous schist also are interbedded with and intergradational with the quartz-muscovite schist. The schist commonly contains porphyroblasts of biotite, garnet, staurolite, and, in the vicinity of the Victory pluton, sillimanite (Woodland, 1965).

The pluton also cuts the Meetinghouse Slate member of the Gile Mountain formation and the metasediments of the Ordovician Albee formation. The Albee formation consists of massive gray quartzite and feldspathic quartzite, which weathers white. Interbedded with these rocks is greenish-gray slate, phyllite, feldspathic phyllite, and quartzose argillaceous phyllite. Sillimanite porphyroblasts are found in the immediate vicinity of the pluton (Doll et al., 1961).

5.3.6.7 Rock Structure

Joints in the Burke Mountain area trend approximately southwest (Woodland, 1965). Folding is common in the metasediments surrounding the Victory pluton. Woodland (1965) described two generations of small-scale folding (usually some centimeters [inches] across). He suggested that the folds represent two separate phases of the same orogeny.

5.3.7 Willoughby Pluton (Plate IV, Pluton 10)

5.3.7.1 Size, Shape, and Location

The Willoughby pluton is located in extreme northeastern Vermont in the Memphremagog, Island Pond, Burke, and Lyndonville 15-min quadrangles. It is an elongated, roughly oblong, narrow-waisted body whose outcrops occur over an area of approx- imately 92 km^2 (36 mi^2). The long axis of the pluton is almost 19 km (12 mi) long and is oriented east and west. Its width is generally 5-6 km (3-4 mi), and it is roughly bisected by Lake Willoughby.

5.3.7.2 Relative and Absolute Age

The Willoughby pluton belongs to the New Hampshire plutonic series, the plutons of which are considered to be Devonian in age (Doll et al., 1961).

5.3.7.3 Origin

The Willoughby pluton probably was emplaced by forceful injection and stoping (Page, 1968). Some of the sedimentary rock inclusions or remnants in the pluton have retained structural conformity with the country rock and may represent roof pendants. Forceful injection of the nearby Derby pluton, which also belongs to the New Hampshire plutonic series, is indicated by the structure of its wall rock. The Willoughby and Derby plutons may have been emplaced by similar processes.

5.3.7.4 Petrography

Mineralogy and Textural Relations. The granite in the Willoughby pluton is medium to coarse grained, granitoid, and light to dark gray in color. It contains feldspar phenocrysts up to 1 cm (0.4 in.) in diameter. Major minerals include oligoclase, microcline, biotite, and quartz. Accessory minerals include muscovite, apatite, sphene, zircon, zoisite, almandine, and allanite. Dennis (1956) gave a modal estimate of plagioclase, 50%; microcline, 20%; quartz, 20%; biotite, 6%; and others, 4%.

When the plagioclase and microcline are in contact with each other, the plagioclase frequently forms myrmekite. The microcline also is commonly cross-twinned and occasionally perthitic. The microcline and quartz usually occur as anhedral grains. The plagioclase and biotite occur as euhedral to subhedral grains.

Near Lake Willoughby, garnet was found adjacent to the contacts in areas having relatively less biotite. Dennis (1956) suggested that this phenomenon could be the result of potassium migration in these rocks.

Alteration. Some of the feldspar in the Willoughby pluton has been altered to kaolin. Chlorite after biotite is also common. The kaolin is found as a coating on feldspar grains.

5.3.7.5 Secondary Intrusions

Pegmatites in the form of dikes and irregular masses are commonly associated with the Willoughby pluton. They appear to have been injected along tension joints. Up to three generations of pegmatites have been observed locally. Prominent minerals include potassium feldspar, quartz, muscovite, black tourmaline, and garnet. Tourmaline and garnet occur more often near the granite contact.

5.3.7.6 Country Rocks

The Willoughby pluton is in contact with the Waits River formation on its western side and the Barton River member of the Waits River formation and the Gile Mountain formation on its eastern side. The metasediments of the Waits River formation consist of a gray quartzose limestone and a micaceous crystalline limestone. The limestone's constituent minerals are actinolite, hornblende, zoisite, diopside, wollastonite, and garnet. The limestone possesses a distinctive brown, earthy crust when weathered. Intimately associated with the limestone is gray quartz-muscovite phyllite or schist. The schist and phyllite contain biotite, garnet, and locally either sillimanite, kyanite, or andalusite.

The Barton River member consists of intimately associated siliceous crystalline limestone and sericite-quartz-chlorite phyllite. Where it is in contact with granite dikes or sills, it forms diopsidic limestone and cordieritic hornfels.

The Gile Mountain formation is a gray quartz-muscovite schist. A gray micaceous quartzite and a calcareous micaceous schist are interbedded and intergradational with the quartz-muscovite schist. The schist commonly contains porphyroblasts of biotite, garnet, staurolite, and, in the immediate vicinity of the pluton, sillimanite (Doll et al., 1961; Goodwin, 1963; Myers, 1964; and Albee, 1968).

5.3.7.7 Rock Structure

A major set of joints occurs in the granite of the Willoughby pluton. The joints dip steeply, almost vertically, and trend N. 20-30 W°. Crystal Lake and Lake Willoughby are glacial lakes oriented parallel to this joint set.

Most fold axes in the surrounding area are roughly parallel to the axis of the Brownington syncline, which merges with the Townshend syncline in southern Vermont. The Strafford-Willoughby arch passes through the eastern portion of the pluton, extending in a southeasterly direction (Doll et al., 1961).

5.3.7.8 Geophysics

Northeastern Vermont is characterized by a strong gravity low whose outer boundary is marked by a Bouguer-anomaly contour of -50 mgal. The lower density of the granitic rocks in the region, as compared to that of the neighboring metasedimentary rocks, can account for the gravity low (Diment, 1968).

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6 CRYSTALLINE ROCKS OF NORTHEASTERN NEW YORK

6.1 GRENVILLE AND AVALONIAN

6.1.1 Alexandria Granite Gneisses (Plate V, Pluton 6)

6.1.1.1 Size, Shape, and Location

The Alexandria granite gneisses comprise at least 40 bodies of alaskite rock; most of them appear to connect at depth to form a thick sheet approximately 87 km (54 mi) long from Canton to Alexandria Bay in the Adirondack lowlands (see Fig. 6.1), and a few are scattered as far east as the Vermont border. The individual bodies are generally narrow and elliptical or irregularly elongated in map plan (see Fig. 6.2). Many of these bodies are 8-10 km (5-6 mi) long, and some are up to 24 km (15 mi) long. Together they form part of a thick stratigraphic unit that has undergone at least three periods of folding, resulting in a continuous sheet of domes and basins. The alaskite is exposed mainly in the cores of major antiforms, forming blunt bodies with plunging ends. The many outcrops of this rock occur over an area of about 635 km^2 (245 mi^2).

6.1.1.2 Relative and Absolute Age

The age of the Alexandria granite gneisses is Precambrian. Controversy has existed over the origin of these gneisses, which has led to varying ideas about their relative age. Buddington (1939) among others believed the alaskite was intruded into Grenville metasediments, transecting the Stark and Diana complexes. This would make the alaskite younger than all of these units. More recent investigators (Carl and van Diver, 1975; Foose and Carl, 1977; and Romey et al., 1980) believed the alaskite to be a conformable stratigraphic unit within the Grenville metasediments, and perhaps even the earliest unit (Romey et al., 1980), making it much older. Silver (1965) dated zircons of some of the alaskite bodies using the U-Pb method and arrived at an age of 1220 m.y.

6.1.1.3 Origin

As mentioned above, controversy has existed over the origin of the Alexandria granite gneisses. Buddington (1939) and Buddington and Leonard (1962) believed the alaskite to be of igneous origin and to have been intruded into the axial region of preexisting folds, forming phacoliths. These rocks were then metamorphosed during the Grenville orogeny. Engel and Engel (1963)

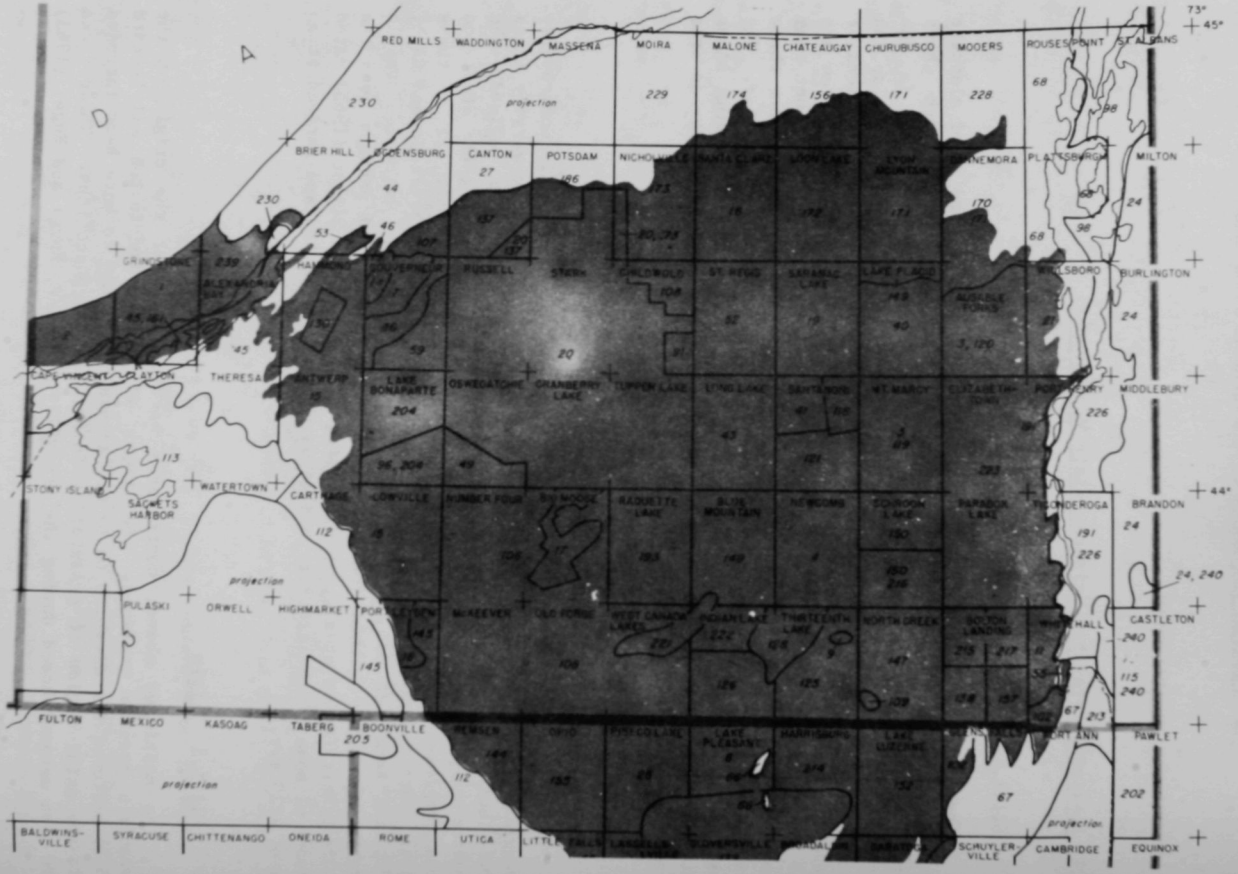

Fig. 6.1 Geologic Quadrangles of the Adirondack Area (Source: Fisher et al., 1971)

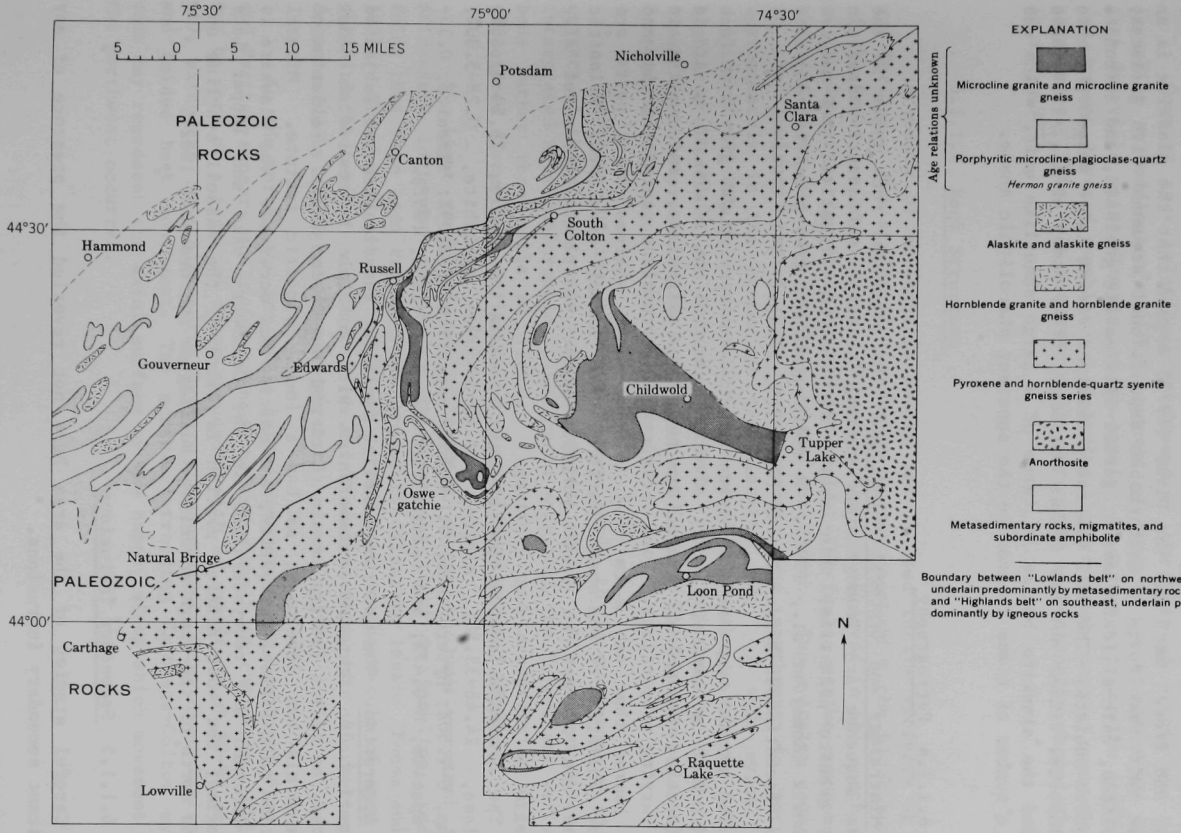

Fig. 6.2 Map of Major Rock Types in the Alexandria, Tupper/Saranac, and Hermon Complexes of the Northwest Adirondacks (Source: Buddington and Leonard, 1962)

believed that the phacoliths were formed by metasomatism of a calcareous arkosic quartzite unit. This would have required the addition of alkalies, alumina, and iron. Carl and van Diver (1975) argued that the alaskite is an ash-flow tuff that forms a conformable depositional sequence with gneisses, amphibolites, calc-silicate layers, garnet-sillimanite gneisses, and carbonate and biotite schists. The tuff was recrystallized during later metamorphism to form alaskite. Other authors (Foose and Carl, 1977; Romey et al., 1980) have explained the alaskite bodies in terms of a large sheetlike unit, which was folded a number of times to produce the apparent phacolithic shape.

6.1.1.4 Petrography

Mineralogy and Textural Relations. The principal rock types in the alaskite sequence are leucogranitic gneiss, granitic and quartz-monzonitic gneiss, granite pegmatite, thin-layered amphibolite, migmatite, and massive vein quartz (Romey et al., 1980). A medium-grained, gray-green trondhjemite forms a discontinuous border facies (Dietrich, 1954).

The granitic gneiss is typically pink to very light gray and fine grained, with an average grain size of about 1 mm (0.04 in.); it is predominantly composed of microcline perthite, oligoclase, and quartz, in widely varying proportions. Some rocks contain microcline perthite grains and quartz leaves that are 2 cm (0.8 in.) or more, whereas other parts are aplitic. Foliation of the sparse mafic minerals is well developed. Mylonitic textures are present in local porphyroblastic varieties (Romey et al., 1980).

The normative composition of this alaskite is quartz, 28.50-41.88%; orthoclase, 24.46-35.5%; albite, 26.20-30.92%; anorthite, 0.28-5.00%; corundum, 0-1.50%; hypersthene, 0.1-2.46%; magnetite, 0-2.09%; ilmenite, 0.30-1.06%; apatite, 0-0.47%; and calcite, 0-0.8% (Buddington, 1957).

Alteration. The upper amphibolite facies of the alaskite was subjected to a period of high-grade metamorphism with multiple phases of folding; therefore, no relict fabrics remain (Carl and van Diver, 1975). Foliation of dark minerals has developed, as has a local mylonitic texture. Mineral alteration to chlorite, antigorite, and calcite has occurred. The alaskite is least deformed and recrystallized in the Cranberry Lake and Tupper Lake quadrangles; the most altered portion is south of the Russell syncline and west and north of the Stark anticline (Buddington and Leonard, 1962).

6.1.1.5 Secondary Intrusions

Careful studies of the area have not revealed the presence of any significant secondary intrusions.

6.1.1.6 Country Rocks

Country rock relationships depend on the mode of origin one prefers. According to Buddington's (1939) concept of origin, the alaskite was intruded into Grenville metasediments, which consist of garnet-sillimanite gneiss, pure and impure layered marble and dolomite, quartz-mesh marble quartzite, and paragneiss. This unit concentrically surrounds each alaskite phacolith (Romey et al., 1980).

6.1.1.7 Rock Structure

A number of strike-slip faults are located in the vicinity of Canton and Pyrites. The strike of the faults is northeasterly; the block northwest of the fault has moved north. Thrust faults are observed in the northern half of the Gouverneur quadrangle (see Fig. 6.1). The faults strike northeast, and the east block has been thrust over the west block. Two mylonite zones occur in the lowlands region. A narrow zone runs southwest to northeast, approximately in the center of the alaskite zone. A wide mylonite zone dipping 30-80° NW. marks the boundary between the lowlands and the Adirondack highlands (Isachsen and McKendree, 1977). The relations of foliation and lineation were presented by Buddington and Leonard (1962), who stressed their contrast to the major folds (see Fig. 6.3).

Foliation in the alaskite bodies tends to be conformable with the contact of the alaskite and surrounding rock units. Carl and van Diver (1975) interpreted this conformable foliation to mean that the layers are sedimentary beds rather than sills or flows. Pegmatitic and crosscutting migmatitic structures disrupt the foliation in places (Romey et al., 1980).

At least three phases of folding appear to have occurred in the Adirondack lowlands, the first two producing tight to isoclinal recumbent folds that have been refolded and eroded to different levels (Foose and Carl, 1977). The dome-and-basin structure in the area is the result of open third-phase refolding of earlier folds. These periods of deformation were accompanied by high-grade metamorphism.

In the north, the intensely folded zone extending southwest from Canton and Pyrites appears to be folded back beneath itself to the south. This zone may be part of a regional nappe structure with an amplitude of 80-97 km (50-60 mi), extending from Alexandria Bay to the northeast, hinged at Parishville, and folded back southwest. The northeast-trending Carthage-Colton mylonite zone may represent the axial surface of the nappe, along which movement toward the northwest occurred.

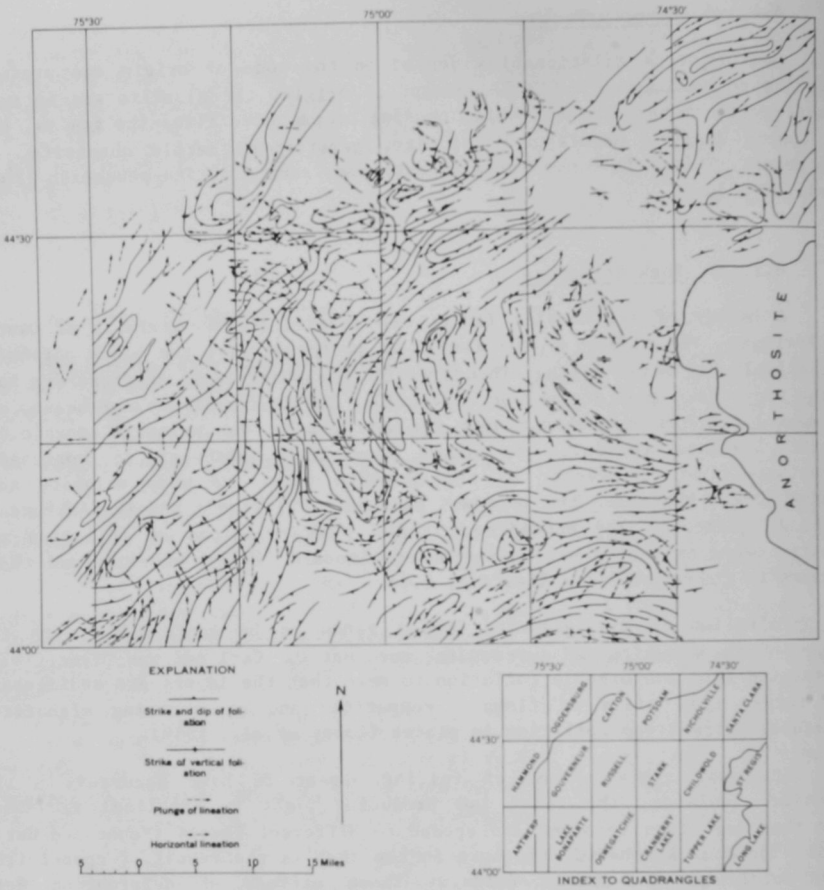

Fig. 6.3 Generalized Trends of Foliation and Linear Structure of the Alexandria, Tupper/Saranac, and Hermon Complexes (note the contrast of lineation with the major folds) (Source: Buddington and Leonard, 1962)

6.1.1.8 Geophysics

The St. Lawrence lowlands show large magnetic lows (Isachsen et al., 1979), but the alaskite body has a distinctive magnetic signature. Breaks in the pattern coincide with zones of brecciation, which are evidence of faulting and displacement (Romey et al., 1980). Magnetic data have also been used to indicate continuation of the unit beneath Paleozoic rocks to the north of the outcrop area.

The Grenville belt along the St. Lawrence River shows a long gravity high through the northwestern lowlands. Simmons (1964) has approximated the shape of the belt as a slab of thickness 3.4-5 km (2.1-3.1 mi), with a minimum density contrast of 0.06 gm/cm^3. Granitic rocks represent local gravity lows in the area.

6.1.2 Diana Gneiss and Tupper/Saranac Metasyenite Complexes (Plate V, Pluton 3)

6.1.2.1 Size, Shape, and Location

Outcrops of the Diana gneiss complex are found over about 700 km^2 (275 mi^2) in the Carthage, Lowville, Antwerp, Lake Bonaparte, Oswegatchie, and Russell quadrangles (see Fig. 6.1). The belt is about 64 km (40 mi) long, about 24 km (15 mi) wide in the southwest, and about 1.6 km (1 mi) wide in the north. The shape of the body is thought to be that of a sheet or sill 10-30 km (6-20 mi) thick that has been complexly folded and eroded. The entire complex is located on the northwestern edge of the Adirondack highland core, along the border where the granitic and charnockitic gneisses of the core meet the metasedimentary rocks of the lowlands.

The Tupper and Saranac complexes to the northeast, and other groups of rock bodies to the southeast of the Diana complex, have been assigned the same pluton number (see Plate V). All these rock bodies have a grossly similar lithology, as indicated in the classic studies by Buddington and his various coworkers since the early 1930s. There may, however, be local differences in the details of lithology and structure of the many rock bodies grouped together for the purposes of this report. Therefore, if more detailed information is needed about any particular one of these bodies, the pertinent geologic quadrangle maps and reports by Buddington and his coworkers should be consulted. For many areas of the Adirondacks, their work is probably the most recent and comprehensive.

The Tupper and Saranac complexes crop out primarily in the Tupper Lake, Long Lake, St. Regis, Saranac, Loon Lake, Lake Placid, Ausable Forks, and Elizabethtown quadrangles (see Fig. 6.1). The Tupper and Saranac complexes are estimated to be no more than 1220 m (4000 ft) thick (Buddington, 1948). All together, they occupy an area of about 2000 km^2 (772 mi^2).

6.1.2.2 Relative and Absolute Age

The Diana and Tupper/Saranac gneisses are Precambrian in age. They have been dated by Silver (1965), using the U-Pb-zircon method, at 1100-1200 m.y. Some believe these gneisses to be younger than both the Adirondack metanorthosite and the metasedimentary rocks of the Grenville sequence, but older than the crystalline rocks of the region (Hargraves, 1968). McLelland

and Isachsen (1980) proposed that syenite dikes in the anorthosites may represent anatectic melt produced from the quartzo-felspathic material during emplacement of anorthosite.

6.1.2.3 Origin

The syenite magma was probably emplaced as a conformable, flat-lying sheet along planes of structural weakness in slightly folded Grenville metasediments (Buddington, 1953; Buddington and Leonard, 1962). Hargraves (1968) suggested that it may have been intruded under an anorthosite sheet, although the anorthosite is no longer present. The varied facies of the complex may be due to gravity stratification within the sheet (Buddington and Leonard, 1962; Hargraves, 1968). The formation of the upper felsic and the basal mafic facies in the Tupper/Saranac complex may be the result of modification by the assimilation of amphibolitic material and/or reaction with residual anorthosite liquid (Buddington, 1948; Davis, 1971).

6.1.2.4 Petrography

Mineralogy and Textural Relations. Four types of gneiss can be distinguished in the Diana gneiss complex, grading from green pyroxene syenite gneiss in the northwest through pyroxene-(hornblende)-quartz syenite gneiss, and pink hornblende-quartz syenite gneiss to hornblende granite gneiss (or syenite gneiss) in the southeast (see Fig. 6.4) (Buddington and Leonard, 1962; Hargraves, 1968; and Isachsen and Fisher, 1970). Buddington and Leonard (1962) argued that the mafic pyroxene-rich facies form the bottom and the more felsic facies the top of the gravity-differentiated sheet.

In general, the Diana gneisses have undergone amphibolite facies metamorphism and exhibit a coarse phacoidal structure. They rarely contain andesine, labradorite xenoliths, or xenocrysts from anorthosite or gabbroic anorthosite. The texture of the foliated rock varies from intensely mylonitized augen gneiss in the northwest to granoblastic gneiss in the southeast (Geraghty et al., 1981).

The green pyroxene-quartz syenite gneiss has a mortar or augen texture. It contains zoned feldspar porphyroclasts up to a few millimeters (a few tenths of an inch) in diameter, with an oligoclase core and a microperthite rim. Quartz is present in leaflike forms or as interlacing veins. The composition of the facies, in percent by weight, is feldspar, 74-82%; quartz, 10-15%; ferroaugite, 4.7%; hornblende, 0.3-1%; and magnetite and ilmenite, 2-5%, with accessory sphene, zircon, and apatite (Buddington and Leonard, 1962).

The green pyroxene syenite gneiss also has a mortar or augen texture, but is slightly coarser grained than the pyroxene-quartz syenite gneiss.

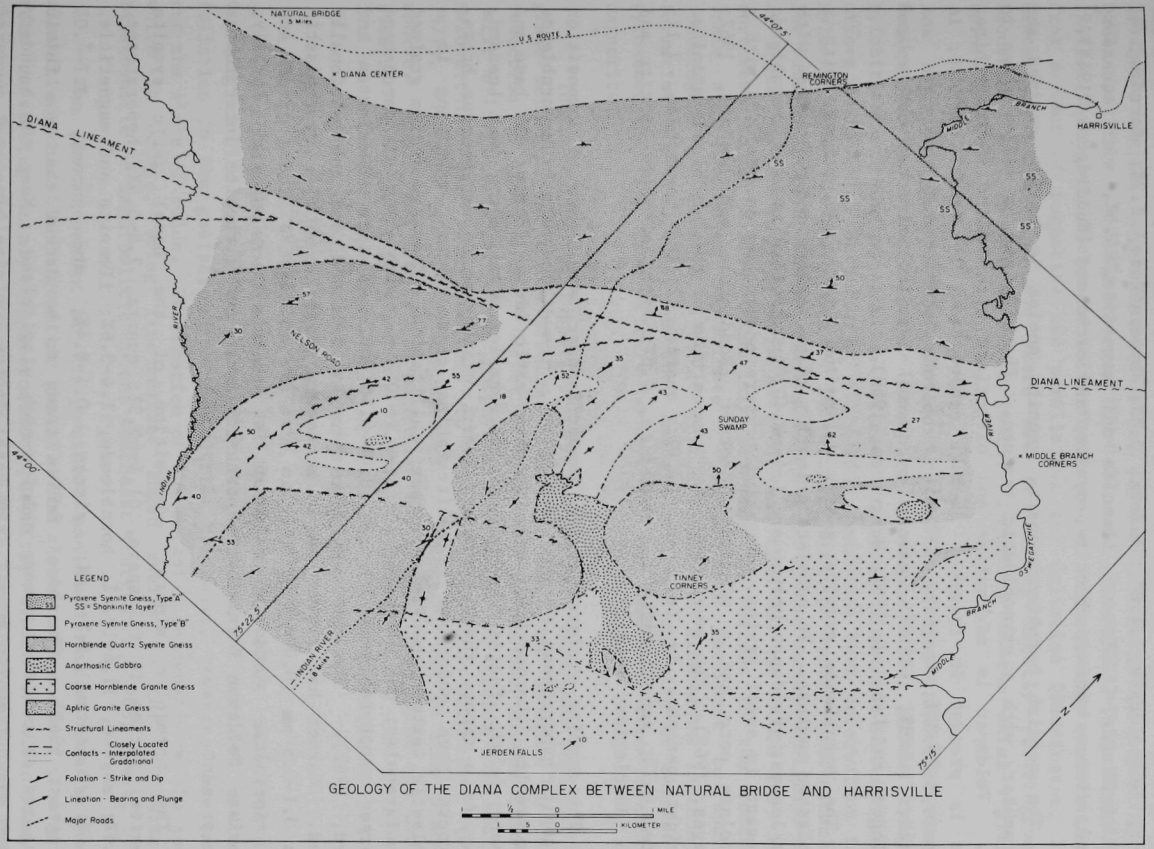

Fig. 6.4 Geological Map of the Diana Syenite Gneiss Complex between Natural Bridge and Harrisville (blank areas within map area signify complete lack of exposure where extrapolation was not justified) (Source: Hargraves, 1968)

Plagioclase feldspars are rimmed with microperthite. The rock contains a number of thin shonkinitic bands or lenses. Average composition, by weight percent, is feldspar, 74.8%; quartz, 5.8%; hypersthene, 1.2%; ferroaugite, 9.2%; hornblende, 1.8%; and ilmenite and magnetite, 4.5%, with accessory apatite, zircon, sphene, chlorite, pyrite, and carbonate (Buddington, 1948).

The reddish hornblende-quartz syenite gneiss is medium grained and equigranular, with a granoblastic texture. Phacoids may be yellow-green in color. Feldspar is microcline and oligoclase with relict microperthite. Quartz is present in leaves. Average composition, by weight percent, is feldspars, 75-80%; quartz, 7-18%; hornblende, 4-8%; and magnetite and ilmenite, 1.5-3%, with accessory pyroxene, apatite, zircon, and sphene (Buddington and Leonard, 1962; Hargraves, 1968).

The hornblende granite gneiss is medium to coarse grained, with a granoblastic texture. It consists primarily of microcline and oligoclase with quartz leaves. Feldspar grains are 1-4 mm (0.04-0.16 in.) in size; hornblende is present in small grains. Rapakivi texture occurs, as do phacoids with a plagioclase core and a microcline border (Buddington and Leonard, 1962; Hargraves, 1969). The average mineral composition of the hornblende granite, by weight percent, is feldspar, 67.8%; quartz, 23.1%; hornblende, 7.1%; biotite, 0.7%; and ilmenite and magnetite, 0.8%; with accessory apatite and zircon (Buddington, 1948).

The syenite gneisses of the Tupper/Saranac area range in composition from augite-hypersthene syenite (mangerite) to augite-hypersthene-hornblende quartz syenite (charnockite), with the more mafic facies forming the base and the more felsic facies the top of the complex (Buddington and Leonard, 1962). As a group, these rocks are green, weathering to maple-sugar brown. The rocks have undergone granulite-facies metamorphism, resulting in a granoblastic texture and variable recrystallization. They are medium grained and equigranular; the average grain size is 0.3 mm, with a varible percentage of larger prophyroblasts 1-4 mm (0.04-0.16 in.) in size (Buddington and Leonard, 1962). The gneisses are strongly foliated and compositionally layered. Garnet has formed as an accessory mineral, both as granoblastic grains 0.1-0.5 mm (0.004-0.02 in.) in size and as coronas around oxides and mafic minerals. Potassium feldspar is present as perthite. Locally, plagioclase feldspar forms large porphyroblasts, and myrmekite intergrowths may be present.

The normative mineral composition of the basal mafic syenite (mangerite) facies is feldspars, 61.1-65.3%; quartz, 2.8-5.5%; hypersthene, 4.6-7.3%; augite, 11.6-12.1%; hornblende, 1.4-3.6%; ilmenite and magnetitie, 5.3-8.4%; apatite, 1.2-1.4%; zircon, 0.1-0.4%; and garnet, 3.7-4.0% (Buddington, 1948). The basal facies tends to be darker than the felsic facies, and it contains a more calcium-rich plagioclase. Near the boundary with the metanorthosite massif, the basal facies contains blue-gray augen of plagioclase and xenocrysts of andesine. The basal facies also contains layers of skarn and schlieren of amphibolite.

The normative mineral composition of the upper felsic syenite (mangerite) facies is feldspars, 66.3-70.3%; quartz, 16.3-18%; hypersthene, 3.6-5.8%; augite, 1.4-3.4%; hornblende, 2.8-5.7%; ilmenite and magnetite, 1.5-2.0%; apatite, 0.5-0.8%; zircon, 0.3-0.5%; and garnet, 0.2-0.5% (Buddington, 1948). The felsic facies contains short, narrow pegmatite veins radiating out from a knot of pegmatite and paralleling the foliation (Buddington, 1948). Lenses of amphibolite and skarn are found only locally (Buddington and Leonard, 1962). Quartz may form pencil lineations in the axial regions of folds.

Alteration. The Diana gneiss complex lies along the northeast-trending Carthage-Colton mylonite zone (Geraghty et al., 1981). Within this zone mylonitization has produced a decrease in grain size and has developed a strong foliation. Geraghty et al. (1981) concluded that deformation in the mylonite zone was due primarily to recrystallization rather than cataclasis. The fabric within the Diana complex varies from almost unrecrystallized mortar or augen texture with a grain size of less than 0.2 mm (0.008 in.) in the northwest to an almost completely granoblastic texture in the southeast. Grain size increases approximately tenfold from northwest to southeast (Buddington and Leonard, 1962). Postdeformational alteration includes the formation of hornblende, biotite, and yellow-green chlorite or calcite from pyroxene; chlorite from primary hornblende; and leucoxene from titaniferous magnetite.

The rocks of the Tupper/Saranac complex have undergone regional high-grade metamorphism, which produced plastic flowage and moderate to intense reduction in grain size. Garnet is developed as coronas around oxide and mafic silicate grains, as as porphyroblasts where the grains are smallest. Mesoperthite was recrystallized to microperthite and oligoclase (Davis, 1971). Primary hornblende was altered to chlorite, and in some places to sphene. Augite was altered to secondary hornblende (Buddington and Leonard, 1962).

6.1.2.5 Secondary Intrusions

The gneisses are frequently cut by metadiabase dikes measuring up to 6 m (20 ft) wide The dikes usually strike northwest and parallel banding and foliation in the gneiss; where they strike northeast, however, they tend to cut the limbs of folds. Deformation of the metadiabase depends on the zone in which it occurs, being fine-grained augen gneiss in the northwest and blastophitic or amphibolitic to the southeast (Hargraves, 1968; Davis, 1971).

The Diana complex at its southern end is intruded into by younger alaskitic and honblende granitic gneiss. The alaskite is a pink, fine-grained rock consisting mainly of microcline perthite, albite or oligoclase, and quartz. The hornblende granite is a reddish, medium-grained, equigranular rock consisting primarily of quartz, orthoclase, albite, anorthite, hypersthene, and magnetite. The vertical joints in Tupper/Saranac complex tend to be intruded by unmetamorphosed diabase dikes (Davis, 1971).

6.1.2.6 Country Rocks

The Diana and Tupper/Saranac magmas were intruded into a Grenville metasedimentary sequence composed of garnet-sillimanite gneiss, pure and impure layered marble and dolomite, quartz-mesh marble-quartzite, and paragneiss. The western border of the Diana complex is still in contact with these Grenville metasediments. Isolated syenite sills are found in the limestones of this sequence.

Hargraves (1968) and Davis (1971) postulated that the gneisses may have been intruded underneath an anorthosite or anorthositic gabbro sheet. Although the anorthosite is no longer present in area, there is a small anorthosite mass to the southwest of the Diana complex, called the Croghan mass. Alaskitic and hornblende granitic gneiss are in contact with the Diana gneiss on its southeastern boundary, as mentioned in the previous section.

6.1.2.7 Rock Structure

The northwestern boundary of the Diana complex is a mylonite shear zone where the gneisses are in contact with Grenville metasediments. The contact consists of a band of intrusive breccia composed of thin-banded pyroxenic quartz syenite and fragments of Grenville rocks (Hargraves, 1968). The mylonite zone extends for 109 km (68 mi) from Carthage to beyond Colton (see Fig. 6.1) to the northeast. The zone ranges in width from a few meters (feet) to 17.7 km (11 mi), and in thickness from 0 m to 5000 m (16,000 ft). The average dip of the mylonite zone is approximately 40° NW. The zone is thought to represent an eastward-directed overthrust zone, in which mylonitization occurred by ductile deformation. Twenty minor faults and fault zones are found within the mylonite zone (Geraghty et al., 1981). Other topographic linear features in this region tend to be oriented north to northeast, but in the northern section of the Diana gneiss complex, some lineaments trend east-west (Isachsen and McKendree, 1977).

Primary layering in the Diana gneiss may be found locally, but it is difficult to see because of metamorphism; secondary foliation paralleling the primary layering is intense (Buddington and Leonard, 1962). Regional mylonite foliation tends to parallel foliation outside of the mylonite zone (Geraghty et al., 1981). Lineation within the plane of foliation takes the form of parallel arrangement of quartz leaves and granulated porphyroclasts. The lineation tends to plunge north or north-northwest, or to the northeast (Hargraves, 1968), and it is observed both parallel to and cutting across the trends of major folds (Geraghty et al., 1981).

The Diana gneiss complex consists of more mafic rocks to the northwest than to the southeast and tends to dip to the northwest. Therefore, if the complex is a gravity-differentiated sheet, it must be overturned to the southeast (Hargraves, 1968). The fold-thrust nappe between Carthage and Colton has apparently produced the overturned, isoclinal-to-recumbent folds of

the complex (Geraghty et al., 1981). In the southwestern part of the complex, the pyroxene syenite and pyroxene-quartz syenite gneisses are exposed on anticlinal cores. Overlying hornblende-quartz syenite and hornblende granite gneisses are revealed in the synclines (Buddington, 1948). Folding tends to trend northeasterly (Buddington and Leonard, 1962).

Faults have been observed in the Tupper/Saranac complex in the vicinity of Tupper Lake. The Tupper Lake fault trends approximately north-south and has a vertical throw of 915-1370 m (3000-4500 ft), with the downthrown block to the west. Jointing in this area is approximately perpendicular to the strike of foliation (Buddington and Leonard, 1962). In the Saranac quadrangle (see Fig. 6.1), two jointing directions are prevalent -- N. 65° E. to N. 80° W., and N. 5° E. and N. 30° W. The first direction parallels diabase dikes in the area. Many of the joints show slickensides, indicating that movement has occurred (Buddington, 1953). An igneous breccia has been noted on the northeastern border of the Tupper/Saranac complex northeast of Saranac Lake (Buddington, 1948).

The gneisses of the Tupper/Saranac complex show intense secondary foliation parallel to primary layering and flow foliation (Buddington and Leonard, 1962). Foliation consists of alternating lamellae of mafic and oxide minerals with quartz and feldspar. Quartz tends to be present in grains that are elongated in the plane of foliation. Foliation is locally wavy and plicated (Buddington and Leonard, 1962). Lineation in these rocks tends to parallel major fold axes (see Fig. 6.3), and reflects parallel, elongated grains of quartz and hornblende.

The Tupper/Saranac complex is a strongly folded sheet. In the Long Lake and Tupper Lake quadrangles (see Fig. 6.1), it forms a broad east-west anticlinal prong composed of a pyroxene syenite gneiss core flanked by hornblende granite. The anticline is asymmetric, with a steeper dip on the north limb. The limbs dip 30-40° and the fold plunges gently to the east (Buddington and Leonard, 1962). In the Saranac quadrangle, the Tupper/Saranac gneisses form a syncline pitching to the north. The trough of the fold is composed of metasediments (Buddington, 1953). In the Loon Lake quadrangle (see Fig. 6.1), the Tupper/Saranac gneisses are folded and border a domical mass of the Stark complex (Postel et al., 1956).

6.1.2.8 Geophysics

Earthquake activity has been recurrent in the area of the Diana gneiss complex (Geraghty et al., 1981). A positive aeromagnetic anomaly, probably due to the presence of ilmenite-magnetite pyroxenite and shonkinite layers, has been observed in the Diana gneiss. Isachsen et al. (1979) reported that the Diana high continues under the Paleozoic cover, whereas magnetic lows are associated with marbles and weakly magnetic sedimentary rocks.

A positive gravity anomaly is associated with the Diana complex. The average density of samples containing less than 10% quartz is 2.74 g/cm^3, that of local granite gneiss is 2.67 g/cm^3, and that of average granite to the southeast is 2.66 g/cm^3 (Buddington, 1968).

The Tupper/Saranac complex is associated with a negative gravity anomaly (Brock, 1980).

6.1.3 Granite Gneiss (Plate V, Pluton 7)

6.1.3.1 Size, Shape, and Location

Pyroxene and/or hornblende granite gneiss is found in a belt approximately 65 km (48 mi) long in the northeastern portion of the Adirondack dome, extending from the Chateaugay quadrangle southeast to the Ausable Forks quadrangle. The quadrangles in which this belt mainly lies include the Churubusco, Loon Lake, Lyon Mountain, and Dannemora quadrangles (see Fig. 6.1). To the west the granite gneiss grades into biotite and/or hornblende gneiss. To the south it grades into leucogranitic (alaskite) gneiss. This body of rock occupies an area of about 925 km^2 (357 mi^2).

6.1.3.2 Relative and Absolute Age

The granite gneiss is of Precambrian age. It is younger than the quartz syenite gneisses, metanorthosite, and Grenville metasediments of the area. Younger still are the diabase dikes that intrude the major joint system in the area (Postel et al., 1956). No absolute age data are available.

6.1.3.3 Origin

Postel (1951) distinguished three types of granite: a microperthite type, of magmatic origin; a microcline and plagioclase type, partly of magmatic origin and partly due to granitization; and a microantiperthite type formed from reaction between late sodic magmatic "emanations" and the earlier types of granite. The physical mode of emplacement was not discussed.

6.1.3.4 Petrography

Mineralogy and Textural Relations. The granite gneiss is a medium-grained rock that is pink when fresh and a light yellow-brown after weathering. It has a xenomorphic granular texture in thin section. It occurs both as a massive, structureless rock and as a gneiss, with gradational contact between them. Foliation is usually well developed, especially in the ferromagnesian silicates. The massive rock also may be gneissic, but its low

mafic content makes it difficult to recognize it as such. Pegmatites are common, both parallel to and crosscutting the foliation. Very large miarolitic cavities are commonly found in close proximity to zones of pegmatite near Lyon Mountain. The cavities contain crystals of all the minerals found in the granite. The mineral composition of the granite gneiss includes varying amounts of potassium feldspar (orthoclase and microcline), plagioclase (albite and oligoclase), microperthite, quartz, green pyroxene, and hornblende or biotite. Accessory minerals include zircon, apatite, titanite, and magnetite (Gallagher, 1937; Postel, 1951; and Postel et al., 1956). Postel (1951) and Postel et al. (1956) distinguished four types of granite gneiss in the Dannemora and Loon Lake quadrangles, including microperthite-, microcline-, microantiperthite-, and plagioclase-rich facies. Disseminated magnetite and magnetite ore bodies may be present in the plagioclase-rich facies.

Alteration. According to Postel (1951), three of the four facies he recognized in the granite gneiss are due to alteration. Gallagher (1937) believed the numerous concentrations of magnetite found in this rock to be due to metasomatic replacement. Iron in chlorides and fluorides may have penetrated the rock along shear zones during a period of deformation (Postel, 1951). Foliation, especially evident in granite of high ferromagnesium mineral content, is also present.

6.1.3.5 Secondary Intrusions

Numerous dikes of two different compositions cut the granite gneiss. Diabase dikes striking east-west and northeast-southwest and dipping almost vertically are inferred to be of Precambrian age. They range from a few millimeters to a few meters (a fraction of an inch to many feet) in thickness. The diabase is fine grained and dark gray; it occurs in both porphyritic and nonporphyritic varieties, and may have a chilled border where it is in contact with the granite. The Lyon Mountain diabase dikes are approximately 50% labradorite, with varying amounts of pyroxene, biotite, magnetite, and accessory apatite. Secondary chlorite and calcite usually are present. The diabase dikes were apparently intruded into major joint and fault systems (Gallagher, 1937).

Syenite dikes also strike east-west. The syenite of Lyon Mountain is dark greenish-black, aphanitic or fine grained, and composed of anorthoclase, albite, pyroxene, amphibole, biotite, magnetite, apatite, calcite, and chlorite. These dikes intrude into sedimentary rocks of Ordovician age (Gallagher, 1937; Postel, 1951).

6.1.3.6 Country Rocks

The Potsdam sandstone, of Cambrian age, overlies the granite gneiss to the north and east. The country rock to the west, as well as a body of rock

enclosed within the outcrop area of the granite in the Lyon Mountain quadrangle, consists of granitic and quartz syenitic gneiss of phacoidal structure. These rocks may be leucocratic and have varying proportions of hornblende, pyroxene, and biotite. They are dark green and coarse grained. Along the margin of these gneisses is dark gray-green, medium-grained, foliated pyroxene-hornblende quartz syenite gneiss (Postel et al., 1956; Isachsen, 1970). To the southeast the granite is in contact with the metanorthosite massif. With the exception of the Potsdam sandstone, these rocks are all older than the granite gneiss.

6.1.3.7 Rock Structure

Rocks of the northeastern Adirondacks seem to be less intensely faulted than those of the west and south (Isachsen and McKendree, 1977). Numerous small faults have been noted at Lyon Mountain. In the Dannemora quadrangle, major faulting has caused brecciation; all the rocks exhibit predominantly north-south and east-west jointing (Postel, 1951), although topographic linear features in this region trend mainly southwest-northeast. A few long linear features in the Lyon Mountain and Churubusco quadrangles run northwest-southeast (Isachsen and McKendree, 1977).

The granite gneiss tends to be well foliated, especially in rocks with prominent ferromagnesian silicates. In more leucocratic rocks, foliation may be recognized by a slight elongation of quartz grains and a slightly coarser texture. The regional strike of foliation in the Dannemora quadrangle is north to northeast, with a dip moderate to steep to the northwest or southeast. Lineation directed to the north or northeast and plunging to the north at a moderate angle is common within the plane of foliation (Postel, 1951).

Deformation appears to have affected the granite gneiss over its entire extent. Large plunging folds, recumbent in places, have been mapped. Deformation is believed to have taken place during or after formation of the gneiss with major stress from the east and southeast. Some shearing and granulation is apparent microscopically in rocks of the Dannemora quadrangle (Postel, 1951). The diabase dikes of the Loon Lake quadrangle do not appear to have been affected by this deformation (Postel et al., 1956).

6.1.3.8 Geophysics

The Adirondack Mountains have been shown to be an area of frequent but weak seismic activity (Sbar and Sykes, 1977). The north flank of the Adirondack massif is particularly active, with earthquakes occurring at midcrustal depths (Aggarwal et al., 1977). Thrusting is the predominant fault mode, with maximum compressive stress approximately N. 70° E. and horizontal. Earthquakes tend to occur on faults running north-northwest and dipping east.

The USGS has published an aeromagnetic map that includes the Adirondack region (Zietz et al., 1980), which shows good correlation between the magnetic and structural grain of the area.

Gravity lows in the Adirondacks are difficult to associate only with granite, as proximal anorthosite also is a contributing factor (Simmons, 1964). Large areas of granite do not individually produce the usual gravity lows, suggesting either that the volume of granite in these areas may be small or that the density contrast between the granitic and surrounding rocks is slight, or both.

6.1.4 Hornblende Granite and Hermon Granite Gneisses (Plate V, Pluton 5)

6.1.4.1 Size, Shape, and Location

The hornblende granite gneiss, formerly known as the Lowville-St. Regis granite gneiss, forms a belt in the northwestern Adirondack Mountains approximately 160 km (100 mi) long that extends from Lowville in the southwest to Chateaugay in the northeast (see Fig. 6.1). Although it includes many small and a few very large outcrops that are discontinuous in their surface exposure, this pluton is thought to be a mass of batholithic proportions that has a maximum thickness of several kilometers (miles) (Buddington, 1948). The pluton is concordant and was intruded into an area of strongly folded and deformed rocks of the Diana, Tupper/Saranac, and Stark complexes as well as the metasedimentary sequence.

The Hermon granite gneiss forms several elongate outcrops (see Fig. 6.2) in an area about 64 km (40 mi) long from Hermon to Evans Mills and about 32 km (20 mi) wide (Isachsen and Fisher, 1970). The granite is the dominant igneous rock type in the Grenville metasedimentary rocks of the Adirondack lowlands. Many outcrops of elongated or distorted form located on the southern and southeastern edge of the Adirondacks have a similar lithology and are classed with the hornblende and Hermon granite gneisses for the purposes of this report. The total area within which this group of rocks crops out is about 4300 km^2 (1660 mi^2). All these rock bodies have grossly similar lithology, as indicated in the classic studies by Buddington and his various coworkers since the early 1930s. There may, however, be local differences in the details of lithology and structure of the many rock bodies grouped together for the purposes of this report. If more detailed information is needed about a particular body, the pertinent geologic quadrangle maps and reports by Buddington and his coworkers should be consulted.

6.1.4.2 Relative and Absolute Age

The hornblende granite gneiss is of Precambrian age. It is younger than the syenitic gneisses with which it shows a transgressive relationship

and in which it has locally formed dikes. The Hermon granite gneiss is also Precambrian in age. It appears to be somewhat older than the alaskite and younger than the syenites and metadiabase dikes of the area (Buddington and Leonard, 1962).

6.1.4.3 Origin

The hornblende granite was probably emplaced as a magmatic body by mechanical displacement of existing folded country rock, which yielded and opened along foliation planes. Although basically conformable, the granite locally crosscuts phacolithic sheets (Buddington, 1948; Buddington and Leonard, 1962).

The magma of the Hermon granite gneiss appears to have been emplaced by successive intrusions into folded metasedimentary rock during a period of deformation (Buddington, 1939). The original biotite-quartz-feldspar gneiss, granulite, and local amphibolite of the Grenville series were altered and replaced by interaction with the magma and its associated solutions. Alteration may have caused migmatization and anatexis of the metasediments as well as hybridization of the magma, which may have been a volatile-rich equivalent of the magma that produced the hornblende granite (Buddington and Leonard, 1962). The contact between the granite gneiss and the metasediments may be sharp where the gneiss is in contact with amphibolite and gradational where it is in contact with mixed gneisses (Buddington and Leonard, 1962).

6.1.4.4 Petrography

Mineralogy and Textural Relations. The hornblende granite gneiss is mostly a reddish, medium-grained, equigranular, gneissic hornblende granite of generally uniform composition. It contains numerous local inclusions of amphibolite and metasedimentary rocks. These inclusions are commonly associated with pegmatite veins that parallel the foliation. The normative mineral composition of the granite is quartz, 19.6-26.76%; orthoclase, 32.80-37.3%; albite, 26.72-31.4%; anorthite, 4.17-7.0%; diopside, 0.11-4.3%; hypersthene, 1.6-2.25%; magnetite, 2.1-4.18%; ilmenite, 0-0.97%; apatite, 0-0.58%; and calcite, 0-0.45%. Microperthite is the main feldspar (40-70%); hornblende is usually less than 5%, and biotite occurs locally. Accessory minerals include magnetite, apatite, zircon, and sphene. The average specific gravity of the rock is 2.640 g/cm^3 (Buddington, 1934 and 1939).

The texture and mineralogy of the hornblende granite in the northwest differs from that in the southeast because of the more intense metamorphism in the northwest. In the north the granite is a strongly deformed granoblastic gneiss, whereas in the south it is a gneissoid hornblende-microperthite granite (Buddington and Leonard, 1962). In the less deformed rocks to the south, the feldspar grains are 1.5-4 mm (0.06-0.16 in.) long and form an interlocking texture; quartz crystals are elongate, and elongate feldspar and

hornblende have their long axes aligned in the plane of foliation. Elsewhere in the south, the granitic rocks display a mortar texture with grains 0.2-0.4 mm (0.0079-0.016 in.) in diameter.

The Hermon granite gneiss is a coarsely porphyritic augen gneiss. The augen are pink or light-colored, Carlsbad-twinned microcline 1.3-2.5 cm (0.5-1 in.) long in a coarse- to medium-grained gray groundmass of oligoclase and quartz. Phenocrysts may also be composed of oligoclase, orthoclase, or microperthite. Considerable granular mortar of oligoclase and quartz may be present. Locally the rock may be recrystallized to a phacoidal aggregate. Rarely, the rock may be massive. The gneiss is usually biotitic, but hornblende may predominate locally.

The composition of the Hermon granite gneiss varies widely where the rock is involved with Grenville metasediments and bands of mixed gneisses, garnet gneiss, limestone, and amphibolite are present. In the latter two cases, the Hermon granite gneiss is associated with a syenite facies. The average mineral composition of the Hermon granite gneiss is potassium feldspar, 0-57%; oligoclase, 11-41.2%; microcline, 0-31.3%; quartz, 16.6-28.1%; hornblende, 0-10.9%; biotite, 0.9-13.4%; magnetite, 0-1.7%; apatite, 0-0.9%; zircon, 0-0.3%; pyrite, 0-0.2%; and sphene, 0-0.4% (Buddington, 1939). In general, the Hermon granite gneiss is less siliceous than the hornblende granite. It also differs in having primary oligoclase and potassium feldspar, rather than microperthite (Buddington, 1939 and 1948; Buddington and Leonard, 1962).

Alteration. Regional metamorphism both preceded and accompanied emplacement of the hornblende granite gneiss. Metamorphism produced a gneissoid rock, with little or no crushing and no garnet in the southwest (Buddington, 1939). The rock consists of microperthite grains 1.5-3.5 mm (0.05-0.14 in.) in diameter, embedded in an approximately equal amount of a granular groundmass of microperthite grains 0.2-0.4 mm in diameter (Buddington and Leonard, 1962). Mortar can form as little as 10% of the rock. In the more deformed rocks to the north, the granite gneiss is completely recrystallized. Average grain size is 0.5-0.8 mm (0.02-0.03 in.). In the Santa Clara quadrangle (see Fig. 6.1), metamorphism caused garnet reaction rims to form around the pyroxene, hornblende, and magnetite. Also, slightly higher percentages of ferric oxide and titanium dioxide also are noted in the gneiss in the Santa Clara quadrangle.

Along the Carthage-Colton mylonite zone, deformation of the Hermon granite produced granular mortar gneiss with a mylonitic groundmass. Local recrystallization produced a phacoidal aggregate. The border facies of the rock were locally deformed and recrystallized. Other alteration effects include a little secondary sphene and the frequent alteration of biotite and hornblende to chlorite (Buddington, 1939).

6.1.4.5 Country Rocks

The hornblende granite gneiss intrudes into rocks of the Diana, Tupper/Saranac, and Stark quartz syenite-granite complexes (see Secs. 6.1.2 and 6.1.7). Although the hornblende granite gneiss transgresses the structure of the country rock locally, it cuts it only at very small angles. For the most part, it conforms to preexisting folds and foliation (Buddington, 1939).

Remnants of older metasediments are found as inclusions in the granite gneiss. These consist of calcite marble, local dolomite marble, quartzite, gneiss, amphibolite, migmatite, granulite, and local schist. Leucogranitic gneiss occurs in some areas at or near the borders of the granite gneiss. This rock may be a differentiate of the hornblende granite.

The Hermon granite magma was intruded into and altered the Grenville metasediments, which consist of biotite-quartz-plagioclase gneiss, amphibolite, related migmatite, and calcitic and dolomitic marble with local calc-silicate rocks.

6.1.4.6 Rock Structure

Regional jointing in the hornblende granite gneiss in the Santa Clara quadrangle strikes N. 60-90° E. or N. 5° E. to N. 25° W., with dips usually more than 70° in either possible direction (Buddington, 1937). Similar jointing patterns seem to prevail to the southwest (Isachsen and McKendree, 1977).

In the lowland belt, where the Hermon granite gneiss crops out, thrust faults (with the east side overthrust), mylonite, fault breccia, and strike-slip faults, all mostly trending southwest and northeast, have been noted (Isachsen and McKendree, 1977).

The hornblende granite gneiss was intruded into rocks that were already folded, and was subsequently involved in isoclinal folding. Foliation and linear structures in the gneiss tend to be oriented appropriately for the area and structure in which they occur. Foliation tends to strike northeast (see Fig. 6.2). Linear structures may be the result of magmatic flow, especially in the southern area, or plastic flow associated with folding (Buddington, 1937; Buddington and Leonard, 1962).

The country rock into which the hornblende granite gneiss was intruded was already highly folded (see Fig. 6.2). The magma crosscut these structures at very small angles and formed local phacoliths in the existing structures. Orogenic deformation accompanied and followed emplacement of the gneiss; such activity was most intense west of the Stark complex, moderate to the southeast, and slight to the south (Buddington and Leonard, 1962).

The Hermon granite gneiss consists of folded, possibly phacolithic sheets that are conformable with synclinal and anticlinal structures in the

country rock (Buddington, 1939). In places the granite transgresses the metasediments (Buddington, 1948).

6.1.4.7 Geophysics

Seismic activity in the northwestern Adirondack Mountains is recurrent but usually weak. The hornblende and Hermon granite gneisses are near the epicenter of the 1944 Massena earthquake (I = 8, M = 5.6) (Geraghty et al., 1981). There are no descriptions in the literature of earthquakes within the granite gneisses.

The USGS has published an aeromagnetic map that includes the Adirondack region (Zietz et al., 1980). Gravity studies of the northwestern Adirondack Mountains fail to reveal gravity anomalies that can definitely be attributed to the presence of the hornblende granite gneiss. Either the thickness of granite is minimal or the density contrast between granite and country rock is too slight (Simmons, 1964).

6.1.5 Main Metanorthosite Massif and the Snowy Mountain, Thirteenth Lake, and Oregon Anorthosite Domes (Plate V, Pluton 1)

6.1.5.1 Size, Shape, and Location

The main metanorthosite massif (Marcy massif) (Isachsen and Moxham, 1968) is roughly heart-shaped in map plan and covers approximately 3433 km^2 (1325 mi^2). The main massif is basically a domical layer with subordinate domes and basins on the upper surface (Simmons, 1964). It is located in the northeastern section of the Adirondack Mountains in an area of granulite facies gneisses and was probably the first of the great Adirondack intrusive bodies to break into the older metasediments (e.g., Miller, 1918). Inward from the margins of the massif, the metasedimentary rocks occur only locally as roof pendants (Isachsen and Fisher, 1970). The massif is surrounded by a variety of rock types, which include metasediments and metagabbro, but among which charnockitic gneisses of uncertain origin are predominant (McLelland and Isachsen, 1980). The entire terrane of metanorthosite consists of the main massif and numerous other bodies to the north and south that form a belt at least 130 km (80 mi) long that strikes north-northeast. Other bodies of significant size occur to the south in the Thirteenth Lake and Indian Lake quadrangles (see Fig. 6.1). The largest, the Oregon dome, is roughly circular in map plan and covers approximately 150 km^2 (58 mi^2). The Snowy Mountain dome (see Fig. 6.5) is somewhat triangular in shape and crops out over an area of about 35 km^2 (14 mi^2).

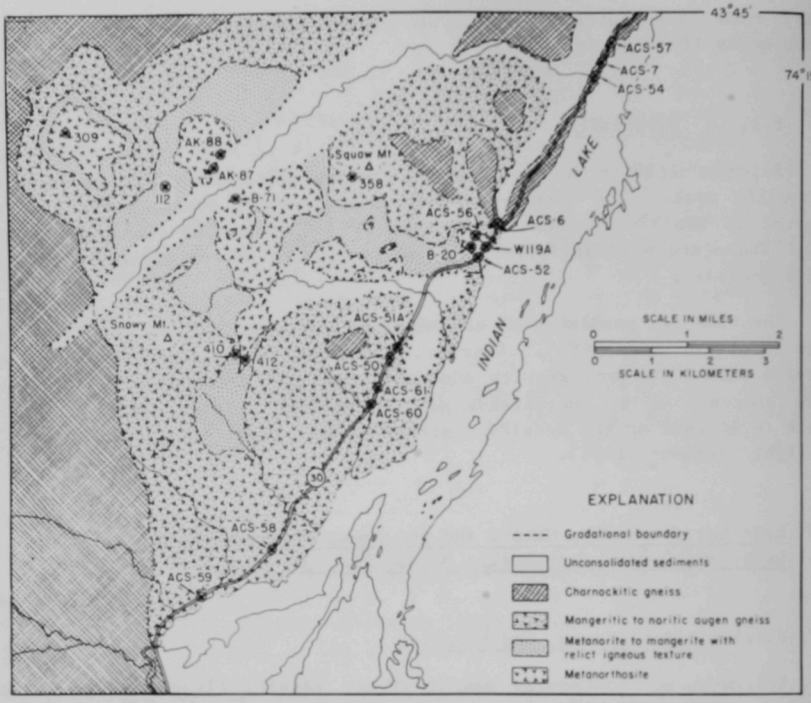

Fig. 6.5 Geology of the Snowy Mountain Dome Area
(Source: de Waard and Romey, 1968)

6.1.5.2 Relative and Absolute Age

The absolute age of the metanorthosite was determined by U-Pb isotope dating of zircons to be about 1130 m.y. (Silver, 1969). This is younger than the metasedimentary rocks into which it intruded and slightly older than the associated charnockitic gneisses. All of these units were formed during the Precambrian.

6.1.5.3 Origin

The main massif was probably emplaced into the metasediments as a large, slablike mass via two underlying feeder tubes (Simmons, 1964). The original magma may have been a gabbroic anorthosite liquid with a high percentage of suspended plagioclase crystals that became concentrated toward the core by flow differentiation. This concentrated the gabbroic anorthosite liquid at the borders (Buddington, 1968). Intrusion into the metasediments

may have taken place by the magma moving conformably along the gliding planes of the metasediments. Continual addition of magma then expanded the originally narrow channels until a large central body was formed that plunges to the north-northeast (Balk, 1930).

6.1.5.4 Petrography

Mineralogy and Textural Relations. The metanorthosite massif consists of two facies -- Marcy and Whiteface. The Marcy facies occurs in the core of the massif and is a bluish-gray, coarse- to very coarse-grained rock with less than 10% dark minerals. The very coarse plagioclase feldspar grains range from 2 cm (0.75 in.) to 8 cm (3 in.) long and lie in a matrix of granular material composed of uncrushed and subordinate crushed feldspar, the latter usually being lesser in amount than the former (Buddington, 1939). The Marcy facies shows indistinct or no foliation, though the feldspar grains may show a crude parallelism (Miller, 1918). Normative mineral composition for this facies is quartz, 1.26-1.92%; orthoclase, 5.56-7.23%; albite, 37.2-40.87%; anorthite, 42.15-49.76%; diopside, 0.65-3.72%; hypersthene, 0.4-4.6%; magnetite, 1.00-1.97%; ilmenite, 0-3.04%; calcite, 0-0.34%; spinel and garnet, 2.7-2.77%; and femics, 3.81-10.73%. The normative plagioclase makeup is Ab-44-50 (Buddington, 1939).

The Marcy facies gradually changes toward the periphery of the massif, where it grades into the Whiteface facies. This unit is a more gabbroic anorthosite marked by an increase in dark minerals. It is medium grained and generally foliated. The plagioclase feldspar grains are white, or nearly so, and are of medium granular texture. There are a few relict phenocrysts of blue-gray plagioclase, and the rock is interbanded and streaked. The range of normative mineral compositions of the Whiteface facies is quartz, 0-7.80%; orthoclase, 3.89-11.12%; albite, 24.89-42.44%; anorthite, 34.19-46.70%; diopside, 0.71-13.22%; hypersthene, 0.83-16.93%; olivine, 0-2.72%; magnetite, 0.81-4.89%; ilmenite, 0-3.12%; apatite, 0-0.60%; calcite, 0-1.10%; femics, 5.95-34.20%; and spinel and garnet, 0-2.95% (Buddington, 1939). Within the border facies a block structure is locally evident, that is, angular blocks of anorthosite enclosed by slightly more mafic anorthosite.

The smaller metanorthosite domes to the south show the same gradual change from a coarse metanorthositic core to a gabbroic (or noritic gabbroic) border in contact with charnockite (Krieger, 1937; Letteney, 1968; and De Waard and Romey, 1969). The metanorthosite core consists of bluish-gray, coarse-grained andesine set in a finer-grained (1-2-mm, or 0.03-0.08-in.) matrix of andesine, orthopyroxene, clinopyroxene, magnetite, biotite, and hornblende. Mafic minerals form about 1-10% of this rock. Foliation is absent or indistinct. Deformation has commonly produced a granulated texture. Away from the core, the rock becomes increasingly foliated and deformed, and the percentages of potassium feldspar, quartz (absent in the core), and mafic minerals increase with a decrease in anorthite plagioclase content.

Alteration. Alteration of the anorthosite consists of reduction in the grain size of the plagioclase, which is attributed either to protoclasis in the later stages of emplacement (Balk, 1931) or to later metamorphism (De Waard and Walton, 1967). Another alteration was the metamorphic reactions that produce garnet plus clinopyroxene and spinel from reaction between primary plagioclase, orthopyroxene, and iron-titanium oxide (McLelland and Whitney, 1980). Regional granulite-facies metamorphism resulted in 25-50% rounded and fractured plagioclase in a matrix of plagioclase and mafic minerals down to 1 mm (0.03 in.) in size. Potassium feldspar occurs as antiperthitic intergrowths or, locally, as interstitial grains. Mafic minerals present include hornblende, garnet, and augite. The garnet appears as thin corona reaction rims around orthopyroxene and iron-titanium oxide. Small porphyroblasts of garnet occur with an increase in metamorphic grade (Davis, 1968). Border facies rocks show extensive crushing and recrystallization, with garnet commonly present. Elsewhere the metanorthosite shows little or no alteration. The border facies also exhibit late saussuritization of the plagioclase and alteration of mafic minerals to chlorite, epidote, zoisite, and carbonate (Buddington, 1939). Rocks in the central and north-central Adirondacks have undergone deformation at greater temperatures and pressures than those of the northwestern Adirondacks (e.g., Buddington, 1953; Bohlen and Essene, 1979).

6.1.5.5 Secondary Intrusions

Metamorphosed gabbroic sheets and lenses were intruded into the roof zone of the metanorthosite as well as into the metasedimentary rocks in the roof zone. A few metagabbro dikes cut across the metanorthosite foliation; others are conformable with it. These metagabbros appear to be younger than the metanorthosite but older than the charnockitic gneisses (Buddington, 1939).

Concordant layers of mangeritic (syenitic) and charnockitic (granitic) gneisses border the metanorthosite and may be of igneous origin. In the Elizabethtown and Port Henry quadrangles, gneisses of this composition also occur within the foliated border of the metanorthosite. Syenite dikes cut the Marcy metanorthosite in the Long Lake, Lake Placid, and Schroon Lake quadrangles.

Postmetamorphic diabase dikes of Precambrian age also intrude the metanorthosite. Most strike within 20° of east-northeast and occupy fractures having the same strike (Buddington, 1939).

6.1.5.6 Country Rocks

The Adirondack anorthosite intrudes into polydeformed rocks of the metasedimentary sequence, the oldest known rock group in the region. These consist of calcite marble, local dolomitic marble, quartzite, gneiss,

amphibolite, migmatite, granulite, and local schist. The foliation is generally parallel to stratification surfaces.

Subsequent to emplacement of the anorthosite, syenitic magma was either intruded into the contact zone between the metasediments and the metanorthosite massif (Buddington, 1939) or generated there from preexisting quartzo-feldspathic gneisses by contact anatexis (McLelland and Isachsen, 1980). In some sections, a thin layer of Grenville rocks remains between the border metanorthosite and syenite gneiss; in other sections, the syenitic gneiss cuts the border facies and even occasionally reaches the Marcy facies. Where the syenitic gneiss contacts the metanorthosite, a transition rock, the Keene gneiss, may be present. It is a medium-grained, granular gneiss with a composition similar to that of the syenite but with scattered xenocrysts of bluish-gray labradorite, which appears to have resulted from partial assimilation of the anorthosite by the syenite magma (Buddington, 1939).

The rock surrounding the smaller metanorthosite domes consists of quartzo-felspathic gneiss, marble, pyroxenic gneiss, and quartzite similar to that surrounding the main massif (Letteney, 1968).

6.1.5.7 Rock Structure

The predominant jointing within the metanorthosite massif strikes north-northeast and is strongly developed. Unmetamorphosed diabase dikes of undetermined age locally occupy fractures with this strike. The few dikes that have been dated so far give K-Ar ages in the Early and Middle Cambrian and the Early and Middle Devonian (Geraghty et al., 1979). The north-northeast jointing is unaffected by the orientation of the local flow structure. East-west jointing is also present and occurs along the northern border of the massif (Balk, 1931). Linear valleys (lineaments) that trend north-northeast and east-northeast generally correspond with the axis of elongation of the Adirondack dome and have been found to be traces of high-angle faults and fracture zones (Isachsen and Wiener, 1981). Such fractures are concentrated in the eastern and central parts of the anorthosite massif (Isachsen, 1974).

In the region of the smaller metanorthosite domes, there are many linear topographic features, and most strike northeast (Isachsen and McKendree, 1977).

In the cores of the anorthosite domes, where the rocks are relatively undeformed, lamination can be seen as a relict magmatic texture (Letteney, 1968; De Waard and Romey, 1969). This texture is primary flow banding due to the parallel orientation of plagioclase grains during flow. Secondary foliation is much less common in the core and less distinct than that in the border Whiteface facies. This facies is nearly always foliated, with an orientation dependent on local factors. However, it does tend to dip outward,

indicating a domical structure (Balk, 1930; Buddington, 1939). Foliation tends to be strong along axial planes, and lineation tends to parallel the strike of the foliation, especially in the south (Buddington, 1939).

Overall, foliation in the border facies of the main massif dips outward, indicating a domical structure. Within the massif, foliation outlines several separate domes, with the most strongly foliated rocks in synclinorial sags (Buddington and Leonard, 1962). Balk (1930) and Buddington (1939) identified the major structural elements of the massif. The St. Regis-Marcy unit of the Marcy anorthosite in the St. Regis, Saranac Lake, Long Lake, Santanoni, Mt. Marcy, Schroon Lake, Paradox, Elizabethtown, Port Henry, Ausable Forks, and Willsboro quadrangles forms a major domical structure. In the eastern part, the so-called Westport structure in the Willsboro and Ausable Forks quadrangles forms another dome. Between the Westport dome and the Marcy/Port Kent metanorthosite structure to the north is a broad saddle of mixed Whiteface-facies anorthosite and belts of metasediments. Between the Westport and the St. Regis-Marcy units is another broad synclinorial belt of mixed metasediments, Whiteface-type, and Marcy-type rocks. The Jay-Mt. Whiteface metanorthosite belt in the Lake Placid and Ausable Forks quadrangles forms a thick sheet dipping to the south.

6.1.5.8 Geophysics

Where the metanorthosite massif crops out, an absence of seismicity in marked contrast to the rest of the Adirondacks was reported by Aggarwal et al. (1977).

Isachsen et al. (1979) concluded from aeromagnetic data that Snow and Gore mountains and Oregon dome are a buried single massif with a diameter of at least 55 km (34 mi). This conclusion is supported by the large gravity low in the area. The mangerite-charnockite domes and associated anorthosite may be an unroofed part of the massif (Isachsen et al., 1979).

A gravity survey of the Adirondack Mountains shows a well-defined anomaly of -12 mgal to -20 mgal over most of the anorthosite massif (Simmons, 1964). An average value for the density of the anorthosite is 2.72 g/cm^3. Since the steep gradients at the edge of the body indicate that the density contrast does not extend to great depth, the anorthosite is probably a layer. The modeled thickness of the layer is approximately 3 km (1.9 mi) in the west and 4.5 km (2.8 mi) in the east. Two local and more negative anomalies of circular shape are interpreted as vertical cylinders of anorthosite extending to a depth of 10-15 km (6.2-9.3 mi). These cylinders are about 6.4 km (4 mi) in diameter and 8-9 km (5-6 mi) long below the main body. Mann and Revetta (1980) found five gravity lows surrounded by gravity highs within the broad low of the massif. These have been interpreted as additional feeder dikes for the anorthosite. Gravity data also indicate that the massif probably extends at shallow levels 16-32 km (10-20 mi) to the north and northwest beyond the present exposure area.

Heat flow through the metanorthosite massif is 0.80 x 10^{-6} cal/cm^2•s at Saranac Lake and 0.81 x 10^{-6} cal/cm^2•s at Elizabethtown; standard-shield crust heat flow is about 1 x 10^{-6} cal/cm^2•s (Pollack and Roy, 1969). Because the metanorthosite is deficient in uranium, thorium, and potassium, its heat flux is about 25% less than that of most surface rocks. The heat flux in the metanorthosite would be reduced by 0.09 x 10^{-6} cal/cm^2•s centrally and 0.07 x 10^{-6} cal/cm^2•s near the boundary due to radioactive deficiency alone. Low thermal conductivity diverts the flux from below to the periphery, reducing the flow by 0.05 x 10^{-6} cal/cm^2•s centrally and 0.12 x 10^{-6} cal/cm^2•s near the boundary.

6.1.6 Olivine Metagabbros (Plate V, Pluton 2)

6.1.6.1 Size, Shape, and Location

Metagabbros occur in a number of locations on the immediate periphery of the metanorthosite massif. This area is near the center of the Adirondack dome, an area that has undergone several periods of igneous intrusion and subsequent deformation and metamorphism.

The largest volume of metagabbroic rocks lies in the northern part of the Elizabethtown (see Figs. 6.1 and 6.6) and Port Henry quadrangles; and continues into the Ausable Forks and Paradox Lake quadrangles to the north and south. This complex is a multiple intrusion composed of two large and several smaller intrusions that form a large basinlike structure. The Hoisington Mountain intrusion is about 12 km (7.4 mi) long and 2-3 km (1.2-1.9 mi) wide, cropping out from Pleasant Valley to Westport to the east. The Iron Mountain intrusion is to the west of Pleasant Valley (Gasparik, 1980). Another group of metagabbro intrusions occurs in the southern Adirondacks in the Newcomb, Schroon Lake, and Paradox Lake quadrangles (see Fig. 6.1). These intrusions occur as oval or round sill-like bodies 3-4 km (1.9-2.5 mi) in diameter (Whitney, 1973). A third intrusion is located near Jay Mountain in the Ausable Forks quadrangle. This body trends southeast-northwest and crops out over an area approximately 3 km (1.9 mi) by 10 km (6.2 mi). The exposed thickness of the body is at least 1000 m (3280 ft) (Whitney, 1972; Whitney et al., 1979). Together these rocks occupy about 89 km^2 (34 mi^2).

6.1.6.2 Relative and Absolute Age

In the Elizabethtown complex, four types of metamorphosed Precambrian rock were apparently intruded in the following sequence (early to late): granitic gneiss, anorthosite, garnet-pyroxene gneiss, and metagabbro (Gasparik, 1980). The metagabbro shows a chilled border against the granitic gneiss, which usually surrounds it; it is therefore younger than the gneiss.

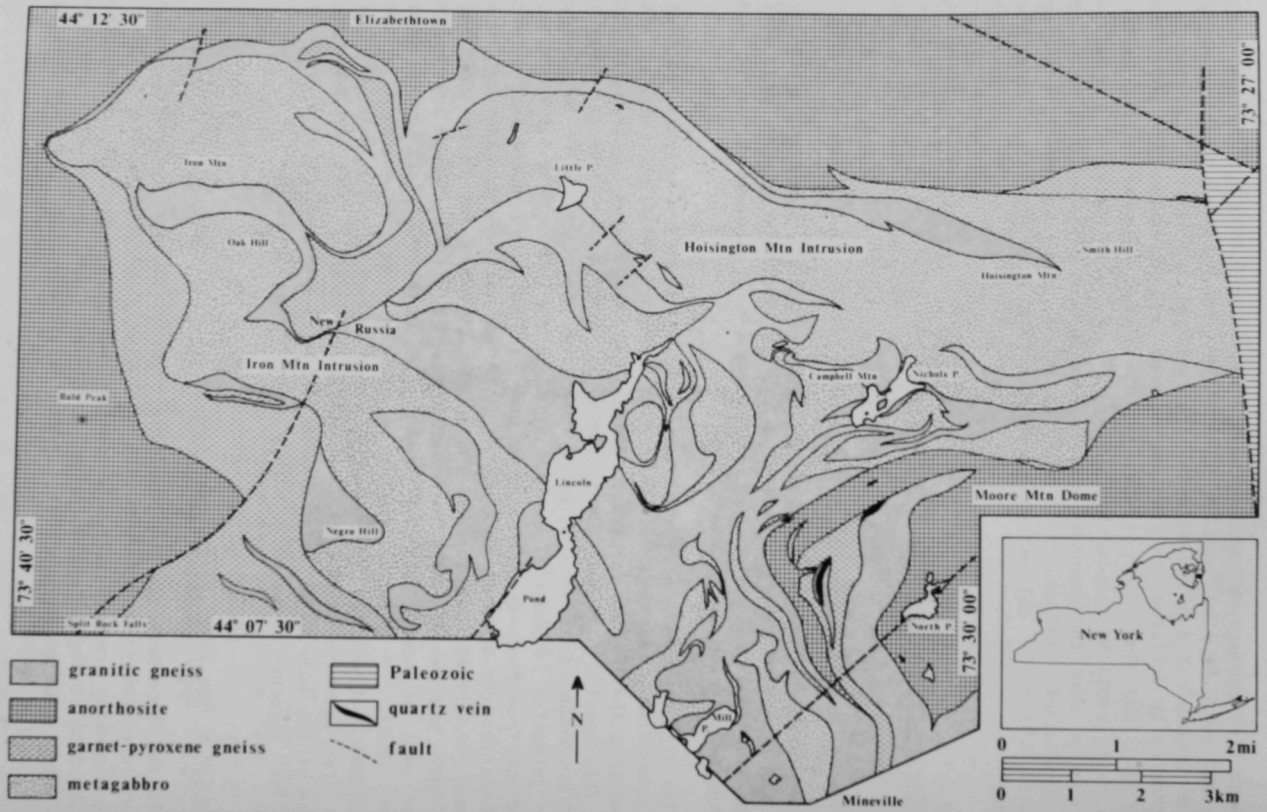

Fig. 6.6 Bedrock Geology of Area between Elizabethtown and Mineville, Eastern Adirondacks (Source: Gasparik, 1980)

The metagabbro rarely is in contact with the anorthosite or the garnet-pyroxene gneiss; their age relationships are therefore uncertain. Other investigators have indicated that the gabbros are younger than the anorthosite but older than the Diana syenite gneiss and the hornblende granite gneiss (Buddington, 1939).

6.1.6.3 Origin

Buddington (1939) interpreted the metagabbro bodies as plugs, stocks, dikes, and laccoliths. If the metagabbro is assumed to be the youngest intrusion in the Elizabethtown complex, it must have been intruded into the center of the granite gneiss, which still forms a roof complex over the metagabbro. The metagabbro is cut by dikes of its own composition, indicating emplacement during multiple periods of intrusion (Gasparik, 1980).

If the metagabbro is assumed to be younger than the syenites and granites, it may have been intruded as lenses or as a sheet within existing, somewhat folded, beds that were later intruded into by the syenites and granites.

6.1.6.4 Petrography

Mineralogy and Textural Relations. The Elizabethtown metagabbro has the composition of an alkali basalt, but differentiation on a scale from a few centimeters to a few meters (inches to several yards) appears to have led to two different mineral assemblages. Type A consists primarily of plagioclase and olivine, with or without ilmenite and clinopyroxene. Type B consists of plagioclase, ilmenite, and clinopyroxene. The rock generally has a subophitic texture, with grains ranging in size from very fine to coarse. Igneous textures and mineral assemblages have been preserved, although the rock has undergone granulite-facies metamorphism that produced garnet. A chilled margin is locally present.

The modal composition of the Type A metagabbro is plagioclase, 25.8-59.9%; olivine, 0-11.1%; primary clinopyroxene. 1.3-15.0%; metamorphic clinopyroxene, 5.0-14.3%; orthopyroxene, trace-3.4%; garnet, 0.4-20.3%; amphibole, 1.0-24.8%; biotite, 0-6.4%; opaque minerals, 0.4-8.5%; spinel, 0-1.0%; apatite, 0.1-5.3%; and serpentine, 0-9.6%. The modal mineralogy of the Type B metagabbro is plagioclase, 16.8-75.6%; primary clinopyroxene, 1.7-30.9%; metamorphic clinopyroxene, 1.5-12.1%; orthopyroxene, 0-2.5%; garnet, 4.0-21.5%; amphibole, 3.3-27.6%; biotite, 0.1-5.9%; opaque minerals, 2.0-8.6%; spinel, trace-0.8%; and apatite, 0.1-0.8% (Gasparik, 1980).

The metagabbros in the southern Adirondacks consist mainly of labradorite, augite, hypersthene, and garnet, with a small amount of olivine. Accessory minerals include amphibole, magnetite, spinel, biotite, and orthoclase or microcline. Along their margins these metagabbros become schistose

and grade into gabbro-amphibolites (Balk, 1932). In the Jay Mountain body, layering is present, striking northwest and dipping at a steep angle to the banding in the surrounding rocks.

Alteration. The metagabbro underwent granulite-facies metamorphism, which led to the formation of amphibole, biotite, garnet, clinopyroxene, orthopyroxene, and spinel, and caused albitization of plagioclase. Fine granular texture also resulted from this process. Reaction coronas occur around grains of olivine and ilmenite (Gasparik, 1980).

6.1.6.5 Secondary Intrusions

The metagabbro is commonly cut by dikes of its own composition.

6.1.6.6 Country Rocks

The metagabbros of the Elizabethtown complex are in direct contact with granitic gneiss, which shows hybrid zones along the contact. Rarely, the metagabbro is in contact with the garnet-pyroxene gneiss and the metanorthosite. The garnet-pyroxene gneiss forms seven sill-like bodies that are fine grained and foliated, and have granoblastic texture. They are usually in contact with the metanorthosite and may contain anorthosite xenoliths. The anorthosite contains phenocrysts of blue plagioclase in a granulated matrix of white plagioclase, commonly with garnet porphyroblasts (Gasparik, 1980).

6.1.6.7 Rock Structure

The Elizabethtown body is part of a large synformal structure. In the vicinity of Elizabethtown, there are several normal faults that trend northeast; displacements are small where they have been observed (Gasparik, 1980). In the Newcomb and Schroon Lake vicinity, fault traces also trend northeast and, where displacement direction can be determined, also are normal faults. Topographic linear features also trending northeast were observed in this region. No significant fractures were noted around Jay Mountain (Isachsen and McKendree, 1977). Foliation was developed as a result of deformation. Whitney et al. (1979) attributed the lineation to later folding events in the area, which was also further disturbed by vertical movements of the gabbros.

6.1.6.8 Geophysics

The metagabbro is closely enclosed on three sides by the metanorthosite massif, which has shown a marked absence of seismicity despite the occurrence of shallow earthquakes in much of the Adirondack region, mostly along north-northwest-trending faults that dip to the east (Aggarwal et al., 1977). An

earthquake swarm was observed in the Blue Mountain Lake area (Sbar et al., 1972), which is approximately 35 km (22 mi) southwest of the metagabbro that crops out in the Newcomb quadrangle.

The density of the metagabbro near Elizabethtown was found to be higher (3.0-3.4 g/cm^3) than that of either the anorthosite (2.8 g/cm^3) or the granitic gneiss (2.7 g/cm^3). The main volume of metagabbro in this area, as outlined by gravity studies, lies between New Russia and Mineville.

6.1.7 Stark Metasyenite Complex (Plate V, Pluton 4)

6.1.7.1 Size, Shape, and Location

The Stark metasyenite complex, composed of charnockitic, granitic, and quartz syenitic gneisses, crops out over an area approximately 95 km (59 mi) long and 3-45 km (2-28 mi) wide (Isachsen and Fisher, 1970). As the floor of the complex is not exposed, its thickness is unknown. It is located on the northern border of the Adirondack highlands in the Stark, Childwold, Nicholville, Santa Clara, Loon Lake, Chateaugay, Saranac, Lyon Mountain, and Lake Placid quadrangles (see Fig. 6.1). This area is part of the Adirondack dome and has been subjected to metamorphism and repeated igneous intrusion. Rocks of similar composition are also found in the southern Adirondacks and have been mapped in Plate V with the same pluton number as the Stark complex for the purposes of this report. All these rock bodies have a grossly similar lithology, as indicated in the classic studies by Buddington and his various coworkers since the early 1930s. There may, however, be local differences in the details of lithology and structure of the many rock bodies grouped together for the purposes of this report. Therefore, if more detailed information is needed about a particular body, the pertinent geologic quadrangle maps and reports by Buddington and his coworkers should be consulted.

6.1.7.2 Relative and Absolute Age

The Stark complex is of Precambrian age. Early investigators (Buddington, 1937 and 1939; Postel et al., 1956; and Buddington and Leonard, 1962) felt that the Stark syenite gneiss was intruded after the Adirondack metasediments and anorthosite were formed but before the granitic gneisses. Heath and Fairbairn (1968) calculated an isochron, based on seven Rb-Sr determinations from various Adirondack syenitic gneisses, that indicates an age of 1055 $\pm$ 30 m.y.

6.1.7.3 Origin

Working in the Santa Clara quadrangle, Buddington (1937) came to the conclusion that the magma of the Stark complex was intruded at moderate depth as a sill or a phacolith along the somewhat folded bedding planes of the Grenville series and along the planes of existing gabbro sills. He postulated that magma close to the wall rock chilled quickly to form a medium-grained pyroxene quartz syenite. The center differentiated as it cooled, producing one facies higher in silica and alkalies (hornblende quartz syenite) and another with less silica but more calcium, iron, and magnesium (pyroxene quartz syenite). He noted that a local, more mafic facies may be a relic of a relatively undifferentiated roof. Brock (1980) disagreed with the concept of a more mafic facies as a chill zone but agreed that it forms the roof of the complex. The entire complex was subsequently metamorphosed during the Grenville orogeny.

6.1.7.4 Petrography

Mineralogy and Textural Relations. The Stark syenite complex has been described as granitic to quartz monzonitic (Brock, 1980, Nicholville quadrangle) and hornblende syenitic to hornblende granitic (Buddington, 1937, Santa Clara quadrangle). Fresh surfaces are pink but locally green, and weather to brown. The textures include medium-grained (cataclastic) mortar, coarse-grained phacoidal (augen), flaser, and granoblastic textures. Although the grain size ranges from 0.25 mm (0.01 in.) to very coarse, the rock is commonly coarse grained and equigranular. The rock tends to be well foliated, as seen in the parallel orientation of quartz ribbons and in the phacoidal texture of feldspar grains.

The mineral composition also is variable, with rocks consisting of 26-63% microperthite, 6-38% oligoclase, 0-7% myrmekite, 8-39% quartz, 0-7% hypersthene, 0-8% augite, 0-11% hornblende, 0-1.8% biotite, 0.2-2.8% magnetite and ilmenite, with accessory sphene, apatite, zircon, and pyrite. Feldspars tend to account for 44-88% of the rock, quartz 8-39%, mafics 0-17%, and garnet 0-3% (Brock, 1980). The rock may have rapakivi texture consisting of 2-4-cm (0.8-1.6-in.) orthoclase grains rimmed with 2-3 mm (0.08-0.1 in.) of oligoclase. Quartz tends to be present as elongate leaves. Mafic minerals often have coronas of garnet and quartz.

The composition of the complex is zoned, with a more mafic periphery (10-20% quartz, 1-18% mafics) and a more felsic core (20-40% quartz, 1-6% mafics).

Alteration. Like all Adirondack Precambrian rocks, the Stark complex was metamorphosed during the Grenville orogeny. To the west the region underwent amphibolite-facies metamorphism, and to the east granulite-facies metamorphism. In the Nicholville quadrangle, metamorphic effects were most intense on the periphery of the complex. Mortar textures developed in the

core, progressing through augen and flaser to granoblastic on the periphery, reducing phenocrysts to a 0.04-mm (0.0016-in.) groundmass. Metamorphism caused symplectic intergrowths of garnet and quartz that form coronas around mafic minerals. Myrmekite formed from microcline, and feldspars changed from pink to green. Biotite and sphene formed by retrograde metamorphism. Weathering has produced clays, chlorite, calcite, hematite, and luxolite (Brock, 1980).

6.1.7.5 Secondary Intrusions

Brock (1980) described basalt dikes in the Nicholville quadrangle that cut all rocks except those forming the core of the Stark complex. The dikes are approximately 25 cm (10 in.) wide, with a dark aphanitic margin and a lighter core. They strike N. 60-80° E. and dip steeply. Brock also described dark to black metagabbro dikes that form narrow bands crossing the foliation of the Stark complex. Metagabbro and fine-grained unmetamorphosed diabase dikes also were noted by Buddington (1937) in the Santa Clara quadrangle. Postel et al. (1956 and 1959) described dikes of pink granite gneiss and quartz syenite; narrow pegmatite and diabase dikes in the Stark complex of the Nicholville quadrangle; and a quartz syenite dike in the Loon Lake quadrangle. The quartz syenite dike is dark gray-green, medium grained, and well-foliated. It is composed of microperthite, sodium plagioclase, quartz, pyroxene, and hornblende, as well as inclusions, schlieren, and layers of amphibolite.

6.1.7.6 Country Rocks

Brock (1980) concluded that the magmas of the Stark complex were intruded into what he termed a layered gneiss complex. This complex consists of a lower unit composed of a medium-grained, granoblastic, well-foliated gneiss (Buddington's St. Regis granite) and an upper unit with high microcline content (Buddington's alaskite, microcline granite, and granite gneiss). Postel et al. (1956 and 1959) and Buddington (1937) argued that these rocks were intruded after the Stark complex. In this view, the Stark magma was intruded into rocks of the Grenville series. The rock surrounding the Stark complex consists of a mix of both. Postel et al. (1959) described a giant migmatite complex in the Nicholville quadrangle that consists of bands 9-18 m (30-60 ft) thick of pink granitic gneiss and older metasediments. This unit consists of hornblende granitic gneiss, biotite granitic gneiss, alaskite, contaminated granitic gneiss, and plagioclase granitic gneiss, with scattered amphibolite layers. Postel et al. (1959) also described quartz syenite gneiss in the Loon Lake quadrangle that includes shreds and layers of Grenville metasediments.

6.1.7.7 Rock Structure

Topographic lineaments within the Stark complex tend to run northeast, although some run northwest or north-south (Isachsen and McKendree, 1977). In the southeastern section of the Nicholville quadrangle and in the northeastern corner of the Lyon Mountain quadrangle, lineaments correspond to the edge of the Stark gneiss outcrop (Isachsen and McKendree, 1977).

Two sets of minor faults are reported in the Nicholville quadrangle; one set strikes northeast and the other southeast (Brock, 1980). No major faults have been noted (Postel et al., 1956 and 1959; Isachsen, 1977). In the Nicholville quadrangle, steeply dipping joints are well developed, with one set striking north and the other east (Brock, 1980). In the Santa Clara quadrangle, jointing is predominantly N. 60-90° E. and N. 5° E. to N. 25° W., with dips usually greater than 70° (Buddington, 1937).

Secondary foliation due to the planar arrangement of streaks of dark minerals, flat quartz leaves, and tabular feldspars is present throughout the Stark complex. In the Nicholville quadrangle, the foliation strikes predominantly east with a near vertical dip (Postel et al., 1959). Lineation due to the alignment of phenocrysts and elongate quartz rods tends to parallel the fold axes of surrounding rocks, thereby forming rod or pencil gneiss in the axial region of the folds.

The Stark complex consists of superposed tight, isoclinal, often plunging or recumbent folds. In the Nicholville quadrangle, two folds are superposed. One trends northeast and is overturned and isoclinal, with its axial plane dipping southeast. The other is a broad, upright fold trending southeast (Brock, 1980). Multiple folding has produced a mass of domes and antiforms. In the Santa Clara quadrangle, the Stark complex consists of close-folded or isoclinally overturned anticlines and synclines (Buddington, 1937). In the Loon Lake quadrangle, the complex forms a domical structure composed of many tightly compressed, plunging isoclinal folds with steep dips (Postel et al., 1956).

6.1.7.8 Geophysics

The Adirondack Mountain region has been shown to be one of recurrent but weak seismic activity (Sbar and Sykes, 1977). The north flank of the Adirondack massif is particularly active, with earthquakes occurring at mid-crustal depths. Thrusting is the predominant fault mode, with maximum compressive stress at approximately N. 70° E. and horizontal. Earthquakes tend to occur on faults running north-northwest and dipping east.

Adirondack Mountains are delineated by an aeromagnetic high over rocks of the Stark and Diana complexes (Isachsen et al., 1979). Brock (1980) reported magnetic values for the Stark complex to be 4750-5250 gamma against a background magnetism in the Adirondacks of 5000 gamma. The Stark complex is associated with a negative gravity anomaly (Brock, 1980).

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7 CRYSTALLINE ROCKS OF MASSACHUSETTS

7.1 GRENVILLE AND AVALONIAN

7.1.1 Berkshire Massif (Plate VI, Pluton 1)

7.1.1.1 Size, Shape, and Location

The Berkshire massif forms the Berkshire highlands in western Massachusetts, west of the Connecticut River valley (see Fig. 7.1). The massif runs roughly north-south for about 70 km (43 mi) in Massachusetts and then continues into Connecticut. It is elliptical in shape. At its extreme northern end, it is only about 4 km (2.5 mi) wide, but it reaches a width of 26 km (16 mi) near the southern border of Massachusetts (Herz, 1961; Norton, 1975). Almost three-fourths of the massif lies in Massachusetts, where it occupies an area of approximately 896 km^2 (346 mi^2).

7.1.1.2 Relative and Absolute Age

The stratigraphic age of the Berkshire massif is Precambrian (Emerson, 1917; Billings, 1956). Its absolute age, based on U-Pb analysis on zircon, is 1000-1080 m.y. (Ratcliffe and Zartman, 1971 and 1976). Surrounding rocks have been affected by at least five, and possibly eight, phases of tectonism, three of them of Paleozoic age, with the earliest of these being Taconic. The last deformation may have been during the Triassic (Ratcliffe and Harwood, 1975; Ratcliffe and Zartman, 1976).

7.1.1.3 Origin

The Berkshire massif is composed of both paragneiss and orthogneiss, which were formed during the thermotectonic events of the Grenville orogeny from sedimentary and volcanic protoliths, respectively. The gneisses and syntectonic intrusions have been subjected to polyphase deformation and metamorphism. Paleozoic metamorphism and deformation produced complex fabrics, folds, and thrusts to form the allochthonous unit. Contacts exhibit thrusts and recumbent folds, which were cut by vertical faults during the postmetamorphic Triassic period (Norton, 1975). The metamorphic grade in the metasedimentary rocks decreases from sillimanite to staurolite to garnet away from the massif. Ratcliffe and Zartman (1976) considered the petrologic and structural characteristics of the gneiss sequence in the Berkshire massif to be similar to those of the Grenville metasedimentary sequence of the Adirondack Mountains.

7.1.1.4 Petrography

Mineralogy and Textural Relations. Ratcliffe and Zartman (1976) recognized 12 distinct lithologic units in the Precambrian gneissic rocks of the Berkshire massif. The mineralogy and textures of these gneissic rocks were produced through complex multiple stages of metamorphism, deformation, and retrogression during the Precambrian and Paleozoic eras. The massif is, therefore, lithologically heterogeneous. Only the major lithologic units are described here. The Berkshire orthogneiss (unit 6 of Ratcliffe and Zartman, 1976) is essentially a granite gneiss derived from a volcanic protolith of felsic composition. Its dominant minerals are pinkish microcline, oligoclase-andesine feldspars, quartz, and biotite-muscovite layers. The rock is homogeneously fine grained, but varies locally. It is generally massive but has interlayered amphibolites. The biotite-quartz-plagioclase paragneiss is commonly medium to dark gray, well banded, fine grained, and locally porphyroblastic. The biotite-quartz-plagioclase-feldspar paragneiss sequence is made of lithologic units (units 7 through 11 of Ratcliffe and Zartman, 1976) composed of calc-silicate rocks, amphibolites, and schistose rocks. The sequence is deficient in potassium feldspar, but contains sillimanite and kyanite. Mafic gneisses and amphibolites (cf. the diorites of Emerson, 1917) occur as interlayered masses and may represent metamorphosed mafic volcanics. They are essentially hornblende and plagioclase rocks with minor biotite and quartz. Layers of feldspathic granulite and biotitic gneiss are frequently present.

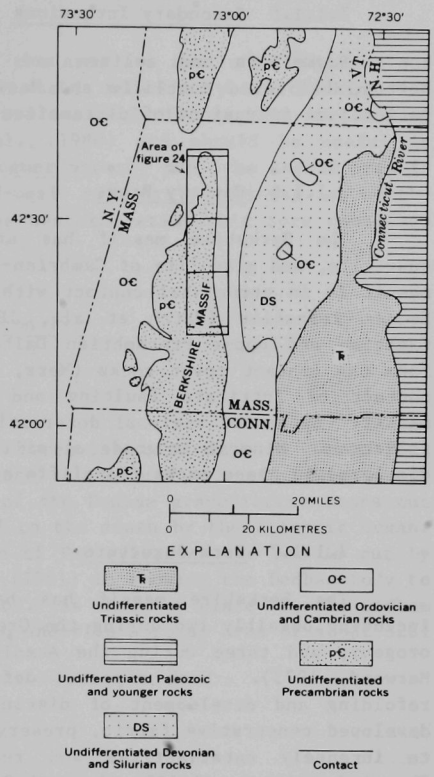


Fig. 7.1 Location of the Berkshire Massif (Source: Norton, 1975)

Alteration. Repeated deformation and metamorphism have greatly altered the state of the rocks. Synkinematic features are overprinted by high-grade thermal metamorphism, erasing traces of earlier and lower-grade static thermal events.

7.1.1.5 Secondary Intrusions

Pegmatites and aplites are later intrusive phases (Emerson, 1917; Norton, 1975; and Ratcliffe and Harwood, 1975). However, xenoliths, segregations, or intrusions of ultramafics occur in the massif.

7.1.1.6 Country Rocks

The Berkshire massif has structural contact with schist, gneiss, quartzite, and granulite of Cambrian-Ordovician age. On its eastern side, the massif is in structural contact with gneiss and schist of the Lower Cambrian Hoosac formation (Hatch et al., 1968). On its western side, the Lower Cambrian and upper Precambrian Dalton formation and the Cheshire quartzite form the contact cover rocks (Herz, 1961; Norton, 1975). These rocks show a contact of intricate faulting and recumbent folding with the overriding massif. Extensive regional deformation and metamorphism have imparted strong foliation. Mineral isograds of sillimanite, staurolite, and garnet are well displayed in these rocks (Ratcliffe and Harwood, 1975).

7.1.1.7 Rock Structure

The Berkshire massif has been affected by multiple deformations, including possibly two during the Grenville orogeny, three during the Taconic orogeny, and three during the Acadian orogeny (Norton, 1975; Ratcliffe and Harwood, 1975). Such repeated deformation caused large-scale folding and refolding and development of distinct thrust faults. Rocks display well-developed penetrative fabric, preserving all stages from incipient fracturing to intensely cataclastized and recrystallized (blastomylonitized) rocks. Shearing and recrystallization are most prominent near faults. Precambrian foliation and lineation are truncated by the unconformity beneath the Dalton formation and are overprinted by later penetrative fabrics.

A number of north-and-south-trending, low-angle thrust faults run through the massif. They are most evident near the border of the massif (see Fig. 7.2). The west edge of the massif, in western and southwestern Massachusetts, contains imbricate thrust slices of recumbently folded Precambrian gneiss and unconformable, younger metasedimentary rocks. The north end also displays thrust relations with the surrounding Ordovician and Cambrian metasediments to the east and Devonian to Cambrian rocks to the west (Norton, 1975). Recumbent folding contemporaneous with thrust faulting and cataclasis is also present. Mylonitization and blastomylonite features are present and associated with intense recumbent folding and thrust faults (Stanley, 1967; Norton, 1975; Osberg, 1975; and Ratcliffe and Harwood, 1975).

7.1.1.8 Geophysics

Airborne electromagnetic data for the area are available (Griscom and Bromery, 1968). The relief contrast is used selectively for marking the boundaries of the plutons. Gravity data are available for certain portions of the massif (Diment, 1968; Diment et al., 1980) and should be useful in defining the intrusives. On a simple Bouguer anomaly map, the massif area is marked by a gravity low (0- to -30-mgal contours) defining a north-northeasterly trend characteristic of the western metamorphic core complexes of the New England Appalachians.

7.1.2 Dedham Granodiorite (Plate VI, Pluton 19)

7.1.2.1 Size, Shape, and Location

The Dedham granodiorite forms an irregular body with at least four major outcrops in southeastern Massachusetts. It occurs as small bodies north and west of the Boston basin, but most of the Dedham granodiorite crops out south of the Boston basin and is bounded to the south by the Atlantic Ocean. It is bisected by the Narragansett basin of Pennsylvanian age and is cut by the Norfolk basin. Plate VI follows Zen (1981) in showing the Dedham body to underlie the surficial deposits on Cape Cod. As shown on Plate VI, the Dedham granodiorite crops out in or is assumed to underlie a total area of about 3261 km^2 (1259 mi^2).

7.1.2.2 Relative and Absolute Age

Whole-rock Rb-Sr analysis of the Dedham granodiorite indicated an age of 608 ± 17 m.y. (Fairbairn et al., 1967). The Dedham granodiorite is Precambrian, because it is comformably overlain by Lower Cambrian sedimentary rocks (Billings, 1929; Dowse, 1950).

7.1.2.3 Origin

The Dedham granodiorite was emplaced during the late Precambrian. Because the country rocks that it intruded have been eroded away, contact relationships are not available to allow determination of the mode of emplacement. However, due to its size and the absence of sedimentary rocks overlying it, it might be possible to interpret the pluton as a forceful emplacement, as opposed to a lopolith.

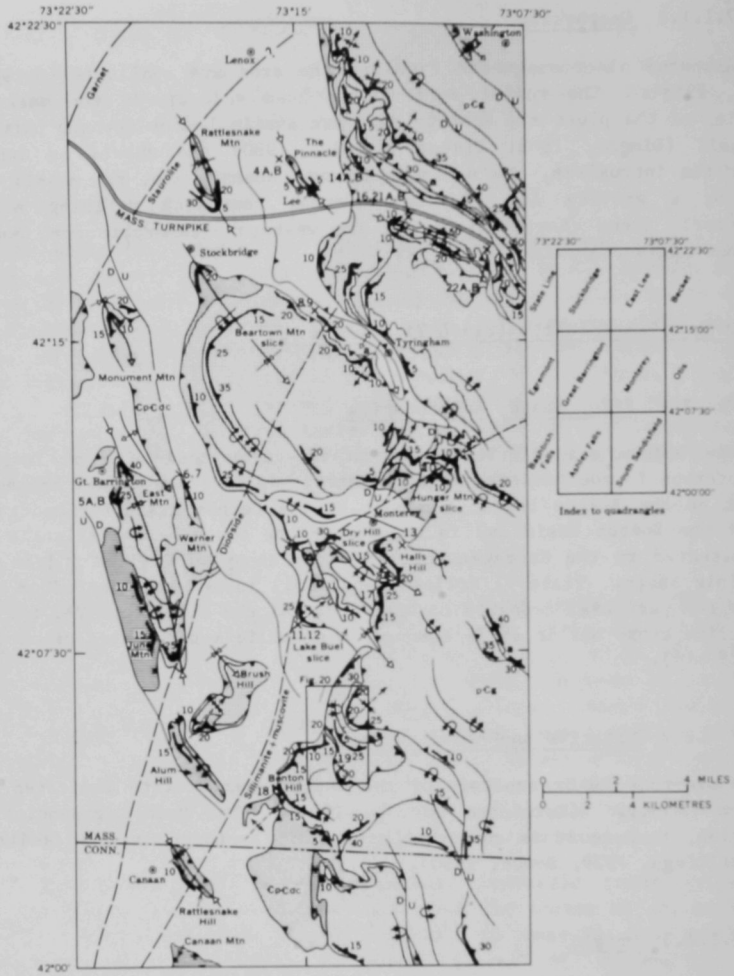


Fig. 7.2 Geologic Map of the Western Edge of the Berkshire Massif Showing Associated Major Thrust Faulting (axial surfaces of isoclinal, strongly overturned-to-recumbent folds with blastomylonitic foliation are also shown) (Source: Ratcliffe and Harwood, 1975)

EXPLANATION

g

Post-thrust granite

a

Pods of alaskite in faults

Rock has weakly developed blastomylonitic foliation. Occurrences north of Massachusetts Turnpike contain crosscutting seams of magnetite and magnetite-rich alaskite

UPPER ORDOVICIAN(?)

ORDOVICIAN(?)

ROCKS FOUND BENEATH THRUST FAULTS OF BERKSHIRE MASSIF

Rocks of the Dalton Formation, Cheshire Quartzite, Stockbridge and Walloomsac Formations, as parautochthonous slices or "autochthonous" rock beneath Berkshire massif. Unit contains allochthonous rocks of the Everett Formation in the northwest

UPPER PRECAMBRIAN TO ORDOVICIAN

UPPER(?) ORDOVICIAN

PRECAMBRIAN(?) TO ORDOVICIAN

ROCKS OF BERKSHIRE MASSIF

Muscovite, biotite, staurolite, quartz or muscovite, biotite, sillimanite schists of June Mountain and Canaan Mountain in thrust contact with other rocks

UPPER PRECAMBRIAN(?) AND LOWER CAMBRIAN

PRECAMBRIAN(?) AND CAMBRIAN

CpCdc

Dalton Formation and Cheshire Quartzite (unconformable cover rocks of the Berkshire massif)

UNCONFORMITY

pCg

Biotite, hornblende, quartz paragneiss, granitic gneiss, calc-silicate rocks, and amphibolite composing the crystalline rocks of the Berkshire massif

PRECAMBRIAN

Contact

Synmetamorphic thrust fault formed during thrusting of Berkshire massif. Sawteeth on upper plate

Low-angle synmetamorphic thrust fault of metamorphic "eugeosynclinal" rocks thrust westward during emplacement of the Berkshire massif. Sawteeth on upper plate

U

D

Postmetamorphic high-angle fault. U, upthrown side, D, downthrown side

Approximate location of isograd in Barrovian zonation, probably Acadian, postdating emplacement of Berkshire massif

25 40

Strike and dip of folded blastomylonitic foliation in zone where particularly well developed. Formed during overthrusting of Berkshire massif, inclined, overturned. Symbol is diagrammatic and applies to rocks above as well as beneath thrusts

Antiform Synform

Axial trace and direction of plunge of overturned fold formed during thrusting of Berkshire massif. F_3 folds of figure 2

Antiform Synform Overturned synform

Axial trace and direction of plunge of first major postthrust folds, with axial surface slip or crenulation cleavage. F_4 folds of figure 2

Antiform Synform

Axial trace and direction of plunge of second major postthrust fold, with axial planar slip or crenulation cleavage. F_5 folds of figure 2

×5A,B

Collection

Locality of specimen figured in text

Fig. 7.2 (Cont'd)

7.1.2.4 Petrography

Mineralogy and Textural Relations. The Dedham granodiorite represents a compositional range from alkalic granite, through granodiorite, to granite and granite porphyry. It is a massive, relatively homogeneous, medium- to coarse-grained rock composed of quartz, 25-30%; plagioclase feldspar, 40-45%; and microcline-microperthite, 25-30%. Apatite, titanite, and magnetite are accessory minerals. A light-gray variety similar in texture and composition to the pink Dedham granodiorite has been reported to grade into the pink facies (Chute, 1950).

Chute (1950) believed that the light-gray rock is equivalent to the pink Dedham granodiorite into which it grades; the pink feldspar is thought to have been bleached by emanations from the Dedham granite. The porphyritic variety of Dedham granodiorite contains pink feldspar phenocrysts 2-3 mm (0.5-0.75 in.) in length. It is scattered throughout the Dedham granodiorite and is thought to be comagmatic. This rock contains about 25% quartz, 50-55% plagioclase feldspar, and 20-25% microcline, which occurs mostly as phenocrysts. This porphyry grades into the Dedham granodiorite by a gradual disappearance of the phenocrysts. A fine-grained albite granite exists within the Dedham granodiorite as dikes.

Alteration. The Dedham granodiorite plagioclase feldspar is altered to sericite and zoisite. Epidote is also a common secondary mineral. Biotite has been altered largely to chlorite. Secondary minerals in the albite granite are biotite, sericite, chlorite, and zoisite. The riebeckite granite shows little alteration.

7.1.2.5 Secondary Intrusions

Dikes of fine-grained albite granite are present in the pink Dedham granodiorite and in the light-gray variety as well. The granite contains about 30% quartz, 55% albite, 10% microperthite, and 5% accessory minerals. This granite has been mapped as small bodies in the Brockton quadrangle (Chute, 1950). A porphyritic riebeckite granite intrudes into the Dedham granodiorite as stocks. The margins of the stocks show finer grain size than the center. This rock contains about 30% quartz, 65% albite, and about 5% riebeckite; the quartz and riebeckite form the larger crystals and phenocrysts. The riebeckite granite is not thought to be comagmatic with the Dedham granodiorite and has been correlated with the Quincy rocks. Fine-grained rhyolite and diabase dikes have been reported in granodiorite north of the Narragansett basin.

7.1.2.6 Country Rocks

The Salem gabbro-diorite, which is thought to be coeval with the Dedham granite, crops out north of the Boston basin and is separated from the main body of Dedham granodiorite, which is south of the Boston basin. Secondary minerals within the Salem gabbro-diorite are chlorite, epidote, sericite, and zoisite. The original feldspars in this unit have been so intensely altered that they are difficult to identify (Toulmin, 1964). As many as eight varieties of granodiorite have been mapped. They vary from coarse to fine grained and all have been highly altered. The Pennsylvanian Wamsutta and Rhode Island formations unconformably overlie the Dedham granodiorite in Rhode Island and southeastern Massachusetts. The Wamsutta formation is composed of reddish, fine-grained quartzitic sandstone with interbedded red slabby shale and gray granule and pebble conglomerate. The typical red sandstone contains 90-95% quartz, 5-10% sericite, and about 1% epidote, titanite, and opaque minerals.

7.1.2.7 Rock Structure

The northern area of the Dedham granodiorite is an imbricate structure characterized by steep dips and thrust faults. Deep-seated thrust faults were reported by Billings (1929). Shear fractures have been reported, with dips varying from 40° to vertical (Chute, 1950). The recent state geological map shows a large number of normal, northeasterly and westerly faults in this area (Zen, 1981). The Bloody Bluff fault zone is the major regional fracture feature. Foliation in the Dedham granodiorite developed parallel to the fault, but was distorted on a small scale (Smith and Barosh, 1983).

7.1.2.8 Geophysics

Part of the area is shown as a gravity low on the Bouguer anomaly map of Diment et al. (1980). The moderate seismicity reported for the area conforms to the general northwesterly distribution of seismicity in the New England states (Sykes, 1978; Diment et al., 1980).

7.1.3 Massabesic Gneiss (Plate VI, Pluton 11)

See Sec. 4.1.1 and Plate III, pluton 20. Only 2 km^2 (0.8 mi^2) of this pluton appears in Massachusetts; most of the body lies in New Hampshire, and therefore it is described with the plutons of that state.

7.1.4 Milford Granite (Plate VI, Pluton 17)

7.1.4.1 Size, Shape, and Location

The Milford granite appears as an irregular body that extends 36 km (22 mi) from Westboro and Southboro to the Massachusetts border, and beyond to Cranston, R.I. Its outcrops are found over an area of about 355 km^2 (137 mi^2) in Massachusetts.

7.1.4.2 Relative and Absolute Age

The Milford granite has been assigned a Precambrian age of about 600-650 m.y. on the basis of analyses of U-Th-Pb ratios in zircon (Zartman and Naylor, 1972; Barosh et al., 1977). This is in comformity with Naylor's (1975) Pb^{207}/Pb^{206} age of 630 m.y.

7.1.4.3 Origin

The Milford granite was emplaced during the late Precambrian. It intrudes into, and contains as roof pendants, some quartzites and gneisses from the Blackstone Group and the Westboro quartzite (Emerson, 1917; Barosh et al., 1977). The granite also intrudes into the Ponaganset gneiss, which is equivalent to parts of the Northbridge granite gneiss (a usage now abandoned [Zen, 1981]). The Bellingham conglomerate (Pennsylvanian, Cambrian, or Proterozoic) lies unconformably over the Milford granite. The contact phase of the granitic magma is shown by a dark hornblendic border, which is described as the Ironstone quartz diorite.

7.1.4.4 Petrography

Mineralogy and Textural Relations. The Milford granite is a light pinkish-gray to medium-gray granite. It is fine to coarse grained, equigranular, and locally porphyritic. The chief minerals are quartz, microcline-perthite, albite-oligoclase, biotite, and muscovite. Apatite, magnetite, garnet, fluorite, sphene, and zircon occur as accessory minerals. The rock is chiefly quartz monzonite, but ranges from granite to granodiorite (Nelson, 1975). Associated small bodies of gabbro and diorite probably were comagmatic with the Milford granite (Barosh et al., 1977).

Alteration. Epidote, chlorite, sericite, and calcite are secondary minerals formed by the alteration of feldspars and mafic minerals. Large dikelike masses of diorite appear in isolated patches. Some are surrounded by a broad zone of aplite, which is interpreted as portions of the peripheral zone spared by erosion.

7.1.4.5 Country Rocks

To the east of the Milford granite is the Dedham granodiorite. The Andover granite (Silurian or Ordovician) lies to the north in the Nashoba zone, and intrudes into sillimanite schists and gneisses. The Milford granite intrudes into the Ponaganset and related gneisses (e.g., the Hope Valley Alaskite) (Zen, 1981). The Quincy granite at Rockport has nearly the same chemical composition as the Milford granite (Emerson, 1917).

7.1.4.6 Rock Structure

Jointing is closely spaced and commonly parallel to the foliation. Numerous faults are shown on a recent geological map of the state (Zen, 1981). The Milford granite displays a platy flow structure. The rock breaks easily into northeast-to-southwest or east-to-west planes. However, the foliation, which may be well developed or absent, strikes northwest, and does not coincide with the general parting tendency of the rock (Balk 1959). The direction of maximum compression, therefore, may have changed after the development of flow structure.

7.1.4.7 Geophysics

The area is enclosed by a gravity low on a Bouguer anomaly map (Diment et al., 1980).

7.1.5 Mt. Holly Complex, Southern Exposure (Plate VI, Pluton 2)

See Sec. 5.1.3 and Plate IV, pluton 2. This body occupies only 9 km^2 (3.5 mi^2) in Massachusetts. Most of this complex occurs in Vermont, and it is therefore described with the plutons of that state.

7.1.6 Pelham Dome (Plate VI, Pluton 7)

7.1.6.1 Size, Shape, and Location

The Pelham dome of western Massachusetts is one of about 20 domes that occupy a belt stretching from Long Island Sound to northeastern New Hampshire. This dome is about 13 km (8 mi) wide and about 45 km (28 mi) long with a strike of north to south (Emerson, 1917; Robinson, 1967). It occupies an area of about 372 km^2 (144 mi^2). The geological setting of the Pelham Dome is shown in Fig. 7.3.

A thickness of 2130 m (7000 ft) of the Pelham gneiss is exposed in the vicinity of the town of Wendell. Whether this thickness is due to repetition

by isoclinal folding or intrusion of granite layers has not been determined. The true thickness is unknown.

7.1.6.2 Relative and Absolute Age

A late Precambrian age for the core of the Pelham dome was determined by Naylor et al. (1973). Within the Pelham dome, the Dry Hill gneiss occurs mainly in the center of the dome, and the Poplar Mountain gneiss makes up most of the outer portion near contacts with younger formations (see Fig. 7.3). Exposures of Pelham gneiss in the core of the Pelham dome are very limited or absent. Zircons separated from a sample of Dry Hill gneiss yielded a nearly concordant U-Pb age of 565 ± 30 m.y. (Robinson, 1967).

7.1.6.3 Origin

The Poplar Mountain gneiss and the Dry Hill gneiss exposed in the core of the Pelham dome are interpreted as metamorphosed felsic volcanic rocks, volcanogenic sedimentary rocks, quartzose sandstones, calcareous sandstones, and shales (Robinson, 1967). Overlying these upper Precambrian rocks and underlying the Middle Ordovician rocks is a sequence of plagioclase gneisses and amphibolites of uncertain age. Some pelitic schists within the upper Precambrian section of the Pelham dome contain armored relics of garnet, sillimanite, and orthoclase, indicating a pre-Devonian, possibly late Precambrian, metamorphism of sillimanite-orthoclase grade (Robinson et al., 1975).

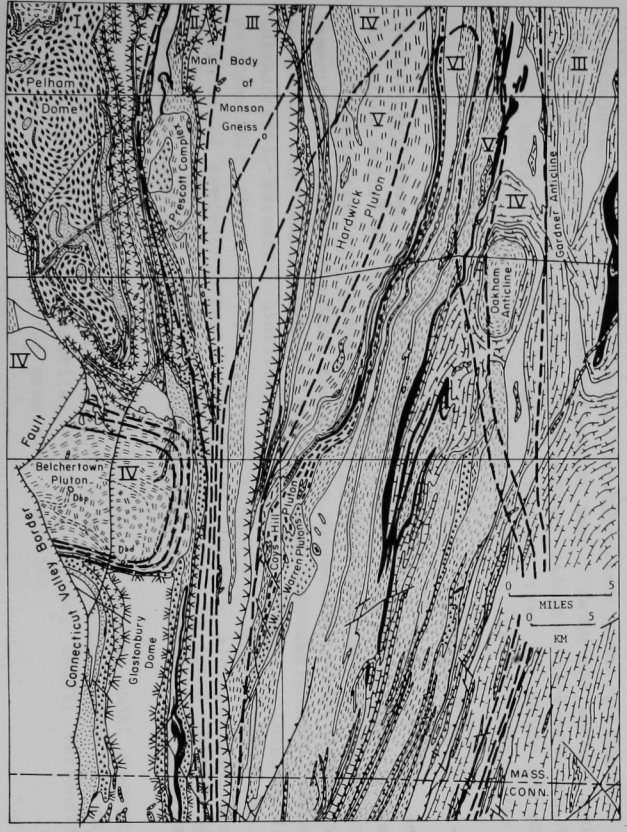
7.1.6.4 Petrography

Mineralogy and Textural Relations. The main body of the Pelham dome occurs in the core and is called the Dry Hill gneiss. This rock is chiefly a fine-grained, moderately foliated, gray-weathering, faintly bedded quartz-rich biotite gneiss with white microcline megacrysts up to 15 cm (6 in.) in diameter. The borders of these megacrysts are granulated. Commonly interbedded with this gneiss is well-bedded quartzite with or without actinolite and banded microcline-biotite-hornblende gneiss. At some localities, the quartzite is interbedded with medium- to coarse-grained epidote amphibolite.

Alteration. The essential minerals of the rock have been moderately altered.

7.1.6.5 Secondary Intrusions

Interlayered with the Dry Hill gneiss are sharply defined zones of Poplar Mountain gneiss, which range from a few feet to tens or hundreds of



Jurassic
Jurassic-Triassic
- Diabase dikes.
- Conglomerate, sandstone, shale, and basalt.

- Foliated muscovite granite gneiss, biotite gneiss.
- Coys Hill Porphyritic Granite.
- Granodiorite, tonalite.
- Gabbro, diorite.

Lower Devonian
- Erving Formation.
- Littleton Formation. Volcanics where separately mapped.

Silurian
- Fitch Formation*
- Clough Quartzite.

(Subzones A and B)

Paxton Formation
- Spss-Sulfidic Schist Member
- Sp-Granulite Member
- Spsq-White Schist Member, Spqr-Quartzite-Rusty Schist Member

(Subzones C and D)

Middle Ordovician
- Partridge Formation. Separately mapped felsic volcanics in different pattern.
- Ammonoosuc Volcanics in Subzone A. Also includes separately mapped mafic volcanics in Partridge Formation in Subzones B and C.

Ordovician
- Massive gneiss of plutonic derivation in dome core.

Ordovician? or older.
- Layered gneiss, Monson Gneiss, Fourmile Gneiss.

Late Precambrian
- Poplar Mountain Gneiss, Mt. Mineral Formation.
- Poplar Mountain Quartzite, Pelham Quartzite.
- Dry Hill Gneiss.

- Mesozoic normal fault, hachures on downthrown side.

* Fitch Formation shown solid black in Subzone B.

Fig. 7.3 Geologic Map of the Pelham Dome (Source: Robinson et al., 1982a)

feet thick. The Poplar Mountain gneiss is composed of moderately foliated, medium- to coarse-grained, white to pink, microcline-biotite gneiss, commonly with widely spaced black hornblende knots or magnetite grains. The thickest of these layers were mapped separately by Balk (1957). Although many of these layers may be intrusive sills, there is no conclusive field evidence to support this claim.

The Dry Hill gneiss in the outer part of the core of the Pelham dome is finer grained than that in the inner portion and consists of medium- to fine-grained, well-foliated layered gneisses, including gray-weathering, biotite-rich gneisses with or without hornblende and yellow-weathering muscovite-biotite gneisses. Megacrysts of microcline are present. Isolated stubby prisms of black hornblende and small garnets up to 2 mm (0.08 in.) in diameter are common. There are a few amphibolite- or biotite-rich bands. Quartzite and quartzose gneiss are unknown.

7.1.6.6 Country Rocks

The country rock to the north, east, and south of the dome consists of Ordovician(?) to upper Precambrian(?) gneisses of uncertain age. To the northeast are the Clough, Fitch, Littleton, Erving, Gile Mountain, and Waits River formations of Lower Devonian-Silurian age. To the west of the dome are the Ammonoosuc volcanics and Partridge formation, as well as the Connecticut Valley border fault, which separates the dome from the Lower Jurassic-Upper Triassic sedimentary formations. Other features within reasonable proximity of the dome include the Warwick dome to the northeast, the Prescott granitoid intrusion and the Hanson gneiss to the east, and the Belchertown intrusive complex to the south.

7.1.6.7 Rock Structure

The only faulting in the immediate area is the Connecticut Valley border fault, which is adjacent to the Pelham dome. In the Pelham gneiss of the Pelham dome, foliation and bedding or layering are parallel except in the few places where there are complex, overturned folds in bedding. Outcrops typically have closely spaced jointing parallel to the foliation.

7.1.6.8 Geophysics

The gravity model of Diment et al. (1980) shows an area enclosed by Bouguer anomaly contours of 0 to -10 mgal, which roughly outlines the gneiss dome.

7.1.7 Salem Gabbro-Diorite (Plate VI, Pluton 18)

7.1.7.1 Size, Shape, and Location

The Salem gabbro-diorite is a northeasterly trending, elongated igneous intrusion primarily in central and southern Essex County. About 30 km (19 mi) long, it extends southwestward into Arlington, Lexington, and Waltham quadrangles, southeastward to the coast from Salem to Swampscott, and north of Ipswich. Outcrops occur in an area of about 290 km^2 (112 mi^2).

7.1.7.2 Relative and Absolute Age

A whole-rock Rb-Sr age of 460 ± 12 m.y., and a K-Ar-biotite age of 414 ± 12 m.y., have been determined for Salem gabbro (Zartman and Marvin, 1971). Zartman and Naylor (1972) reported a K-Ar-hornblende date of 640 m.y. for a coarse-grained, lithologically similar quartz diorite; it is therefore possible that igneous activity occurred in eastern Massachusetts in the late Precambrian. Zen (1981) assigned a Proterozoic Z age to the Salem gabbro-diorite.

7.1.7.3 Origin

Chute (1966) proposed that the Salem gabbro-diorite is of magmatic origin and has undergone crystallization and differentiation that culminated in a pegmatitic phase. However, Boutilier (1963 and 1964) observed relict volcanic textures in Salem rocks in the area from Medford to the Merrimack River, which led him to conclude that these rocks represent metalavas derived by a transformation of the Lynn volcanics.

7.1.7.4 Petrography

Mineralogy and Textural Relations. The Salem hornblende gabbro is greenish-gray to medium-dark gray, fine to medium grained, and generally massive. The chief minerals are hornblende, plagioclase (andesine to labradorite), pyroxene, biotite, spherite, and magnetite. Locally, lenses of a pegmatitic phase are present and grade into a medium-grained diorite composed of randomly oriented hornblende (10-15%), albite (70%), and quartz (5%). Boutilier (1964) considered the block-faulted gabbro-diorite phases as metalavas.

Toulmin (1964) gave the following ranges for the modal composition for intrusive rocks of the area: quartz, 0-5%; plagioclase, 50-75%; augite, 0-15%; hornblende, 2-30%; biotite, 0-10%; chlorite, 0-15%; carbonate, 0-1%; apatite, trace-1%; and opaque minerals, trace-5%.

Alteration. In the Salem gabbro-diorite, pyroxene is altered to hornblende, which in turn is altered to biotite (Nelson, 1975). Most of the plagioclase is highly altered to an aggregate of sericite, clay minerals, and calcite (Toulmin, 1964). In zoned crystals of plagioclase, calcic cores were converted more completely to sericite and clay than were the sodic rims. Thin layers of leucoxene (and possibly sphene) occur along the cleavage of biotite in the Salem gabbro-diorite, as a result of alteration. In the pegmatitic Salem gabbro-diorite, hornblende cores are altered to epidote and chlorite. In the Norwood quadrangle, calcic plagioclase in the Salem gabbro-diorite has been saussuritized, and altered to albite and sodic oligoclase.

7.1.7.5 Secondary Intrusions

In Medford, the Reading granite intrudes into the metalavas. Shearing and undulatory extinction is well developed in the feldspar. Dikes of porphyritic microgranite, syenite, and lamprophyre cut the diorites near Salem (Zartman and Marvin, 1971). Toulmin (1964) found that pegmatite dikes and pods made of pink feldspar, quartz, and epidote are common; calcite pods occur but are rare. Basalt and diabase dikes cut the Salem gabbro-diorite abundantly.

7.1.7.6 Country Rocks

The Salem gabbro-diorite lies in the Milford-Dedham zone. To the east is the Atlantic Ocean. To the north-northwest lies the Nashoba zone. Towards the south is the Northern Boundary Fault and the Boston basin.

The Salem gabbro-diorite intrudes into the Marlboro formation and is unconformably overlain by the Lynn volcanics. Altered contact relations exist between the Salem gabbro-diorite, the Dedham granodiorite, and the Westwood granite.

7.1.7.7 Rock Structure

The Salem gabbro, between Salem and Lynn, is highly and complexly jointed, with minor faults and shear zones cutting the rock. Locally, small-scale foliation occurs in association with the shear zones. Schlieren masses of diorite are found in the granitic phase of the Salem quartz diorite. In some places, the Salem rock is gneissoid or foliated, and hornblende is so abundant that the rock becomes schistose. The Clinton-Newbury and Bloody Bluff faults are major fracture features in the general region.

7.1.7.8 Geophysics

Section 7.2.4.8 would apply to the Salem Gabbro-Diorite as well.

7.1.8 Sterling and Ponaganset Gneisses (Plate VI, Pluton 16)

See Sec. 8.1.3 and Plate VII, pluton 3. Although outcrops of these rocks are found over an area of about 306 km^2 (118 mi^2) in Massachusetts, they occupy an even larger area in Rhode Island and are therefore described with the plutons of that state.

7.2 TACONIC

7.2.1 Andover Granite (Plate VI, Pluton 15)

7.2.1.1 Size, Shape, and Location

The Andover granite occurs as three disconnected bodies, elongated northeasterly, that crop out over an area of about 237 km^2 (92 mi^2) in northeastern Massachusetts. These bodies trend northeast between the Clinton-Newbury fault zone to the north and the Bloody Bluff fault zone to the south. The Dracut diorite, the Worcester phyllite, and the Merrimack group lie to the north of the Clinton-Newbury fault zone.

7.2.1.2 Relative and Absolute Age

Because the Andover granite cuts every rock type with which it is in contact, it is probably the youngest intrusive in the area. However, it may be slightly older than the Sharpners Pond diorite, which lies to the east (Castle, 1965). Handford (1965) reported a whole-rock Rb-Sr age of 450 ± 22 m.y. for the Andover granite.

7.2.1.3 Origin

Koza and Wroe (1974) proposed that the Andover pluton, due to its sheetlike lithologies, may have formed by "autochthonous transportation of a geosynclinal sequence rather than the allocthonous intrusion of granite magma."

7.2.1.4 Petrography

Mineralogy and Textural Relations. The Andover granite is predominantly a white, gneissose or banded, peraluminous granite-adamellite. It interfingers complexly with gneisses to the west and gabbro-diorite to the east. The texture ranges from typically granitic, medium grained, and allotriomorphic, through typically pegmatitic, extremely coarse grained, and locally hypidiomorphic (Dennen, 1943; Castle, 1965). Textural and

lithological changes are most noticeable perpendicular to the regional strike. Koza and Wroe (1974) divided the unit into five parallel lithic zones (probably representing sheets) of (1) augen gneiss, (2) a metasediment, (3) leucocratic granite with subordinate pegmatite and quartz monzonite, (4) granodiorite, and (5) pegmatite and subordinate granite. The contacts within zones are vertical.

Alteration. Although gneissic, the Andover granite is only slightly altered. Since the original minerals are unchanged, the gneissic texture is considered to be primary and partly the result of orogenic movement during intrusion of the granite.

7.2.1.5 Secondary Intrusions

Since the Andover granite intrudes into all the surrounding rocks, it is considered to be the youngest intrusion. No secondary intrusions have been recognized.

7.2.1.6 Country Rocks

The Andover granite intruded into the Merrimack quartzite. On the south side, it is in contact with the biotite-hornblende tonalite facies of the Sharpners Pond tonalite, undivided metasedimentary and metavolcanic rocks. Several small stocks intruded into gabbro-diorite in the area (Clapp, 1921).

7.2.1.7 Rock Structure

The Andover granite lies between the northeast-striking Clinton-Newbury fault zone to the north and the Bloody Bluff fault zone to the south. Extensive mylonitic zones are associated with these fault zones. Regional relations suggest that the Bloody Bluff fault zone forms the western border of the southeastern mobile belt -- the Avalonian belt -- in eastern Massachusetts. Regional considerations also strongly suggest that the Bloody Bluff fault zone and the related Clinton-Newbury fault form an integral part of a major fracture system extending from Connecticut to the Gulf of Maine and possibly into New Brunswick (Nelson, 1976).

The country rock into which the Andover granite is intruded is folded. This folding appears to have been taking place at the time of intrusion (Emerson, 1917).

7.2.1.8 Geophysics

The pattern of mapped bedrock geology in northeastern Massachusetts bears a striking similarity to lineaments on aeromagnetic maps of the area. This correspondence is especially well shown northwest of Boston between the Clinton-Newbury and Bloody Bluff fault zones. Along the Clinton-Newbury fault zone, a zone of northeast-trending positive anomalies lies to the south, which contrasts with the broad magnetic low on the north side of the fault (Alvord et al., 1976).

The area of the rock body is characterized by broadly spaced Bouguer anomaly contours (0 to +10 mgal) (Diment et al., 1980), indicating association with mafic rocks.

7.2.2 Ayer Intrusive Complex (Plate VI, Pluton 12)

7.2.2.1 Size, Shape, and Location

The Ayer intrusive complex is an elongated, discontinuous belt of rocks in east-central Massachusetts, on the west side of the Clinton-Newbury fault zone in the Merrimack synclinorium. The most recent division of the complex is into three units: (1) a facies that ranges from granite to tonalite, (2) the Clinton facies, and (3) the Devens-Long Pond facies (Zen, 1981). Five separate bodies in the northern part of the state stretch for about 50 km (31 mi) in Massachusetts and extend on into New Hampshire; across a 20-km (12-mi) gap to the south, another elongated area of outcroppings appears and extends 20 km (12 mi) to the Connecticut border. The complex extends over an area of 203 km^2 (78 mi^2) in Massachusetts; it crops out primarily in the Ayer, Clinton, Hudson, and Westford quadrangles. Skehan et al. (1976) stated that an intrusion of three plutonic bodies (the Ayer granodiorite, the Rattlesnake Hill muscovite granite, and the Andover granite) have a collective thickness of 1422 m. Individual thicknesses are not known. Barosh (1977) stated that the Long Pond gneiss member has a maximum thickness of 1160 m.

7.2.2.2 Relative and Absolute Age

The age of the Clinton facies is uncertain. On the basis of lithostratigraphic correlations, Gore (1976) proposed a Devonian age for the Clinton facies and a pre-Silurian age for the Devens-Long Pond facies, as it is intruded into by the Clinton facies. On the basis of U-Th-Pb analyses of zircon, Peck (1976) reported an age of 435 ± 10 m.y., which is in conformity with the statigraphic correlation. However, Zen (1981) designated both facies as at least of Lower Silurian age on the state geologic map, and Devens-Long Pond as perhaps Upper Ordovician.

7.2.2.3 Origin

Because of the local discordant contacts of the Clinton facies and the extension of its apophyses into the country rock, the presence of sharply bounded xenoliths, the fine-grained and chilled contact relations, the evidence of contact metamorphic effects in the country rock, the textural and mineralogic homogeneity, and a chemical composition similar to that for the low-temperature trough in the albite-orthoclase-silica-water system, Gore (1976) suggested that the Clinton facies has a plutonic origin. Gore also suggested that the Devens-Long Pond rocks are a metamorphic product of a layered sequence, which is reflected in the variations in texture and composition of the rocks.

7.2.2.4 Petrography

Mineralogy and Textural Relations. The granite-to-tonalite rocks of the Ayer complex are partly porphyritic, locally gneissic, and locally muscovitic (Zen, 1981).

The Clinton facies ranges from light to medium gray, green-gray, or dark gray; the rocks are a porphyritic, coarse-grained, slightly to moderately foliated biotite granite to quartz monzonite, with a nonporphyritic border phase. The border phase is fine to medium grained and forms a lit-par-lit complex with the porphyritic phase (Barosh, 1977; Zen, 1981). The major minerals are plagioclase (albite and sodic oligoclase), quartz, and biotite. Apatite, zircon, and opaque minerals occur as accessory minerals. Epidote, allanite, clinozoisite(?), chlorite, sphene, and a carbonate are present in varying amounts. Well-developed flakes of muscovite are present, and sericite is found throughout the rock. The modal composition of the Clinton facies is potassium feldspar, 29%; plagioclase, 36%; quartz, 27%; and mafics, 8% (Gore, 1976).

Megacrysts of microcline-microperthite are unique to the Clinton facies and appear to have originated from magma that crystallized after consolidation (Gore, 1976). The crystals are euhedral and up to 10 cm (4 in.) long. They exhibit Carlsbad twinning and may contain unaltered, euhedral, poikilitic plagioclase crystals 0.2-0.5 mm (0.008-0.02 in.) long. Blue quartz is found in the granite near Millstone Hill (Emerson, 1917).

The Devens-Long Pond facies is a belt of intermixed equigranular and porphyroblastic feldspathic gneisses, composed of biotite granite-quartz monzonite and granodiorite. Intermediate textures are subordinate.

The porphyroblastic gneiss is light to medium gray, with lenticular prophyroblasts, 1-2 cm (0.4-0.8 in.) long, that have xenoblastic microtextures and are composed primarily of microcline, but some plagioclase porphyroblasts are present. The major minerals are albite, microcline perthite, quartz, and biotite. Accessory minerals are chlorite, sericite, sphene, and an opaque

mineral. Zircon, apatite, muscovite, a carbonate, and limonite(?) are common. Allanite and tourmaline are rare. The modal composition is quartz, 26%; plagioclase, 40%; potassium feldspar, 29%; and mafics, 4% (Gore, 1976).

The nonporphyroblastic gneiss ranges from light gray to almost black, but is usually a medium-gray, medium-grained, well-foliated, crystalloblastic, slightly cataclastic biotite gneiss. A lenticular or massive amphibole-bearing biotite gneiss is also present. The major minerals are quartz, albite, and biotite. Microcline, sphene, and allanite are accessory minerals. Blue-green pleochroic amphibole (actinolite?), muscovite (commonly sericite), epidote, zircon, apatite, garnet, tourmaline(?), and chlorite are present. The modal composition is quartz, 29%; plagioclase, 53%; potassium feldspar, 5%; muscovite, 0.7%; and mafics, 12% (Gore, 1976). The rock ranges in composition from a quartz monzonite to a quartz diorite. Sriramadas (1966) reported that coarse-grained, foliated, gray granodiorite is the dominant facies represented in the Manchester quadrangle in New Hampshire.

Alteration. In the Clinton facies, the plagioclase is saussuritized, and the nonporphyritic border phase has undergone granulation and recrystallization in a major shear zone (Gore, 1976). Locally, in the fault zones, biotite replaces muscovite. Strong chloritization occurs where there is shearing and mylonization in the fault zones (Barosh, 1979).

7.2.2.5 Secondary Intrusions

The Devens-Long Pond facies is intruded into by the Chelmsford granite. The contact is transitional over a zone (possibly an aureole) varying from 30 m (100 ft) to more than 100 m (330 ft) wide (Gore, 1976). Dikes of pegmatite and aplite are common and may be related to the Chelmsford.

7.2.2.6 Country Rock

Geologic relations in the Harvard-Ayer and Clinton areas are shown in Figs. 7.4 and 7.5, respectively. The Clinton facies intrudes into the Oakdale formation, the Straw Hollow diorite, the Worcester phyllite, and the Berwick formation (Gore, 1976; Zen, 1981). Emerson (1917) stated that the Ayer complex occurs in dikes and veins in the Carboniferous strata of the Worcester and Merrimack troughs. The Clinton facies is in fault contact, across a narrow mylonite zone, with the Harvard conglomerate, and is also in fault contact with the Brimfield schist. Gore (1976) believed that the Oakdale formation may be unconformably deposited on parts of the Devens-Long Pond rocks.

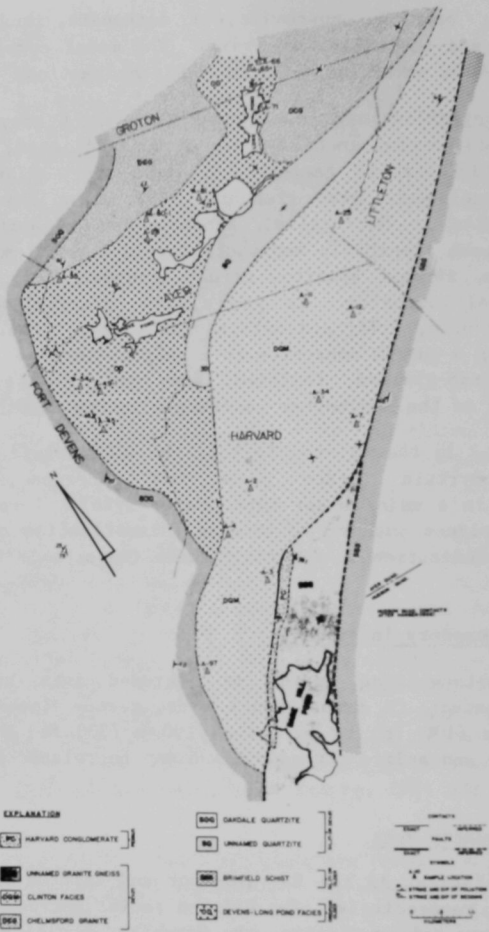


Fig. 7.4 Geologic Map of the Harvard-Ayer Area (Source: Gore, 1976)

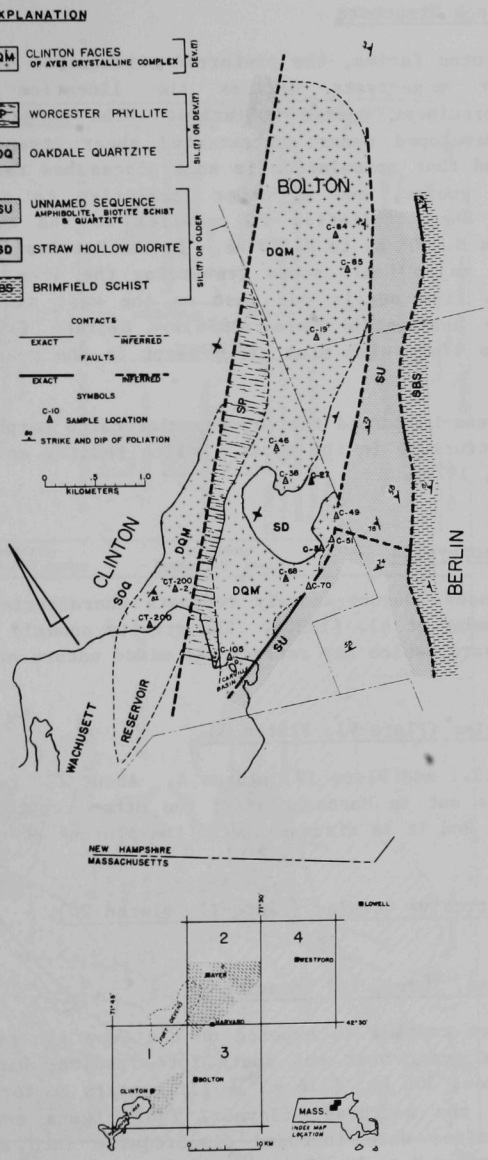


Fig. 7.5 Geologic Map of the Clinton Area (Source: Gore, 1976)

7.2.2.7 Rock Structure

In the Clinton facies, the preferred orientation of the xenoliths and potassium-feldspar megacrysts defines the lineation and a secondary development of prominent coarse foliation. The second foliation in the Clinton facies developed under concentrated shear stress in narrow zones. Gore (1976) stated that granulation in some places has reduced the rock to a thinly laminated gneiss, and at other localities to an augen gneiss or blastomylonite. Shear foliation is parallel to the major foliation and regionally strikes N. 30° E. to N. 60° E. In the Rattlesnake Hill fault zone, one of the three major fault zones traversing the area, the strike of the foliation changes from nearly due east on the west side of the fault to northeast on the east side (Gore, 1976). Emerson (1917) reported that vertical fractures lined with mica are present in the granite near Millstone Hill.

In the Devens-Long Pond facies, the microcline porphyroblasts produced an ophthalmitic structure in the rock. Relict bedding and cross-laminations are found (Barosh, 1977).

7.2.2.8 Geophysics

Broadly spaced Bouguer-anomaly contours characterize the area on the gravity map of Diment et al. (1980). The gravity anomaly increases from -10 mgal to 0 mgal inward, which may reflect the mixed nature of the rocks.

7.2.3 Barnard Gneiss (Plate VI, Pluton 3)

See Sec. 5.2.1 and Plate IV, pluton 4. About 130 km^2 (50 mi^2) of this body of rock crops out in Massachusetts; the other two-thirds of the pluton occurs in Vermont, and it is discussed with the plutons of that state.

7.2.4 Cape Ann Intrusive Complex (Plate VI, Pluton 20)

7.2.4.1 Size, Shape, and Location

The Cape Ann complex is exposed on the Cape Ann peninsula (see Fig. 7.6), and outcrops occur west and south into Ipswich, Danvers, and Beverly over an area of about 300 km^2 (116 mi^2). It appears to form a sheetlike body dipping gently to the northeast (Joyner, 1963; Lyons and Krueger, 1976). These alkalic granites were intruded discordantly into an area of folded igneous and metamorphosed rocks (Toulmin, 1964).

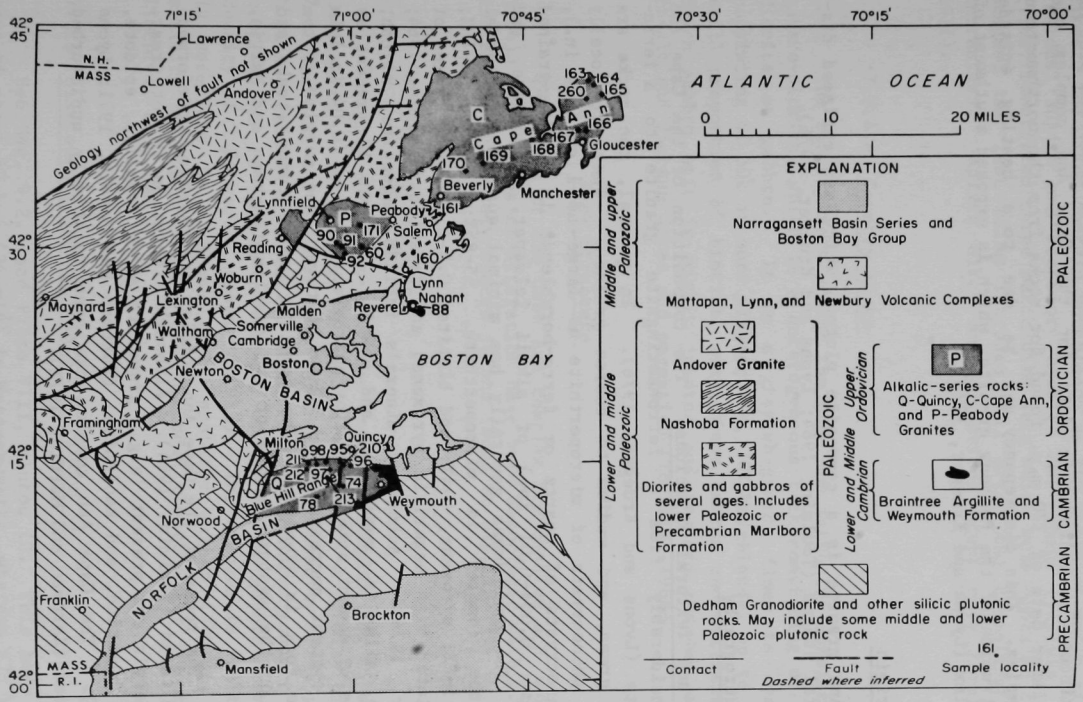


Fig. 7.6 Alkalic Granitic Plutons in Boston Bay Area
(Source: Zartman and Marvin, 1971)

7.2.4.2 Relative and Absolute Age

The Cape Ann complex is believed to be Late Ordovician in age. Zartman and Marvin (1971) dated the complex at 452 ± 10 m.y. with a concordia plot of Pb^{207}/Pb^{206} in zircons. Radiometric data for the Cape Ann complex are complicated by a later date of 370 m.y. found for rocks from the northwestern section of the complex. This discrepancy may be due to a heating episode related to the intrusion of the Peabody granite, which is exposed southwest of the Cape Ann complex (Lyons and Krueger, 1976).

7.2.4.3 Origin

The Cape Ann granite is a composite pluton, which was emplaced discordantly in several stages (Toulmin, 1964; Lyons and Krueger, 1976).

7.2.4.4 Petrography

Mineralogy and Textural Relations. The composition of the Cape Ann complex varies considerably from a ferrohastingsite granite to a ferrohastingsite syenite (Lyons and Krueger, 1976). In general, the rocks are massive, gray to green, and medium to coarse grained. They are locally porphyritic, with phenocrysts of microperthite as large as 1 cm (0.4 in.), and sometimes phenocrysts of quartz or ferro-hornblende in a fine-grained mosaic (0.1-0.3 mm; 0.004-0.01 in.) of alkali feldspar, oligoclase, and possibly quartz. The groundmass generally has minimal quartz. The modal composition of the complex is microperthite, 55-85%; quartz, 8-40%; ferrohastingsite, 5-8%; pyroxene, 0-4%; and biotite, 0.8-1.5%; with traces of riebeckite (associated with aegeritic pyroxene) and fayalite (Toulmin, 1964; Lyons and Krueger, 1976). Accessory minerals include magnetite, zircon, fluorite, sulfides, allanite, muscovite, and apatite. Quartz is present as mosaic intergrowths of equant anhedral grains approximately 1 mm (0.04 in.) in size. The quartz content decreases from 35% to 5% toward the margins of the complex, where the rock is syenitic. The syenitic facies forms stocks and dikes, especially on the western margin of the complex (Zartman and Marvin, 1971).

Alteration. The rocks of the Cape Ann complex are massive, undeformed, and relatively unaltered (Toulmin, 1964). Although alteration is not great, some secondary minerals are present (e.g., grunerite after fayalite, epidote, chlorite, calcite, hematite, and kaolinite) (Zartman and Marvin, 1971; Lyons and Krueger, 1976). The Cape Ann complex has been relatively undisturbed since it crystallized (Zartman and Marvin, 1971).

7.2.4.5 Secondary Intrusions

The Cape Ann complex is cut by a large number of highly altered diabase dikes. The dikes contain hornblende as the dominant mafic silicate, with some residual augite and accessory apatite and sphene (Toulmin, 1964). Roulston (1983) identified three mafic dike sets which were emplaced in three different stages, the last being intruded well after the consolidation of the other intrusives. Aplite dikes and pegmatites occur only rarely (Buma et al., 1971).

7.2.4.6 Country Rocks

The country rock around the Cape Ann intrusion consists mainly of Salem gabbro-diorite, which is a heterogeneous rock consisting of hornblende-augite diorite or gabbro. It is a grayish-green, fine- to coarse-grained rock consisting of 50% or more plagioclase, 20% hornblende, up to 20% augite, and a small proportion of biotite, with minor quartz and potassium feldspar. The Salem gabbro unit is highly jointed and faulted and has a large number of basalt and diabase dikes and pegmatite (Toulmin, 1964).

7.2.4.7 Rock Structure

The rocks of the Cape Ann complex are massive and undeformed (Toulmin, 1964). Some zones of alteration have been associated with ground-water action in regions of closely spaced joints (Buma et al., 1971).

7.2.4.8 Geophysics

Gravity data show that the Cape Ann complex has an average density of 2.66 gm/cm^3, with a range of 2.62-2.74 gm/cm^3. A gravity high has been noted over the east end of Cape Ann, which is not caused by the Cape Ann rocks. The high may reflect a concealed high-density intrusive younger than the Cape Ann complex, or it may be a remnant of the Salem gabbro-diorite, which has a density of 2.96 gm/cm^3 (Joyner, 1963). The complex is relatively shallow and is underlain by a dense rock, probably gabbro (Joyner, 1963). Seismic activity of moderate magnitude has been reported for the area (Sykes, 1978; Diment et al., 1980).

7.2.5 Glastonbury Gneiss (Plate VI, Pluton 5)

See Sec. 9.2.3 and Plate VIII, pluton 10. This body crops out over an area of about 62 km^2 (24 mi^2) in Massachusetts, whereas 222 km^2 (86 mi^2) occur in Connecticut. This body is therefore discussed with the plutons of Connecticut.

7.2.6 Monson Gneiss (Plate VI, Pluton 8)

7.2.6.1 Size, Shape, and Location

The Monson gneiss consists of three bodies located in north- and west-central Massachusetts -- the main body of Monson gneiss, the Warwick dome, which is also referred to as the Pauchaug gneiss, and the Tully body (see Fig. 7.3) (Robinson, 1963 and 1967; Robinson et al., 1982a). The main body extends from the southern portion of the Mount Grace 15-min quadrangle through Massachusetts into Connecticut. It has an elongate shape that narrows southward. The Warwick dome is located in the eastern half of the Northfield, the south-central portion of the Keene, the western portion of the Mount Grace, the northeastern corner of the Millers Falls, and the northwestern corner of the Orange 15-min quadrangles. It is an elongated oval about 18 km (11 mi) long and 5 km (3 mi) wide. The Tully body is located in the southeastern portion of the Mount Grace and southwestern portion of the Royalston 15-min quadrangles. It has a bulbous outline, with a maximum width of 5 km (3 mi) and a length of about 14 km (9 mi) (Thompson et al., 1968). The Monson gneiss as a whole is found over an area of about 528 about km^2 (204 mi^2) in Massachusetts.

7.2.6.2 Relative and Absolute Age

Brookins and Hurley (1965) reported a possible 472 ± 15 m.y. age of formation for the Monson gneiss, on the basis of Rb-Sr analysis of whole rock from the vicinity of the Killingworth dome south of Middletown, Conn. That gneiss is thought to be continuous with the Monson gneiss in Massachusetts (Rodgers et al., 1959). However, the samples used were relatively rich in strontium and poor in rubidium. Pegmatites that cut the Monson gneiss yield U-Pb, Th-Pb, K-Ar, and Rb-Sr ages that average about 250 m.y. (Rodgers, 1952; Wasserburg and Hayden, 1955; and Brookins et al., 1967). This Permian age is interpreted as coeval with the last strong thermal metamorphism affecting the Monson gneiss in west-central Massachusetts (Brookins, 1969). However, most core rocks of the Bronson Hill anticlinorium are of Ordovician age.

7.2.6.3 Origin

There is much controversy over the origin of the Monson gneiss. Some workers have proposed that the Monson gneiss grades upward into the overlying Ammonoosuc volcanics of Middle Ordovician age (Lundgren, 1962; Naylor, 1967), but Robinson (1963, 1967, and 1977) suggested an unconformity between these two contacts. He recognized three principal lithologies that are about equal in abundance. In types 1 and 2, the variety and detail of bedding and mineralogy indicate derivation from interbedded felsic and mafic volcanics, largely tuffs, which had undergone a minimum of chemical change by sedimentary

processes. Type 3 could represent either intrusive sills or unstratified portions. No regular pattern of distribution of the types has been found upon which to base an internal stratigraphy.

7.2.6.4 Petrography

Mineralogy and Textural Relations. The petrological characteristics of the principal components of the Monson gneiss as described by Balk (1946), Robinson (1963, 1967, and 1977), and others are as follows. The Monson gneiss is composed of about 15% amphibolite and 85% feldspar-quartz gneiss. The formation consists roughly of three principal types: (1) portions in which amphibolite and feldspar-quartz gneiss are closely interbedded; (2) portions made up entirely of feldspar-quartz gneiss, with well-defined layering due to minor compositional variations; and (3) portions composed of nonlayered feldspar-quartz gneiss.

Main Body of Monson Gneiss. Layered gneisses of types 1 and 2 are abundant along the western side of the main body. A large zone along the eastern border east and southeast of Kelton Hill consists of poorly layered, weakly foliated feldspar-quartz gneiss of type 3 (commonly with microcline megacrysts up to 2.5 cm [1 in.] across) and some feldspar-quartz gneiss with hornblende. These outcrops are massive and rounded, and amphibolite layers are rare.

Warwick Dome. The core of the Warwick dome is a medium-grained, nearly massive granitic gneiss. The modal composition is quartz, 30.5-40.8%; plagioclase, 24.1-50.8%; microcline, 12.9-39.6%; and biotite, 1.9-3.9%; with traces of magnetite, apatite, chlorite, and epidote (Robinson, 1967). The core rock differs from the border phase of the dome in its predominance of microcline over plagioclase, its lack of hornblende and chlorite, and its uniformity in mineral composition and texture. In many places, foliation is lacking, but lineation may be seen on surfaces striking north to south.

The border gneiss is a yellowish, medium- to fine-grained, moderately well foliated rock composed of sodium plagioclase (albite or oligoclase), quartz, biotite, and hornblende, with considerable amounts of secondary muscovite and chlorite; epidote, apatite, and sporadic grains of euhedral or anhedral garnet also are present.

In the northwestern corner of the dome is a large amphibolite lens. It is a well-foliated, grayish-green amphibolite, which contrasts with more siliceous, chlorite-bearing lenses.

Tully Body. The Tully body generally has the same layered characteristics as much of the main body. It is formed of massive, moderately foliated, feldspar-quartz gneiss. On the west face of the mountain is a zone of well-bedded amphibolite and feldspar-quartz gneiss containing a large ultramafic zone.

Alteration. In thin section, much of the gneiss of the Warwick dome shows alteration. The plagioclase is cloudy and commonly sericitized, and primary hornblende may be partly or wholly replaced by chlorite. The light-colored border gneiss contains many chlorite veinlets and small masses of coarse-leaved chlorite associated with dismantled masses of pyrite and magnetite.

7.2.6.5 Secondary Intrusions

No significant secondary intrusions are reported in the literature for the area.

7.2.6.6 Country Rocks

To the east, the Warwick dome is in contact with the Ammonoosuc volcanics of Ordovician age, which are fine-grained quartz-feldspar gneisses containing small amounts of layered quartz-feldspar-diopside granulites, hornblende gneiss and schist, and quartz-mica-sillimanite schist. On the west, it is bordered by the Crag Mountain formation of middle Paleozoic age. This formation is a dark-gray, well-foliated mica schist, commonly containing garnet, pyrite, and staurolite.

The Tully body is in contact with the Ammonoosuc volcanics on the upper western side. Elsewhere it is in contact with the Partridge formation, which is a mica schist containing black carbonaceous matter (commonly graphite) and iron sulfide. Less abundant rock types include calc-silicate and quartz granulite.

On the upper western and entire eastern side of the main body, the Monson gneiss is in contact with the Partridge formation. This formation is a mica schist of Middle Ordovician age containing graphite, magnetitic pyrrhotite, and minor inclusions of calc-silicate and quartz granulites (Emerson, 1917; Robinson, 1967).

7.2.6.7 Rock Structure

Two faults striking northeast cut into the Warwick dome in the upper western area. One fault striking from north to south cuts the Tully body in the eastern portion. One fault striking east-southeast cuts into the main body of Monson gneiss on the western side (Thompson et al., 1968).

Linear structural features are represented by oriented minerals, mineral aggregates, stretched pebbles, and minor fold axes, which are parallel to major folds. In the west, linear features gently plunge to the north, whereas in the east, they plunge south and southeast at highly varied angles.

The north end of the main body plunges 20°-60° S. The west limb is nearly vertical, and the east limb dips 30°-60° W. The north end of the Warwick dome plunges east-southeast. The dome is divided into two lobes by a transverse syncline overturned to the south (Thompson et al., 1968). The south lobe is overturned in a southwestern direction. The narrow belts of Monson gneiss, peripheral to the east side of the main body and wrapping around the south end of the Tully body, are interpreted as the core of a recumbent fold. The Tully body has both of its ends plunging southwest beneath the main body. The recumbent northward overturning of the north ends of the main and Tully bodies took place after the formation of the Skitchewaug nappe and before the regional tight folding synchronous with dome formation. The narrow band of the Partridge formation in the middle of the main body in the Quabbin reservoir is a normal syncline (Robinson, 1967).

7.2.6.8 Geophysics

Airborne electromagnetic surveys have been carried out to differentiate mafic bodies and magnetite-rich silicic felsic facies of the various intrusive and metamorphic phases. Particular information is available on the Shelbourne Falls area rocks (Griscom and Bromery, 1968).

Gravity surveys have been conducted to define the dome margins (Ashwal et al., 1979). On a Bouguer map, the contours of gravity anomaly define the outline of the gneiss bodies (Diment et al., 1980).

Limited data on heat flow are available for intrusives in west-central Massachusetts.

7.2.7 Newburyport Granodiorite (Plate VI, Pluton 13)

7.2.7.1 Size, Shape, and Location

The Newburyport granodiorite (also referred to as quartz diorite by Castle, 1960) is a dual pluton in Essex County, northeastern Massachusetts, that brackets the mouth of the Merrimack River and extends into New Hampshire. The Newburyport granodiorite lies in the Merrimack synclinorium. It occupies a roughly square area in Massachusetts about 10 km (6 mi) on a side; outcrops occur over an area of about 98 km^2 (38 mi^2) in Massachusetts.

7.2.7.2 Relative and Absolute Age

Lithologically similar rocks intrude into the Silurian Kittery quartzite in New Hampshire. The K-Ar biotite age of 321 ± 10 m.y. reported by Zartman et al. (1970) for its extension in New Hampshire may be due to thermal

modification durng the Acadian orogeny. Zen (1981) identified the Newburyport granodiorite as being Silurian or Ordovician.

7.2.7.3 Origin

The Newburyport granodiorite is of magmatic origin and is assumed to be related to the widespread mafic igneous activity associated with the different phases of the Taconic orogeny.

7.2.7.4 Petrography

Mineralogic and Textural Relations. The Newburyport body is a gray, medium-grained granodiorite-tonalite. The major minerals are andesine-labradorite, 45-70%; orthoclase, <5%; quartz, 5%; and hornblende, 20-25%. Biotite, augite, ilmenite, magnetite, apatite, rutile, and titanite are accessory minerals (Emerson, 1917). Toulmin (1964) and Chute (1966) have reported small bodies of lithologically similar rocks in the Salem and Norwood quadrangles.

Alteration. The alteration of hornblende to biotite and chlorite, and of feldspar to sericite and epidote is prominent (Toulmin, 1964; Chute, 1966) and may have resulted from low-grade metamorphism in the area.

7.2.7.5 Secondary Intrusions

Dikes of granite and aplite are found in the Newburyport pluton.

7.2.7.6 Country Rocks

The Newburyport granodiorite extends eastward to the Atlantic Ocean. The Sharpners Pond diorite lies in fault contact to the south. To the west and north of Newburyport, a gray, medium-grained porphyritic granite with microcline phenocrysts, designated as part of the Newburyport complex, intrudes into the Kittery formation (Zen, 1981).

7.2.7.7 Rock Structure

The Essex fault is the only major, regionally developed fault bordering the Newburyport granodiorite. It has an east-northeast trend.

7.2.7.8 Geophysics

Birch (1983) interprets high magnetic anomalies in the area as small Newburyport intrusions extending into the continental shelf.

7.2.8 Sharpners Pond Diorite (Plate VI, Pluton 14)

7.2.8.1 Size, Shape, and Location

The Sharpners Pond tonalite (referred to as diorite by Zen, 1981), formerly mapped with the Salem gabbro-diorite, is an irregular subalkalic intrusive series in northeastern Massachusetts. Its outcrops are distributed through the Lawrence, South Groveland, Wilmington, Reading, and Salem quadrangles, over an area of about 190 km^2 (73 mi^2). In the vicinity of Sharpners Pond (see Fig. 7.7), located approximately 7 km (4.5 mi) west of Topsfield, only one of the three facies in the formation is well developed (Castle, 1965). Other undivided alkalic intrusives in the vicinity may be correlative.

7.2.8.2 Relative and Absolute Age

Zartman et al. (1970) assigned a K-Ar isotopic age of Devonian or older to the Sharpners Pond tonalite. The Assabet quartz diorite described by Hansen (1956) correlates with the Sharpners Pond body in its lithology and its relationship with the Andover granite. Stratigraphic correlation with other plutons in the area indicates a Silurian age (Zen, 1981).

7.2.8.3 Origin

According to Castle (1965), the Sharpners Pond magma may have formed from fractional melting of mantle material. The intrusive igneous rocks in the area may be interrelated, although Toulmin (1964) rejected the correlation between the Sharpners Pond tonalite and the Salem gabbro-diorite. Castle (1965) believed that the possible inclusion of the Salem rocks in the subalkalic intrusive series should not be dismissed. However, the Sharpners Pond tonalite is an I-type granite that intruded into the metavolcanics of the Nashoba block (Hepburn et al., 1983).

7.2.8.4 Petrography

Mineralogic and Textural Relations. According to Castle (1965), the Sharpners Pond rocks contain three facies: hornblende diorite; biotite-hornblende tonalite; and biotite tonalite. All three facies are transitional with each other, and the younger rocks tend to be more felsic. The Sharpners Pond body becomes more mafic towards the Salem gabbro-diorite.

The Sharpners Pond tonalite is dark greenish-gray or black to light gray, medium grained and equigranular, and primarily massive but somewhat foliated. The chief minerals are microcline, 0-33%; biotite, 5-17%; hornblende, 0-4%; plagioclase, 28-56%; and quartz, 26-31%. In the biotite tonalite facies, microcline is abundant. Chlorite, muscovite, epidote,

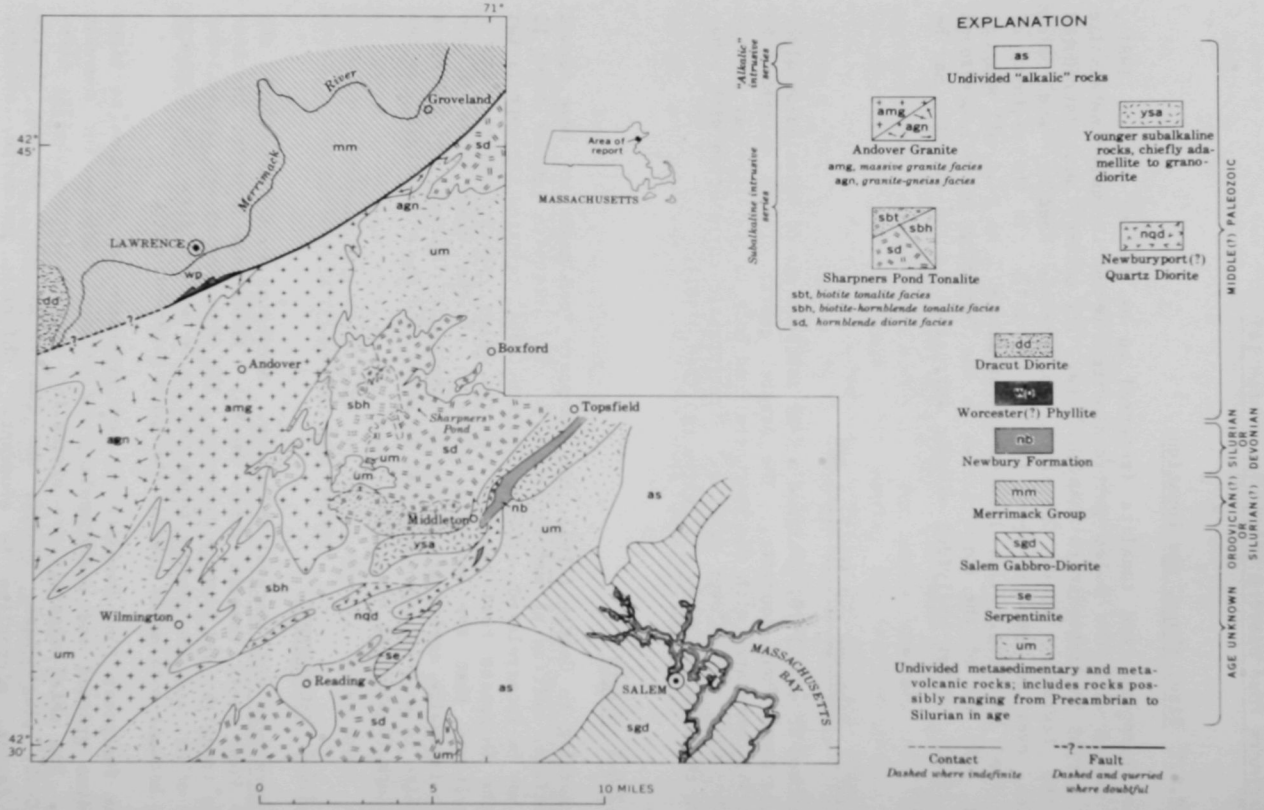


Fig. 7.7 Generalized Geologic Map Showing Distribution of Major Rock Units in Part of Northeastern Massachusetts (Source: Castle, 1965)

sphene, and apatite are accessory minerals. With an increase in hornblende to 19-37% and a decrease in microcline, the tonalite grades into hornblende diorite, whereas, with an increase in biotite to 10-23%, it becomes biotite-hornblende tonalite.

Alteration. Toulmin (1964) stated that the Sharpners Pond rocks are less altered than the Salem gabbro-diorite.

7.2.8.5 Secondary Intrusions

Andover granite and other Silurian subalkalic rocks, chiefly adamellite, granodiorite, and granite, intrude into the Sharpners Pond body. A few trap dikes are found in outcrops of Sharpners Pond (Castle, 1965).

7.2.8.6 Country Rock

The Sharpners Pond rocks are included in the Nashoba fault zone. They intrude into Ordovician units of sillimanite schist and gneiss, amphibolite, biotite gneiss, calc-silicate gneiss, and marble located near Andover (Zen, 1981). The Salem gabbro-diorite lies to the southeast. In the northernmost section of the hornblende diorite facies is a contact with the Bloody Bluff fault; on the other side lies the Merrimack Group.

7.2.8.7 Rock Structure

The Sharpners Pond rocks are less deformed than those of the Salem gabbro-diorite. Zen (1981) showed regional faulting in the area. Secondary faults are transverse to the major faults. The faults disrupted all of the stratified and plutonic rocks in post-Pennsylvanian and pre-Cretaceous times (Bell and Alvord, 1976).

7.2.8.8 Geophysics

Section 7.2.4.8 would apply to the Sharpners Pond diorite as well.

7.3 ACADIAN AND POST-ACADIAN

7.3.1 Belchertown Intrusive Complex (Plate VI, Pluton 6)

7.3.1.1 Size, Shape, and Location

The Belchertown pluton is located in west-central Massachusetts and is bordered on the west by the Connecticut Valley border fault (see Fig. 7.3).

It is roughly circular in map plan, with a diameter ranging from 12 km (7 mi) to 18 km (11 mi); outcrops occur over an area of approximately 123 km^2 (48 mi^2). On the basis of ground and airborne magnetometer studies, Hall (1975) considered the pluton to be disc-shaped with a thickness of 500-1000 m (1650-3300 ft).

7.3.1.2 Relative and Absolute Age

The Belchertown pluton has been recognized as a Devonian intrusion by its relationships to other rocks and recent age determinations on zircon. A fresh sample of anhydrous, two-pyroxene quartz monzodiorite typical of the inner core of the pluton was chosen as representative of the magmatic stage of emplacement and least affected by metamorphic hydration. The U-Pb data for this sample indicate an age of 380 ± 5 m.y. (Ashwal et al., 1979). This age is in excellent agreement with the geological relations of the Belchertown pluton, which deforms the Erving and Waits River formations of Early Devonian age.

7.3.1.3 Origin

The Belchertown pluton is one of a series of intrusions emplaced into domes of Precambrian gneiss along the Bronson Hill anticlinorium during the peak of regional kyanite-zone metamorphism during the Acadian orogeny (Thompson et al., 1968). It was perhaps emplaced as a syntectonic intrusion well before the end of the Acadian orogeny. Robinson et al. (1982b) reported that a hornblende-biotite-gneiss mantle around the Belchertown body was produced during the dome stage of formation while it preserved an unaltered core.

7.3.1.4 Petrography

Mineralogy and Textural Relations. Ashwal et al. (1979), Hall (1973), and Leo et al. (1977) considered the zoned nature of the Belchertown pluton to represent the effects of regional metamorphism. The inner zone of the pluton is composed predominantly of massive, medium-grained hypersthene-augite quartz monzodiorite, which is believed to be the area least affected by metamorphism. On a fresh surface, the rock appears pink due to abundant plagioclase with minute inclusions of iron-titanium oxides. This rock contains as much as 35% mafic minerals, including poikilitic brown biotite flakes as much as 2.5 cm (1 in.) across and stubby crystals of green augite and brown hypersthene. Other minerals include hornblende, quartz, potassium feldspar, and accessory magnetite, titanohematite, apatite, and zircon. In some places, this rock shows weak subparallel alignment of plagioclase grains. The modal composition of this rock is quartz, 4.4%; plagioclase, 46.9%; potassium feldspar, 11.9%; hypersthene, 5.2%; augite, 15.6%; biotite 12.8%; and other, 5%.

The transitional zone of the pluton contains rocks that are defined as having a massive texture with no tectonic fabric and no hypersthene; they may or may not contain augite. The average modal composition is quartz, 10.1%; plagioclase, 48.1%; potassium feldspar, 8.8%; augite, 2.2%; amphibole 17.3%; and biotite, 13.3%. The boundary between the outer zone and the transitional zone is defined as the outer limit of transitional rocks containing augite. The boundary between the inner zone and transitional zone is the outer limit of inner-zone rocks containing hypersthene.

The outer zone of the pluton is composed of the fully recrystallized counterpart of the primary quartz monzodiorite. This recrystallized rock is a hornblende-biotite gneiss, which has a well-developed tectonic foliation and lineation formed by alignment of biotite, hornblende, quartz, and feldspar. This rock type is medium grained, gray to greenish-gray in color, and contains no pyroxenes or opaque minerals but does contain important amounts of bluegreen hornblende, olive-green biotite, epidote, and sphene. It has a greater proportion of quartz and a smaller proportion of plagioclase than the primary quartz monzodiorite. The average modal composition is quartz, 9%; plagioclase, 41.5%; potassium feldspar, 10.3%; amphibole, 21.3%; and biotite, 13.3%.

Alteration. In the rocks of the transitional zone, olive-green biotites commonly contain concentrations of euhedral or subhedral sphene crystals along their edges, suggesting breakdown of brown, titanium-rich igneous biotite to form titanium-deficient green biotite and sphene. Green biotite also forms rims between orthopyroxene and orthoclase microperthite. The rocks of the outer zone have no pyroxenes or opaque minerals. These minerals have been completely consumed during a chemical redistribution involving hydration and recrystallization, to become a hornblende-biotite gneiss (Ashwal et al., 1979).

7.3.1.5 Secondary Intrusions

At the northwestern contact between the inner and transition zones of the pluton is an elliptical mass of ultramafic and mafic rock measuring roughly 200 m (650 ft) by 300 m (980 ft). The rock in this ultramafic body is a dark-green, medium- to coarse-grained hornblende-biotite-olivine pyroxenite, with a texture dominated by euhedral to subhedral augite and smaller grains of bronzite and subrounded magnesian olivine. The origin, and therefore the age, of this olivine-bearing ultramafic body is uncertain (Robinson, 1967).

7.3.1.6 Country Rocks

The country rocks into which the Belchertown pluton intrudes consist of a series of gneiss domes and ridges that define the Bronson Hill anticlinorium (Billings, 1956; Robinson, 1967). The anticlinorium extends from northern New Hampshire to Long Island Sound. The cores of the domes are

composed of late Precambrian and Ordovician gneisses overlain by a sequence of metamorphosed Middle Ordovician volcanic and sedimentary rocks, which are in turn unconformably overlain by a sequence of metamorphosed Silurian and Lower Devonian sedimentary and volcanic rocks (Billings, 1956; Thompson et al., 1968). Lower Jurassic and Upper Triassic clastic sedimentary rocks and basalts are the country rocks west of the Connecticut Valley border fault.

On the north margin, the Belchertown pluton discordantly intrudes into the Lower Devonian Erving formation (Guthrie and Robinson, 1967; Guthrie, 1972; and Robinson et al., 1973) and the Fourmile gneiss (Ashenden, 1973) of possible late Precambrian age in the Pelham dome. To the east, the pluton is concordantly intruded into the east limb of a syncline of the Middle Ordovician Partridge formation. To the south, it is nearly concordant at a stratigraphic level near the base of the Erving formation. To the west, the Connecticut Valley border fault separates the pluton from the Lower Jurassic East Berlin, Granby Tuff, and Portland formations of the Connecticut valley.

7.3.1.7 Rock Structures

The only major fault in the area is the Connecticut Valley border fault, which is adjacent to the pluton on the west. According to Ashwal (1974), mylonitization and cataclasis is present in zones. Most zones of mylonitized gneisses have undergone recrystallization, as evidenced by mineral lineation and coaxial minor folds. In some places, the rock of the inner zone shows weak subparallel alignment of plagioclase grains, which is indicative of igneous flow structure.

The pluton intrudes into the Bronson Hill anticlinorium, and there has been no folding of either the intrusion or the country rocks in post-Acadian time (Thompson et al., 1968; Ashwal et al., 1979).

7.3.1.8 Geophysics

According to Ashwal and Hargraves (1977), the plutonic rocks possess a natural remnant magnetization having a southerly direction of polarization. Computer modeling of the three-dimensional shape of the pluton based on magnetic studies reveals that the inner zone is a slab about 500 m (1630 ft) thick having a root about 5000 m (16,400 ft) deep beneath a 2772-gamma low lying southeast of the surface exposure of the plutonic rocks. This shape may be interpreted as either a deep downfold of primary plutonic rocks in the synclinal corner between the Glastonbury dome and the Monson gneiss or as a cylindrical feeder vein for the entire pluton.

An idea of the three-dimensional shape of the pluton may be obtained from the calculated residual gravity anomaly (Kane and Bromery, 1968). By using an average density of 2.84 g/cm^3 for the Belchertown pluton and 2.67 g/cm^3 for the dome gneisses and the rocks of the Triassic and Jurassic basin,

the best fit of the calculated gravity anomaly is produced by a thin, 1500-m (4900-ft) thick, disc-shaped intrusive mass (Hall, 1975). This model is not applicable, however, to the eastern and southeastern parts of the pluton, because there is no appreciable density difference between the rocks of the pluton and the adjacent metamorphosed Paleozoic stratified rocks.

7.3.2 Fitchburg Intrusive Complex (Plate VI, Pluton 10)

7.3.2.1 Size, Shape, and Location

The Fitchburg intrusive complex is a body, somewhat elongated from north to south, in central Massachusetts, portions of which lie in the Ashburnham, Gardner, Wachusett Mountain, Paxton, Ashby, Fitchburg, Sterling, and Worcester North quadrangles. It includes many igneous sheets and irregular masses of diverse composition that crop out over an area of about 444 km^2 (171 mi^2).

7.3.2.2 Relative and Absolute Age

The Fitchburg intrusive complex intruded into sedimentary and metamorphic rocks of Silurian and Lower Devonian age (Zen 1981), thus placing the age of the crystalline rocks as Devonian or younger. Maczuga (1981) quoted R.E. Zartman as communicating a zircon age of 382-412 m.y. for the weakly foliated biotite-muscovite granite.

7.3.2.3 Origin

The high potassium and aluminum content of the Fitchburg complex, and the absence of large amounts of relatively mafic material, led Maczuga (1981) to propose that granitic rocks of the complex were derived from partial melting of a subducted continental crust. The granodiorite probably represents more extensive partial melting than the granite. The tonalitic inclusions may be early differentiates from such a melt. Chemical variation within the units may be a reflection of a melt-segregation process, but more likely it is the interaction of complex mechanisms. The peaking of the Acadian metamorphism obviously was responsible for local anatectic melting of the granodiorite, which formed migmatites.

7.3.2.4 Petrography

Mineralogy and Textural Relations. Maczuga (1981) provided petrographic details for the threefold lithologic classification proposed for the Fitchburg complex by Hepburn (1976), Peper (1976), and Tucker (1978).

Foliated biotite granodiorite-tonalite gneiss is a prominent sill-like intrusion in the South Monoosnoc Hill and Wachusett Mountain, representing the eastern and western limbs, respectively, of the broad structural syncline containing the Fitchburg rocks. At South Monoosnoc, the complex consists of a dark-gray, fine-grained, granoblastic and moderately foliated rock containing plagioclase, 32-40%; potassium feldspar, 0.1-9%; quartz, 22-25%; and biotite, 29-32%. Accessory minerals are sphene, muscovite, allanite, epidote, zircon, apatite, magnetite, and ilmenite. At Wachusett Mountain, the rock is mineralogically similar but strongly foliated. Sillimanite-grade metamorphism has caused extensive recrystallization, thus obliterating primary textures.

Biotite-muscovite-granite-granodiorite gneiss forms a moderately to strongly foliated rock unit that varies in texture and mafic mineral content. Principal minerals are plagioclase, 23-48%; potassium feldspar, 6-32%; quartz, 24-28%; biotite, 6-20%; and muscovite, 3-9%; with accessory minerals similar to the granodiorite-tonalite gneiss.

The biotite-muscovite granite is massive to weakly foliated and medium to coarse grained, with uniform texture and mineralogy over large areas. It probably intruded, concurrent with or shortly after the major deformational episode, into the other two units described above, which are older. The major minerals are plagioclase, 21-29%; potassium feldspar, 29-32%; quartz, 27-32%; muscovite, 3-7%; and biotite, 2-8%. Zircon, ilmenite, apatite, and allanite are the chief accessory minerals.

7.3.2.5 Secondary Intrusions

Pegmatitic dikes are abundant in the structurally lower gneissic units and may be anatectic melts (Maczuga, 1981). Tourmaline pegmatites are present in the granites. Mesozoic diabase dikes are also present.

7.3.2.6 Country Rocks

The complex is located in the Merrimack synclinorium, a belt of eugeosynclinal sedimentary and volcaniclastic rocks. It intrudes into the Littleton and Paxton formations. To the northeast of the complex lies the Massabesic gneiss of the Avalon platform rocks.

7.3.2.7 Rock Structure

The granodiorite-tonalite gneiss and granite-granodiorite gneiss form structurally lower and higher units, respectively, of the broad syncline in which the complex is contained. These units were probably emplaced as sheets before or during the three major episodes of westward-directed recumbent folding (Field, 1975; Tucker, 1977 and 1978; and Peterson, 1983) during the Acadian. Mineral lineations are parallel to axial surfaces in the metamorphic rocks (Tucker, 1978). Robinson and Hall (1980) proposed nappe formation,

backfolding, and doming as a sequence of evolution of the Acadian crystalline rocks of the area. The massive granite represents emplacement during or after the end stages of the Acadian deformation.

7.3.2.8 Geophysics

Diment et al. (1980) indicated that a gravity low characterizes the area on the Bouguer anomaly map.

7.3.3 Hardwick-Coys Hill Granite (Plate VI, Pluton 9)

7.3.3.1 Size, Shape, and Location

The Hardwick-Coys Hill body is a group of rather symmetrical, elongated intrusions located in central Massachusetts in the Royalston and Athol 15-min quadrangles (see Fig. 7.3). The largest of these begins with a blunt point in Marlboro, N.H. (where it is known as the Fitzwilliam granite), and extends southward, ending with a similar blunt point in Monson, Mass. Some short, narrower outcrops parallel the largest body; the group extends about 70 km (43 mi) from north to south in Massachusetts, and its outcrops occur over an area of about 358 km^2 (138 mi^2).

7.3.3.2 Absolute and Relative Age

Billings (1956) assigned the Hardwick and Coys Hill granites to the New Hampshire plutonic series, thus indicating a Devonian or younger age. Robinson (1967) indicated a similar age based on field relations. The pluton traces northward into the Spaulding series, 393 m.y. old, of New Hampshire (Lyons et al., 1982). Hayward (1983) reported a Rb-Sr whole-rock age of 383 m.y. for the Fitzwilliam granite in New Hampshire.

7.3.3.3 Origin

The Hardwick granite is of magmatic origin and contains local inclusions of sillimanite-mica schist and calc-silicate country rock. "The abundance of iron in the border rocks is attributed to differentiation. If it had been derived from absorption of the rusty schist, graphite and fibrolite would be present" (Emerson, 1917).

7.3.3.4 Petrography

Mineralogy and Textural Relations. The composition of the Hardwick pluton varies from granite to tonalite, and is commonly porphyritic. It is

low in quartz and high in feldspars, usually containing orthoclase, a little plagioclase, much biotite, much magnetite, hypersthene, zircon, and allanite (Emerson, 1917). It is generally very coarse and blotched with biotite. On cleavage faces, broad, equant, wavy-surfaced flakes of biotite, many of them more than a centimeter (less than an inch) square, or aggregations of small flakes, cover the surface. The rock has a dead black color. Except where it is in contact with the Coys Hill granite, hornblende is generally absent, the dark color being due to abundant biotite and magnetitie. The Coys Hill granite is a coarsegrained, foliated granite gneiss with mica quartz monzonite phases (Emerson, 1917). It is regularly sheeted and contains very large microcline megacrysts. These crystals generally have a granulated border. Feldspar is chiefly represented by microcline and anorthoclase.

Where the Coys Hill granite is in contact with the Hardwick granite, it is a clear-gray granite; where it rests against the Brimfield schist, small red garnets or large secondary garnets and biotite become abundant. Graphite scales appear in the feldspar, and amber fibrolite appears in silky white masses.

Alteration. No specific alteration features are reported. However, secondary minerals have developed at the contact of the Hardwick and Coys Hill bodies.

7.3.3.5 Secondary Intrusions

No significant secondary intrusions are described for the area, but fine-grained granite dikes cut across the tectonic foliation.

7.3.3.6 Country Rocks

The west side of the Hardwick granite is in intrusive contact with the Partridge formation of Middle Ordovician age (Robinson, 1967; Thompson et al., 1968). This formation consists predominantly of mica schist containing black carbonaceous matter, commonly graphite, and an iron sulfide that is magnetic pyrrhotite in most occurrences. Less abundant rock types include calc-silicate and quartz granulites. On the west side, the granite also is in contact with the Littleton formation of early Devonian age. This formation is typically a gray mica schist or phyllite with a variable proportion of sandy or quartzitic beds. Minor rock types include rare calc-silicate rocks and some concordant biotite gneiss. The east side of the granite is in contact with the intimately interfolded Partridge and Littleton formations.

7.3.3.7 Rock Structure

Hall and Robinson (1982) proposed the development of the Bronson Hill anticlinorium and Merrimack synclinorium in central Massachusetts and adjacent

New Hampshire through several stages. The early stage was marked by nappes which progressed through backfolding and culminated in the final stage of doming low-density gneisses into denser mantle rocks. The Coys Hill pluton shows a tectonic foliation that defines west-plunging folds. Ductile mylonite zones are also present. In the Hardwick formation, the foliation dips gently to the west and dome-stage lineation is also present (Robinson et al., 1982b).

7.3.3.8 Geophysics

Section 7.1.6.8 would apply to the Hardwick-Coys Hill granite as well.

7.3.4 Williamsburg Granodiorite (Plate VI, Pluton 4)

7.3.4.1 Size, Shape, and Location

The Williamsburg granodiorite consists of a series of irregular masses outlining, and scattered through, a roughly oval area about 30 km (19 mi) long and 12 km (7 mi) wide on the west side of the Connecticut valley in western Massachusetts (see Fig. 7.8). Outcrops occur through an area of 130 km^2 (50 mi^2).

7.3.4.2 Relative and Absolute Age

Field relationships suggest an age of Early Devonian for this intrusive (Norton, 1967).

7.3.4.3 Origin

The Williamsburg granodiorite intrudes into formations of Early Devonian to Ordovician age. Intrusion may have been contemporaneous with the regional deformation and metamorphism associated with the Devonian Acadian orogeny. On the east, the intrusive is limited by structural contact with younger Triassic and Jurassic formations (Norton, 1967; Hatch et al., 1968).

7.3.4.4 Petrography

Mineralogy and Textural Relations. The Williamsburg granodiorite is texturally variable. In the central areas, it is fine to medium grained, with the compositional characteristics of biotite granite. It locally has large masses and sheets of coarse-grained and very coarse grained varieties with abundant muscovite, which grade into the main mass of granite without showing distinct boundaries. Sodium plagioclase, microcline, quartz, and muscovite-biotite are the principal minerals. Plagioclase is more abundant than potassium feldspar (Emerson, 1917; Bain, 1955).

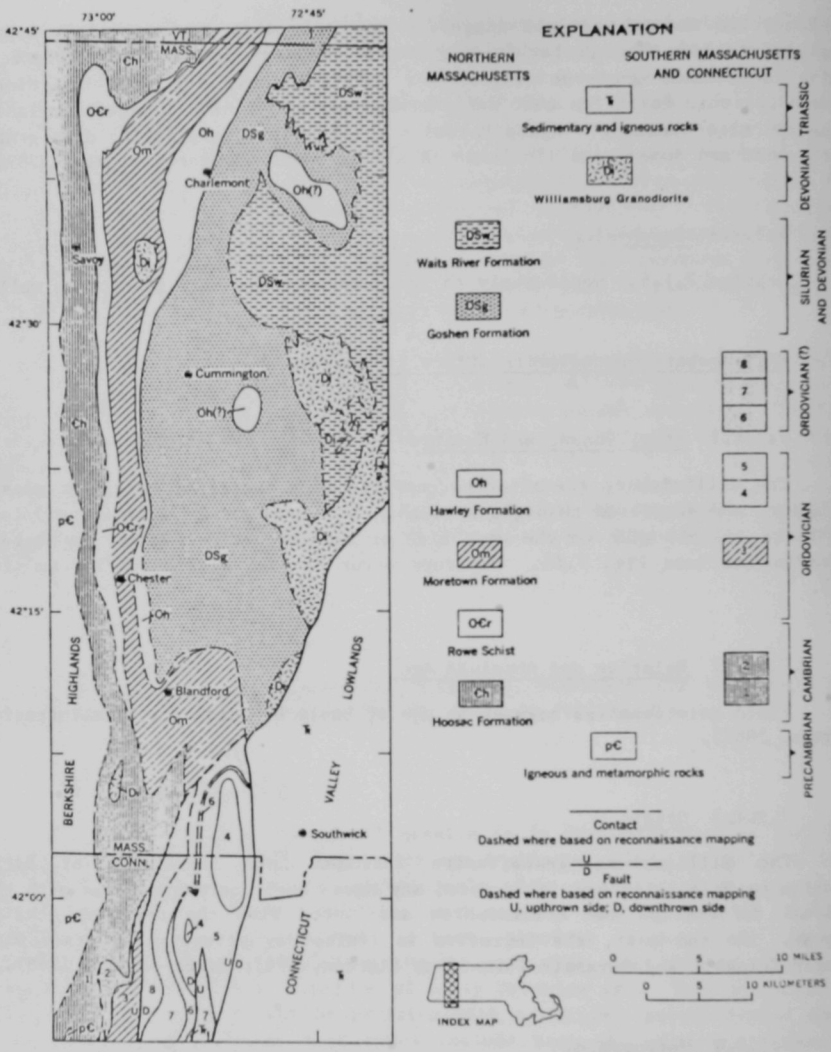


Fig. 7.8 Geologic Map of Williamsburg Granodiorite Area (Source: Hatch et al., 1968)

Alteration. The feldspars of the intrusive show sericitization. Pegmatitic phases show moderate hydrothermal alteration.

7.3.4.5 Secondary Intrusions

Pegmatites and aplites profusely penetrate into and beyond the main granite mass. They form a halo around the intrusive. Pegmatites also intrude into the schist. Here they are rich in rare-earth element minerals: it is common to find tourmaline, spodumene, zircon, spessartine, and many varieties of gem-quality feldspars.

7.3.4.6 Country Rocks

To the west and south, the country rock is a series of Middle Ordovician to Lower Devonian formations consisting of granulites, schists, calc-silicate rocks, and volcanics. To the east, the Williamsburg granodiorite is in structural contact with the Portland formation of Triassic-Jurassic age, which consists of conglomerates, arkoses, and volcanics. To the northwest, the granodiorite has a less distinct contact with the Belchertown tonalite (Hatch et al., 1966).

7.3.4.7 Rock Structure

The Williamsburg granodiorite exhibits foliation and lineation akin to that of the country rock. In addition, the Great Connecticut Valley fault is the major structural feature to the east.

7.3.4.6 Geophysics

The area is characerized by widely spaced 0- to -10-mgal gravity-anomaly contours on a simple Bouguer map (Diment et al., 1980).

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8 CRYSTALLINE ROCKS OF RHODE ISLAND

8.1 GRENVILLE AND AVALONIAN

8.1.1 Bulgarmarsh Granite (Plate VII, Pluton 6)

8.1.1.1 Size, Shape, and Location

The Bulgarmarsh granite is exposed in the Fall River, Tiverton, and Sakonnet Point 7.5-min quadrangles of Rhode Island, extending east and northeast into Massachusetts. Its outcrops in Rhode Island are found over an area of roughly 100 km^2 (39 mi^2), in a belt about 24 km (15 mi) long and no more than 6 km (4 mi) wide.

8.1.1.2 Relative and Absolute Age

On the basis of stratigraphic relations, Pollock (1964) proposed the Bulgarmarsh granite to be of Devonian(?) age or older. Galloway (1973) determined a Rb-Sr whole-rock age for the Bulgarmarsh granite of 534 ± 13 m.y. Mineralogically, texturally, and structurally, the Bulgarmarsh granite is similar to the Dedham granite in Massachusetts (see pluton 19 on Plate VI), for which Zartman and Naylor (in press) report a late Precambrian U-Pb zircon age.

8.1.1.3 Origin

The intrusive contact of Bulgarmarsh granite with mica-chlorite schist indicates a magmatic origin for the body. Xenoliths of the schist have been turned and variably reoriented with respect to the original bedding. Roof pendants and many inclusions of schist indicate that stoping played a strong role in the emplacement of the pluton (Pollock, 1964).

8.1.1.4 Petrography

Mineralogy and Textural Relations. The Bulgarmarsh granite is best described as a light-colored, coarse-grained, foliated rock that is very rich in quartz and feldspar. It is often pink to gray and usually contains quartz, microperthite-microcline, albite, biotite, chlorite, and muscovite. Minor constituents and accessory minerals include magnetite, ilmenite, leucoxene, rutile, apatite, zircon, epidote, garnet, sphene, and allanite. The albite tends to contain small epidote crystals. Pollock (1964) suggested that this may mean that the albite was derived from a somewhat calcic plagioclase. The

average modal composition of the Bulgarmarsh granite is quartz, 36-66%; microperthite-microcline, 17-38%; albite (An-3-7), 7-32%; biotite and chlorite, 0.2-5%; and muscovite, 0.2-13%.

Bulgarmarsh granite, in places, contacts mica-chlorite schist to the west in a transition zone that is as much as 183 m (600 ft) wide. The granite gneiss of the zone is rich in quartz and feldspar like the main body of granite, but is slightly porphyroblastic. It has an increased amount of quartz and correspondingly less feldspar, that is, quartz, 63%; microcline-microperthite, 14%; albite, 10%; muscovite, 8%; and biotite and chlorite, 5%. Apatite, epidote, sphene, garnet, zircon, and magnetite are accessory minerals. The porphyroblasts consist of pink microperthite-microcline and average about 2.5 cm (1 in.) long (Pollock, 1964; Quinn, 1971). This zone may have formed when the nearby granite soaked into and replaced some of the schist (Pollock, 1964).

Alteration. Biotite is altered to chlorite. The opaque minerals also show varying degrees of alteration.

8.1.1.5 Secondary Intrusions

There are many small aplite dikes and quartz veins in the Bulgarmarsh granite (Pollock, 1964).

8.1.1.6 Country Rocks

The Bulgarmarsh granite is in contact with the Precambrian mica-chlorite schist of the Sakonnet region to the west of its southern end, and the Pennsylvanian sedimentary rocks of the Narragansett basin to the west. It is also in contact with the chlorite-biotite schist of the Tiverton region and a small body of metadiorite to the west of its center and its northern end.

The mica-chlorite schist of the Sakonnet region is light gray to greenish, fine grained, and thinly bedded. Major constituent minerals are biotite, chlorite, and quartz. Interbedded with the schist are numerous thin marble and quartzite beds. The chlorite-biotite schist of the Tiverton region is green to gray, fine grained, and poorly foliated. Major minerals include chlorite, biotite, and quartz. Beds of quartzite and amphibolite are interbedded with the schist. The Dighton conglomerate and the Rhode Island formation of the Narragansett basin lie unconformably over the western edge of the Bulgarmarsh granite. The Rhode Island formation is about 3050 m (10,000 ft) thick and consists of a large variety of sedimentary rock types. The Dighton conglomerate is the youngest member in the Narragansett basin.

8.1.1.7 Rock Structure

The Bulgarmarsh granite shows flow foliation over most of its extent, which may have been modified by metamorphism. The foliation results from the parallelism of biotite and chlorite flakes. The foliation, though quite variable, generally trends to the northeast, oblique to the margin of the Narragansett basin. No folds have been reported in the Bulgarmarsh granite.

8.1.1.8 Geophysics

Hildreth's (1979) regional Bouguer anomaly map for New England shows low gravity for the Bulgarmarsh area.

8.1.2 Esmond Granite (Plate VII, Pluton 5)

8.1.2.1 Size, Shape, and Location

The Esmond granite comprises two major and seven smaller bodies of granite in northern Rhode Island. The largest body is almost 19 km (12 mi) long and 5.5 km (3.4 mi) wide, covering approximately 67 km^2 (26 mi^2) in the Georgiaville, North Scituate, Providence, and Pawtucket 7.5-min quadrangles. Outcrops occur in a total area of approximately 123 km^2 (47 mi^2), and include exposures in the Crompton, East Greenwich, Franklin, Blackstone, and Chepachet 7.5-min quadrangles, in addition to those mentioned above.

8.1.2.2 Relative and Absolute Age

Due to poor outcrop exposure and continuity complex structures, and variable lithologic features, age relationships of bedrock units in Rhode Island are not clearly known. Age estimates have been based on crosscutting rela- tionships of igneous bodies and the degree of foliation in gneissic rocks. The degree of foliation is not a reliable indicator, as foliation is grada- tional and variable through single bodies. Since the major period of geological investigations from 1940 to 1971 (Quinn et al., 1949; Quinn, 1971), the ages of plutonism have been reevaluated. The suggested Mississippian(?) or older age of the Esmond granite (Quinn, 1971) has been revised to late Precambrian (Hermes et al., 1981). Hermes et al. (1981) gave a U-Pb zircon age of 620 m.y. (Avalonian) for the Esmond granite (also see Hermes and Zartman, 1982).

8.1.2.3 Origin

The mode of emplacement of the Esmond granite was by magmatic intrusion. Evidence for igneous intrusion includes many xenolithic inclusions and

large roof pendants of rocks of the Blackstone series (Hermes et al., 1981). The angular inclusions of schist and quartzite were entrained and then settled in a liquid magma, as evidenced by the random orientation and arrangement of inclusions of variable lithology (Quinn, 1963). A comagmatic event produced the Grant Mills granodiorite, quartz diorite, and Esmond granite. The dominant lithology in the Esmond group is a calc-alkaline rock series having rare earth element patterns similar to rocks from other calc-alkaline granitic suites known to have formed at a convergent plate boundary (Hermes et al., 1981).

8.1.2.4 Petrography

Mineralogy and Textural Relations. The Esmond group is an assemblage of related plutonic rocks ranging from quartz diorite to two-feldspar granite and possibly more mafic hypabyssal and volcanic rocks spatially related to the Hunting Hill greenstone member of the Blackstone series (Hermes et al., 1981). The rocks have calc-alkaline affinities and are massive to highly fractured (Hermes and Zartman, 1982), The Esmond granite is the dominant lithology of the Esmond group. The Grant Mills granodiorite, a porphyritic rock, grades into the Esmond granite.

The Esmond granite is a light-pink, medium-grained, massive to faintly foliated granite. The mineralogical composition (as volume %) can be described by an average of five typical specimen modes presented by Quinn (1971): potassium feldspar, 40%; quartz, 34%; plagioclase, 22%; biotite, 2%; muscovite and other accessory minerals, 2%, which may include magnetite, sphene, apatite, monazite, zircon, garnet, and calcite. Granitic varieties of rocks in the group are light gray, light tan, or green (in the presence of abundant fine-grained epidote), medium to coarse grained, or distinctly foliated. The associated gray to green-gray, porphyritic granodiorite is composed of a medium- to coarse-grained matrix of albite-oligoclase, microcline microperthite, quartz, and biotite, with microcline microperthite phenocrysts.

Alteration. Typical chemical alterations of the Esmond granite area rocks include plagioclase to albite, epidote, and sericite, or to less calcic plagioclase with abundant small inclusions of muscovite and clinozoisite (Quinn, 1967), and the alteration of biotite to chlorite. The plagioclase alteration may be a retrograde metamorphic effect or deuteric alteration. The change was due to permeating solutions and low pressure during waning metamorphism of the Blackstone series, or was a late result of plutonic activity (Quinn, 1959), or occurred during the cooling of the granite (Quinn, 1963).

In the Esmond group, foliated rocks (e.g., secondary chlorite, epidote, muscovite, and clinozoisite) are concentrated in linear patterns along annealed brittle fractures (Hermes et al., 1981). Calcic plagioclase has altered to albite, sericite, and epidote, and much biotite has altered to chlorite (Quinn, 1971).

8.1.2.5 Secondary Intrusions

The Quincy granite intrudes into the Esmond granite in the Pawtucket quadrangle, and diabase dikes of possible Triassic age cut the Grant Mills granodiorite in the same quadrangle. In the Blackstone quadrangle, a diabase dike cuts the Esmond granite and extends into Massachusetts. This relatively large mafic dike intersects sedimentary rocks of Pennsylvanian age, suggesting that it may correlate with Triassic diabase units in the Connecticut valley (Quinn, 1971). According to Quinn (1952), the Esmond granite in the East Greenwich quadrangle has been intruded irregularly by fine-grained granite. Inclusions of possible Scituate granite gneiss may be reinterpreted as secondary intrusions into the Esmond granite.

8.1.2.6 Country Rocks

The intrusion of the Esmond granite into the Blackstone series caused variable metamorphism and granitization of the Mussey Brook schist, the Quinnville quartzite, the Sneech Pond schist, and the Hunting Hill greenstone. Inclusions of blocks of quartzite and schist in the granite indicate that magmatic stoping during emplacement of the grani e bro e o pieces of Blackstone-series country rocks. The scarcity of bedroc o cro and the thick blanket of glacial deposits prevent observations of intrusive relationships and contacts of the Esmond rocks with country rocks, although Quinn (1971) mentioned some minor contact metamorphism resulting from intrusion of plutonic rocks in Rhode Island.

8.1.2.7 Rock Structure

The Esmond rocks are massive to highly fractured. A normal fault, extending 8 km (5 mi) in a northeastern direction in north-central Rhode Island and into Massachusetts, forms the contact of the Esmond granite with the Bellingham conglomerate. The eastern block, containing the Esmond granite, is upthrown relative to the western block (Quinn, 1971). A narrow, elongate body of Esmond granite to the south is bounded on all sides by faults.

No joint sets or fracture patterns have been described in the Esmond granite, although locally sheared and annealed brittle fracture zones, along which secondary minerals (chlorite, epidote, and muscovite) are concentrated, are common (Hermes et al., 1981).

Flow of the magma is indicated by the random orientation of block inclusions and variably developed gneissic foliation. The Esmond granite is less foliated than other granite gneisses of comparable age in the region, suggesting that it intruded as a late tectonic event. Foliation strikes north and dips east in the Chepachet quadrangle; is vertical and strikes north in the Providence quadrangle; strikes north and dips east in the southern part of

the Pawtucket quadrangle; strikes nearly east-west and dips north in the northern part of that quadrangle; strikes mostly northwest and dips northeast in the North Scituate quadrangle; and strikes and dips in various directions in the Grant Mills granodiorite in the Providence quadrangle. Foliation may be due to flow or compression during or immediately after intrusion (Quinn, 1959). Flow structure is more distinct near the margins of the rocks. The Esmond granitic rocks have not been significantly folded.

8.1.3 Sterling Pluton Group (Includes Hope Valley Alaskite and Ponaganset Gneisses) (Plate VII, Pluton 3)

8.1.3.1 Size, Shape, and Location

The Sterling pluton group includes the Ponaganset gneiss, the Hope Valley alaskite gneiss, and related crystalline rocks that also continue into Massachusetts. The Ponaganset gneiss includes a range of gneissic and granitic rock varieties, which crop out discontinuously in a north-south trend over a large area in the Chepachet, Clayville, Uxbridge, Oxford, Thompson, East Killingly, Oneco, and Voluntown quadrangles in northwestern Rhode Island and eastern Connecticut and Massachusetts. The Hope Valley alaskite gneiss is exposed in southwestern and northwestern Rhode Island, with continuations into Connecticut and Massachusetts. The largest unbroken expanse is in the Hope Valley, Carolina, Slocum, Kingston, Voluntown, and Ashaway quadrangles. Smaller masses occur in the Narragansett Pier, Wickford, Coventry Center, East Killingly, Thompson, and Oxford 7.5-min quadrangles. In Rhode Island, there are at least 14 mappable masses of Hope Valley alaskite gneiss that total approximately 260 km^2 (100 mi^2). The Potter Hill granite gneiss, which is mapped as four small exposures in southwesternmost Rhode Island, may be a lithologic variant of the Hope Valley alaskite gneiss (Hermes, 1982). The Potter Hill granite gneiss crops out in the Ashaway quadrangle. All together, these masses of presumably related rock occur over an area of about 535 km^2 (207 mi^2).

The region has been extensively glaciated, and glacial lakes, some of which have been modified into reservoirs, overlie the Ponaganset gneiss. Glacial deposits almost completely obscure large areas of bedrock. There is relatively great topographic relief in the outcrop area, ranging from large, smooth hills to swamplands.

8.1.3.2 Relative and Absolute Age

The intrusive, metamorphic, stratigraphic, and temporal relationships of many of the bedrock units in Rhode Island are complex and not yet definitely known. The difficulty in interpreting such relationships is due, in part, to poor exposure, gradational and variable lithologic features, and

complex structures. The Sterling pluton group intrudes into the older Blackstone series and is intruded into by dikes and sills of the younger Scituate granite gneiss, the Westerly granite and the Narragansett Pier granite (Moore, 1959). According to Hermes et al. (1981), the rocks of the Sterling pluton group are probably of Precambrian (Avalonian) age, on the basis of U-Pb age determinations on zircons.

Hermes (1982) stated that preliminary and unpublished Rb-Sr whole-rock dates are compatible with the Avalonian age interpretation. The Ponaganset gneiss continues into Massachusetts, where it may correlate with the Northbridge granite gneiss, now known to be of late Precambrian age. Uranium-lead age determinations on zircons from the Hope Valley alaskite give a late Precambrian age (Hermes et al., 1981).

8.1.3.3 Origin

The mode of emplacement of the Sterling plutonic group is probably by syntectonic magmatic intrusion, with associated emanations of magma into the Precambrian Blackstone series and metasomatic granitization (Quinn, 1967; Moore, 1958). The rock material invaded the Blackstone-series terrane as a liquid or partly liquid intrusion, pushing blocks of schist and quartzite ahead of it and rotating them within the liquid (Quinn, 1971). The granitic magma may have been derived through anatexis of nearby feldspathic sedimentary rocks (Quinn, 1967 and 1971).

8.1.3.4 Petrography

Mineralogy and Textural Relations. The rocks of the Sterling pluton group are lithologically very heterogeneous and anisotropic. The Ponaganset gneiss is made up of many types of granitic rocks, generally gray to light gray, medium to coarse grained, and gneissic or foliated and lineated. Variations of the rock type include lenses or layers that are pinkish or dark gray in color and fine grained or porphyroblastic. Quinn (1967) reported that the Ponaganset gneiss unit includes granite, granodiorite, tonalite, adamellite, trondhjemite, and quartz monzonite that are all distinctly gneissic. Banding in the rocks is defined by lineation of biotite flakes in foliation planes, which generally strike and plunge to the north. A second foliation occurs in some places, striking to the west and dipping to the north. A light-colored facies of the gneiss resembles the Hope Valley alaskite gneiss, but is pink and contains more microcline and quartz and less plagioclase and biotite.

The Ponaganset gneiss, in most occurrences, has a granitic composition. The potassium feldspar is finely twinned microcline, which may be perthitic (Quinn, 1967). Plagioclase varies from oligoclase to andesine. Accessory minerals include chlorite, epidote, clinozoisite, magnetite, ilmenite, pyrite, garnet, sphene, leucoxene, zircon, allanite, fluorite, and

apatite. An average of eight modal analyses (in volume %) of the Ponaganset gneiss yielded quartz, 35.4%; potassium feldspar, 23.0%; plagioclase, 29.2%; biotite, 8.8%; hornblende, 1.2%; epidote, 0.6%; muscovite, 0.1%; and accessory minerals, 1.4% (Quinn, 1971).

White to light-pink subhedral porphyroblasts of microcline in many places deform biotite foliation and enclose large magnetite octahedra. Where well developed, this unusual texture is made up of microcline porphyroblasts up to 2 cm (0.8 in.) in length, which constitute up to 20% of the rock (Harwood and Goldsmith, 1971). The Ponaganset gneiss also contains scattered interlayers of dark-gray, nonporphyroblastic gneiss up to a meter or two (a few feet) thick.

The Ponaganset gneiss contains many lenses of fine-grained felsic and mafic rocks, which are different from the gneiss but which are lineated and foliated continuously with the host rock. The contacts are sharp. The lenses are interpreted to be relics of older gneissic rocks. An increase in mylonitization in the Ponaganset gneiss near its contact with the Hope Valley unit has been interpreted by O'Hara (1983) to indicate that they may have been tectonically juxtaposed as allochthonous units.

The Hope Valley alaskite is usually a medium-grained, light-gray, lineated gneiss. The lineation is formed by flattened, rod-shaped aggregates of quartz grains 15-20 mm (0.6-0.8 in.) long. Euhedral to subhedral grains of magnetite up to 3 mm (0.1 in.) in diameter are common. Biotite may be present to the south, where the rock appears moderately foliated. Modal analysis yielded microcline and microperthite, 24-39%; albite and oligoclase, 20-40%; quartz, 29-42%; and biotite, trace-4%; with accessory amounts of magnetite, sphene, apatite, zircon, muscovite, hornblende, epidote, allanite, rutile, fluorite, sericite, hematite, leucoxene, and calcite (Moore, 1958).

Alteration. Weathered Ponaganset gneiss is characteristically gnarled and nubby, a texture developed by the differential weathering of porphyroblasts of microcline. Chlorite, sericite, hematite, leucoxene, and calcite are alteration minerals in the Hope Valley alaskite (Moore, 1958). The Sterling plutonic group has been modified by metamorphism.

8.1.3.5 Secondary Intrusions

Dikes and sills of Scituate granite cut the Ponaganset gneiss. In the Oneco quadrangle, a sill emanating from the Scituate granite to the west has a surface exposure 2.6 km (1.6 mi) long and 0.16 km (0.1 mi) wide in the Ponaganset gneiss (Harwood and Goldsmith, 1971).

Sills and dikes of aplite and pegmatite may be present in the Hope Valley alaskite, which was intruded into by the Westerly and Narragansett Pier granites. The fine-grained aplite granite is present mainly as sills and dikes. It is light gray to pinkish in color; some parts are lineated and some

are massive. Locally, it may be porphyritic, with phenocrysts of pink microcline. The main constituents are microcline or microperthite, albite-oligoclase, quartz, and biotite, with minor amounts of zircon, apatite, magnetite, allanite, epidote, muscovite, sericite, chlorite, and hematite (Moore, 1958).

A small amount of the Westerly granite is also present as sills and dikes in the alaskite gneiss. It is fine grained, bluish gray, and uniform. Main constituents include smoky quartz, plagioclase, potassium feldspar, and biotite, with an average grain size of 0.6-0.8 mm (0.02-0.03 in.). Although usually massive, the granite may be weakly foliated along the margins of dikes (Moore, 1959).

The Narragansett Pier granite is typically tan, pink, or gray, medium to coarse grained, massive, and porphyritic. Its main constituents are biotite, quartz, potassium feldspar, and plagioclase. Magnetite is a common accessory mineral. Weak foliation is present locally (Moore, 1959).

8.1.3.6 Country Rocks

The Sterling pluton group discordantly intrudes into the Blackstone series, the Chepachet gneiss, the Absalona formation, an amphibolite unit, and the Plainfield formation of Connecticut and western Rhode Island. Contacts with the Blackstone series are characterized by a zone grading from gneiss through migmatitic rock to the Blackstone series. Small pieces or blocks of schist and quartzite originally of the Blackstone series have been metasomatized and granitized, and rotated in the gneiss such that foliation wraps around the included bodies. Most marginal areas of the gneisses near country rocks are stongly foliated. Generally, intrusion of the gneiss resulted in varying degrees of metamorphism, metasomatism, and granitization of the inclusions and marginal country rock.

8.1.3.7 Rock Structure

In the Oneco and Voluntown quadrangles in western Rhode Island and eastern Connecticut, in which the Ponaganset gneiss is exposed, Perhac (1958) noted a major set of joints that strikes northeast and dips steeply to the northwest and southeast, and two vertical sets of joints that strike nearly perpendicularly, that is, east-west and north-south.

According to Quinn (1967), the foliation and layering in the Ponaganset gneiss developed from modification, under stress, of the original flow structure. Foliation in the Ponaganset gneiss is steep and north trending. In some parts of the Chepachet quadrangle, a second foliation strikes to the west and dips north. In the Oneco quadrangle, foliation formed by the lineation of biotite grains increases in dip inclination from west to east. Foliation in the Ponaganset gneiss and in the fine-grained facies of the

Scituate granite dips to the west and north and flattens, at the north end of the Scituate coarse-grained lobe, towards the southeast. This structural feature is interpreted to be a structural nose of foliation plunging to the northwest in the central gneisses of the Oneco quadrangle in Connecticut, indicating significant folding (Harwood and Goldsmith, 1971).

Where foliation is present in the Hope Valley alaskite gneiss, it is a representation of the primary flow structure. Foliation is moderate to the south and weak to absent to the north. The trend of the foliation is northwest, with a moderate dip to the northeast in the Hope Valley quadrangle. Lineation, as outlined by rod-shaped quartz aggregates, plunges to the north at low angles; the lineation becomes intense and north trending near its contact with the Ponganset gneiss on the east (O'Hara, 1983).

Deformation involving folding, faulting, and metamorphism of the surrounding rocks predated the emplacement of the Ponaganset gneiss. Patterns of change in foliation structure suggest minor folds in the gneiss.

Shear zones are present locally in the Hope Valley alaskite gneiss. There is such a zone that trends northeast in the Carolina quadrangle. The gneiss in this quadrangle is cut by steep joints striking N. 10-30° E. and about N. 20° W. (Moore, 1959).

8.1.3.8 Geophysics

The area is characterized by a gravity low, as seen from the regional gravity data reported by Hildreth (1979).

8.1.4 Ten Rod Granite Gneiss (Plate VII, Pluton 2)

8.1.4.1 Size, Shape, and Location

The Ten Rod granite gneiss appears in southern Rhode Island as a large, crystalline rock body of somewhat irregular shape in the Slocum, Kingston, Narragansett Pier, and Wickford 7.5-min quadrangles, and as small inliers in the Hope Valley alaskite gneiss and Scituate granite in the Coventry Center, Hope Valley, and Carolina 7.5-min quadrangles. The large body of the Ten Rod granite gneiss measures about 17 km (10.6 mi) long and 12 km (7.6 mi) wide, occupying approximately 131 km^2 (50 mi^2). The city of Kingston and 14 small lakes lie within the outcrop area. Bodies of the Precambrian Hope Valley alaskite gneiss and Blackstone series metamorphic rocks crop out in the gneiss.

8.1.4.2 Relative and Absolute Age

Quinn (1971) classified the Ten Rod granite gneiss with the "older plutonic rocks" of Rhode Island, citing previous investigators (e.g., Moore, 1958), who had suggested that the Ten Rod granite gneiss was younger than the Scituate granite gneiss of Devonian(?) age or older. Scituate granite gneiss, Hope Valley alaskite gneiss, and Ten Rod granite gneiss were formerly included in the Sterling group of plutonic rocks, until it was recognized that the granite gneisses were of diverse ages and rock types. The relative age relationships interpreted from early field investigations and absolute ages from isotopic analyses have been reevaluated, as workers recognize the Ten Rod granite gneiss as part of the recently established Avalonian plutonism at convergent plate margins.

Zircon geochronologic analysis yielded a late Precambrian (Avalonian) date of emplacement and a 275-m.y. metamorphic event (Hermes et al., 1981). The same isotope-disturbing event probably explains the 285-m.y. age of the Ten Rod granite gneiss obtained by Quinn et al. (1957). Williams (1964) discussed a 303-m.y. age. Day (1968) determined the Ten Rod granite gneiss and Hope Valley alaskite gneiss to be 638 ± 42 m.y. old by Rb-Sr whole-rock dating.

8.1.4.3 Origin

The Ten Rod granite gneiss was emplaced by magmatic intrusion, but details are confusing. For example, Moore (1963 and 1964) described dikelike intrusions of the Ten Rod granite gneiss into the Middle Devonian Scituate granite gneiss. (This must be incorrect, because the Ten Rod granite gneiss has been dated at roughly 600 m.y., while the Scituate granite gneiss has been dated at 370 m.y.) In the Slocum quadrangle, on the other hand, the Scituate granite gneiss was reported to intrude into the Ten Rod granite gneiss (Power, 1959). Barosh and Hermes (1981) stated that locally the Sterling group rocks in southeastern Connecticut were emplaced as sills parallel to foliation and layering in the country rock and then folded to fit the configuration of fault zones.

8.1.4.4 Petrography

Mineralogy and Textural Relations. The Ten Rod granite gneiss is described as a pink to gray, fine- to medium-grained granite gneiss with strong lineation and moderate to weak foliation. An average of five typical modes presented by Quinn (1971) shows a mineralogic composition (in volume %) of quartz, 32.9%; plagioclase, 33.2%; potassium feldspar, 27.2%; biotite, 5.3%; hornblende, 0-2.8%; and accessory minerals, 1.0%. Large phenocrysts of pink to tan perthitic microcline, which may contain inclusions of plagioclase and quartz or be concentrated into aggregates with inclusions, give the rock a distinctive appearance. The microcline phenocryst grains and aggregates are

scattered or sometimes arranged in a well-developed lineation in a groundmass of fine-grained, albite-rimmed plagioclase, quartz, and microcline.

The petrology of the Ten Rod granite gneiss varies greatly from body to body, with significant variations even in a single body. In the Kingston quadrangle, the rock is a medium-gray and pink-gray, medium- to coarse-grained gneiss of smoky quartz, flesh-colored potassium feldspar (10-15% microcline grains and aggregates), white plagioclase (An-22), 5-10% biotite, and magnetite (Moore, 1964). Feldspar and quartz augen are typically developed, so much so that the Ten Rod granite gneiss is labeled augen gneiss in the Slocum (Power, 1959) and Narragansett Pier (Nichols, 1956) quadrangles. The augen gneiss is described as a strongly lineated and foliated gneiss, with augen of quartz and feldspar imbedded in a micaceous matrix and with inclusions and intercalations of feldspathic biotite schist. It was interpreted to be the result of feldspathization-granitization of masses of Blackstone series rocks, but evidence of intrusion, indicating a magmatic origin, refutes that hypothesis (Williams, 1964).

In the Wickford quadrangle, the Ten Rod granite gneiss is represented by three mapped facies -- porphyritic, nonporphyritic, and fine grained -- all of which are characterized by abundant biotite and augen-shaped microcline. In the Kingston quadrangle, the granite gneiss is coarser grained and less strongly or persistently foliated, being a gray, medium-grained gneiss of a plagioclase and quartz groundmass with microcline phenocrysts.

Alteration. Metamorphism of the Ten Rod granite gneiss produced granulation, metamorphic foliation, and late porphyroblasts (Quinn, 1971). Deformation produced strain shadows and granulated margins in quartz grains. Feldspar has been partly altered to sericite (Power, 1959).

8.1.4.5 Secondary Intrusions

Fine-grained granite sills and dikes occur in the Ten Rod granite gneiss in the Kingston quadrangle (Moore, 1964). The granite is a light-gray to pink, partly lineated and partly massive rock. Locally, it may be porphyritic with phenocrysts of pink microcline. Main constituents are microcline or microperthite, albite-oligoclase, quartz, and biotite, with minor amounts of zircon, apatite, magnetite, allanite, epidote, muscovite, sericite, chlorite, and hematite (Moore, 1958). Locally, dikes and veins of Narragansett Pier granite cut the Ten Rod granite.

8.1.4.6 Country Rocks

On the south, the Ten Rod granite gneiss was intruded into metasediments of the Blackstone series, which consists of interlayered gray to pinkish-gray, medium-grained, lenticular to weakly banded quartz-feldspar-biotite gneiss; light- to dark-gray, medium- to fine-grained quartz-biotite

schist; medium- to dark-gray, medium-grained hornblende feldspar schist; and local dark hornblende-biotite-quartz feldspar gneiss or fine-grained feldspathic quartzite (Moore, 1959). The Ten Rod granite gneiss also is in contact with the Permian Narragansett Pier granite, the Pennsylvanian Rhode Island formation, the mid-Devonian Scituate granite gneiss, and the Precambrian Hope Valley alaskite gneiss. The Scituate granite gneiss may be pink, tan, or gray, and medium to coarse grained. It is often lineated and locally foliated. The main constituents are microcline, microperthite, albite-oligoclase, quartz, and biotite (Quinn, 1971).

8.1.4.7 Rock Structure

Jointing of the Ten Rod granite gneiss is common in the Slocum quadrangle (Power, 1959). Local shear zones and cleavage in shear zones (striking northwest and dipping west) have been observed in the Coventry Center quadrangle (Moore, 1963).

Foliation is generally well developed in the Ten Rod granite gneiss and usually strikes northeast to northwest and dips north. Moderate to strong lineation formed by streaks of biotite and large microcline grains occurs in the Kingston quadrangle. The lineation plunges northeast at a moderate angle.

8.2 TACONIC

There are no rocks associated with the Taconic orogeny in Rhode Island.

8.3 ACADIAN AND POST-ACADIAN

8.3.1 Scituate Granite (Plate VII, Pluton 4)

8.3.1.1 Size, Shape, and Location

The Scituate granite is exposed over a large area of central Rhode Island. It crops out largely in the North Scituate, Clayville, Crompton, Coventry Center, Slocum, and Hope Valley 7.5-min quadrangles. There are small, scattered exposures in the Oneco, Voluntown, Carolina, Georgia- ville, and Chepachet quadrangles. The largest continuous area within which outcrops are found is an irregularly deformed ellipse approximately 40 km (25 mi) long and 24 km (15 mi) wide. Outcrops of the Scituate granite are found over an area of about 490 km^2 (189 mi^2).

8.3.1.2 Relative and Absolute Age

Zircon ages for Scituate granite are mid-Devonian (Hermes et al., 1981). However, rocks in Massachusetts, just north of northwestern Rhode Island, have been identified as similar to Scituate granite but have been dated as late Precambrian (Zartman and Naylor, in press). Therefore, rocks presently mapped as Scituate granite may consist of intermingled plutons of mixed ages (Hermes et al., 1981). In Rhode Island, Scituate granite is 370 m.y. old and is equivalent in age to the Cowesett granite. Contacts between the 370-m.y.-old Scituate granite and the 630-m.y.-old Hope Valley gneiss are not gradational in Rhode Island (Hermes, 1982).

8.3.1.3 Origin

Evidence for the origin of the Scituate granite cannot be observed in most places. However, in the North Scituate and Hope Valley quadrangles, there is evidence of magmatic intrusion into the Blackstone series and other rocks (Quinn, 1971). In the Chepachet quadrangle, the Scituate granite appears to have intruded along the foliation of the Absalona formation. Because the foliation of the Scituate granite plunges to the north, the Scituate magmas probably originated to the north and moved upward toward the south (Quinn, 1967).

8.3.1.4 Petrography

Mineralogy and Textural Relations. Jordan and Hermes (1983) consider the whole area of Scituate granite to be a Devonian intrusive complex that includes the petrogenetically related East Greenwich pluton newly described by them. The Scituate granite is medium to coarse grained and pink, tan, or gray in color. It is commonly lineated and locally foliated. Lineation is marked by oval splotches of biotite flakes that range from 20 mm (0.8 in.) by 50 mm (2 in.) to 40 mm (1.6 in.) by 75 mm (3 in.), and which are 2-10 mm (0.08-0.4 in.) apart (Moore, 1963). Much of the granite is subporphyritic to porphyritic, with phenocrysts consisting of microcline.

Modal analysis yielded microcline and microperthite, 31.6-49.3%; albite-oligoclase, 4.7-30.2%; quartz, 19.3-39%; and biotite, 1.7-11.5%. Hornblende and magnetite are common locally (Quinn, 1971). Accessory minerals include sphene, zircon, epidote, apatite, chlorite, muscovite, allanite, ilmenite, pyrite, and fluorite. Microcline is present as grains 12-20 mm (0.5-0.8 in.) in cross section by 40 mm (1.6 in.) in length; the long axis is usually oriented parallel to the biotite lineation. Much of the quartz is present as lenticular aggregates within the plane of foliation. Intrusions of a finer-grained granite of similar composition are found in the same area as the Scituate granite and are thought to represent a late-stage event (Quinn, 1971). According to Jordan and Hermes (1983), the East Greenwich pluton displays a calc-alkaline to alkaline trend. It is characterized by quartz

diorite-monzodiorite, fine-grained granite-granodiorite, and abundant equigranular granite phases.

Alteration. The Scituate granite underwent a period of metamorphism, alteration, and protoclasis, as seen in the presence of epidote chlorite and bent biotite grains, plagioclase twins, strain shadows and sutured borders of quartz grains, and granulated or broken microcline (Moore, 1963).

8.3.1.5 Secondary Intrusions

The Ten Rod granite gneiss has been reported to intrude into the Scituate granite in the southern part of the Scituate outcrop area. The Ten Rod granite gneiss is pinkish to medium gray, medium to fine grained, and porphyritic. It consists mainly of microcline, albite-oligoclase, quartz, and biotite, with well-developed foliation. The microcline is present in large, pink to tan crystals (Quinn, 1971). Because Hermes et al. (1981) assigned a Precambrian age to the Ten Rod granite gneiss, the relationship of the Scituate and the Ten Rod granite gneisses is uncertain.

8.3.1.6 Country Rocks

The Scituate granite intrudes into the Absalona formation at Snake Hill in the Chepachet quadrangle (Quinn, 1971). The Absalona formation is a porphyroblastic, dark-gray biotite schist or gneiss. The main mineral constitutents are microcline, microperthite, albite, quartz, and biotite. The porphyroblasts are composed of microcline (Quinn, 1971).

8.3.1.7 Rock Structure

The Alleghenian orogeny probably induced some degree of deformation and metamorphic features in the Scituate granite belt. Many outcrops of Scituate granite in the Coventry Center quadrangle are cut by shear zones that generally strike northward and dip to the west. Relative movement across the zones is uncertain, although in places the west side appears to have moved down and to the north (Moore, 1963).

Foliation in the Scituate granite appears in places to be a primary flow structure. Elsewhere it could be a relict or a later metamorphic structure (Quinn, 1963). Lineation in the Coventry Center quadrangle plunges 15-30° N. or NNW. Foliation tends to strike northward, dipping 20-40° W. or NW. (Moore, 1963).

8.4 ALLEGHENIAN AND POST-ALLEGHENIAN

8.4.1 Narragansett Pier Granite (Plate VII, Pluton 1)

8.4.1.1 Size, Shape, and Location

The Narragansett Pier granite forms a roughly rectangular intrusive body extending about 32 km (20 mi) from Narragansett Pier west to Westerly. In the Narragansett Pier quadrangle, the granite body is about 10 km (6 mi) wide from north to south and is exposed over an area of about 153 km^2 (59 mi^2). The granite was intruded into folded, metamorphosed clastic sediments of the Narragansett basin (Nichols, 1956; Quinn, 1971).

8.4.1.2 Relative and Absolute Age

The Narragansett Pier granite has been dated at 234 ± 23 m.y. by use of the Pb-α method (Quinn et al., 1957) and 273 m.y. by use of the U-Pb method on monazite (Kocis et al., 1978). It intruded into Pennsylvanian metasediments of the Rhode Island formation (Nichols, 1956; Quinn, 1971) and is the youngest plutonic rock in southeastern New England. Hermes and Zartman (1982) related it to Permian plutonic activity.

8.4.1.3 Origin

Emplacement by forceful injection is suggested by the bending of schistosity in the Rhode Island formation, parallel to the rounded ends of converging sills of the Narragansett Pier granite (Nichols, 1956).

8.4.1.4 Petrography

Mineralogy and Textural Relations. The composition of the Narragansett Pier granite is that of a quartz monzonite to granodiorite. It is typically reddish in color but may be tan or gray. The granite is medium grained and equigranular but locally porphyritic. It may be massive or weakly foliated. The composition of the granite is fairly uniform, consisting of microcline, 30-35%; oligoclase, 30-35%; quartz, 25-30%; biotite, 3-5%; and minor muscovite (Quinn, 1971). Accessory minerals include magnetite, pyrite, and spessarite. A border facies of the granite may be distinguished by its gray-white color and higher content of potassium feldspar, garnet, and muscovite, and lower content of oligoclase and biotite. Rotated inclusions of schist from the surrounding country rock are found locally and are associated with a higher biotite concentration and a gneissic structure in the vicinity of the contact (Nichols, 1956).

Alteration. Secondary alteration may be minimal, but data are sparse.

8.4.1.5 Secondary Intrusions

In addition to the pegmatite and aplite scattered in patches through the Narragansett Pier granite, secondary intrusions consist of dikes of Westerly granite and lamprophyre.

8.4.1.6 Country Rocks

The Narragansett Pier granite was intruded into the folded clastic rocks of the continental Rhode Island formation. This formation consists of schistose conglomerate and pebbly sandstone, with interstratified muscovite schist and meta-anthracite. The primary constitutents are quartz, muscovite, and biotite, with minor plagioclase, garnet, chlorite, zircon, graphite, ilmenite, and apatite. Quartzite and quartz pebbles are elongated (Nichols, 1956). The Narragansett Pier granite also was intruded into the Hope Valley alaskite gneiss, and amphibolite and other rocks of the Plainfield formation.

8.4.1.7 Rock Structure

Irregularly spaced joints have been observed to cut the Narragansett Pier granite (Nichols, 1956). A local, weakly developed planar and linear structure observed in the Narragansett Pier granite is believed to be a primary flow structure (Quinn, 1971). The Narragansett Pier granite was emplaced into folded rocks of the Narragansett basin, but does not appear to have been folded itself.

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9 CRYSTALLINE ROCKS OF CONNECTICUT

9.1 GRENVILLE AND AVALONIAN

9.1.1 Bear Hill Massif (Plate VIII, Pluton 6)

9.1.1.1 Size, Shape, and Location

The Bear Hill massif comprises three irregular bodies, which are generally elongated to the northeast. Outcrops occur over an area of about 148 km^2 (57 mi^2) in western Connecticut in the Kent, New Milford, New Preston (see Fig. 9.1), Washington, Brookfield, New Fairfield, Warren, and Danbury quadrangles. The massif was referred to as the gneiss of the highlands massifs by Rodgers (1982) and as part of the Hudson Highlands by Jackson and Hall (1982) and Clark and Kulp (1968). It is probably a northeastern extension of the much more extensively described Manhattan Prong.

9.1.1.2 Relative and Absolute Age

The Bear Hill massif is composed of several gneissic lithologies of Proterozoic age (Rodgers et al., 1959; Jackson and Hall, 1982; and Rodgers, 1982). The granite gneisses have been correlated with the Fordham and Yonkers gneisses of the Manhattan Prong. The gray biotite gneiss unit is stratigraphically equivalent to the Fordham gneiss, for which a Pb^{207}-Pb^{206} age of about 980 m.y. was reported by Grauert and Hall (1973). Long (1969) reported a Rb-Sr whole-rock age of 575 ± 30 m.y. for the Yonkers gneiss. Clark and Kulp (1968) and Dallmeyer and Sutter (1976) stated that a K-Ar biotite age of 355 m.y. shows that a Devonian thermal event affected the rocks of the area.

9.1.1.3 Origin

Rocks of the Bear Hill massif are part of the Precambrian basement of the area. The granite gneisses are probably the product of high-grade metamorphism of rhyolitic and arkosic material (Mose and Hayes, 1975). The amphibolites are probably metamorphosed mafic volcanics.

9.1.1.4 Petrography

Mineralogy and Textural Relations. The rocks of the Bear Hill massif are lithologically heterogeneous. Dana (1977) described the massif as composed of gray biotite gneiss, augen gneiss, and pink granitic gneiss. Rodgers (1982) expanded on this classification. The following three descriptions are excerpted from Dana's 1977 study.

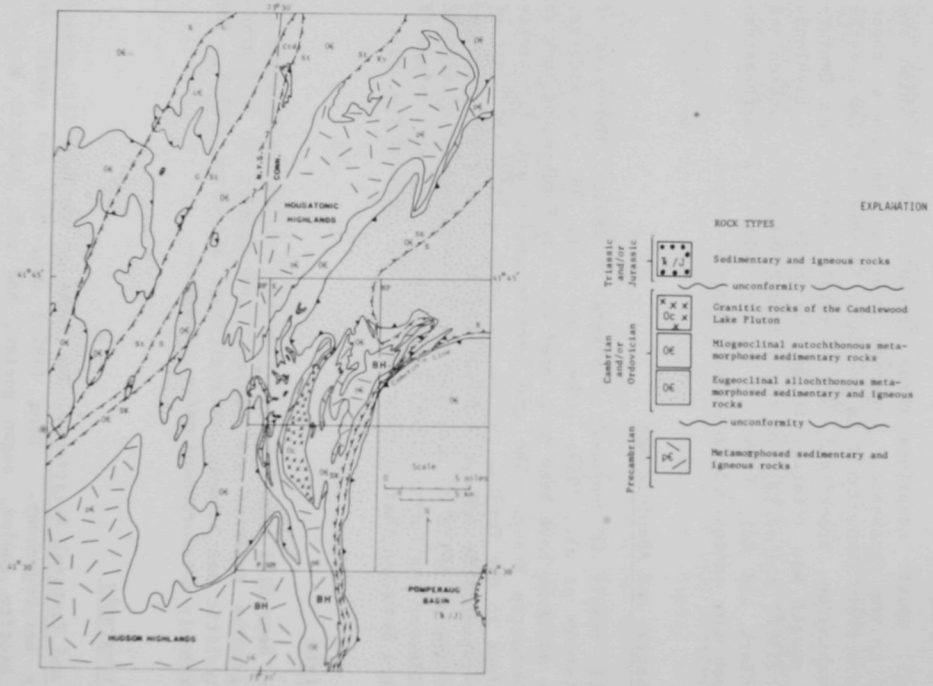

Fig. 9.1 Geologic Map of Western Connecticut and Adjacent Eastern New York ("BH" identifies units of the Bear Hill massif in Connecticut) (Source: Adapted from Jackson and Hall, 1982)

The gray biotite gneiss unit is compositionally heterogeneous, but the pale-pink to gray, fine- to medium-grained biotite-quartz-plagioclase gneiss is the most common facies. Its modal composition is quartz, 21%; plagioclase, 20%; and biotite, 2%. Layers 10-25 cm (4-10 in.) thick of black quartz-plagioclase-biotite schist are common in the unit. A dark-greenish-gray, medium-grained variety has a modal composition of quartz, 75%; plagioclase, 58%; microcline, 3%; biotite, 19%; magnetite, 2%; and sphene; with trace amounts of muscovite, garnet, hornblende, apatite, and zircon. Greenish-gray, medium- to coarse-grained biotite-hornblende gneiss that contains graphite and lavender garnets are found locally. Its modal composition is plagioclase, 55%; biotite, 20%; epidote, 10%; and hornblende, 15%; with trace amounts of apatite, sphene, quartz, and magnetite. Pale-pink or white, fine- to medium-grained biotite-muscovite-quartz-plagioclase-microcline gneiss, common in the southwest, is interbedded with dark-green, medium-grained, well-foliated amphibolite layers 0.1-5 cm (0.04-2 in.) thick and with granitic biotite-feldspar-quartz gneiss layers. The modal composition of this gneiss is quartz, 14-40%; plagioclase, 20-35%; and microcline, 39-49%; with trace amounts of biotite, muscovite, chlorite, epidote, apatite, zircon, sphene, and magnetite. The modal composition of the amphibolite is quartz, 4%; plagioclase, 38%; and hornblende, 57%; with minor to trace amounts of biotite, epidote, apatite, zircon, and sphene.

The augen gneiss is a layer of pale-pink to light-gray, medium-grained, well-foliated biotite-plagioclase-quartz-microcline granitic gneiss with a consistent thickness of 50 m (164 ft). Pink microcline augen up to 4 cm (1.6 in.) long are common. Quartz, plagioclase, and garnet augen are also found. Three percent of the rock unit is made up of ilmenite and magnetite crystals up to 1.5 mm (0.06 in.) across. The pale-pinkish-gray type of gneiss contains more plagioclase than the pink type. Its modal composition is quartz, 39%; plagioclase, 2-7%; microcline, 35-49%; biotite, 8-9%; and muscovite, 1-7%; with trace amounts of chlorite, pistacite, apatite, magnetite, and ilmenite.

The pink granitic gneiss is a fine-grained biotite-plagioclase-quartz-microcline granitic gneiss. Two percent of the rock unit is made of ilmenite and magnetite crystals less than 0.5 mm (0.02 in.) across. Locally, hornblende is present. The modal composition of the gneiss is quartz, 33%; plagioclase, 17%; microcline, 44%; and biotite, 3%; with minor amounts of muscovite, sphene, magnetite, and ilmenite.

Alteration. Retrogressive metamorphic effects in the Bear Hill gneisses are preserved in the partial alteration of biotite to chlorite and of sillimanite to muscovite.

9.1.1.5 Secondary Intrusions

Granitic and pegmatite intrusions associated with the Acadian orogeny are present.

9.1.1.6 Country Rock

The Bear Hill massif is probably part of the Grenville metamorphic terrane and has an uncomformable contact with early Paleozoic metamorphic rocks. To the southeast, it is separated from the early Paleozoic metamorphosed eugeosynclinal sequence (Hartland formation) by Cameron's line. The Precambrian rocks of the Hudson Highlands and the Manhattan Prong lie to the south and southwest; those of the Housatonic Highlands lie to the north (see Fig. 9.1).

9.1.1.7 Rock Structure

Although Precambrian thermotectonic events may have modified the rocks of the Bear Hill massif, evidence thereof is meagerly preserved because of overprinting and obliteration associated with the Taconic and Acadian orogenies. The Taconic orogeny probably caused the early, first-stage isoclinal folding and sillimanite-grade metamorphism, but the present structural patterns evolved largely during the Acadian orogeny (Dana, 1977). In the Lake Waramaug area, the major Bear Hill-stage folds (from southeast to northwest) are the Woodville anticline, the Aspetuck syncline, the Bear Hill anticline, and the Eel Pond syncline (see Fig. 9.2). Axial planes strike northeast and dip 71-82° NW. All folds except the Eel Pond syncline are isoclinal and overturned to the east.

The Bear Hill anticline has a curved axial plane that strikes N. 49-60° E. and dips 74-81° NW. The curve may be due to late-stage deformation. The orientations of the minor fold axes are consistent with the hinge of the anticline (Dana, 1977). Fold axes generally trend N. 16 W. and dip 56° NW. Minor fold axes often fan out within the steeply dipping axial plane. Some nearby folds have a counterclockwise sense of rotation. The Waramaug, the Above All, and the Baldwin Hill are major thrust faults formed near the Bear Hill anticline, but they do not affect the Precambrian gneisses.

9.1.1.8 Geophysics

Cameron's line, a major tectonic and stratigraphic boundary, lies to the east and southeast of the Bear Hill massif. Regional gravity data show that the area of the massif is enclosed by gravity contours of -10 mgal to -20 mgal (Hildreth, 1979). The regional gravity trend is to the northeast.

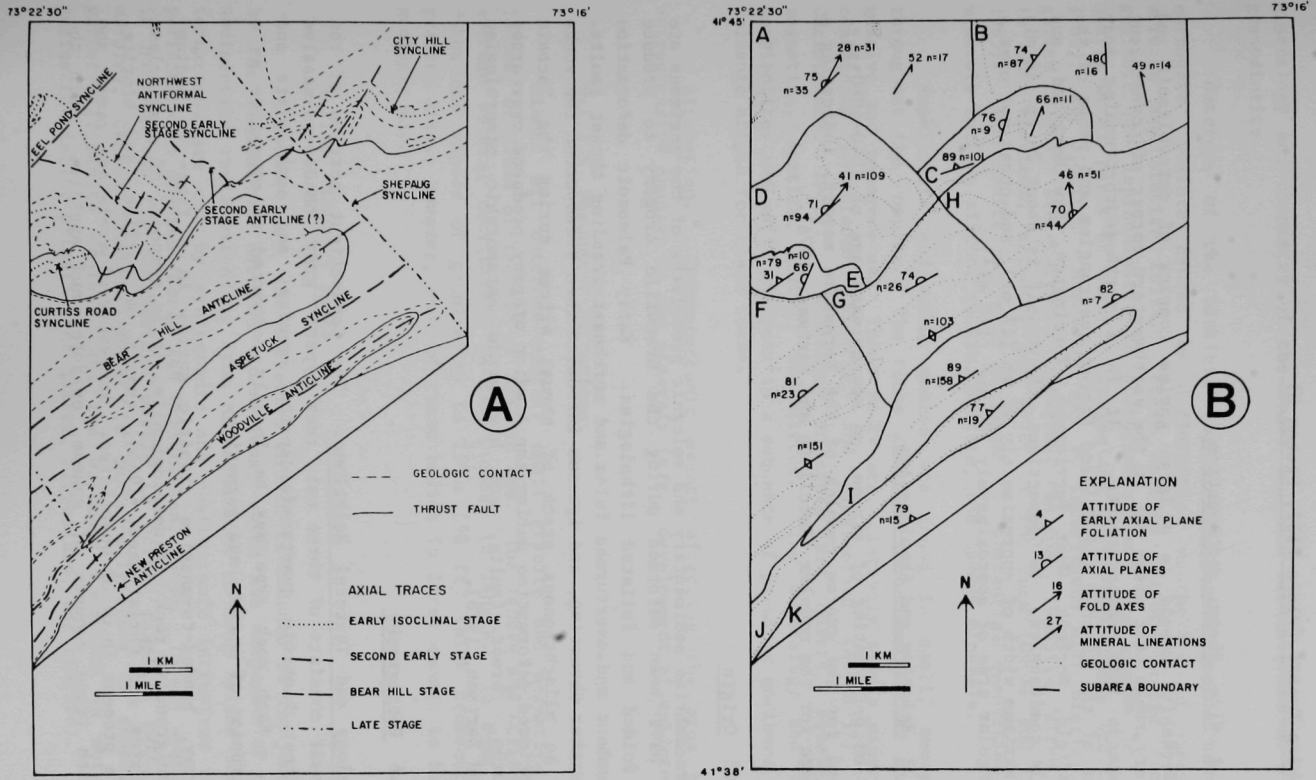

Fig. 9.2 Geologic and Tectonic Maps of the Lake Waramaug Area (A: axial traces of major folds; B: axial plane foliation and Bear Hill-stage structural trends; and "n" = ..." denotes number of observations averaged for each orientation datum) (Source: Dana, 1977)

9.1.2 Berkshire Massif (Plate VIII, Pluton 8)

9.1.2.1 Size, Shape, and Location

The Berkshire massif is a large gneiss complex of metavolcanic and metasedimentary rocks that crops out in the northwestern corner of Connecticut but also extends into Massachusetts. It is a north-south-trending body covering most of the Winsted and Tolland Center quadrangles and parts of the South Sandisfield, Norfolk, West Torrington, Torrington, New Hartford, and West Granville quadrangles. In Connecticut, the massif is about 26 km (16 mi) long, has a maximum width of approximately 27 km (17 mi), and crops out over a roughly triangular area of approximately 290 km^2 (110 mi^2).

9.1.2.2 Relative and Absolute Age

Based on U-Pb dating of zircon, an approximate age of 1040-1080 m.y. was calculated for the gneiss complex of the Berkshire massif, indicating a Precambrian age for the complex (Ratcliffe and Zartman, 1976).

9.1.2.3 Origin

The Precambrian sedimentary and volcanic protoliths of the gneisses are believed to have been deformed during the Grenville orogeny to produce isoclinally folded and foliated lithologies. Early Paleozoic deformation produced recumbent and overturned folds and northeast-trending thrust faults. Refolding, further thrusting, and igneous intrusion then transported the rocks westward as an allochthonous stack of thrust slices during the Taconic orogeny. Continued deformation during the Acadian orogeny produced overturned fold structures, thrust faults, and high-grade metamorphic mineralogies (Ratcliffe and Zartman, 1976).

9.1.2.4 Petrography

Mineralogy and Textural Relations. The Connecticut portion of the Berkshire massif consists of three main rock types: gray, banded-to-massive granitic gneiss; biotitic quartzo-feldspathic gneiss; and amphibolite and mafic gneiss. Each rock type may be locally converted to migmatite as a result of intrusion of granite and pegmatite.

The gray, banded-to-massive granitic gneiss is a fine- to medium-grained, inequigranular rock. Mineralogically, it is composed of plagioclase (An-27-29) (Gates and Christensen, 1965), quartz, microcline, and biotite, with muscovite present in minor quantities. Accessory minerals are garnet and magnetite, with trace amounts of sphene, epidote, and apatite. Both the

mineralogy and chemistry of the rocks indicate metamorphism of a granite or granodiorite.

Outcrops of the biotitic quartzo-feldspathic gneiss have a rough, corrugated texture caused by differential weathering of quartz-feldspar streaks and biotite folia. Fresh surface textures range from strongly gneissic and lepidoblastic to porphyroblastic. The gneiss is composed of nearly equal parts of quartz, plagioclase (An-25-33) (Gates and Christensen, 1965), and biotite. Muscovite may be interleaved with biotite, and small garnet crystals are scattered throughout the rock. Kyanite, although not seen in hand specimen, accounts for 2-10% of the rock (Gates and Christensen, 1965). Sillimanite may also be found in substantial quantities; accessory minerals include zircon, rutile, apatite, and opaque minerals.

Amphibolite and mafic gneiss are found in small, unmapped layers throughout the complex. The rocks are fine to medium grained and typically massive and homogeneous. These rocks are relatively simple mineralogically, containing hornblende, biotite, and plagioclase (An-28-37) (Gates and Christensen, 1965). Accessory minerals are quartz, sphene, epidote, magnetite, apatite, garnet, and zircon. The mineralogy and layering of amphibolites and mafic gneisses in a sequence of granitic gneisses indicates a volcanic origin for these rocks.

Alteration. Thermotectonic events affecting the area have caused retrogressive changes in the minerals (e.g., development of epidote).

9.1.2.5 Secondary Intrusions

Almost all outcrops exhibit small granite and pegmatite stringers, lenses, or dikes of Taconic or Acadian age. The lenses and stringers tend to average several centimeters (a few inches) in length. In some localities, dikes and masses of granite up to 15 m (50 ft) thick crosscut the small granite dikes; however, none of these bodies is large enough to be shown on a geologic map.

9.1.2.6 Country Rocks

In Connecticut, the Berkshire massif is bounded to the east and south by the Hoosac and Hartland formations, which are composed of mixed metasedimentary rocks extending from Massachusetts to Long Island Sound. These formations are cut by syntectonic and posttectonic intrusives ranging from peridotite to granite (Gates and Christensen, 1965). To the west, the massif is bordered by the Stockbridge marble (a partly clastic, partly dolomitic marble), Canaan Mountain schist (a quartz-oligoclase or andesine-biotite-garnet-sillimanite schist), and the Waramaug formation (a metasedimentary gneiss consisting primarily of quartz, oligoclase, and biotite).

9.1.2.7 Rock Structure

Repeated deformation has produced stacked and folded thrust fault slices; northwest-trending, high-angle reverse faults; and northwest- and north-trending normal faults in the gneiss complex (Ratcliffe and Zartman, 1976). All rock types described exhibit secondary foliation caused by regional metamorphism. In some instances, the different units can be distinguished by slight differences in foliation.

The gneiss complex of the Berkshire massif forms a large southward-plunging antiform. Intense deformation and multiple folding are observed in outcrops of all of the units described, indicating that the internal structure of the massif is extremely complex.

9.1.2.8 Geophysics

An aeromagnetic map of the Torrington quadrangle was compiled by Philbin and Smith (1966). It shows a pronounced north-south magnetic trend in the gneiss complex, ending rather abruptly along an almost east-west line in the north-central part of the quadrangle (Martin, 1970). Aeromagnetic investigations of the regional structure of Precambrian rocks in this area have been done by Harwood and Zietz (1974).

9.1.3 Housatonic Massif (Plate VIII, Pluton 7)

9.1.3.1 Size, Shape, and Location

The Housatonic massif, located in western Connecticut, is a well-layered, metamorphosed, acidic volcanic sequence, with subordinate interbedded mafic volcanics and terrigenous clastics. These rocks are probably an extension of the Green Mountains of Vermont (Gates and Martin, 1976). The massif is somewhat elongated to the northeast; outcrops occur over an area in Connecticut of about 243 km^2 (94 mi^2). It is estimated that the gneiss unit contained within the isoclinal fold is at least 5 km (3 mi) thick, or thicker.

9.1.3.2 Relative and Absolute Age

The rocks of the Housatonic massif are designated Precambrian on the basis of lithostratigraphic relations, and may be Proterozoic Y, on the basis of U-Pb age estimates on zircon from the Fordham gneisses (Grauert and Hall, 1973; Rodgers, 1982). The Housatonic gneisses are lithologically similar to those of the Berkshire allochthon and many of the rocks of the Fordham terrane (Hall, 1980). It is likely that there is equivalence among several of these units (Gates and Martin, 1976; Hall, 1980).

9.1.3.3 Origin

Rocks of the Housatonic massif are considered part of the Grenville basement and represent gneisses derived from sedimentary, volcanic, and intrusive rocks. Gates and Martin (1976) stated that (1) the layered granitic gneiss represents metamorphosed felsic volcanic tuffs, (2) the amphibolite and mafic gneiss represent mafic metavolcanics, and (3) the well-layered gneissic lithology is indicative of a metamorphosed volcaniclastic and minor terrigenous clastic sequence.

9.1.3.4 Petrography

Mineralogic and Textural Relations. In general, the rocks of the Housatonic massif are highly heterogeneous and anisotropic. They are mostly fine to medium grained, but their textures and compositions are highly variable. Gates and Martin (1976) recognized three mineralogical types: granitic gneisses, quartz-plagioclase-biotite gneisses, and mafic to amphibolitic gneisses. Their modal analyses are shown in Fig. 9.3.

The granitic gneisses are rich in potassium feldspar compared with the quartz-plagioclase-biotite gneisses (see Fig. 9.3). They are particularly well layered, with layers ranging from a few centimeters to meters (inches to yards) thick. The quartz-plagioclase-biotite gneisses, on the other hand, are not well layered but are relatively rich in mica that occurs as layers, wisps, or knots. The biotite/muscovite ratio is 3:1 in the gneisses. The differences between the gneisses can be attributed to the composition of the source material or to the effects of subsequent thermotectonic adjustments.

Amphibolites and mafic gneisses are ubiquitous, but relatively subordinate, and characteristically contain plagioclase, hornblende, and sphene. When hornblende is a minor mineral, biotite becomes a major one. The subordinate diopside-quartz-graphite granulites, the calc-silicate rocks, and the kyanite-sillimanite-mica-quartz-plagioclase gneisses are all interlayered with the granitic gneisses and amphibolite. Such sequences are lithologically similar to the Waramaug formation.

9.1.3.5 Secondary Intrusions

No significant secondary intrusions are described in the literature.

9.1.3.6 Country Rock

The rocks overlying the Housatonic massif are the Manhattan schist, including the Waramaug formation and Canaan Mountain schist. To the southeast, Tyler Lake gneiss occurs above the Waramaug; to the northwest lies Salisbury schist, Stockbridge marble, and Poughquag quartzite (Rodgers, 1982).

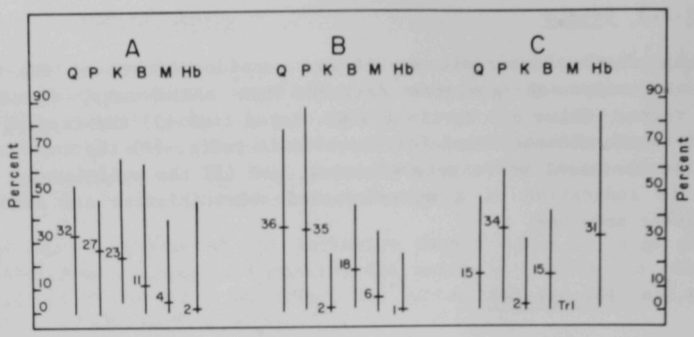

Fig. 9.3 Modal Analyses of the Housatonic Gneisses (range of composition indicated by A: granitic or potassium-feldspar-rich gneiss; B: biotite-feldspar-quartz gneiss; C: amphibolitic and mafic gneiss; with Q = quartz, P = plagioclase, K = potassium feldspar, B = biotite, M = muscovite, and Hb = hornblende) (Source: Adapted from Gates and Martin, 1976)

The geologic relations in the west-central part of Connecticut containing part of the Housatonic massif have been studied by Gates (1961 and 1975) and are shown in Fig. 9.4. In the study area, the Housatonic gneisses are unconformably overlain by the metamorphic equivalent of a sandstone-carbonate-shale sequence of the Dalton, Stockbridge, and Walloomsac formations. Layering in the gneiss complex is usually parallel to the structurally overlying Waramaug formation; however, at some locations there seems to be an interlayering of the two formations.

9.1.3.7 Rock Structure

Gates and Martin (1976) reported that intralayer isoclinal folds are present in the Housatonic Highlands, but the duplication of mappable units is not recognizable. Hall (1980) assessed the basement and cover-rock relationships south of the Hudson Highlands and interpreted their deformational history. Hall stressed that five phases of deformation during the Taconic and Acadian orogenies caused multiple folding and faulting involving both the basement and cover rocks. Similar deformations probably affected the Housatonic area as well, but the intensity of deformation may have been different in this region.

High-angle reverse faults border the northwest side of the massif (Rodgers et al., 1959). The contact with the Dalton formation in the South Canaan quadrangle is possibly a thrust feature (Gates and Martin, 1976).

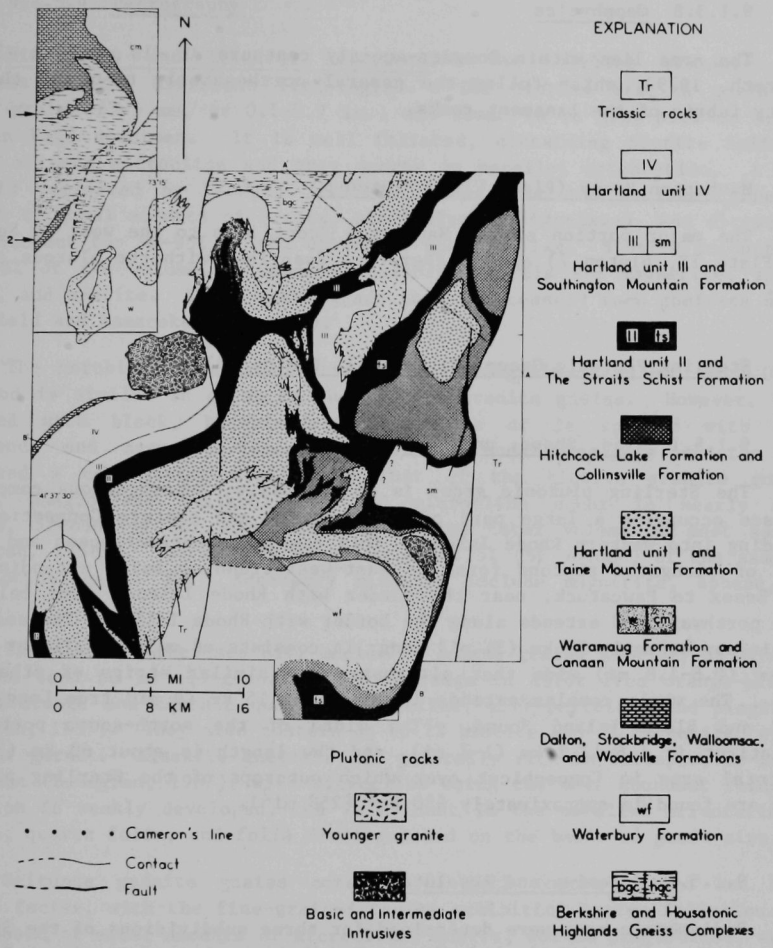

Fig. 9.4 Geologic Map of West-Central Connecticut (the Housatonic massif crops out in parts of the South Canaan [1] and Cornwall [2] quadrangles) (Source: Adapted from Gates and Martin, 1976)

9.1.3.8 Geophysics

The area lies within Bouguer-anomaly contours of -10 mgal to -30 mgal (Hildreth, 1979), which follow the general northeasterly trend of the low-gravity fabric of the basement rocks.

9.1.4 Manhattan Prong (Plate VIII, Pluton 5)

The major portion of the Manhattan Prong lies to the west in New York (see Plate IX, pluton 7) and is therefore described with the plutons of that state.

9.1.5 Sterling Plutonic Group (Plate VIII, Pluton 13)

9.1.5.1 Size, Shape, and Location

The Sterling plutonic group is an extensive, heterogeneous complex of gneisses occupying a large part of southeastern and eastern Connecticut and extending into western Rhode Island. This complexly metamorphosed and folded group of elongated plutons forms an east-west-trending belt, extending from near Essex to Pawcatuck, near the border with Rhode Island. The belt then turns northward and extends along the border with Rhode Island. The east-west belt is approximatly 55 km (34 mi) long; it consists of multiple linear strips 1-3 km (0.6-1.8 mi) wide that alternate with similar strips of other rock units. The whole complex extends inland about 15 km (9 mi) from Long Island Sound and Block Island Sound. The width of the north-south portion in Connecticut is 1.6-11.2 km (1-7 mi), and the length is about 60 km (37 mi). The total area in Connecticut over which outcrops of the Sterling plutonic group are found is approximately 650 km^2 (250 mi^2).

9.1.5.2 Relative and Absolute Age

Radiometric ages were determined for three subdivisions of the Sterling plutonic group (Hope Valley alaskite, Ten Rod granite gneiss, and Scituate granite). Zircons from these rocks yielded a late Precambrian age, with the data indicating emplacement during the Avalonian orogeny (Hermes et al., 1981). Other plutons of the Sterling plutonic group are believed to be contemporaneous with these.

9.1.5.3 Origin

The plutons of the Sterling plutonic group were probably emplaced as sills parallel to foliation and layering in the country rock, and then folded to follow the present configuration of the Lake Char and Honey Hill fault zones, which curve from south to west in southeastern Connecticut (Barosh and Hermes, 1981).

9.1.5.4 Petrography

Mineralogy and Textural Relations. The Sterling plutonic group comprises several different lithologies. The biotite granite gneiss has coarse grains (3-5 mm, or 0.1-0.2 in.) and tends to be gray-pink to orange-pink in hand specimen. It is well foliated, containing biotite folia and grains of pink microcline and gray quartz in parallel orientation. A modal analysis performed by Lundgren (1966a) showed that the gneisses generally consist of equal amounts of quartz, plagioclase (oligoclase), and microcline, which account for 90-95% of the volume of most samples. Biotite accounts for about 5% of the volume. Common accessory minerals are magnetite, zircon, garnet, and apatite. These rocks grade into the country rock gneisses of the Plainfield and Mamacoke formations.

The hornblende-granite gneiss has medium grains (1-3 mm, or 0.04-0.1 in.) and is similar in color to the biotite granite gneiss. However, it is streaked with black, hornblende-rich laminae or is spotted with black hornblende and magnetite-ilmenite grains. Modal analysis of the rock disclosed a composition similar to that of the biotite granite gneiss. Quartz, plagioclase (oligoclase), and microcline occur in nearly equal amounts, accounting for more than 90% of the total volume of most samples (Lundgren, 1966a). Hornblende makes up about 5% (by volume) of the samples, and biotite up to 3%. Accessory minerals include magnetite, sphene, and zircon.

The alaskite gneisses have medium to coarse grains (1-4 mm, or 0.04-1.6 in.) and tend to be equigranular. They range in color from orange-pink to grayish orange and contain quartz, microcline, microperthite, and plagiocase (Lundgren, 1967). They also contain up to 2% biotite and 1% magnetite, with a trace of garnet. Alaskite gneisses are generally rich in feldspar, up to 50% by volume (Lundgren, 1967), with microcline being the most abundant feldspar. Foliation is weakly developed, but is evident in the parallel orientation of biotite, quartz folia, and folia distinguished on the basis of grain size.

Scituate granite gneiss consists of both coarse-grained and fine-grained facies, with the fine-grained facies exhibiting better foliation. It is composed of equal amounts of microcline, quartz, sodium plagioclase, and 5-15% biotite, with minor amounts of magnetite, zircon, apatite, allanite, and chlorite after biotite. Some samples are locally porphyritic (Harwood and Goldsmith, 1971).

The Hope Valley alaskite gneiss is generally light gray to flesh-colored and fine to medium grained. It is composed of nearly equal amounts of microcline, quartz, and plagioclase. Biotite (1-5%) is fine grained, and the gneiss may contain up to 5% microcline phenocrysts as long as 5 mm (0.2 in.). Samples of these rocks usually display strong foliation and lineation formed by the parallel orientation of biotite flakes (Harwood and Goldsmith, 1971).

The Ten Rod granite gneiss is a medium- to fine-grained, porphyritic, strongly foliated granite gneiss. It is composed of approximately equal amounts of microcline, quartz, and sodium plagioclase. Biotite content ranges from 2% to 8%; accessory minerals are magnetite, apatite, zircon, allanite, chlorite, and epidote. This gneiss commonly contains large phenocrysts of microcline, ranging in abundance from 5% to 30%.

Alteration. Weathering imparts a rusty and friable character to the alaskite rocks; most of the biotite has been altered to chlorite in the Westerly-type granite dikes.

9.1.5.5 Secondary Intrusions

Granitic and pegmatitic dikes intruding into the Sterling plutonic group are of the Westerly or the Black Hall type. Westerly-type dikes are generally fine grained and equigranular. They are considered to be quartz monzonites, with plagioclase (oligoclase) and microcline being nearly equal in abundance. Muscovite is present, formed by extensive alteration of oligoclase; micrographic intergrowths of microcline and quartz may be present. Locally, the rock contains graphic granite. Most of these dikes are 10 cm (4 in.) to 1.2 m (4 ft) wide, but some are wider.

Dikes of the Black Hall type are the most common (Lundgren, 1967). They are pegmatitic biotite granites with a large variation in grain size and texture. The rock is characteristically massive and exceptionally coarse grained, and contains microcline, plagioclase, quartz, and randomly oriented biotite laths. Plagioclase and quartz tend to be subordinate to microcline, occurring as interstitial grains or intergrown with larger microcline crystals. Most discordant parts of these dikes are aligned east-west and dip steeply.

9.1.5.6 Country Rocks

The gneisses of the Sterling plutonic group are surrounded by the Plainfield and Mamacoke formations. The Plainfield formation comprises many different interbedded metasedimentary and metavolcanic rocks. The Mamacoke formation consists predominantly of medium-grained, gray, banded gneiss. In some places, the formation also includes biotite gneisses and schists, quartzite, and amphibolite.

9.1.5.7 Rock Structure

The area containing the Sterling plutonic group is highly faulted; two major faults are recognized -- the Honey Hill fault and the Lake Char fault. The Honey Hill fault is a major thrust fault that trends in a northwestern to western direction. Evidence indicates that the hanging wall was underthrust

northward relative to the footwall. The dip of the fault plane ranges from 15° to 50° N. in the Fitchville quadrangle (Snyder, 1964b). The trace of the fault is marked by a zone of mylonitic and blastomylonitic gneisses. The Lake Char thrust fault is believed to be part of the same fault system. The Lake Char fault trends north; its hanging wall is displaced easterly relative to its footwall.

The Sterling plutonic group has two prominent joint sets. The best-developed set is nearly vertical and strikes approximately north (±20°). The second set also is steeply dipping but strikes east (±10°) (Lundgren, 1967). The east-trending joints are usually filled with pegmatite or granite, whereas the north-striking joints are not.

All of the rock types described exhibit various degrees of metamorphic foliation. The granitic gneisses may exhibit an augen texture, particularly in the highly biotitic phases, but foliation in the alaskite gneiss is well developed only near its contact with biotitic gneisses of the Plainfield formation (Lundgren, 1967).

The rocks of the Sterling plutonic group and other formations in the area have undergone at least two episodes of folding. The plutons located along the Connecticut-Rhode Island border are part of a series of north-trending, north-plunging folds known as the West Rhode Island fold belt (Barosh and Hermes, 1981). Harwood and Goldsmith (1971) showed that the biotite lineation of the Sterling plutonic group within the Oneco quadrangle plunges parallel to fold axes within the Plainfield formation, indicating folding about a common axis. In the northern reaches of the belt, the folds are broad and open; in the southern sections, the folds are compressed and are eventually overturned and intercepted by thrust faulting.

9.1.5.8 Geophysics

Aeromagnetic maps of all Connecticut quadrangles have been published by USGS. The following specific maps may be of significance:

1. Boynton, G.R., and C.W. Smith, Aeromagnetic Map of the Jewett City Quadrangle, GP-539 (1966).

2. Boynton, G.R., and C.W. Smith, Aeromagnetic Map of the Plainfield Quadrangle, GP-541 (1966).

3. Boynton, G.R., and C.W. Smith, Aeromagnetic Map of the Oneco Quadrangle, GP-542 (1966).

4. Boynton, G.R., and C.W. Smith, Aeromagnetic Map of the Norwich Quadrangle, GP-543 (1966).

5. Boynton, G.R., and C.W. Smith, Aeromagnetic Map of the Old Mystic Quadrangle, GP-544 (1966).

6. Boynton, G.R., and C.W. Smith, Aeromagnetic Map of the Voluntown Quadrangle, GP-545 (1966).

7. Boynton, G.R., and C.W. Smith, Aeromagnetic Map of the Uncasville Quadrangle, GP-546 (1966).

8. Boynton, G.R., and C.W. Smith, Aeromagnetic Map of the Asbury Quadrangle, GP-547 (1966).

9. Philbin, P.W., and C.W. Smith, Aeromagnetic Map of the East Killingly Quadrangle and Part of the Oneco Quadrangle, GP-591 (1966).

10. Author unknown, Aeromagnetic Map of the Thompson Quadrangle and Part of the East Killingly Quadrangle, GP-653 (1969).

No interpretations are provided with these maps.

9.1.6 Waterford Gneiss Group (Plate VIII, Pluton 12)

9.1.6.1 Size, Shape, and Location

Rocks of the Waterford gneiss group occur as discontinuous, irregular bodies and multiple thin strips 1-2 km (0.6-1.2 mi) wide, which are complexly interdigitated with rocks of the Sterling plutonic group (see Sec. 9.1.5) in the southern and southeastern part of Connecticut. Collectively, the rocks of the Waterford gneiss group crop out over an area of about 385 km^2 (149 mi^2). Section 9.1.6 is based on studies in the New London area, where the total thickness of the upper sequence described by Goldsmith (1976), including the Monson and New London gneisses and the Mamacoke formation, is about 3100 m (10,170 ft). However, other parts of this complexly folded and metamorphosed rock group would likely differ in some respects.

9.6.1.2 Relative and Absolute Age

No radiometric age data are available for the group, and the structural and metamorphic complexity makes stratigraphic correlation difficult. Rodgers (1982) assigned a late Precambrian (Proterozoic Z) age to the Waterford gneiss group, but it may include some units equivalent to the Monson gneiss, which is thought to be Ordovician. Goldsmith (1976) considered the age to range from Precambrian to Ordovician.

9.1.6.3 Origin

The Waterford gneiss group originated through metamorphism of predominantly volcanic protoliths of mafic to felsic composition and some related intrusive rocks to gneisses, amphibolites, and schists. Metamorphosed volcaniclastic and sedimentary rocks form subordinate schistose and quartzite units.

9.1.6.4 Petrography

Mineralogy and Textural Relations. In the New London area, Goldsmith (1976) recognized that the upper sequence of the Waterford gneiss group consists of three major units: the Mamacoke formation, the New London gneisses, and the Monson gneiss.

The upper part of the Mamacoke formation contains thin white quartzite layers, which are underlain by sillimanite-bearing biotite gneiss containing lenses of massive amphibolite. In the lower but thicker part, the Mamacoke formation is composed of sparsely layered to well-layered biotite-quartz-feldspar gneiss. Plagioclase is common, but potassium feldspar is significant in the granite layers. Amphibolite layers are subordinate.

The New London gneiss unit is quite heterogeneous and is composed of at least three different facies. The first, or layered facies, consists of alternating layers of biotite-quartz-plagioclase gneiss and amphibolite, probably representing dolerite and basalt protoliths, respectively. The upper units within this facies are more prominently layered than the bottom units. The second, or massive, facies is relatively uniform in texture and composed of granodiorite, with ubiquitous magnetite. Rotated amphibolite blocks and inclusions indicate deformational flow. The third, or aegirine-augite gneiss unit, is essentially composed of alkalic granite.

The Monson gneiss unit is discussed here as well as in Sec. 9.2.6, for completeness, because Rodgers (1982) indicated that the Waterford gneiss group contains units equivalent to the Monson. The Monson gneiss consists of well-developed lens-shaped layers that are relatively uniform in composition and texture. The gray hornblende-biotite-quartz-plagioclase gneiss phase is locally massive, whereas the biotite-quartz-microcline-plagioclase gneiss phase contains lenticular bodies of amphibolite. Major minerals form clots and streaks. The upper part of the Monson gneiss on the east side of the Niantic River is thinly layered and displays wide variation in mafic mineral content.

Alteration. The metamorphic grade of the Waterford gneiss group in New London is higher than that of equivalent units in northeastern Massachusetts (Goldsmith, 1976). Major minerals show varying degrees of alteration, but details are lacking.

9.1.6.5 Secondary Intrusions

Secondary intrusions are not discussed in the literature, which may indicate that they are not significant.

9.1.6.6 Country Rocks

The Waterford gneiss group, which is part of the crystalline basement in Connecticut, is interdigitated with the Sterling plutonic group. It is truncated, in places, by younger Paleozoic units.

9.1.6.7 Rock Structure

The prominent layering characteristic of units of the Waterford gneiss group accounts for their generally high degree of heterogeneity. Although details of fabric analysis are lacking, the units are very intricately folded. The Honey Hill and Lake Char faults are regional features; other, locally intense, fault zones are observed.

9.2 TACONIC

9.2.1 Ayer Intrusive Complex (Plate VIII, Pluton 15)

A major portion of the Ayer Intrusive Complex lies in Massachusetts. Therefore, it is described in Sec. 7.2.2 (see Plate VI, pluton 12).

9.2.2 Brookfield Gneiss (Plate VIII, Pluton 4)

9.2.2.1 Size, Shape, and Location

The Brookfield gneiss is a discontinuous body located in western Connecticut. It includes five areas of outcroppings arranged in an arcuate pattern extending from Hawleyville northward to West Torrington. Their combined surface area is approximately 170 km^2 (66 mi^2). The largest outcrop is located southwest of the Housatonic River.

9.2.2.2 Relative and Absolute Age

The Brookfield gneiss is one of a group of gneisses associated with the Taconic orogeny in Connecticut and is therefore considered to be of Ordovician age (Rodgers, 1982).

9.2.2.3 Origin

The Brookfield gneiss, which was probably emplaced as a syntectonic igneous intrusion (Pease, 1982), has been subsequently modified by thermotectonic events.

9.2.2.4 Petrography

Mineralogy and Textural Relations. The Brookfield gneiss is a typical layered rock body. Some phases of it resemble the Harrison gneiss (see Sec. 9.2.5), a speckled, fine- to medium-grained diorite gneiss. Darker layers are composed of oligoclase-andesine, hornblende, variable amounts of biotite or epidote, small amounts of quartz and diopside, and minor amounts of garnet, sphene, and magnetite. Lighter layers are composed of oligoclase-andesine, biotite, and minor amounts of potassium feldspar. The layers of diorite gneiss are commonly cut by bodies of light-gray rock that have higher orthoclase content (Rodgers et al., 1959). Stocks of norite and local bodies of hornblende pyroxenite and hornblende peridotite also are present.

Alteration. Epidote formation indicates limited alteration of feldspar.

9.2.2.5 Secondary Intrusions

Near Mt. Prospect and West Torrington, dikes of biotite hornblendite and pyroxenite are associated with the gneiss (Rodgers et al., 1959).

9.2.2.6 Country Rocks

The Brookfield gneiss is mainly in contact with the Hartland formation, an interlayered mica quartzite and schist. No internal stratigraphic sequence has yet been applied to the Hartland formation.

Exposures of Bookfield gneiss near Bantam are in contact with the Waramaug formation as well as the Hartland formation. The former is composed of two gneissic rock types with gradational contacts. One of these is a coarse-grained and intensely deformed biotite-quartz gneiss. The latter is a less-deformed, light-colored, feldspathic mica quartzite or gneiss. Both are intruded by granite and pegmatite (Rodgers et al., 1959).

9.2.2.7 Rock Structure

Foliation in the area dips 20-85° W. or NW. and is parallel to the bedding, if visible. Minor isoclinal folding is present. A strong lineation plunges 10-50° NW. and may be related to late-stage deformation or to dome formation and intrusive activity (Rodgers et al., 1959).

9.2.2.8 Geophysics

No specific data are available. Regional Bouguer anomaly maps show gravity anomalies of +30 mgal to +40 mgal (Hildreth, 1979).

9.2.3 Glastonbury Gneiss (Plate VIII, Pluton 10)

9.2.3.1 Size, Shape, and Location

The Glastonbury gneiss occurs as a belt of northeasterly trending bodies that have maximum individual widths of approximately 6 km (4 mi). This belt extends from southern Massachusetts down through much of Connecticut (see Fig. 9.5). Within Connecticut, outcrops of Glastonbury gneiss are found over an area of about 222 km^2 (86 mi^2).

9.2.3.2 Relative and Absolute Age

In the Hampton quadrangle, the Glastonbury gneiss intrudes the Ammonoosuc volcanics of Middle Ordovician age. Brookins (1968) reported a Rb-Sr whole-rock age of 450 ± 10 m.y. for Ammonoosuc samples from New Hampshire. Also, a minimum age of 350 m.y. is indicated by pegmatite dikes in the gneiss (Brookins and Methot, 1971). Therefore, the Glastonbury gneiss is considered to be of Ordovician age.

9.2.3.3 Origin

On the basis of the parallelism of the Glastonbury gneiss with the Bolton, Hebron, and Amherst belts, which are all of known sedimentary origin, Collins (1954) proposed an origin by granitization of sedimentary and Ammonoosuc volcanic rocks. The gneiss also has certain similarities with the Oliverian domes studied by Naylor (1968 and 1969).

An intrusive igneous origin is indicated by the dips of the foliation, which suggests that the Bolton schist and the mafic gneiss common in the Middletown area have been cut off at depth by intrusion of the Glastonbury granite gneiss (Stugard, 1958). Stugard (1958) noted that there are inclusions of Bolton schist in the Glastonbury gneiss and that diopside occurs at the Bolton-Glastonbury contact. The presence of dark, oval, biotite-rich patches or schlieren along foliation planes in the bed of Roaring Brook, and the low percentage of potassium (Aitken, 1955), also support this view.

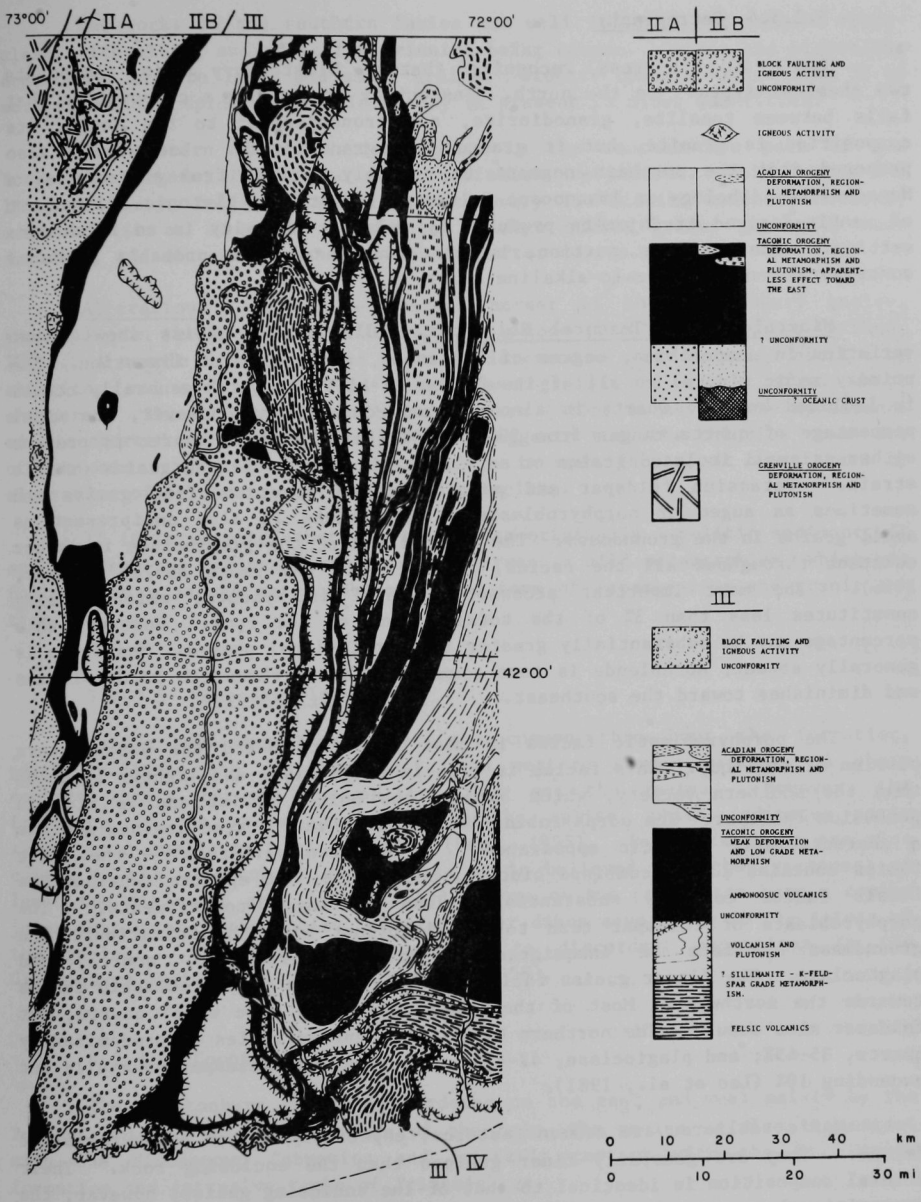

Fig. 9.5 Geologic Relations of Glastonbury Gneiss in Massachusetts and Connecticut (Source: Adapted from Robinson and Hall, 1980)

9.2.3.4 Petrography

Leo et al. (in press) recognized that the Glastonbury gneiss represents two chemical types: to the north, Glastonbury gneiss has a composition that falls between tonalite, granodiorite, and trondhjemite; to the south, its composition is granite, but it grades into granodiorite. Leo et al. also proposed that the northern segment is probably derived from anatexis of a Monson-like lithology or Ammonoosuc volcanics, or it is a plutonic counterpart of mantle-derived trodhjemite produced by volcanic activity in an island-arc setting. The southern portion, on the other hand, was probably produced contemporaneously from calc-alkaline crust.

Mineralogy and Textural Relations. Glastonbury gneiss shows great variation in composition, degree of foliation, and degree of lineation. The primary mafic mineral in all of its facies is biotite, which generally occurs in isolated knots. Quartz is also a common constituent; however, the modal percentage of quartz ranges from 20% to 60% (Herz, 1955). Quartz is present either as small isolated grains or as flaser, in which case the grains exhibit strain. Potassium feldspar and perthite often occur as fine grains and sometimes as augen or porphyroblasts. Plagioclase is also omnipresent as small grains in the groundmass. The composition of the plagioclase is rather constant throughout all the facies, remaining in the range Ab-72-75 (Herz, 1955). The most important accessory mineral is epidote, which usually constitutes less than 3% of the total volume of the rock. However, this percentage may be substantially greater in certain local phases. Muscovite is generally absent; hornblende is more common in the northwestern border facies and diminishes toward the southeast.

The porphyroblastic facies is most abundant toward the northwestern portion of the body. This facies is rich in plagioclase and lighter in color than the southern variety, which borders the Bolton schist and is rich in potassium feldspar. The porphyroblastic facies is very coarse grained and has a characteristic granitic appearance. It has two varieties -- the mafic facies contains 20-40% combined biotite, hornblende, and epidote, whereas the felsic facies contains substantially less of these constituents. The porphyroblasts of feldspar tend to be discordant with the foliation. The groundmass consists of inequigranular potassium feldspar, quartz, and plagioclase. The flaser gneiss facies grades into the porphyroblastic facies towards the northwest. Most of the flaser facies is made up of quartz, but feldspar also occurs. The northern Glastonbury gneiss facies is predominantly quartz, 35-45%; and plagioclase, 42-47%; with potassium feldspar generally not exceeding 10% (Leo et al., 1981).

Mafic schlieren are common features, especially in the porphyroblastic facies. They are generally finer grained than the enclosing rock. Their mineral composition is identical to that of the enclosing gneiss; however, the compositional percentages may differ.

The rocks in the southern facies are well foliated and contain oligoclase, with albite and pericline twinning being common. Quartz and microcline are common as both large and small grains. Biotite content varies from 3% to more than 20%. Epidote and zoisite may be present in minor quantities.

The eastern border facies occurs in the southeastern part of the formation. Again, felsic and mafic subfacies are present. This facies has little, if any, foliation and a comparatively low biotite content; however, muscovite is as abundant as biotite. Large quartz grains generally show suturing, and microcline is most abundant after quartz.

Alteration. Along the western border of the Glastonbury gneiss, especially in the Rockville quadrangle, is a zone of chloritic schist (Aitken, 1955). A sample from this zone shows replacement of very deformed biotite by chlorite, and some quartz and plagioclase grains are highly strained or shattered. The rock was described by Aitken (1955) as having the appearance of a contorted mylonite. This evidence suggests that the chlorite zones are of tectonic origin. Sericite also occurs in most facies; it was formed by alteration of feldspar, particularly plagioclase.

If it is accepted that the gneiss was originally a pelitic sediment, it appears that the sediments originally underwent a mild metamorphism, which was followed by a second metamorphism in the form of metasomatism and granitization.

9.2.3.5 Secondary Intrusions

Aplite and pegmatite dikes are common throughout the formation, particularly in the porphyroblastic and western border facies. A sharp contact exists between the aplite dikes and the gneiss. However, the foliation of the gneiss runs through the aplite, which indicates early syntectonic emplacement of the dike (Herz, 1955). Pegmatite dikes are also common in the Glastonbury gneiss, and it is believed that they represent at least two different ages (Herz, 1955). Some of the pegmatites have a common joint pattern with the gneiss, whereas other dikes seem to be in the joints of the gneiss. The ratio of concordant to discordant pegmatites in the Glastonbury gneiss is 1.5 to 1 (Stugard, 1958).

9.2.3.6 Country Rocks

The Glastonbury gneiss is bordered on the east and west mainly by the Bolton schist. On the west side of the body, the gneiss is also in contact with the Middletown formation and in fault contact with the Mt. Clough formation and intrusive rocks of Triassic age.

The Bolton schist is primarily a mica-quartz schist, with a typical modal composition of biotite, 50%; muscovite, 30%; quartz, 15%; oligoclase,

3-4%; and accessory minerals, 1%. Accessory minerals include pyrite, staurolite, and garnet (Herz, 1955). The contact between the Bolton schist belt and the Glastonbury gneisses is generally marked by quartzite that locally grades into quartz schist, depending on the sericite content.

The gneisses of the Middletown formation are generally well banded and fine grained, and contain biotite, hornblende, potassium feldspar, plagioclase, and quartz. The gneisses tend to have well-developed amphibolite schlieren; augen of feldspar are common.

9.2.3.7 Rock Structure

Fractures occur throughout the Glastonbury gneiss. Aitken (1955) grouped them into five main groups: a set trending N. 48° W. and dipping 67° SW.; a set trending N. 63° E. and dipping 47° NW.; a set trending N. 88° W. and dipping 87° NE.; a set trending N. 43° E. and dipping 40-60° NW.; and a set trending N. 7° E. and dipping 75° SE.

Herz (1955) described two joint sets in the Glastonbury quadrangle. One set strikes N. 75° W. and dips 87° SE., and the other strikes N. 35° E. and dips 75° SE.

The Triassic border fault lies along the contact between the volcanic rocks of Triassic age and the older gneisses, including the Glastonbury gneiss. The fault dips 60° W. (Herz, 1955). Other faults trend more or less parallel to this direction.

Foliation or schistosity is well developed in most facies. The schistosity follows the bedding and wraps around the noses of folds. This indicates that the foliation was formed not only by alignment of minerals normal to the direction of a compressive force, but also by growth of platy minerals parallel to bedding planes during recrystallization, and possibly by movement along shear planes within the rock (Collins, 1954). In the Rockville quadrangle, the foliation changes its trend from north-south to N. 57° E., with westerly dips ranging from 18° to vertical (Aitken, 1955).

There is also a quite noticeable lineation within the Glastonbury gneiss formed by the alignment of felsic minerals and biotite streaks. The lineation strikes approximately north-south and plunges north at no more than 22° (Aiken, 1955). Herz (1955) reported that the lineation in the Glastonbury quadrangle plunges N. 35° E.

The Glastonbury gneiss is part of a major anticlinal dome that can be traced into Massachusetts. As the axis of the dome is traced southward, it shows overturning to the east. The gneiss is part of a zone of crystalline rocks believed to be an overturned syncline to the east of the Glastonbury gneiss and an anticline in the gneiss itelf (Herz, 1955).

9.2.3.8 Geophysics

Aeromagnetic maps of all Connecticut quadrangles have been published by USGS. The following specific maps may be of significance:

1. Philbin, P.W., and C.W. Smith, Aeromagnetic Map of the Southern Coventry Quadrangle, GP-586 (1966).

2. Philbin, P.W., and C.W. Smith, Aeromagnetic Map of the Rockville Quadrangle, GP-587 (1966).

3. Author unknown, Aeromagnetic Map of the Ellington Quadrangle and Part of the Rockville Quadrangle, GP-648 (1969).

4. Quthor unknown, Aeromagnetic Map of the Stanford Springs Quadrangle and Part of the South Coventry Quadrangle, GP-649 (1969).

5. Author unknown, Aeromagnetic Map of the Marlborough Quadrangle and Part of the Rockville Quadrangle, GP-849 (1973).

6. Author unknown, Aeromagnetic Map of the Columbia Quadrangle and Part of the South Coventry Quadrangle, GP-850 (1973).

7. Author unknown, Aeromagnetic Map of the Moodus Quadrangle, GP-858 (1973).

No interpretations are provided with these maps. On the map compiled by Hildreth (1979), the Bouguer anomaly of the area is characterized by contour values of about -10 mgal to -20 mgal, which is in accordance with the lithologic setting and regional gravity trend or fabric.

9.2.4 Granitic Gneisses (Includes Mine Hill, Tyler Lake, Ansonia, and Siscowit Gneisses) (Plate VIII, Pluton 3)

9.2.4.1 Size, Shape, and Location

The Mine Hill granite gneiss and Tyler Lake granite are individual plutons. The former is a southwest-trending body in the Roxbury quadrangle in west-central Connecticut. The latter is an elongate, irregular body found mostly in the Cornwall and Warren quadrangles in northwestern Connecticut. Six other granitic bodies of probable Ordovician age crop out in western Connecticut and are considered to be related to the Mine Hill and Tyler Lake

plutons (Rodgers, 1982). The six outcrop areas were originally mapped as the same unit, but Gates (1959 and 1961) and Pease (1978) separated them in their later mapping. The largest of these units is roughly J-shaped and is located in the Ridge, Redding, Wilton, and Weston quadrangles in southwestern Connecticut (Rodgers, 1982). This southwesterly trending mass encompasses an area approximately 21 km (13 mi) wide from east to west and 24 km (15 mi) long from northeast to southwest. The whole group of eight bodies crops out over a total area of about 263 km^2 (101 mi^2). Because of limited information on the six smaller bodies, only the Mine Hill and Tyler Lake plutons are described in this section. The other bodies are presumed to be lithologically similar.

9.2.4.2 Relative and Absolute Age

The Tyler Lake, Mine Hill, and associated plutons are considered to be of Ordovician age and are apparently associated with the Taconic orogeny (Pease, 1982).

9.2.4.3 Origin

The Mine Hill granite gneiss was probably emplaced as a syntectonic intrusion, as indicated by its semiconcordant contacts with country rock (Gates, 1959). The Tyler Lake granite was probably emplaced as syntectonic or posttectonic intrusion.

9.2.4.4 Petrography

Mineralogy and Textural Relations. The Mine Hill granite gneiss is a uniform, fine-grained muscovite gneiss with minor biotite. Folia are characterized by variations in texture, relative abundance of sodium plagioclase or microcline feldspar, and planes of coarse muscovite. A xenomorphic, granular texture is typical of all but the micaceous layers of the rock. Cataclastic textures and replacement of microcline by plagioclase are visible in thin section (Gates, 1959).

The Tyler Lake granite is a massive, fine- to medium-grained granite with megacrysts of plagioclase, microcline, and quartz. It exhibits cataclastic, granitic, or gneissic textures, with lenses of granulated feldspars apparent in the gneissic texture (Gates, 1961).

Alteration. No data are available.

9.2.4.5 Secondary Intrusion

Gates (1959) reported that there are no crosscutting dikes in the Mine Hill granite gneiss. No information on secondary intrusions in the Tyler Lake granite is available.

9.2.4.6 County Rocks

The Mine Hill pluton is surrounded by members of the Hartland formation that include mica quartzite, mica-quartz schist, feldspathic mica quartzite, and feldspathic mica-quartz schist. These members are interbedded and gradational, so that the rock exhibits a massive to schistose texture. Planar elements range from slightly developed foliation to schistosity to primary layering. The thickness of individual layers ranges from several centimeters (a few inches) to about 61 m (200 ft). In the vicinity of the Mine Hill granite gneiss, the foliation of the Hartland formation generally strikes parallel to and dips away from the border of the pluton. However, southeast of the Mine Hill body, the foliation in the Hartland formation is more often crumpled, strikes generally to the northwest, and dips toward the southwest (Gates, 1959).

The Tyler Lake pluton is surrounded by gneiss of the Waramaug formation. Also, a small body of Waramaug gneiss is located within the pluton. The gneiss of the Waramaug formation includes interbedded and intergraded quartzo-feldspathic biotite gneiss and sillimanite-garnet-biotite quartzo-feldspathic gneiss. Concordant lenses of amphibolite are present locally and are most common near the northern end of the Tyler Lake body. The foliation of the gneiss is steep to vertical, and its strike generally parallels the outline of the Tyler Lake granite body (Gates, 1961).

9.2.4.7 Rock Structure

No description of jointing within the Mine Hill granite gneiss, Tyler Lake granite, or related rocks is available. A curved fault or coalescing faults, believed to follow the northern boundary of the Mine Hill pluton, have been mapped by Gates (1959). The throw of the fault may have been vertical, with numerous small, vertical faults are associated with it. No description of any faulting within the Tyler Lake granite is available.

9.2.4.8 Geophysics

The area displays a local gravity high on the Bouguer map. This configuration conforms to the regional gravity fabric (Hildreth, 1979).

9.2.5 Harrison Gneiss (Plate VIII, Pluton 1)

9.2.5.1 Size, Shape, and Location

The Harrison gneiss forms elongate, discontinuous masses cropping out over an area of approximately 146 km^2 (56 mi^2). The largest of these masses, in the Glenville quadrangle, is approximately 17 km (11 mi) long and 10 km (6 mi) wide, with an estimated thickness of more than 610 m (2000 ft), as recently estimated by Hall (1976). The Harrison gneiss trends northeast through the Mt. Vernon, Mamaroneck, Glenville, Stamford, and Norwalk South 7.5-min quadrangles of southeastern New York and southwestern Connecticut. To the southeast, it is bordered by Long Island Sound; on the other sides, it is bordered by the Hartland formation.

9.2.5.2 Relative and Absolute Age

Younger than the Hartland formation, the Harrison gneiss is considered to be Middle Ordovician (Gates and Martin, 1976; Hall, 1976; and Rodgers, 1982). Potassium-argon radiometic data indicate that the Harrison gneiss has gone through more than one thermal event -- one about 360 m.y. ago and one about 260 m.y. ago (Clark and Kulp, 1968). Correlation of the pre-Devonian lithostratigraphic units in western Massachusetts, Connecticut, and southeastern New York is shown in Fig. 9.6.

9.2.5.3 Origin

Hall (1976) correlated the eugeosynclinal Upper Hartland formation sequence with the miogeosynclinal Manhattan Schist, Member A. The protolith for the Harrison gneiss was probably mafic and felsic volcanic units that attained their present configuration through repeated thermotectonic modification (Hall, 1976).

9.2.5.4 Petrography

Mineralogy and Textural Relations. Harrison gneiss is dark gray to greenish; its major minerals are andesine (commonly as augen), quartz (generally subordinate), hornblende and biotite in variable amounts, and potassium feldspar. Minor minerals include garnet, augite, sphene, epidote, chlorite, and microcline. The Harrison gneiss has locally developed augen structures. The lighter layers are marked by feldspathic schists 5-15 cm (2-6 in.) thick. Dark-gray gneiss layers up to 60 cm (2 ft) thick are locally well developed. The Harrison gneiss is massive in some places but locally contains feldspar megacrysts as long as 5 cm (2 in.) (Hall, 1976).

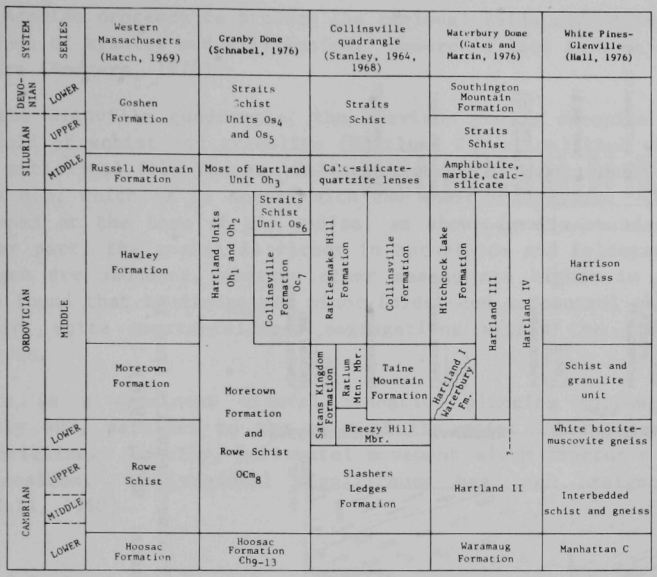

Fig. 9.6 Proposed Correlation of Stratified Units in Western Massachusetts, Western Connecticut, and Southeastern New York (Source: Adapted from Stanley and Hatch, 1976)

Alteration. Development of epidote, chlorite, and sericite in the gneiss may be related to alteration of its principal minerals.

9.2.5.5 Secondary Intrusions

In the Glenville area, pegmatites outline the pattern of folds (Hall, 1968) but have not been folded themselves. The pegmatites do not display any definite secondary foliation.

9.2.5.6 Country Rock

The Harrison gneiss has both intrusive and structural contacts with the Cambrian-Ordovician Hartland formation (see Fig. 9.7).

9.2.5.7 Rock Structure

Although the Taconic orogeny certainly modified the eugeosynclinal rocks in western Connecticut, strong deformation and metamorphism occurred

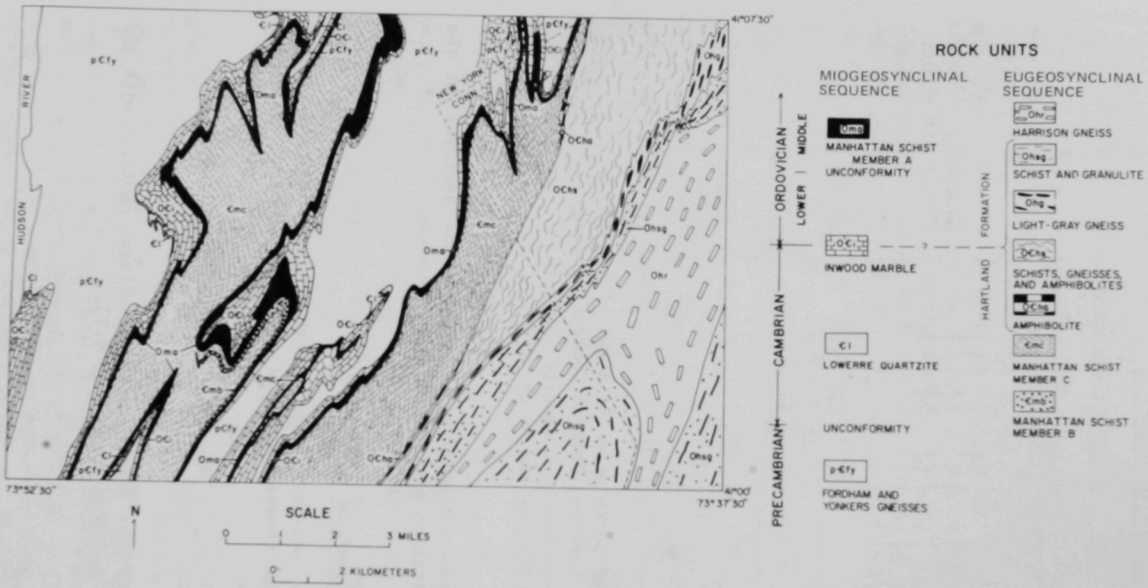

Fig. 9.7 Geologic Map of the White Plains-Glenville Area in Connecticut and New York (Source: Hall, 1976)

during the Acadian orogency to produce the regional folds and nappes and high-grade terranes of kyanite and sillimanite. Cameron's line is a major regional fault feature (Rodgers, 1970).

In the Glenville quadrangle, the Harrison gneiss occupies an earlier isoclinal fold of schist and granulite (Hartland formation) that was refolded by a later anticlinal fold phase. Foliation strikes almost east-west and has a northerly dip, which is in accord with the anticlinal crest. Foliation is well developed at the base of the gneiss, as shown by its abundant biotite. In the upper part, the gneiss is richer in hornblende and feldspar. In some layers, augen are abundant, whereas other masses are higher in microcline. Hall (1968) found that boudinage and minor folds seem to control the structure of irregular, white quartz-feldspar segregations within the gneiss of the Glenville area.

There is a prominent mineral lineation plunging northwesterly, at approximately 40°, parallel to the axis of the major later-phase (probably Acadian) anticline. Locally, horizontal movement along fractures is evident from slickensides. No regional significance has been assigned to this movement (Hall, 1968).

9.2.5.8 Geophysics

On a regional Bouguer gravity map, Hildreth (1979) showed contours of +30 mgal to +40 mgal in the area.

9.2.6 Monson (Haddam) Gneiss (Plate VIII, Pluton 11)

9.2.6.1 Size, Shape, and Location

The Monson (Haddam) gneiss crops out along a long, narrow, north-south anticlinal belt that extends about 85 km (53 mi) down from Massachusetts almost to Long Island Sound. This belt terminates in a large, roughly triangular body called the Haddam gneiss, which is considered to be part of the Monson gneiss (Pease, 1982). This triangular body has a base, parallel to Long Island Sound, of about 20 km (12 mi) and a height to the north of about 24 km (15 mi). The estimated area within which the Monson and Haddam gneisses crop out in Connecticut is about 317 km^2 (122 mi^2).

9.2.6.2 Relative and Absolute Age

On the basis of Rb-Sr whole-rock age determinations, Brookins and Hurley (1965) reported an age of 472 ± 15 m.y. for the Monson gneiss. The contact between the Monson gneiss and the overlying Middletown formation is gradational. The Middletown formation has been radiometrically dated at 440 ± 15 m.y. by whole-rock Rb-Sr methods, which agrees with observed field

relationships (Brookins and Hurley, 1965). Leo et al. (in press) assigned a U-Pb age of 428-450 m.y. (also see Sec. 7.2.6).

9.2.6.3 Origin

The Monson (Haddam) gneiss originated during Early to Middle Ordovician time as a sequence of volcanics and intrusives, including andesites, quartz keratophyres, and basalts, as well as intrusive rocks of similar composition. This sequence was then covered by sulfidic shales and siltstones of the Collins Hill formation. Deformation during the Taconic orogeny produced the structure that is now referred to as the Monson anticline (Lundgren, 1964). The area was also affected by later orogenics. Probable displacement of some protoliths by metasomatism or anatexis is mentioned by Pease (1982).

9.2.6.4 Petrography

Mineralogy and Textural Relations. The Monson gneiss is generally a medium-grained, light- to dark-gray, biotitic and hornblendic quartz-plagioclase gneiss with amphibolite layers. It is a one-feldspar (plagioclase) gneiss, with microcline, if present, occurring in minor quantities. Quartz and oligoclase or andesine are the major constituents, ranging from 25% to 40% and 40% to 55%, respectively (Lundgren, 1963; Lundgren and Thurrell, 1973). Biotite is omnipresent in various quantities, and hornblende is common. Accessory minerals include magnetite-ilmenite, garnet, and anthophyllite.

The Haddam gneiss is variable in fabric and composition. It is essentially a single-feldspar (oligoclase) gneiss with minor biotite, garnet, and rare microcline (Rodgers et al., 1959; Lundgren, 1964). It is foliated to massive and granitoid in the core area near Killingworth. It contains some hornblende gneiss and amphibolite units.

Alteration. Local hydrothermal alteration associated with pegmatite dikes was reported by Stugard (1958). Also, local mylonitization is present in fault zones.

9.2.6.5 Secondary Intrusions

The Monson gneiss is cut by pegmatite dikes believed to be approximately 260 m.y. old (Stugard, 1958). Some dikes are white granitic pegmatites, with plagioclase being the most abundant feldspar. Pink pegmatites are more common and tend to be rich in pink microcline. Quartz is an essential but interstitial mineral, and muscovite is an abundant accessory mineral. Garnet, biotite, lepidolite, spodumene, magnetite, and tourmaline may also be present in smaller quantities. A northeast-trending Triassic diabase dike cuts the northern part of the Haddam unit.

9.2.6.6 Country Rocks

In northern Connecticut, the Monson gneiss is bordered on the east by the Hamilton Reservoir formation, the Southbridge formation (Ammonoosuc volcanics), and younger rocks of the Great Hill syncline (Pease, 1975). In its southern area of exposure, on the flanks of the Killingworth dome, the Haddam gneiss is bordered on the east by the Tatnic Hill formation and on the west by the Middletown formation (Lundgren, 1963, 1964, and 1970).

The upper schist member of the Hamilton Reservoir formation is in contact with the Monson gneiss. This member consists mainly of a rusty-weathering, dark- to light-gray quartzo-feldspathic gneiss and brownish-gray biotite-sillimanite schist. Garnet, sulfide minerals, and graphite are characteristic.

The Southbridge formation is poorly exposed adjacent to the Monson gneiss. It consists chiefly of well-layered, medium- to dark-gray, fine- to medium-grained quartz-plagioclase-biotite schist and gneiss. Calc-silicate minerals and layers rich in calc-silicate minerals are characteristic.

The Tatnic Hill formation, mapped as the Brimfield schist by Snyder (1970), is mostly rusty-weathering, graphitic sillimanite schist.

Where it is in contact with the Monson (Haddam) gneiss, the Middletown gneiss is characterized by a variety of interlayered amphibolites and orthoamphibolite gneisses. The varieties include hornblendite, biotite gneiss and schist, feldspathic gneiss and granulite, and garnet granulite (Lundgren and Thurrell, 1973).

9.2.6.7 Rock Structure

Late-stage, small-scale faulting is noted by Lundgren (1963). This late stage represents essentially down-dip movement, in which the Rope Ferry gneiss was displaced northward relative to the overlying Putnam, Canterbury, and Hebron formations. The Monson gneiss is also jointed, with the most prominent joints corresponding to the major regional joints affecting most of the pre-Triassic rocks of the area. One set of joints is rather steep and trends north-south. There is much evidence, such as slickensides, of movement along these planes. Another, less prominent set of joints has a nearly vertical attitude but an east-west strike.

All of the Monson and Haddam gneisses are foliated to some extent, with the degree of foliation ranging from almost indistinguishable to well developed. Boundinage structure is well developed in areas where amphibolite layers are present within the gneiss. Lundgren (1964) described some boudins seen in the Essex quadrangle. It is evident that the amphibolite layers were thinned by plastic flow, which eventually led to separation of individual boundins from one another. This process also explains the blocklike masses of

amphibolite that appear to float within the dominant gneiss in the Chilton quadrangle. Numerous small faults and joints occur throughout the Haddam body (Lundgren and Thurrell, 1973; Pease, 1982).

The Monson gneiss occurs at the core of an elongated anticlinal structure, a major recumbent anticline that extends south through Connecticut from Massachusetts (Lundgren, 1964). Although the axial plane of the fold dips eastward in the Essex quadrangle, the attitude of the axial plane varies significantly from one location to another.

The Monson (Haddam) gneiss also forms the core of the Killingworth dome. Throughout the dome there are numerous, smaller-scale isoclinal folds with amplitudes of 2 m (7 ft) or more (Lundgren and Thurrell, 1973). The axial surfaces of the folds have a common orientation with the foliation in the dome. The southeastern part of the dome consists of two antiforms with north-south axial surfaces.

9.2.6.8 Geophysics

Aeromagnetic maps of all quadrangles in Connecticut have been published by USGS. The following specific maps may be of significance:

1. Philbin, P.W., and C.W. Smith, Aeromagnetic Map of the Rockville Quadrangle, GP-587 (1966).

2. Author unknown, Aeromagnetic Map of the Marlborough Quadrangle and Part of the Rockville Quadrangle, GP-849 (1973).

3. Author unknown, Aeromagnetic Map of the Moodus Quadrangle, GP-858 (1973).

4. Author unknown, Aeromagnetic Map of the Deep River Quadrangle, GP-868 (1973).

5. Author unknown, Aeromagnetic Map of the Essex Quadrangle, GP-877 (1973).

No interpretations are provided with these maps. Hildreth (1979) gave Bouguer-anomaly contour values of -10 mgal to -20 mgal for the area.

9.2.7 Norwalk Area Gneisses (Plate VIII, Pluton 2)

9.2.7.1 Size, Shape, and Location

The gneisses of the Norwalk area form an irregular pluton cropping out over an area of about 201 km^2 (78 mi^2) in southwestern Connecticut. The gneisses are primarily in the Norwalk North and Norwalk South 7.5-min quadrangles. Figure 9.8 shows the lithology of these two areas.

9.2.7.2 Relative and Absolute Age

Hall (1971) suggested that the Norwalk area gneisses correlate with the Cambrian and Ordovician sequence in Vermont. The mafic gneiss of Norwalk resembles the Harrison gneiss. Kroll (1977) reported the following stratigraphic order of the major units (see Fig. 9.8) in the Norwalk area:

Youngest	mafic gneiss unit III
↑	mafic gneiss unit II
	mafic gneiss unit I
	interlayered schist and granulite
	massive felsic gneiss
	mixed felsic gneiss
Oldest	biotite-plagioclase-quartz gneiss

This sequence is equivalent to that described by Rodgers (1982).

Clark and Kulp (1968) obtained K-Ar ages of 255 m.y. and 290 m.y. on mica in these gneisses, suggesting that they were subjected to a thermal event after their original formation.

9.2.7.3 Origin

Crowley (1968) and Kroll (1977) suggested a metavolcanic origin for the biotite-plagioclase-quartz gneiss, on the basis of its chemical and lithological correlation with the Collinsville formation. Kroll (1977) further proposed that the mixed felsic gneisses were derived from the intermingling of arkosic or felsic volcanic rocks with shaly, carbonate, and basaltic rocks. This would explain the gradational character among the granitic, metasedimentary, and metaigneous parts of the unit. The microcline-quartz-plagioclase gneiss is of granitic composition, the schist bodies are aluminous metasedimentary rocks, and the amphibolites and porphyroblastic biotite schists are metaigneous rocks.

Interlayered schist and granulite are a metamorphosed unit of interlayered sands or silts and shales. Impure carbonate rocks are represented by calc-silicates. Mafic gneisses represent metavolcanic rocks. The

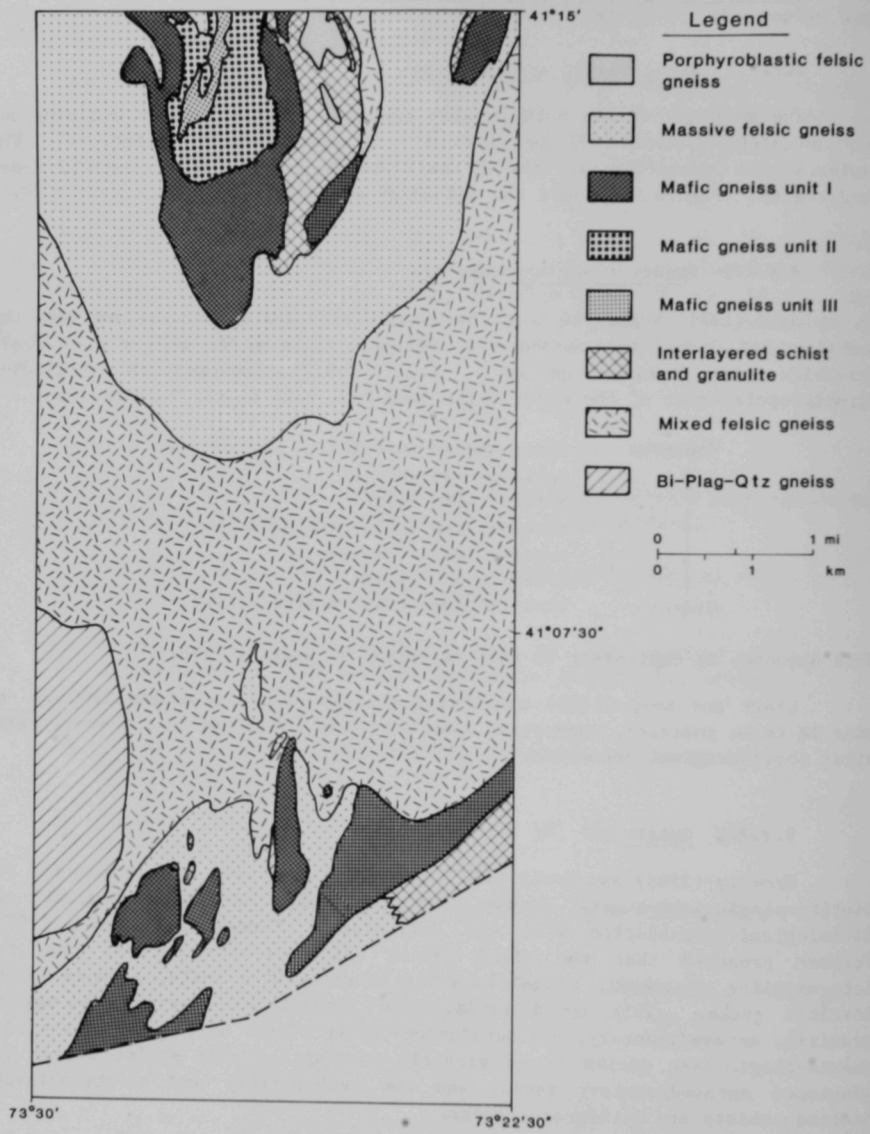

Fig. 9.8 Lithologic Map for the Norwalk North and Norwalk South Quadrangles (Source: Adapted from Kroll, 1977)

porphyroblastic microcline-quartz-plagioclase gneiss is granitic in origin and was intruded into the complex before or during metamorphism and deformation. The massive felsic gneisses probably represent remobilized felsic volcanic-sedimentary rocks, or silicic intrusions.

9.2.7.4 Petrography

Mineralogy and Textural Relations. Lithologic heterogeneity is very prominent in the rocks of the Norwalk area (see Fig. 9.8).

Biotite-Plagioclase-Quartz Gneiss. This unit is gray to buff, medium to coarse grained, and slightly to moderately foliated. Laminae of biotite are found in a granular matrix. The modal composition is quartz, 43-57%; plagioclase, 22-33%; biotite, 7-17%; and microcline, 2-14%; with minor amounts of magnetite, muscovite, apatite, garnet, zircon, and chlorite (Kroll, 1977).

Mixed Felsic Gneiss. This unit is composed of interlayered gneiss, amphibolite, and schist. The microcline-quartz-plagioclase gneiss, locally pegmatitic, is gray-white to buff-gray to yellowish-gray and medium to very coarse grained. The modal composition is plagioclase, 22-46%; quartz, 21-36%; microcline, 14-34%; muscovite, 1-13%; and biotite, 2-7%; with minor amounts of opaque minerals, chlorite, apatite, garnet, and zircon (Kroll, 1977). Fragmented xenoblastic and idioblastic grains of garnet are common, with chlorite along the fractures and rims. Locally, porphyroblasts of microcline a few centimeters long (an inch or so) are found in the gneiss.

The amphibolites are shiny black, medium to coarse grained, and massive; they contain quartz lenses and have thin, subparallel felsic stringers. Locally, garnet or biotite or both are conspicuous porphyroblasts.

The porphyroblastic biotite gneiss is dark, medium to coarse grained, and well foliated. Xenoblastic microcline porphyroblasts 1-2 cm (0.4-0.8 in.) long are abundant. The chief minerals are biotite, plagioclase, microcline, and quartz. Locally, pyroxene is a primary mineral. Apatite, zircon, and opaque minerals are accessory minerals.

The calc-silicate gneiss is fine to medium grained and has rusty micaceous layers separating light-green and dark-green bands 1-8 cm (0.4-3.2 in.) thick. The darker layer is richer in amphiboles; the biotite-rich layers are really biotite-quartz-plagioclase schists. Some layers are pink and rich in garnet. Locally, quartz-feldspar pegmatites and quartz pods are found.

Interlayered Schist and Granulite. This unit is a medium-grained, slightly foliated, mica-poor, dark-gray quartz-plagioclase or light-pink granulite interlayered with well-foliated, medium- to coarse-grained, dark-brown gedrite schist, containing gedrite blades up to 4 cm (1.6 in.) long. The granulite layers are usually thicker than the schist layers, ranging from 1-2 cm (0.4-0.8 in.) to 1 m (3 ft). Locally, some of the rocks are graphitic

and contain tourmaline. Quartz pods are common. On Calf Pasture Island, in the Norwalk South quadrangle, quartz-kyanite pegmatite pods occur. The modal composition of the schist is plagioclase, 0.5-53%; quartz, 25-62%; microcline, 0-28%; muscovite, 2-36%; and biotite, 0.2-22%; with minor garnet, sillimanite, opaque minerals, apatite, zircon, sphene, and chlorite (Kroll, 1977).

Mafic Gneisses. There are three units of porphyroblastic and nonporphyroblastic mafic gneisses. Unit I is a mafic gneiss that is up to 300 m (1000 ft) thick, averaging 180 m (600 ft). The gneiss is equigranular, medium to coarse grained, layered, and cross-laminated. Locally, plagioclase and hornblende porphyroblasts are found. There is interlayering of a 10-cm (4-in.) thick, granular, medium-grained rock, and a thicker, more foliated, coarse-grained rock. The modal composition of the mafic gneiss is plagioclase, 3-47%; quartz, 0-30%; microcline, 0-48%; biotite, 0-33%; hornblende, 0-40%; clinopyroxene, 0-9%; and sphene, 0.3-4%. Apatite, zircon, allanite, opaque minerals, chlorite, and calcite are present in minor amounts (Kroll, 1977).

Mafic granular pods, ultramafic rocks, garnet-gedrite schist, and a thin layer of calc-silicate rock are also found in unit I. The mafic granular pods are pillow structures; they are black, spherical to ellipsoidal, medium grained, nonfoliated, granular, and range from 4 cm (1.6 in.) to 1 m (3 ft) long. The modal composition is plagioclase, 16.8%; microcline, 6.1%; hornblende, 7.1%; pyroxene, 55.3%; and sphene, 2.8%. Apatite, zircon, opaque minerals, calcite, and clinozoisite are present in minor amounts (Kroll, 1977). The ultramafic rocks include chlorite-talc-enstatite and talc-serpentine-tremolite schists. The garnet-gedrite rocks are coarse grained, massive, and slightly to well foliated, and contain cordierite locally.

The felsic interlayers are dark, 0.5-4 cm (0.2-1.6 in.) thick, granular, and slightly foliated. Their modal composition is plagioclase, 17-55%; quartz, 12-36%; microcline, 11-69%; and biotite, 0-20%. Hornblende, sphene, apatite, zircon, allanite, opaque minerals, chlorite, and garnet are present in minor amounts.

Unit II of the mafic gneisses is 60-300 m (180-1000 ft) thick and massive, with some layering. Gray, subhedral microcline porphyroblasts form up to 10% of the unit. Mafic, foliated pods and layers are also found. The modal composition of unit II is plagioclase, 19-41%; quartz, 3-29%; microcline, 4-25%; biotite, 9-29%; hornblende, 0-42%; clinopyroxene, 0-6%; sphene, 0.3-5%; apatite, 0.1-1%; zircon, 0-0.2%; allanite, 0-0.4%; opaque minerals, trace to 0.7%; and chlorite, 0-0.6% (Kroll, 1977).

Unit III is similar to unit I, with a modal composition of plagioclase, 43-48%; quartz, 11-23%; microcline, 0-2%; biotite, 21-28%; hornblende, 4-15%; sphene, 1-2%; apatite, 0.6-0.9%; zircon, trace; and opaque minerals, 0.1-0.5% (Kroll, 1977).

Alteration. Crowley (1968) reported that there is an increase in regional metamorphic grade from central to western Connecticut, ranging from chlorite grade near New Haven to sillimanite grade near Bridgeport. Some retrograde metamorphism may have occurred (Kroll, 1977).

9.2.7.5 Secondary Intrusions

Pegmatitic intrusions form dikes and sills throughout the crystalline rock body. Small hornblendite and diabase dikes are also reported. Dark, fine-grained diabase cuts across the quartz-plagioclase-microcline gneiss in the mixed felsic gneiss.

9.2.7.6 Country Rock

To the east of the Norwalk area gneisses lie large refolded and recumbent folds and nappes of metamorphosed eugeosynclinal rocks, including the granitic gneisses described in Sec. 9.2.4. To the west are refolded, steeply inclined isoclinal folds of metamorphosed miogeosynclinal rocks, including the Harrison gneiss described in Sec. 9.2.5 (Kroll, 1977; Rodgers, 1982). Hornblendites, pegmatites, and diabase occur in the Norwalk area gneisses.

9.2.7.7 Rock Structure

Kroll (1977) suggested that the major folding and metamorphism are largely related to Acadian thermotectonic events of 360 m.y. ago. A second event, about 260 m.y. ago, was primarily thermal in nature. Structurally, the rocks of the Norwalk area are highly anisotropic.

Folding in the Norwalk North area is complex and multiple. Kroll (1977) studied the structural fabric of the area. In the Norwalk North area, the Wilton fold incorporates the schist, granulite, and mafic gneiss units. In the Wilton fold, the S_2 foliation is parallel to the bedding surfaces but becomes an axial plane foliation in the mafic gneiss; lineation is well developed in the schist and granulite unit; and the minor fold axes F_1 postdate the S_2 foliation (Kroll, 1977). In the Norwalk South area, a major fold is recognized. The lineation and minor fold axes plunge southward and southwestward, in contrast to the north- and west-plunging orientation in the northern section.

Fractures are common in all lithologic units and frequently show slickensides on steeply dipping, northwest-trending fracture surfaces (see Fig. 9.9). Such slickensides are generally indicative of shear fractures caused by stress systems with near-horizontal maximum compression and near-vertical intermediate compression. Kink bands are the result of north-south compression (Dietrich, 1968).

In the massive felsic gneiss, foliation is not well developed, except in local micaceous rocks. Lineation is absent or not well developed. The gneiss is fractured, with slickensides coated with fine-grained tourmaline.

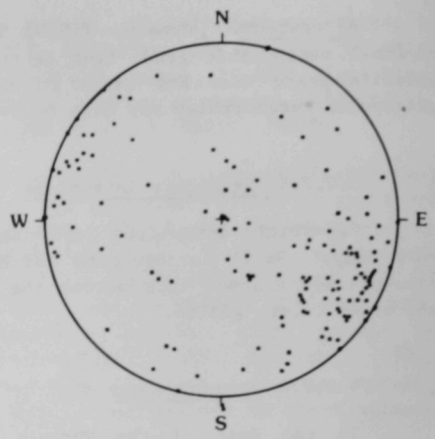

Fig. 9.9 Orientation Diagram of 133 Sets of Slickensides in the Norwalk North and Norwalk South Quadrangles (Source: Kroll, 1977)

9.2.7.8 Geophysics

The contact between the biotite-plagioclase-quartz gneiss and the amphibolite-bearing unit of the mixed felsic gneisses in the Norwalk South quadrangle trends parallel to the magnetic-intensity lines on the aeromagnetic map (U.S. Geol. Survey, 1971). The magnetic intensity of the biotite-plagioclase-quartz gneiss is greater than that of the other rock units, because of the magnetite content. Hildreth (1979) showed -10-mgal to -20-mgal gravity-anomaly values on a regional Bouguer anomaly map for the area.

9.3 ACADIAN AND POST-ACADIAN

9.3.1 Canterbury Gneiss (Plate VIII, Pluton 14)

9.3.1.1 Size, Shape, and Location

The Canterbury gneiss is a fairly homogeneous, medium-grained, quartzo-feldspathic gneiss that crops out in eastern Connecticut. The pluton forms a belt that trends north-northeast and extends nearly across the state (see Fig. 9.10). The approximate total length of the belt of six elongated, roughly parallel strips is 78 km (48 mi). The approximate area occupied by the body is 210 km^2 (81 mi^2).

9.3.1.2 Relative and Absolute Age

The youngest rocks that the gneiss cuts are of Ordovician age. Whole-rock Rb-Sr analysis suggests that the age of the Canterbury gneiss is

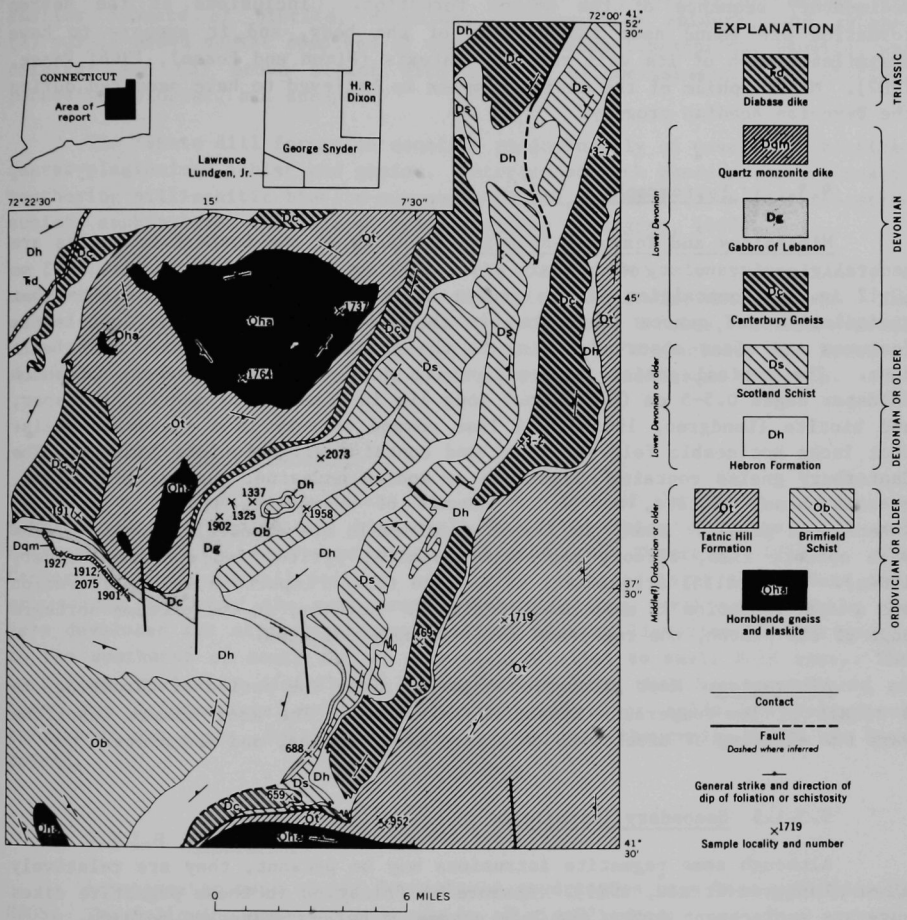

Fig. 9.10 Geologic Map of Canterbury Gneiss Area
(Source: Zartman et al., 1965)

approximately 329 ± 9 m.y. (Pease, 1982); it is possible, however, that this reflects a later thermal event that reset the ratios of rubidium and strontium in the measured rocks.

9.3.1.3 Origin

The Canterbury gneiss is a metamorphosed intrusive body. The antecedent of the gneiss was probably emplaced as an intrusive sill into the

sedimentary sequence of the Hebron formation. Inclusions of the Hebron formation are found near the margins of the body, and it appears to have assimilated much of its wall rock by anatexis (Dixon and Pessel, 1966; Pease, 1982). Metamorphism of the entire complex is believed to have occurred during the Devonian Acadian orogeny.

9.3.1.4 Petrography

Mineralogy and Textural Relations. Rocks of the Canterbury gneiss are generally equigranular, with grain sizes ranging from 1 mm (0.04 in.) to 3 mm (0.12 in.). Composition of the gneiss varies from granite to tonalite, but granodiorite and quartz monzonite are the most common facies. Cataclastic features are often observed, owing to the proximity of the Honey Hill fault zone. The typical gneiss is a medium- to dark-gray augen gneiss, in which feldspar augen 0.5-5 cm (0.2-2 in.) long lie in a matrix of quartz, feldspar, and biotite (Lundgren, 1966b). A less common variety is a muscovite gneiss that lacks noticeable feldspar augen and contains slightly less biotite. The Canterbury gneiss contains oligoclase or sodium andesine, quartz, microcline, greenish-black biotite, and minor amounts of muscovite, iron-rich epidote, sphene, and garnet. Among the accessory minerals are euhedral allanite grains with epidote rims, zircon, magnetite-ilmenite, pyrite, and apatite (Snyder, 1964a). Subparallel arrangement of biotite flakes imparts a strong lineation and a weak foliation in most hand specimens. However, in the extreme northern edge of the pluton, the rocks are more strongly foliated.

Alteration. Most alteration occurred after the period of thrusting, as a result of low-temperaure hydration reactions. The most common reactions were the altering of biotite to chlorite, minor sphene, and muscovite.

9.3.1.5 Secondary Intrusions

Although some pegmatite intrusions may be present, they are relatively minor (Lundgren et al., 1971). Absence of foliation in these pegmatite dikes suggests emplacement during the late stages of metamorphism and folding.

9.3.1.6 Country Rocks

The Canterbury gneiss is bounded mainly by the Hebron formation and the Tatnic Hill formation, but it also comes in contact with the Scotland schist and the Lebanon gabbro (Dixon and Shaw, 1965).

The Hebron formation, where it is adjacent to the Canterbury gneiss, consists of various granular and micaceous schists and gneisses that represent metamorphosed calcic sedimentary rocks (Snyder, 1964a). Three different rock types are considered part of the Hebron formation: (1) layered, green-gray,

calc-silicate rock composed of quartz, andesine-labradorite, hornblende, and smaller amounts of biotite, diopside, aluminum-rich epidote, and sphene; (2) purple-brown schist composed of quartz, andesine-labradorite, biotite, and calcite; and (3) brown schist with quartz, oligoclase-andesine, biotite, potassium feldspar, and sphene.

The Tatnic Hill formation consists predominantly of muscovitic biotite-quartz-plagioclase schist and gneiss. Interbedded with these rocks are rusty-weathering sillimanitic biotite-muscovite schist, garnetiferous quartz-biotite schist, amphibolite, and calc-silicate gneisses (Lundgren, 1962).

The Scotland schist of Dixon and Shaw (1965) is predominantly a muscovite schist with well-developed foliation and indistinct bedding. The schist is composed of fine-grained quartz and biotite, coarse-grained muscovite, and minor amounts of garnet, staurolite, oligoclase, and kyanite. Accessory minerals are tourmaline, apatite, zircon, and opaque minerals.

9.3.1.7 Rock Structure

Two types of faults affect the outcrop area of the pluton. These are low-angle thrust faults and high-angle faults striking north to south and north-northwest to south-southeast (Lundgren and Thurrell, 1973). The Canterbury gneiss is well jointed, containing two major joint sets. Both sets are nearly vertical; the most prominent set strikes north-south, while the less developed set strikes east-west. Most mineral lineation plunges gently to the southwest or south and is generally parallel to small fold axes. The Canterbury gneiss is distributed throughout an area that has experienced at least two deformational events, and major and minor folds are recognized throughout the area. Outcrop patterns suggest that the gneiss itelf has been folded.

9.3.1.8 Geophysics

Aeromagnetic maps of all Connecticut quadrangles have been published by USGS. The following specific maps may be of significance:

1. Boynton, G.R., and C.W. Smith, Aeromagnetic Map of the Norwich Quadrangle, GP-543 (1966).

2. Boynton, G.R., and C.W. Smith, Aeromagnetic Map of the Scotland Quadrangle, GP-540 (1966).

3. Philbin, P.W., and C.W. Smith, Aeromagnetic Map of the Hampton Quadrangle and Part of the Scotland Quadrangle, GP-593 (1966).

4. Philbin, P.W., and C.W. Smith, Aeromagnetic Map of the Danielson Quadrangle and Part of the Plainfield Quadrangle, GP-592 (1966)

5. Author unknown, Aeromagnetic Map of the Eastford Quadrangle and Part of the Hampton Quadrangle, GP-651 (1969).

6. Author unknown, Aeromagnetic Map of the Willimantic Quadrangle and Parts of the Spring Hill and Scotland Quadrangles, GP-851 (1973).

7. Author unknown, Aeomagnetic Map of the Fitchville Quadrangle and Part of the Norwich Quadrangle, GP-860 (1973).

No interpretations are provided with these maps.

9.3.2 Nonewaug Granite (Plate VIII, Pluton 9)

9.3.2.1 Size, Shape, and Location

The Nonewaug granite forms four discontinuous areas of outcroppings that define a lens-shaped body trending in a southwesterly direction, in the Woodbury, Litchfield, Thomaston, and Waterbury quadrangles. The two largest of the four areas are each about 12 km (7 mi) long and 3-5 km (2-3 mi) wide; the two smaller areas are each about 5 km (3 mi) long and 2 km (1.2 mi) wide. Outcrops occur over a total area of about 78 km^2 (30 mi^2) (Gates, 1954).

9.3.2.2 Relative and Absolute Age

The Nonewaug granite intrudes into the Hartland formation of Ordovician age and is intruded into by a basalt of Late or post-Triassic age. Whole-rock Rb-Sr dating indicates an age of 382 ± 64 m.y. (Besancon, 1970).

9.3.2.3 Origin

The Nonewaug granite probably intruded into the Hartland formation as a magma. The body is also associated with dikes that cut into the Hartland formation.

9.3.2.4 Petrography

Mineralogy and Textural Relations. The Nonewaug granite is a banded igneous rock with layers ranging in thickness from a centimeter or less (a fraction of an inch) to a few meters (several feet). Its texture ranges from very fine grained granite to coarse pegmatite. Boundaries between layers are gradational. The most common rock is a patchy mixture of granite and pegmatite in the central part of the body. There is no regular distribution of the fine- or coarse-grained types. The general rock fabric is of an intrusive breccia, with fine-grained granite forming clasts and with more pegmatitic granite as the host. A broad zone with plumose muscovite and graphic granite crosses the central part of the granite body and trends N. 60° E. The graphic granite grains range in size from less than 2.5 cm (1 in.) to 0.6 m (2 ft).

The chief minerals of the Nonewaug granite are microcline, sodium plagioclase, quartz, muscovite, and biotite. Apatite and garnet occur as accessory minerals. Microcline and albite normally account for 65% of the rock, with quartz 30% and mica about 5%. Microcline may predominate in one sample and plagioclase in another. All proportions between a nearly pure potassium feldspar granite and a nearly pure sodium plagioclase granite occur. In general, plagioclase is more abundant, and the Nonewaug is termed a sodic granite. Muscovite is the predominant mica (Gates, 1954).

Alteration. Secondary alteration may be minimal, as there is no mention of it in the literature.

9.3.2.5 Secondary Intrusions

No secondary intrusions have been mapped within the Nonewaug granite.

9.3.2.6 Country Rocks

The Nonewaug granite intrudes into the Hartland formation, which lies on the eastern flank of the Green Mountain anticlinorium and extends from Vermont through Massachusetts and Connecticut, to Long Island Sound. The Hartland formation is considered to be a Cambrian-Ordovician sequence consisting of various metasediments, interbedded mica quartzites, mica quartz schists, and granitic gneisses. Granite and pegmatite dikes and sills are prominent features in the Hartland formation near its contacts, particularly along its southern border.

9.3.2.7 Rock Structure

The Nonewaug granite body is transected by a north-south-trending Mesozoic fault. The textural layering of the Nonewaug granite indicates pervasive flow banding. The Nonewaug granite is not folded.

9.3.2.8 Geophysics

On Hildreth's (1979) map, the pluton lies in an area characterized by low-value positive gravity contours.

9.4 ALLEGHENIAN AND POST-ALLEGHENIAN

9.4.1 Narragansett Pier Granite (Plate VIII, Pluton 16)

See Sec. 8.4.1 and Plate VII, pluton 1. Only 1 km^2 (0.4 mi^2) of the Narragansett Pier granite occurs in Connecticut; the rest of it lies in Rhode Island. Therefore, it is discussed with the plutons of that state.

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10 CRYSTALLINE ROCKS OF SOUTHEASTERN NEW YORK, NORTHERN NEW JERSEY, AND SOUTHEASTERN PENNSYLVANIA

10.1 GRENVILLE AND AVALONIAN

10.1.1 Baltimore Gneiss Complex (Plate IX, Pluton 2)

10.1.1.1 Size, Shape, and Location

The Baltimore gneiss complex extends from just southeast of Trenton, N.J., southwestward to the Delaware border. The complex is elongated parallel to a general strike of N. 68° E., stretching in this direction more than 92 km (57 mi). Its maximum width, near the southern end, is 14 km (9 mi); its minimum width, near the northern end, is 0.5 km (0.3 mi). Baltimore gneiss crops out in many 7.5-min quadrangles, including the Trenton West, Langhorne, Hatboro, Ambler, Frankford, Germantown, Norristown, Valley Forge, Malvern, Philadelphia, Lansdowne, Media, West Chester, Unionsville, Coatesville, Marcus Hook, Wilmington North, Kennett Square, and West Grove quadrangles. Outcrops of this body are found throughout an area of about 477 km^2 (185 mi^2).

10.1.1.2 Relative and Absolute Age

On the basis of U-Pb analyses on zircon, Lapham and Root (1971) assigned an age of about 1100 m.y. to the Baltimore gneiss. However, Wagner and Crawford (1975) proposed that this is probably the date of an event of granulite-facies metamorphism, and that the rocks may actually be older than 1100 m.y.

10.1.1.3 Origin

The Baltimore gneiss may have resulted from metamorphism of both igneous and sedimentary Precambrian rocks (Wagner and Crawford, 1975). However, the rocks have been so severely altered that their origin is uncertain.

10.1.1.4 Petrography

Mineralogy and Textural Relations. Baltimore gneiss is synonymous with the term "basement complex" (Higgins, 1972). The gneiss comprises several different metamorphic assemblages. On the basis of studies within a 65-km^2 (25-mi^2) area of the Westchester prong section of the complex, Wagner and Crawford (1975) divided the complex into four main lithologies: (1) pyroxene

granulite, (2) quartzo-feldspathic granulite, (3) quartzose rock, and (4) amphibolite and migmatite. Within this study area, which is only about one-seventh of the total surface area of the Baltimore gneiss, Wagner and Crawford (1975) noted several major compositional differences between the western and eastern sections. These differences are probably related to differences in original rock type and metamorphic history. In places, the Baltimore gneisses have some similarities with the Reading Prong lithologies; this is particularly true in areas north and south of the Huntington and Cream Valley faults.

The pyroxene granulite (also referred to as gabbroic gneiss) composes about half the surface area of the complex. The granulite is a two-pyroxene plagioclase rock containing accessory magnetite and/or ilmenite, zircon, and apatite. The orthopyroxene is hypersthene, and the clinopyroxene is salite. Plagioclase is twinned and often antiperthitic. Some pyroxene granulite contains hornblende or biotite or both, apparently in equilibrium with the pyroxenes. Foliation is faint to nonexistent.

The composition of the quartzo-feldspathic granulite is extremely variable; in the past it has been classed with felsic or granite gneisses. The quartz grains contain garnet, presumably from the first of several metamorphic stages. Biotite is common, and apatite is present as an accessory mineral (Wagner and Crawford, 1975). The amphibolite-migmatite is a crystalloblastic rock composed mostly of hornblende and plagioclase, with small amounts of quartz and accessory minerals (Armstrong, 1941). Pyroxene is also present and may constitute as much as 5% of the rock.

Rocks of the quartzose facies are probably minor, underlying less than 5% of the surface area in the region studied. Most of these rocks are kyanite quartzites. Some samples contain as much as 40% garnet. Phlogopite occurs in some of the kyanite quartzites, accompanied by minor amounts of plagioclase, orthoclase, or very fine perthite. Rutile is often present, and graphite and zircon are accessory minerals. Hematite may be present, possibly the result of weathering of magnetite (Wagner and Crawford, 1975).

Alteration. Both Amenta (1974) and Wagner and Crawford (1975) agreed that the Baltimore gneiss underwent three stages of metamorphism. However, Amenta believed the area was also subjected to five discrete phases of deformation. It has not been determined whether the rocks now forming the Baltimore gneiss were originally igneous or sedimentary. The Grenville orogeny, which metamorphosed the original rocks beyond recognition (Wagner and Crawford, 1975), is probably the oldest event preserved in the region.

A second metamorphism associated with intense deformation probably occurred about 440 m.y. ago during the Taconic orogeny. Metamorphism under anhydrous conditions produced garnet coronas in granulite-facies rocks and diabase dikes. Metamorphism under hydrous conditions resulted in amphibolite-type facies. This metamorphic event was also responsible for altering the Wissahickon sediments to amphibolite-facies rocks (Wagner and Crawford,

1975). However, the effects of post-Taconic events on the Baltimore gneiss are difficult to differentiate from those of the Taconic events (Drake, 1982).

10.1.1.5 Secondary Intrusions

Three secondary intrusive events are recorded in the Baltimore gneiss. The first was intrusion of diabase dikes sometime between 1000 m.y. and 440 m.y. ago. It has been suggested that these dikes are Precambrian, but a Cambrian or Ordovician age is also possible (Wagner and Crawford, 1975). Several of these pre-Triassic dikes occur in the lower west center of the complex, with a lone dike in the upper east area near the Newark basin border. The dikes show well-preserved ophitic textures, sharp contacts, and chilled margins. These dikes were also partially recrystallized during post-Cambrian metamorphism. During the early Paleozoic, the gneiss was intruded by several pegmatite dikes, three at the southwestern terminus of the complex. All have similar thicknesses and are from 0.8 km (0.5 mi) to 5 km (3 mi) in length. Two more massive pegmatite bodies are located within the complex. One is a small elliptical formation in the southwestern region; the other is a V-shaped structure in the north-central portion (Gray et al., 1960) that is about 2.4 km (1.5 mi) long and 0.3 km (0.2 mi) wide at its widest point. Drake (1982) suggested that the pegmatites were probably generated within the gneiss during subsequent metamorphism.

The gneiss has also been intruded into by Triassic diabase of the York Haven quartz tholeiitic type. At least 10 of these dikes, 1.6-6.4 km (1-4 mi) long, intersect the lower third of the complex (Smith et al., 1975). All three types of dikes show a general strike of N. 45° E.

10.1.1.6 Country Rocks

For the most part, the gneiss is unconformably overlain by members of the Wissahickon formation, presumably along a tectonic contact. Along the southern and northwestern edges, contact is predominantly with an oligoclase-mica schist (assumed to be a facies of the Wissahickon) and also hornblende gneiss, some augen gneiss, and quartz-rich and feldspar-rich members that show varying degrees of granitization (Gray et al., 1960). At the western extremity, the gneiss is overlain, respectively, by Cockeysville marble, a white to light-bluish-gray marble, and by the Setters formation, a white feldspathic quartzite to gray mica gneiss and schist. All three of these layers are part of the provincial series called the Glenarm supergroup, which overlies the basement Precambrian rocks throughout the region.

Country-rock relationships to the east are more complex, where fairly extensive contacts with the Cambrian Chickies formation are recorded both north and south of the Baltimore gneiss. The Chickies formation is a light-gray, hard, massive, scolithus-bearing quartzite and quartz schist, with a conglomerate at its base and thin interbeds of dark slate at its upper margin.

In the extreme northeast, the Stockton formation defines the edge of the Newark-Gettysburg Triassic basin, which is bounded by faults on the southeast. It is a light-gray to buff, coarse-grained arkosic sandstone and conglomerate, with red and brown siliceous sandstone and red shale. This contact suggests covering of the Glenarm supergroup within the Newark basin.

10.1.1.7 Rock Structure

Foliation within the felsic rocks of the complex is defined primarily by biotite; steep joints have developed parallel to these foliation planes (Armstrong, 1941). Near Jenkintown, these joints strike about N. 60° E. The second most prominent joint system throughout the gneiss is nearly vertical and strikes north (Armstrong, 1941). However, cross jointing is not uncommon.

Fault lines define almost all of the southern edge and more than half of the northern edge of the complex. The Rosemont and Cream Valley fault zones that border the Baltimore gneiss were active as late as the Alleghenian orogeny. South of the Rosemont fault, the Alleghenian deformation is not evident (Amenta, 1974).

There are five separate generations of folds, resulting from five discrete deformational events. Within the banded hybrid rocks (gneisses), the fold axes plunge 10-30°.

10.1.1.8 Geophysics

Diment et al. (1980) compiled Bouguer gravity data for the northeastern United States. The gravity gradient represented by the Baltimore gneiss and related gneisses separates the metamorphic core of the Appalachian Mountains from the miogeosynclinal deposits, which are characterized by sharp gravity lows.

10.1.2 Honey Brook Complex (Plate IX, Pluton 1)

10.1.2.1 Size, Shape, and Location

The Honey Brook complex is located between the Susquehanna and Schuylkill rivers, directly west of Phoenixville in southeastern Pennsylvania. The Phoenixville, Pottstown, Elverson, Malvern, Downington, Wagontown, Honey Brook, New Holland, Coatesville, Parkesburg Gap, and Quarryville quadrangles are the 7.5-min quadrangles in which the complex crops out. The Honey Brook complex underlies an area of about 507 km^2 (195 ^{2} mi), which includes the Honey Brook upland and the Glenarm terrane, on opposite sides of the largely Paleozoic Chester Valley region (see Fig. 10.1). The anorthosite body, which underlies about 9% of the total surface area of the complex (Gray et al.,

1960), is often referred to as the "Honey Brook anorthosite." According to Gray et al. (1960), the remainder of the area is underlain by granite gneiss, monzonite, granodiorite, graphitic gneiss, and gabbroic gneiss. However, the geological understanding of the area has been substantially reappraised by Crawford (1979), Crawford and Crawford (1980), and Crawford and Hoersch (1983).

The southern edge of the complex is fairly straight, extending approximately 58 km (36 mi) and striking about N. 71° E. From its northeastern terminus to its center, the complex is fairly massive, with an average width of 16 km (10 mi). From the center westward, the complex bifurcates into two narrow prongs, each with an average width of 4 km (2.5 mi).

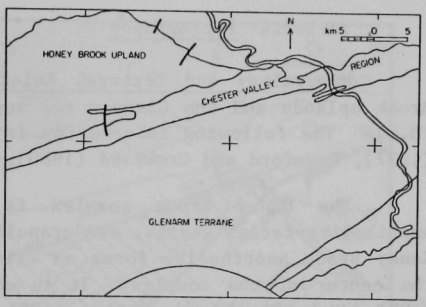


Fig. 10.1 Geographical Relationship between the Honey Brook Upland, Chester Valley Region, and Glenarm Terrane of the Southeastern Pennsylvania Piedmont (Source: Crawford and Crawford, 1980)

10.1.2.2 Relative and Absolute Age

Tilton et al. (1960) and Grauert et al. (1973) found U-Pb-zircon ages of 980-1100 m.y. for gneisses from the Honey Brook upland and the Glenarm terrane, ages that probably indicate the effect of the Grenville thermotectonic event. Crawford and Hoersch (1983) reported Ar^{40}-Ar^{39} ages of 850-880 m.y. on coexisting hornblende and biotite from the granulite gneisses, which suggests cooling and uplift following the Grenville event.

10.1.2.3 Origin

Mafic and felsic gneisses of amphibolite facies in the Honey Brook complex probably are of metavolcanic and metagraywacke origin (Crawford et al., 1971; Drake, 1982; and Crawford and Hoersch, 1983). The graphitic gneisses may be graphite-bearing phases of the felsic gneiss (Crawford and Hoersch, 1983), rather than being of metasedimentary origin as Gray et al. (1960) and Berg et al. (1980) suggested. The anorthosite body is of magmatic origin; its present status is the result of flow differentiation (Crawford et al., 1971; Raman and Hewing, 1979).

10.1.2.4 Petrography

Mineralogy and Textural Relations. The complex comprises the Honey Brook uplands and the Glenarm terrane of the Pennsylvania piedmont (see Fig. 10.1). The following information is based on Crawford et al. (1971), Demmon (1977), Crawford and Crawford (1980), and Crawford and Hoersch (1983).

The Honey Brook complex is principally composed of anorthosite, amphibolite-facies gneiss, and granulite-facies gneisses (see Fig. 10.2). The Honey Brook anorthosite forms an elliptical body of about 47 km^2 (18 mi^2) at the center of the complex. It is composed of an association of anorthosite (40%), leuco-hornblende gabbro (40%), and hornblende gabbro (20%), which are distinguished by differences in their hornblende content -- about 0-8% in the anorthosite and up to 55% in the gabbros. The texture is typically hypidiomorphic granular. Biotite, quartz, apatite, and magnetite are accessory minerals. Crawford et al. (1971) suggested that flow differentiation may have concentrated the plagioclase in the core, while forcing the gabbroic anorthosite liquid to the borders, where it crystallized last.

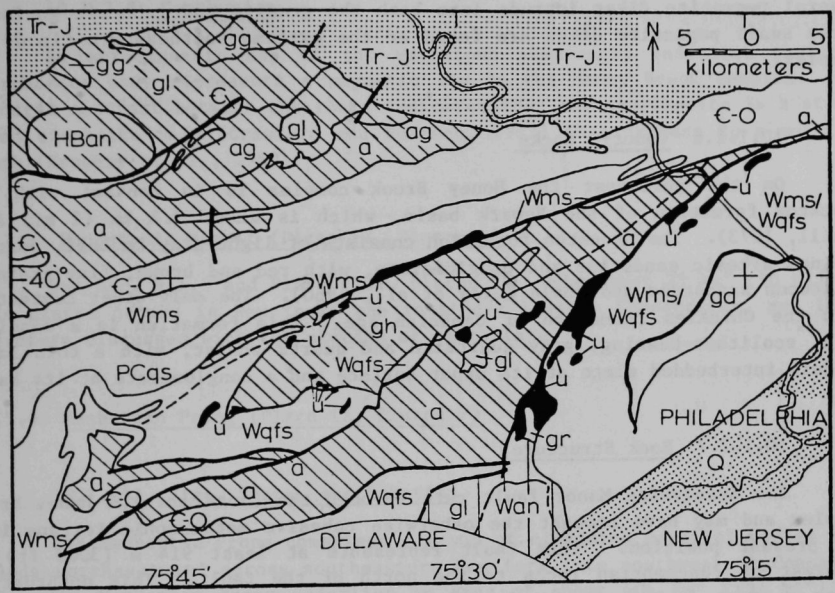
The amphibiolite-facies gneisses are predominantly massive gneisses of felsic and intermediate composition. The principal minerals are quartz, plagioclase, hornblende, biotite, and muscovite, in varying proportions. In contrast, the mafic gneisses, composed of hornblende and plagioclase, are well foliated but subordinate.

The graphite-bearing gneisses form important constituents of both the amphibolite- and granulite-facies rocks of the uplands. The granulite-facies gneisses are medium- to coarse-grained, weakly foliated, and somewhat layered felsic and mafic granulites (charnockites). The presence of mesoperthite, hypersthene, and augite, in addition to quartz and plagioclase, differentiates them from the amphibolite-facies gneisses. The upland gneisses are largely of felsic to intermediate composition, whereas the Glenarm terrane gneisses are more mafic and lack graphite.

Alteration. Three separate stages of metamorphism are identified with the youngest greenschist facies, which caused noticeable retrogression in the metamorphic grade (Demmon, 1977). Metamorphism caused the formation of uralite, sericite, clinozoisite, zoisite, garnet, and rare epidote and allanite (Crawford et al., 1971). Tectonic movement or later magmatic stress caused granulation. Plagioclase in the felsic gneisses is sericitized.

10.1.2.5 Secondary Intrusions

Precambrian diabase dikes intrude into the granite gneiss portion of the Honey Brook complex in four places at its southern tip. Also, a long diabase dike of uncertain age (probably either Precambrian or Triassic) intrudes into the north-central section of the body. All of these dikes strike approximately N. 20° E. At the point where the complex narrows,



Q	Quaternary alluvium
Tr-J	Triassic–Jurassic clastic sedimentary rocks intruded by diabase dykes and sills
€-O	Cambro-Ordovician carbonates and clastic sedimentary rocks
€	Cambrian quartzites
	?Lower Palaeozoic?
gl	Lima granite
u	Utramafic rocks
gd	Springfield granodiorite
PCqs	Peters Creek quartzitic schist
Wms	Wissahickon mica schist
Wqfs	Wissahickon quartzofeldspathic schist
Wan	Wilmington anorthosite
	?Precambrian and Precambrian—Basement gneiss
	amphibolite grade
a	undifferentiated gneiss
ag	graphitic gneiss
	granulite grade
HBan	Honey Brook anorthosite
gg	graphitic gneiss
gh	undifferentiated gneiss, high P
gl	undifferentiated gneiss, low P

Fig. 10.2 Geologic Map of the Piedmont, Southeastern Pennsylvania (diagonal line pattern = basement gneiss; vertical line pattern = Wilmington complex; no pattern = Glenarm supergroup and Paleozoic rocks; black = ultramafic rocks; and stipple = sedimentary cover) (Source: Crawford and Crawford, 1980)

several pegmatite dikes intrude into both the granite and gabbroic gneisses. Three small pegmatite dikes are found at the extreme southeastern tip of the complex.

10.1.2.6 Country Rocks

On the northeast the Honey Brook complex is in contact with the Stockton formation of the Newark basin, which is about 3.2 km (2 mi) deep (Faill, 1973). The Stockton formation consists of light-gray to buff, coarse-grained arkosic sandstone and conglomerate, with red and brown, fine-grained, siliceous sandstone and shale (Gray et al., 1960). The only other contact is with the Chickies formation of Cambrian age. This formation is a massive, hard, scolithus-bearing, gray quartzite and quartz schist, with a thin layer of dark interbedded slate at its upper surface and a conglomerate at its base.

10.1.2.7 Rock Structure

The Brandywine Manor fault cuts through the center of the Honey Brook complex and may have brought the otherwise cohesive metamorphic terrane into its present position. This fault represents at least 914 m (3000 ft) of upthrust of Precambrian rocks to the north of the fault. This upthrusting along a steep fault accounts for the anorthositic core being exposed at the surface. The Gap overthrust fault has a general east-west strike; it cuts through the southern third of the complex and runs subparallel to the Brandywine Manor fault. Upthrusting occurred in the region south of this fault (Bascomb and Stose, 1938).

According to Gray et al. (1960), several short faults exhibiting little vertical movement intersect the northeastern boundary of the Precambrian outcrop. These faults all strike about N. 30° E. and are all less than 6 km (4 mi) long.

This area is currently interpreted as the result of a Taconic nappe structure, with an Alleghenian ramp-and-flat-decollement deformation superposed on the earlier extreme deformation (Drake, 1982). The complex is composed of three adjacent anticlines that are bounded by the faults discussed above. Furthest north is the Welsh Mountain anticline, which extends northward past the Precambrian rocks to include exposures of Cambrian metasediments. This anticline is believed to be severely eroded. It terminates abruptly at the Brandywine fault.

At the center of the body is the Barren Hills anticline, which occupies a slight depression defined by the Brandywine and Gap faults. This anticline is westward-plunging. Finally, the southernmost third of the Honey Brook region is known as the Mine Ridge anticline. Folding in this region was extremely intense, and the structure is complicated by extensive faulting. It is possible that faulting and folding were contemporaneous developments (Bascomb and Stose, 1938).

10.1.2.8 Geophysics

An aeromagnetic map of the Wagontown quadrangle indicates that the anorthosite body represents a magnetic low in the Honey Brook complex. The magnetic contrast between the rocks indicates that the anorthosite is a steep-sided stocklike body (Crawford and Hoersch, 1983) that extends to a moderate depth (Bromery et al., 1960).

10.1.3 Housatonic Massif (Plate IX, Pluton 6)

See Sec. 9.1.3 and Plate VIII, pluton 7. Only about 25 km^2 (10 mi^2) of this pluton occurs in New York; the remainder lies in Connecticut, and it is therefore discussed with the plutons of that state.

10.1.4 Manhattan Prong (Plate IX, Pluton 7)

10.1.4.1 Size, Shape, and Location

The Manhattan Prong comprises a group of elongated parallel bodies that trends northeastward across southeastern New York into southwestern Connecticut. The Manhattan Prong underlies an area of about 406 km^2 (157 mi^2) and includes several different lithologies.

10.1.4.2 Relative and Absolute Age

On the basis of lithostratigraphic relations, crystalline rocks of the Manhattan Prong (Fordham gneiss and Yonkers gneiss) are assigned a late Proterozoic age and are considered part of the "Grenville Terrain" (Hall, 1968a; Rodgers 1982). On the basis of Rb-Sr whole-rock data, Long (1969) reported that the Yonkers gneiss of the Manhattan Prong is 450-1000 m.y. old and similar to rocks associated with the Avalonian event. Radiometric data suggest that rocks of the Manhattan Prong underwent a major thermal event about 360 m.y. ago, which indicates possible influence by the Acadian orogeny (Long and Kulp, 1962; Hall, 1968a).

10.1.4.3 Origin

The Yonkers gneiss is considered to be a metamorphosed rhyolite or arkose. Most of the Fordham gneiss consists of metamorphosed clastic sedimentary rocks. Amphibolites in the Fordham gneiss are believed to be metamorphosed mafic volcanics, whereas granitic gneisses are believed to be metamorphosed rhyolitic volcanic rocks. These gneisses probably represent a metamorphosed eugeosynclinal sequence that contained intrusive rocks (Hall, 1968a and 1976).

10.1.4.4 Petrography

Mineralogy and Textural Relations. Metasedimentary rocks of the Manhattan Prong include the Fordham gneiss, Lowerre quartzite, Inwood marble, and Manhattan schist. Metaigneous or metasedimentary rocks include the Yonkers and Pound Ridge gneisses. Igneous rocks consist of the Cortlandt complex, the Peekskill granite, and smaller intrusive bodies (Mose and Hayes, 1975).

Rocks of granitic composition are primarily the Fordham gneiss, the Yonkers gneiss, and the Manhattan schist. The Fordham gneiss has been extensively studied in the White Plains quadrangle (Hall, 1968b) and Glenville quadrangle (Hall, 1968a and 1968b). In the White Plains quadrangle, the Fordham gneiss consists of five members referred to as A through E:

Member A is composed of garnet-biotite-quartz-feldspar gneiss, amphibolite, and gray biotite-hornblende-quartz-feldspar gneiss. Member B consists, for the most part, of gray garnet-biotite-quartz-feldspar gneiss that is locally sillimanite bearing, and amphibolite. Member C is similar to Member B, but is hornblende-bearing; it also contains layers of pinkish gneiss similar to the Yonkers gneiss. Member D is sulfide-bearing sillimanite-garnet-biotite-quartz schist or schistose gneiss, with local layers of quartzite or biotite gneiss and calc-silicate granulite. Member E consists of gray garnet-biotite-quartz-feldspar gneiss, biotite-quartz-feldspar gneiss, and amphibolite, all interlayered.

In the Glenville area, amphibolite and calc-silicate rocks are recognized as important components of the five lithologic units (Hall, 1968b and 1976). The sequential and spatial relations of lithological subdivisions in the two quadrangles have not yet been fully deciphered.

Brock and Brock (1983) reported that the granulite facies assemblage is also present in the Fordham gneiss of the Manhattan Prong and the Hudson Highlands. In contrast to the highly heterogeneous nature of the Fordham gneiss, the Yonkers gneiss is commonly pinkish, homogeneous biotite-hornblende-quartz-feldspar gneiss.

Alteration. The rocks of the Manhattan Prong were metamorphosed during the main phase of the Taconic orogeny. All of the Manhattan Prong region, except the western margin, underwent a second thermal event during the Acadian orogeny (Hall, 1968a).

10.1.4.5 Secondary Intrusions

The rock formations of the Manhattan Prong are complex, and some contain pegmatitic intrusions that are probably anatectic.

10.1.4.6 Country Rocks

The Manhattan Prong is composed of rocks of metasedimentary or metavolcanic origin; they were not igneous intrusives. The geological relations of the Manhattan Prong are shown in Fig. 10.3. It is bounded on the west by the Mesozoic sedimentary rocks of the Newark basin, on the north by the Precambrian rocks of the Hudson Highlands, on the east by Cambrian and Ordovician eugeosynclinal rocks of the Hartland formation east of Cameron's line, and on the south by Cenozoic and younger sedimentary rocks and sediments.

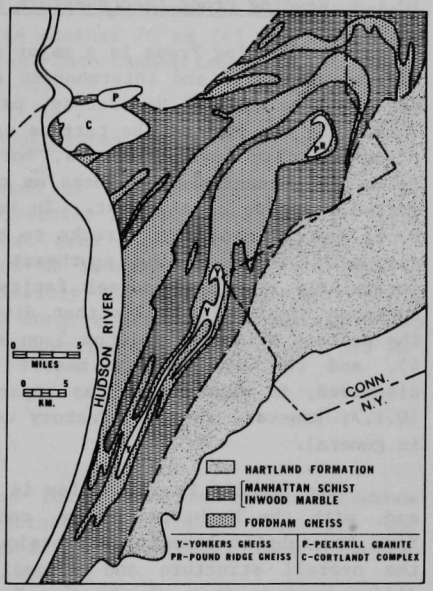


Fig. 10.3 Generalized Geologic Map of Manhattan Prong (Source: Mose and Hayes, 1975)

10.1.4.7 Rock Structure

There is some evidence of faulting in the Manhattan Prong (Ratcliffe, 1971), but it is not always structural evidence.

At least three separate episodes of folding occurred in the Manhattan Prong. The first deformation involved intense folding and high-grade metamorphism at some time during the Precambrian. This was followed by episodes of deformation in the Ordovician, the latest of which was extremely intense and occurred in or later than the Middle Ordovician. Brock and Brock (1983) proposed that the Fordham gneiss originally was the floor for Paleozoic cover rocks and was later isoclinally folded. It is considered possible that the major deformational events in the Manhattan Prong occurred during the Taconic and Acadian orogenies (also see Merguerian, 1983).

10.1.4.8 Geophysics

The Manhattan Prong is defined by -20-mgal to -40-mgal Bouguer-anomaly contours on the gravity map of Diment et al. (1980); the Prong is part of the east-to-northeasterly gravity trend characterized by the metamorphic core rocks of the Appalachian belt.

10.1.5 Reading Prong, Northwestern Massif (Plate IX, Pluton 4)

The Reading Prong is a major northeasterly trending complex of gneissic ridges, plateaus, and intermontane sedimentary sequences that covers much of the northern part of New Jersey, parts of Pennsylvania, and southeastern New York. The map plan of the terrane exhibits nearly continuous streaks or belts of metamorphic rock sequences, bordered along its entire length by lower Paleozoic metasedimentary rocks on the northwest and Mesozoic sedimentary and intrusive rocks on the east. In New Jersey the contact between the Reading Prong and the sedimentary rocks to the northwest is defined by reverse faults and overthrusts; to the southeast that contact is marked by a system of irregularly developed normal faults. Within the Reading Prong itself are numerous high-angle faults that divide the complex into three main regions: the western extension (pluton 3 on Plate IX), the northwestern massif (pluton 4), and the southeastern massif (pluton 5). These three regions are discussed, in this report, as separate plutons in Secs. 10.1.5, 10.1.6, and 10.1.7; however, the introductory comments below describe the entire complex in general.

The Reading Prong complex is associated with the Green Mountain massif and, with the Manhattan Prong, constitutes the Pennsylvania-New Jersey-New York Precambrian basement expression of the Appalachian Mountains. However, the overall structure and composition of the Reading Prong terrane are different from that of the New England Berkshire and Housatonic highlands (Smith, 1969). Drake (1982) concluded that the Berkshires are more clearly characterized by nappe and thrust structures; Drake further noted that many Reading Prong rocks are metasedimentary and metavolcaniclastic, and that the granitic rocks are probably syntectonic anatectic sheets.

The Precambrian-Paleozoic contact of the Green Mountain-Reading Prong arc is overturned and overthrust to the west. The crystalline part of the entire Reading Prong has been interpreted to be a far-travelled allochthonous mass, or klippe, of Precambrian and early Paleozoic rocks overthrust westward into northern New Jersey during the Taconic orogeny. Drake (1980) included the Reading Prong in the "Taconides" terrane. Helenek (1983) proposed that the quartz-plagioclase gneisses were derived from volcanic island-arc protoliths.

In Pennsylvania the Reading Prong is known as the Reading and Durham Hills; in New Jersey as the New Jersey Highlands; and in New York as the Hudson Highlands. Although a terminal moraine obscures a middle section of the terrane, the region was not extensively glaciated. Because the rocks are deeply weathered, fresh outcrops are rare, except in the southeastern massif.

The Reading Prong province runs approximately 240 km (150 mi). It extends from near the Schuykill River in eastern Pennsylvania about 70 km (43 mi) to the New Jersey border, where it is approximately 16 km (10 mi) wide. It continues northeast across New Jersey about 100 km (62 mi), and widens to 35 km (22 mi) at the New York border between the Kittatinny valley and the

Piedmont plateau. The northwestern massif tapers out about 20 km (12 mi) into New York, but the eastern branch continues another 70 km (43 mi) across New York, even extending a few kilometers (miles) into Connecticut.

10.1.5.1 Size, Shape, and Location of Pluton 4

The northwestern massif is composed of five distinct, subparallel blocks, which trend in a northeasterly direction and are separated by high-angle faults. Young (1978) reported that each block is a succession of complexly deformed paragneisses and sheets of metamorphosed salic intrusive rocks. The northwestern massif lies primarily in New Jersey, where it is about 15 km (9 mi) wide and 55 km (34 mi) long; a narrow strip extends about 18 km (11 mi) into New York. The massif underlies an area of about 572 km^2 (220 mi^2).

10.1.5.2 Relative and Absolute Age

The entire Reading Prong has been interpreted as an allochthonous thrust sheet and, therefore, determinations of relative age are not possible. Absolute ages, determined by Lapham and Root (1971), were 1100-1150 m.y., which places these rocks in the Precambrian.

10.1.5.3 Origin

The entire crystalline belt of the Reading Prong has been interpreted to be a far-travelled allochthonous mass overthrust in a westerly direction during the Taconic orogeny (Drake, 1970). The nature of the original magmatic emplacement is therefore obscured by the tectonic events that moved the granitic rocks from their original environment.

10.1.5.4 Petrography

Mineralogy and Textural Relations. The northwestern massif consists of five structural blocks isolated by high-angle faults. The intrusive rocks of the Pochuck Mountain block are isolated stocklike and sill-like masses of buff-colored, structureless, medium- to coarse-grained biotite or hornblende granite and leucogranite (Young, 1978). The major intrusive rocks of this block consist of quartz, microperthite microcline, oligoclase, hornblende, and biotite. Magnetite, zircon, apatite, sphene, allanite, and fluorite occur as accessory minerals.

The Pimple Hills-Allamuchy block is made up of various gneisses that are not of intrusive origin. In the Sparta Mountain-Cranberry Lake block, gneisses dominate and are intruded into by thick sheets of syenitic to monzonitic rocks, with quartz syenite, quartz monzonite, and granitic variants.

The Beaver Lake block is dominated by stratigraphically layered gneisses that have been folded into an isoclinally plunging antiform. A sheet of hornblende granite occurs in the nose of the Beaver Lake antiform. It is a medium-grained, hypidiomorphic-granular to granoblastic rock, with microperthite set in a groundmass of finer-grained plagioclase and microcline. Hornblende, quartz, magnetite, zircon, apatite, and (locally) biotite are also present.

The Wawayanda-Hopatong block is composed chiefly of gneissose rocks including granitic and quartz syenitic variants. These are medium grained and foliated, with hypidiomorphic texture. They are composed mainly of clinopyroxene, quartz, mesoperthite, microperthite, plagioclase, magnetite, and sphene. Hornblende, zircon, and apatite are found locally.

The Mase Mountain block is also composed mainly of gneissose rocks, the principal intrusive component of which is a hornblende-bearing leucogranite. This is a coarse-grained, well-foliated, pinkish rock, with locally prominent lineation defined by quartz rods and hornblende needles. Where it is uncontaminated, the leucogranite has hypidiomorphic-granular texture and is composed of quartz, microcline microperthite, and oligoclase, with minor chloritized hornblende, magnetite, zircon, and apatite.

Gneisses of the northwestern massif include quartz-oligoclase gneiss, microcline gneiss, biotite-bearing gneiss, hornblende-bearing gneiss, biotite-quartz-plagioclase gneiss, diopside-epidote-quartz gneiss, quartz-oligoclase leucogneiss, quartz-monzonite to quartz-syenite orthogneiss, quartz-microcline gneiss, epidote-scapolite-quartz gneiss, hypersthene-quartz-plagioclase gneiss, and clinopyroxene-quartz-plagioclase gneiss.

Alteration. Secondary features are related to retrograde metamorphism during one or more Paleozoic orogenic events. Dallmeyer (1973) reported that a common feature is sericitic replacement of plagioclase. Clinozoisite and epidote are commonly part of this saussuritic alteration, but they also occur locally as discrete, interstitial grains that are often intergrown with secondary calcite.

10.1.5.5 Secondary Intrusions

The Mase Mountain block has been intruded into by a sheet of fayalite-bearing clinopyroxene-perthite quartz syenite (Young, 1972). This rock is a folded sheet of dark-greenish-gray, coarse-grained, massive rock with hypidiomorphic-granular texture, and is composed of ferrohedingbergite, quartz, and mesoperthite, with minor fayalite, zircon, apatite, and magnetite (Young, 1972).

10.1.5.6 Country Rocks

The northwestern massif is unconformable beneath Cambrian and Ordovician sedimentary rocks. These rocks include orthoquartzite carbonate shelf deposits (Lower Cambrian through Lower Ordovician), deeper-water neritic carbonate deposits, and graywacke-shale flysch deposits (Middle and Upper Ordovician).

10.1.5.7 Rock Structure

The entire northwestern massif is bisected and transected by seven major high-angle fault zones. This faulting has created five major structural blocks. The general geologic map of the massif outlines these blocks and the faulting associated with them. The Pochuck Mountain block is isolated on its southeastern flank by the East fault zone. Contacts with Paleozoic rocks to the north and south are interrupted by high-angle faults. The Pimple Hills-Allamuchy block is bounded on the northwest by the Hamburg fault zone. Along the entire southeastern margin, the block is bounded by the Zero fault zone and is on the upthrown side of the zone. The Sparta Mountain-Cranberry Lake block is bounded on the northeast by the Zero fault zone and along its southeastern margin by the East fault zone. The two major slices within the block are separated by the Wright Pond fault zone, and the downthrown side is on the southeastern side of the zone. The Beaver Lake block is bounded by the East fault zone to the northwest. Along the northwestern end of the zone, the Beaver Lake block has dropped downward; farther south the block has moved relatively upward, owing to the wedging in of the Sparta Mountain-Cranberry Lake block. Along the southeastern margin, the block is bounded by the Vernon fault zone. The block is bounded by the Reservoir fault to the southeast. The Reservoir fault is a high-angle reverse fault that has been seismically active recently. The Mase Mountain block is bounded by the Reservoir fault to the northwest and by the Longwood Valley fault to the southeast.

The Beaver Lake block has been folded into a large isoclinal, northeasterly plunging antiform. The gneisses of the Mase Mountain block have been cast into a series of open, northeasterly plunging folds and are syntectonically intruded by hornblende-bearing leucogranite.

10.1.5.8 Geophysics

The Reservoir fault has been seismically active recently. Sbar et al. (1970) reported a 3-mo occurrence of seismicity in the Wawayanda-Hopatcong block. The activity, which occurred around the Hopatcong block, lowered the level of Hopatcong Lake by 81 cm (32 in.) (Sbar et al., 1970). The epicenters were very shallow (0.5-3 km [0.3-2 mi]), the magnitude low (1.25), and the intensity IV. The origin of the seismicity is unknown.

Aeromagnetic surveys have been conducted for every 7.5-min quadrangle in the Reading Prong area.

The northwestern massif of the Reading Prong lies on a gravity gradient that is characteristic for Precambrian rocks north through Vermont and south to the Blue Ridge anticlinorium. Most of the Reading Prong lies over a negative gravity anomaly (Diment et al., 1980).

10.1.6 Reading Prong, Southeastern Massif (Plate IX, Pluton 5)

10.1.6.1 Size, Shape, and Location

The southeastern massif is composed of three blocks outlined by high-angle faults that trend in a northeasterly direction. As with the northwestern massif, the blocks are composed of a complex succession of deformed paragneisses and sheets of metamorphosed salic intrusive rocks. The southeastern massif, roughly 150 km (93 mi) long and, on average, 18 km (11 mi) wide, falls about equally in New Jersey and New York, and continues a small distance into Connecticut. The total area occupied is about 1990 km^2 (768 mi^2).

10.1.6.2 Relative and Absolute Age

Because the Reading Prong has been interpreted as an allochthonous thrust sheet, determination of relative age is not possible. Absolute ages determined by Lapham and Root (1971) are about 1100-1150 m.y., which places these rocks in the Precambrian.

10.1.6.3 Origin

The entire Reading Prong crystalline belt has been interpreted as a far-travelled allochthonous mass overthrust in a westerly direction during the Taconic orogeny (Drake, 1970 and 1980). The nature of the original magmatic emplacement is therefore obscured by the tectonic events that moved the rocks from their original environment.

10.1.6.4 Petrography

Mineralogy and Textural Relations. The southeastern massif consists of three structural blocks separated by high-angle faults. The Monroe-Dover block is composed of paragneisses intruded into by thick sheets of gneissoid hornblende granite. The granites are medium to coarse grained, foliated, or lineated, with hypidiomorphic-granular texture. They are composed of

hornblende, quartz, microcline microperthite, and oligoclase, with minor magnetite, zircon, and apatite. Biotite-bearing facies occur locally.

The Ramapo block is composed of gneisses that have been injected by large sheets of massive, foliated, or lineated hornblende granite. The granitic rocks are medium to coarse grained with hypidiomorphic-granular texture. The major minerals are hornblende, quartz, microcline microperthite, and oligoclase.

The Brewster block also is composed of gneisses that were invaded by large masses of hornblende or biotite granite. The granite is granoblastic to hypidiomorphic-granular and is composed essentially of quartz, microcline microperthite, sodium oligoclase, and hornblende or biotite, with accessory magnetite, zircon, apatite, and sphene.

The gneisses of the southeastern massif include pyroxene-hornblende-plagioclase gneiss, hypersthene-quartz-plagioclase-gneiss, biotite-quartz-plagioclase gneiss, quartz-oligoclase leucogneiss, and sillimanite- and cordierite-bearing biotite-quartz-feldspar gneiss.

Alteration. Secondary features are related to retrograde metamorphism during one or more Paleozoic orogenic events. Dallmeyer (1973) reported that a common feature is sericite replacement of plagioclase. Clinozoisite or epidote or both are commonly part of this saussuritic alteration, but they also occur as discrete, interstitial grains that are often intergrown with secondary calcite.

10.1.6.5 Secondary Intrusions

No significant secondary intrusions have been reported in the southeastern massif.

10.1.6.6 Country Rocks

The Reading Prong is flanked by sedimentary rocks of Cambrian and Ordovician age that were deposited in a Paleozoic syncline (Drake, 1970). These rocks consist of orthoquartzite shelf deposits, deeper-water neritic carbonate deposits, and a graywacke-shale flysch deposit.

10.1.6.7 Rock Structure

The Monroe-Dover block borders the Green Pond syncline to the northwest. In New York, this boundary is a high-angle fault along which the Precambrian crystalline rocks are on the upthrown side and the Paleozoic rocks are on the downthrown side. The block is bounded in New York on the southeastern side by a high-angle fault that follows the Ramapo River to the

Wanaque Reservoir at Pompton, N.J. In central New Jersey, the entire southeastern massif consists of a single block. A few small, east-trending cross faults transect the Monroe-Dover block.

The Ramapo block is bounded by a reverse fault that separates it from the Monroe-Dover block towards the southwest and from lower Paleozoic rocks toward the northeast. The block is bounded to the southeast by the Ramapo fault zone. This zone of normal faulting separates Precambrian crystalline rocks from Triassic basalts, red beds, and lacustrine deposits. Northeast of the Hudson River, the block is bounded by the Canopus fault zone (Ratcliff, 1971), which separates it from rocks of the Reading Prong that were severely affected by Paleozoic metamorphism and deformation (Long and Kulp, 1962; Dallmeyer, 1974). Near the Hudson River, the Ramapo block has been faulted into a number of smaller subparallel slices (Helenek, 1965 and 1971; Fisher et al., 1970). The blocks in the northern portion are either in fault contact with or unconformably overlain by Paleozoic rocks in Dutchess County, N.Y. The southern terminus occurs near Pompton, N.J., where the two bounding fault zones converge. The Ramapo-Canopus fault zone subdivides the Hudson Highlands into eastern and western lithologic units (Hall, 1980). The Brewster block lies east of the Canopus fault zone. The southern boundary has been interpreted by Prucha (1968) as a high-angle fault.

The Precambrian crystalline rocks of the entire Reading Prong have been complexly folded. Folds range in size up to 11 km (7 mi) long parallel to the axis and 1.6 km (1 mi) in width. The folds range from upright and open to isoclinal and overturned, the latter being the most common. Most folds plunge northeast, essentially parallel to the regional lineation.

10.1.6.8 Geophysics

Seismicity in northern New Jersey and southeastern New York is attributed to thrust motion (e.g., of the Ramapo fault along northeast-trending planes) (Aggarwal and Sykes, 1978). The Precambrian ridges are bounded on the north by strong negative gravity anomalies and on the south by gravity gradients that drop off into the trench of negative anomaly north of an adjacent Precambrian ridge. Drake (1970) interpreted this configuration as due to the absence of Precambrian rocks beneath the Paleozoic rocks on the north sides of the ridges but to their presence, buried beneath Paleozoic rocks, on the southern sides. His model suggests a series of southeast-dipping, rootless Precambrian bodies that have nonmagnetic material in front, below, and behind.

The southeastern massif lies on a gravity gradient northeast of a gravity low in the western extension, The gravity gradient continues north along a belt of Precambrian crystalline rocks to Vermont, and to the south along the Blue Ridge anticlinorium.

The rocks of the Reading Prong mark the western edge of the gravity trend characterized by the metamorphic core of the Appalachian belt (Diment et al., 1980).

10.1.7 Reading Prong, Western Extension (Plate IX, Pluton 3)

10.1.7.1 Size, Shape, and Location

The western extension of the Reading Prong is composed of a group of northeasterly trending ridges isolated by valleys underlain by Paleozoic sedimentary rocks (Young, 1978). This part of the prong is about 125 km (78 mi) long and extends about 70 km (43 mi) along its strike into Pennsylvania. The general shape of the group of ridges is elongated; they underlie an area of about 822 km^2 (317 mi^2). Drake (1970) showed that the Precambrian crystalline rocks underlying the ridges are remnants of the basement core of the Musconetong nappe, emplaced toward the northwest during the Late Ordovician.

10.1.7.2 Relative and Absolute Age

The relative age of the crystalline rocks is Precambrian. Although the crystalline basement rocks are in contact with Paleozoic rocks, these are unconformable contacts. Absolute age determinations have yielded ages of 1100-1150 m.y. (Lapham and Root, 1971).

10.1.7.3 Origin

Because the crystalline basement was tectonically emplaced as a thrust sheet during the Taconic orogeny (Drake, 1970), evidence of its original magmatic emplacement is obscure.

10.1.7.4 Petrography

Mineralogy and Textural Relations. The intrusive rocks of the western extension of the Reading Prong and the southeastern massif consist predominantly of gneissoid hornblende granite or hornblende granite gneiss. The rocks have relatively little clinopyroxene and mesoperthite and are relatively rich in quartz. In the northwestern part of the prong, hornblende and clinopyroxene granite is common, but hornblende- or clinopyroxene-bearing syenitic and monzonitic rocks or their variants are equally abundant. These crystalline rocks are relatively enriched in clinopyroxene and mesoperthite and somewhat poor in quartz.

Alteration. Retrograde features induced during Paleozoic orogenies are present.

10.1.7.5 Secondary Intrusions

No significant secondary intrusive phases are reported for the area.

10.1.7.6 Country Rocks

To the west, Reading Prong rocks are flanked by the miogeosynclinal sedimentary rocks of the Appalachian mountains.

10.1.7.7 Rock Structure

Young (1978) suggested that the western extension of the Reading Prong was on the move prior to Paleozoic time. Reconstruction indicates the existence of two parallel intrusive belts in the Precambrian -- a northwestern syenitic belt and a southeastern granitic belt. The syenitic and monzonitic rocks were generated mainly by very deep-level partial melting of quartzo-feldspathic and anorthositic rocks in a high-temperature, relatively anhydrous environment. The massif is composed of a series of subparallel, northeasterly trending blocks apparently separated by high-angle faults. Within each block is a succession of complexly deformed paragneisses and sheets of metamorphosed silicic intrusive rocks.

10.1.7.8 Geophysics

See Sec. 10.1.6.8.

10.2 TACONIC

10.2.1 Bedford Gneiss (Plate IX, Pluton 10)

See Sec. 9.2.5 and Plate VIII, pluton 1. Only about 32 km^2 (12 mi^2) of this body of crystalline rock lies in New York; a much larger area of outcroppings occurs in Connecticut, where it is identified as the Harrison gneiss.

10.2.2 Granitic Gneiss (Plate IX, Pluton 8)

See Sec. 9.2.4 and Plate VIII, pluton 3. Only about 10 km^2 (4 mi^2) of this gneiss occurs in New York, whereas it underlies a much larger area in Connecticut.

10.2.3 Harrison Gneiss (Plate IX, Pluton 11)

See Sec. 9.2.5 and Plate VIII, pluton 1. Only about 38 km^2 (15 mi^2) of this rock lies in New York. This is the southern end of the Harrison gneiss of Connecticut, which has a sizeable exposure there; the northern end of Connecticut's Harrison gneiss also appears in a small area of New York as the Bedford gneiss (see Sec. 10.2.1).

10.2.4 Norwalk Area Gneiss (Plate IX, Pluton 9)

See Sec. 9.2.7 and Plate VIII, pluton 2. The largest part of this pluton occurs in Connecticut; only about 10 km^2 (4 mi^2) is found in New York.

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APPENDIX

AGES OF CRYSTALLINE ROCK BODIES DESCRIBED IN THIS REPORT

Table A.1 Ages of Crystalline Rock Bodies of Maine

Plate II Number	Name	Age (m.y.)	Dating Method	Reference[a]
	Grenville/Middle Proterozoic (1600-1000 m.y.)			
21	Chain Lakes Massif	1500-1600	Pb-Pb zircon	Naylor et al. (1973)
	Taconic/Cambrian-Silurian (570-408 m.y.)			
22	Attean Quartz Monzonite	470	U-Pb zircon	Boudette and Boone (1976)
		395-500	Pb-α	Faul et al. (1963)
		356 ± 15	K-Ar biotite	Faul et al. (1963)
	Acadian-Post-Acadian/Devonian-Pennsylvanian (408-286 m.y.)			
3	Biddeford Pluton	345 ± 12	Rb-Sr whole rock	Gaudette et al. (1975)
27	Bottle Lake Quartz Monzonite	393 ± 11	Pb-Pb zircon	Ayuso and Arth (1983)
		380 ± 5	Rb-Sr whole rock	Ayuso and Arth (1983)
		360 ± 16	Rb-Sr whole rock	Ayuso and Wones (1980)
		380	Pb-α zircon	Faul et al. (1963)
39	Cadillac Mountain and Somesville Plutons	400-389	K-Ar biotite	Faul et al. (1963)
29	Calais Gabbro-Diorite Complex	Devonian	Lithostratigraphic correlation	Spooner and Fairbairn (1970)
26	Center Pond Quartz Monzonite	380 ± 5	Pb-Pb zircon	Loiselle et al. (1983b)
		372 ± 10	Rb-Sr whole rock	Loiselle et al. (1983b)
		340-370	Pb-α zircon	Larrabee et al. (1965)
		349-359	K-Ar biotite	Larrabee et al. (1965)
		350 ± 100	Rb-Sr whole rock	Larrabee et al. (1965)

Table A.1 (Cont'd)

Plate II Number	Name	Age (m.y.)	Dating Method	Reference
	Acadian-Post-Acadian/Devonian-Pennsylvanian (408-286 m.y.) (Cont'd)			
28	Chiputneticook Quartz Monzonite	389 ± 20 400	Rb-Sr whole rock K-Ar biotite	McCutcheon et al. (1981) Faul et al. (1963)
5	Concord and Related Plutons	359 ± 11	Rb-Sr whole rock	Lyons and Livingston (1977)
35	Cranberry (Wabassus) Lake Pluton	356 420	K-Ar biotite Pb-α zircon	Faul et al. (1963) Faul et al. (1963)
19	Flagstaff Lake Complex	Devonian	Lithostratigraphic relations	Boone (1973)
31	Gardner Lake Gabbro-Diorite Complex	Devonian	Lithostratigraphic correlation	Crats (1978); Doyle (1967)
38	Gouldsboro Pluton	Devonian	Lithostratigraphic relations	Doyle (1967); Chapman (1962)
32	Granodiorite of Whitneyville	Devonian	Lithostratigraphic correlation	Westerman (1973 and 1979); Doyle (1967)
9	Hallowell and Togus Plutons	387 ± 11 394 ± 8	Rb-Sr whole rock	Dallmeyer and Van Breeman (1978 and 1981)
10	Hartland Pluton	360 ± 8 375 356	Rb-Sr whole rock K-Ar biotite K-Ar biotite	Dallmeyer et al. (1982) Hurley (1964) Faul et al. (1963)
24	Katahdin Batholith	414 ± 4 388 ± 5 355-390	Pb-Pb zircon Rb-Sr whole rock K-Ar biotite	Loiselle et al. (1983a) Loiselle et al. (1983a) Faul et al. (1963)

Table A.1 (Cont'd)

Plate II Number	Name	Age (m.y.)	Dating Method	Reference
	Acadian-Post-Acadian/Devonian-Pennsylvanian (408-286 m.y.) (Cont'd)			
36	Lead Mountain Pluton	393 ± 17	Rb-Sr whole rock	Loiselle et al. (1983b)
17	Lexington Pluton	360 ± 12	Rb-Sr whole rock	Naylor (1975)
41	Lucerne Pluton	371 ± 21	Rb-Sr whole rock	Loiselle et al. (1983b)
		380 ± 5	Pb-Pb zircon	Loiselle et al. (1983b)
		365	K-Ar biotite	Faul et al. (1963)
4	Lyman Pluton	322 ± 12	Rb-Sr whole rock	Gaudette (in press)
30	Meddybemps and Related Granite	400	Rb-Sr whole rock	Spooner and Fairbairn (1970); Faul et al. (1963)
13	Mooselookmeguntic Pluton	379 ± 6	Rb-Sr whole rock	Moench and Zartman (1976)
42	Mt. Waldo Pluton	390 ± 10	Rb-Sr whole rock	Sweeney (1974)
		325	K-Ar biotite	Zartman et al. (1970)
23	Moxie Pluton	393	K-Ar biotite	Faul et al. (1963)
33	Peaked Mountain Pond Gabbro-Diorite Complex	Devonian	Lithostratigraphic correlation	Doyle (1967); Westerman (1979)
		332	K-Ar biotite	Faul et al. (1963)
12	Phillips Pluton	379 ± 6	Rb-Sr whole rock	Moench and Zartman (1976)
18	Pierce Pond Pluton	Devonian	Lithostratigraphic relations	Boone (1973)

Table A.1 (Cont'd)

Plate II Number	Name	Age (m.y.)	Dating Method	Reference
	Acadian-Post-Acadian/Devonian-Pennsylvanian (408-286 m.y.) (Cont'd)			
34	Pocamoonshine Gabbro-Diorite	Devonian	Lithostratigraphic relations	Doyle (1967)
7	Quaker Hill Pluton	Devonian	Lithostratigraphic correlation	Doyle (1967)
15	Redington Pluton	Devonian	Lithostratigraphic correlation	Doyle (1967); Moench (1970)
11	Rome Pluton	314-281	Ar-Ar biotite	Dallmeyer (1979)
6	Sebago Lake Batholith (includes Songo granodiorite)	340-252 227	K-Ar biotite Ar-Ar biotite	Guidoti (1965) Dallmeyer (1979)
25	Seboeis Complex (or Schoodic Lake granite)	Devonian	Lithostratigraphic correlation	Doyle (1967)
40	Sedgwick Granite	395 ± 15	Rb-Sr whole rock	Brookins (1967)
20	Seven Ponds Pluton	Devonian	Lithostratigraphic correlation	Harwood (1973)
16	Sugarloaf Pluton	397 ± 12	K-Ar biotite	Zartman et al. (1970)
37	Tunk Lake Pluton	369	K-Ar biotite	Faul et al. (1963)
14	Umbagog Granodiorite (includes Adamstown pluton)	Ordovician to Devonian	Lithostratigraphic relations	Hussey and Pankiwskyj (1980); Doyle (1967)

Table A.1 (Cont'd)

Plate II Number	Name	Age (m.y.)	Dating Method	Reference
	Acadian-Post-Acadian/Devonian-Pennsylvanian (408-286 m.y.) (Cont'd)			
8	Waldoboro Pluton	295 ± 9 Devonian	K-Ar biotite Lithostratigraphic correlation	Zartman et al. (1970) Doyle (1967)
2	Webhannet Pluton	400	Rb-Sr whole rock	Gaudette et al. (1975)
	Alleghenian-Post Alleghenian/Permian-Cretaceous (286-66 m.y.)			
1	Agamenticus Complex	228 ± 5 216 ± 4	K-Ar biotite K-Ar biotite	Foland and Faul (1977) Foland et al. (1971)

[a]Reference details are given in Sec. 3.5 or in the references section of the adjoining state.

Table A.2 Ages of Crystalline Rock Bodies of New Hampshire

Plate III Number	Name	Age (m.y.)	Dating Method	Reference[a]
	Avalonian/Late Proterozoic (1000-570 m.y.)			
20	Massabesic Gneiss	646	U-Th-Pb zircon	Aleinikoff et al. (1979)
	Taconic/Cambrian-Silurian (570-408 m.y.)			
23	Ayer Intrusive Complex	435 ± 10	U-Th-Pb zircon	Peck (1976)
10	Jefferson Dome	455	U-Th-Pb zircon	Zartman and Leo (1981)
		441 ± 5	Rb-Sr whole rock	Foland and Loiselle (1981)
9	Lost Nation Group	444-451	U-Th-Pb zircon	Lyons (1981)
6	Mascoma and Lebanon Domes	440 ± 20	Rb-Sr whole rock	Naylor (1969)
22	Newburyport Granodiorite	Silurian or Ordovician	Lithostratigraphic correlation	Zen (1983)
		321 ± 10	K-Ar biotite	Zartman et al. (1970)
3	Westmoreland-Swanzey and Related Domes	455	K-Ar biotite	Zartman and Leo (1981)
	Acadian-Post-Acadian/Devonian-Pennsylvanian (408-286 m.y.)			
2	Ashuelot Pluton[b]	411 ± 19	Rb-Sr whole rock	Lyons and Livingston (1977)
5	Cardigan Pluton[b]	402 ± 19	Rb-Sr whole rock	Lyons and Livingston (1977)

Table A.2 (Cont'd)

Plate III Number	Name	Age (m.y.)	Dating Method	Reference
	Acadian-Post-Acadian/Devonian-Pennsylvanian (408-286 m.y.) (Cont'd)			
7	Concord and Related Plutons	359 ± 11	Rb-Sr whole rock	Lyons and Livingston (1977)
21	Exeter Pluton	390	Rb-Sr whole rock	Gaudette et al. (1975)
24	Fitchburg Intrusive Complex	382-412	U-Pb zircon	Maczuga (1981)
1	Fitzwilliam Granite	383	Rb-Sr whole rock	Hayward (1983)
19	Granite Pluton near Barrington	Devonian	Lithostratigraphic correlation	Lyons and Livingston (1977)
8	Lincoln Pluton[b]	402 ± 19	Rb-Sr whole rock	Lyons and Livingston (1977)
12	Long Mountain Granite	323 ± 11	Rb-Sr whole rock	Lyons and Livingston (1977)
4	Mt. Clough Pluton (includes Bethlehem gneiss)	405 ± 78	Rb-Sr whole rock	Lyons and Livingston (1977)
15	Sebago Lake Batholith	340-252 227	K-Ar biotite K-Ar biotite	Guidotti (1965) Dallmeyer (1979)
13	Umbagog Granodiorite	Ordovician to Devonian	Lithostratigraphic correlation	Hussey and Pankiwskyj (1980); Doyle (1967)
18	Weare and Spaulding Plutons[b]	411 ± 19	Rb-Sr whole rock	Lyons and Livingston (1977)
17	Winnipesaukee Pluton[b]	402 ± 19	Rb-Sr whole rock	Lyons and Livingston (1977)

Table A.2 (Cont'd)

Plate III Number	Name	Age (m.y.)	Dating Method	Reference
	Alleghenian-Post Alleghenian/Permian-Cretaceous (286-66 m.y.)			
16	Ossipee Mountains Pluton	121	K-Ar biotite	Foland and Faul (1977)
11	Pilot-Pliny Complex (includes Percy Peaks and Gore Mountain stocks)	Jurassic	Lithostratigraphic correlation	Foland and Faul (1977)
14	White Mountain Batholith (also Belknap Mountain ring-dike complex)	170-195 175-196	Rb-Sr whole rock Rb-Sr whole rock	Creasy (1974) Eby and Creasy (1983)

[a]Reference details are given in Sec. 4.5 or in the references section of the adjoining state.

[b]Intrusive bodies related to the Kinsman quartz monzonite.

Table A.3 Ages of Crystalline Rock Bodies of Vermont

Plate IV Number	Name	Age (m.y.)	Dating Method	Reference[a]
	Grenville/Middle Proterozoic (1600-1000 m.y.)			
3	Chester Dome	~1000	Lithostratigraphic correlation	Faul et al. (1963); Rosenfeld (1954 and 1968)
5	Mt. Holly Complex, Northern Exposure	~1000	Lithostratigraphic correlation	Faul et al. (1963)
2	Mt. Holly Complex, Southern Exposure	~1000	Lithostratigraphic correlation	Faul et al. (1963)
1	Sadawga and Ray Pond Domes	~1000	Lithostratigraphic correlation	Skehan (1961)
	Taconic/Cambrian-Silurian (570-408 m.y.)			
4	Barnard Gneiss	Middle Ordovician	Lithostratigraphic correlation	Ern (1963)
8	Lost Nation Group[b]	444-451	U-Th-Pb zircon	Lyons (1981)
	Acadian-Post-Acadian/Devonian-Pennsylvanian (408-286 m.y.)			
13	Averill Pluton	Mississippian	Lithostratigraphic correlation	Billings (1956)
12	Echo Pond Pluton	Devonian	Lithostratigraphic correlation	Doll et al. (1961)

Table A.3 (Cont'd)

Plate IV Number	Name	Age (m.y.)	Dating Method	Reference
	Acadian-Post-Acadian/Devonian-Pennsylvanian (408-286 m.y.)			
6	Knox Mountain Pluton	380	Rb-Sr whole rock	Naylor (1971)
		330 ± 25	Rb-Sr whole rock	Murthy (1957)
9	Maidstone Pluton	Devonian	Lithostratigraphic correlation	Doll et al. (1961)
11	Nulhegan Pluton	Devonian	Lithostratigraphic correlation	Doll et al. (1961)
7	Victory Pluton	Devonian	Lithostratigraphic correlation	Doll et al. (1961)
10	Willoughby Pluton	Devonian	Lithostratigraphic correlation	Doll et al. (1961)

[a]Reference details are given in Sec. 5.4 or in the references section of the adjoining state.

[b]Equivalent to the Highlandcroft Series.

Table A.4 Ages of Crystalline Rock Bodies of Northeastern New York

Plate V Number	Name	Age (m.y.)	Dating Method	Reference[a]
	Grenville/Middle Proterozoic (1600-1000 m.y.)			
6	Alexandria Granite Gneisses	1220	U-Pb zircon	Silver (1965)
3	Diana Gneiss and Tupper/Saranac Metasyenite Complexes	1100-1200	U-Pb zircon	Silver (1965)
7	Granite Gneiss	Middle Proterozoic	Lithostratigraphic correlation	Silver (1969)
5	Hornblende Granite and Hermon Granite Gneisses	Middle Proterozoic	Lithostratigraphic correlation	Buddington and Leonard (1962)
1	Main Metanorthosite Massif and the Snowy Mountain, Thirteenth Lake, and Oregon Anorthosite Domes	1130	U-Pb zircon	Silver (1969)
2	Olivine Metagabbros	Middle Proterozoic	Lithostratigraphic correlation	Buddington (1939)
4	Stark Metasyenite Complex	1055 ± 30	Rb-Sr whole rock	Heath and Fairbairn (1968)

[a]Reference details are given in Sec. 6.2.

Table A.5 Ages of Crystalline Rock Bodies of Massachusetts

Plate VI Number	Name	Age (m.y.)	Dating Method	Reference[a]
		Grenville/Middle Proterozoic (1600-1000 m.y.)		
1	Berkshire Massif	1000-1080	U-Pb zircon	Ratcliffe and Zartman (1971); Ratcliffe and Zartman (1976)
2	Mt. Holly Complex, Southern Exposure	Middle Proterozoic	Lithostratigraphic correlation	Faul et al. (1963)
		Avalonian/Late Proterozoic (1000-570 m.y.)		
19	Dedham Granodiorite	608 ± 17	Rb-Sr whole rock	Fairbairn et al. (1967)
11	Massabesic Gneiss	646	U-Th-Pb zircon	Aleinikoff et al. (1979)
17	Milford Granite	600-650	U-Th-Pb zircon	Zartman and Naylor (1972); Barosh et al. (1977)
		630	Pb^{207}/Pb^{206} zircon	Naylor (1975)
7	Pelham Dome	565 ± 30	U-Pb zircon	Robinson (1967)
18	Salem Gabbro-Diorite	Late Proterozoic	Lithostratigraphic relations	Zen (1983)
		640	K-Ar hornblende	Zartman and Naylor (1972)
		460 ± 12	Rb-Sr whole rock	Zartman and Marvin (1971)
		414 ± 12	K-Ar biotite	Zartman and Marvin (1971)
16	Sterling and Ponaganset Gneisses	Late Proterozoic	Lithostratigraphic correlation	Zen (1983); Hermes et al. (1981)

Table A.5 (Cont'd)

Plate II Number	Name	Age (m.y.)	Dating Method	Reference
		Taconic/Cambrian-Silurian (570-408 m.y.)		
15	Andover Granite	450 ± 22	Rb-Sr whole rock	Handford (1965)
12	Ayer Intrusive Complex	Silurian or Ordovician	Lithostratigraphic correlation	Zen (1983)
		435 ± 10	U-Th-Pb zircon	Peck (1976)
3	Barnard Gneiss	Middle Ordovician	Lithostratigraphic correlation	Ern (1961)
20	Cape Ann Intrusive Complex	452 ± 10	Pb-Pb zircon	Zartman and Marvin (1971)
5	Glastonbury Gneiss	Ordovician	Lithostratigraphic correlation	Rodgers (1982)
		350	Rb-Sr whole rock	Brookins and Methot (1971)
8	Monson Gneiss	472 ± 15	Rb-Sr whole rock	Brookins and Hurley (1965)
13	Newburyport Granodiorite	Silurian or Ordovician	Lithostratigraphic correlation	Zen (1983)
		321 ± 10	K-Ar biotite	Zartman et al. (1970)
14	Sharpners Pond Diorite	Silurian or Ordovician	Lithostratigraphic correlation	Zen (1981)
		Devonian or older	K-Ar biotite	Zartman et al. (1970)

Table A.5 (Cont'd)

Plate VI Number	Name	Age (m.y.)	Dating Method	Reference
	Acadian-Post-Acadian/Devonian-Pennsylvanian (408-286 m.y.)			
6	Belchertown Intrusive Complex	380 ± 5	U-Pb zircon	Ashwal et al. (1979)
10	Fitchburg Intrusive Complex	382-412	U-Pb zircon	Maczuga (1981)
9	Hardwick-Coys Hill Granite	393	Lithostratigraphic correlation	Lyons (1982)
4	Williamsburg Granodiorite	Devonian	Lithostratigraphic correlation	Norton (1967)

[a]Reference details are given in Sec. 7.4 or in the references section of the adjoining state.

Table A.6 Ages of Crystalline Rock Bodies of Rhode Island

Plate VII Number	Name	Age (m.y.)	Dating Method	Reference[a]
	Avalonian/Late Proterozoic (1000-570 m.y.)			
6	Bulgarmarsh Granite	Late Proterozoic	U-Pb zircon	Zartman and Naylor (in press)
		534 ± 13	Rb-Sr whole rock	Galloway (1973)
5	Esmond Granite	620	U-Pb zircon	Hermes et al. (1981)
3	Sterling Pluton Group (includes Hope Valley alaskite and Ponaganset gneisses)	Late Proterozoic	Lithostratigraphic correlation	Hermes et al. (1981)
2	Ten Rod Granite Gneiss	638 ± 42	Rb-Sr whole rock	Day (1968)
	Acadian-Post-Acadian/Devonian-Pennsylvanian (408-286 m.y.)			
4	Scituate Granite	370	U-Pb zircon	Hermes (1982)
	Alleghenian-Post Alleghenian/Permian-Cretaceous (286-66 m.y.)			
1	Narragansett Pier Granite	273	U-Pb monazite	Kocis et al. (1978)

[a]Reference details are given in Sec. 8.5 or in the references section of the adjoining state.

Table A.7 Ages of Crystalline Rock Bodies of Connecticut

Plate VIII Number	Name	Age (m.y.)	Dating Method	Reference[a]
		Grenville/Middle Proterozoic (1600-1000 m.y.)		
6	Bear Hill Massif	Middle Proterozoic	Lithostratigraphic correlation with Fordham gneiss	Grauert and Hall (1973)
		575 ± 30	Rb-Sr whole rock	Long (1969)
		355	K-Ar biotite	Dallmeyer and Sutter (1976)
8	Berkshire Massif	1040-1080	U-Pb zircon	Ratcliffe and Zartman (1976)
7	Housatonic Massif	Middle Proterozoic	Lithostratigraphic correlation with Fordham or Berkshire gneiss	Grauert and Hall (1973); Rodgers (1982)
5	Manhattan Prong	Middle to Late Proterozoic	Rb-Sr whole rock; lithostratigraphic correlation	Rodgers (1982)
		Avalonian/Late Proterozoic (1000-570 m.y.)		
13	Sterling Pluton Group	Late Proterozoic	Lithostratigraphic correlation	Hermes et al. (1981)
12	Waterford Gneiss Group	Late Proterozoic	Lithostratigaphic correlation	Rodgers (1982)

Table A.7 (Cont'd)

Plate VIII Number	Name	Age (m.y.)	Dating Method	Reference
		Taconic/Cambrian-Silurian (570-408 m.y.)		
15	Ayer Intrusive Complex	435 ± 10	U-Th-Pb zircon	Peck (1976)
4	Brookfield Gneiss	Ordovician	Lithostratigraphic correlation	Rodgers (1982)
10	Glastonbury Gneiss	Ordovician	Lithostratigraphic correlation	Rodgers (1982)
		>350	Rb-Sr whole rock	Brookins and Methot (1971)
3	Granitic Gneisses (includes Mine Hill, Tyler Lake, Ansonia, and Siscowit gneisses)	Ordovician	Lithostratigraphic correlation	Pease (1982); Rodgers (1982)
1	Harrison Gneiss	Middle Ordovician	Lithostratigraphic correlation	Hall (1976); Gates and Martin (1976); Rodgers (1982)
		360	K-Ar biotite	Clark and Kulp (1968)
11	Monson (Haddam) Gneiss	428-450	U-Pb zircon	Leo et al. (in press)
		472 ± 15	Rb-Sr whole rock	Brookins and Hurley (1965)
2	Norwalk Area Gneisses	Ordovician	Lithostratigraphic correlation	Hall (1971)

Table A.7 (Cont'd)

Plate VIII Number	Name	Age (m.y.)	Dating Method	Reference
	Acadian-Post-Acadian/Devonian-Pennsylvanian (408-286 m.y.)			
14	Canterbury Gneiss	329 ± 9	Rb-Sr whole rock	Pease (1982)
9	Nonewaug Granite	382 ± 64	Rb-Sr whole rock	Besancon (1970)
	Alleghenian-Post Alleghenian/Permian-Cretaceous (286-66 m.y.)			
16	Narragansett Pier Granite	273	U-Pb monazite	Kocis et al. (1978)

[a]Reference details are given in Sec. 9.5 or in the references section of the adjoining state.

Table A.8 Ages of Crystalline Rock Bodies of Southeastern New York, Northern New Jersey, and Southeastern Pennsylvania

Plate IX Number	Name	Age (m.y.)	Dating Method	Reference[a]
		Grenville/Middle Proterozoic (1600-1000 m.y.)		
2	Baltimore Gneiss	~1100 >1100	U-Pb zircon Lithologic characteristics	Lapham and Root (1971) Wagner and Crawford (1975)
1	Honey Brook Complex	980-1100	U-Pb zircon	Tilton et al. (1960); Grauert et al. (1973)
6	Housatonic Massif	980-1080	Lithostratigraphic correlation	Grauert and Hall (1973)
7	Manhattan Prong	~1000	Rb-Sr whole rock	Long (1969)
4	Reading Prong, Northwestern Massif	1100-1150		Lapham and Root (1971)
5	Reading Prong, Southeastern Massif	1100-1150		Lapham and Root (1971)
3	Reading Prong, Western Extension	1100-1150		Lapham and Root (1971)
		Taconic/Cambrian-Silurian (570-408 m.y.)		
10	Bedford Gneiss	~467 Middle Ordovician	K-Ar and U-Pb Lithostratigraphic correlation	Long and Kulp (1962) Gates and Martin (1976); Hall (1976); Rodgers (1982)

Table A.8 (Cont'd)

Plate IX Number	Name	Age (m.y.)	Dating Method	Reference
		Taconic/Cambrian-Silurian (570-408 m.y.) (Cont'd)		
8	Granitic Gneiss	Middle Ordovician	Lithostratigraphic correlation	Rodgers (1982); Pease (1982)
11	Harrison Gneiss	Middle Ordovician	Lithostratigraphic correlation	Gates and Martin (1976); Hall (1976); Rodgers (1982)
9	Norwalk Area Gneiss	Ordovician	Lithostratigraphic correlation	Hall (1971)

[a]Reference details are given in Sec. 10.3 or in the references section of the adjoining state.

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