Guidebook for Field Trips in Western Maine and Northern New Hampshire

Beverly Johnson
Bates College

J. Dykstra Eusden
Bates College, deusden@bates.edu

Follow this and additional works at: http://scarab.bates.edu/neigc2017

Part of the Geology Commons

Recommended Citation

This Event is brought to you for free and open access by the Conferences and Events at SCARAB. It has been accepted for inclusion in New England Intercollegiate Geological Conference 2017 by an authorized administrator of SCARAB. For more information, please contact batesscarab@bates.edu.
Guidebook for Field Trips in Western Maine and Northern New Hampshire

New England Intercollegiate Geological Conference
109th Annual Meeting

September 29 - October 1, 2017

Edited by Beverly Johnson and Dykstra Eusden

Hosted by Bates College Department of Geology
Maine Mineral and Gem Museum
NEW ENGLAND INTERCOLLEGIATE GEOLOGICAL CONFERENCE
109th Annual Meeting

Guidebook for Field Trips in Western Maine and Northern New Hampshire

Edited by
Beverly Johnson and J. Dykstra Eusden Jr.

Hosted by
The Bates College Department of Geology and The Maine Mineral and Gem Museum

September 29 - October 1, 2017
Copies of this guidebook may be purchased for $25 from the following address:
Chair, Geology Department
Carnegie Science
44 Campus Ave.
Bates College
Lewiston, ME 04240

**Front and Back Cover Credit**

The photo is from Rumford Whitecap more or less in the middle of the region covered by this guidebook and itself part of the subject of trip B5 (Spigel et al, this volume; photo credit Eusden).
The back cover is a collage of author-submitted representative images for the trips that define this year's NEIGC conference. The NEIGC logo is courtesy of Lindley Hanson.

The field guides for this Conference are offered under the terms of the Creative Commons Attribution-Non-Commercial 3.0 Unported

All field trips are available for free download through the Bates College Repository SCARAB (scarab.bates.edu/neigc2017).

**DISCLAIMER**

**Before visiting any of the sites described in New England Intercollegiate Geological Conference guidebooks, you must obtain permission from the current landowners.**

Landowners only granted permission to visit these sites to the organizers of the original trips for the designated dates of the conference. It is your responsibility to obtain permission for your visit. Be aware that this permission may not be granted.

Especially when using older NEIGC guidebooks, note that locations may have changed drastically. Likewise, geological interpretations may differ from current understandings.

Please respect any trip stops designated as "no hammers", "no collecting" or the like.

Consider possible hazards and use appropriate caution and safety equipment.

NEIGC and the hosts of these online guidebooks are not responsible for the use or misuse of the guidebooks.
# TABLE OF CONTENTS

Welcome and Foreword ........................................................................................................... iv

Dedication to John W. Creasy .................................................................................................. v

List of Previous Meetings of the NEIGC ................................................................................ vi


A4. Possible Post-Laurentide Cirque Glaciation in the Great Gulf, Presidential Range, New Hampshire – Brian Fowler and Ian Dulin ........................................................................................................... 81

A5. Grafton Notch State Park: Glacial Gorges and Streams Under Pressure in the Mahoosic Range, Me – Alice M. Doughty and Woodrow B. Thompson ................................................................. 95


B1. Geology of the Bald Mountain-Saddleback Wind Range, Me – Doug Reusch and Jake Hansen .......................................................................................................................................................... 119


B3. Paleo-Dunes and Other Post-Glacial Oddities in the Woods and Fields of New Sharon and Chesterville, Me – Patricia Millette, Benjamin Andrews, Anna Glass, Thaddeus Gunter and Roshan Luick .......................................................................................................................... 161


B5. Geology of the Lower Ellis River Valley and Rumford Whitecap Mountain, Andover and Rumford, Maine – Lindsay Spigel, Amber Whittaker, and Ryan Gordon .............................. 197

B6. Devonian Granite Melt Transfer in Western Maine: Relations Between Deformation, Metamorphism, Melting and Pluton Emplacement at the Migmatite Front Gary S. Solar, Paul B. Tomascak, and Michael Brown .............................................................................................................................. 217


C2. Field Relations, Petrography and Provenance Of Mafic Dikes, Western Maine – David Gibson, Donald Osthoff and Chase Rerrick ........................................................................................................... 273

C3. Transect from the Migmatized Central Maine Belt to the Bronson Hill Anticlinorium – J. Dykstra Eusden, Sarah Baker, Jordan Cargill, Erik Divan, Ian Hillenbrand, Paul O'Sullivan, and Audrey Wheatcroft ...................................................................................................................... 287


C5. The Sandy River Revisited – Julia Daly, Tom Eastler and Daniel Locke .................................................................................................................................................................................. 317

WELCOME

Greetings and welcome to the 109th New England Intercollegiate Geological Conference! The "footprint" of the field trips is the foothills and mountains of western Maine and the adjacent White Mountains of New Hampshire. Think of an area including Bethel, Farmington, Rumford, and Rangeley, Maine, and Gorham, Lancaster, and Berlin, New Hampshire. Your host this year is Department of Geology, Bates College with special help from the Maine Mineral and Gem Museum and Gould Academy.

With a highpoint of 6,288 ft. on Mt. Washington, NH and a low elevation of 289 ft. in New Sharon, ME, the terrain of the region covered by NEIGC 2017 is as varied as its geology. The meeting this year has a complete geologic spectrum of fieldtrips and we give our sincere thanks to all the leaders for working so hard and enthusiastically on their excursions. It is very time consuming to prepare and organize these trips and to then write the manuscripts. For that we are grateful to all the authors. Please thank your fieldtrip leaders for their efforts!

There are great bedrock trips (A1, A2, A6, B1, B4, B6, C2, C3, and C6) examining the complexly deformed and metamorphosed Paleozoic stratigraphy and the wide variety of cross cutting felsic to intermediate plutons, abundant pegmatites, and mafic dikes. There are wonderful glacial trips (A4, B2, B5, and C1) exploring aspects of the glaciation and deglaciation of the Laurentide ice sheet. There are excellent trips for those interested in post-glacial landscape evolution and the modern river systems (A5, B3, C5). Finally, there is a unique hydrogeology trip to a site in Berlin (A3) and a geo-archaeology trip to Paleo-indian sites in New Hampshire (C4).

FOREWORD

We would like to thank the many people who have made this NEIGC possible. Brenda Pelletier (Bates Conferences and Campus Events) and Genevieve Robert (Bates Geology) designed the on-line registration form; a first for NEIGC! Krystie Wilfong (Bates Associate College Librarian) developed the web site and uploaded the entire guidebook as well as individual fieldtrips that can be freely downloaded. Thanks to Michelle Holbrook-Pronovost (Academic Technology Consultant) for her expert assistance in editing the final manuscripts and assembling the guidebook. The staff of Bates Post and Print did a wonderful job printing and binding this full color guidebook, another first for NEIGC. Barbra Barrett and staff of the Maine Mineral and Gem Museum kindly offered to host the Friday night reception at the Museum. Beth McWilliams and George Carone of Gould Academy have done excellent work on the logistics of the Saturday night banquet at Ordway Hall. Thanks to Lindley Hanson of Salem State for updating and maintaining the official NEIGC web site. Joe Kopera's efforts as NEIGC Secretary to develop the NEIGC Facebook page are much appreciated. Marita Bryant and Mike Retelle (Bates Geology) have been very helpful with the overall logistics of meeting planning and registration, and in keeping us both relatively sane! We dedicate this year's NEIGC to our colleague, mentor, and friend John W. Creasy.

Beverly Johnson

Dykstra Eusden

Bates College, Department of Geology
This 109th meeting of the New England Intercollegiate Geological Conference is dedicated to

John W. Creasy
1945 - 2017

Professor of Geology, Bates College

John was an excellent mentor to all of us younger faculty and staff in the Geology Department, giving needed advice but also letting us discover on our own how to navigate the waters of being a researcher and teacher. With his trademark great dry wit, steady and friendly personality, intense focus on always improving and tweaking his classes, and purposeful leadership as department chair, we all grew to rely on him.

John's research added to the knowledge of the geologic history of the Mesozoic White Mountain Magma Series including his fieldwork in the Crawford Notch and Conway region of New Hampshire. His work on the Moat Volcanics was critical in fostering our understanding of these rare volcanic rocks and their place in the evolution of calderas and caldera collapse. John mapped many quadrangles in Maine and always involved scores of Bates undergraduates, teaching them the ways of bedrock field geology.

John's signature course was a five-week Short Term trek in May to map the geology of the U.S. Southwest. John and students explored the Four Corners region starting along the Rio Grande near the Texas/New Mexico/Mexico border mapping mantle rocks at Kilborne Hole, then moving to the stunningly beautiful Chiricahua National Monument in SE Arizona to study the caldera and its balanced rhyolite rocks, and then on to Sunset Crater just north of Flagstaff to map cinder cones and lava flows. The mid point of the trip is a 3-day overnight to the Grand Canyon where the magnificent geology is so well displayed. The trip continued with a visit to Mesa Verde National park and a project on the stratigraphy of Durango, Colorado, and ended with a project on a massive anticlinal fold in San Ysidro New Mexico before heading back to Albuquerque. Many students over the years cited this trip as a pivotal point in helping them find their love for geology.

We will miss him and his sage advice. He'd love to be leading a trip in this year's NEIGC!
<table>
<thead>
<tr>
<th>Year</th>
<th>Location</th>
<th>Organizer(s)</th>
</tr>
</thead>
<tbody>
<tr>
<td>2017</td>
<td>Bethel, ME</td>
<td>Dykstra Eusden and Beverly Johnson</td>
</tr>
<tr>
<td>2016</td>
<td>Bath, ME</td>
<td>Henry Berry and David West</td>
</tr>
<tr>
<td>2015</td>
<td>Middletown, CT</td>
<td>Martha Gilmore</td>
</tr>
<tr>
<td>2014</td>
<td>Wellesley, MA</td>
<td>Margaret Thompson and David Hawkins</td>
</tr>
<tr>
<td>2013</td>
<td>Millinocket Lake, ME</td>
<td>Lindley Hanson</td>
</tr>
<tr>
<td>2012</td>
<td>Sunapee, NH</td>
<td>Peter Thompson and Timothy Allen</td>
</tr>
<tr>
<td>2011</td>
<td>Middlebury, VT</td>
<td>David West</td>
</tr>
<tr>
<td>2010</td>
<td>Orono, ME</td>
<td>Martin Yates, Chris Gerbi, Alice Kelley, Dan Lux</td>
</tr>
<tr>
<td>2009</td>
<td>Lyndon, VT</td>
<td>David Westerman and Alison Lathrop</td>
</tr>
<tr>
<td>2008</td>
<td>Westfield, MA</td>
<td>Mark Van Baalen and Mike Young</td>
</tr>
<tr>
<td>2007</td>
<td>Quebec City, PQ</td>
<td>Louise Corriveau, Tom Clark and Michel Malo</td>
</tr>
<tr>
<td>2006</td>
<td>Rangeley, ME</td>
<td>David Gibson, Julia Daly and Douglas Reusch</td>
</tr>
<tr>
<td>2005</td>
<td>New Haven, CT</td>
<td>Brian Skinner and Tony Philpotts</td>
</tr>
<tr>
<td>2004</td>
<td>Salem, MA</td>
<td>Lindley Hanson</td>
</tr>
<tr>
<td>2003</td>
<td>Amherst and Northampton, MA</td>
<td>John Brady and Jack Cheney</td>
</tr>
<tr>
<td>2002</td>
<td>Lake George, NY</td>
<td>James McLelland and Paul Karabinos</td>
</tr>
<tr>
<td>2001</td>
<td>Fredericton, NB</td>
<td>Dave Lentz and Ron Pickerill</td>
</tr>
<tr>
<td>2000</td>
<td>Orono, ME</td>
<td>Martin Yates, Daniel Lux and Joseph Kelley</td>
</tr>
<tr>
<td>1999</td>
<td>Burlington, VT</td>
<td>Barry Doolan</td>
</tr>
<tr>
<td>1998</td>
<td>Kingston, RI</td>
<td>Dan Murray</td>
</tr>
<tr>
<td>1997</td>
<td>Killington-Pico, VT</td>
<td>Timothy Grover and Helen Mango</td>
</tr>
<tr>
<td>1996</td>
<td>Mount Washington, NH</td>
<td>Mark Van Baalen</td>
</tr>
<tr>
<td>1995</td>
<td>Brunswick, ME</td>
<td>Arthur Hussey and Robert Johnston</td>
</tr>
<tr>
<td>1994</td>
<td>Millinocket, ME</td>
<td>Lindley Hanson and Dabney Caldwell</td>
</tr>
<tr>
<td>1993</td>
<td>Boston, MA</td>
<td>National GSA: Jack Cheney and Chris Hepburn</td>
</tr>
<tr>
<td>1992</td>
<td>Amherst, MA</td>
<td>Peter Robinson and John Brady</td>
</tr>
<tr>
<td>1991</td>
<td>Princeton, ME</td>
<td>Allan Ludman</td>
</tr>
<tr>
<td>1990</td>
<td>La Gaspésie, PQ</td>
<td>Walter Trzcienski</td>
</tr>
<tr>
<td>1989</td>
<td>Farmington, ME</td>
<td>Archie Berry</td>
</tr>
<tr>
<td>1988</td>
<td>Keene, NH</td>
<td>Wallace Bothner</td>
</tr>
<tr>
<td>1987</td>
<td>Montpelier, VT</td>
<td>David Westerman</td>
</tr>
<tr>
<td>1986</td>
<td>Lewiston, ME</td>
<td>Donald Newberg</td>
</tr>
<tr>
<td>1985</td>
<td>New Haven, CT</td>
<td>Robert Tracy</td>
</tr>
<tr>
<td>1984</td>
<td>Danvers, MA</td>
<td>Lindley Hanson</td>
</tr>
<tr>
<td>1983</td>
<td>Greenville-Millinocket, ME</td>
<td>Dabney Caldwell and Lindley Hanson</td>
</tr>
<tr>
<td>1982</td>
<td>Storrs, CT</td>
<td>Ray Joesten and Sidney Quarry</td>
</tr>
<tr>
<td>1981</td>
<td>Kingston, RI</td>
<td>Jon Boothroyd and Don Hermes</td>
</tr>
<tr>
<td>1980</td>
<td>Presque Isle, ME</td>
<td>David Roy and Richard Naylor</td>
</tr>
<tr>
<td>1979</td>
<td>Troy, NY</td>
<td>Gerald Friedman</td>
</tr>
<tr>
<td>1978</td>
<td>Calais, ME</td>
<td>Allan Ludman</td>
</tr>
<tr>
<td>1977</td>
<td>Quebec City, PQ</td>
<td>René Béland and Pierre LaSalle</td>
</tr>
<tr>
<td>1976</td>
<td>Boston, MA</td>
<td>Barry Cameron</td>
</tr>
<tr>
<td>1975</td>
<td>Great Barrington, MA</td>
<td>Nicholas Ratcliffe</td>
</tr>
<tr>
<td>1974</td>
<td>Orono, ME</td>
<td>Philip Osberg</td>
</tr>
<tr>
<td>1973</td>
<td>Fredericton, NB</td>
<td>Hugo Grenier and Nick Rast</td>
</tr>
<tr>
<td>1972</td>
<td>Burlington, VT</td>
<td>Barry Doolan and Rolfe Stanley</td>
</tr>
<tr>
<td>1971</td>
<td>Concord, NH</td>
<td>John Lyons and Glenn Stewart</td>
</tr>
<tr>
<td>1970</td>
<td>Rangeley Lakes-Dead River, ME</td>
<td>Gary Boone</td>
</tr>
<tr>
<td>1969</td>
<td>Albany, NY</td>
<td>John Bird</td>
</tr>
<tr>
<td>1968</td>
<td>New Haven, CT</td>
<td>Phil Orville</td>
</tr>
<tr>
<td>1967</td>
<td>Amherst, MA</td>
<td>Peter Robinson, David Drake and Richard Foose</td>
</tr>
<tr>
<td>1966</td>
<td>Katahdin, ME</td>
<td>Dabney Caldwell</td>
</tr>
<tr>
<td>1965</td>
<td>Brunswick, ME</td>
<td>Arthur Hussey</td>
</tr>
<tr>
<td>1964</td>
<td>Chestnut Hill, MA</td>
<td>James Skehan</td>
</tr>
</tbody>
</table>

*Note: individual trip leaders are listed for earlier meetings*
<table>
<thead>
<tr>
<th>Year</th>
<th>Town, State</th>
<th>Names</th>
</tr>
</thead>
<tbody>
<tr>
<td>55 - 1963</td>
<td>Providence, RI</td>
<td>Quinn, Mutch, Shafer, Agron, Chapple, Feiniger &amp; Hall</td>
</tr>
<tr>
<td>54 - 1962</td>
<td>Montreal, PQ</td>
<td>Gill, Clark, Kranck, Stevenson, Stearn, Elson, Eakins, &amp; Gold</td>
</tr>
<tr>
<td>53 - 1961</td>
<td>Montpelier, VT</td>
<td>Doll, Cady, White, Chidester, Matthews, Nichols, Baldwin, Stewart, Dennis, Smith, and Ferrris</td>
</tr>
<tr>
<td>52 - 1960</td>
<td>Rumford, ME</td>
<td>Griscom, Milton, Wolfe, Caldwell, and Peacor</td>
</tr>
<tr>
<td>51 - 1959</td>
<td>Rutland, VT</td>
<td>Zen, Kay, Welby, Bain, Theokritoff, Osberg, Shumaker, Berry, and Thompson</td>
</tr>
<tr>
<td>50 - 1958</td>
<td>Middletown, CT</td>
<td>Rosenfeld, Eaton, Sanders, Porter</td>
</tr>
<tr>
<td>49 - 1957</td>
<td>Amherst, MA</td>
<td>George Bain</td>
</tr>
<tr>
<td>48 - 1956</td>
<td>Portsmouth, NH</td>
<td>Novotny, Billings, Chapman, Bradley, Freedman and Stewart</td>
</tr>
<tr>
<td>47 - 1955</td>
<td>Ticonderoga, NY</td>
<td>Rodgers, Walton, MacClintock, Bartolome</td>
</tr>
<tr>
<td>46 - 1954</td>
<td>Hanover, NH</td>
<td>Elston, Washburn, Lyons, McKinstry, Stoiber, McNair, and Thompson</td>
</tr>
<tr>
<td>45 - 1953</td>
<td>Hartford, CT</td>
<td>Flint, Gates, Peoples, Cushman, Aitken, Rodgers and Troxell</td>
</tr>
<tr>
<td>44 - 1952</td>
<td>Williamstown, MA</td>
<td>Perry, Foote, McFaden, and Ramsdell</td>
</tr>
<tr>
<td>43 - 1951</td>
<td>Worcester, MA</td>
<td>Lougee and Little</td>
</tr>
<tr>
<td>42 - 1950</td>
<td>Bangor, ME</td>
<td>Trefethen and Raisz</td>
</tr>
<tr>
<td>41 - 1949</td>
<td>Boston, MA</td>
<td>Nichols, Billings, Shrock, Currier, and Stearns</td>
</tr>
<tr>
<td>40 - 1948</td>
<td>Burlington, VT</td>
<td>Charlie Doll</td>
</tr>
<tr>
<td>39 - 1947</td>
<td>Providence, RI</td>
<td>Alonzo Quinn</td>
</tr>
<tr>
<td>37 - 1941</td>
<td>Northampton, MA</td>
<td>Balk, Jahns, Lochman, Shaub, and Willard</td>
</tr>
<tr>
<td>36 - 1940</td>
<td>Hanover, NH</td>
<td>J. W. Goldthwait, Denny, Shaub, Hadley, Bannerman, and Stoiber</td>
</tr>
<tr>
<td>35 - 1939</td>
<td>Hartford, CT</td>
<td>Troxell, Flint, Longwell, Peoples and Wheeler</td>
</tr>
<tr>
<td>34 - 1938</td>
<td>Rutland, VT</td>
<td>George W. Bain</td>
</tr>
<tr>
<td>33 - 1937</td>
<td>NYC-Dutchess Co., NY</td>
<td>O'Connell, Kay, Fluh, Hubert and Balk</td>
</tr>
<tr>
<td>32 - 1936</td>
<td>Littleton, NH</td>
<td>Marland P. Billings, Hadley, Cleaves and Williams</td>
</tr>
<tr>
<td>31 - 1935</td>
<td>Boston, MA</td>
<td>Morris, Pearsall, and Whitehead</td>
</tr>
<tr>
<td>30 - 1934</td>
<td>Lewiston, ME</td>
<td>Lloyd Fisher and Edward Perkins</td>
</tr>
<tr>
<td>29 - 1933</td>
<td>Williamstown, MA</td>
<td>Herdman Cleland, Perry, and Knopf</td>
</tr>
<tr>
<td>28 - 1932</td>
<td>Providence-Newport, RI</td>
<td>C. W. Brown</td>
</tr>
<tr>
<td>27 - 1931</td>
<td>Montreal, PQ</td>
<td>O'Neill, Graham, T. M. Clark, Gill, Osborne, and McGerrigle</td>
</tr>
<tr>
<td>26 - 1930</td>
<td>Amherst, MA</td>
<td>F. B. Loomis and Gordon</td>
</tr>
<tr>
<td>25 - 1929</td>
<td>Littleton, NH</td>
<td>Irving B. Crosby</td>
</tr>
<tr>
<td>24 - 1928</td>
<td>Cambridge, MA</td>
<td>Marland P. Billings, Kirk Bryan, and Kirtley Mather</td>
</tr>
<tr>
<td>23 - 1927</td>
<td>Worcester, MA</td>
<td>Perry, Little, and Gordon</td>
</tr>
<tr>
<td>22 - 1926</td>
<td>New Haven, CT</td>
<td>C. R. Longwell</td>
</tr>
<tr>
<td>21 - 1925</td>
<td>Waterville, ME</td>
<td>Edward H. Perkins</td>
</tr>
<tr>
<td>20 - 1924</td>
<td>Providence, RI</td>
<td>C. W. Brown</td>
</tr>
<tr>
<td>19 - 1923</td>
<td>Beverly, MA</td>
<td>A. C. Lane</td>
</tr>
<tr>
<td>18 - 1922</td>
<td>Amherst, MA</td>
<td>Ernst Anteves</td>
</tr>
<tr>
<td>17 - 1921</td>
<td>Attleboro, MA</td>
<td>J. B. Woodworth</td>
</tr>
<tr>
<td>16 - 1920</td>
<td>Hanging Hills, Meriden, CT</td>
<td>W. N. Rice and Wilbur Foye</td>
</tr>
<tr>
<td>15 - 1917</td>
<td>Gay Head, Martha's Vineyard, MA</td>
<td>J. B. Woodworth and E. Wigglesworth</td>
</tr>
<tr>
<td>14 - 1916</td>
<td>Blue Hills, MA</td>
<td>W. O. Crosby and C. H. Warren</td>
</tr>
<tr>
<td>13 - 1915</td>
<td>Waterbury-Winsted, CT</td>
<td>Jos. Barrell</td>
</tr>
<tr>
<td>12 - 1912</td>
<td>Higby-Lamentation Blocks, CT</td>
<td>W. N. Rice</td>
</tr>
<tr>
<td>11 - 1911</td>
<td>Nahant-Medford, MA</td>
<td>A. C. Lane and D. W. Johnson</td>
</tr>
<tr>
<td>10 - 1910</td>
<td>Hanover, NH</td>
<td>James W. Goldthwait</td>
</tr>
<tr>
<td>9 - 1909</td>
<td>Northern Berkshires, MA</td>
<td>H. F. Cleland</td>
</tr>
<tr>
<td>8 - 1908</td>
<td>Long Island, NY</td>
<td>Jos. Barrell</td>
</tr>
<tr>
<td>7 - 1907</td>
<td>Providence, RI</td>
<td>C. W. Brown</td>
</tr>
<tr>
<td>6 - 1906</td>
<td>Meriden-East Berlin, CT</td>
<td>H. E. Gregory</td>
</tr>
<tr>
<td>5 - 1905</td>
<td>Boston-Nantasket, MA</td>
<td>D. W. Johnson and W. O. Crosby</td>
</tr>
<tr>
<td>4 - 1904</td>
<td>Worcester, MA</td>
<td>Benjamin Emerson</td>
</tr>
<tr>
<td>3 - 1903</td>
<td>West Peak, Meriden, CT</td>
<td>W. N. Rice</td>
</tr>
<tr>
<td>2 - 1902</td>
<td>Mount Tom, MA</td>
<td>Benjamin Emerson</td>
</tr>
<tr>
<td>1 - 1901</td>
<td>Westfield River Terraces, MA</td>
<td>William Morris Davis</td>
</tr>
</tbody>
</table>
A1: LITHIUM-BORON-BERYLLIUM GEM PEGMATITES, OXFORD CO., MAINE: HAVEY AND MOUNT MICA PEGMATITES

LEADERS: 1William B. Simmons, 1Alexander U. Falster, 1Karen L. Webber, 1Myles M. Felch and 2Dwight C. Bradley
1MP2 Research Group, Maine Mineral & Gem Museum, 99 Main Street, Bethel, Maine
211 Cold Brook Road, Randolph, New Hampshire 03593

Contact information: alexander.falster@gmail.com (504) 220-5260

INTRODUCTION

This fieldtrip will visit two world renowned gem-producing pegmatites, in the Oxford County pegmatite field of western Maine: Havey and Mount Mica pegmatites. This is intended to be primarily an instructional fieldtrip lead by pegmatite experts from the Maine Mineral and Gem Museum (MMGM) MP2 (Mineralogy, Pegmatology, Petrology) research group. This is an opportunity to learn about the latest research and advances in the field of pegmatology in Maine. Mt. Mica and the Havey pegmatites are enriched in Be, B, and Li and over the last ten years have produced significant finds of gem tourmaline. The Havey produces a very consistent mint green elbaite with lesser amounts of pink and some rare blue gem elbaite. Mt. Mica has produced a huge quantity of gem elbaite in a wide range of colors. Some of the elbaite crystals recently mined from Mt. Mica are exceptionally large and many are fine museum-quality mineral specimens.

This field trip guide is divided into the following four sections:
1. Field trip summary with the leaders identified (A1-1)
2. General geology background information – Falster, Simmons & Bradley (A1-2)

TIME, PLACE, LOGISTICS: Friday September 29th at 8:30 am meet in the back-parking lot of the Maine Mineral and Gem Museum located at 99 Main Street in Bethel, Maine (357595.99 m E, 4918691.40 m N). The field trip will begin with a brief discussion in the parking lot. We encourage carpooling to limit the number of vehicles entering the mines. A convoy of vehicles will travel approximately one hour to the Havey Quarry located off of Levine Road in Poland (395997.39 m E, 4880613.82 m N). Plan to spend approximately 2hrs at this location, starting with a discussion and later an opportunity to collect. Departing the Havey Quarry, the convoy will travel north approximately 40 minutes to the Mt. Mica Quarry located off of Mt. Mica road in Paris (382422.66 m E, 4902813.15 m N). Plan to spend approximately 2 hrs. at this site, which will begin with a discussion and end with an opportunity to collect on the dumps afterwards. There will not be an opportunity to go underground. At 4:00 pm, participants will leave the mine and travel north approximately 40 minutes back to Bethel. We encourage all participants to attend the Friday night reception at the Maine Mineral and Gem Museum. For this field trip, be prepared with a lunch as we will not be stopping along the way. There will be bathroom facilities available at each of the mine sites. Enrollment will be limited to 25 participants. Expect cold and unpredictable weather and dress accordingly. Please note: both quarries we are visiting are not open to the public except via previous arrangement.
GENERAL GEOLOGY OF THE OXFORD COUNTY PEGMATITE FIELD IN MAINE

1Alexander U. Falster, 1William B. Simmons & 2Dwight C. Bradley
1MP2 Research Group, Maine Mineral & Gem Museum, 99 Main Street, Bethel, Maine
211 Cold Brook Road, Randolph, New Hampshire 03593

REGIONAL GEOLOGY

The New England Appalachians formed during a prolonged Wilson Cycle between the breakup of the supercontinent Rodinia in the Neoproterozoic and the staged assembly of the supercontinent Pangea in the Phanerozoic. The pre-Silurian rocks mainly fall into two groups: those of Laurentian or peri-Laurentian affinity, toward the west, and those of peri-Gondwanan affinity, toward the east (Hibbard et al., 2006). These rocks were juxtaposed via mainly convergent plate motions, deformed, metamorphosed, and intruded by granitic plutons during a succession of Paleozoic orogenies, the main ones being the Taconic (Ordovician), Acadian (Late Silurian-Early Devonian), Neoacadian (Late Devonian-Early Mississippian), and Alleghanian (Pennsylvanian-Permian) (Robinson et al., 1998). By the end of the Alleghanian orogeny, the Appalachian chain stretched thousands of kilometers across the interior of the now-assembled Pangea supercontinent. Supercontinent breakup began soon after, with Triassic to Early Jurassic within-plate magmatism and rifting leading, eventually, to the opening of the Central Atlantic from Early Jurassic to present.

The Oxford pegmatites of western Maine are situated in the Central Maine Belt (Figure 1) in a belt of metasedimentary rocks (Figure 2.) that were deposited in the late Ordovician, Silurian, and earliest Devonian. These rocks were deposited in a deep-water basin, the Central Maine Basin, immediately before and during the Acadian orogeny (Bradley et al., 2000). Deformation and metamorphism took place during the Acadian, Neoacadian, and Alleghanian orogenies. Eusden et al. (2017, this volume) present a summary of the geologic history.

Plutons were emplaced during multiple pulses spanning the late Neoproterozoic to the Cretaceous (Bradley et al., 2015). The Oxford pegmatite district includes plutons related to four igneous episodes. Examples include the Songo pluton (Neoacadian granodiorite, 364±1.3 Ma, Gibson et al., 2017), the Sebago pluton sensu lato (Alleghanian granite, 288 ± 13 to 297 ± 14 Ma, Solar and Tomascak, 2016), the Whale’s Back pluton (a White Mountain Series granite, 184±3 Ma, Foland and Faul 1977), and the Pleasant Mountain pluton (Cretaceous, 112±3 Ma, Foland and Faul, 1977).
Figure 1. General geology of New England (from Brown & Solar, 1998).
Figure 2: Metamorphic facies in the Oxford pegmatite field and the locations of pegmatites. Modified from Roda-Robles (2011), after Guidotti et al. (1989).

LOCAL GEOLOGY

The Oxford pegmatite district largely coincides with the area formerly mapped as the Sebago batholith (Osberg et al., 1985). The mapped extent of the Sebago batholith has gotten much smaller in recent years. This tract has now been subdivided into the Sebago pluton and the Sebago Migmatite Domain (SMD) (Solar and Tomascak, 2009), (Figure 3). As noted above, the Sebago pluton has been dated at 288 ± 13 to 297 ± 14 Ma, Solar and Tomascak, 2016), whereas the migmatites have been dated at 376 ± 14 Ma (Solar and Tomascak, 2016). The metasedimentary, Central Maine Basin country rocks have been metamorphosed to middle- to upper-amphibolite facies (Tomascak et al., 1996).
Figure 3. Map showing Oxford pegmatite field in the Sebago Migmatite Domain and the location of the Havey and Mt. Mica Pegmatites. Modified from Solar & Tomascak (2009).

OXFORD FIELD PEGMATITES

Overall, pegmatite bodies in the SMD are concordant with the foliation of the host rocks, although some display irregular and locally discordant contacts. Generally, pegmatites in the SMD exhibit internal mineralogical and textural zonation. The outermost wall-zone is typically less than 1m thick, with a homogeneous pegmatic texture. K-feldspar, albite, quartz, biotite, muscovite, garnet ± schorl are the most common minerals in this zone, except in Mt. Mica which has only albite in the wall zone with microcline appearing only in the core zone. Many pegmatites exhibit a comb structure of wedge-shaped schorl crystals, some up to 50 cm in length, that grow directly from the contact with the country rock in the hanging wall, inside a homogeneous pegmatite matrix (particularly well developed in the Emmons pegmatite). The wall-zone is gradational into the intermediate zone of graphically intergrown feldspar and quartz (graphic granite), with accessory garnet, biotite, muscovite and black tourmaline. In some pegmatites, the intermediate zone is asymmetric and thicker under the core zone, e.g. at the Havey pegmatite. The core-zones consist of meter-sized masses of blocky K-feldspar and/or quartz with a more evolved assemblage of finer-grained, irregular pods consisting of albite, muscovite, lepidolite, Li-tourmaline, Li-Al and Fe-Mn-phosphates, Sn-, Nb- and Ta-oxides, beryl and Cs-beryl, spodumene, petalite and pollucite. Within the core-zone pods, pockets are relatively common. The pockets contain gemmy elbaite associated with quartz, lepidolite, albite, cassiterite and clay minerals. Another notable feature in the foot wall of Oxford Field pegmatites is the existence of a garnet layer or line under the core zone roughly paralleling the contact of the pegmatite. The garnet line helps the miners to determine the limits of the core and pockets, as pockets never occur below this layer. Prismatic schorl crystals intergrown with or close to the garnet line are oriented perpendicular to the contact and indicate the sense of upward crystal growth. In some cases, a second line of schorl occurs below the garnet line. Overall, the composition of Oxford Field pegmatites is not significantly enriched in rare-elements, as these are concentrated only in the
innermost parts of the bodies and constitute a low percentage of the volume of the pegmatite. Textural and mineralogical criteria indicate that the pegmatites crystallized from the borders inward. Fractionation processes were highly effective during the crystallization of these pegmatites. In general, the composition of the primary minerals changes progressively from the contact to the core zone where compositions change more sharply. The Li/(Fe+Mg) ratio of tourmaline increases, the K/Rb of micas and K-feldspar and the Fe/(Fe+Mn) of phosphates decreases, Li and F in mica and Cs in beryl increases as well as a general increase in the proportion of Li-, F- and/or Cs-bearing minerals, parallel to a decrease in Fe-Mn-Mg-bearing phases (Roda-Robles et al. 2015, Wise and Brown 2010).

Pegmatites of the Oxford field are enriched in Be, B, Li and Cs, with low Ta. Wise and Francis (1992) and Wise (1995) classified them as LCT type (Černý & Ercit 2005) pegmatites, but pegmatite classification is currently being revised and for this paper we use the actual elements for the classification. The inferred relationship between the Oxford field pegmatites and surrounding rocks has evolved over the years. The parent bodies of Oxford field pegmatites were once thought to be the Sebago batholith in the south and some smaller plutons in the north, such as the Rumford, Phillips and Mooselookmeguntic (Wise and Francis, 1992 & Wise, 1995). As the Sebago Pluton is now about 30 km distant from Oxford pegmatite field and is about 30 Ma older than the pegmatites, we no longer consider it a viable source for the origin of the pegmatic melts. Our MP2 group has proposed an anatectic origin for some of the Oxford pegmatites (Simmons et al., 1995, 1996, 2016), Roda-Robles (2015).

A recent geochronology study of Appalachian rare-element pegmatites included a few in western Maine. There are many single-crystal U-Pb zircon ages from these rocks because of a combination of abundant xenocrystic grains, severe lattice damage in high-uranium zircons, and common-lead-bearing inclusions. The tightest age control is from the Irish Pit at Mt. Mica. Its age is bracketed between 263.89 +0.18/-0.35 Ma (U-Pb zircon, CA-TIMS) and 259.9±2.3 Ma (40Ar/39Ar muscovite, plateau age), ranking it one of the youngest igneous rocks in New England before the onset of igneous activity related to breakup of Pangea. A pegmatite sample from the Main Pit at Mt. Mica yielded only xenocrystic zircons but did give a 40Ar/39Ar muscovite plateau age of 253.5±2.2 Ma. The Emmons pegmatite yielded a 40Ar/39Ar muscovite plateau age of 250.4±2.3 Ma. The Lord Hill pegmatite yielded a 40Ar/39Ar muscovite plateau age of 231.8±3.0 Ma. Farther north, the Black Mountain pegmatite, near Rumford, yielded a 40Ar/39Ar muscovite plateau age of 253.5±2.2 Ma. The Lord Hill pegmatite is therefore significantly older than the Oxford pegmatites. All of the dated pegmatites are of LCT (lithium-cesium-tantalum) type, except Lord Hill, which is a hybrid LCT-NYF (niobium-yttrium-fluorine) type.

REFERENCES


MOUNT MICA PEGMATITE, PARIS, OXFORD COUNTY, MAINE

William B. (Skip) Simmons¹, Alexander U. Falster¹, Karen Webber¹, Myles M. Felch¹ and Gary Freeman ²
¹MP² Research Group, Maine Mineral & Gem Museum, 99 Main Street, Bethel, Maine
²48 Lovejoy Road, Paris, ME

ABSTRACT

Mt. Mica pegmatite is famous for gem tourmaline production for nearly 200 years. The dike, ranging in thickness from 1 to 8 meters and dipping 20° SE, has a simple zonal structure consisting of a wall zone and core zone. The wall zone is essentially devoid of K-feldspar. The outer portion of the pegmatite consists of quartz, muscovite, albite (An 1.8) and schorl. Muscovite is the dominant K-bearing species in the outer portion of the pegmatite. K-feldspar only appears in the core zone adjacent to pockets. The pegmatite is subparallel to the foliation of the enclosing migmatite, and leucosomes show a gradational contact with the pegmatite where juxtaposed. Texturally, the pegmatite and leucosomes appear to be in equilibrium with no change in grain size or composition where the two are in contact. Garnet-biotite thermometry of the migmatite at the contact yields an average temperature of 630°C, which is consistent with the P-T conditions inferred for a Sebago Migmatite Domain (SMD) assemblage of sillimanite, quartz, muscovite, biotite and alkali feldspar of 650°C and 3 kb. Gradational contact between leucosomes and pegmatite suggests that the pegmatic melt was at the same temperature. Coromoto Minerals began mining in 2003 and the mine now extends down dip for over 100 meters to a depth of 33 meters. A very detailed and accurately surveyed geologic map produced by owner/operator Gary Freeman during mining shows the total area of pegmatite removed, the spatial distribution and aerial extent of pockets, massive lepidolite (compositions near trilithionite) pods, microcline, and xenoliths. The map was analyzed using image analysis and thickness values of the units to calculate the total volumes of pegmatite mined, lepidolite pods and all pockets found. Forty-five drill cores were taken across the pegmatite from the hanging wall to foot wall contacts along a transect intentionally avoiding lepidolite pods and miaroles. Cores were pulverized, thoroughly mixed and homogenized and the percent Li content calculated from the mapped volume was added to produce a sample that was representative of the bulk composition of Mt. Mica. The sample was then analyzed by fusion ICP spectroscopy for major and trace elements and DCP spectroscopy for B and Li. Structural water was determined by LOI. Water content was calculated using the calculated volume of open space (pocket volumes), assuming that the pockets were filled with water-rich fluid. This fluid content was added to LOI water (above 500°C) to estimate a maximum H2O content of 1.16 wt. % of the pegmatite melt. REE plots of bulk pegmatite vs. leucosomes from the migmatite are strikingly similar. Chondrite normalized REE patterns of leucosomes and pegmatite are very flat with no Eu anomaly, whereas Sebago granite is more strongly LREE-enriched and displays a pronounced negative Eu-anomaly. Spider diagrams of leucosomes and pegmatite vs. average crust show very similar patterns. These results suggest that the Mt. Mica pegmatic melt did not form by fractional crystallization of the older Sebago pluton, but instead was derived directly from partial melting of the metapelitic rocks of the SMD. Batches of anatectic melt accumulated and coalesced into a larger volume that subsequently formed the pegmatite. This is the first chemical evidence presented for the formation of an LCT type pegmatite by direct anatexis.

INTRODUCTION

Mt. Mica pegmatite (Figure 1) is the site of the first reported occurrence of tourmaline in North America (Hamlin 1873, 1895) (Figures 2 and 3) and is famous for gem tourmaline production for nearly 200 years. It is experiencing a remarkable new chapter in its long-lived and historically important mining history. Coromoto Minerals LLC acquired the property and began mining in 2003. Recent mining by owner Gary Freeman has produced a large amount of quantitative information about the pegmatite. Careful mapping of the location and sizes of lepidolite (used as the series name for trioctahedral micas near trilithionite in composition) masses, pockets and the volume of pegmatite mined underground has provided an unprecedented database which we utilized in this study. Over the last dozen years of mining several hundred pockets have been carefully documented. Their sizes and contents have been carefully measured and recorded. These pockets have yielded large gem quality crystals of green and pink tourmaline rivaling the best material ever produced from Mt Mica in its almost 200-year history. The new
phase of mining activity by Coromoto Minerals, which began as an open trench mine in 2003, now continues down dip underground for over 100 meters to a depth of about 30 meters. The pockets range in size from a few cm$^3$ to one in excess of 500 m$^3$. Several dozens of the intermediate to larger pockets have produced thousands of carats of gem quality tourmaline and lesser quantities of morganite. Pocket density averages about one every 3 meters with larger pockets having greater spacing and small ones having less, making this one of the most pocket-rich pegmatites in North America. In addition to the gem material, thousands of high-quality mineral specimens including tourmaline, beryl, apatite, lepidolite, rose and smoky quartz, hydroxylherderite, cassiterite, pollucite, and kosnarite have been recovered.

**GENERAL GEOLOGY OF THE PEGMATITE**

Mt. Mica intrudes stromatic migmatite of the SMD (Figure 1). The migmatite consists of felsic leucosomes of quartz and feldspar and melanosomes of biotite-quartz-feldspar schist. The contact is sharp between the pegmatite and melanosomes but completely gradational between the leucosomes and the pegmatite. In a few places along the contact, a weakly developed 2-4 cm comb structure of oriented muscovite crystals is present. In most places the contact is parallel to foliation, but in places where the pegmatite cuts foliation, ductile deformation is clearly evident. In general, zoning in the Mt. Mica pegmatite is indistinct. Internal zoning is asymmetric and not well developed and, basically, in most places no intermediate zone can be distinguished. The wall zones extend inward from the hanging wall and foot wall contacts to the core zone. Grain size increases gradually up to the core zone which is marked by an abrupt increase in grain size, especially of muscovite which forms in large 4 to 10 cm “A” shaped twinned books and the appearance of more evolved minerals and miarolitic cavities. The dike ranges in thickness from 1 to 8 meters and dips about 20° to 25° to the SE. To date, the mining extends underground down dip about 100 m to a depth of 33 m beneath the surface. The outer zones of the pegmatite consist of nearly end-member albite (An ~ 0.5), quartz, muscovite, and schorl. Visual estimates and point counts of polished slabs yield a modal composition of the wall zone of roughly 40 % quartz, 39 % albite and 19 % muscovite.

**Figure 1.** Mt. Mica pegmatite with entrance to the underground workings, showing the contact with the overlying stromatic migmatite host rock. Mine opening 2.5 m.
Figure 2. A photograph of the mining at Mt. Mica ca. 1890. Sticks with red flags mark positions of pockets. The persons in the image are L. Kimball Stone (left) and Loren B. Merrill (right). Modified from Bastin (1911).

Figure 3. Elbaite crystal reproduced from The Tourmaline by A. C. Hamlin, 1873.

The wall zone of Mt. Mica is unusual as it is essentially devoid of K-feldspar; and the major K-bearing mineral is muscovite. K-feldspar only appears as large masses in the core zone of the pegmatite adjacent to some of the larger pockets, where it serves as a pocket indicator in some cases. Notably, one, or in some instances, two distinctive garnet and schorl lines occur along the somewhat finer grained footwall portion of the dike (Figure 4). The lines are about 0.5 to 1.0 m above the base of the pegmatite and are roughly parallel to the pegmatite-country-rock contact. The line consists of an undulating, somewhat sinusoidal to irregular concentration of 1 to 5 cm garnet crystals in a matrix of quartz and feldspar. The most evolved portion of the pegmatite is always above this horizon in the more incompatible-rich core zone of the pegmatite, where the pockets occur. Thus, this line is an important marker that the miners interpret to be the bottom of the productive zone, below which no pockets occur.

The core consists mainly of quartz, albite, microcline, and schorl with local pods of white cleavelandite and, less commonly, pods of deep purple lepidolite with elbaite, spodumene, pollucite, cassiterite, columbite group minerals, and rare beryl (Figures 5-17). Thin section analyses of the lepidolite masses revealed that bulk modal composition is 71% lepidolite and 29% quartz and albite.

Miarolitic cavities or pockets are relatively common in Mt. Mica and are the source of the gem elbaite that this pegmatite is so famous for. The pockets are ovoid in shape and tend to be most elongated in the horizontal plane and range in size from a few centimeters to one chamber in excess of 11 meters across.
MINERALOGY

Feldspars

The dominant feldspar in Mt. Mica is a white albite. K-feldspar is virtually absent in the outer zone of the pegmatite. The Ca content of the albite is uniformly low in the wall zone, averaging about An$_{2.0}$. Albite in the core zone approaches pure end-member albite, averaging An$_{0.2}$. K-feldspar is only present in the core zone, where large crystals of microcline occur, up to a meter in maximum dimension, near larger pockets and rare pollucite masses. The crystals are perthitic and determined by X-ray diffraction to be near maximum microcline. The average K/Rb ratio of 150 for K-feldspar (Marchal et al. 2014) is somewhat high for a B-rich LCT pegmatite, suggesting that Mt. Mica is not very evolved. However, the lower K/Rb ratio of 40 for K-feldspar located in and around pockets and pollucite pods reveals that Mt. Mica is moderately evolved in the core regions (Marchal et al. 2014).

This suggests a Na-dominant pegmatitic melt where micas are the dominant K minerals. Feldspars throughout the pegmatite are dominantly Na-rich plagioclase with the highest An content (An$_{14.45}$) occurring at the hanging-wall contact with the country rock. The sodic plagioclase from the wall zones (An$_{0.05}$–An$_{5.71}$) and core zone (An$_{0.13}$–An$_{0.88}$) correspond to nearly pure albite. The composition of the K-feldspar shows that the Ab content ranges between 2.37% near the pocket, to 4.79% next to a pollucite mass, and is confirmed by X-ray diffraction to be near maximum microcline. The Rb content ranges from 0.001 to 0.023 apfu, and Cs content ranges from 0 to 0.005 apfu. Overall, K-feldspar is not very enriched in either Rb or Cs, but there is a population of microcline spatially associated with miarolitic cavities and pollucite masses that has higher Rb and Cs. The K/Rb (apfu) ratio of microcline ranges from about 40 to 730.

Quartz

Quartz is the second most abundant rock-forming mineral in the pegmatite. It forms scattered pods up to m-size in maximum dimension in the core region, but no distinct quartz core is present. The color ranges from white to colorless to smoky brown. Crystals of colorless, smoky and white quartz occur in miarolitic cavities and can reach several dm in maximum dimension. Rose quartz crystals are found rarely in a few pockets. Unusual rings of rose quartz crystals encircle large smoky quartz crystals. An example of a quartz crystal-filled pocket is shown in Figure 5.
Figure 5. A miarolitic cavity, primarily filled with quartz crystals. About 50 cm field of view.

**Muscovite and Lepidolite**

Muscovite and lithium muscovite are the principal micas and are the dominant K-species in the wall zone of the pegmatite. Micas are small and show little change in composition from the pegmatite contact up to the core zone margin, with Li content ranging from >0.01 to 0.9 apfu (Marchal et al. 2014), (Figure 6). There is an abrupt increase in Li-muscovite crystal size at the core zone where large, euhedral, twinned mica crystals, up to 17 cm across occur. Crystals in close proximity or extending into pockets are commonly rimmed with lepidolite. The rims consist of a mosaic overgrowth of 2 to 5 mm lepidolite crystals. There is a sharp boundary in composition between the muscovite and the lepidolite rim (Figure 7). Pods of lepidolite, up to several meters in size, are intermittently distributed within the core zone. The lithium content of all lepidolite from pods and rims ranges from 2.0 to 3.4 apfu (Marchal et al. 2014). In a few instances, macroscopic interlayering of muscovite and lepidolite on a several micron scale occurs where a muscovite crystal extends into pockets or lepidolite pods. The details of the mica compositional ranges are presented in Marchal et al. (2014).

Figure 6. A-type twinned lithian muscovite crystal rimmed with lepidolite from a miarolitic cavity.
Tourmaline

Tourmaline is present in all the pegmatite zones. In the wall zone, tourmaline occurs as fine- to coarse-grained anhedral to subhedral prisms of black schorl and is occasionally graphically intergrown with quartz. Most of the tourmaline at Mt. Mica is schorl, which occurs abundantly in the wall zone and the core zone. Colored gemmy tourmalines are mainly restricted to the miarolitic cavities (pockets) (Figures 8-10), where they occur as blue to green to pink crystals, almost entirely of elbaite composition or color zoned from black to green to pink. Rossmanite has been found rarely in pink or colorless portions of pocket tourmaline. Foitite occurs sometimes as well, typically as the more or less fibrous black termination on some elbaite tourmalines. Very rarely, tourmalines have been found in miarolitic cavities that have a base of schorlitic composition that was overgrown by elbaite tourmaline of various colors and eventually had a rossmanitic portion near the termination and a black cap of foititic composition (Simmons et al., 2005a and 2005b).

In places, a distinct comb structure of tapered or wedge-shaped crystals fan out from the contact into the wall zone, pointing toward the interior of the pegmatite. This texture is similar to that described from San Diego Co. California pegmatites that have been attributed to rapid crystal growth (Webber et al. 1997, 1999). Schorl crystals also radiate around some pockets and the convergence direction of their long axes serves as a pocket indicator. Crystals that extend close to or into a pocket are typically color zoned grading from black to green to pink in color (Figure 8). In the core zone, around and near pockets, tourmaline is dominantly elbaite that occurs as opaque to translucent green, mm- to cm-size prisms. In the pods of fine-grained lepidolite masses, small pinkish tourmaline crystals are relatively common, but most are altered to clay minerals. Inside the pockets in the core zone, spectacular, gem-quality, prismatic crystals of blue to green to pink, pink and green color zoned prisms and watermelon tourmaline occur, some as large as 15 cm in length (Figures 9 and 10). This elbaite is associated with albite var. cleavelandite, quartz crystals up to 30 cm and medium–to–very coarse crystals or books of Li-muscovite that may be rimmed by lepidolite (Figure 6). The tops of some pink elbaite crystals are capped with a thin layer of black tourmaline (foitite) and the crystals are similar to the Mohrenkopf crystals from Elba, Italy.
Tourmaline exhibits the expected trend of Fe-rich schorl in the outer portion of the pegmatite evolving to greater contents of Al and Li in the core zone of the pegmatite (Simmons et al. 2005a & b) (Figures 11 and 12). The pink and green color of elbaite in pockets is mainly a consequence of Fe content. Colors grade from black to blue and blue-green to light green to pink as Fe content drops in the core zone and particularly in pockets (Simmons et al. 2005a & b).
Beryl

Beryl is not as abundant in the massive pegmatite as in many other Oxford field pegmatites. However, in miarolitic cavities, superb specimens have been found (Figures 13 and 14). Beryl found in the massive pegmatite is bluish-green to yellowish whereas in miaroles, beryl is white. Most of the crystals found in the miarolitic cavities consist of white to colorless or pinkish cesium-enriched beryl or pink morganite. Compared to other pegmatites in the Oxford field, notably the Orchard, the Emmons and the Bennett pegmatite, Mt. Mica pegmatite appears to have a paucity of common beryl. It is more abundant in the miarolitic cavities, suggesting that Be was retained until the pocket stage was reached.

Pollucite

Pollucite has been found in m-sized masses (Figure 15) in the inner zones of the pegmatite, associated with cleavelandite, microcline, montebrasite, lepidolite and spodumene. Recently, exceptional gem-quality extensively etched pollucite (Figure 16) has also been found in miarolitic cavities and is associated with crystals of cleavelandite, quartz, lepidolite, fluorapatite, gem-quality elbaite and beryl. Etched pollucite is easily mistaken for etched beryl.
Some domains in the etched pollucite have anomalous birefringence, which can further obscure their proper identification.

![Figure 15. A meter-sized pollucite mass.](image)

**Figure 15.** A meter-sized pollucite mass.

![Figure 16. Etched, gem-quality, pocket pollucite crystals](image)

**Figure 16.** Etched, gem-quality, pocket pollucite crystals

**Montebrasite**

Montebrasite occurs as masses and crude white crystals in the core zone, commonly with lepidolite and spodumene (Figure 17). The maximum dimension can reach several dm. No amblygonite has been found to date in any Oxford field pegmatites, even though in the older literature, amblygonite is commonly cited.

![Figure 17. Large montebrasite pods with lepidolite and altered spodumene in the core zone. About 2 m field of view (Gary Freeman photo).](image)

**Figure 17.** Large montebrasite pods with lepidolite and altered spodumene in the core zone. About 2 m field of view (Gary Freeman photo).
Spodumene

Spodumene has been observed in dm-sized crystals but is typically completely replaced by clay minerals or mica species in most cases (Figure 17).

Garnet

Garnet, typically almandine-spessartine solid solution composition, occurs dominantly in the garnet layer in the foot wall of the pegmatite (Figures 4 and 18). Garnet as isolated crystals is rare, but somewhat more abundant near the contact of the pegmatite; it is sparse in the rest of the pegmatite. The garnet line consists of an undulating, somewhat sinusoidal to irregular concentration of 1 to 5 cm garnet crystals in a matrix of quartz and feldspar. The most evolved portion of the pegmatite is always above this horizon in the more incompatible-rich core zone of the pegmatite, where the pockets occur. Thus, this line is an important marker that the miners interpret to be the bottom of the productive zone, below which no pockets occur. The garnet layer marks a distinct boundary layer of the pegmatite. In microenvironments, a highly fractionated mineral assemblage is present (Felch, 2014; Felch et al., 2016). Garnets may reach up to 2-3 cm in diameter and in garnet layer portion below miarolitic cavities, a characteristic alteration rim of dark blue tourmaline (Figure 18) is typically present (see Felch et al. (2016) for more details about the garnet line and the associated evolved mineral assemblage).

Figure 18. A 1.5 cm wide garnet with tourmaline rim from the garnet layer.

Columbite Group Mineralogy

Columbite group minerals show only minor to moderate enrichment in Mn and Ta, reaching only columbite-(Mn) (Simmons et al. 2013). The Mt. Mica pegmatite is not very rich in columbite group minerals. Among the high field-strength element bearing minerals, cassiterite and zircon are the most abundant. In fact, Mt. Mica seems rather depleted in Ta relative to other B-rich LCT pegmatites.

Cassiterite

Masses of cassiterite are abundant at Mt. Mica. In the past, a short-lived attempt was made to mine cassiterite for its tin content.
List of mineral species identified in the Mount Mica pegmatite

Albite          | Heterosite | Plagioclase group
Almandine      | Hureaulite  | Pollucite
Arsenopyrite   | Hydroxylapatite | Purpurite
Autunite       | Hydroxylherderite | Pyrite
Beraunite      | Jahnseite-(CaMnMn) | Quartz
Bertrandite    | Kaolinite group | Reddingite
Beryl          | Kosnarite   | Rhodochrosite
‘Biotite’      | ‘Lepidolite’ | Roscherite group
Cassiterite    | Laueite     | Rossmanite
Columbite-(Fe)| Löglingite  | Schorl
Columbite-(Mn)| Mccrillisite | Scorodite
Cookeite       | ‘Manganese oxides’ | Siderite
Crandallite    | Meta-autunite | Sphalerite
Dahlilite      | Metatorbernite | Spodumene
Elbaite        | Microcline  | Stewartite
Fairfieldite   | Microlite group | Strunzite
Fluorapatite   | Mitridatite | Tantalite-(Mn)
Foitite        | Montebrasite | Tapiolite-(Fe)
Glucine        | Montmorillonite group | Torbernite
Goethite       | Moraesite   | Tourmaline group
Goyazite       | Muscovite   | Triphylite
Graphite       | Opal        | Uraninite
Greifensteinite| Phosphosiderite | Zircon
Hematite       | Phosphouranylite |

BULK COMPOSITION OF MT. MICA

A very detailed and accurately surveyed geologic map produced by owner/operator Gary Freeman during mining shows the total area of pegmatite removed, the spatial distribution and aerial extent of pockets, massive lepidolite (compositions near trilithionite) pods, microcline, and xenoliths. The map was analyzed using image analysis and thickness values of the units to calculate the total volumes of pegmatite mined, lepidolite pods and all pockets found. Forty-five drill cores were taken across the pegmatite from the hanging wall to foot wall contacts along a transect intentionally avoiding lepidolite pods and miaroles. Cores were pulverized, thoroughly mixed and homogenized and the percent Li content calculated from the mapped volume was added to produce a sample that was representative of the bulk composition of Mt. Mica. The sample was then analyzed by fusion ICP spectroscopy for major and trace elements and DCP spectroscopy for B and Li. Structural water was determined by LOI. Water content was calculated using the calculated volume of open space (pocket volumes), assuming that the pockets were filled with water-rich fluid. This fluid content was added to LOI water (above 500°C) to estimate a maximum H₂O content of 1.16 wt. % of the pegmatite melt. REE plots of bulk pegmatite vs. leucosomes from the migmatite are strikingly similar. Chondrite normalized REE patterns of leucosomes and pegmatite are very flat with no Eu anomaly, whereas Sebago granite is more strongly LREE-enriched and displays a pronounced negative Eu-anomaly (Figures 19 and 20). Spider diagrams of leucosomes and pegmatite vs. average crust show very similar patterns (Figure 21). These results suggest that the Mt. Mica pegmatitic melt did not form by fractional crystallization of the older Sebago pluton, but instead was derived directly from partial melting of the metapelitic rocks of the SMD. Batches of anatectic melt accumulated and coalesced into a larger volume that subsequently formed the pegmatite. This is the first chemical evidence presented for the formation of an LCT type pegmatite by direct anatexis.

We believe that we have obtained a very accurate estimate of the whole rock composition of Mt. Mica Pegmatite and it illustrates that the pegmatite was not close to water saturation at the time of emplacement and that overall the pegmatite is only moderately evolved (Simmons et al. 2016), only areas within the core zone and the
pockets are highly evolved. Also, although Mt. Mica is generally considered to be a relatively pocket-rich pegmatite, the actual volume of open space is quite small, less than 1.0 % and the calculated amount of water contained in the melt that formed the pegmatite is about 1.16 wt.%.

**Figure 19.** Whole-rock chondrite normalized REE plot of Mt Mica Pegmatite (red), SMD Leucosomes (blue), Sebago Pluton Granitic rock (green). Chondrite values of McDonough and Sun (1995).

**Figure 20.** Whole-rock spider diagram sample of Mt. Mica pegmatite (red) and SMD leucosomes (blue) and Sebago Pluton Granite (green) relative to upper continental crust. Normalization values from Taylor and McLennan (1985).

**Figure 21.** Whole-rock spider diagram of Mt. Mica pegmatite (red) and SMD leucosomes (blue) and Sebago Pluton Granite (green) relative to total crust. Normalization values from Rudnick and Fountain (1995).

**MODEL FOR POCKET FORMATION**

The model for pocket formation employed in the calculations of pocket volume and total water content of the pegmatite is that described by Simmons *et al.* (2003, 2012 and 2016), which involves exsolution of a second fluid phase after most of the pegmatic melt has crystallized. The residual melt accumulates toward the center of the dike and becomes progressively enriched in water and other fluxing materials until ultimately, a second aqueous fluid
exsolves forming more or less spherical segregations ("bubbles") of a flux-rich aqueous fluid that constitutes a protopocket. According to the most widely accepted theory of pocket formation (Jahns & Burnham 1969), once the supercritical aqueous fluid starts to exsolve, diffusion of ions from the coexisting silicate melt into the fluid supplies nutrients to the crystals growing in the protopocket. Continued rapid diffusion of ions from the silicate melt to the growing crystal surfaces in the fluid of the protopocket is proposed to explain the greater volume of crystals found in the pockets than what could have grown from the less dense, aqueous, pocket-forming fluid alone. Thus, the flux-rich aqueous fluid is the medium through which ions diffuse to the growing crystal surfaces. We note that this model of pocket formation is directly opposed to that of London, (2013), but we feel this model best supports the observations of pockets found at Mt. Mica.

CONCLUSIONS

In contrast to suggestions of previous authors (Wise & Brown 2010, Wise & Francis 1992) that this pegmatite was derived by fractional crystallization from the Sebago Granite pluton (or any other pluton), we believe the chemical evidence strongly suggests that the Mt. Mica pegmatic melt could be derived directly from partial melting of the metapelitic rocks of the Sebago Migmatite Terrain. We suggest that the Mt. Mica pegmatic melt did not form in situ, but that batches of anatectic melt accumulated and coalesced into a larger volume that subsequently formed the Mt. Mica pegmatite. This is new chemical evidence for the formation of an LCT type pegmatite by direct anatexis.

REFERENCES

Taylor & McLennan (1985) The Continental Crust; Its composition and evolution; an examination of the
Oscillatory Nucleation In The Formation Of Line Rock In Pegmatite-Aplite Dikes. *Journal Of Petrology* 38,
1777-1791.
Maine. *Northeastern Geology, 14*, 82-93.
THE HAVEY PEGMATITE, POLAND, ANDROSCOGGIN COUNTY, MAINE

Encar Roda-Robles¹, Alexander U. Falster ², William B. Simmons ², Jeff Morrison³, Myles M. Felch² and James W. Nizamoff⁴

¹Dpto. Mineralogía & Petrología, UPV/EHU, Bilbao, Spain, encar.roda@ehu.es
²MP² Research Group, Maine Mineral & Gem Museum, 99 Main Street, Bethel, Maine
³20 East Elm Street, Yarmouth, Maine 04096
⁴64 Mineral Drive, Hebron, ME 04238

INTRODUCTION

The Havey pegmatite (Figure 1) in Androscoggin County, Maine belongs to the Oxford pegmatite field. The pegmatite is named the Berry-Havey and is owned by different parties. The Havey pegmatite portion is operated by Jeffrey Morrison. The pegmatite is one of the more evolved Oxford field rare-element pegmatites. It is enriched in Li, F, B, Be and P and contains miarolitic cavities with gem tourmaline (Figures 2-4 and 6-8). Lepidolite masses are common, but pollucite is rare. The pegmatite is complexly zoned with a border zone, two intermediate zones, a core margin and a core zone (Figure 1). Tourmaline is abundant throughout the pegmatite. The pegmatite has been cut by several younger basaltic composition dikes (Figure 5).

Figure 1. Map and cross sections of the Havey pegmatite, 2011.
The Havey pegmatite is a tabular to irregular body, conformable to the host-rock (biotite-amphibole-rich schist), to the SW of the quarry (~40º SSE), whereas in the northern part the body is more horizontal. The lack of outcrop of the foot-wall makes it difficult to determine the thickness of the body but, according to the present level of exposure, it is over 30 m. The pegmatite shows a well-developed internal structure, with five different zones: wall zone, intermediate zone I & II, core margin and core zone. These zones are subparallel, with quite irregular limits among them, mainly for the core margin-core zone transition. From the contacts inward, the following zones are distinguished (Roda-Robles et al., 2015) (Table 1): (1) Wall zone, commonly with a fine to medium sized pegmatitic texture, although a gneiss-like facies is locally common. Main minerals are quartz, K-feldspar, plagioclase, biotite and muscovite, with tourmaline and garnet as common minor phases. (2) Intermediate zone volumetrically is the most important, mainly in the lower portion of the dyke, with more than the 85 % belonging to graphic intergrowths of quartz-K-feldspar (intermediate zone I). Biotite, garnet and black tourmaline are minor phases. In the lower part of the intermediate zone, locally the texture and mineralogy change to coarse quartz, K-feldspar and black prismatic tourmaline (intermediate zone-II). Moreover, a > 30m³ “xenolitic” gneiss-like body is located in the lower intermediate zone-I, showing similar mineralogy and gradual textural changes with the pegmatitic material. (3) Core margin, volumetrically important, this unit hosts the different pods that constitute the core of the pegmatite. Main minerals are albite (cleavelandite), quartz and tourmaline. Moreover, garnet commonly occurs as medium sized reddish-brownish crystals concentrated in a layer just below the core pods. This “garnet layer” is common to most of the pegmatites from this region. (4) Core zone is not a continuous unit, but several pods of different sizes (2 to ~10m), frequently interconnected and hosted by the core margin.

Table 1. Main characteristics of the Havey pegmatitic units and the associated tourmaline.

<table>
<thead>
<tr>
<th>ZONE</th>
<th>MINERALOGY</th>
<th>GENERAL TEXTURES</th>
<th>TOURMALINE TEXTURE</th>
<th>COMPOSITIONAL VARIATION</th>
</tr>
</thead>
<tbody>
<tr>
<td>WALL</td>
<td>Qtz, Kfs, Pl, Bt, Ms ± Grt ± BLACK Tur</td>
<td>Homogeneous, very fine to medium grained facies. Locally greenish Kfs</td>
<td>Very fine grained, black prismatic crystals. Very scarce</td>
<td>Dravite-Schorl</td>
</tr>
<tr>
<td>INTERMEDIATE-I</td>
<td>Qtz, Kfs ± Grt ± Bt ± BLACK Tur</td>
<td>Qtz-Kfs graphic intergrowths (&gt; 90% volume)</td>
<td>Fine to medium subhedral crystals</td>
<td>Schorl-foitite</td>
</tr>
<tr>
<td>INTERMEDIATE-II</td>
<td>Qtz, Kfs, BLACK Tur</td>
<td>Blocky Kfs and Qtz</td>
<td>Coarse prismatic black crystals</td>
<td>Schorl-foitite</td>
</tr>
<tr>
<td>CORE MARGIN</td>
<td>Ab, Qtz, BLACK ± GREEN Tur</td>
<td>Matrix of tabular crystals of Clv, where coarse tourmaline crystals occur</td>
<td>Coarse black tourmaline prisms (&lt; 70 cm length) crowned by an intergrowth of black±green Tur and Ab</td>
<td>From Schorl -foitite to elbaite</td>
</tr>
<tr>
<td>CORE</td>
<td>Lpd, Ms, Qtz, Kfs, Mon, Ab, Brl, Fe-Mn-Pho GREEN, PINK &amp; MULTICOLORED Tur</td>
<td>Irregular pods of fine grained Lpd, coarse book Ms, Mon, Fe-Mn-Pho-Cst±Col-Tan Coarse morganite (Brl) sub-euhedral crystals are common. Pockets with elbaite in a Cookeite matrix</td>
<td>In the pods, subhedral, fine to medium zoned xls watermelon, pink or green crystals with lepidolite and muscovite. In pockets, green, teal or watermelon gem-quality xls.</td>
<td>Pods: Elbaite-Rossmanite Pockets (gem): Elbaite -rossmanite</td>
</tr>
</tbody>
</table>

Qtz=quartz; Kfs=feldspar; Pl=plagioclase; Ms=muscovite; Bt=biotite; Grt=garnet; Tur=tourmaline; Ab=albite; Clv=cleavelandite; Mon=montebrasite; Cst-cassiterite; Col-Tan=columbite-tantalite; Lpd=lepidolite; Pho=phosphates; Brl=beryl; Cook=cookeite. *Grain size: very fine = <6 mm; fine = 6 mm to 2.5 cm; medium 2.5 cm to 10 cm; coarse = >10 cm.
CRYSTALLIZATION SEQUENCE AND CHEMICAL EVOLUTION

The asymmetry indicates that the crystallization from the footwall and from the hanging-wall proceeded in different ways. Textures such as the quartz-K-feldspar and quartz-tourmaline graphic intergrowths, and the comb tourmaline crystals, suggest that crystallization proceeded under disequilibrium conditions from an undercooled melt (Simmons et al. 2003). Crystallization of the tourmaline layer in the core margin followed different steps. It starts with the crystallization of the tapered prisms, in general perpendicular to the contacts. Then, it follows with the formation of the crowns of quartz-tourmaline graphic intergrowth around the tapered prisms, where the composition and color often change from black schorl to greenish (or bluish) elbaite. Finally, it ends with the breaking of some of the tourmaline crystals, mainly belonging to the crowns. The occurrence of a significant volume of tourmaline in this layer implies a pronounced depletion of B in the melt during its crystallization. The saturation in volatiles in the melt increased the crystallization of anhydrous and non-volatile-bearing minerals, such as quartz and feldspars, in the wall and intermediate zones of the pegmatite, which would have increased the mole fraction of the volatile components in the melt, ultimately resulting in the exsolution of a fluid phase, and the formation of pockets. The presence of B₂O₃ enhances the solubility of water in the melt (Holtz et al. 1993; London 2009). Accordingly, the formation of the tourmaline layer dramatically lowers the solubility of water in the remaining melt, which also promotes exsolution of a water-rich fluid to form pockets. Assuming a closed system, we speculate that the breaking of some of the last formed tourmaline crystals in the tourmaline layer of the core margin could be related to a sudden increase in the fluids pressure provoked by the exsolution of the water-rich fluids from the melt, which imply a significant volume increase that could cause brecciation (Phillips 1973; Burnham and Ohmoto 1980; Burnham 1985). At this point, the concentration of Fe was low enough and Li was high enough to allow the crystallization of the first colored tourmalines in the core margin, followed by the crystallization of the core zone, where all the tourmaline corresponds to elbaite, with variable amounts of F, Li, Al, Mn, and vacancies. On the other hand, garnet crystallization ends in the core margin, below the core zone. This is likely related to the increase of the Li and F content in the pegmatite-forming system making Mn compatible in micas and tourmaline, which would destabilize garnet, that had evolved to Mn-richer compositions (Čerňy et al. 1985; London et al. 2001; London 2008). Overall, tourmaline compositions evolve from schorl in the outer zone of the pegmatite to elbaite in the core zone of the pegmatite.

Figure 2. Pockets in Havey Pegmatite
Figure 3. The “Spaniard” pocket being worked in 2011.

Figure 4. Jim Nizamoff in “Otto’s” pocket.

Figure 5. Large spray of colorful pink elbaite in massive pegmatite.

Figure 6. Basaltic dike cutting Havey pegmatite.

Figure 7. Open gem tourmaline pocket. Green and pink elbaite visible along bottom of pocket.

Figure 8. Close up of Figure 6 showing bi-colored gemmy elbaite.
MINERALS

List of mineral species identified in the Havey pegmatite

<table>
<thead>
<tr>
<th>Albite</th>
<th>Hydroxylherderite</th>
<th>Rhodochrosite</th>
</tr>
</thead>
<tbody>
<tr>
<td>Almandine</td>
<td>Lacroixite</td>
<td>Rockbridgeite</td>
</tr>
<tr>
<td>Apatite-(CaF)</td>
<td>Landesite (TL)</td>
<td>Rossmanite</td>
</tr>
<tr>
<td>Arsenopyrite</td>
<td>Lepidolite</td>
<td>Schoepite</td>
</tr>
<tr>
<td>Autunite</td>
<td>Lithiophilite</td>
<td>Schorl</td>
</tr>
<tr>
<td>Bertrandite</td>
<td>Meta-autunite</td>
<td>Siderite</td>
</tr>
<tr>
<td>Beryl</td>
<td>Metaschoepite</td>
<td>Spodumene</td>
</tr>
<tr>
<td>Biotite</td>
<td>Microcline</td>
<td>Stewartite</td>
</tr>
<tr>
<td>Cassiterite</td>
<td>Microlite group</td>
<td>Strunzite</td>
</tr>
<tr>
<td>Columbite-(Fe)</td>
<td>Mitridatite</td>
<td>Tantalite-(Mn)</td>
</tr>
<tr>
<td>Columbite-(Mn)</td>
<td>Montebasite</td>
<td>Todorokite</td>
</tr>
<tr>
<td>Cookeite</td>
<td>Montmorillonite</td>
<td>Tourmaline</td>
</tr>
<tr>
<td>Cryptomelane</td>
<td>Muscovite</td>
<td>Triplite</td>
</tr>
<tr>
<td>Dahllite</td>
<td>Phosphosiderite</td>
<td>Uraninite</td>
</tr>
<tr>
<td>Dickinsonite-(KMnNa)</td>
<td>Plagioclase group</td>
<td>Uranophane</td>
</tr>
<tr>
<td>Elbaite</td>
<td>Pollucite</td>
<td>Wardite</td>
</tr>
<tr>
<td>Eosphorite</td>
<td>Purpurite</td>
<td>Zircon</td>
</tr>
<tr>
<td>Fairfieldite</td>
<td>Quartz</td>
<td></td>
</tr>
<tr>
<td>Hureaulite</td>
<td>Reddingite</td>
<td></td>
</tr>
</tbody>
</table>

Tourmaline

Tourmaline is the most sought after mineral species in this pegmatite. In the past few years, some very fine gem tourmalines have been produced (Figures 11-15). Whereas the size is not as large as the tourmalines from Mt. Mica, the crystals consist of a high proportion of gem material. The pocket tourmalines are primarily green but some pink and watermelon colors have also been produced. In Figure 4, a pink tourmaline spray is shown that occurred in the inner zones of the pegmatite, seemingly of fracture-controlled origin. Schorl is the common tourmaline of the outer, massive zones in the pegmatite. Compositionally, tourmaline from the Havey pegmatite ranges from schorl to elbaite, with some rossmanite (Figure 16).
Figure 11. Green elbaite just recovered from a 2011 find.

Figure 12. A selection of gem tourmaline from the Havey pegmatite. 1-2 cm range.

Figure 13. Two gem crystals from the Havey pegmatite, about 3 cm tall.

Figure 14. Multicolored elbaite. (J. Scovil photo)

Figure 15. Fractured gem elbaite prism in albite.
Figure 16. Plots of the chemical composition of tourmalines from the different units in the Havey pegmatite (Roda-Robles, 2011): (a) X-site plot of (Na+K)-Ca-X-site vacancy; (b) variation in R²⁺/(R²⁺+2Li) vs. o/(o+Na+K) as proposed by Henry et al. (2011) for the classification of tourmaline; (c) triangular plot of Li-Al(Y)-Fe+Mn; and (d) X-site charge vs. F. All data in apfu.
Fluorapatite

Fluorapatite occurs as an accessory phase in all the zones of the Havey, except in some pockets of the core zone, where it may be abundant. In the wall zone and intermediate zone apatite appears as fine-grained (< 5 mm) greyish crystals (Table 1). Bigger bluish prismatic to anhedral crystals are relatively abundant in the core margin, frequently intergrown with garnet in the garnet layer. In the core zone apatite has only been found in pockets, frequently coating some coarse smoky quartz crystals. In the pockets apatite may occur as bluish or purple short hexagonal prisms (Figure 17), and as euhedral, sharply zoned, hexagonal lenses, with a clear blue to greyish core and a narrow white rim. As for garnet, chemical composition of apatite changes notoriously regarding the pegmatite zone (Table 1). All the analyzed apatite crystals are F-rich fluorapatite. Apatites from the wall zone, intermediate zone and core margin are the F-poorest; whereas those from the pockets show the highest F contents. Manganese contents change in a great extent (Table 1). Mn-rich apatite is only found in the inner zones of the pegmatite (core margin and core zone), with the highest values in the apatite from the garnet layer of the core margin, and intermediate values in the core of the zoned lenticular crystals of the pockets. The rest of the apatites from the pockets, as well as those from the wall zone and intermediate zone, are Mn-poor. Iron contents are low in all the analysed crystals. Interesting zoned crystals (Figure 18) have recently been found.

![Figure 17. Large fluorapatite matrix specimen](image)
About 25 cm maximum dimension

![Figure 18. Zoned fluorapatite](image)
4 cm across.

Quartz

Quartz is abundant as smoky quartz crystals in miarolitic cavities. Complexly grown crystal groups (Figure 15) or well-crystallized single crystals (Figure 19) have been found in the past few years.

Feldspars

Feldspars are the major mineral group in the pegmatite. Potassium feldspars are typically microcline and are white to off-white unless they have been metasomatized near miarolitic cavities, then the color changes into beige to tan. Albite is abundant and in replacement units and in miarolitic cavities it is generally found as the platy variety cleavelandite (Figure 9).

Garnet

Garnet occurs as a minor phase in all the zones of the Havey pegmatite, with the exception of the core zone, where it has not been found. Most of the garnet crystals exhibit subrounded forms and reddish to brownish colours in hand sample. Only some crystals are irregular or poikilitic, mainly in the gneiss-like material from the wall zone. The biggest crystals are found in the garnet layer of the core margin, just below the core zone. There is a marked and continuous chemical variation in the garnet composition from the wall zone inwards, with a Mn increase parallel to the Fe decrease (from Mn-rich almandine in the wall zone to Fe-rich spessartine in the core margin) (Table 1). In contrast, only garnets from the wall zone show a core-to-rim systematic variation, with a Fe-, Mg-, and Ca-decrease, parallel to a Mn-increase. In the other zones of the pegmatite garnet chemical profiles of single crystals are flat or saw-shaped. In one case, extensive replacement of almanditic garnet by löllingite has been documented (Simmons & Falster, 2017)
**Hydroxylherderite**

Hydroxylherderite is a common accessory mineral in miarolitic cavities in the Havey pegmatite (Figure 20).

![Figure 19. Mosaic smoky quartz crystal](image1)

![Figure 20. Hydroxylherderite (J. Scovil photo)](image2)

**Lepidolite**

Lepidolite is abundant in the pegmatite as masses in the inner zones of the pegmatite as well as unusual, pillar-shaped crystals in the miarolitic cavities (Figures 21-22). Notably, the massive lepidolite is very colorful with attractive lilac hues (Figure 10).

![Figure 21. Columnar lepidolite from a pocket.](image3)

![Figure 22. Columnar lepidolite from a pocket.](image4)

**Beryl** has been found as large (25 cm+) etched masses of morganite in the pegmatite (Figure 23). It also occurs as common, bluish to greenish beryl in the massive pegmatite.
SUMMARY

The continuous increase in the Mn content in the garnet from the wall zone to the core margin, parallel to the decrease in Fe, Mg and Ca (Table 1), is a typical fractionation-induced trend, with almandine in the less fractionated outer zones of the pegmatite, and spessartine in the core margin. No garnet has been found in the core zone. In this zone crystallization of phosphates (both, Fe-Mn-rich and Al-F-rich) becomes important, occurring as common subrounded pods of montebrasite and/or Fe-Mn-phosphates. The lack of garnet once the phosphates start crystallizing has been reported in many other pegmatites (e.g. Roda-Robles et al., 2013). This is interpreted to be the result of the sequestering of the remaining Fe and Mn by the phosphates, which prevents the crystallization of garnet. The increase of Mn from core to rim in the garnet indicates that minerals other than garnet control the Fe and Mn contents of the melt (Müller et al., 2012). Iron would incorporate preferentially into tourmaline, whereas Mg partitions into biotite and Mn into garnet. The main chemical variations for major elements in apatite correspond to the F content. The incompatible character of F makes that melt becomes Fricher with fractionation. Thus, the minerals growing at the final stages of the Havey crystallization tend to be F-rich, as it is the case of apatite, tourmaline, micas and montebrasite.

The development of pockets inside the core zone was most probably related to the exsolution of a fluid phase from the melt. The pockets represent the space that was once filled by accumulated supercritical fluid (Nabelek et al. 2010; Simmons et al. 2012), and show that the exsolved fluid was collected in discrete spaces instead of one continuous space between the hanging wall and lower portions of the dikes (Maloney et al. 2008). Fluid inclusions in the quartz crystals from the pockets at the Havey are mainly aqueous (Fuertes-Fuente, personal communication), which supports this model. Taking into account the mineralogy of the pegmatite, the presence of pockets, and that the regional metamorphism occurred at low pressure (Tomascak et al. 2005), we can assume that the crystallization of the Havey pegmatite developed under pressures in the range 2–3 kb. The maximum water solubility in silicate melts at those pressures is ~6 wt% in the absence of boron (Holtz et al. 1995). This amount of H2O, given its molar volume at 400 °C and 2 kbar, would occupy ~28% of the chamber volume (London 2008; Maloney et al. 2008). Based on these estimations, and on the relation between the volume of the pockets and the volume of the core zone, it seems plausible that the exsolution of fluids from the pegmatitic melt took place close to the end of the crystallization of the core margin. According to London (2008), the crystallization of granitic melt containing 6 wt% H2O promotes a volume increase of 21% at constant pressure, and this release of vapor could cause the rupture of the pegmatite. In our case, the rupture of the tourmaline crystals close to the core zone and pockets therein.

Figure 23. Etched beryl, var. morganite. 6 cm across.
REFERENCES


ROAD LOG

STOP 1: HAVEY PEGMATITE. UTM: 396013.86E 4880609.96N

Mileage

0.0  Meeting Point. Back parking lot of the Maine Mineral & Gem Museum, the Museum address is 99 Main Street, Bethel. The back-parking lot can be accessed from Chapman Street, off of Main street. UTM coordinates: 357596.60E 4918695.79N.

0.0  Cars will line up going north on Main street by the Museums front entrance.

0.2  Head north on Main Street toward Mechanic Street

23.6  Main Street becomes ME-26. Continue on ME-26 S.

24.8  Turn Right onto Main Street in Paris (This is still ME-26 S)

25.2  At the stop light Turn Left onto Fair street (This is still ME-26 S)

31.4  Continue travelling down ME-26 S (Main street)

36.1  Turn Left onto ME-121 N

39.8  After the traffic light in Mechanic Falls turn left onto ME-121 N (ME-11, Lewiston Street)

40.0  Take slight Right off of ME-121 N onto Lower Street

40.2  Turn Right onto Empire Road

41.8  Turn Left onto Hardscrabble Road

41.9  DANGEROUS TURN BEFORE HILL, PROCEED WITH CAUTION. Turn Left onto Levine Road

50.0  Turn Left off of Levine Road to access Havey Quarry entrance. Parking spaces will be located along the left side of the mine road. WHEN PARKING, DO NOT BLOCK ACCESS ROAD.

After parking the group will convene and walk down into the quarry.

STOP 2: MT. MICA PEGMATITE. UTM: 382422.43E 4902825.20N

50.1  Exit the Havey Quarry and Turn Right onto Levine Road

50.2  Turn Right onto Hardscrabble Road

51.8  Turn Right onto Empire Road

52.1  CAUTION DANGEROUS INTERSECTION. Turn Left onto ME-121/ME-11 (Minot Ave).

52.8  Turn Right onto ME-119N (Woodman Hill Road).

59.1  In the village of West Minot, Turn Left onto ME-119 N
67.6 **Turn Right** onto ME-117 (Buckfield Road). Be prepared to turn left.

384ft **Turn Left** onto Hill Street

68.5 Merge onto Old Route 26/Paris Hill Road

69.1 Take a slight **Right** onto Paris Hill Road

70.3 In the village of Paris, **Turn Right** onto Lincoln Street.

70.7 Continue straight, Lincoln Street turns into Mt. Mica Road.

72.1 Continue driving down Mt. Mica Road

**Turn Left** onto a gated dirt road (UTM 382373.27E 4902591.54N), this is the access road to the Mt. Mica Mine. Drive up the access road and find parking in the lots on the right. After parking the group will convene before walking down into the mine. **WHEN PARKING, DO NOT BLOCK ACCESS ROAD.**

**End of Field Trip Stops**

The road log continues to return to the meeting point at the Maine Mineral & Gem Museum, where the Welcoming Reception is being held.

72.1 Exit the Mt. Mica Mine access road by taking a **Right Turn** onto Mt. Mica Road, heading west.

73.4 Continue going straight onto Lincoln Street

73.3 **Turn Right** onto Paris Hill road

75.6 **Turn Right** onto ME-26 N and continue north.

95.1 Continue straight onto ME-26 N/Main Street, crossing the railroad tracks.

95.3 The Maine Mineral & Gem Museum is located on the left at 99 Main Street.

---

Friday night from 5-7 pm (hors d’oeuvres and cash bar) there will be an *open house at the Maine Mineral and Gem Museum*. See world class Maine minerals and discover the otherworldly Stifler Collection of Meteorites. Led by the MMGM staff.
INTRODUCTION

Organic-rich marine sedimentary rocks, a.k.a. black shales (e.g., Wignall, 1994), are the subject of great public and scientific attention nowadays. They are source rocks for the world's oil and gas. Current public interest is due in large part to advances in hydrofracturing or "fracking" that have changed the economics of the oil industry. For example, the Marcellus Shale, underlying portions of New York, Pennsylvania, Maryland, West Virginia, and Ohio, is a source rock for oil and natural gas, and is currently exploited by fracking in several of the above states. In Maine and New Hampshire, the Smalls Falls Formation, which once resembled the Marcellus, now appears as sulfidic schist and quartzite, having undergone metamorphism and deformation during the Acadian orogeny. As a result, Smalls Falls rocks have been thermally promoted far beyond the "oil window" of ~60 to 120°C and hence are not candidates for oil and gas exploration. This applies also to black shales and schists elsewhere in Maine, which similarly lack fracturing potential for oil and/or gas due to high temperatures experienced during metamorphism.

"The rocks don't change from one generation of geologists to the next" (J. Haller, pers. comm., 1984). Nonetheless, our perceptions of the rocks, and our interpretation of their relations, do change. Figure 1 shows the pre-plate tectonic conception of regional geology as expressed fifty years ago, whereas Figure 2 shows, after decades of work, our current understanding of tectonic subdivisions. Prior to 1970 and the plate tectonic revolution, the geology of the northern Appalachians generally followed the stratigraphic model of Billings (1956) that was based on rocks in New Hampshire. Extending this system to the east into Maine proved problematic, however. At the 1970 NEIGC meeting, Boone, Boudette, and Moench established the now-classic Rangeley to Seboomook stratigraphic model (Boone et al., 1970) that is now widely accepted. Later workers (e.g., Hatch et al., 1983) successfully extended the Rangeley-based stratigraphic system to the west into New Hampshire. In other words, it proved easier to allow our interpretations to evolve from east to west than the reverse, due largely to the extensive deformation and metamorphism in western regions.

In the Silurian, following the Taconic arc-continent collisions, marine conditions persisted to the southeast between the enlarged Appalachian margin of Laurentia and several peri-Gondwanan terranes. The Central Maine Trough deepened rapidly to the east in the vicinity of Rangeley, Maine, as shown by significant increases in the stratigraphic thickness of formations compared to the area west of Rangeley. This abrupt deepening was named the Silurian Tectonic Hinge by Boone et al. (1970). The earliest Silurian sediments were the thinly laminated sandstones and siltstones of the Silurian (?) Greenvale Cove Formation, overlain by coarse clastic sediments of the Silurian (?) Perry Mountain Formation, overlain by coarse clastic sediments of the Rangeley Formation, followed by quartz-rich sandstone turbidites of the Perry Mountain Formation. At this time, portions of the basin became so cut off from the general oceanic circulation that bottom waters became oxygen-deprived, including locally (but not pervasively) anoxic conditions and apparently rare euxinic (sulfidic) conditions. This was the depositional environment of the Smalls Falls Formation seen on this trip. Smalls Falls rocks are in turn overlain by the calcareous Madrid Formation (Siluro-Devonian) and the pelitic turbidites of the Devonian Seboomook Group, the latter correlated with the Littleton Formation of New Hampshire (Billings, 1956). Appendix A of this field trip guide contains a formation-by-formation description of the metasedimentary rocks of this region.

The area encompassed by this field trip in western Maine is underlain by metamorphic rocks of Paleozoic age that have undergone complex deformation to the southwest, and been intruded by numerous plutons of mostly Devonian age. In addition to the Bedrock Geologic Map of Maine (Osberg et al., 1985), the reader is referred to the geologic map of Western Interior Maine (Moench and Pankiwskyj, 1988b) that covers all of the field trip stops. A more recent regional geologic map by Moench et al. (1995) also covers the
localities we will visit, but at a smaller scale. The reader is also referred to Boone et al. (1970), Moench and Boudette (1970), Moench (1971), and Moench (2006), for more extended descriptions of stratigraphy and structure in this region.

Figure 1. Lithofacies and paleogeography of the Ludlow Stage in the northern Appalachian region (after Boucot, 1968). Box shows area of this field trip.

TECTONIC FRAMEWORK

The Smalls Falls Formation crops out in the Central Maine Trough (CMT), primarily in the northwestern part. The Central Maine Trough, essential to understanding the regional tectonics, comprises marine sedimentary rocks of Silurian-Devonian age (Reusch and van Staal, 2012 and references therein). It extends from Connecticut through Massachusetts and New Hampshire to Maine. On its northwest flank, CMT strata conformably overlie Ordovician marine sedimentary and volcanic rocks of the Bronson Hill belt. Note that the Middle Ordovician volcanic rocks are unconformable on the Dead River Formation that has Gondwanan provenance and displays Penobscottian deformation. Contact relationships along the southeastern margin of the CMT are varied. In southern Maine, CMT strata are difficult to distinguish from strata of the Merrimack Trough. In south-central Maine, CMT strata are in fault contact with the Liberty-Orrington belt. In New Brunswick, CMT strata overlie the Miramichi belt via a Salinic (Silurian) unconformity. Paleocurrent data and stratigraphic relationships strongly suggest the CMT received sediment from the northwest during the Early Silurian, and subsequently also received sediment from the southeast (e.g., the Rangeley Formation along the northwestern margin has northwestern provenance, and the Vassalboro Group along the southeastern margin has southeastern provenance).
Figure 2. Generalized geologic map of portions of Maine and adjacent regions showing locations of major lithotectonic terranes (after Hibbard et al., 2006). Note correspondence between this map and the Boucot map: Calcareous Mudstone Belt has become Connecticut Valley-Gaspé Trough, Central Clastic Belt has become Central Maine Trough; Coastal Volcanic Belt retains its name. Box shows area of this field trip. See Hibbard et al. (2006) for description of numbered features.
Regional stratigraphic relationships suggest the CMT occupied a forearc setting with respect to a post-Taconic continental arc that lay to the northwest on the Taconic-modified Appalachian margin of Laurentia (Hussey et al., 2010; Reusch and van Staal, 2012). Rocks of the Liberty-Orrington-Miramichi belt constitute a segment of the Brunswick subduction complex between the forearc basin and trench located in the Fredericton Trough farther southeast. Recent detrital zircon results indicate the Fredericton Trough marks the locus of a formerly wide basin (Tetagouche-Exploits). Closure of the basin during the Silurian resulted in emergence of the Brunswick subduction complex, documented by an erosional influx along the southeastern margin of the CMT (e.g., Vassalboro Group).

Early Devonian strata of central and northern Maine, including the Madrid Formation and Seboomook Group, are widely interpreted as a clastic wedge deposited in a foreland basin that lay northwest of a growing Acadian orogen (Bradley, 1983; Bradley et al., 2000; Bradley and Tucker, 2002). The basin migrated northwestward in front of the growing orogen, eventually emerging above sea level and culminating with deposition of the Catskill delta. The Smalls Falls Formation was deformed during this Early Devonian event. The northwestward migration of the basin and the deformation front are constrained by a host of dated plutons (Bradley et al., 2000). All of these rocks display a steeply dipping foliation that is axial planar to map-scale folds such as the Brimstone Mountain anticline and Bear Hill syncline between Rangeley and Phillips (Moench and Pankiwskyj, 1988b). Moench was much interested in early, syn-sedimentary structures that he related to down-to-basin extensional faults such as the Rumford allochthon (Moench and Pankiwskyj, 1988a; Reusch et al., 2010)

"REGIONAL CONTACT METAMORPHISM"

The rocks of west-central Maine are polymetamorphosed and intruded by plutons1 so numerous that their overlapping contact metamorphic aureoles have in places become regional features. Charlie Guidotti (pers. comm.) coined the term "regional contact metamorphism" to describe this high-T and low-P pattern. Figure 3 shows in outline the major plutons of western Maine and the general nature of metamorphic isograds that surround them. In detail, the metamorphic history in western Maine is highly complicated and involves several thermal events that are variably superimposed on each other. These are commonly referred to as the M1, M2, and M3 events. Guidotti et al. (1996, p. 176) describe M1 as "a regionally pervasive greenschist event, possibly associated with a deformation that produced a crenulation cleavage." They continue: "... these observations indicate development of a micaceous foliation before any of the later events which attained high grades. This implies at least some early recrystallization of these rocks -- probably syntectonically." There is no firm date for the M1 event other than that it preceded the higher grade M2 metamorphism associated with the emplacement of Devonian plutons. The M2 was a low P (~3 kb) - high T static event that produced in pelitic compositions widespread staurolite- and andalusite-bearing assemblages. However, Guidotti et al. (1996, p. 176) caution that "M2 has been associated with the intrusion of early Devonian granites, but this has been demonstrated only locally for the Lexington pluton to the east." M3, which overprinted M2, was a slightly higher P (~3.5 kb) event that produced sillimanite-bearing assemblages. M3 was also associated with emplacement of Devonian plutons, leading to overlapping contact aureoles and the "regional contact metamorphism" described here. To the south of our field trip area, close to the margin of the Sebago batholith, there was also an M4 event that reached sillimanite-K-spar facies. The most recent review that conveys an idea of this complexity is Guidotti and Johnson (2002).

---

1 We note that many igneous intrusions in Maine are called batholiths in the older literature, but plutons in more recent papers. For this field trip there is no practical difference; we use the terms interchangeably.
Figure 3. Map showing metamorphic isograds and Devonian intrusives in area of the field trip. This map may be correlated with geologic map in Figure 2 by matching pluton shapes. Metamorphic isograds are M2 (solid) staurolite; M2 (dashed) cordierite; M3 staurolite or cordierite; Carb.Kfs.Sil (dash-dot) Carboniferous sillimanite-K-feldspar. Abbreviations as follows: HG, Hallowell pluton group; LB, Lexington batholith; LG, Livermore Falls pluton group; MB, Mooselookmeguntic batholith; NB, Norridgewock batholith (now called Rome pluton); PB, Phillips batholith; SE, Sebago batholith; SoB, Songo batholith; RB, Redington batholith. After Guidotti and Holdaway (1993).
THE SMALLS FALLS FORMATION

The Smalls Falls Formation was named after its type locality: the Smalls Falls Picnic Area in the Sandy River Plantation between Rangeley and Phillips (Fig. 4). See Appendix A for the original description of this formation. The rocks at these falls, possibly named after Jesse Small, a miller who resided in the area around the time of the Civil War, were mapped as Silurian, Wenlock to Ludlow, by Osberg et al. (1968). This age assignment has been recently challenged (albeit with very limited sampling) by detrital zircon data that suggest a younger Early Devonian (Pragian) age (Bradley and O'Sullivan, 2016). On this trip we will offer ideas on what the basin may have looked like during Smalls Falls time, and how the organic-rich sediments responded to the Acadian orogeny. There are many rusty-weathering schists in Maine, but we only consider rocks part of the Smalls Falls Formation if they have the proper stratigraphic position (see Appendix A). Guidotti and Van Baalen (2001) and Van Baalen (2006) provide a general introduction to many aspects of the Smalls Falls Formation, including environmental issues related to arsenic.

The Smalls Falls and correlative formations units crop out for a distance of over 400 km along strike in New England. In far northern Maine, at the limit of exposures that can reasonably be correlated with the Smalls Falls Formation, the rocks have never been deeply buried, and hence the metamorphic grade is very low (lower greenschist and below). Proceeding along strike to the south and west, however, the metamorphic grade rises considerably, because the rocks now on the surface reflect successively deeper levels in the crust. At its type locality, the metamorphic grade is staurolite (for pelitic rocks) and in western Maine and for much of central New Hampshire, in the so-called sillimanite plateau, the metamorphic grade is sillimanite. Far to the southwest of the area of this field trip, on the Massachusetts-Connecticut border, where rocks from deep in the crust are now exposed on the surface, the metamorphic grade is pyroxene-granulite. As a result of this uplift and erosion, a tilted cross section of a large portion of the crust in northern New England presents a natural laboratory where the response of similar rocks to varying degrees of metamorphism can be studied.

Figure 4. Smalls Falls with lower plunge pool. Note rusty nature of outcrops, due to weathering of abundant Fe-sulfide minerals (pyrrhotite and pyrite).
METAMORPHISM OF SULFIDE-RICH PELITIC ROCKS

The petrologic aspects of black shales are mainly related to the relative abundance of Fe-sulfide minerals commonly present, even if deposition did not involve a truly anoxic environment. Several interesting considerations arise due to the presence of Fe-sulfides in shales that undergo metamorphism. In addition to the more narrowly mineralogical features discussed in the next section, these include: (1) impacts on the silicate bulk composition and resultant silicate and opaque mineral assemblages, (2) effects on reactions between the silicates and Fe-sulfides, and (3) impacts on the composition of the fluid phase present during metamorphism. For metapelitic rocks in New England, descriptive aspects of (1) have been considered by Guidotti (1970), Guidotti et al. (1975, 1977), and Robinson et al. (1982). Thompson (1972) developed the theoretical aspects of (2), and details of specific reactions have been presented by Tracy and Robinson (1988). For rocks in New England, (3) has been discussed by Guidotti (1970) and especially Henry (1981).

The impact of sulfide content on the effective silicate bulk composition is most easily appreciated by considering as a first approximation that Fe$^{2+}$ in a rock that is "tied up" in sulfides is unavailable to enter the silicate minerals. As a consequence, minerals that are Fe$^{2+}$-rich relative to Mg will be absent (e.g., garnet, staurolite, chloritoid, and ilmenite) and the minerals that are present will be Mg-rich (e.g., phlogopite instead of biotite). As a result, at a given grade of metamorphism, metamorphosed black shales will, in addition to an Al-silicate, contain various combinations of assemblages of Mg-rich minerals: cordierite, phlogopite or Mg-rich biotite, and Mg-rich chlorite. Where the assemblages of such rocks are merged with assemblages in associated metapelites having bulk compositions of common shale, one can define more precisely the metamorphic mineral facies present (Thompson, 1957). By using this approach, Guidotti et al. (1975) defined rigorously the metamorphic mineral facies for the middle metamorphic grades of the low-P and high-T metamorphism that occurred in western Maine.

As an example, referring to Figure 5, "normal" metapelites having bulk composition $x$ will contain staurolite, whereas sulfide-rich rocks in which Fe is not available to form silicates, have an effective bulk composition $y$ and thus will contain cordierite instead of staurolite. This mineralogical difference explains why Smalls Falls rocks at their type locality, at staurolite grade metamorphism, lack staurolite. The metamorphic facies concept holds that, for a given bulk composition, the mineral assemblage is a function of $P$ and $T$, whereas for a given $P$ and $T$ the mineral assemblage is a function of bulk composition. The Smalls Falls rocks provide an excellent example of this principle.

A further effect of the increased sequestering of Fe by Fe-sulfide is that it reveals a change in the Ti-saturating phase that occurs in the rocks, resulting in the formation of rutile instead of ilmenite. Accordingly, whereas Smalls Falls rocks may contain rutile, adjacent rocks of the Perry Mountain Formation contain ilmenite. The approach for a theoretical understanding of this change has been developed by Thompson (1972). A key aspect of this development is that as metamorphic grade increases, the pyrite originally present in black shales reacts with the silicate minerals and is converted to pyrrhotite, by the model reaction:

$$\text{FeS}_2 + \text{Fe} \rightarrow 2\text{FeS}$$

As a consequence, an even larger fraction of the Fe in the rock becomes sequestered into sulfide and so is unavailable to enter the silicates. Thus, the bulk composition and mineral assemblage aspects described above become even more amplified, and as a result the newly formed silicates tend to be Mg-rich. Details of some of these sulfide-silicate reactions are presented by Tracy and Robinson (1988). Such reactions explain why pyrrhotite predominates in metapelites, and why pyrite persists only in rocks that at the sedimentary stage contained particularly large modal amounts of pyrite, e.g. $>> 10$-15 vol %. Overall, it becomes clear that the specific Fe-sulfide contained in a metamorphosed black shale is a function of both original S content of the rock, and metamorphic grade.

Note also, following Thompson (1972), that the nature of the Ti-saturating phase in sulfide- and graphite-rich rocks changes during metamorphism, from ilmenite to rutile. Once again, the siderophile behavior of Fe is responsible.
Finally, the sulfide-silicate equilibria typically involve the gas species CO₂, H₂S, CH₄, and SO₂ as well as H₂O. Hence, these gases impact the composition of the metamorphic C-O-H-S fluid phase, especially with regard to its deviation from being solely H₂O. The study by Henry (1981) was especially interesting in this context, in showing that sulfidation equilibria affected the composition of the fluid phase such that on the scale of only cm, continuous reactions among silicates were systematically facilitated, thereby demonstrating very local, buffering control by the rocks on the fluid composition.

![AFM diagram after Thompson (1957) for metapelites at upper staurolite zone.](image)

Figure 5. AFM diagram after Thompson (1957) for metapelites at upper staurolite zone. A, Al₂O₃; F, FeO; M, MgO; KAS, either kyanite, andalusite, or sillimanite; St, staurolite; Crd, cordierite; Bt, biotite; Ms, muscovite; Qtz, quartz.

WHOLE-ROCK GEOCHEMISTRY

Whole-rock analyses for major, minor, and trace elements (including rare earth elements, REE) were acquired on 21 samples of black shale and black schist from the Smalls Falls Formation. Most samples were obtained from small cores (drilled in 2015) in order to minimize oxidation and weathering. Basic statistics are listed in Table 1, together with data for average black shale, for comparison. The analyzed samples, from areas of lower to upper greenschist grade and amphibolite grade, were collected in order to evaluate effects of metamorphism on bulk compositions of these strata, and to constrain the redox state of bottom waters and pore fluids during sedimentation (Slack et al., 2016). Distinctive features of these compositions include generally low MnO and P₂O₅ (<0.1 wt % each), plus low TOC (<1.5 wt %) but high total S (1.2-9.7 wt %). Despite the generally high S contents, base and related transition metals are uniformly low (Co <40 ppm, Cu <65 ppm, Ni <85 ppm, Pb <35 ppm, Zn <135 ppm). Relative to median or average contents of black shales worldwide (Table 1), the Smalls Falls samples have significantly lower TOC and V, moderately lower Cu and Zn, and higher Fe₂O₃, Sn, and As.
Table 1. Basic statistics for bulk compositions of black shale and black schist from the Smalls Falls Formation, northern and western Maine

<table>
<thead>
<tr>
<th>Oxide/Element</th>
<th>Mean</th>
<th>Stdev</th>
<th>Minimum</th>
<th>Maximum</th>
<th>Avg Black Shale*</th>
</tr>
</thead>
<tbody>
<tr>
<td>Al₂O₃ (wt %)</td>
<td>13.89</td>
<td>3.44</td>
<td>7.35</td>
<td>19.65</td>
<td>13.23</td>
</tr>
<tr>
<td>Fe₂O₃</td>
<td>7.16</td>
<td>2.30</td>
<td>2.24</td>
<td>12.38</td>
<td>2.86</td>
</tr>
<tr>
<td>MnO</td>
<td>0.06</td>
<td>0.06</td>
<td>0.01</td>
<td>0.31</td>
<td>0.05</td>
</tr>
<tr>
<td>MgO</td>
<td>2.30</td>
<td>0.77</td>
<td>0.85</td>
<td>3.23</td>
<td>1.16</td>
</tr>
<tr>
<td>CaO</td>
<td>1.85</td>
<td>0.88</td>
<td>0.42</td>
<td>3.78</td>
<td>2.10</td>
</tr>
<tr>
<td>K₂O</td>
<td>3.06</td>
<td>0.78</td>
<td>1.33</td>
<td>4.34</td>
<td>2.41</td>
</tr>
<tr>
<td>P₂O₅</td>
<td>0.06</td>
<td>0.04</td>
<td>0.00</td>
<td>0.11</td>
<td>0.32</td>
</tr>
<tr>
<td>TiO₂</td>
<td>0.68</td>
<td>0.15</td>
<td>0.33</td>
<td>0.92</td>
<td>0.33</td>
</tr>
<tr>
<td>TOC**</td>
<td>0.82</td>
<td>0.24</td>
<td>0.43</td>
<td>1.48</td>
<td>3.20</td>
</tr>
<tr>
<td>CO₂</td>
<td>0.38</td>
<td>0.55</td>
<td>0.01</td>
<td>1.45</td>
<td>n.r.</td>
</tr>
<tr>
<td>Total S</td>
<td>3.28</td>
<td>1.82</td>
<td>1.22</td>
<td>9.67</td>
<td>n.r.</td>
</tr>
<tr>
<td>As (ppm)</td>
<td>92.5</td>
<td>84.3</td>
<td>30</td>
<td>280</td>
<td>30 ± 3</td>
</tr>
<tr>
<td>Au (ppb)</td>
<td>9.1</td>
<td>9.2</td>
<td>5</td>
<td>38.0</td>
<td>7 ± 1</td>
</tr>
<tr>
<td>Ba</td>
<td>446</td>
<td>156</td>
<td>187</td>
<td>801</td>
<td>500 ± 20</td>
</tr>
<tr>
<td>Ce</td>
<td>68.6</td>
<td>22.3</td>
<td>25.7</td>
<td>107</td>
<td>58 ± 3</td>
</tr>
<tr>
<td>Co</td>
<td>22.6</td>
<td>7.48</td>
<td>12.6</td>
<td>38.6</td>
<td>19 ± 1</td>
</tr>
<tr>
<td>Cr</td>
<td>69.5</td>
<td>17.7</td>
<td>30</td>
<td>100</td>
<td>96 ± 3</td>
</tr>
<tr>
<td>Cu</td>
<td>43.5</td>
<td>11.6</td>
<td>24</td>
<td>64</td>
<td>70 ± 3</td>
</tr>
<tr>
<td>La</td>
<td>31.9</td>
<td>10.4</td>
<td>11.7</td>
<td>48.9</td>
<td>28 ± 1</td>
</tr>
<tr>
<td>Mo</td>
<td>28.7</td>
<td>17.8</td>
<td>3</td>
<td>61</td>
<td>20 ± 1.5</td>
</tr>
<tr>
<td>Ni</td>
<td>56.7</td>
<td>14.5</td>
<td>34</td>
<td>82</td>
<td>70 ± 2</td>
</tr>
<tr>
<td>Pb</td>
<td>19.5</td>
<td>6.24</td>
<td>11</td>
<td>34</td>
<td>21 ± 1</td>
</tr>
<tr>
<td>Sb</td>
<td>1.6</td>
<td>0.76</td>
<td>0.1</td>
<td>2.6</td>
<td>5.0 ± 0.5</td>
</tr>
<tr>
<td>Sc</td>
<td>13.8</td>
<td>4.01</td>
<td>6</td>
<td>21</td>
<td>12 ± 1</td>
</tr>
<tr>
<td>Sn</td>
<td>5.9</td>
<td>7.44</td>
<td>1</td>
<td>29</td>
<td>3.9 ± 0.3</td>
</tr>
<tr>
<td>Th</td>
<td>11.4</td>
<td>3.28</td>
<td>4.2</td>
<td>17.4</td>
<td>7.0 ± 0.4</td>
</tr>
<tr>
<td>U</td>
<td>7.49</td>
<td>3.86</td>
<td>2.87</td>
<td>15.2</td>
<td>8.5 ± 0.8</td>
</tr>
<tr>
<td>V</td>
<td>114</td>
<td>39.4</td>
<td>61</td>
<td>217</td>
<td>205 ± 15</td>
</tr>
<tr>
<td>Y</td>
<td>26.7</td>
<td>7.86</td>
<td>13.8</td>
<td>45.6</td>
<td>26 ± 1</td>
</tr>
<tr>
<td>Zn</td>
<td>65.5</td>
<td>30.4</td>
<td>6</td>
<td>132</td>
<td>130 ± 10</td>
</tr>
<tr>
<td>Zr</td>
<td>186</td>
<td>56.0</td>
<td>97.3</td>
<td>331</td>
<td>120 ± 5</td>
</tr>
</tbody>
</table>

*Major elements including TOC (medians) from Vine and Turtelot (1970);
MnO, P₂O₅, trace elements, and REE (means) from Ketris and Yudovich (2009)

**Includes organic C + graphitic C
n.r., not reported; all analyses by fusion ICP-AES (no data for SiO₂ or Na₂O)
In Figure 6 the high S/non-carbonate C ratios of the Smalls Falls samples are especially distinctive. Such uniformly low contents of non-carbonate C suggest one of three processes, as shown in the inset (1) hydrothermal S overprint, (2) thermal or metamorphic loss of C, or (3) euxinic marine deposition. Absence in all samples (and outcrops) of epigenetic, sulfide-bearing veins rules out the first possibility. The third possibility is considered unlikely because all of the Mo concentrations (Table 1) are below the threshold of 100 ppm that reflects persistently euxinic (sulfidic) conditions (Scott and Lyons, 2012). A caveat in this context is the presence of uniformly low Mo in modern sediments of the euxinic Black Sea, but this is linked to the isolated nature of this water body, which is essentially cut off from recharge by Mo-rich oxygenated waters of the open ocean (including the Mediterranean). Given these constraints, the most logical explanation for the distribution of data on Figure 6 is a loss of C via thermal or metamorphic effects. The high temperatures (and pressures) of the greenschist- to amphibolite-grade metamorphic conditions of the samples are consistent with this mechanism, but a loss of C (and some metals) after deposition but prior to metamorphism must also be considered. Scott et al. (2017) have recently determined a loss of ca. 30% organic C in one sample from the Late Devonian-Early Mississippian Bakken Formation in North Dakota, attributed to thermal effects during late diagenesis at or above the thermal window for oil generation (~60-120°C). An even greater loss of C may have occurred in the Smalls Falls Formation, reflecting heating during both late diagenesis and multiple episodes of regional metamorphism. Also relevant to this issue are the high contents of V, Ni, and to a lesser extent Mo that occur in many crude oils (e.g., Lewan, 1984; Ventura et al., 2015), which are seldom discussed in studies of black shale geochemistry, but are important to consider because the likely removal of these metals (in unknown proportions) during oil generation means that common redox proxies such as V/(V+Ni) must be used with caution in applications to ancient black shales such as those of the Smalls Falls Formation. This caution is supported by the low mean V content of 114 ± 39.4 ppm in our samples (Table 1), relative to that of average black shale worldwide (205 ± 15 ppm), consistent with a loss of substantial V during oil generation and/or later regional metamorphism of Smalls Falls rocks.

Figure 6. Organic + graphitic C vs Total S plot of bulk compositions of black shale and schist from the Smalls Falls Formation. All samples have high ratios of S/(organic+graphitic C), consistent with thermal and/or metamorphic loss of C (see text). Inset from Lyons et al. (2000).
Numerous geochemical proxies have been used for evaluating the redox states of bottom waters and pore fluids during sedimentation. However, in many cases these proxies yield conflicting results, for a variety of reasons. Among various commonly applied proxies, one of the most robust is a plot of total organic C (including graphitic C) vs Mo (Fig. 7). Based on work by Scott and Lyons (2012), the relatively low Mo contents of the Smalls Falls samples—despite mostly high to very high total S (Fig. 6)—suggest that the bottom waters during sedimentation were mainly suboxic to anoxic, and lacking persistently euxinic (sulfidic) conditions given the absence of Mo concentrations >100 ppm. It is possible that samples having <25 ppm Mo record oxic bottom waters (Scott and Lyons, 2012), but this could also reflect dilution by quartz or other clastic components. Importantly, the above interpretations assume that little if any Mo was lost from the sediments during oil generation, and/or during later metamorphism in the region.

Figure 7. Organic + graphitic C vs Mo plot of bulk compositions of black shale and schist from the Smalls Falls Formation. Relatively low Mo contents suggest a range of redox conditions in bottom waters, from suboxic to anoxic (see text). Trends for anoxic to euxinic/sulfidic sediments of the Black Sea, Framvaren Fjord (Norway), Cariaco Basin (offshore Venezuela), and Saanich Inlet (British Columbia) are from Algeo and Lyons (2006). Mo limit for persistently sulfidic/euxinic conditions is from Scott and Lyons (2012).

PROVENANCE AND ORIGIN OF THE SMALLS FALLS FORMATION

The Smalls Falls Formation originated as organic-rich clastic sediments that were deposited into a basin whose bottom waters underwent seemingly abrupt changes at the beginning and end of Smalls Falls time. There is good evidence that the source region for these sediments probably lay to the north and west (present day coordinates), but the cause of the abrupt changes in bottom water chemistry is more speculative. We consider some modern-day analogues in the context of Silurian tectonics in order to suggest a likely scenario.
Insights into the provenance of the original sediments come from plots of immobile trace elements. Figure 8 shows that, with one exception, all samples have Th/Sc ratios >0.5, which suggests predominantly felsic sources. The trend and range of Zr/Sc ratios is consistent with sediment recycling and zircon addition. A ternary plot of Th-Sc-Zr/10 (Fig. 9) shows that data for all samples from greenschist-grade outcrops have bulk compositions that fall in the field for continental arcs. A source from a predominantly felsic arc is thus proposed, an interpretation supported by the elevated Sn contents of some of the analyzed samples, which range up to a maximum of 29 ppm (Table 1). For comparison, average Sn in black shale globally is 3.9 ± 0.3 ppm (Ketris and Yudovich, 2009). These mostly high Sn contents, together with the elevated Th/Sc ratios, suggest an evolved felsic plutonic and/or volcanic source within a former continental arc. In an earlier study, Cullers et al. (1997) reached a similar conclusion for the Smalls Falls Formation, on the basis of trace element ratios and REE patterns, albeit for a small data set (two samples).

Figure 8. Plot of Zr/Sc vs Th/Sc of bulk compositions of black shale and schist from Smalls Falls Formation. Relatively high Th/Sc ratios of >0.5 suggest predominantly felsic sources; Zr/Sc ratios >30 imply zircon addition via sediment recycling. Plot from McLennan et al. (1993).
A source region lying to the northwest is considered most likely (Bradley and O’Sullivan, 2006). Bradley and Hanson (2002) reported predominantly northwest paleocurrent directions for the Smalls Falls Formation. Given the early Ludlow graptolite age of this unit, a Middle Silurian or older arc is required. Potential candidates include Silurian felsic igneous rocks to the northwest, such as the 443 Ma Attean pluton in northwestern Maine (Gerbi et al., 2006) and the Ascot Complex in southeastern Québec. Older Ordovician arcs of this region also are plausible sources, but are undermined by the lack of Ordovician ages in the detrital zircon histogram for the Smalls Falls Formation (Bradley and O’Sullivan, 2006).

Several possible modern analogues can be invoked for the basin—or basins—in which the Smalls Falls sediments accumulated. Certain geochemical signatures (e.g., generally low MnO and locally elevated Mo concentrations) suggest that the basin(s) had suboxic to periodically anoxic bottom waters. Euxinic (sulfidic) conditions are implied, at least locally, by the presence of very small (~5-6 μm) pyrite framboids in one sample from northern Maine (Slack et al., 2016). However, it is important to note that these redox interpretations are greatly constrained by the effects of high temperatures reached prior to and during regional metamorphism, possibly involving the selective removal of unknown quantities of trace elements such as Ni, V, and possibly Mo. Another important point is that the Smalls Falls Formation extends for over 400 km along strike (Osberg et al., 1985), although not in continuous outcrop; this discontinuity may reflect tectonic dismemberment from an original contiguous basin. Alternatively, non-deposition, a Devonian tectonic overprint, or deposition within separate basins may be the explanation². However, the great strike length of the Smalls Falls Formation cannot be attributed solely to tectonic events prior to or during the Acadian orogeny, hence any proposed analogue involving a modern anoxic basin should have similar maximum dimensions. Few candidates are known that exclude sites of oceanic spreading centers floored by basalt (e.g., Red Sea) or that have relatively small diameters (e.g., 150 km for Cariaco Basin north of Venezuela). The large Black Sea (~1200 km diameter) is also ruled out as an analogue because it

² We cannot rule out diachronous deposition into separate basins. The detrital zircon data from Bradley and O’Sullivan (2016) allow this interpretation.
has a restricted setting as an intracontinental rift basin in which the bottom waters are uniformly euxinic/sulfidic (more reducing than anoxic). We suggest that a modern analogue could be the Southern California Borderland, west of Los Angeles, where seven separate suboxic to anoxic basins and intervening rises in total extend over a distance of nearly 400 km (Christensen et al., 1994; Chong et al., 2012). The sediments in most of these basins have several percent or more of organic C, thus being similar to inferred protoliths of the black shales and black schists that make up the Smalls Falls Formation.

More detailed work on the bulk geochemistry of this formation, both stratigraphically at previously sampled sites and at new sites, could help to refine our understanding of how the sediments of this distinctive Paleozoic unit accumulated, and the effects of subsequent diagenesis and regional metamorphism.

The question of what caused abrupt changes in bottom water chemistry, from fully oxygenated to oxygen-deprived, locally anoxic, must be considered in light of what we know of Silurian tectonics in the region. Moench (2006, p. 84) wrote "The abrupt contact between the Perry Mountain and Smalls Falls Formations marks a change from aerated to strongly reducing conditions, probably the result of abrupt subsidence [sic] that produced a closed, euxinic basin." And then, re the sharp contact between Smalls Falls and Madrid, "These features signal the arrival of energetic, probably oxygenated currents that ended the euxinic conditions of Smalls Falls deposition." Reusch and van Staal (2012, p. 254) in turn suggested that "A mechanism for the black shale deposition invokes isolation of the forearc basin as a result of collision-induced uplift of the adjacent, partly subjacent subduction/collision complex (Dickinson and Seeley, 1979). The Central Maine–Matapedia forearc experienced a general shallowing, due to a combination of basin infilling and uplift."

We suggest four processes that may have been at work, either individually or in combination:

1. rising sea level;
2. increased productivity, including a widespread algal bloom that enhanced eutrophication, possibly triggered by changes in marine circulation due to tectonic events;
3. changing marine circulation that greatly limited, or eliminated, recharge with oxic seawater due to various factors (e.g., newly formed submarine sill or horst as in Southern California Borderland and Cariaco Basin), or different patterns of seawater circulation as in Arabian Gulf; and
4. a change to very slow sedimentation with minimal detrital influx that produced a 'condensed section.'

For the lower contact of the Smalls Falls Formation, we suggest that uplift accompanied by extension could have produced horsts and grabens (similar to the modern-day North Sea). The grabens would be an ideal site for black shale deposition, as in the Southern California Borderland discussed above. Another mechanism to explain the abrupt transition across the Perry Mountain Formation–Smalls Falls Formation contact is the mid-Silurian emergence of the Brunswick subduction complex (BSC) on the southeastern margin of the Central Maine Trough (Reusch and van Staal, 2012). This tectonic event is critical to understanding the Silurian geology of Maine, as it is the basis for the Salinic system that dominated the Silurian, and culminated with closure of the Tetagouche-Exploits basin on the Dog Bay Line (Reusch and van Staal, 2012). The BSC emergence may have affected sedimentation on the northwestern margin of the basin in several ways: First, it would have modified internal circulation by restricting the exchange with the open ocean to the southeast. Second, extensive erosion as documented by strata of the Vassalboro Group implies nutrient loading and eutrophication. Third, an emergent landmass might have shed a buoyant freshwater lid across the narrowing Central Maine Trough to the northwest, thus further amplifying eutrophication. An alternative mechanism for the onset of black shale deposition entails the formation of isolated basins within an overall extensional upper plate setting (Tremblay and Pinet, 2005; Reusch and van Staal, 2012).

The transition across the Smalls Falls Formation–Madrid Formation contact also begs for an explanation. The map pattern suggests the possibility of sedimentary interfingering, but we believe the map pattern is tectonic and not depositional, as Boone et al. (1970) noted that the contact is sharp and conformable. A simple mechanism is an increase in sedimentation rate, such that dilution by calcareous and siliciclastic sediment terminated black shale deposition in advance of the Acadian orogeny. However, the calcareous
nature of the Madrid, and in fact the calcareous uppermost portion of the Smalls Falls, also requires consideration of a source region for the carbonates. Offshore reefs from the eastern approaching landmass are one possibility; another is a carbonate bank akin to the Brionçonnais zone in the Alps (e.g., Hsu, 1994). Increased sedimentation from the source region evidently brought black shale deposition to an end, as reflected by the siliciclastic strata of the Madrid Formation.

ACKNOWLEDGMENTS

The authors would like to particularly acknowledge the contributions of Gary Boone and the late Bob Moench, who led the way for decades in efforts to unravel the enduring mysteries of this sedimentary basin. We also thank Wally Bothner for helpful comments on the manuscript.

APPENDIX A - STRATIGRAPHY

(a) Greenvale Cove Fm. [Sg], Silurian, Lower Llandoverian? Mainly a slightly rusty weathering, light grey, laminated and thinly bedded feldspathic meta-sandstone and meta-siltstone, commonly calcareous. The thickness is 200 m. Originally mapped as Upper Ordovician by Boone et al. (1970), Moench et al. (1995) mapped it as Silurian.

(b) Rangeley Fm. [Sra, Srb, Src], Llandoverian to Wenlockian? A thick formation subdivided into three members, A, B, and C with a total thickness of about 3 km in its type area near Rangeley. A coarse conglomerate followed by moderately rusty-weathering, thin to thick, interbedded metapelite and impure quartzites; commonly showing graded interbeds. Some massive arkosic sandstone and thin conglomerate beds throughout. Member A, containing coarse boulder conglomerate, is 1200 m thick, and correlates with the Clough Quartzite of western New Hampshire. Rangeley A conformably overlies the Greenvale Cove. Member B has a similar thickness, and represents a generally finer grained, more distal facies. Member C is yet finer-grained, with a thickness of about 600 m. The rusty weathering reflects the presence of minor-up-to-5% pyrrhotite. Graphite and ilmenite are other typical opaque accessory minerals. Many Rangeley rocks weather to a characteristic brick-red color.

(c) Perry Mountain Fm. [Sp], Silurian, Wenlockian? Thin to thick, graded interbeds of metapelite and fairly pure quartzite. Ilmenite is the typical opaque mineral, with pyrrhotite and graphite sparse or absent in some samples. Hence, outcrops have a grey weathering surface. A very few samples have magnetite present. Thickness is about 400 m. The boundary between this unit and the overlying Smalls Falls Fm. is sharp and conformable.

(d) Smalls Falls Fm. [Ssf], Silurian, middle Wenlockian to lower Ludlovian?.

As originally defined by Boone et al. (1970),

The Smalls Falls Formation is a rusty-weathering unit composed of black sulfidic metashale cyclically interbedded with sulfidic quartz-rich metasandstone. The upper few hundred feet is calcareous. Thick graded beds of commonly calcareous quartz granule metaconglomerate are abundant in northern outcrops, particularly near the faults northwest of Madrid (Moench, 1970). The Smalls Falls Formation is about 2500 feet thick west and south of Madrid, but thins and wedges out northward. The contact between the Smalls Falls and Madrid Formations is sharp and conformable.

The age of the Smalls Falls has recently been challenged: see discussion in trip text. Thin to thick interbeds of calcareous quartzite and very rusty-weathering metapelite characterize the formation. The pelitic beds especially have very abundant amounts of pyrrhotite and graphite, typically 5-10% and 1-3 modal%, respectively, with rutile as the Ti phase rather than ilmenite. Moreover, due to abundant muscovite and a lesser amount of Mg-rich biotite, the pelitic samples are basically white in color on unweathered surfaces. Typical metapelitic minerals such as garnet and staurolite are absent, but Mg-rich cordierite is common. On this field trip we will encounter the graptolite-bearing Parkman Hill Fm at Stop 1, correlated with
Smalls Falls. In Western New Hampshire, the well-known arenaceous Clough Quartzite (Billings, 1956) is considered a time-equivalent, proximal facies of the more distal Smalls Falls Fm.

(e) Madrid Fm. [DSm], Upper Silurian to Lower Devonian. Medium to thick, immature interbeds of biotite granofels and calc-silicate granofels, with actinolite common at lower metamorphic grades, diopside at higher. The Madrid formation is correlated with the Fitch Formation of western New Hampshire. The age of the Madrid is somewhat in doubt: originally it was considered Silurian, but now (Moench et al., 1995) it is mapped as Devonian. See discussion in Moench (2006). The thickness of the Madrid is about 600 m.

(f) Carrabassett Fm. of the Seboomook Group [Dc], Devonian? This is the basal unit of the Seboomook Group, whose type locality is in the Little Bigelow Mountain quadrangle (Boone, 1973). We will not visit this formation on this trip. Variable thickness interbeds of metapelite and impure feldspathic quartzites, commonly showing graded bedding. The accessory opaque minerals are ilmenite, graphite, and minor to moderate amounts of pyrrhotite, the latter in some cases being reflected by moderately rusty weathering surfaces.

(g) Seboomook Fm. [Ds], Devonian. Thickly bedded turbidites with primary sedimentary structures preserved, showing an easterly source. Seboomook Group is correlated with the well-known Littleton Formation of New Hampshire.
Starting Point: Coburn Park, a public park along the Kennebec River in Skowhegan. 44°46.3 N, 69°42.6 W. The park is about 0.5 miles east of the center of town along US 2 (Water St.); it has a gazebo and a Porta Potty. Muffin alert: there is a fine bakery on Water St. called The Bankery! There is also a Dunkin Donuts on Water St.

From the park entrance, walk 500 ft. east on Water St. to a trail that enters the woods on the right at a sign "433". Follow obvious path through the woods along the river to large stream outcrops on the north shore of the Kennebec River. Watch out for poison ivy here!

STOP NO. 1, Great Eddy. (30 minutes) We are at the Great Eddy of the Kennebec River below the Skowhegan Dam (formerly Falls). Exposed here are the steeply dipping, rusty weathering rocks mapped as the Silurian Parkman Hill Formation by Ludman (1977), but remapped as the eastern facies of the Smalls Falls Formation on the Moench and Pankiwskyj (1988b) map. Here we also see the underlying metagreywacke and pelite of the Silurian Sangerville Formation, in the same stratigraphic position as the Perry Mt. Formation. The contact between these formations is exposed here. We are on the north limb of the Currier Hill Syncline; the tops point south, towards the river. Therefore the north limb of the syncline is overturned. This syncline was first recognized by Charlie Guidotti (unpublished mapping, 1964). The Parkman Hill ~ Smalls Falls rocks, closest to the river, are rusty weathering schists, at biotite grade of metamorphism. Here the sulfides include pyrrhotite (dominant) and pyrite (subsidary), as well as minor amounts of chalcopyrite and sphalerite. Table 4 contains analyses of these sulfides. Higher up and in the woods are the Sangerville ~ Perry Mt. rocks.

After examining the outcrops, return to the vans in Coburn Park. Exiting the park, reset odometers to 0.0 as you turn left onto Water St.

0.0 Exit Coburn Park, turn left onto Water St.

0.2 Bear right onto High St., pass through a traffic light.

0.5 At second traffic light turn right onto Madison St. (US 201). Move into the left lane.

0.8 Turn left onto Spring St. with a car wash on the corner.
0.9 Turn right onto Russell Rd. at old warehouse, then pass Skowhegan fairgrounds at 1.4 miles.

6.2 Carefully turn left onto Rt. 148 at crossroads with unique radar-assisted stop sign.

7.7 Note distant views to the north towards the Saddleback massif, underlain by the 407 Ma Redington Pluton (Solar et al. (1998).

9.6 Pass straight through the town of Madison and its now-closed paper mill, cross Kennebec River.

10.3 Turn right onto US 201A northbound. Continue along the scenic Kennebec River. Note sign commemorating the route of Benedict Arnold’s invasion of Quebec in 1775.

14.9 Cross bridge as you enter village of North Anson, turn left onto Rt. 16, noting outcrops of Madrid Formation in the river. This is the Carrabassett River. We will see more of these rocks at Stop 3.

16.6 Note antique wooden observation tower in cornfield.

17.9 Note more roadside outcrops of Madrid Formation on right.

20.3 Pull into small parking lot on the left side of highway, taking care for high speed traffic on this road. There is parking for 6-8 vehicles here. 44°53.6 N, 69°59.4 W. Cross road very carefully and walk back along highway 300ft to:

STOP NO. 2, Carrabassett River Valley (20 minutes) Here we have 3m high vertically dipping, rusty weathering outcrops of the Smalls Falls Formation, and at the eastern end of the outcrop, massive Perry Mt. Formation lacking the normal repeating quartzite beds. This locality is close to the western margin of the Lexington Batholith, dated at 404±2 Ma, Solar et al. (1998). We are within the contact aureole, with nearby calcisilicate rocks of the Madrid Formation containing diopside, and farther out, actinolite. More pelitic rocks might contain garnet and staurolite, except that in the sulfide-rich Smalls Falls there is no garnet! The contact between Smalls Falls and Perry Mt. is exposed in this outcrop. Look for garnet and staurolite in the Perry Mt. rocks - so far we have not found them, but they may be there.

Cross road very carefully and return safely to the vans in the parking lot. Continue westbound on Rt. 16 for a short distance, preparing for immediate left turn.

20.4 Turn left onto Katie Crotch Rd. (derivation unknown). Follow along the Carrabassett River.

22.4 At stop sign, travel straight ahead, joining Rt. 146 in the village of East New Portland. Cross the river on a concrete bridge that is in poor repair.

22.7 At far end of bridge, turn left onto dirt road and park beside river. 44°53.9 N, 70°01.3 W. Scramble down to the river at:

STOP NO. 3, East New Portland (30 minutes) Extensive exposures of the Devonian Madrid Formation form the river bottom here in an area of rapids. These calcareous rocks are gently dipping to the north, as we are on the north limb of the Strickland Hill Anticline, containing Smalls Falls rocks at its core. The rock type is biotite granofels and calcisilicate granofels. As we are still close to the western margin of the Lexington Batholith, the metamorphic grade here is actinolite. A prominent granitic dike crosscuts the Madrid Formation in the river bed. This dike is a two-mica granite, probably anatectic. Look for tourmaline, garnet, and other aluminous phases in the granite. Although there is no date for this dike, it is likely related to emplacement of the Lexington Batholith at 404±2 Ma (Solar et al., 1998), and if so, places an important constraint on the depositional age of the Madrid Formation.

Return to the vans, turn around and head back to the paved road at the end of the bridge.
22.8 Continue straight ahead, following Rt. 146 to the town of New Portland, where road winds around a bit. Note sign in village for the Wire Bridge, a local curiosity.

27.2 At stop sign, turn left onto Rt. 27 southbound, a high speed road. Watch for trucks!

31.4 Note low, dark outcrops of Madrid Formation on left.

33.3 Pass the "Maine Wood Turning" factory, that produces handles, dowels, and other parts as forest products of the North Woods. At the south end of the factory turn right onto Rt. 234 towards Strong.

35.9 Strong town line.

38.6 At bottom of a steep hill as we enter the village of Strong, turn into parking lot of the Harry Gordon Lumbering Co., and park. 44°48.2 N, 70°13.0 W. Carefully walk back uphill 200 ft. to some low outcrops at roadside. Watch out, this is a blind corner, so keep off the road! Fortunately there is relatively little traffic on this road.

**STOP NO. 4, Strong** (10 minutes) We may skip this stop if the group size is too large. The outcrops here are very graphitic, steeply dipping rocks that have been confused with Smalls Falls Formation. The metamorphic grade here is garnet but of course there is no garnet due to the high sulfide content of the rock. There are horizons of pyrrhotite that is only weakly magnetic. In pyrrhotite, a non-stoichiometric compound, the magnetic susceptibility is related to the Fe concentration, with lower concentrations more strongly attracted to magnetic fields. Our structural position here is just west of the Winter Brook Fault (back up the hill, but not exposed), marking the boundary of the Rumford allochthon (Moench and Pankiwskyj, 1988a). Temple Stream rocks are lithologically similar to Smalls Falls rocks, but lie in a higher structural setting.

Return safely to the vans, watching out for traffic on the road as vehicles descend the hill.

Continue west on Rt. 234 into the village of Strong.

38.9 Turn left onto Rt. 145 in center of village.

39.3 Cross Sandy River on modern bridge, bear right and at a yield sign join Rt. 4 towards Rangeley. The highway crisscrosses the Sandy River as it passes through the towns of Phillips and Madrid.

53.1 Note the tiny village of Madrid, with an abandoned hotel and Mobil station. On the right Reeds Mill Rd. crosses the river at a bridge. Under this bridge is the type locality of the Madrid Formation, but as we saw excellent exposures of this formation at Stop 3 we will not stop here. 44°51.8 N, 70°27.7 W for reference, at the former Mobil gas station.

55.1 Note 5-m-high road cuts of Smalls Falls rocks on right. Slow down, approaching a left turn that comes up suddenly.

56.0 Turn left into the Smalls Falls Picnic Area (sign) and park vehicles. 44°51.4 N, 70°30.9 W.

**STOP NO. 5, Smalls Falls Picnic Area** (90 minutes, including lunch) The scenic picnic area (with bathrooms) is also the type locality of the Smalls Falls Formation, where the Sandy River descends over a series of cascades. This is a no-hammer stop! There are several waterfalls here, the highest of which is just above the picnic area. During the warm period of the year this is a popular swimming hole. The river has forced its way through the weakest portions of the rusty weathering Smalls Falls Formation. We will cross the footbridge and ascend the trail that follows the right bank of the river, with chain link fences protecting the most dangerous places. On smoothed surfaces look for a variety of pseudomorphs. In the bottom of an enlarged pothole there are some very well displayed pseudomorphs after cordierite. They consist of coarse-grained muscovite, Mg-rich
chlorite, and phlogopite. Please do not hammer on them - take a picture instead! Higher up, look for large chiastolite crystals - except that, as shown by Guidotti and Cheney (1976) they are now pseudomorphs composed mainly of margarite. As your eyes become more focused on the details of the textures present, you will see that other 2 cm pseudomorph knots are present. These consist of aggregates of chlorite and phlogopite after cordierite. The original chiastolite and cordierite formed during M2 but during M3 they have been pseudomorphed (ibid.)

Continuing along the right bank of the river, within sight of a wooden bridge, there are smoothed surfaces containing examples of submarine soft sediment deformation (see photo).

Cross the wooden bridge on a dirt road that joins the paved highway. If time permits, we will follow the former route of Rt. 4, now grass, to a set of 5 m high outcrops at the boundary between the Smalls Falls Formation and the underlying Perry Mt. Formation. The contact lies in the woods and is unfortunately not exposed here. Here the rock has thin to thick, graded interbeds of metapelite and fairly pure quartzite. Perry Mountain itself, which gives its name to the Perry Mt. Formation, lies just across Rt. 4. We will see the Perry Mt. Formation again at Stop 8.

Return to the vans via the trail we used during the ascent.

56.3 Exit the parking lot and turn left, westbound, on Rt. 4 towards Rangeley. Note large roadcuts of Smalls Falls rocks on the right.

58.8 Appalachian Trail crossing.

61.4 Note 3-m-high outcrops of Perry Mt. Formation on the left.

61.8 Pass Long Pond on the left. The bedrock geology here is now Rangeley C Formation, the upper, more distal portions of the Rangeley Formation. We will see these rocks again at Stop 7.

64.8 Note sign for Rangeley Lakes State Park, as road ascends to a height of land. Slow down.

65.7 Near height of land, at 3-m-high roadcuts on the right, turn left into a dirt driveway, taking care for oncoming traffic. Park here. 44°55.9 N, 70°37.4 W.

**STOP NO. 6, Rangeley Conglomerate** (20 minutes). The roadcuts here display the spectacular polymictic conglomerate of the Silurian Rangeley Formation. This stop is the same as Stop 4 of NEIGC 1970 Trip A1 (Moench and Boudette, 1990), as well as Stop 1 of NEIGC 2006 Trip C4 (Van Baalen, 2006). The rocks are the Rangeley A Formation, close to their lower contact with Silurian (?) and pre-Silurian rocks to the west. The rock here includes a spectacular polymictic boulder conglomerate with clasts derived from
source regions to the northwest of here. At first glance one might assume this to be a basal conglomerate above the Taconic Unconformity, but Moench and Boudette (1970) argue that the sequence is in fact conformable. Some of the clasts are coarse-grained igneous rocks. There are also graded beds here, indicating tops to the east. In general, the lowest part of the Rangeley Formation (Rangeley A) contains the coarsest conglomerates and sandstones; the formation fines upward and eastward into deeper portions of the basin into which it was deposited, probably as a result of a marine transgression (see Fig. 1).

Cross back to the vans, watching out for traffic. The dirt loop where the vans are parked rejoins Rt. 4 in a short distance. Turn right on Rt. 4, back the way we came.

66.4 Turn right onto South Shore Rd., noting sign for Rangeley Lakes State Park. Follow this road along the shore of Rangeley Lake, with occasional views and many lake camps.

74.0 Turn left at stop sign intersection onto Rt. 17 south, towards Rumford.

75.5 Pass scenic overlooks on left, at which we will stop only if time permits.

80.9 At height of land, pull into well-marked scenic overlook on right and park, 44°50.2' N, 70°42.6' W.

**STOP NO. 7, SCENIC OVERLOOK** (10 minutes) This scenic overlook with expansive views to the west over Mooselookmeguntic Lake and other distant lakes, underlain by the 388 Ma Mooselookmeguntic Pluton (Solar et al., 1998) and beyond, the pre-Silurian rocks that are beyond the scope of this trip. Across the road are outcrops of Rangeley C, the relatively fine grained, most distal portion of the Rangeley Formation. The bedrock here contains staurolite. Note perched two-mica granite boulder of the Mooselookmeguntic Pluton on top of the outcrop.

Re-board vans and descend south along Rt. 17 towards Rumford.

92.4 Pull into parking lot for Coos Canyon rest area in the small village of Byron, 44°43.2'N, 70°37.9' W. There is a public restroom here, across the bridge on the left. The small store across the road has a nice collection of hand lenses and mineral specimens, as well as ice cream.

**STOP NO. 8, COOS CANYON** (30 minutes). This stop is the same as Stop 4 of GSA 2001 Trip F1, Guidotti and Van Baalen.

The beautiful rocks in the river bed are the Perry Mt. Formation that we saw at its type locality but actually better displayed here. Note the well-defined thin bedding of metapelite and fairly clean quartzite, commonly in a gradational fashion. Although these rocks have been subjected to two high-grade metamorphisms, some of the quartzite beds still display very nice delicate, fine-scale cross beds. The present grade here is Upper Staurolite Zone and the main AFM mineral assemblage is staurolite + garnet + biotite + chlorite with muscovite + plagioclase + quartz also abundant. Accessory opaque minerals are ilmenite and very sparse pyrrhotite and graphite. The main thing to note is the very abundant staurolite, garnet, and dark colored biotite, all Fe-rich minerals, something that never occurs in the Mg-enriched silicate bulk compositions of the Smalls Falls Formation. The occurrence of pseudomorphs (now composed of coarse muscovite) (to 1 cm x 5 cm) after andalusite indicates that the earlier M2 event was somewhat higher grade than the second M3 event to which the rocks are now approximately equilibrated.

If time permits, we will continue 0.2 miles south along Rt. 4 to a small dirt road on the right immediately before a bridge over the Swift River. There is a parking area, possibly with some construction equipment here. Descend to the river by a rough trail to some additional outcrops of the Perry Mt. Formation showing variable Al-content as evidenced by presence or absence of staurolite on different horizons. Additionally, there are some lovely open folds here.

Return to the vans, continue to Bethel via Rt. 17 and US 2, about 50 minutes.

**END OF TRIP**
REFERENCES


Photo: Peter Lyttle, Gary Boone, Bob Moench, and unidentified geologist, ca. 1968
INTRODUCTION

Field trip stops will illustrate the hydrologic and geologic setting near Berlin, New Hampshire, and how the setting affects a Superfund site. The trip will include a visit to the former Chlor-Alkali Facility Superfund Site (hereinafter called the “site”), bedrock outcrops, and river reaches. The site’s industrial history, current environmental and regulatory issues, as well as how the geology and groundwater hydrology affect the site cleanup will be discussed. Nearby outcrops of the Ordovician Oliverian Plutonic Suite and Ordovician Ammonoosuc Volcanics, representative of the site and regional geology, will be visited. Stops at several river reaches, with varying settings, will include descriptions of hydraulics, surface water, and sediment chemistry.

Field studies were completed by the U.S. Geological Survey (USGS), in cooperation with the U.S. Environmental Protection Agency (EPA) in August and September 2009 (Chalmers and others, 2013, https://pubs.usgs.gov/of/2013/1076/; Degnan and others, 2011, https://pubs.usgs.gov/sir/2011/5158/), and by the USGS and the New Hampshire Department of Environmental Services (NHDES), Waste Management Division, from October 2002 through February 2004 (Degnan and others, 2005, https://pubs.usgs.gov/sir/2004/5282/). The studies were designed to provide geologic information, provide a conceptual understanding of hydrogeology of the site, and further understand the riverbed sediment and potential contaminant distribution downstream from the site in Berlin, N.H. The site was associated with a pulp and paper mill on the bank of the Androscoggin River (fig. 1). The Chlor-Alkali Facility produced chlorine gas for the papermaking industry. Mercury was released and seeped into the soil, till, and underlying fractured bedrock because of site activities (U.S. Environmental Protection Agency, 2011) and may represent a risk to human health and the environment.

Former Chlor-Alkali Facility industrial history

From 1899 to 1965, a chlor-alkali facility on the east bank of the Androscoggin River was used to produce chlorine gas for the papermaking industry in Berlin, N.H. (fig. 1). The site was associated with a pulp and paper mill, and a sawmill. Chlorine was produced at the site using electrolytic diaphragm and potentially mercury cells. Chlorine was produced primarily to supply the papermaking industry for paper bleaching. Mercuric chloride also may have been produced on site. The sawmill included a wood preserving operation from 1888 to 1930, which used mercuric chloride in a process known as “Kyanization” to preserve the wood (Gove, 1986).

In the 1990s, elemental mercury in the form of a silver-colored liquid was observed, and vapor was detected in and near bedrock fractures along the left riverbank near the site and in river sediment. This prompted the New Hampshire Department of Environmental Services to investigate the site. Since the late 1990s, efforts have been made to contain mercury at the site and eliminate the seepage of contaminated groundwater to the river (Margaret A. Bastien, New Hampshire Department of Environmental Services, written commun., 2003). The site was placed on the EPA National Priorities List in 2005 (U.S. Environmental Protection Agency, 2011; New Hampshire Department of Environmental Services, 2017).

Remedial efforts and environmental concerns

Remedial efforts at the site included (1) removal and demolition (debris left as landfill on site) of buildings associated with the chlor-alkali facility, (2) installation of a subsurface bentonite-soil slurry (barrier) wall on the site perimeter that is connected to the bedrock surface (fig. 2), (3) installation of a synthetic cap over the site to prevent precipitation infiltration, and (4) pressure grouting bedrock fractures along the riverbank. The intent of these remedial actions was to substantially reduce groundwater flow through the site’s overburden and reduce this driving
force for contaminant migration. Despite earlier actions to address the source of contamination, mercury continues to be present in the Androscoggin River at bedrock fractures at the edge of the site; the mechanism of transport “flouring” from deeper or adjacent mercury deposits is unknown. Between 1999 and 2006, about 135 pounds of elemental mercury and sediment containing mercury were removed from the river and riverbank (U.S. Environmental Protection Agency, 2011; New Hampshire Department of Environmental Services, 2017).

Figure 1. A) Location of the study area, Berlin, New Hampshire, with lineaments from high-altitude photography. (Lineaments from Ferguson and others, 1999) and B) circa 1920 photograph of the former Chlor-Alkali Facility Superfund Site on the left, looking south from Sawmill Dam.

An understanding of the extent of fine sediment and mercury contamination in the Androscoggin River also was needed to determine the potential effects on the environment. Eggleston (2009) indicated that resuspension of mercury-contaminated sediment can account for a large percentage of the annual downstream mercury load in a point-source, mercury-impaired watershed of the Shenandoah River in Virginia. An understanding of sediment distribution in the Androscoggin River downstream from the site could identify potential zones of contaminant deposition. Elemental mercury, such as that emanating from the site, can be transported with fine-grained, organic carbon-rich, riverbed sediments. Deposition of these sediments pose a concern because they provide an optimal environment for mercury methylation (Marvin-DiPasquale, Lutz, and others, 2009; Marvin-DiPasquale, Agee, and
others, 2011). Methylation is the conversion of elemental mercury to the organic form (methyl mercury) through microbial activity. Methylmercury is the toxic form of mercury and more mobile within the food chain. This conversion enables the bioaccumulation of methyl mercury in fish.

**Geologic setting**

Berlin, N.H., is on the eastern edge of the Bronson Hill anticlinorium and, more specifically, on the southeast flank of the Jefferson Dome (Billings and Billings, 1975). Billings and Billings (1975) describe three stages of rock deformation and associated development of foliation in the area. Bedrock in the Berlin area is primarily composed of the Oliverian Plutonic Suite and the Ammonoosuc Volcanics. The Androscoggin River channel and narrow valley in Berlin is underlain by metamorphosed biotite-quartz monzonite of the Ordovician Oliverian Plutonic Suite (Billings and Billings, 1975; Lyons and others, 1997). Gneisses and amphibolites of the Ordovician Ammonoosuc Volcanics lie immediately east and west of the valley floor (Billings and Billings, 1975; Lyons and others, 1997). Pegmatite of the Devonian New Hampshire Plutonic Suite locally cuts the Oliverian and Ammonoosuc rocks (Billings and Billings, 1975).

Billings and Billings (1975) describe the Oliverian Plutonic Suite as primarily pink, foliated, medium- to coarse-grained, granoblastic-textured quartz monzonite and pink pegmatite. The Ammonoosuc Volcanics are fine-grained, light-gray, foliated biotite gneiss locally interlayered with hornblende amphibolites. The units are distinguished by the finer grained, locally fragmental and bedded nature of the Ammonoosuc Volcanics and pink character of the Oliverian gneisses. The Oliverian Plutonic Suite intrudes the rocks of the Ammonoosuc Volcanics. Pink pegmatites of the New Hampshire Plutonic Suite locally cut the Oliverian and Ammonoosuc rocks (Billings and Billings, 1975).

Examination of 20 outcrops within a 1-mi radius of the site (fig. 3) support the division of units presented by Billings and Billings (1975). Amphibolites of the Ammonoosuc Volcanics, believed to be xenoliths, are within the area mapped as Oliverian Plutonic Suite rocks. For example, amphibolite (dark green to black, fine-needle, hornblende-plagioclase gneiss) xenoliths are present along Route 16, about 1,000 feet (ft) north of the hospital shown on the 7.5-minute, Berlin, N.H., topographic map. Dark-gray to black hornblende-biotite gneiss with biotite porphyroblasts also are present in rocks mapped as Oliverian in outcrops east of the entrance to the Berlin pulp mill. Unmetamorphosed lamprophyre dikes contain xenoliths of gray gneiss and pink feldspar-bearing gneiss and are present in outcrops east of the entrance to the Berlin pulp mill (about 1,000 ft south of the site).

Analysis of steeply dipping fractures in the Berlin area identifies two distinct fracture domains (fig. 4), which are areas defined by and containing distinct fracture patterns. South and east of the Androscoggin River, principal peaks on normalized azimuth-frequency plots trend west or northwest. West and north of the river (fig. 4), principal peaks trend north or north-northeast. At two localities west of the river, west-northwest and west-southwest principal peaks are present in addition to the north-trending principal peaks. East of the river (fig. 4), however, no north-trending principal peaks are present. The presence of generally west- and north-trending principal peaks at the two localities indicates that domain overlap is happening at these outcrops. The domain boundary and transition zone generally trends northeast along the river. The site is near the boundary between the domains observed in Berlin, N.H. (fig. 4).

Two lineaments (fig. 1) identified from high-altitude areal photographs pass near the site along the riverbank, and towards the southern end of the site (Ferguson and others, 1999). Certain types of lineaments, particularly certain fracture-correlated lineaments, have been associated with more transmissive bedrock in New Hampshire (Moore and others, 2002).

Overburden geologic materials just upstream from the site include stratified sand, gravel, and silt alluvium deposited by glacial outwash. Overburden near the site and downstream to the backwater of the Brown Dam (Cotton, 1975; Gerath, 1978; Olimpio and Mullaney, 1997) was generally less than 20 ft thick and consists of thin deposits of glacial till (an unsorted mixture of clay, silt, sand, cobbles, and boulders). Glacial till covers bedrock on the valley floor and walls where slopes are gentle, but is absent along the Androscoggin River at the site and for 2 mi downstream from the site (U.S. Environmental Protection Agency, 2011).
Figure 2. Location of reference grid, geophysical survey lines, mapped bedrock area, and water-level monitoring points at the former Chlor-Alkali Facility Superfund Site, Berlin, New Hampshire.
Figure 3. Bedrock geology, azimuth-frequency plots of fractures, fracture domains, and foliation near the former Chlor-Alkali Facility Superfund Site, Berlin, New Hampshire.

HYDROGEOLOGY OF THE SITE

The site setting is primarily fractured-rock where little was previously known about the geology and hydrogeology. Data collected during the fall of 2002 through the winter of 2004 were used to produce a geologic description and preliminary understanding of hydrogeology to guide ongoing remedial investigations for mercury contamination and is described in Degnan and others (2005, https://pubs.usgs.gov/sir/2004/5282/).
Site geology

Bedrock in the Androscoggin River, along the west side of the site, consists of gneiss with thin discontinuous lenses of chlorite schist and discontinuous tabular pegmatites as shown in cross section (fig. 4). Two distinct fracture domains overlap near the site (fig. 3). One domain, south and east of the Androscoggin River, is characterized by steeply dipping fractures with principal trends to the northwest. The second domain, north and west of the river, is characterized by steeply dipping fractures with principal trends to the north. The site is near the transition zone between these domains and locally has principal trends common to both fracture domains.

The fractured rock of the site consists of steeply dipping fractures in gneiss that terminate on subhorizontal fractures along contacts with pegmatite, on nonplanar fractures, or at moderately dipping contacts with chlorite schist. The rock types and fractures observed at the riverbank are believed to extend to the east beneath the site. Fractured-rock characteristics of the site include:

- Weakly developed gneissosity with nonplanar (irregular) fractures with a 1-millimeter (mm) (0.04 inch [in.]) aperture that generally conform to the gneissosity. Gneissosity is folded about an axis that plunges N.48°E. at 34°. Fractures that are subparallel to the gneissosity are folded about an axis that plunges N.55°E. at 47°.
- Foliated chlorite-schist lenses, and fractures that parallel these lenses, are folded about an axis that plunges N.37°E. at 26°SE. Vugs in the chlorite schist may contribute to the porosity of fractures in these lenses.
- Fracture intensity is a function of rock type. Rocks at the site, from most to least fractured, include (a) chlorite schist, (b) fine-grained nonfoliated gneiss, (c) coarse-grained weakly foliated gneiss, and (d) pegmatite.
- Fracture aperture varies with fracture type. In general, the aperture of parallel fractures sets are generally less than 1 mm (less than 0.04 in.). Nonplanar (irregular) fractures, locally associated with chlorite schist, often have apertures of about 1 mm (0.04 in.). En échelon fractures have the greatest aperture, generally 1–2 mm (0.04–0.08 in.) and as much as 5 mm (0.2 in.); however, the individual fractures that make up en échelon fracture sets are generally not through going.
- Steeply dipping en échelon fracture zones, parallel fracture zones, and silicified brittle faults show consistent strikes to the NE and on average dip NW.
- Gently dipping to subhorizontal fractures in the gneissic rocks have an average strike and dip of N.43°E., 09°SE.

The geology and hydrology at the site represent a highly complex hydrogeologic environment in terms of groundwater flow and mercury transport. Because mercury is present in the dissolved, elemental, and amalgamated form at the site, the effect of groundwater flow on mercury occurrence and transport will vary according to the form of the mercury. In addition, resuspension and deposition of mercury from turbulent flows in the Androscoggin River may also affect the presence of elemental mercury along bedrock outcrops adjacent to the river.

Figure 4. Generalized cross section of the bedrock geology at the former Chlor-Alkali Facility Superfund Site outcrop from grid 0 to 36, (360 ft) looking east, Berlin, New Hampshire.
Groundwater

Monthly measurements at additional sites were used to supplement spatial coverage of continuous measurement sites. Synoptic surveys of all the wells and piezometers at the site were used to create potentiometric-head maps of groundwater in the overburden and bedrock representing high and low water-level conditions. The bedrock aquifer near the river is well connected to the river, and head gradients in the bedrock across the site are large (more than 10 ft). Water movement between the river and the bedrock aquifer is greatest during periods of river stage fluctuations. A bulk horizontal hydraulic conductivity of the bedrock was estimated, from stage and well water-level responses, to be about 0.2 to 20 ft per day (ft/d). Individual fractures or fracture zones likely have hydraulic conductivities much greater than the bulk rock and affect the higher hydraulic conductivity estimates. Groundwater may move readily through near horizontal, or shallow to moderately dipping fractures, along chlorite schist lenses or through near horizontal fractures at the pegmatite contacts near the river. The near-horizontal features may serve as conduits to the bulk of the site for groundwater in steeply dipping fractures in gneiss. The horizontal, or gently dipping, fractures are discontinuous; therefore, the effective hydraulic conductivity across the site is likely to be closer to the low range of the estimated values (0.2 ft/d).

An unsaturated zone in the middle of the capped area caused by a high bedrock surface separates flow in the overburden into a northern and southern area. The flow is to the west toward the river in the northern and southern areas. Because overburden water-level fluctuations are small, head gradients in the overburden remain fairly constant across the cap area, partly because of the result of the relatively stable head in the canal and flow out of the discharge pipe in the concrete wall, in addition to the geomembrane cap on the site. The discharge pipe drains water at the base of the overburden near the river towards an altitude of 1,087.4 ft (the altitude of the base of the pipe). The alternating lenses of pegmatite, chlorite schist, and gneiss in the underlying bedrock may impose a vertical anisotropy so that the saturated overburden is perched above unsaturated bedrock adjacent to the river.

Bedrock water levels measured on March 20, 2003, during low water-level conditions, indicate a large head difference (about 14 ft) across the site, 8 ft greater than the overburden. Head contours indicate a northwesterly bedrock groundwater-flow direction for much of the cap area and a westerly flow direction for the southern part of the cap area. Calculated bedrock head gradients show small variations in maximum azimuthal direction but large variations in the slope of the gradient. The direction of the bedrock head gradient shifts further downstream as river stage rises. This shift also indicates that the river and bedrock are closely connected hydraulically. Water-level data from overburden and bedrock wells in the capped area indicate a combination of upward and downward gradients on April 2, 2003. Upward gradients from the fractured bedrock to the overburden could provide groundwater recharge to the overburden.

Groundwater-level fluctuations measured at continuous sites fluctuated most at bedrock well MW-7 and least at overburden well MW-2 (10 ft south of MW-7) for the period of record. The average vertical head gradient between the overburden and bedrock (at MW-2 and MW-7, fig. 2) is large (more than 8 ft) and indicates a poor vertical connection between the overburden and bedrock near the riverbank (fig. 2). Heads in MW-7 are slightly lower than SW-1, for 180 of 191 days monitored in 2002 and 2003, indicating a connection to the river downstream of the pool at SW-1 (fig. 2). Groundwater specific conductance at MW-7 shows small variations associated with a water-level rise on June 15, 2003 (fig. 5). The specific conductance of groundwater at MW-7 is high, exceeding 4,000 micromohs per centimeter (μmos/cm). A general decrease in the groundwater specific conductance at MW-7, with sustained high river stage, indicates that low specific conductance river water may temporarily flow into the bedrock aquifer with the increased gradient from the river. Average daily river water specific conductance for the period of record (December 2002–July 2003) shows a pattern of dilution with increased streamflow. Gradual rises in specific conductance during low-flow periods may indicate that high-conductivity groundwater is seeping into the pool from the site. Flow reversals are followed by more conductive water in the pool.
**Figure 5.** Five-minute interval water level and specific conductance from bedrock well MW-7 and river stage SW-1 at the former Chlor-Alkali Facility Superfund Site, Berlin, New Hampshire. (Location of sites shown on figure 2.)

**Geophysics**

Exposed bedrock at the riverbank provided ideal conditions for ground-penetrating radar (GPR) signal transmission and data collection; GPR surveys were also done on the site cap (fig. 6). The capped area at the site contains demolition debris, including metal, which limited the GPR survey potential. GPR data were collected using a point-survey mode because the rough surfaces at the site prevented data collection in a continuous-survey mode. Survey lines (fig. 2), collected directly on rock on the riverbank, indicate mostly shallow and a few steeply dipping reflections. Several reflections indicating fractures at depth are consistent with the fracture patterns observed at the riverbank; for example, horizontal and shallow-dipping reflectors at the northern end represent contacts with fractures associated with gneiss and pegmatite lenses. Inspection of bedrock cores from boreholes in the capped area (boreholes drilled for wells MW-4 and MW-10) indicate that alternating lenses of gneiss, chlorite schist, and pegmatite, similar to that seen at the riverbank, are present in the capped area.

**Figure 6.** Ground-penetrating radar profiles from line 26 on the capped area of the former Chlor-Alkali Facility Superfund Site, Berlin, New Hampshire. (Depth scale is material dependent. Location of line shown on figure 2.)
Resistivity surveys were completed on the riverbank and around the perimeter of the site (fig. 2). Electrical anomalies with resistivity values relatively lower than the surrounding rock apparent in most lines (less than about 1,000 ohm-ft) are fractures. Fracture zones along the site perimeter project towards the site under the barrier wall. The 60° and 85° trending anomalies are of interest because they are parallel to a dominant fracture peak. Various anomalous areas along resistivity line 1 on the riverbank (fig. 7) have resistivities less than 150 ohm-ft and represent fracture zones containing contaminated groundwater. A folded chlorite-schist lense in this area, with vuggy fractures and chlorite alteration, is a major structural feature on the riverbank and is associated with the low-resistivity anomaly. Water levels in bedrock well MW-7 indicate a hydraulic connection between the fractured bedrock and the river in this area (fig. 2). Results indicate electrically conductive anomalies in the bedrock that are in the same areas as the reflections seen in the GPR results.

**Figure 7.** Resistivity profiles from lines 1 and 2 on the bank of the Androscoggin River, former Chlor-Alkali Facility Superfund Site, Berlin, New Hampshire. (Location of lines shown on figure 2.)

Borehole-geophysical logs indicate fewer fractures in MW-14 (fig. 8) than in MW-15; however, shallow dipping fractures are present in both wells at about 28 and 30 ft (at an elevation of about 1,074 ft) below the top of the casing and show the strongest indications of groundwater flow. Fluid property logs of wells MW-14 and MW-15 show fluid temperature and conductance inflections at this depth, which indicate hydraulically active fractures. The shallow subhorizontal fracture at about 1,074 ft elevation likely provides the hydraulic connection to the river noted in the water-level analysis of wells MW-7 and MW-8. Electrically conductive groundwater (greater than 4,000 micromhos per centimeter [μS/cm]) was detected at wells MW-7 (fig. 5) and MW-14 (fig. 8) near the largest resistivity anomaly at about 235 ft along line 1 and 225 ft along line 2 (fig. 7) with a moderate apparent dip to the northeast.
Conceptual model

A conceptual model and preliminary hydrogeologic characterization of groundwater flow at the site indicates that groundwater flows east to west and follows a stair-step pattern within the bedrock toward the river (fig. 9). The overburden aquifer, which consists of till and fill materials, is perched in places and poorly connected to the bedrock aquifer and, therefore, isolated from short-term changes in river stage downstream of the dam; however, the overburden is likely recharged from groundwater inflow that either moves through or under the barrier wall or from the underlying bedrock aquifer. The site cap limits direct recharge from precipitation. Regional groundwater flow may enter the site at the site perimeter. The bedrock aquifer at the site is connected to the Androscoggin River as indicated by results of geologic mapping and hydraulic analysis. The implications of site hydrogeology on mercury storage and transport require additional studies. Geologic mapping at the riverbank, where elemental mercury has repeatedly been found on the outcrop and along fractures exposed at the bank, shows that fracture frequency varies with rock type (plate 1).

Geologic mapping determined that gneiss, 6–9 ft thick, containing near vertical fractures contains moderately dipping, vug-filled, chlorite-schist lenses bounded by fractures and subhorizontal unfractured pegmatite. Subhorizontal or moderately dipping fractures connect the vertical fractures of the bulk rock. Inspection of borehole cores (appendix 1) and geophysical-survey results indicate this geologic pattern likely extends east from the riverbank across the site. Data on fracture density indicate that groundwater in bedrock at the site is stored in the near vertical fractures in the gneiss that comprise most of the bedrock aquifer. The low hydraulic conductivity of the near-vertical fracturing is probably the limiting hydraulic conductivity in the bedrock aquifer at the site. In fractured rock, the hydraulic conductivity of the aquifer, over a scale similar to the field site, is controlled by the small fractures of the dominant fracture network (Tiedeman and others, 1997); therefore, the bulk hydraulic conductivity for the bedrock aquifer across the site is estimated to be about 0.2 ft/d. The hydraulic conductivity of individual fractures near the river is relatively high as indicated by rapid water-level fluctuations in some observation wells.
The high hydraulic conductivity estimate (2–20 ft/d) from well MW-7 is affected by a direct flow path in a more open subhorizontal fracture similar to those noted in the nearby borehole associated with MW-14 (fig. 8). Conversely, a low hydraulic-conductivity estimate of 0.2 ft/d, based on water-level data at wells MW-8 and MW-9, is representative of the steeply dipping fractures in gneiss that make up the bulk of the rock.

Figure 9. Conceptual model of the hydrogeology at the former Chlor-Alkali Facility Superfund Site, Berlin, New Hampshire.

DOWNSTREAM RIVER REACHES

Flow in the Androscoggin River is regulated by eight hydroelectric dams in the study area; flows are controlled to respond to power demands, floods, and structure maintenance. The mean annual flow measured at USGS gage 01054000 in Gorham, N.H., is 2,110 cubic ft per second (ft³/s). The month of May has the highest average annual groundwater flow at 4,210 ft³/s, and August has the lowest at 1,960 ft³/s. Dams and lakes in the headwaters of the Androscoggin River provide storage for a substantial amount of runoff and reduce flood peaks (Federal Emergency Management Agency, 1994). Though hydroelectric dams in Berlin, Gorham, and Shelburne (fig. 2) control flow and sediment transport during normal flows, because of minimal storage volume they have little effect on controlling flood flows (Federal Emergency Management Agency, 1981). The average channel slope in the study area from the site to the Maine State line is 26.1 ft per mile (ft/mi). The slope is much greater (100 ft/mi, fig. 2) between the site and the Cascade Dam (fig. 2) and greatly increases the river’s capacity to produce hydroelectric power, and transport sediment and contaminants in this reach. Most of the dams in the study area make use of the head drop available at the dam site to generate power. The Riverside and Brodie Smith Dams in Berlin (fig. 2) divert water out of the river channel and into a penstock (in this case, a large pipe) to increase heads on the turbines that are farther downstream. Steeper parts of the channel downstream from the Riverside and Brodie Smith Dams receive limited flow during...
average flows, because of the penstock diversion, but carry large flood flows. Sediment may accumulate in deeper pools in these sections of river; however, these areas are not navigable because of steep channel slopes and were not surveyed in this investigation. The arrangement of dams and penstocks, from the Saw Mill Dam at the site to the Berlin–Gorham town line, creates areas of backwater where sediment can accumulate upstream from the Riverside, Brodie Smith, Cross Power, and Cascade Dams (fig. 2).

Alluvial fan deposits, consisting of sand, gravel, and silt are on the left bank (east) of the river upstream from the Cascade and Brown Dams and on the right bank near the Gorham Dam and Shelburne Reservoir. Stratified sand, gravel, and silt alluvium is the dominant deposit beneath the river channel from the backwater behind the Brown Dam to the New Hampshire–Maine State border. Ice-contact deposits of sand and gravel are downstream from the Shelburne Dam in the river channel in the form of eskers, channel fillings, kames, and kame terraces. Undifferentiated glacial drift, consisting mostly of till, was along the left bank of the river in Gorham and Shelburne (Gerath, 1978; Gerath and others, 1985). The Androscoggin River is incised and boulder filled, and average flows form rapids downstream of the Cascade, Brown, and Gorham Dams in Gorham for about 2, 1.5, and 1.5 mi, respectively. Downstream from the Brown Dam to the Shelburne Reservoir, the channel gradient decreases and grades into a slightly braided sediment-filled channel with anabranching sections. The river runs unobstructed downstream from the Shelburne Dam in Shelburne and has braided and meandering sections for about 6.5 mi to the New Hampshire–Maine State border. A significant process that may alter sediment distribution is ice-scour during the winter. At Riverside Dam, about 1,000 ft south of the site, large amounts of ice have been observed abrading the bottom of the river and being transported over the dam.

Figure 10. Generalized locations of sediment and pore-water sampling sites on the Androscoggin River downstream from the former Chlor-Alkali Facility Superfund Site in Berlin, New Hampshire. Stream reaches are signified by AR followed by number. The reference reach (AR2) is 16 kilometers upstream from the site and is not shown. Sampling locations are indicated by red circles, and dams are indicated by black squares. Elevation and distance data from Google Earth, February 17, 2012.

Geophysical bed sediment characterization
Surface geophysical surveys, such as GPR and multifrequency electromagnetic (FDEM) surveys, were used to determine the extent and nature of riverbed sediments in the Androscoggin River downstream from the site in Berlin, N.H. A full description of results, processing methods, and presentation in tables, maps, and cross sections is provided in Degnan and others (2011, https://pubs.usgs.gov/sir/2011/5158/). Results are discussed in terms of
sediment electrical conductivity, pore-water specific conductance (SC), and potential contaminant distribution in riverbed sediments.

The river reaches surveyed in this study ranged from pooled water conditions to fast moving water-flow conditions that varied with channel geometry, dam operation, and runoff. Results of GPR and FDEM surveys were used to estimate the extent and nature of riverbed sediments (tables 1 and 2). In general, wider, less steep gradient reaches have more fine sediment, and were measured with additional surveys in this study, whereas narrow steep gradient reaches had less sediment. Specific conductance measured during surface-water, pore-water, and sediment sampling (with subsequent grain-size analysis), collected as part of a parallel USGS investigation (Chalmers and others, 2013), were used in processing and interpreting surface geophysical surveys. The electrical resistivity of sediment samples was measured in the laboratory with pore water intact for comparison with FDEM survey results. Geophysical surveys of the reference reach, Wheeler Bay (upstream end of reach AR–2), were completed to assess equipment responses in an environment unaffected by site-related contaminants.

To help understand conductivity variations with depth, the most stable five frequencies for a given survey were used along selected survey paths for inverse modeling of the data for select lines to associate resistivity values with depth (fig. 12). Results of the inversion are given in terms of resistivity (inverse of conductivity). Resistivity values measured from bed sediments (about 200 ohm-meters) and surface-water SC values (converted to about 300 ohm-meters) were used to construct a two-layered starting model. GPR riverbed depth interpretations overlaid on the FDEM inversion indicate a correlation with more resistive river water (green layer on top) and less resistive (blue layer) sediment through about 70 percent of the cross section (fig. 12).

At very low frequencies, electromagnetic induction response is due more to the magnetic properties of the subsurface than electrical properties, and the FDEM survey magnetic susceptibility responses are similar to a magnetometer survey. Magnetic susceptibility was calculated using the raw in-phase component of the lowest frequencies used in the surveys (570; 990; 1,770; 3,090 hertz [Hz]). When a magnetic response occurred, generally all four frequencies gave a similar response, though the lower frequencies indicate magnetic material more often than the higher frequencies. Magnetic susceptibility responses (lower frequency FDEM) were plotted on maps in reaches near the site in Berlin, N.H., to search for metal debris that may affect FDEM responses.

The reach between the former Site and the Riverside Dam (AR-3), had small areas of fine sediment on the upstream left bank and the downstream right bank, with an elevated FDEM conductivity (31.4 milliseimens per meter (mS/m) maximum). The larger FDEM conductivity likely was because of elevated riverbed pore-water SC. Reaches AR–4 and AR–5 were downstream from steep gradient (100 ft/mi) bedrock gorges that convey high flows from the reach adjacent to the site to pooled areas behind dams. Reach AR–4, upstream from the Brodie Smith Dam, had the largest pore-water SC, FDEM, and lab-measured sediment conductivity values measured in the study. Pore-water SC measured in this reach was 279 and 324 mS/m at sediment sample locations AR–4_1 and AR–4_2, respectively (fig. 10), on a sandbar near the left bank. These sediment samples had laboratory-measured sediment conductivity values of 67.4 and 76.8 mS/m, similar to nearby estimated FDEM values of 73.2 and 72.8 mS/m. The largest conductivity measured with FDEM in reach AR–5 was 10.2 mS/m on the downstream left side (fig. 8).

Reach AR–9 between the Gorham and Shelburne Dams contains the largest body of pooled water in the study area, the Shelburne Reservoir. The first one-half of the reach has cobble, boulder, and bedrock bed material; although more than 77 percent of the area of the riverbed surveyed in reach AR–9 is estimated to be covered by gravel or finer material (table 2). The sediment in reach AR–9 had a maximum estimated FDEM conductivity of 12.6 mS/m (table 1), greater than all of the other reaches except for AR–3 and AR–4 (nearby and within 1 mile downstream from the site). In addition to large FDEM values, this reach had the second greatest pore-water SC measured, 45.8 mS/m (table 1).

Through combining results and analysis from GPR and FDEM surveys, with sediment pore water and laboratory measured conductivity, detailed riverbed-sediment characterizations were made. Results from GPR surveys were used to image and measure the depth to the riverbed, depth to buried riverbeds, riverbed thickness and to interpret material-type variations in terms of relative grain size (fig. 11). Fifty-two percent of the riverbed in the study area was covered with gravel and finer sediments. GPR surveys are affected by contrasts in the electrical properties of water and sediment. The electrically resistive river water and sediment in this study area were conducive to the penetration of the GPR and FDEM signals and allowed for effective sediment characterization by geophysical methods.
Figure 11. Example of a cross section showing ground-penetrating radar profile and interpretation in reach AR–7 upstream from the Brown Dam, Androscoggin River, Gorham, New Hampshire (location shown in figure 10).

Figure 12. Example of a cross section showing inverted electromagnetic induction profile in reach AR–9 between the Gorham and Shelburne Dams, Androscoggin River, Shelburne, New Hampshire (location shown in figure 10). Solid black line represents riverbed measured with ground-penetrating radar.

Table 1. Estimates of electromagnetic conductivity summarized by reach and correlation with water depth.

<table>
<thead>
<tr>
<th>Reach name</th>
<th>Reach code</th>
<th>Conductivity, in millisiemens per meter</th>
<th>Water depth and conductivity correlation</th>
</tr>
</thead>
<tbody>
<tr>
<td>Wheeler Bay</td>
<td>AR–2</td>
<td>2.6</td>
<td>9.4</td>
</tr>
<tr>
<td>Upstream of the Riverside Dam</td>
<td>AR–3</td>
<td>3.1</td>
<td>31.4</td>
</tr>
<tr>
<td>Upstream of the Smith Dam</td>
<td>AR–4</td>
<td>10.1</td>
<td>194.9</td>
</tr>
<tr>
<td>Upstream of the Power Dam</td>
<td>AR–5</td>
<td>4.2</td>
<td>10.2</td>
</tr>
<tr>
<td>Power Dam to Cascade Dam</td>
<td>AR–6</td>
<td>2.7</td>
<td>7.1</td>
</tr>
<tr>
<td>Cascade Dam to Brown Dam</td>
<td>AR–7</td>
<td>2.2</td>
<td>8.9</td>
</tr>
<tr>
<td>Brown Dam to Gorham Dam</td>
<td>AR–8</td>
<td>0.9</td>
<td>7.1</td>
</tr>
<tr>
<td>Gorham Dam to Shelburne Dam</td>
<td>AR–9</td>
<td>2.5</td>
<td>12.6</td>
</tr>
<tr>
<td>Downstream of the Shelburne Dam</td>
<td>AR–10</td>
<td>1.7</td>
<td>3.5</td>
</tr>
</tbody>
</table>
Table 2. Percentage of fine sediment summarized by reach.

<table>
<thead>
<tr>
<th>Reach name</th>
<th>Reach code</th>
<th>Area of gravel or finer sediment, in square feet</th>
<th>Percent of reach area</th>
</tr>
</thead>
<tbody>
<tr>
<td>Wheeler Bay</td>
<td>AR–2</td>
<td>933,256</td>
<td>97</td>
</tr>
<tr>
<td>Upstream from the Riverside Dam</td>
<td>AR–3</td>
<td>29,894</td>
<td>11</td>
</tr>
<tr>
<td>Upstream from the Smith Dam</td>
<td>AR–4</td>
<td>85,109</td>
<td>25</td>
</tr>
<tr>
<td>Upstream from the Power Dam</td>
<td>AR–5</td>
<td>354,077</td>
<td>61</td>
</tr>
<tr>
<td>Power Dam to Cascade Dam</td>
<td>AR–6</td>
<td>342,268</td>
<td>51</td>
</tr>
<tr>
<td>Cascade Dam to Brown Dam</td>
<td>AR–7</td>
<td>1,345,012</td>
<td>42</td>
</tr>
<tr>
<td>Brown Dam to Gorham Dam</td>
<td>AR–8</td>
<td>432,820</td>
<td>20</td>
</tr>
<tr>
<td>Downstream from the Shelburne Dam</td>
<td>AR–9</td>
<td>5,794,177</td>
<td>77</td>
</tr>
<tr>
<td>Total</td>
<td>AR–10</td>
<td>6,915,104</td>
<td>86</td>
</tr>
<tr>
<td>Total</td>
<td>AR–10</td>
<td>16,231,717</td>
<td>52</td>
</tr>
</tbody>
</table>

Mercury contamination

Total mercury (THg) and methylmercury (MeHg) concentrations in Androscoggin River sediment, pore water, and biota were elevated downstream from the site relative to reference sites (figs. 13 and 14); methods and results are described in Chalmers and others (2013, https://pubs.usgs.gov/of/2013/1076/). Sequential extraction of surface sediment showed a distinct difference in mercury speciation upstream compared with downstream from the site. The reference site was dominated by potassium hydroxide-extractable THg consistent with organic mercury or particle-bound divalent mercury (Hg(II)), whereas sites downstream from the point source were dominated by concentrated nitric acid-extractable THg, indicative of Hg0 or mercurous chloride. Mercury metrics from the study indicated Hg(II) at the reference site was more available for Hg(II)-methylation compared with sites downstream from the point source. In addition, whole sediment Hg(II)R and smallmouth bass THg concentrations seemed to increase farther downstream from the point source. The farthest downstream reach (AR9 from Gorham Dam to Shelburne Dam) had larger mass of fine sediment and larger estimated mass inventory of mercury species than any other stream reach by an order of magnitude for both masses.

Toxicity tests and invertebrate community assessment suggest that impairment of invertebrates is not occurring at the 2009 and 2010 levels of mercury contamination downstream from the point source. Concentrations of THg and MeHg in most water and sediment samples from the Androscoggin River were below Federal and consensus-based guidelines, whereas smallmouth bass mercury concentrations were above U.S. Environmental Protection Agency and regional guidelines in all samples. Smallmouth bass THg concentrations from the Androscoggin River downstream from the point source were substantially higher than those reported in a national survey, but only smallmouth bass mercury concentrations from the farthest downstream stream reaches (Cascade Dam to Shelburne Dam) were substantially higher than those in Northeastern region studies.

The apparent greater potential for Hg(II)-methylation and mercury bioaccumulation in the lower gradient stream reaches of the Androscoggin River may reflect changes in the type and size of particles deposited to the benthos and the speciation and availability of mercury for Hg(II)-methylation associated with those particles. These findings suggest that an even greater potential for Hg(II)-methylation and mercury bioaccumulation may exist as the river gradient continues to flatten downstream from Shelburne Dam.
Figure 13. Concentrations of sediment A, total mercury (THg), and B, methylmercury (MeHg), from the Androscoggin River, Coos County, New Hampshire. Samples were collected in 2009 and 2010. Samples from the reference reach (AR2) are from 16 kilometers (km) upstream from a Former Chlor-Alkali Facility Superfund Site in Berlin, N.H. Samples from near-stream reaches (AR4–AR6) are from 2 to 4 km downstream from the Former Chlor-Alkali Facility Superfund Site, and samples from far-stream reaches (AR7–AR9) are from 8 to 16 km downstream from the Former Chlor-Alkali Facility Superfund Site. THg and MeHg nondetect data are excluded from the plot because of high detection levels. The dashed blue line indicates the median reference (AR2) sediment concentration, and the dashed red line indicates the threshold effects level (TEL) concentration of 180 nanograms per gram (ng/g; MacDonald and others, 2000). dw, dry weight.

Figure 14. Concentrations of pore-water A, total mercury (pw.THg), and B, methylmercury (pw.MeHg) percentage of THg as MeHg (pw% MeHg) from the Androscoggin River, Coos County, New Hampshire. Samples were collected in 2009 and 2010. Samples from the reference reach (AR2) are from 16 kilometers (km) upstream from a Former Chlor-Alkali Facility Superfund Site in Berlin, N.H. Samples from near-stream reaches (AR4–AR6) are from 2 to 4 km downstream from the Former Chlor-Alkali Facility Superfund Site, and samples from far-stream reaches (AR7–AR9) are from 8 to 16 km downstream from the Former Chlor-Alkali Facility Superfund Site. Total mercury nondetect data are excluded from the plot because of high detection level. The dashed blue line indicates the median reference (AR2) pore-water concentration. ng/L, nanogram per liter.

ACKNOWLEDGMENTS

The authors wish to thank the co-authors of the reports that document the data collection and analysis behind the results presented in this field trip: Stewart F. Clark, Jr. (geologic mapping and analysis), Philip T. Harte, Thomas J. Mack, Andrew P. Teeple, Craig M. Johnston, Mark C. Marvin-DiPasquale, James F. Coles, and Jennifer L. Agee. The authors also wish to thank many others who helped support the work in the field, as cooperators, and reviewers including: Fred McGarry, John Cotton, and Margaret Bastien of the NHDES; Vincent DelloRusso, James Soukup, Kathleen Soukup, Joseph Souney, and Joseph Schmidl of Weston Solutions, Inc.; Gregory Walsh, Jeffrey Deacon, Glenn Hodgkins, Robert Flynn, Marc Zimmerman, Jon Denner, Jamie Shanley, Brandon Fleming, Thor Smith, and
Laura Hayes of the USGS; Vivien Taylor of Dartmouth College; Cornell Rosiu of the EPA; and Tammie Lavoie, Dennis Pednault, and David Bolstridge of the Berlin-Gorham Operations of Fraser Paper, Inc.

ROAD LOG

MEETING POINT, Adjacent to the Brodie Smith Dam, Berlin, NH. (326596.00 m E, 4926285.00 m N)
Friday, September 29th, 10:00 AM, in the parking lot on the south side of Mason St. on an unnamed island adjacent to the Smith Dam and Hydrostation water intake canal in the Androscoggin River. The Smith Dam is about 30 miles (40 minutes) west of Bethel, ME. From Bethel follow Rte. U.S. 2 west to Gorham, N.H., then take Rte. NH 16 north to Berlin. In Berlin, turn right onto Unity St. (truck Rte. 16 N) at the James Cleveland Bridge, then left onto Mason St. The parking lot will be on your left after crossing the first bridge over the Androscoggin River on Mason St. Cumulative mileages given below may differ from those shown on your odometer, but the indicated distances between stops are generally accurate.

Mileage and directions to STOP 1.
0.0 Head southeast on Mason St toward Unity St.
0.1 Turn left onto Unity St.
0.2 Continue onto Coos St.
0.3 Turn left onto Hutchins St.
1.5 Turn left on Bridge St.
1.6 Turn left through gate (access with field trip leaders only)
1.9 To penstock gatehouse at site

STOP 1. Former Chlor-Alkali Facility Superfund Site Site (327732.00 m E, 4927301.00 m N)
The riverbank outcrop at the site contains a variety of brittle structures that display a consistent pattern of orientation and association. Fracture density and style are related to rock type along the riverbank outcrop. The tabular body of pegmatite at the north end of the riverbank is relatively unfractured compared to gneiss. Fractures present in the gneiss below this pegmatite terminate at the pegmatite-gneiss contact. Coarse-textured weakly foliated gneiss is fractured but contains fewer isolated fractures and fewer fracture sets than fine-grained gneiss, which is highly fractured (fig. 5). Closely spaced individual fractures are shown as zones on the geologic map (plate 1, Degnan and others, 2005). Fracturing in the chlorite-schist lenses is present along the boundaries of the lenses and along parting parallel to foliation within the lenses. En échelon fracture zones, parallel fracture zones, and faults cut the folded foliation of the gneiss at the site. Steeply dipping en échelon fracture zones, parallel fractures, and faults have a similar trend throughout the riverbank outcrop. Two faults form the contacts of a cataclastic pegmatite (plate 1, Degnan and others, 2004).

Surface- and groundwater levels were used to assess hydraulic connections and provide a hydraulic analysis of the bedrock-river aquifer system. The location of the canal; the Androscoggin River; remedial modifications (barrier wall and cap); and relic foundations and plumbing, including the concrete retaining wall and discharge pipe, affect spatial and temporal variability in water-level fluctuations and flow patterns at the site.

Mileage and directions to STOP 2.
1.9 penstock gate house at site to Bridge St.
2.17 Turn right onto Bridge St.
2.2 Turn right onto Hutchins St.
2.5 Turn left onto Success Pond Rd.
3.1 Park at power lines.

STOP 2. Success Pond Rd. Ammonusic Volcanics (328744.00 m E, 4927196.00 m N)
Analysis of fracture measurements define inclusion in the domain of west- and northwest-trending fractures.

Mileage and directions to STOP 3.
3.1 Turn left onto Success Pond Rd.
3.7 Turn right onto Hutchins St.
4.0 Turn left onto Bridge St.
4.1 turn right, continue onto Hutchins St.
4.5 Turn left onto 12th St.
4.55 Continue onto 12th St. Bridge
4.6 Turn left onto Main St.
5.8 Turn left into court house parking lot, outcrop is in the right back corner

STOP 3. Court House, Oliverian Plutonic Suite
(327165.00 m E, 4926737.00 m N)
Analysis of fracture measurements define inclusion in the domain of north-trending fractures.

**Mileage and directions to STOP 4.**
5.8 Turn left onto Main St.
6.0 Continue onto Pleasant St.
6.8 Turn right onto Glen Ave.
7.2 Turn right into gas station parking lot, right side

STOP 4. Irving Station, schist, gneiss, Oliverian Plutonic Suite and Description
(325975.00 m E, 4925215.00 m N)
Analysis of fracture measurements define inclusion in the domain of west- and northwest-trending fractures.

**Mileage and directions to LUNCH STOP.**
7.2 Turn left onto Glen Ave.
7.3 Turn right onto James Cleveland Bridge/Unity St.
7.8 Left at park

LUNCH STOP. Park on Unity St. on the left bank of the Androscoggin River between the Brodie Smith and Cross Power Dams
(326523.00 m E, 4925948.00 m N)

**Mileage and directions to STOP 5.**
7.8 Head southwest on Unity St.
8.3 Turn left onto Glen Ave. (Rt. 16)
8.6 Left into gravel parking lot by dam

STOP 5. Cross Power Dam
(326002.00 m E, 4924914.00 m N)

**Mileage and directions to STOP 6.**
8.6 Head south on NH-16 S/Glen Ave. toward Watson St.
11.9 Turn left into gravel parking area

STOP 6. Brown Dam
(325179.00 m E, 4919681.00 m N)

**Mileage and directions to STOP 7.**
11.9 Turn left onto NH-16 S/Main St.
14.4 Turn left onto Power House Rd.
14.6 Slight right and park

STOP 7. Gorham Dam
(327457.00 m E, 4917291.00 m N)

**Mileage and directions to STOP 8.**
14.6 Slight left onto Power House Rd.
14.8  Turn left onto US-2 E/Main St.
18.0  Turn left onto North Rd.
18.3  Turn right onto North Rd.

STOP 8.  Shelburne Dam
(331575.00 m E, 4918981.00 m N)

REFERENCES CITED


POSSIBLE POST-LAURENTIDE CIRQUE GLACIATION IN THE GREAT GULF PRESIDENTIAL RANGE, NEW HAMPSHIRE

Brian Fowler, NH Geologic Resources Advisory Committee
Ian Dulin, Dept. of Geology, Bates College

INTRODUCTION

This trip continues to explore the possibility that active local glaciers existed in the cirques on the Presidential Range after departure of the Late Wisconsinan Laurentide Ice Sheet (LIS), a subject debated for now more than 145 years (Fig. 1; Fig. 2). Debate arises because the surficial geologic features of these cirques are not consistent with those generally considered together as diagnostic of recent cirque glaciation: 1) fresh, sharply-defined cirque basins and 2) associated moraines within and below them. Here the cirques are sharply-defined, but so far associated moraines are missing.

Figure 1: Tuckerman (r) and Huntington (l) Ravines; glacial cirques, eastern flank of Mt. Washington. Bradford Washburn photograph used by permission.

Over the years, debate has been prolonged by difficult field access in and below the cirques that hampered the search for moraines. Recently however, recreation trail development improved such access below the Great Gulf cirque and permitted new mapping there. This work, together with laboratory-based petrographic study of stone clast provenance and recently published post-LIS regional climate indications, led to the identification of landforms that may be moraines. If confirmed by further study, the presence of these features will combine with the freshness of the cirque’s morphology as the first consistent evidence of active cirque glaciation in the region following departure of the LIS.
Figure 2: Glacial cirques on the Presidential Range showing extent of estimated cirque glaciers. From Goldthwait, 1970.
PREVIOUS STUDIES

Louis Agassiz (1870) proposed that ice caps existed over the region’s higher elevations after the LIS and that alpine glaciers flowed downward from them through the cirques and cirque-like basins into the region’s valleys. In support he cited their sharply-defined morphology along with valley features he proposed were their terminal moraines.

C. H. Hitchcock (1876, 1877, 1878) performed field work on the region’s highest peaks and refuted this ice cap proposal, citing the lack of striations oriented in the multi-directional patterns that would arise from such ice caps. However, he did not further consider the formation of the cirques.

J. W. Goldthwait (1913) undertook field reconnaissance in the cirques, concluding that no post-LIS activity had occurred within them. He observed no striations or erosional features produced by local ice moving along their axes, no moraines on or beyond their floors, and evidence that the symmetrical U-shape of several had been last modified by erosion he asserted could only be attributed to the obliquely overriding LIS. He also found erratic cobbles on their floors and inferred from their presence that the cirques were formed before and not after the LIS, postulating that post-LIS cirque glaciation would have removed them. Later (1916), he convincingly reinterpreted the deposition of Agassiz’s terminal moraines, showing they were related to the LIS and not alpine glaciers flowing downward from regional ice caps.

Ernst Antevs (1932) questioned Goldthwait’s conclusion regarding post-LIS cirque glaciation. He proposed that the absence of moraines resulted not from a lack of post-LIS activity but from a lack of till deposits on their floors, noting that till was rarely observed adjacent to or beneath abundant deposits of thick talus. He postulated that cirque glaciers may have been diminutive and largely immobile; only able to undermine and steepen their basins.

D.W. Johnson (1933) agreed with Antevs arguing that lack of end moraines was not sufficient evidence to conclude no post-LIS activity had occurred. He cited alpine regions elsewhere that had never undergone continental glaciation but whose cirques lacked moraines.

R. P. Goldthwait (1936, 1940, 1970) performed further field work in the cirques. He confirmed the conclusion that no post-LIS cirque glacier activity had occurred within them by adding to his father’s evidence the northwest to southeast orientations of various “groove-like features” and roche mouintonees on cirque floors and headwalls. Despite important differences in the nature and elevation of these features, he asserted they could have been formed and preserved only if the overriding LIS was the last erosional agent to affect the pre-existing cirques. At the same time, however, he suggested the possibility that residual ice masses might have persisted after departure of the LIS in the deeper most favorably oriented cirques. In 1970, he prepared the first morphometric comparison of the cirques, estimating firm line and bergschrund elevations along with estimated terminus positions for their possible glaciers (Fig. 2).

W. F. Thompson (1960, 1961) used early techniques of photogrammetry to argue the sharply defined morphology in the cirques could only result from active cirque glaciation that post-dated the LIS and that any moraines in or below them had been obliterated by intense wasting. He did not, however, support these assertions with field observations.

D.M. Eskenasy (1978) completed photogrammetric analyses and field work in the King Ravine cirque and proposed that residual ice-based activity had occurred there on and within what was then proposed by some to be a relict rock glacier developed in post-LIS time. The study was, however, unable to establish if the feature was a rock glacier or if any post-LIS glacial activity had occurred in the cirque.

P.T. Davis and R. B. Davis (1980) and later P.R. Bierman, et al. (2000) and P.T. Davis, et al. (2003) investigated the possibility and timing of post-LIS activity in the cirques first using minimum radiocarbon ages from the limited number of available tarns and peat bogs and later cosmogenic nuclide ages from quartz veins in boulders and bedrock surfaces in the cirques. Samples from Tuckerman Ravine (near the Great Gulf; Fig. 2) yielded an age range consistent with regional dates for deglaciation by the LIS but not sufficiently specific to establish if post-LIS cirque activity had occurred there.
D.C. Bradley (1981) challenged the absence of post-LIS cirque activity by proposing that surficial deposits below the mouth of the north-facing King Ravine cirque were a composite moraine emplaced there by a local glacier flowing out of the cirque in post-LIS time. He supported this by citing the well-known presence on the feature of boulders whose lithologies outcrop on the cirque’s headwall, along with the postulation that such deposits could not have survived overriding erosion by the LIS and thus must post-date it.

B.K. Fowler (1984) and R.B. Waitt and P.T. Davis (1988) independently examined this proposal and its supporting evidence. They each concluded the deposits are not a moraine related to a post-LIS cirque glacier, but are instead a complex of massive debris-flows likely related to rapidly melting residual ice in the cirque.

D. J. Thompson (1999) re-examined what were believed by some to be moraines in the Tuckerman Ravine cirque. He concluded from the depositional fabric of their large bouldery clasts that one of the deposits could possibly be a relict rock glacier, but the others were simply talus accumulations beneath steep slopes. He further concluded their presence did not support the presence or reactivation of post-LIS glaciation in the cirque.

P. T. Davis (1999) re-examined the morphometry of the cirques by expanding the techniques used earlier by Goldthwait (1970). He continued to find no convincing evidence for post-LIS cirque activity in the cirques, generally postulating that post-LIS regional climate warmed so rapidly that equilibrium-line altitudes rose too quickly above the cirques floors to support cirque glaciers.

Thompson and Fowler, 1989; Thompson et al., 1999; Fowler, 1999, 2011, 2012; and Thompson et al., 2017 established that moraines deposited by late-glacial readvance or standstill of the LIS 5 to 8 km northwest of the Presidential Range are part of the White Mountain Moraine System (Fig. 3) that extends irregularly west to east across the region. Minimum-limiting radiocarbon ages and direct correlation to the North American Varve Chronology (Ridge, et al., 2012) demonstrate these moraines were deposited after the departure of the LIS from the immediate region during the Older Dryas Cold Interval (~14.0 cal ka BP).

Figure 3: Location and extent of the White Mountain Moraine System (gray-shaded solid and dashed line) and its proximity to Mt. Washington and the Great Gulf Cirque complex. From Thompson, et al., 2017.

B. K. Fowler (2011) completed compilation of the surficial geology of the Mt. Washington East 7.5-minute quadrangle that includes the region’s most prominent cirques. This work located for the first time landforms below
the Great Gulf that may be moraines related to active glaciation in the cirque after the departure of the LIS. The mapping also established that morphologic asymmetry in the Great Gulf cirque itself results from local rock structure and not erosion by an overriding ice sheet. The work also demonstrated that moraines located northwest of the Presidential Range are part of the White Mountain Moraine System.

I. T. Dulin (2011) completed field and laboratory work that established that the provenance of a significant percentage of stone clasts in the lower landform identified by Fowler (2011) lies within the Great Gulf cirque complex. He concluded that this evidence suggests the lower landform was deposited by ice moving out of the cirque and not from wasting ice masses nearby.

NATURE AND GENESIS OF THE POSSIBLE GREAT GULF MORAINES

There are two landforms proposed to be moraines, one near the mouth of the Great Gulf cirque complex and the other in the lower central area of its principal cirque basin (Fig. 4).

The lower elevation landform is an extensive group of cross-valley hillocks strewn with very large, sharply to moderately faceted boulders (to 20 ft. diameter) of rock types outcropping within the cirque above. Its deposits consist of loosely consolidated, poorly to moderately sorted, sandy till and well sorted coarse to medium sandy gravel with scattered lenses of silty sand that could not have survived overriding erosion by the LIS. Thus, its location, hummocky cross-valley topography, loosely-consolidated deposits, and apron of abandoned distributary drainage away from the cirque mouth suggest it could be a terminal moraine emplaced by a cirque glacier after departure of the LIS from the area. This proposal is also supported by evidence below the landform on the valley’s floor. Deposits there show its emplacement displaced and temporarily dammed the pre-existing Peabody River creating an ephemeral lake. This lake quickly filled with deposits of moderately well-sorted loosely-consolidated
sand and gravel derived from the feature’s nearby slopes that could similarly not have survived overriding LIS erosion (Fowler, 2011, 2013; Fig. 4).

The higher elevation and less extensive landform is a group of similarly boulder-strewn (to 12 ft. diameter) gravelly hillocks that form a cross-valley ridge on the central floor of the cirque (Fig. 4; Fig. 5). Its deposits again consist of loosely consolidated, poorly to moderately sorted, sandy till and well sorted coarse to medium sandy gravel and silty sand. Its location, hummocky cross-valley topography, and abandoned down-cirque distributary drainage apron suggest it could be a recessional moraine deposited after departure of the LIS and during retreat of the cirque glacier.

Figure 5: View west from Great Angel Station, Great Glen Trails system. Proposed recessional moraine is visible as the flat cross-valley feature to the immediate left of the lone pine tree in the right-center of the image. B.K. Fowler photograph.

Fowler (2013) proposed these landforms are moraines using the following rationale. Earlier but just recently published findings of Thompson, et al. (2017 and Fig. 3), suggest that near-glacial conditions may have existed in the region immediately following departure of the LIS and due to its close proximity during the now confirmed nearby stillstand. Such conditions may have permitted equilibrium-line altitudes to remain beneath the floors of the favorably oriented higher elevation cirques of the Great Gulf complex for time sufficient to maintain or reactivate residual ice movement. The cirques in the complex (Fig. 2; Great Gulf, Sphinx Basin, Jefferson Ravine, and Madison Gulf) have the highest elevations and areally largest composite basins in the region and their individual cirque axes face away from maximum solar isolation. Thus, the newly identified landforms below and within the complex could be moraines related to post-LIS cirque glaciation, especially since their cross-valley topography and loosely consolidated deposits could not have survived, and thus must post-date, its overriding erosion.

Fowler (2013) investigated and preliminarily discarded two alternate proposals for the emplacement of these landforms (Fig. 6). The first was that the lower feature was emplaced by readvance of LIS ice in the valley below the cirque. This was rejected for three reasons. First, surficial mapping there failed to detect evidence of readvance...
between the feature and the White Mountain Moraine System to the north (Fowler, 2012). Next, reconnaissance observation of stone clast provenance on the landform failed to detect elevated percentages of the distinctive two-mica granite over which readvancing ice would have passed (Dulin, 2011; Fig. 6). And finally, the loosely-consolidated deposits of the ephemeral lake dammed by the landform could not have survived erosion by such a valley readvance.

![Figure 6: Bedrock geology, lower Peabody River valley showing the locations of the proposed terminal moraine and the circular body of two-mica granite to its north northeast in the valley through which any LIS readvance would have passed. From Fowler, 2013.](image)

The second proposal was that the lower landform was deposited from masses of stagnating LIS ice in its immediate vicinity following retreat of the ice sheet (Fowler, 2013). This idea was also discarded for three reasons. First, the very large angular boulders on its surface do not reflect the degree of comminution and surface abrasion typical of long-term entrainment in an ice sheet. Next, its constituent deposits are not the largely unsorted, densely consolidated, and heavily comminuted till typically created within an ice sheet. And finally, the well-developed, unidirectional distributary drainage pattern across the feature is not consistent with the multi-directional patterns typically observed in the region where large masses of stranded ice are known to have down-wasted (Goldthwait and Mickelson, 1982).

Meanwhile, Dulin (2011) completed a stone clast provenance study of samples taken from till across the lower landform. The study included 250 unsorted clasts from 5 locations and determined their provenance by laboratory comparison with known lithologies in the region as established by Eusden (2010; Fig. 6). Results showed that the clasts include lithologies of both local and regional provenance, but clasts uniquely sourced from within the Great Gulf cirque support the interpretation that they were deposited by processes dependent on the presence of a local glacier in the cirque after the departure of the LIS as proposed by Fowler (2013). The study concluded that its results along with the pronounced topography and cross-valley location of the lower landform support the interpretation that it is a terminal moraine.
Much work remains to be done on these landforms and their genesis. Further stone clast provenance studies are being considered, as is cosmogenic nuclide exposure dating of boulders on the features to determine the date of their melt-out and emplacement. The coming availability of LiDAR imagery for the region will be of great value for further mapping and interpretation.

FIELD TRIP HIKING LOG

The trip consists of an easy to sometimes moderate ~ 4.5-mile roundtrip hike to the Great Angel Station Lookout on the Great Glen Trails System to examine landforms traversed by the hike and visible from the Lookout (weather-permitting) that may represent terminal and recessional moraines in the cirque. The trip begins in the North Parking Lot at the Mt. Washington Auto Road (MWAR: 322363.00 m E / 4906302 m N) Base Lodge on the east side of NH Rte. 16, 8 miles south of Gorham, NH. Figure 7 shows this location along with the trail stops included in the field trip.

PLEASE NOTE:

Motorized vehicles are not permitted on the Great Glen Trails System. Because of frequent trail maintenance, periodic sporting events, and sometimes unfavorable weather, parties interested in foot access to the system must check-in and obtain an up-to-date trail system map at the MWAR Base Lodge.

Sturdy footwear and a backpack with water, snack items, and clothing appropriate for the day and potential seasonal weather are required for this trip.

Meanwhile, the following geologic maps will be helpful:

Bedrock Geology of the Presidential Range, New Hampshire (Eusden, 2010)


Stop 1: MWAR North Parking Lot.

This Stop provides a dramatic view into the lower reaches of the Great Gulf cirque complex and of the surrounding peaks of northern Presidential Range (left to right: Mt. Washington, 6,288 feet; Mt. Jefferson 5,712 feet; Mt. Adams 5,774 feet; and Mt. Madison, 5,367 feet). Refer to the text above for general descriptions of the Great Gulf cirque complex, possible local glaciation within the complex, the nature and formation of the possible moraines, and the ephemeral lake the lower landform dammed during its emplacement.

Walk south to the MWAR Base Lodge building and then right to its west side and enter the pedestrian tunnel under NH 16. After emerging from the tunnel, follow short but intersecting segments of Nordic ski trails across the bottom of the now-dry bottom of the ephemeral lake to Intersection 2 (see Fig. 7). From this Intersection, continue along a combination of Nordic ski trails over the surface of the proposed terminal moraine landform, passing trail Intersections 5, 8, 14,19, 27, 33, 51 and 55 to the Great Angel Station at the northwesterly end of, and at highest point on, the trail system and the proposed terminal moraine (see Figs. 4 and 7).

Stop 2: Great Angel Station.

This Stop provides a viewpoint into the central portion of the Great Gulf cirque complex and in particular of the cross-valley location of the proposed recessional moraine (see Figs. 4 and 5). Refer to the text above for descriptive information. Here the West Branch of the Peabody River has been diverted to the north by a bedrock outcropping and the ice-proximal slope of the landform that appears to have been emplaced against it.

Return to Intersection 55 and proceed along the ski trail to Intersection 56 and Stop 3.
Stop 3: Drifter Stone Count Site.

This Stop and Stops 4 and 5 visit sites typical of clast provenance in cirque-ice proximal and distal locations on the proposed terminal moraine and then at a location outside its topographic limits (see Figs. 6 and 7, Eusden, 2010, and Dulin, 2011). The following information is pertinent to each of these Stops.

In order to determine where the material in the landform was derived, the study conducted by Dulin (2011) attributed stone clast provenance to Cirque, Northern, Adjacent, or Erratic sources based on their petrography compared with those mapped in the region by Eusden (2010). The two-mica granite that outcrops directly to the north of the landform was used as the Northern indicator (Fig. 6; DCtmg). The Rangeley Formation (Sr and Src) was used as the Adjacent indicator even though it outcrops both within the cirque and to its immediately adjacent north. Only lithologies outcropping exclusively within the cirque were used for the Cirque indicator. These include the Littleton (DI), Madrid (Sm), Smalls Falls (Ssf) Formations, a granite-diorite (Dwd), and various pegmatites. Clasts not identifiable or not associated with these indicator lithologies were considered erratic as shown in Table 1 and discussed below.

<table>
<thead>
<tr>
<th>Table 1</th>
<th>Clast Provenance - Proposed Terminal Moraine (%)</th>
</tr>
</thead>
<tbody>
<tr>
<td>Test Pit</td>
<td>Erratic</td>
</tr>
<tr>
<td>Drifter</td>
<td>39</td>
</tr>
<tr>
<td>Wilding</td>
<td>52</td>
</tr>
<tr>
<td>Thumper 1</td>
<td>26</td>
</tr>
<tr>
<td>Thumper 2</td>
<td>57</td>
</tr>
<tr>
<td>Combined</td>
<td>48</td>
</tr>
<tr>
<td>Libby*</td>
<td>36</td>
</tr>
<tr>
<td>* Off The Landform</td>
<td></td>
</tr>
</tbody>
</table>

The relatively high percentage of erratic clasts in the landform is attributable to the presence of residual LIS till left within and near the cirque. If post-LIS cirque activity took place, this till would be eroded out of the cirque first and before any of its bedrock, and would thus make up a significant part of the material in a post-LIS landform.

Adjacent clasts (e.g. Rangeley Fm.) are located both north of the landform and on its cirque-side suggesting they could have been deposited by either or both the LIS and post-LIS cirque activity. Thus such clasts are not diagnostic of either mode of deposition, and reliance must be placed instead on percentages of Cirque and North clasts to establish the source locations.

The Libby Site is presented separately at the bottom of Table 1 to show clast provenance typical of terrain that lies beyond the mapped limits of the proposed terminal moraine. The Site’s clasts are dominated by those of northerly provenance, principally two-mica granite (see Fig. 6; DCtmg), suggesting they were deposited by the LIS as it advanced into the area from the north. This contrasts distinctly with the Combined results (Table 1) of samples from the landform which show a marked paucity of tw-o-mica granite and an abundance instead of clast types related to emplacement processes that appear unrelated to the LIS. If the landform were emplaced by the LIS moving from the north in a readvance, the provenance of clasts across its surface would resemble the Libby Site’s distribution.

The Drifter and Wilding Sites are located at positions that would have been proximal to glacial ice possibly advancing out of the cirque, with the Wilding Site (not visited by this trip) located several hundred meters south southwest of Drifter. Clasts at both Sites show a distinct abundance of Cirque clasts, but both also show the presence of clasts likely arising from residual till.
The Thumper 1 and 2 Sites are located at positions that would have been distal to glacial ice advancing out of the cirque, with the Thumper 1 Site (also not visited) located a few hundred meters northeast of Thumper 2. Clasts at Thumper 2 show an abundance from the Cirque, while those at Thumper 1 show abundance from residual till and nearby outcrops.

When results from the Drifter, Wilding, Thumper 1 and 2 Sites are combined (Table 1), the influence of cirque ice appears more clearly. Clasts of Cirque origin comprise 26% of the total with only 7% clasts of North origin and 19% of Adjacent origin. This significantly higher percentage of Cirque clasts as compared to North clasts supports the proposal that the landform was emplaced by active ice processes within the Great Gulf as opposed to a readvance of the LIS from the north.

From Intersection 56, proceed along a combination of ski trails passing Intersections 49, 44, 40, 50, and 54 to Intersection 38 near the Thumper 2 Site.

Stop 4: Thumper 2 Stone Count Site.

Please refer to Table 1 and discussion above.

From Intersection 38, proceed along a combination of ski trails passing Intersections 40 and 33 to Intersection 27 near the Libby Site.

Stop 5: Libby Stone Count Site.

Please refer to Table 1 and discussion above.

From Intersection 27, proceed along the Libby Trail to Intersection 19 and Stop 6.

Stop 6: Rangeley Boulder and Adjacent Boulder Fields.

This enormous boulder is comprised of rusty schist of the Lower Silurian Rangeley Formation (Sr and Src), whose closest outcrops capable of producing such a massive block are located on open cliffs approximately 1/2 mile to the west along the lower southerly walls of the cirque mouth. Its moderately subdued facets and generally but lightly abraded surfaces, along with its moderately weathered appearance, suggest it was not transported to its present location by a fast-moving mechanism like a debris flow. Blocks within such flows in the region generally appear much more sharply faceted, with only locally abraded areas on their surfaces, and with little surficial weathering. The process that transported this block appears to have consisted of a much slower-moving mechanism that permitted the block’s original facets to be moderately subrounded, its surfaces to be generally not locally abraded, and in contact with sufficient moisture to begin weathering its iron-bearing minerals. Being plucked from or otherwise falling onto, and then becoming incorporated within, a mass of active glacial ice moving toward the east from a cliff location appears a more viable candidate for this deposition mechanism with the time spent in the moving ice sufficient to create the features and general condition observed.

Meanwhile, smaller boulders found on boulder fields scattered elsewhere on this landform consist of this same lithology along with erratic lithologies that outcrop at locations higher up in the lower portion of the cirque (see Eusden, 2010). Most distinctive of these are members of the Devonian Littleton Formation (Dl), some of which have been transported to locations east of NH 16 in the vicinity of Nineteen Mile Brook (see Fig. 7 and Fowler, 2011). As indicated, much work remains to be done related to these “relocated” and erratic boulders before their use as positive evidence for post-LIS cirque glacial activity can be confidently established.

From Intersection 19, proceed along a combination of ski trails passing Intersections 14, 8, 5, and 2 and then proceed north about 200 yards up the Clementine Wash Trail.
Stop 7: The Peabody River Channel.

This Stop provides a viewpoint across the present-day channel of the Peabody River that was relocated to this position by the emplacement of the proposed terminal moraine and then eroded to nearly its current dimensions at the time the ephemeral lake catastrophically drained. The residual and largely flat bottom of the lake is located immediately south of the Stop, and the loose gravelly nature of the deposits eroded by the drainage process can be observed on the channel slopes.

From Stop 7, proceed back to Intersection 2 and turn left onto the “Glen Meadows Sluice” trail (see Fig. 7) and follow it to its intersection with the “Geepers” trail on the right. Turn right on this trail and follow it south to the pedestrian tunnel which leads back to the MWAR Base Lodge and parking lot.

Figure 7: Map of the Great Glen Trails System.

REFERENCES CITED


INTRODUCTION

The focus of this field trip will be the glacial geology (U-shaped valley, gorges, and an esker) but the exposed bedrock offers glimpses of Devonian granite with pegmatite intrusions. We will drive as a caravan in our own vehicles along the scenic Route 26 into Grafton Notch State Park, northwest of Bethel, Maine. Carpooling is highly recommended. The field stops are at five designated park landmarks and one sand pit. There is an optional strenuous hike with a 1000 foot elevation gain to Table Rock over 1.9 miles. This optional hike is best for experienced and adventurous hikers who enjoy lots of rock steps, bolted rebar steps, boulder scrabbles, and extraordinary views (not available in bad weather).

Logistics: Wear hiking shoes or boots with good traction, pants to prevent bug and tick bites, and layers in case the weather turns cold or rainy. Avoid steep or slippery places on ledges next to cliffs and fast-moving streams. There are toilets/outhouses at three of the stops (Screw Auger, Spruce Meadows and Old Speck) and at the initial meeting place. No shops are available on the way or in the state park, so please bring your food and water with you.
BEDROCK GEOLOGY

Metamorphic rocks
Grafton Notch cuts through part of the Mahoosuc Range, which includes metamorphic bedrock ~420 million years old. We will not be visiting any locations with metamorphic rocks, but the area includes:

OCAd: Ordovician Cambrian Dear River formation (slate, quartzite, phyllite)
Sp: Silurian Perry Mountain formation (sandstone, shale, quartzite)
SrAc: Silurian Rangeley formation "A" member lithic sandstone (sandstone, conglomerate)
Sr: Silurian Rangeley formation (mudstone, sandstone)

Intrusive rocks

Mafic and Felsic Intrusions. Highly altered Oam: Ordovician Ammonoosuc Volcanics (mafic and felsic volcanics, medium rank amphibolite facies) outcrop in the region, but we will be in the D1b(m),3: Devonian granite (muscovite accessory mineral) (Milton, 1961; Moench et al., 1995; Thompson, 2001). Three major orogenies lead to the formation of the Appalachian Mountains: Taconic, Acadian, and Alleghenian. The granite we will visit formed during the Acadian orogeny in the Devonian (410 to 400 Ma) when the collision of continental fragments, called Avalonia, and the North American paleocontinent, Laurentia, closed of the southern Iapetus Ocean (Eusden et al., 2013).

Bedrock features. Thrust faults and sheet jointing (caused by erosion of overlying rock above the granite releases stress) may be visible at Mother Walker Falls. Episodic glaciers eroded this landscape to create features such as small cirques, steep cliffs on the valley walls, and a U-shaped valley (Fig. 2 & 3). Rivers continue to flow through gorges and potholes created during the last ice age when the rivers were under pressure and contained large quantities of sediment, such as glacial flour.

![Figure 2. Photograph of the U-shaped valley in Grafton Notch from the Appalachian Trail on Old Speck. Route 26 and the Old Speck parking lot are visible for scale.](image)

SURFICIAL GEOLOGY

Quaternary geology

Fluvial deposits. Floodplains exist in the relatively wide portions of the Grafton Notch valley and adjacent to the Androscoggin River. From the meeting place to the first stop, we will drive over flat areas of farmland that take advantage of the rich and fertile soil of the floodplain.

Glacial deposits. The Laurentide Ice Sheet deposited till with many large erratics on the valley floor (Marvinney and Thompson, 2000). Some of the boulders in the valley resulted from rock fall from the steep valley walls, but there are plenty of rock lithologies different from those of the local bedrock. Glacial-fluvial deposits, such as eskers and outwash, contain large quantities of sand and gravel (Fig. 3). The Grafton Notch area includes the following Quaternary features:
Qs - Quaternary swamp, peat, silt, clay and sand, poorly drained areas leading to wetland formation
Qt - Quaternary till, poorly sorted glacial debris including large erratics
Qg - Quaternary gravel, glacial stream deposit, well-sorted sand and gravel, moderate to high permeability

Figure 3. Surficial Geologic Map of Grafton Notch State Park (modified from Caldwell, 1975).

ROAD LOG

MEETING POINT:
9:00 AM at the Androscoggin Rest Area (UTM 356200 m E 4923600 m N) along Routes 26, 2, and 5 just northeast of the Sunday River Brewing Company in Bethel. **Warning:** Large mammals are very common in this park, so please take caution as you are driving these roads. There are toilet facilities available at some but not all stops.

Lunch: Bring your lunch, water, snacks, bug spray, sunscreen, and cameras because there will not be any shops in the park.

Physical demands: Avoid steep or slippery places on ledges next to cliffs and fast-moving streams. The optional 2-3 hour hike is steep and includes rock steps, bolted rebar, boulder scabble, and a steep drop off at the top. By continuing on the optional hike, you acknowledge that you are a fit, capable hiker and are hiking at your own risk.

Entrance fees: There is a **$3 fee per vehicle** for the park; please pay this at the first stop (Screw Auger Falls).

Directions: From Rt 26 in Bethel: Drive north on Rt 26 and turn left in Newry to continue on Rt 26 (Bear River Rd).

This field trip is covered by Maps 10 & 18 in DeLorme's Maine Atlas and Gazetteer. Much of the information below comes from a glacial geology guide to Grafton Notch State Park (Thompson, 2001).

Mileage: Cumulative mileages given here may differ from those shown on your odometer, but the indicated distances between stops are generally accurate.
MEETING PLACE: Androscoggin River Rest Area (outhouse available), turn right out of parking lot and go N on Rt 26

2.7 Drive north to Newry, turn left at Bear River to continue on Route 26 and Bear River Road

12.1 STOP 1: Screw Auger Falls (entrance to parking lot on the left, outhouse available)

13.3 STOP 2: Mother Walker Falls (on the right)

14.1 STOP 3: Moose Cave (on the right)

15.8 STOP 4: Spruce Meadows Picnic Area (on the left, outhouse available)

17.6 STOP 5: Sand and gravel pit (on the left)

20.4 OPTIONAL STOP 6: Old Speck parking lot and Table Rock trailhead (turn around and drive south, look for the Appalachian Trail crossing and the parking lot on the right, outhouse available)

MEETING POINT: ANDROSCOGGIN RIVER REST AREA (45 min)
(UTM 356200 m E 4923600 m N, outhouse available) From Bethel, drive north on Rt. 26, cross over the Androscoggin River, pass the Sunday River Brew Pub and pull into the rest area on the right.

![Figure 4. Map of the Androscoggin River from Magalloway River and Lake Umbagog to Merrymeeting Bay and the Atlantic Ocean.](image)

The mighty Androscoggin River flows west from the New Hampshire side of Lake Umbagog, bends south, then east to flow back into Maine, bends north to Rumford, then southeast through Lewiston and mixes with the Kennebec River waters in Merrymeeting Bay and ultimately into the Gulf of Maine (Fig. 4). The Androscoggin is the third largest river in Maine and drops more than 1500 vertical feet over a 169 mile distance from Lake Umbagog to Merrymeeting Bay. This river was once known for its plentiful fish populations, inspiring my grandfather to write a story about a giant salmon that lived in the headwaters of Lake Umbagog:

"The Magalloway River has its birth within the sylvan mountains at the northwest corner of our great state of Maine, where it has a common meeting place with New Hampshire and Canada. Craggy mountains with gray granite ledges surmounted by tangles of spruce, fir, and cedars lead downward along hardwood ridges to valleys of hidden ponds and cascading crystal clear streams. This country of
magnificent splendor and serenity, far from the din of cities and madding crowds, is the setting for the story of MAGALLOWAY SAM!"

-Excerpt from Magalloway Sam by Dr. Lowell E. Barnes

The Androscoggin was a vital pathway for many fish species, including the Atlantic Salmon (Watts, 2017). After years of intense fishing, settlers built textile, lumber, and pulp and paper mills along the river and dumped pollutants, such as sewage, wastewater, phosphorous, dioxin, chlorine byproducts, lead and others into the river. By the 1960s, the Androscoggin had become one of the most polluted rivers in the United States, with dissolved oxygen levels reaching zero in the summer. At present, the Androscoggin has over 100 dams (32 for recreation, 16 for hydropower generation, 9 for flood control and storm water, 9 for reservoir storage, and 28 for ‘other’ purposes).

U.S. Senator Ed Muskie (a.k.a. ‘Mr. Clean’; native of Rumford, Maine; Bates alum ’36) was instrumental in the 1970 Clean Air Act and 1972 Clean Water Act. The Clean Water Act provided funding and legal mandates for sewage treatment plants along the Androscoggin and the river health and clarity have improved dramatically. The river is almost up to Class C standards (the lowest allowed in Maine) because the Gulf Island Pond section still has oxygen levels that are too low for fish (Watts, 2017). Recreation on and near the Androscoggin thrives now that the river looks and smells cleaner.

Research questions to ponder:
• How have dissolved oxygen levels changed over time (years and seasons)?
• Does flooding and erosion cause contaminated sediment to ‘reenter’ the environmental system?
• What are the research opportunities for water quality on Maine rivers?

STOP 1: SCREW AUGER FALLS. (1 hour)
(UTM 348900 m E 4937100 m N, outhouse available) Turn right out of rest area, drive north on Rt. 26 for 2.7 mi, turn left onto Bear River Road, which is a continuation of Rt. 26. Continue driving north on Rt 26 and at 12.1 mi, turn left into the Screw Auger Falls parking area. Please remember to pay your $3 park fee here.

Figure 5. Screw Auger Falls, Grafton Notch State Park, Maine (people for scale in both photos).
This location includes a 23-foot waterfall in a narrow, twisting gorge along the Bear River. Water levels in the spring are high and the gorge become louder and more dangerous to viewers in flood conditions. There are many pothole features within mountain streams in New England, and at this location we can see where two potholes connected forming a natural bridge. If the water is low, we may be able to see well-rounded stones of local and foreign origin.

The waterworn bedrock above the waterfall shows pegmatite vein intrusions in the granite. It is likely that the gorges follow a weakness in the bedrock, whether that weakness is a fault, pegmatite vein, joint, or something else, is uncertain. There are two main hypotheses regarding the formation of these gorges. They may have formed while the ice sheet retreated north of the region and contributed a large quantity of meltwater, or, more likely, they formed while the ice sheet covered the area and the subglacial water was under very high pressure. Either way, the stream was likely sediment laden (as evidenced by local eskers and outwash deposits), which causes more intense erosion than the modern stream.

Research questions to ponder:
- Do the gorges follow some type of weakness in the bedrock? If so, what is that weakness (fault, joint, pegmatite, etc.)?
- Does pegmatite erode more easily than granite?
- To form a gorge, does there need to be an overburden of thick ice to increase water pressure or could the sediment-laden streams create gorges without the pressure of an ice sheet?
- How long would it take to form a gorge of this depth under different scenarios of sediment content and pressure conditions of the water if we assume all of the erosion happened during the most recent ice age (~20,000 years ago).

STOP 2: MOTHER WALKER FALLS (40 min)
(UTM 3472700 m E 4937700 m N) Turn left out of the parking lot. Drive north on Rt. 26 for 1.2 mi and pull into the Mother Walker Falls turn off on the right. If possible, park at a slight diagonal to accommodate more vehicles.

If the weather is good from the parking lot, look south to the tree line and bald top of the nearby mountain. We are at a higher altitude now and these mountains support boreal plant and animal species. If possible, view the nearby cliffs from the parking area. The steep cliffs and somewhat flat and wide valley floor are indicative of glacial erosion. The classic U-Shaped glacial valley can be seen in many mountainous regions in the middle-latitudes, but typically due to mountain glaciers. Was this valley carved by relatively thin ice that was confined by the valley walls, or can ice sheets carve U-Shaped valleys while overtopping the peaks? The terminus of an ice sheet may appear lobate and confined by topography (e.g. Greenland Ice Sheet), and these lobes would appear to erode and flow more like mountain glaciers. Alternatively, when the Laurentide Ice Sheet covered Grafton Notch during the Last Glacial Maximum, ice would have been thicker in the valleys than over the mountain peaks, potentially leading to more erosion in the valley bottoms. Subglacial meltwaters might have funneled through these lower areas and caused more substantial erosion to the bedrock compared with glacial erosion on the peaks.

As we walk down to Mother Walker Falls, notice the abandoned carriage road. This early passageway was more dangerous and difficult than transporting materials along the Androscoggin River.

Mother Walker Falls is a narrow gorge more than 40 ft deep and 980 ft long. There is a series of cascading pools with a total drop of 100 ft. Along the path, if you find pegmatite outcrops, notice the crumbly feldspar.

Research questions to ponder:
- Could feldspar be the weak underbelly of the pegmatite?
- Is the bedrock weakness that the stream exploited here the same weakness at Screw Auger Falls?

STOP 3: MOOSE CAVE (40 min)
(UTM 346300 m E 4938500 m N) Please use caution while pulling back onto Rt. 26 north (right). Drive north 0.8 mi and pull into Moose Cave parking area on the right.
From the parking lot, you may be able to see the glacially eroded cliffs on both sides of the valley, including Table Rock (STOP 6) above. Along the path, note several large boulders (probably left by the ice sheet, but some could be rock fall). The gorge in Moose Cave was created in part by sub-glacial rivers under pressure, and the ceiling of the cave is a large granite slab that fell onto the gorge. Moose Cave is about 600 ft long and 50 ft deep. Brewer (1978) speculated that the gorge may have developed along a fault in the granitic bedrock, but this origin has not been confirmed. Major faults in Maine are sometimes marked by prominent quartz veins or zones in which the rock is broken or contorted. Perhaps these gorges follow other types of weaknesses in the bedrock, such as pegmatite veins.

Research questions to ponder:
- What is the age and origin of the Moose Cave gorge?
- Is there evidence for past stream erosion across the valley floor or is it confined to the gorges?
  Look for striations on quartz veins, determine direction (east) parallel to the river and the valley.

STOP 4: SPRUCE MEADOWS PICNIC AREA (1 hour)
(UTM 345600 m E 4940800 m N, outhouse available) Please use caution while pulling back onto Rt. 26 north (right). Drive north 1.7 mi and turn left into the Spruce Meadows Picnic Area.

From the parking lot, walk along the path to the west side of the picnic area. Notice the abundance of sand and how flat this area is compared with what we've seen today. At the western edge of the picnic area, walk up onto a small ridge. This ridge is part of an esker. From one of the picnic sites, you can see views of Old Speck and the 'Eyebrow' cliffs on the mountain. From another picnic site, you can look out over the wetland, which is poorly drained, possibly due to fine sediments left by glacial meltwater.

Figure 6. Topographic map of the Spruce Meadow Picnic Area, Rt. 26, the Swift Cambridge River and the red lines represent ridges of gravel deposits (eskers). Modified topographic map.

Research questions to ponder:
- What determines whether glacial meltwater will erode into bedrock or result in an esker deposit?
- What did the proglacial environment look like as the glacier retreated through this area?
- What do the wetland - fine sediment deposits relate to?
- Is the flat picnic area part of a level outwash surface? Or perhaps a delta?

STOP 5: SAND AND GRAVEL PIT (1 hour)
(UTM 344300 m E 4943000 m N) From the picnic area, turn left onto Rt. 26 north and drive 2.2 mi where you will turn left onto York Pond Road. Drive slowly on this dirt road as there may be other recreational drivers. The parking area is in a gravel pit on the left not far from the road.
This gravel pit is not in the park and we ask that you are respectful to the owners and others on the field trip while exploring these deposits. Please do not climb to the headwall and be careful of rocks rolling from the walls.

There are several features to notice in this pit:
1. Graded gravel beds
2. Finer, silty/sand deposits
3. Rounded cobbles

This sand and gravel pit (Fig. 6 & 7) is part of an esker deposit (Fig. 3, 6 & 7) that is probably a continuation of the Spruce Meadows esker seen at the previous stop. Here we can see a diversity of lithologies, and a relatively high concentration of sand as opposed to cobbles.

Research questions to ponder:
• Is there a way to tell the location of an esker within an ice sheet (subglacial, englacial, supraglacial)?
• Are eskers only likely to be deposited in areas of subdued topography and not in narrow valleys?
• Does the grain size distribution of sediment within an esker depend on water speed?
• What is the significance of the finer, laminated sediments near the top of the esker deposit?

STOP 6: OLD SPECK PARKING LOT AND TABLE ROCK TRAILHEAD (2-3 hours)
(UTM 345400 m E 4939200 m N, outhouse available) If you are not planning on participating in the hike, you may now return to Bethel by following Rt. 26 south. If you are joining us on the hike, turn right onto Rt. 26 south and drive 2.8 mi then turn right into the Old Speck parking lot.

The optional strenuous hike begins here. Cross the road to the east side of the valley (use caution while crossing the road) and follow the signs to Table Rock (Caution: Do not follow the Appalachian Trail). The loop hike is about 1.9 mi long with a 1000 ft elevation gain and takes ~2-3 hours (Fig. 8). This hike is best for experienced and adventurous hikers who enjoy lots of rock steps, bolted rebar steps and scenic views. The connection to the AT and return to the parking lot is relatively gentle and easy to walk down.

Along the hike, keep your eyes out for various bedrock and surficial features. You may notice large erratic boulders deposited by the Laurentide Ice Sheet or more recently dislodged from the valley wall. If you are a well-rounded naturalist, note the boreal species (e.g. White-throated sparrow, balsam fir) along the trail. Closer to the lookout, we will climb over a scree slope. Think about the size of these boulders and what might be causing the scree to form and why the boulders are a typical size (e.g. jointing). From the lookout, we may be able to see the Grafton Notch State Park in its entirety, as well as the U-shaped valley. View the summit of Old Speck (4170 ft, 1270 m elevation), the highest peak of the Mahoosuc Range, and the steep, glistening cliffs below it. Take a close look at the Table Rock lithology, note the wide pegmatite vein and how it is essentially flush with the granite surface.
Figure 8. Topographic map of the Table Rock loop trail, contours are 20 foot increments, side of boxes represent 1 mile. From the parking lot on the west side of Rt. 26, we will cross the road to the east side and bear right onto the Table Rock trail. After summiting, we will connect with the Appalachian Trail and walk back to the parking lot. Modified topographic map.

Figure 9. Left image shows Grafton Notch from Table Rock and right image shows ‘the Eyebrow’, Rt 26, and the steep drop off from Table Rock.

Research questions to ponder:
- Does the prominent pegmatite vein in Table Rock suggest that pegmatite is as strong as granite?
- Does a U-shaped valley only form when ‘thin’ ice is constrained by the valley walls and does not overtop them? Or can U-shaped valleys form while under a mile-thick ice sheet?
- What caused the valley to be in this orientation (fault, contact, weaker lithology, etc)?
- Where there mountain glaciers in this area and if so, when did they melt away completely?
REFERENCES


All photographs were taken by A. Doughty between 2015 and 2017.
STRATIGRAPHIC AND STRUCTURAL TRAVERSE OF MOUNT MORIAH AND THE WILD RIVER WILDERNESS AREA

Tim Allen, Department of Environmental Studies, Keene State College, Keene, NH 03435-2001

INTRODUCTION

East of Pinkham Notch and south of the Androscoggin River valley lies a large roadless area of the White Mountain National Forest containing the Carter–Moriah mountain range and the Wild River valley (Fig 1). Original mapping in this and adjacent areas was done by M. P. Billings and others in the 1940’s and 1950’s (1941, 1946, 1975, 1979), and Billings’ work lays the foundation upon which all subsequent work in this area is based. Since the time of Billings’ maps, however, further developments have been made in understanding the litho-stratigraphy of Siluro-Devonian metasedimentary rocks throughout New Hampshire and western Maine (Osberg et al., 1985; Moench & Pankiwskyj, 1988; Lyons et al., 1991). Attempts to apply the new lithostratigraphy to this region met with limited success, particularly in the migmatites that make up the Carter–Moriah and Wild River area (CM–WR; Hatch et al., 1983; Hatch & Moench, 1984; Hatch & Wall, 1986). The CM–WR area is shown on current maps as “undifferentiated sedimentary rocks in areas of extreme migmatization” (Osberg et al., 1985; Lyons et al., 1991).

The rocks are indeed migmatized, although in general the line bounding these “undifferentiated” rocks does not necessarily represent a “migmatite front” but rather the limits of easily mappable terrain (Hatch & Wall, 1986, page 146). Locally, however, sharp “migmatite fronts” can be clearly defined, separating un-migmatized schists from intensely migmatized gneisses of the same parent lithology. Associated with detailed studies of such a front in Pinkham Notch (trip C6 of this volume; Allen, 1992, 1996b, Allen et al, 2001), I have undertaken a third-generation mapping effort attempting to differentiate the stratigraphy and structure of the migmatites in the CM–WR area.

REGIONAL GEOLOGIC SETTING

The migmatites in the CM–WR region lie along the axis of the Central Maine Terrane (CMT, Fig. 2; Zen et al., 1986) and are central to the broad region of high grade metamorphism of the Acadian Orogen. To the west and north, the CMT abuts the Bronson Hill Anticlinorium, which represents a magmatic arc of Ordovician age with a thin cover of Silurian and Devonian sediments. The CMT is regarded as an eastward thickening sedimentary basin adjacent to the arc, filled with Silurian age shales, quartzites, and calcareous rocks deposited in a deep water anoxic environment, and topped by early Devonian volcanics and turbidites from an eastern source (Moench & Pankiwskyj, 1988). This basin, together with the Bronson Hill arc, was multiply deformed and metamorphosed during large scale crustal thickening of the Acadian Orogeny.

The structure of the CMT consists of two major synclinoria separated by the Central New Hampshire Anticlinorium (Eusden, 1988). This anticlinorium acts as a “dorsal zone” from which originate west-vergent structures to the west and east-vergent structures to the east (Eusden, 1988). In fact, large scale west-vergent fold nappes carried high-grade rocks from the western CMT over the Bronson Hill Anticlinorium (Thompson et al., 1968; Chamberlain et al., 1988). The Central New Hampshire Anticlinorium is marked not only by exposures of the oldest rocks, but is also the locus of anomalous metamorphic “hot spots” (Fig. 2; Chamberlain & Lyons, 1983; Chamberlain & Rumble, 1988) and migmatite zones (Fig. 2; Wilson, 1969; Billings & Fowler-Billings, 1975; Englund, 1976; Eusden, 1988).

The plutonic rocks associated with the Acadian metamorphic high, and the Central New Hampshire Anticlinorium structure of the CMT, belong to the New Hampshire Plutonic Series. Petrologic and geochemical studies suggest that they are anatectic crustal melts (Duke, 1978; Clark & Lyons, 1986; Lathrop, et al., 1994). The oldest of three generations (400 to 390 Ma; Lyons & Livingston, 1977; Barreiro & Aleinikoff, 1985) are the synmetamorphic and syntectonic Kinsman, Bethlehem, and Spaulding groups. These are large, shallow sheet-like bodies that are intimately involved with Acadian nappé structures (Nielsen et al., 1976; Thompson et al., 1968), and have isograds mapped across them (Chamberlain & Lyons, 1983). At about 380 Ma, post-tectonic plutonism resulted in abundant small bodies of two-mica granites known as the Concord group (Lyons et al., 1982; Harrison et al., 1987). The last group of the New Hampshire Plutonic Series yield ages of about 320 Ma (Lyons et al., 1991; Osberg et al., 1985), volumetrically significant only in Maine, causing late thermal metamorphism.
Figure 1: Geologic map of the Carter–Moriah and Wild River area, from this study and from Billings & Fowler-Billings (1975); Hatch & Wall (1986); and Hatch & Moench (1984). There are numerous granites and pegmatites too small to show. Note direction of north arrow. Field trip stops are indicated by the circled numbers.
Figure 2: Map of the Acadian metamorphic high in New England, showing shallow level plutonism in Maine, “hot spots” and migmatite zones (schematic) and the Central New Hampshire Anticlinorium in New Hampshire, and deep-level, high-grade gneiss in Massachusetts. CMT labels the axis of the Central Maine Terrane; and BHA the Bronson Hill Anticlinorium. After Chamberlain & Robinson (1989).

METASEDIEMNTARY ROCKS

Although Billings & others (1941, 1946, 1975, 1979) did not recognize the stratigraphic sequence as it is now understood (Hatch et al., 1983), they did recognize some of the important lithologies upon which the modern lithostratigraphic sequence is based. Thus their maps provide a useful starting point for work in this region. All of the
rocks previously assigned to the Devonian Littleton Formation (Billings & Fowler-Billings, 1975), are here subdivided into a Siluro-Devonian lithostratigraphy similar to that now recognized elsewhere in New Hampshire and western Maine, based on correlations made by Hatch et al., (1983). Units include the Silurian Rangeley, Perry Mountain, Smalls Falls, and Madrid Formations, as well as the Devonian Littleton Formation. The Perry Mountain Formation had not previously been recognized in much of this area (Hatch et al., 1983; Hatch & Wall, 1986). The metamorphism of these rocks is discussed elsewhere (trip C6 of this volume; Allen 1992, 1996b).

Rangeley Formation

The Rangeley Formation is the oldest and the most abundant rock in the CM-WR area. The Rangeley consists of gray and rusty orange weathering coarse-grained pelitic schists and migmatite gneisses, with minor interlayered quartzites, and abundant, distinctive calc-silicate pods. These rocks are generally biotite schists or gneisses, containing quartz, biotite, muscovite, albite plagioclase, garnet and sillimanite. An important accessory mineral is pyrrhotite, whose presence is responsible for the orangish weathering color these rocks often have. It is the Rangeley that is most often migmatized in this area, although the intensity of migmatization is highly variable. I have not further subdivided the Rangeley in this area, as has been done elsewhere (Lyons et al., 1991; Eusden, 1988; Moench & Boudette, 1970).

Original bedding can occasionally be recognized where well-defined beds of contrasting composition occur. Graded beds or other topping indicators have not been observed. More often, original bedding cannot be recognized, although the schistose foliation and migmatitic layering are generally bedding-parallel. Many weathered outcrops present a massive appearance, however, and the fabric of the rock is best seen on water-washed polished surfaces. The character of the foliation in these rocks is very rough, as the rock is very coarse grained, often contains large, un-oriented muscovite spangles, and is typically migmatitic. Billings & Fowler-Billings (1975) described three types of migmatite fabrics: “layered”, “podded”, and “wispy.” These generally grade into one another, and are often highly distorted and incoherent—“swirly.”

The calc-silicate pods often occur in groups along bedding planes—often associated with more quartz-rich beds—or in clusters within larger quartz-rich pods or blocks. The pods resist migmatization, and in the migmatites they usually remain parallel to the foliation and gneissosity, although occasionally they are at odd angles to it. The pods are usually concentrically zoned in composition and rimmed by a weathered-out moat, while the cores stand up in relief above the surface of the surrounding schist or gneiss. These pods are thought to be metamorphosed calcareous concretions (Billings & Fowler-Billings, 1975), although some may be rip-up clasts or blocks of reef material carried from the shelf into the deep basin (Guthrie, 1984; Guthrie & Burnham, 1985), during the rapid sedimentation characteristic of the Rangeley (Moench, 1970). Elsewhere, calc-silicate pods in the Rangeley have been interpreted as boudins of once continuous calcareous sandy beds or lenses (Eusden, 1988). The appearance of “pods within block” features suggests that they are concretions or clasts and not boudins.

Also distinctive in the Rangeley of this area are occasional exotic quartz, quartzite, or granitoid pebbles and cobbles, usually as isolated individuals rather than in recognizable conglomeratic horizons, but clearly of sedimentary origin. Zones of matrix-supported polymictic conglomerate are observed in the Rangeley in other regions of Maine and New Hampshire (e.g., Moench & Pankiwskyj, 1988; Allen, 1984).

In addition to the calc-silicate pods and the cobbles, larger (1 to 5 meter) exotic blocks have also been observed. Often the lithology of these blocks is suggestive of the rock units that overlie the Rangeley. These blocks may have an origin similar to the rip-up clast model proposed for the calc-silicate pods. These features may represent a sedimentary or olistostromal mélangé indicative of rapid, sometimes chaotic, sedimentation in a submarine fan environment (Eusden et al., 1996). Moench (1970) discusses extensive pre-metamorphic deformation in the Rangeley Formation due to extremely rapid sedimentation and the build-up of high fluid pore pressures.

Perry Mountain Formation

The Perry Mountain Formation consists of interbedded gray quartzites and schists, often bearing pytymagically folded coticules, and occasionally, calc-silicate pods. The quartzite and schist interbeds range from 1–2 cm to 5–10 cm in thickness, and occasionally quartzite dominates. The contacts between the quartzite and schist interbeds occasionally show grading, but are more often sharp. As in the Rangeley, the calc-silicate pods occur strung out
along bedding planes, although the “pods within block” features are unique to the Rangeley. In the Perry Mountain, some of the pods are clearly boudins. Locally the schist layers have well developed sillimanite nodules, pseudomorphs after andalusite. The Perry Mountain is rarely migmatized in this area.

Most of the rocks shown as Perry Mountain Formation in Fig. 1 were not recognized as such by Hatch & others (1983, 1984, 1986), but instead were assigned to the Rangeley Formation. I have assigned these rocks to the Perry Mountain, however, because of the abundance of quartzites and the nature of the bedding, and because they appear in the proper sequence between Rangeley schists and migmatites and Madrid or Smalls Falls rocks.

Smalls Falls Formation

The Smalls Falls Formation is a highly graphitic and sulfidic schist with sulfidic micaceous quartzites. These rocks often weather a dull brown color to very dark rusty-red. The intense weathering due to the abundant pyrrhotite makes the rock very crumbly—as a result outcrops are not very resistant to erosion. The micaceous quartzite beds often breakdown to a characteristic gritty sand. These rocks are often finely laminated—flaggy—with bedding less than 1 cm thick. In many places, however, any semblance of bedding has been destroyed due to incompetent structural behavior.

Madrid Formation

The Madrid Formation consists of green calc-silicates, and fine grained plagioclase-biotite-quartz “salt & pepper” granofels. These rocks are very well bedded, weathering to produce distinctive tabular blocks and slabs, from 2 to 10 cm thick. The Madrid is generally not well exposed, being less resistant to erosion than adjacent quartzites and sillimanite schists; and is fairly thin throughout the region (never more than a few tens of meters). The best exposures are found in streams, and are quite distinctive. The Boott Member of the Littleton Formation and other lime-silicate rocks as mapped by Billings & others (1941, 1946, 1975, 1979) are now thought to be the Madrid Formation (Hatch et al., 1983).

Littleton Formation

The Littleton Formation in this area is composed of aluminous schists with interbedded quartzites, with generally very good graded bedding. These rocks are generally silver-gray in color, and have distinctive, abundant sillimanite pseudomorphs after andalusite, often up to 5 cm long. Locally, these sillimanites define a strong lineation; elsewhere they may occur as “turkey tracks” on foliation planes. Rocks of the Littleton Formation that have been migmatized often have a very different texture or fabric from the migmatites of the Rangeley. This may be best described as a “stringy” or “sinewy” texture, as the migmatite leucosomes appear to define a lineation within the melanosome/mesosome matrix. The calc-silicate pods abundant in the Rangeley are absent from the Littleton Formation.

In the Moriah Brook section (Stop 6, Fig. 1) there are conglomeratic horizons in contact with exposures of the Madrid Formation. Sequentially, these conglomerates appear to belong to the lower Littleton Formation. These conglomeratic horizons may be similar to the “Wild Goose Grits” mapped within the Littleton by Eusden & others (1987; 1988) south of this region. These conglomerates, and the apparent local absence or extreme thinning of the Madrid Formation, may indicate a local unconformity at the base of the Littleton.

IGNEOUS ROCKS

Billings & Fowler-Billings (1975) mapped several igneous rock types in the CM–WR area, belonging to the Devonian New Hampshire Plutonic Series, and to the Jurassic–Cretaceous White Mountain Plutonic-Volcanic Series. Rocks of this later series consist of volcanic vent agglomerate and diabase dikes, of minor importance to this study. Plutonic rocks in this region occur in two main modes—as large, mappable plutons such as the Peabody River Stock (Fig. 1; Billings & Fowler-Billings, 1975), and as smaller, more heterogeneous granitic and pegmatitic bodies and dikes that occur throughout the migmatite zone. As Billings & Fowler-Billings note (1975, p. 64), these rocks are difficult to portray on the geologic map because of their small size, wide distribution, and intricate contact relationships.
A larger body of this latter type occupies an area of about 10 km², extending from Pinkham Notch proper north nearly to Emerald Pool, and east almost to the summit of Wildcat Mountain (Fig. 1, Stop 9). It underlies the slopes of the Wildcat Mountain Ski Area, from which is derived the name I have assigned to this type of rock—the “Wildcat Granite.” The Wildcat Granite can be described as granite only in generalities—there are clearly at least two different phases. One consists of medium grained, whitish-weathering clean two-mica granite (hereafter, the “G” phase, for granite). The second is much coarser grained, orangish-weathering granitoid (hereafter, the “R” phase, for the Rangeley Formation), also bearing both muscovite and biotite, but with much more abundant biotite than in the “G” phase. Within this second phase are abundant calc-silicate pods, identical to those found in the metasediments, rimmed by strong reaction zones. Textures and mineralogy of the “R” phase suggests that it may be formed from completely melted and recrystallized Rangeley schists. Both the “G” and “R” phases are extensively intermingled in a highly complex fashion. Wispy biotite-rich schlieren can be observed throughout. The contact between the granite and the surrounding migmatitic metasedimentary rocks is gradational—not a sharp intrusive contact. Similar occurrences of Wildcat-type granitoids are found throughout the migmatite zone, usually associated with pegmatites. The Wildcat Granite is very similar to some exposures of the Blackwater Pluton of the Spaulding Group of the New Hampshire Plutonic Series (Lyons, 1988; Duke, 1978). The Spaulding Group is considered to be late-tectonic, and has been dated at 392 ± 5 Ma (Lyons & Livingston, 1977). Beyond the explanation given above, it is interesting to speculate why Billings & Fowler-Billings (1975) might not have shown the Wildcat Granite on their map—granitization and the origin of granites were “hot topics” at the time they were doing this mapping (1950’s).

Two generations of granitic pegmatite are observed throughout the migmatite zone. The first is generally gray in color and contains quartz, albitic plagioclase, muscovite, and spessartine garnet. These pegmatites are sometimes slightly foliated, and often have gradational contacts with adjacent migmatites. The second generation pegmatites, which cross-cut the earlier pegmatites and the granites, are white in color, and contain quartz, albitic plagioclase, muscovite ± potassium feldspar, and abundant tourmaline. These white pegmatites usually have sharp contacts with adjacent rocks. Both sets of pegmatites can be either cross-cutting or parallel to the structural trends in the adjacent rock—the second-generation pegmatites also cross-cut the Wildcat Granite. Eusden (1988) has also observed two stages of pegmatite similar to that described here, associated with another migmatite zone to the south.

The Peabody River Stock (Fig. 1) is a homogeneous and undeformed two-mica granite or quartz monzonite of the Concord Group of the New Hampshire Plutonic Series (Billings & Fowler-Billings, 1975). This homogeneous granite is distinct from the Wildcat Granite, which is extremely heterogeneous. It appears to have no relationship to the migmatization. In fact, at the migmatite front to be visited on trip C6 (this volume), the rocks become more intensely migmatized as one moves away from the Peabody River Stock. The Concord Group granites represent post-tectonic plutonism at about 380 Ma (Lyons et al., 1982; Harrison et al., 1987), which is consistent with the undeformed nature of the granites in Peabody River Stock, and suggests that its emplacement post-dates the migmatization, and the associated Wildcat Granite.

**STRUCTURE**

Billings & others (1941, 1946, 1975, 1979) undertook extensive detailed study of geologic structures in the CM–WR area and adjacent Mount Washington region. Their analysis of the small scale structures and structural fabrics is excellent, however their interpretation of the larger structure was limited by the poor stratigraphic control (Billings & Fowler-Billings, 1975, page 57).

The mesoscopic, measurable structural features observed in these rocks are dominated by a foliation or schistosity defined by the alignment of micas. This foliation is generally bedding-parallel. Where the rock is migmatized, the leucosomes often lie in planes parallel to this foliation. Many of the rocks also have a strong lineation that is the result of small scale crenulation folds of the primary foliation, crenulation and alignment of micas, and the alignment of minerals such as sillimanite. The mineral and crenulation lineations are parallel to the axes or hinge lines of minor folds. Rarely can a good determination of the orientation of the axial plane to these fold features be made. A spaced cleavage, that may be axial planar to F3 folds, is observed only along the western boundary of the migmatite zone.

Overall, the structural trend in the migmatite zone is similar to that shown on Billings & Fowler-Billings (1975) map of the Gorham 15° quadrangle, as rock units and planar features strike northeast throughout the migmatite zone (Fig. 1). This trend is due to tight upright anticlinal and synclinal folding. This folding is demonstrated by the
Figure 3: Equal Area stereonet projections of: Contoured density of poles to bedding and foliation planes from the Carter Moriah and Wild River area (A) and from Pinkham Notch (D); Fold axes, crenulation lineations and mineral lineations (undifferentiated) from the Carter–Moriah and Wild River area (B) and Pinkham Notch (E); poles to planar pegmatite contacts from the Carter–Moriah and Wild River area (C); and Poles to spaced cleavage (+) and F3 axial planar cleavage (+) from Pinkham Notch (F).
Figure 4: Geologic and structural map of the Pinkham Notch, NH, study area. F2 and F3 folds labelled. Rocks at Emerald Pool are part of a disrupted block of the Rattle River – Mt. Moriah Syncline. This area is the subject of Trip C6, this volume (see also Allen, 1996b).

distribution of poles to bedding and foliation planes in an equal area stereonet diagram, or pi diagram (Fig. 3A). This folding must be at least F2, as it folds a previous foliation. However, no F1 fold axes have yet been identified. This
Cryptic F1 folding event may have produced large recumbent isoclinal fold and thrust nappe structures, probably eastward vergent, similar to the F1 events observed elsewhere throughout New Hampshire (e.g., Eusden, 1988). F2 fold axes and lineations are plotted on an equal area stereonet in Fig. 3B, and plunge shallowly alternately to the northeast and to the southwest.

The F2 folds would have refolded the cryptic F1 nappes, and further mapping in the area may be able to identify these nappes through the map pattern. For example, the younger rocks (Littleton Formation) in the cores of these synclines do not cross the ridgelines or mountain tops. Where these synclines intersect ridgelines or mountain tops, older rocks (Rangeley Formation) outcrop. These older rocks form the upper plate of a recumbent synclinal nappe that has been refolded down into the cores of the crossing synclines.

The F2 folding event appears to have produced three major synclines, with intervening anticlines. Our traverse (Fig. 1) will encounter the first syncline in the upper reaches of the Rattle River north of Mt. Moriah, and the second syncline in the middle stretches of Moriah Brook. The third syncline is on the slopes between the Wild River and the Basin Rim. A small subsidiary syncline may occur between Howe Peak and Shelburne Moriah Mountain. These structures are difficult to trace along strike, some evidence in fact suggests that these structures may be discontinuous or disrupted.

This structural disruption is depicted in Fig. 1, notably by the block of rocks at Emerald Pool, shown as an extension of the F2 Rattle River—Mt. Moriah Syncline (Fig. 4). My interpretation differs significantly from previous maps of this area (Billings and others, 1941, 1975; Hatch & Wall, 1986). Past interpretations of these rocks have tried to relate them to the belt of Madrid and Smalls Falls just to the west (Fig. 4; the Boot Member of the Littleton Formation, Billings & Fowler-Billings, 1975), that separates the un-migmatized schists of the Littleton Formation on Mt. Washington from the migmatites of the Rangeley Formation in Pinkham Notch and to the east. Detailed mapping shows that the rocks at Emerald Pool are isolated within migmatites of the Rangeley Formation. There are numerous other examples of such isolated blocks along F2 fold trends (Fig. 1), with migmatites truncating the bedding of less-migmatized units. One possible explanation might be that of granitic magmas migrating upward through the crust preferentially along these upright F2 axial planes (Fig. 5; although pegmatite dikes in the area show no preferred orientation (Fig. 3C)). This upwelling may have driven the migmatization process (trip C6, this volume) creating highly mobile rocks; disrupting the F2 structures and skewing F2 fold axes (Fig. 3B).

Along the western boundary of the migmatite zone, structures are somewhat different, as there is a strong sense of folding about west directed fold axes (Fig. 4; and Billings & others, 1941, 1975). This folding is deemed to be a third generation of folding (F3), and probably post-dates the migmatization of the rocks to the east. The relative timing of the folding is indicated by the truncation of the migmatite front against map scale F3 folds (Fig. 4). Here also, the bedding and foliation planes dip only moderately to the west (Fig. 3D)—in the migmatite zone the dip tends to be much steeper (Fig. 3A). The poles to bedding and foliation planes from Pinkham Notch also roughly define a girdle on the equal area stereonet (Fig. 3D), that is consistent with the trend and plunge of fold axes and lineations in the area (Fig. 3E). A spaced cleavage, that may be axial planar to these F3 folds, is observed only along this western boundary of the migmatite zone. Poles to these cleavage planes are shown on an equal area stereonet projection in Fig. 3F. Comparison with Fig. 3D suggests that the cleavage is roughly parallel with the dominant trend of bedding and foliation.

There appears to be some fault motion, with cutting out of units, along the western boundary of the migmatite zone, perhaps associated with this F3 folding (Fig. 4). The axial planar cleavage of the F3 folds is roughly parallel to the presumed orientation of the fault surface. It is possible that this faulting and folding may represent downsloping of the un-migmatized rocks to the west related to the upwelling of granite magmas through the migmatite zone. On the other hand, the faulting may be related to intrusion of small necks of White Mountain Magma Series volcanic vent agglomerate that occur along this trend (Billings & others, 1975, 1979). Of course, these intrusions may have re-activated pre-existing structures related to the boundary of the migmatite zone.

**DISCUSSION AND CONCLUSIONS**

In summary, I recognize a four-stage structural history consisting of a possible east or northeast vergent isoclinal fold and thrust nappe (F1), refolded by a series of nearly upright anticlines and synclines with axes plunging gently alternately north-northeast and south-southwest (F2). These folds are disrupted and the F2 fold axes
skewed apparently by intrusion of granitic magmas (probably along F2 axial planes) and the development of highly mobile partially melted migmatites. Finally, there is open folding about moderately westward plunging axes (F3) developing an axial planar spaced cleavage, along the western margin of the migmatite zone. Faulting along this boundary may be related to the F3 folding and the upwelling of granitic magmas in the migmatite zone, or may be due to later volcanic activity.

With the exception of late faulting, all the deformation and metamorphism in this area is presumed to be Acadian in age. Eusden & Lux (1994) report Ar40/Ar 39 ages for metamorphic muscovites from this area (including migmatite outcrop #036 of Allen, 1992) of 300 to 275 Ma. The muscovite samples were collected over a vertical relief of 1.5 km, and suggest very slow cooling and uplift rates at that time (Eusden & Lux, 1994). These results suggest that metamorphism and deformation in this area could not have been related to late stage magmatism, such as the Carboniferous Sebago Batholith (Osberg et al., 1985), but must be Acadian.

Eusden (1988) and Lyons et al. (1991) drew the trace of the Central New Hampshire Anticlinorium through the CM–WR area (Fig. 1). My mapping has confirmed that the metasedimentary rocks in this zone are predominantly the Silurian Rangeley Formation, the oldest unit in the Central Maine Terrain of New Hampshire. This is consistent with the placement of the Central New Hampshire Anticlinorium here. In addition, the dominant structures in the CM–WR zone are very nearly upright (Fig. 3A), which is consistent with the upright structures of Eusden’s (1988) central “dorsal zone.” Structures to either side of the central dorsal zone tend to be inclined or recumbent. Other migmatite zones similar to the one studied here, and other metamorphic “hot spots,” are also centered on the Central New Hampshire Anticline or “dorsal zone” (Eusden, 1988; Chamberlain & Lyons, 1983; Chamberlain & Rumble, 1988). Most of the plutons of the Spaulding and Concord groups of the New Hampshire Plutonic Series are also found within the Central New Hampshire Anticlinorium. This suggests that anticlinorium structures forming the cores of orogenic belts may provide a structural control on pluton migration through the crust, or vice versa (Fig. 5). Additionally, this pluton migration might be responsible for the migmatization of these rocks (trip C6, this volume, Allen, 1992, 1996b). Comparison of the cartoon depicted in Fig. 5 with the map of the Acadian Orogen in Fig. 2, furthers the concept of the New England Appalachians as a surrogate crustal section (Chamberlain & Robinson, 1989; Rodgers, 1970, p. 114).

![Cartoon depicting granitic magma migration from deep crustal levels through the crust along preferential pathways related to the “Dorsal Zone” (Eusden, 1988), disrupting structures and driving anomalous “hot spot” metamorphism and migmatization.](image)

**Figure 5:** Cartoon depicting granitic magma migration from deep crustal levels through the crust along preferential pathways related to the “Dorsal Zone” (Eusden, 1988), disrupting structures and driving anomalous “hot spot” metamorphism and migmatization.

**ACKNOWLEDGEMENTS**

This research was supported by National Science Foundation Grant EAR–8957703 awarded to C. Page Chamberlain, by National Science Foundation Grant EAR–9104553 awarded to Joel Blum and C. Page Chamberlain, and by Geological Society of America Research Grant 4357-90 awarded to Tim Allen. Structural orientation data were analyzed using the Stereonet program developed by Rick Allmendinger of Cornell University. Thanks to the following for their advice, comments on previous versions of related manuscripts and/or for

---

**Figure 5:** Cartoon depicting granitic magma migration from deep crustal levels through the crust along preferential pathways related to the “Dorsal Zone” (Eusden, 1988), disrupting structures and driving anomalous “hot spot” metamorphism and migmatization.
discussions on earlier field trips: Page Chamberlain, Joel Blum, Leslie Sonder, Doug Rumble, Dyk Eusden, John Lyons, and Bob Moench, and students and colleagues at Dartmouth and Keene State.

**“ROAD” LOG**

**MEETING POINT:** Assemble (7:00 AM!) first at the Wild River Campground at the end of the Wild River Road off of the Evans Notch Road (ME/NH 113) (see (A) on Fig. 1), consolidate into as few vehicles as possible—leaving as many as possible at Wild River—and proceed to the Rattle River Trailhead (Appalachian Trail) on Route 2 three miles east of Gorham. Almost the entire trip will be on foot, over the Rattle River, Moriah Brook, and connecting trails (approximately 13 miles, with 3300 feet of vertical relief—“book time” of almost 9 hours). The area between the Carter-Moriah ridge and the Wild River is a federally designated Wilderness Area, with restrictions on the size of hiking groups. Be prepared to spend a full day hiking in the mountains—wear appropriate boots and clothing, and bring plenty of food, water, and extra clothing. At the end of the hike, we will consolidate into the vehicles we left at Wild River and return to Rattle River and thence to Bethel.

Maps (USGS 7.5 minute 1:24,000 quadrangles): Shelburne, NH-ME; Wild River, NH-ME; Carter Dome, NH. The AMC Carter-Moriah Range Trail Map and the AMC White Mountain Trail Guide are also useful. The best map for the Pinkham Notch area is Washburn’s “Mount Washington and the Heart of the Presidential Range” (1988, 1:20,000).

**Mount Moriah Traverse, Trail Mileage**

<table>
<thead>
<tr>
<th>Mileage</th>
<th>Description</th>
</tr>
</thead>
<tbody>
<tr>
<td>0.0</td>
<td>Rattle River Trailhead (Appalachian Trail), three miles east of Gorham on US route 2. Proceed up the Rattle River trail parallel to the Rattle River (south).</td>
</tr>
<tr>
<td>1.7</td>
<td>Rattle River Shelter</td>
</tr>
</tbody>
</table>

**STOP 1. PERRY MOUNTAIN or LITTLETON and RANGELEY** (45 minutes): Outcrops in the river downstream of Shelter of pegmatite and aluminous gray migmatite gneiss with preserved bedding layers, possibly the Perry Mountain formation. At the shelter and upstream, extensive outcrops of orange Rangeley migmatite gneiss with pods, and extensive pegmatite intrusions.

<table>
<thead>
<tr>
<th>Mileage</th>
<th>Description</th>
</tr>
</thead>
<tbody>
<tr>
<td>3.2</td>
<td>Stream crossing below pool and falls</td>
</tr>
</tbody>
</table>

**STOP 2. SMALLS FALLS, MADRID and/or LITTLETON** (15 minutes): Bordering the pool, rusty and flaggy quartzites and schists of the Smalls Falls formation, and at the falls above the pool, layered gray aluminous schist cyclicly interbedded with fine laminations of granofels, possible graded bedding, of upper Madrid or lower Littleton affinity.

<table>
<thead>
<tr>
<th>Mileage</th>
<th>Description</th>
</tr>
</thead>
<tbody>
<tr>
<td>4.3</td>
<td>Rattle River trail ends, follow Kenduskeag trail right (west) towards Mount Moriah.</td>
</tr>
<tr>
<td>4.5</td>
<td>Trail slabs across the south side of Middle Moriah Mountain</td>
</tr>
</tbody>
</table>

**STOP 3. MADRID and RANGELEY** (15 minutes): small outcrops of light gray to grayish green tabular bedded calc-silicates, occasionally including dark green amphiboles. These rocks will persist as float as we move into gray to orange pod bearing migmatite gneiss with grit horizons.

<table>
<thead>
<tr>
<th>Mileage</th>
<th>Description</th>
</tr>
</thead>
<tbody>
<tr>
<td>5.7</td>
<td>Kenduskeag trail ends, follow short side trail to the summit of Mount Moriah (outcrops of Rangeley gneiss) then follow the Carter–Moriah Trail south towards Imp &amp; the Carters.</td>
</tr>
<tr>
<td>6.7</td>
<td>Ledges overlooking the headwaters of Moriah Brook</td>
</tr>
</tbody>
</table>

**STOP 4. RANGELEY** (15 minutes): orange migmatitic gneiss, abundant pods generally aligned parallel to foliation, locally blood red (sulfurous) “pods” with a flaggy schistose foliation, abundant diffuse pegmatites, and possible zones of re-crystallized quartz pebble conglomerate.

<table>
<thead>
<tr>
<th>Mileage</th>
<th>Description</th>
</tr>
</thead>
<tbody>
<tr>
<td>7.1</td>
<td>Stony Brook trail comes in from right (west). Follow the Moriah Brook trail left (east) towards the Wild River.</td>
</tr>
<tr>
<td>9.8</td>
<td>Trail crosses from northeast bank to southwest bank of Moriah Brook</td>
</tr>
</tbody>
</table>
STOP 5. LITTLETON and MADRID within RANGELEY (45 minutes): upstream from where the trail crosses the stream, rhythmically bedded aluminous schist and quartz granofels and laminated green calc-silicates, cut at a low angle by migmatite gneiss.

11.2 Trail crosses from southwest back to northeast bank of Moriah Brook

STOP 6. LITTLETON CONGLOMERATES, MADRID, and SMALLS FALLS above RANGELEY (45 minutes): gray pebble/cobble conglomerates, gray to rusty brown schists, and laminated green-brown calc-silicates with 0.1 to 1 meter-scale chevron and similar folding. Within the gorge are punky weathering rusty sulfidic schists, and pod bearing orange migmatite gneiss, as well as abundant pegmatites.

12.2 Highwater Trail joins from the right (southwest), follow the Moriah Brook trail northeast, to the Wild River.
12.4 Outcrop in Wild River of orange to gray migmatite gneiss with pods of sulfidic rusty schist and possible relict bedding represented by thin granular horizons separated by aluminous horizons.
12.6 Cross bridge over the Wild River. Follow the Wild River Trail northeast 0.3 miles to campground and autos.

REFERENCES CITED


GEOLOGY OF THE BALD MOUNTAIN-SADDLEBACK WIND RANGE, WEST-CENTRAL MAINE

By
Douglas N. Reusch, Division of Natural Sciences, University of Maine at Farmington, Farmington, Maine 04938
Jake Hansen, 437 Harris Street, Hendersonville, NC 28792
e-mail: reusch@maine.edu

INTRODUCTION

Bald Mountain, between Wilton and Weld in west-central Maine (Fig. 1), hosts one of the largest continuous exposures of bedrock in this part of the state. It is a popular hiking destination and field site for undergraduate students. Bald has been a stop on an International Geological Congress (IGC) field trip (Moench, 1989) and two previous New England Intercollegiate Geological Conference (NEIGC) field trips (Reusch and Powers, 2006; Reusch et al., 2010).

Breakthroughs in understanding the geology of the Bald Mountain field site occurred during the fall of 2009, when digital photography was first employed to identify several marker beds with certainty, and during the fall of 2010, when repetitions of marker beds demanding cryptic thrust faults were first recognized (Hansen and Reusch, 2011).

Here, we describe the details of the microstratigraphy and structures within a field site located on the northeast ridge of the mountain, extending several hundred meters along the ridge. This site, which can be framed pedagogically as a microcosm of William Smith’s study area, has a great potential for introducing geology students to the principles of stratigraphy, structural geology, and geologic mapping. The bulk of this paper is based on a field guide for the Geological Society of Maine summer 2012 field trip (Reusch et al., 2012). Detailed (1cm = 10 m [1:1000 scale]) mapping is presented for the northeastern side of Bald. Mapping of the far side of Bald and Saddleback Wind is in progress.

Figure 1. Saddleback Wind (center), Bald Mountain (right), and wind towers (left) viewed from Farmington.

GEOLOGIC SETTING

The Central Maine Trough comprises marine strata of Silurian and Devonian age (Fig. 2, units 27 and 32a) deposited on fragments of Ganderia (unit 18, e.g., Dead River Formation, Grand Pitch Formation, and Miramichi Group), which share a common Gondwanan provenance and Early Ordovician “Penobscot” deformation (Reusch and van Staal, 2012). Post-Penobscot dispersal of Ganderian fragments occurred behind the Popologan-Ammonoosuc arc (unit 13b), which resulted in a wide back arc basin blanketed by Middle-Upper Ordovician pelagic sediments (too thin to separate but included within units 13b and 27). Following collision of the post-Penobscot arc (unit 13b) with Laurentia, subduction flipped and Central Maine (unit 27) occupied a southeast-facing forearc setting during the Silurian. Closure of the Ordovician back arc basin during the Salinic orogeny (Dokken et al., 2014) set the stage for Acadian orogeny as the next peri-Gondwanan element—Avalonia—accreted.

The Bald Mountain-Saddleback Wind range lies within the Rumford outlier (Fig. 3; Reusch et al., 2010), largely composed of marine strata assigned to the Seboomook Group (unit 32a) of presumed Early Devonian age (Osberg et al., 1985; Moench and Pankiwskyj, 1988). These strata are dominantly thinly bedded and fine-grained, suggesting a distal site of deposition on a submarine fan. They contrast with generally lighter-colored, thicker-
bedded and somewhat coarser sedimentary rocks of the surrounding unit 27. All strata of the Central Maine Trough have been deformed into upright folds, and variably metamorphosed up to sillimanite grade near Devonian plutons.

Figure 2. Regional setting of Bald Mountain-Saddleback Wind (Hibbard et al., 2006). The range is centrally located within the Rumford outlier (Unit 32a, mostly west of Farmington). In turn, the Rumford outlier, ~70 km · 35 km (more than 2/3 the area of Rhode Island), lies within the northwestern part of the Central Maine Trough (unit 27). The embryo-shaped pluton is the Phillips Granite. In clear weather, Mt. Washington, 54 miles to the WSW, is easily visible. Rectangular boxes indicate the locations of Figs. 3 and 4.

**GENERAL GEOLOGY**

In brief, a distal turbidite sequence includes several sets of marker beds that display remarkable lateral continuity. These beds are considered Devonian based on correlation with the Littleton Formation and Seboomook Group. They are repeated along cryptic thrusts in a structurally complex zone interpreted as a thrust duplex formed at the Acadian deformation front. Fold geometry of early isoclinal folds and the cryptic nature of the thrusts indicate wet sediment deformation, and mullions imply rotation of the maximum compressive stress from vertical to subhorizontal. Upright folding overlapped with late-stage intrusion of granite and sillimanite-grade metamorphism.

**Stratigraphy**

The Day Mountain Formation of the Seboomook Group (Moench and Pankiwskyj, 1988) crops out extensively in the eastern part of the Rumford outlier (Fig. 3). We have identified several sets of marker beds consisting of distinctive “bar codes” of gray pelite and quartzite. We have adopted informal names (Fig. 5), intended to be useful locally. Details are presented in inferred ascending stratigraphic order. Nearby, the Temple Stream Formation, a rusty-weathering graphitic schist, supposedly underlies the Day Mountain Formation, followed farther below by the Mount Blue Formation, with lithology resembling that of the Day Mountain Formation.

**Southeastern member.** The Southeastern member is well exposed on the cliffs, a dip slope, southeast of the main ridge (Fig. 15). Sandstone beds are upright and dip 35° to the southeast. Several meters of section are exposed between the Southside fault and trees at the bottom of the ridge. Relative age is poorly constrained but this member may be oldest according to its position in the hanging wall of a suspected thrust duplex.
Figure 3. Elements of the Rumford outlier (brick pattern, ribbon limestone; dots, granule conglomerate; gray, Temple Stream Formation; vertical lines widely spaced, Mount Blue Formation; vertical lines narrowly spaced, Day Mountain Formation) and surrounding region (a, Anasagunticook Formation; c, Carrabassett Formation; m, Madrid Formation; s, Sangerville Formation; sf, Smalls Falls Formation; x, plutons). F, Farmington. Bald Mountain is Stop 8 on NEIGC Trip C3 (Reusch et al., 2010). Gray lines north and south of Stop 8 are, respectively, Routes 156 and 2.
Stripes beds. The type locality of the Stripes beds is near the base of the Northeast Ledges (Fig. 15). A triplet of medium thick quartzite beds overlies a single thin bed (Fig. 5C). Also note the very thin beds between the thick beds of the triplet. Around 6 meters of section is present between these marker beds and The Meeting Place beds. The Stripes beds have also been recognized at the Triangle, where they are twice repeated by the 3 a.m. thrust and an unnamed thrust to the northwest.

Meeting Place beds. The Meeting Place beds are defined at the Meeting Place, where they can be followed to the southwest across a branch of the OMG thrust and also to the northeast into the Northeast Ledges. They consist of

Figure 5A. 2xl marker beds exposed on lower Northeast Ledges (left) and near the Triangle (right). See text for description. Brunton compass for scale.

Figure 5B. The Meeting Place marker beds exposed on the Northeast Ledges (left) and at the Meeting Place (right). Natasha Manuel’s knee for scale.

Figure 5C. The Stripes marker beds exposed at bottom of Northeast Ledges (left), southeast side of the Triangle (center) and middle of the Triangle (right). Scale varies.
Figure 4. Stop locations and reconnaissance geology. Stop 10 is off map near base of access road. Caution: for inclined beds, tops are mostly unknown. Stop 9 is the entire access road (~2 miles, elevations 1700’ to 2100’).
four prominent thin beds spaced several tens of centimeters apart underlain by a thickening-upward triplet of very thin beds (Fig. 5B). Approximately 2-3 meters of mainly pelite is present between these beds and the overlying 2xl beds.

**2xl beds.** The 2xl beds (Fig. 5A) are defined in the ledge just above the Meeting Place. They consist of two thick, cross-laminated (Bouma C) sandstone beds underlain by 5 thin laminations, which are in turn underlain by 4 laminations. This member can be traced southwest into the Triangle area. At the Meeting Place, it is repeated across the OMG thrust, and both occurrences can be followed northeastward into the Northeast Ledges. Cross laminations here suggest flow from northeast to southwest (Reusch and Powers, 2006).

**Northwestern member.** The Northwestern member is exposed along the trail below the Meeting Place and to the northwest of the Meeting Place fault. It consists of thinly bedded graded quartzite-pelite couplets, all topping to the southeast. Again, relative age is poorly constrained but this unit may be the youngest according to its position in the footwall of a suspected thrust duplex.

**Age relationships.** The 2xl member overlies, presumably on a conformable contact, the Meeting Place member, which in turn overlies the Stripes member. All of these units may be older than the Northwestern member, interpreted to occupy the footwall of an Older thrust duplex.

**Other units.** The Dark Ledges unit comprises rusty-weathering, very thinly laminated sulfidic, graphitic schist. It crops out to the southeast on the upper half of the near face of Bald Mountain, and is interpreted to be in fault contact with typical gray schist-quartzite to the northwest (Fig. 15). It may correlate with the Temple Stream Formation. The gray rocks that extend from above the Triangle to the summit have not been subdivided, although some quartzite beds are traceable over tens of meters.

**Structural geology**

The lower field site is centered on a structurally complex belt sandwiched between uniformly southeast-dipping, southeast-topping sections. We interpret the complex zone as a duplex. Early cryptic thrusts and rare isoclinal folds formed as wet sediment entered the deformation front, and were later steepened in conjunction with upright F2 folds. Minor granite intrusions (G1 and G2) are both folded by F2 folds (G1) and also truncate the folds (G2).

**OMG thrust.** The OMG thrust was first recognized at the Meeting Place based on repetition of two southeast-topping sections of the 2xl member. Quartz veins may mark its precise location here but to the northeast, the thrust is entirely cryptic within the Northeast Ledges. It truncates the southeasterly belt of 2xl on the lower Northeast Ledges, and truncates the northwesterly belt of 2xl at the Meeting Place, where it offsets the Meeting Place member a few meters in a left-lateral sense.

**3 a.m. thrust.** The 3 a.m. thrust was recognized at 3 a.m. by comparing digital photographs taken from the Triangle area. The Stripes member crops out along the southeastern margin of the Triangle and also in the middle of this ledge. The thrust is entirely cryptic, as with the OMG thrust, which suggests that either the repeated sedimentary sections were still wet or that metamorphic recrystallization has obliterated the fault surface. A third exposure of the Stripes member occurs still farther northwest, hence an additional (unnamed) cryptic thrust is required at the Triangle location.

**Meeting Place fault.** A quartz vein-decorated fault truncates the Northwestern member at the Meeting Place (Fig. 6A). The fault is approximately parallel to the base of the Meeting Place unit. This fault, interpreted as the floor thrust of a duplex, extends northeast into the Northeast Ledges and may be followed uphill to the southwest where it is covered by trees to the northwest of the Triangle. Folded quartz veins in the vicinity of Broken Arm Rock (Fig. 15) suggest that it predates F2 folds.

**Southside fault.** At the top of the Southside Ledges, a steeply-dipping fault separates the Southeastern member from the structurally complex central zone (Fig. 6B). Drag folds in the Southeastern member suggest that the southeast block moved relatively down during subhorizontal extension.
**F1 folds.** Two prominent examples of early isoclinal folds (F1), both synclines, are present in the map area. The upper example is a textbook refolded fold (Fig. 7A). The lower example displays strong asymmetry with the southeastern (northwest-topping) limb strongly attenuated (Fig. 8 of Reusch and Powers, 2006). Both folds occur in proximity to thrust faults. In the upper example, sandstone is thickest in the hinge, indicating the sand was liquefied when the fold formed (Waldron and Gagnon, 2011).

**F2 folds.** Asymmetric upright folds are the most obvious type of structure in the field area. Long southeast-topping limbs alternate with short upright limbs. Fold axes are subhorizontal, plunging gently both to the northeast and southwest. Small-scale F2 folds are common, and map-scale (>10 m) flat limbs are present in the upper part of the Northeast Ledges, just above Broken Arm Rock, and in the Three Temples area. Steeply northwest-dipping schistocity is locally axial planar to the folds but in most places, while the schistocity does not truncate F2 folds, it dips more steeply northwest than the axial planes.

**Mullions.** A textbook example of mullions was found in the Southeastern member within a 30-cm thick sandstone bed (Fig. 7B). The mullions display a wavelength of 5-10 cm. Quartz veinlets penetrate the sandstone bed at the junctions of the mullions, and display northwest vergence.

**Thrust duplex.** Together, the pattern of cryptic thrusts and bedding orientations strongly suggests a thrust duplex located between uniformly southeast-dipping (~35°), upright sections in the Northwestern and Southeastern members, and bounded by the Meeting Place fault, a likely floor thrust, and Southside fault. Bedding within the duplex is subvertical, consistent with a northwest vergent sense of imbrication.
Metamorphism and granites

Metamorphism. Pelitic strata are metamorphosed to muscovite-biotite-garnet-sillimanite-staurolite schist (Fig. 8). Moench and Pankiwskyj (1988) show the sillimanite isograd located just southeast of the Bald summit. Calc-silicate pods are present along the trail between The Wall and summit.

Granites. Medium-grained leucogranites are present in the upper half of the mountain between the lower field site (Triangle) and summit. Twin Dikes (Fig. 15) refers to a pair of parallel meter-wide dikes that extends from southeast to northwest across the ridge. These G2 dikes cut F2 folds. Other smaller granite pods are clearly folded (G1).

Figure 8. Staurolite-garnet schist from northeast side of Bald (left, PPT; right, XPT). Central garnet ~1 mm across.

GEOLOGIC HISTORY

Turbidite deposition

The environment of deposition was clearly marine, bedding and grain size characteristics suggesting a distal location on a submarine fan (Fig. 9A). Sandstones are quartzose, but otherwise little is known about the provenance. While paleocurrents were dominantly from northeast to southwest (Reusch and Powers, 2006), and locally to the southeast (Reusch et al., 2010), other interpretations of the Seboomook Group strongly suggest a southeastern source (e.g., Bradley et al., 2000; Bradley and Hanson, 2002). An explanation for the single example of southerly flow (Fig. 9B) invokes a meandering channel on a very gentle slope. The extraordinary continuity of bedding with no changes over hundreds of meters is quite remarkable and consistent with a distal setting. Cross laminations are interpreted as Bouma C flow regime and a turbidity current mechanism of deposition. It is possible that some very thin, ungraded laminations are contourites.

Cryptic thrusts and isoclinal folds

Cryptic thrusts and the geometry of isoclinal F1 folds (Waldron and Gagnon, 2011) strongly suggest deformation of wet sediment as it was transported through a convergent deformation front. Flow of sand into fold hinges indicates that it was liquefied at this time. The surface slope, initially flat, may have increased as the sediments approached the deformation wedge. While difficult to distinguish between pure gravity and tectonic mechanisms, we prefer rooted deformation, as opposed to slumping, based on the presence of quartz veins along the Meeting Place fault.

Subsequent deformation

The mullions record a two-step history (Kenis et al. 2005) beginning with hydrofracturing of the sandstone bed, resulting in the quartz veinlets, under a vertical maximum compressive stress due to either sedimentary or tectonic loading. Subsequently, the maximum stress rotated to cause bedding-parallel shortening (mullions). Microstructural observations suggest a dislocation mechanism of deformation in the quartz veinlets, and pressure solution mechanisms in very weak sediment (Kenis et al. 2005).
F2 folds indicate continued horizontal shortening, leading to a general steepening of the strata. Schistocity is generally steeper than the F2 axial planes, and is likely to have formed later, mimicking an earlier pressure solution cleavage.

**Figure 9.** A) Graded beds at the Meeting Place. B) Cross-laminated bed in 3 Temples area.

**Metamorphism and plutonism**

The intrusion of granite overlapped F2 folding and outlasted it (Solar et al., 1998; Solar and Brown, 2001). Metamorphism is regionally correlated with the pattern of plutons, and likely enhanced an older tectonic fabric. Staurolites appear to be randomly oriented, suggesting they grew at a quite late stage.

**Note on carbon cycling and glacial geology**

The rock cycle and carbon cycle are coupled. Graphitic schists of the Dark Ledges unit record organic carbon burial, hence the basin was a sink for carbon dioxide during the time when these sediments were deposited (Fig. 10A). Calc-silicate rocks record thermal decomposition of carbonates and reaction with silicates, hence during metamorphism this region was a source of carbon dioxide (Fig. 10B). And, note glaciated surface (Fig. 11).

**Figure 10.** A) Northwestern contact of Dark Ledges unit (sneaker) and close-up of rare unweathered surface of graphitic schist. B) Calc-silicate pod, formerly carbonate concretion.
Figure 11. A) Glacial grooves oriented east-southeast. Ice flow was roughly perpendicular to the ridge. B) Crag-and-tail feature in same orientation doubles as trail marker for entrance to “Secret Path” to the Northeast Ledges.

REGIONAL SIGNIFICANCE

Ganderia, a key peri-Gondwanan element of New England, and the Salinic forearc

The Central Maine Trough comprises marine sedimentary rocks deposited on various fragments of Ganderia, the first of several peri-Gondwanan elements to be accreted to Laurentia (van Staal et al., 2002; Hibbard et al., 2006; Reusch et al., 2006). Ganderia is defined on the basis of Cambro-Ordovician quartz-rich strata (e.g., Miramichi Group, Grand Pitch Formation, Dead River Formation, and evidently the Moretown Formation [MacDonald et al., 2014]) that display Gondwanan provenance (arguably the Amazon craton) and Early Ordovician “Penobscot” deformation (in Newfoundland clearly related to ophiolite obduction). Based on relationships best preserved in the Bathurst area of New Brunswick, subsequent to the Penobscot collision, Ganderia was dispersed in a back arc setting (Tetagouche-Exploits basin of van Staal et al., 2009). After a subduction flip (Karabinos et al., 2017) and Late Ordovician accretion of the Popologan-Ammonoosuc arc to Laurentia, a new southeast-facing continental arc (“Salinic arc,” e.g., Quimby Formation [Bronson Hill] and Attean pluton [Boundary Mountains]) grew on the post-Taconic Laurentian margin (Moench and Aleinikoff, 2002). In this context, the Central Maine region then occupied a forearc setting (Reusch and van Staal, 2012; Dokken et al., 2014).

Figure 12. Location of Ganderia within the Appalachian-Caledonide orogen (Waldron et al., 2014).

Acadian foreland basin

The Seboomook Group is widely interpreted as a clastic wedge deposited in a foreland basin sourced from erosion of a growing Acadian orogen during the Early Devonian (Bradley, 1983; Bradley et al., 2000; Bradley and Tucker, 2002). The basin migrated northwestward in front of the growing orogen, eventually emerging above sea level and culminating with deposition of the Catskill delta. Note that Hibbard et al. (2006) interpret this foreland
basin to be in a retroarc position with respect to the Silurian Coastal arc of coastal Maine, New Brunswick, and southern Newfoundland. In other words, in their scenario, northwest vergence is opposite that related to presumed northwest-dipping subduction of the Avalonian plate beneath Laurentia.

**Acadian orogenesis**

The geology of the Bald Mountain field site is entirely consistent with the tectonic model of Bradley et al. (2000; Fig. 13). It is a snapshot of the processes that characterize Acadian orogenesis at a high structural level. A northwest vergence is consistent with this model of a northwest-migrating orogenic front, but may be at odds with the well-documented southeast-vergent nappes of the Presidential Range in New Hampshire (Eusden and Lyons 1993). Reconciling the structural geology of these classic sites constitutes a first-order problem in northern New England geology.

![Figure 13. Tectonic model (Bradley et al. 2000) showing formation of F1 folds and cryptic thrusts as foreland basin sediments cross the deformation front.](image)

**ACKNOWLEDGMENTS**

The first author appreciates the patience of students in structural geology over the years. He also thanks Scott Johnson, Dyk Eusden, Robert Marvinney, Dwight Bradley, and Lauren Bradley for site visits as well as the participants of NEIGC trips during 2006 and 2010. Thanks to Carrier Timberlands and Rebecca Howard of Saddleback Ridge Wind for allowing access to Bald and Saddleback, respectively. Mike Pakulski contributed to mapping in 2014, and Wyatt McCurdy is thanked for help with Fig. 4.

**ROAD LOG**

This trip is a point-to-point hike from the Bald Mountain trailhead on Rte. 156 to the Saddleback Ridge wind farm in Carthage.

**MEETING POINT.** Saturday September 30th, 8:00 AM in the gravel parking area behind the Saddleback Ridge Wind office building (390373.00 m E, 4938625.00 m N). Driving east on Route 2, around 50 minutes from Bethel and near the crest of a long hill, turn left on Winter Hill Road (easily missed, look for signs “Winter Hill Antiques” and “Rocky Mountain Terrain Park”). Proceed 0.9 miles west to the office site. Bathroom facilities will be available. Following introductory remarks, two UMF vans will deliver 20 participants to the Bald Mountain trailhead on Route 156 (http://www.mainetrailfinder.com/trails/trail/bald-mountain-and-saddleback-wind-trail 393477 m E, 4945902 m N), a 20-minute drive. (Vans will return to the starting point, continue to the end of the wind farm access road, and drivers then hike northeast to re-join the group.) Note that while the UMF vans can accommodate 20, others are welcome to participate should they be willing to make similar travel arrangements (preferably in advance). Bring lunch, lots of water, and appropriate clothing for a long, all-day hike with significant portions exposed to the elements. Trail conditions vary considerably. The initial steep ascent has suffered from erosion. The top half of Bald is ledge that under some conditions can be dangerously slippery. If weather is atrocious, the contingency plan is to repeat our 2010 NEIGC trip.

**Mileage.**

0.0 Turn L out of parking lot and proceed E on Winter Hill Road.
**STOP 1. MEETING PLACE, NORTHEAST LEDGES, AND SOUTHIDE.** (393446.00 m E, 4945002.00 m N, 1 HOUR)

We will examine the type localities of the Meeting Place beds, the 2xl beds, and Meeting Place fault. Participants will be set loose to find the “Rosetta Stone” repetition of the 2xl beds. We will then proceed to the Northeast Ledges to inspect extensions of the geology displayed at the Meeting Place, the type locality of the Stripes, and a spectacular F1 isocline. At Southside, we will point out the Southside fault and discuss remotely the mullions that are present in a dangerous location half way down the cliff.

![Figure 14. Google Earth image of Meeting Place ledges taken on mid-afternoon of 2013/9/17. Red line is 10 m and oriented north-south. At north end (right) of red line, UMF students and first author (circle of white spots) are examining the type locality of the 2xl beds. The chances of being captured on Google Earth were high because we visit this site frequently. The “Rosetta Stone” stone, where repeated 2xl beds were first recognized, is located approximately just right of center.](image)

**STOP 2. THE TRIANGLE.** (393303.00 m E, 4944893.00 m N, 15 MINUTES)

The Stripes beds are repeated here. Both sets of beds top to the southeast. This is the type location for the 3 a.m. thrust. A third repetition was subsequently recognized, hence there must be at least two cryptic thrusts here. A short distance to the southeast of the Triangle ledge may be found an excellent example of the 2xl beds. Fine details of the bar code are the same as in outcrops hundreds of meters to the northeast. Presumably, the Meeting Place beds are present beneath the intervening vegetated area.

**STOP 3. TWIN DIKES AND DARK LEDGES.** (393175.00 m E, 4944782.00 m N, 15 MINUTES)

Two parallel granite dikes strike northwest and cross cut the F2 folds. Elsewhere, identical granite bodies are significantly deformed, which suggests that emplacement of magma was contemporaneous with F2 deformation. The Dark Ledges unit crops out to the southeast, where good examples of rusty weathering, thinly laminated graphitic schist may be observed. The northwestern contact of the Dark Ledges unit cuts bedding and F2 folds, hence is interpreted to be a fault (Dark Ledges fault on Fig. 15).

**STOP 4. THREE TEMPLES.** (393115.00 m E, 4944736.00 m N, 15 MINUTES)

The site name refers to a triplet of roche moutonées. This area coincides with a large flat limb of an F2 fold. Look for a small parasitic fold with uncommon S-asymmetry to confirm this. On the face of a small cliff (middle roche moutonée), cross laminated quartzite indicates transport from north to south. This current direction is anomalous, in that at face value it suggests a northern provenance. However, it might be reconciled with the prevailing view of a southeastern provenance for Devonian sediments if viewed as a meandering channel on a very low-gradient part of a submarine fan). Alternatively, these strata potentially might correlate with the Perry Mountain
Figure 15. Precisely located geologic data, northeastern side of Bald Mountain (see text for explanation).
Formation that has a well-established northwestern provenance. Moench (written communication, 2006) was able to distinguish Perry Mountain and Seboomook strata on the basis of “fast” versus “slow” grades. Perry Mountain sediment was mature, and the transition from sand to clay is relatively sharp (“fast grade”). Seboomook sediment was mud-rich, sedimentation rate high, and the transition from sand to clay less sharp (“slow grade”).

STOP 5. BALD MOUNTAIN SUMMIT. (392948.00 m E, 4944662.00 m N, 30 MINUTES)
On the approach to the summit, the path traverses a small body of granite and the northwestern contact of the Dark Ledges unit. On a clear day, one can see Saddleback (east of Rangeley, 340°, 22 miles distant), the Presidential Range (Mt. Washington, 242°, 54 miles distant), and the Camden Hills (Mt. Megunticook, 124°, 69 miles distant). The view spans most of Ganderia (Fig. 12), and this is a great location to discuss the regional tectonics. On a more local scale, in the foreground to the northwest, the Phillips Granite is centered on the valley of Webb Lake; rocks nearly identical to Bald form the Tumbledown massif on the far side of the lake. Prevailing geomorphic wisdom attributes differential erosion to the contrast between the granite and resistant metamorphic country rock around it. On a very local scale, why is the ridge where it is? Is it a geomorphological accident, or is there something about the bedrock that makes it preferentially resistant to erosion?

STOP 6. STAUROLITE CITY. (392632.00 m E, 4944078.00 m N, 30 MINUTES)
A spectacular set of ledges is bounded to the northeast by a sheer cliff (Fig. 16A). The northwest-southeast cross section nicely captures the geometry of F2 folds (long southeast-topping limbs and short upright limbs). Staurolites abound in the predominantly pelitic schists, which commonly are very thinly laminated (Fig. 16B).

Figure 16. A) Cliff face at Stop 6 displays representative F2 folds. B) Very thinly laminated pelitic schist.

STOP 7. SADDLEBACK WIND SUMMIT AREA. (392280.00 m E, 4942495.00 m N, 1 HOUR)
Uncommon horizons of medium-bedded quartzites display asymmetric fold geometries (Fig. 17A). As on Bald, most beds top to the southeast in this area. The thinned northwest-topping beds are consistent with an overall top-to-northwest vergence (e.g., Fig. 13). Likely, we will have much more to communicate about these outcrops pending the results of an undergraduate mapping project in September.

STOP 8. EASTERN SPUR. (392982.00 m E, 4943062.00 m N, 1 HOUR includes round trip hike on ancient trail)
A set of quartz veins decorates a fault duplex here (Fig. 17B). Note truncation of beds on the right, and right-lateral sense of quartz veins in the duplex.

STOP 9. WIND FARM ACCESS ROAD. (Tower 12: 391918.00 m E, 4942035.00 m N, Tower 3: 391161.00 m E, 4940244.00 m N, 90 MINUTES)
Moench and Pankiwskyj (1988) show the Bald Mountain detachment fault passing through this ridge, but reconnaissance mapping has not found evidence for a major discontinuity. Some beds top to the northwest (Fig. 18A). Rusty-weathering, very thinly bedded to laminated pelitic schist is common (Fig. 18B).

STOP 10. PEGMATITE. (390155.00 m E, 4939314.00 m N, 10 MINUTES)
No field trip in west-central Maine would be complete without a pegmatite. Near the base of the access road, new road cuts expose a granitic pegmatite, presumably an offshoot of the nearby Phillips Granite with an age of ca. 403 Ma (Solar et al., 1998).
Figure 17. A) Thinned NW-topping limbs, Jesse Powers ~2 m tall. B) 1m-wide right-lateral duplex, looking NE.

Figure 18. A) Northwest (left)-topping quartzite bed in road cut on main access road below Tower 8. B) Thinly laminated schist near Tower 4.

REFERENCES CITED


GLACIAL DEPOSITS AND LATE-GLACIAL TO POSTGLACIAL ALLUVIAL FANS IN THE NORTHWESTERN WHITE MOUNTAINS, NEW HAMPSHIRE

By

Woodrow B. Thompson¹, Maine Geological Survey (retired), 93 State House Station, Augusta, ME 04333
Gregory Barker², New Hampshire Geological Survey, P.O. Box 95, 29 Hazen Dr., Concord, NH 03302-0095
E-mail addresses: ¹iceagemaine@myfairpoint.net, ²gbarker@des.state.nh.us

INTRODUCTION

This trip visits the northwestern White Mountains, where a wide variety of glacial and glacial-lake deposits formed during recession of the Laurentide Ice Sheet (Thompson et al., 2017). We will examine key sites that turned up during recent field studies and see examples of new LiDAR imagery that is being used to refine the mapping of glacial features in the area. The results of detailed surficial quadrangle mapping for the New Hampshire Geological Survey’s STATEMAP program have likewise contributed much to our understanding of the area’s glacial history (e.g. Hildreth, 2009; Fowler and Barker, 2015; Thompson, 2016).

The trip ranges from Randolph west to Bethlehem, and north to Jefferson and Lancaster (Figs. 1, 2). Topics of discussion include ice-marginal deposits of the White Mountain Moraine System, their age and climatic significance, glacial lake sequences and drainage channels, alluvial fans, and a saprolite occurrence.

Note that all gravel pit stops are on private property, and permission must be obtained from the owners for any future visits.

Figure 1. Map of the northern White Mountains showing locations of Trip B2 stops 1–7. Modified from Thompson et al. (2017). Long solid gray line indicates limit of Littleton-Bethlehem (L-B) readvance and related moraines of the White Mountain Moraine System (WMMS). Dashed line shows correlation with the Berlin moraines. Open circles mark ponds cored by Christopher Dorion. Black squares indicate sample locations for cosmogenic-nuclide exposure dating by Greg Balco: Sleeping Astronomer Moraine (SAM) and Beech Hill moraines (BHM). Rectangle shows area of geologic map (Fig. 2).
There are three principal rivers in the field trip area. The Ammonoosuc River originates on the side of Mount Washington in the Presidential Range. It flows west to Littleton and then southwest to the Connecticut River. Neighboring rivers to the north – the Johns River and Israel River – flow northwest and join the Connecticut River at Dalton and Lancaster respectively. The valleys of all these rivers hosted ice-dammed glacial lakes during recession of the Laurentide Ice Sheet. They also were affected by a major late-glacial readvance of the ice sheet, during which an extensive series of moraines were deposited across northern New Hampshire. The stops on our trip were chosen to present stratigraphic and geomorphic evidence of these glacial events, as well as the modification of the landscape by postglacial streams.

**Figure 2.** Geologic map of the field trip area showing moraine clusters (green lines) deposited during the L-B readvance and recession from the readvance maximum, associated glacial lakes (blue), lake spillways (blue arrows), recessional ice-margin positions (purple lines), and area covered by LiDAR image in Fig. 5 of Thompson *et al.* (2017). Lake stages shown here include: Crawford (Cr), Gale River 2 (G2), and Bethlehem 2 (B2) stages of Lake Ammonoosuc; Lake Carroll stages C1, C2; and the Bowman (Bo), Pine Knob (PK), and Baileys (Ba) stages of Lake Israel. Black dots mark coring sites and approximate basal radiocarbon ages in cal ka BP for Pond of Safety (POS), Carroll spillway (CS), Cherry Pond (CP), Martin Meadow Pond (MMP), and York Pond (YP). Black triangles are locations of the Carroll delta (C) and Lake Crescent delta (LCD). Unlabeled glacial lakes include small parts of Lake Franconia (southwest corner of map) and two arms of Lake Hitchcock on the western border. Topographic base map contour interval is 20 m. Modified from Thompson *et al.* (2017).
PREVIOUS WORK

Many geologists have investigated the glacial history of the White Mountains, starting in the 1800s. The following account summarizes and updates reviews by Thompson (1999) and Thompson et al. (2009a). Visits by Louis Agassiz, Edward Hitchcock, and Charles Lyell stimulated interest in glaciation of the region. Early investigations in the present field trip area focused on clusters of drift ridges and hummocks in the Ammonoosuc River valley that eventually became known as the Bethlehem Moraine. Agassiz published the first observations on these deposits following his 1847 visit to the White Mountains. He thought that the morphology and boulder provenance of moraines in the Bethlehem area proved they were deposited by a local glacier flowing north from the vicinity of Mount Lafayette in Franconia.

Charles Hitchcock (1878a) agreed with Agassiz's theory of local ice depositing moraines from the south. Upham likewise concurred with this theory and formally named the Bethlehem Moraine (Upham, 1904, p. 12). His description of the Bethlehem Moraine is very similar to what recent workers have observed: "The material of this belt is chiefly till, with some modified drift, as kames, or knolls of gravel and sand. The contour is very irregular, in multitudes of hillocks and little ridges, grouped without order or much parallelism of their trends. Everywhere in and upon these deposits boulders abound, ...being far more plentiful than in and on the adjoining smoother tracts of till throughout this region" (Upham, 1904, p. 11–12).

James Goldthwait (1916) reinterpreted the Bethlehem Moraine. Agassiz's model was found to be flawed because it proposed a topographically unrealistic ice-flow path, lacked adequate documentation of northward erratic transport, and was not supported by striation evidence. Goldthwait pointed to the lack of recessional moraines in the Franconia Range as another problem with Agassiz's and Upham's local-ice models. He said that the geometry and provenance of the Bethlehem Moraine favor deposition from the north by the continental ice sheet. Goldthwait’s 1916 paper included the first map of the Bethlehem Moraine.

Ernst Antevs (1922) inferred from his Connecticut Valley varve records that a glacial readvance occurred west of Littleton where Comerford Dam is now located (Fig. 1). Crosby (1934) reached the same conclusion, supported by the two-till stratigraphy that he found at the dam site. Antevs and Crosby equated this readvance with that which deposited the Bethlehem Moraine complex. Louisee (1935) referred to this event as the "readvance at Littleton", and Thompson et al. (1999) named it the "Littleton-Bethlehem readvance" to stress the connection with the deposits historically known as the Bethlehem Moraine. Richard Lougee (1935) described a new section next to Comerford Dam showing deformed varves between two till units. He correlated the Comerford varves with Antevs’ nearby sections and inferred from a gap in the varve sequence that the readvance covered the site for 151 years.

In 1930 Lougee assisted J. W. Goldthwait in a gravel inventory funded by the New Hampshire Highway Department. He was assigned to map several 15-minute quadrangles in the White Mountains. Lougee (n.d.) prepared a manuscript that included a wealth of new information on glacial deposits, ice-dammed lakes, and meltwater drainage routes in the region. A copy of this paper resides in the Special Collections of the Dartmouth College Library. It is unfortunate that it was never published, since it contains the first analysis of the stages of glacial Lake Ammonoosuc and corresponding spillways.

Following a research hiatus of several decades, the 1996 NEIGC trip to the study area presented results of renewed work on deglaciation of the northwestern White Mountains (Thompson et al., 1996). In the guidebook for that trip, Christopher Dorion included his analysis of sediment cores from Surplus Pond in western Maine and Pond of Safety in Randolph, New Hampshire, showing clear stratigraphic evidence of Younger Dryas cooling. Jack Ridge and his students led a companion trip presenting new results from the glacial Lake Hitchcock varve sequence in the vicinity of Comerford Dam (Ridge et al., 1996). During this same trip, Thompson described a stream bluff exposure near the New Hampshire end of the dam that showed further stratigraphic evidence of glacial readvance.

Thompson et al. (1999) discovered moraine clusters in Carroll and Randolph, which correlate with the Bethlehem moraine complex, and subsequently found another moraine series of similar age in the Berlin area (Fig. 2). The overall moraine belt spans northern New Hampshire and is now called the White Mountain Moraine System (Thompson et al., 2009a, 2017). Dorion’s radiocarbon ages from basal pond sediments in the region suggested an age of 12,000 radiocarbon years (~14 cal ka) for the Littleton-Bethlehem (L-B) readvance and associated moraines, coinciding with the brief interval of cold climate called the Older Dryas Chronozone. Ridge pinpointed the
readvance age much more precisely and confirmed the Older Dryas connection through his work on the New England Varve Chronology and its relation to the Comerford Dam readvance site. His varve correlations in the Connecticut River valley west of Littleton showed that the readvance occurred at 11,900–11,800 14C ka BP (13.9–13.8 cal ka BP) (Ridge et al., 1999, 2004). The Lake Hitchcock varve chronology is now incorporated in the North American Varve Chronology of Ridge et al. (2012). Ridge’s latest refinement of the L-B readvance age places it at ~14.0–13.8 cal ka BP, and has established a firm connection to the Older Dryas event (GI-1d) in the Greenland ice core record (Thompson et al., 2017). Greg Balco used this readvance age to help calibrate cosmogenic-nuclide production rates for exposure dating in New England. He discussed this work and its application to the White Mountain moraines during the 2009 NEIGC (Thompson et al., 2009a) and in the paper by Thompson et al., 2017. Balco’s study sites are indicated on Figure 1.

MORAINES AND ICE-DAMMED GLACIAL LAKES

Moraines

The following information is from Thompson et al. (2017). Most of the moraines in the field trip area are concentrated in three clusters that constitute large parts of the White Mountain Moraine System (WMMS). From west to east, these are the Bethlehem moraine complex, the Beech Hill moraines, and the Randolph moraines (Fig. 2). (The Berlin moraines are also briefly described here for sake of completeness.) The moraines in all these areas are generally similar to one another. They are composed predominantly of loose sandy till with abundant stones including many granitic boulders derived from local plutons. The moraine ridges typically are 3–30 m high and rarely as much as 50 m. Individual segments are up to 1300 m long and most are sharp crested. The spacing between moraines varies, having depended on ice retreat rate and sediment supply. In the tight cluster of the Beech Hill moraines, it ranges from 30 m to about 200 m.

Bethlehem moraine complex. The Bethlehem moraines (Fig. 2) were deposited during ice recession from the upper Ammonoosuc valley. This is the most diffuse group of moraines in the WMMS, spanning up to 7 km of northward to northwestward ice-margin retreat. A few outlying moraines of the Bethlehem complex occur in the western part of Littleton, 12 km east of the Comerford Dam readvance site. Exposures and well logs indicate the Bethlehem moraines consist chiefly of till. Thompson et al. (1999) described a moraine cross-section in Littleton village where shear structures indicated ice shove from the north. The high (30–40 m) sharp-crested moraines just southeast of Littleton were deposited in terrestrial settings at the terminus of the moraine belt, whereas some of the other moraines in the Bethlehem moraine complex formed where the ice margin stood in glacial Lake Ammonoosuc.

New information on the stratigraphy of the Bethlehem moraines has come from test borings by Sanborn, Head, & Associates (2014) at the regional landfill on Trudeau Road in Bethlehem, about 30 km east of Comerford Dam (Fig. 1). The landfill is situated on one of the most distal moraines in this part of the complex. Three of the deepest borings at the Trudeau Road site are located in the proximal part of the moraine, along an E-W line over a distance of ~340 m. The western boring (B-916D) penetrated 31 m of surficial sediments overlying bedrock. It encountered 13 m of readvance till overlying 10 m of sand, silt, and minor gravel. The latter unit in turn overlies 8 m of till. The same units occur to the east, where boring B-918D ended at ~38.5 m without reaching bedrock (3.7 m till / 29.3 m silt-sand / 5.5 m till). Still farther east, boring B-919D encountered 4.5 m of till / 42.1 m of silt-sand / 3.0 m of till / 2.2 m of silt-sand.

Two sand and gravel pits near the Trudeau Road landfill show readvance till overlying water-laid glacial sediments. The section in one pit exposes a moraine consisting of till overlying coarse gravel with sand lenses. We will visit this site at Stop 4. The other section is 1.25 km southeast of the first, in the distal flank of the moraine on which the landfill is located. It shows silty-sandy diamict with small rounded stones overlying well-stratified glaciofluvial sand and fine gravel. The diamict is interpreted as till derived from recycling of glaciolacustrine sediments during local ice readvance. The landfill borings and nearby pit exposures collectively show that at least some of the distal Bethlehem moraines are not composed solely of till. The upper (readvance) till at all of these Bethlehem sites is thought to be equivalent in age to the upper till at Comerford Dam.

Beech Hill moraines. Thompson et al. (1999) discovered a cluster of end moraines just north of Beech Hill in the town of Carroll (Fig. 2). These moraines are located about 7 km east of the Bethlehem moraine complex. The Beech Hill moraine cluster is at least 0.8 km wide from its distal to proximal margin. New LiDAR imagery shows
additional moraine ridges and hummocks in the wooded area just to the north, which are currently being field-
checked. Elevations of the well-defined moraines are 396–427 m. They comprise ten till ridges that are 4–12 m
high, up to 700 m long, and trend ENE-WSW. Granite boulders up to 3 m wide are abundant on the surfaces of the
moraines. Considering the location of the Beech Hill moraines and their relationships to local glacial-lake stages, it
is reasonably certain that they correlate with the northern part of the Bethlehem moraine complex.

**Randolph moraines.** Another cluster of moraines, which we correlate with the Bethlehem and Beech Hill
moraines, occurs to the northeast in the Israel River valley in Randolph (Fig. 2). The Randolph moraines span about
3.4 km in the direction of downvalley ice retreat (toward the west-northwest). The largest and most distal of these
moraines are located on the drainage divide at the head of the Israel valley, in the Bowman area of Randolph, where
their crests reach elevations of about 540 m (see Stop 1). A major outlying moraine near the Pond of Safety coring
site in the hills of northern Randolph (Figs. 1, 2) rises to 720 m. Its location relative to the moraines at the head of
the Israel valley suggests the moraines in both places were deposited by the same ice tongue and perhaps at the same
time. The difference in their elevations may reflect a steep ice-surface profile near the glacier margin during the L-
B readvance. These distal moraines are up to 50 m high and associated with proglacial channels that fed meltwater
into neighboring river basins to the north and east.

Exposures in Randolph have recorded a glacial readvance of at least several hundred meters in the upper Israel
valley. The Corrigan pit, located on the south side of Valley Road and 0.8 km east of the Jefferson-Randolph town
line (Fig. 1), shows glaciolacustrine deltaic sand and gravel overlain by 6 m of stony glacial till comprising a
moraine ridge. Recumbent folds and thrust faults in the upper part of the lacustrine unit indicate an ice readvance
from the west, concurrent with deposition of the moraine.

**Berlin moraines.** A series of moraines near the town of Berlin in the northeastern White Mountains have been
mapped with the aid of LiDAR imagery (Fig. 2; Thompson et al., 2009b; Thompson and Svendsen, 2015). The
most clearly defined part of the moraine cluster is 3.5 km wide. Moraine crest elevations range from up to 492 m
south of Jericho Lake to about 415 m in the Upper Ammonoosuc River valley. The Berlin moraines are low but
very distinct till ridges that mostly trend northwest-southeast. They were deposited by ice retreating northeast from
the headward part of the Upper Ammonoosuc basin. The moraines are 3–10 m high and strewn with large granitic
boulders.

Hildreth’s mapping (2009) showed a prominent moraine just west of the Pond of Safety coring site in Randolph
(Fig. 2), on the divide between the Israel River and Upper Ammonoosuc River basins. As noted above, this deposit
is an outlying member of the Randolph moraines. LiDAR imagery reveals multiple channels that drained meltwater
north from the Pond of Safety moraine and supplied sediment to the glacial Lake Crescent delta described below.
The position and elevation of this delta require that the Upper Ammonoosuc valley was dammed by the ice margin
that deposited the Berlin moraines, thus showing that the Berlin moraines are coeval with the Randolph moraines
and likewise part of the WMMS (Fig. 1).

**Glacial lakes**

**Glacial Lake Ammonoosuc.** Warren Upham’s work with the Hitchcock survey led him to propose that a lake
had existed in the Fabyan area in the upper part of the Ammonoosuc River valley (Upham, 1878). Goldthwait
(1916) named this water body “Lake Ammonoosuc“. It resulted from damming of the west-draining valley by a
tongue of late Wisconsinan ice receding from the Bethlehem area. As the ice margin withdrew, successively lower
spillways for the lake were uncovered and the lake level fell.

Goldthwait (1916) identified two levels of glacial Lake Ammonoosuc: a higher level into which a "pitted
outwash plain" and other ice-contact deposits at Carroll and Twin Mountain were built, followed by a lower level
into which the Bethlehem moraine complex was deposited. Lougee (n.d.) realized that an earlier and higher stage of
Lake Ammonoosuc (his "Crawford Stage") drained east through Crawford Notch at an elevation of approximately
573 m (1880 ft). This was the same route followed by a subglacial tunnel drainage which formed the esker in the
upper Ammonoosuc Valley (Goldthwait and Mickelson, 1982).

Thompson et al. (1999) named the post-Crawford stages of Lake Ammonoosuc. However, it was Lougee's
undated manuscript (ca. 1930) that first identified some of these lake levels and their outlets. After the Crawford

Stage, glacial Lake Ammonoosuc drained southwestward through five progressively lower spillways (G1-G5 in Figure 3) into the Gale River valley. The spillway for the Gale River 2 Stage is a prominent channel that can be seen along U. S. Route 3 southwest of Twin Mountain village. Later spillways north of Bethlehem village drained the Bethlehem and Wing Road Stages of Lake Ammonoosuc into Indian Brook and later directly into the Ammonoosuc River. The Gale River and younger stages of the lake generally were not deep, so the ice margin probably was grounded on the lake floor. The widest and deepest stage may have been Gale River 2. A well (CFW 53) just west of Twin Mountain encountered a contact between thick glaciolacustrine clay and the underlying till at an elevation of 401 m (Flanagan, 1996). Comparison with the nearby G2 spillway elevation of 445 m (Fig. 3) indicates an initial water depth of at least 44 m.

Figure 3 also shows proposed recessional positions of the late Wisconsinan ice margin that were contemporaneous with the Gale River and later stages of Lake Ammonoosuc. Thompson et al. (1999) inferred these ice margins from the orientation of segments of the Bethlehem moraine complex, together with ice blockages of the valley that would have been required to hold the lake at elevations corresponding to known deltas and spillways. The agreement between elevations of Lake Ammonoosuc deltas and terraces in Twin Mountain, and the matching spillways of the Gale River 2-4 Stages, suggests that the receding glacier margin was a tight dam for the lowering lake. During the evolution of Lake Ammonoosuc, water carried sediment into the lake not only from the melting glacier but also from the early Ammonoosuc River and smaller streams draining the surrounding mountains, as shown by Lougee (1940). Both the lake and the mouth of the river shifted westward as the lake level dropped, and Lake Ammonoosuc ultimately disappeared when the ice margin receded from the Alderbrook area in northernmost Bethlehem.

Figure 3. Part of the Whitefield 15-minute quadrangle, showing inferred ice-margin positions (gray lines) and meltwater spillway channels (arrows). “B”: Beech Hill moraines. “C”: Carroll Delta (Stop 3). Labeled arrows (G1 etc.) show spillways for the Gale River stages of glacial Lake Ammonoosuc listed below. Spillway elevations are based on contours from the newer Bethlehem 1:25,000 metric quadrangle. From Thompson et al. (1999).
Lake stage: Elevation:

<table>
<thead>
<tr>
<th>Lake River</th>
<th>G</th>
<th>Elevation</th>
</tr>
</thead>
<tbody>
<tr>
<td>Gale River 1</td>
<td>G1</td>
<td>477 m (1565 ft)</td>
</tr>
<tr>
<td>Gale River 2</td>
<td>G2</td>
<td>445 m (1460 ft)</td>
</tr>
<tr>
<td>Gale River 3</td>
<td>G3</td>
<td>435 m (1427 ft)</td>
</tr>
<tr>
<td>Gale River 4</td>
<td>G4</td>
<td>423 m (1387 ft)</td>
</tr>
<tr>
<td>Gale River 5</td>
<td>G5</td>
<td>405 m (1328 ft)</td>
</tr>
</tbody>
</table>

Lakes Carroll, Israel, and Whitefield. Three closely related glacial lakes developed as the ice receded northward from the Ammonoosuc valley. Lake Carroll formed north of Twin Mountain village, just south of the Beech Hill moraines. The first stage of this lake (C1, Fig. 2) was very small and briefly drained south through the 429-m channel that incises the Carroll delta (Stop 3). Lake Carroll then dropped to the C2 stage as it drained west along the ice margin.

At the same time that glacial Lake Carroll existed, deglaciation of the neighboring Israel valley to the east resulted in the first stage of ice-dammed Lake Israel. This stage (Bo, Fig. 2) drained east across the divide at Bowman and into the Moose River valley. Deltaic and subaqueous fan deposits were built into the Bowman stage lake. With further ice retreat, lake waters in the Israel valley may have briefly expanded west as the Pine Knob (PK) stage and drained through a 387-m col into the Ammonoosuc valley, as shown in Figure 2. Slight additional recession led to the creation of two separate lakes: the Baileys stage of Lake Israel to the east and Lake Whitefield to the west. The Baileys stage (Ba) spilled west across a low divide at Cherry Pond in Jefferson and thence into Lake Whitefield. Lake Israel terminated when ice retreat to Lancaster caused it to merge with glacial Lake Coos in the Connecticut River valley. Lake Coos (Fig. 2) was a large drift-dammed lake just north of glacial Lake Hitchcock. During ice retreat from the study area, it extended from Dalton, New Hampshire, up the Connecticut valley to North Stratford (Thompson et al., 2011). Lake Whitefield occupied the Johns River valley as the ice margin receded north from Whitefield village. This lake initially drained south into the Ammonoosuc River, and with further ice recession likewise dropped and merged with Lake Coos.

Lake Crescent. Recessional ice margin positions indicated by the Berlin moraines suggest that an ice-dammed lake existed in the Upper Ammonoosuc River valley, in the northeast part of the study area (Figs. 1, 2) and distinct from the Ammonoosuc River basin discussed earlier. The headwaters of this river occupy a mountain-rimmed basin that drains to the north and would have been blocked by the NE-retreating ice lobe that deposited the moraines. A series of nine meltwater channels drained the basin along its eastern border. These channels are lower to the north, recording a succession of probable lake spillways that opened as the ice receded (Thompson and Svendsen, 2015). The dense forest cover and poor exposure of Pleistocene deposits in the Upper Ammonoosuc valley hindered recognition of lacustrine deposits when the area was mapped by Hildreth (2009). However, LiDAR imagery shows a delta complex at the head of the basin that was built by glacial meltwater streams flowing northward into the former lake. Elevations of terraces on the delta top correspond with some of the spillway channels mentioned above, confirming the lowering of the lake level with time. Based on this evidence, we have identified an ice-dammed lake called glacial Lake Crescent after the neighboring Crescent Range (Fig. 2).

ICE-FLOW DIRECTIONS

There are virtually no bedrock exposures within the Bethlehem moraine complex, but striated outcrops in surrounding areas of Bethlehem and Littleton indicate ice-flow directions mostly in the range of 170–190° (Hitchcock, 1878b, 1905; J. W. Goldthwait, 1916; R. P. Goldthwait et al., 1951). J. W. Goldthwait’s map shows a few representative examples of the southward flow. Thompson et al. (1999) recorded striaion trends in this area ranging from 125° to 203°. In the few places where the latter authors could determine relative ages of multiple striaion sets, the striaions trending S to SSW are usually youngest.

Just north of Bethlehem, Thompson et al. (1999) found striaions trending 174-185° on the north side of the Ammonoosuc River, and 178-179° on the crest of Dalton Mountain. These data show that late-glacial southward ice flow crossed both the high hills and valley floor north of the Bethlehem moraine complex. Striaions between Littleton and the Connecticut River likewise indicate generally southward ice flow. Most striaions seen in the latter area trend between 170° and 190°, though a few are more southeasterly. However, south-trending striaions also
occur in many places farther south in the Connecticut River basin (Goldthwait et al., 1951), so they are not associated exclusively with the Bethlehem moraine complex.

At 10 localities in the Whitefield-Lancaster-Jefferson area, Thompson et al. (2009a) recorded striation sets trending SE to SSE (135–172°). One of these sites clearly shows three successive ice-flow directions. It is located on the NW end of a long road cut on the SW side of U. S. Route 2, 2.6 km SE of Lancaster. This ledge has a protected lee surface with the oldest striation set trending 241° and a younger 165° set. The youngest of the three sets trends 155° on the stoss surface of the outcrop. Recent mapping of the Jefferson quadrangle documented striation sets indicating SE ice flow at multiple sites on the mountainside NE of Jefferson village, and ESE ice flow on slopes SW of Pliny Mountain (Thompson, 2016).

ACKNOWLEDGEMENTS

We are especially grateful to the following persons and companies who granted permission to visit the stops on this field trip and enabled us to carry out research visits on other occasions: Jim Brianas, John Scarinza (Randolph Town Forest), and Judith Asetta (Stop 1); Tim Bradstreet of Pike Industries, Inc. (Stop 3); Mark Champagne of Bethlehem Earth Materials (Stop 4); Nigel Manley of the Society for Protection of New Hampshire Forests in Bethlehem (Stop 5); Craig Kluckie (Stop 6); and Paul Crane (Stop 7). Many other residents of the White Mountains have kindly assisted us over the years, especially Dennis Field of Lancaster, whose knowledge of the North Woods has been most helpful.

Surficial geologic mapping funded by the New Hampshire Geological Survey – USGS STATEMAP cooperative enabled us to gather much of the new data reported here. The authors are indebted to Brian Fowler for sharing his knowledge of the White Mountains and regional geology during many discussions in the field. We also thank Kristen Svendsen of the New Hampshire Department of Environmental Services for digitizing the maps from which Figures 1 and 2 were extracted. Al Falster at the Maine Mineral and Gem Museum conducted the sediment analysis needed to help explain the clay occurrence at Stop 7-A. Recent bedrock mapping of the Jefferson-Randolph area by Dyk Eusden and his students (Bates College) helped us to identify rock types in the study area. Douglas Rankin (U. S. Geological Survey) likewise assisted W. B. Thompson with understanding the bedrock geology of the Littleton area and sources of glacially transported rocks.

ROAD LOG

Meeting point: Assemble at 8:00 am on September 30 at the Randolph Fire Station, 0.5 mile from U. S. Route 2 via the Pinkham B Road (aka Dolly Copp Rd.) (319571 m E; 4915340 m N). The turnoff onto this road is about 28 miles / 40 minutes west of Bethel, ME; and from Gorham, NH, it is 4.5 miles / 7 minutes west of where you turn at the traffic light on the west end of downtown. The trip will start promptly at 8:30 am. Carpooling at the fire station is recommended, and moderate to high clearance vehicles are best for the woods roads. Come prepared with your lunch, water, and clothing suitable for predicted weather. We’ll be walking through the woods at Stops 1 and 2 (short to moderate hikes including rocky ground and uneven terrain); all other stops will be accessed by car. The trip will end in late afternoon at a point east of Lancaster, from which participants can drive back to Route 2 in Jefferson and east to NEIGC headquarters in Bethel. Topographic map coverage of the field trip area is provided by the Bethlehem, Jefferson, Mount Dartmouth, and Mount Washington 7.5-minute quadrangles, and the Mount Washington 1:100,000 map. The latter map gives an excellent overview of the White Mountain region.

All stops are on private property, and permission must be obtained from the owners for any future visits!

Mileage Note that cumulative mileages given here may differ from those shown on your odometer, due to variations in driving around pits and woods roads, but the distances between stops are generally accurate.

0.0 Exit parking lot and turn L on Pinkham B Road.
0.5 Turn L on U. S. Route 2.
3.4 Just beyond Lowe’s Store, turn R on short connector to Durand Road.
3.45 Turn L on Durand Road, then go W about 0.45 mile to end of road and park.
STOP 1. BOWMAN MORAINES (Randolph).

At this stop we will examine moraines on the divide between the east-flowing Moose River and west-flowing Israel River. These are the earliest of the Randolph Moraines described above. They are interpreted to have formed at the distal limit of the White Mountain Moraine System during Older Dryas climate cooling and glacial readvance ca. 14,000 years ago.

Leaving the parking area, we will see meltwater channels that drained the first stage of glacial Lake Israel. As we go up the driveway to the Brianas residence, we’ll pass a couple of small NW-trending moraines that are crossed by the pipeline, and then will climb onto one of the prominent moraines seen from Lowe’s Store (Fig. 4). This ridge has a slightly arcuate shape on the topo map, perhaps because its E side has been partly eroded by meltwater flowing off the ice margin. The moraine is up to ~30 m high and trends between NW (in the southern part) and NNW. There are many 1–3 m boulders along its crest, suggesting the ridge is composed of till. The proximal (W) side of the ridge is locally very bouldery and appears more so than the E side. The boulders are white to pinkish-gray, medium-coarse grained, massive to foliated granitic rock with some pegmatite. A deep channel that curves around the N end of this moraine segment was cut by a glacial stream that carried meltwater off to the E. The channel has a prominent boulder lag on its floor. A second large moraine is located SW of the first and is not so high or sharp-crested. Till was formerly exposed in cuts along a logging road on the latter ridge. There are also many quartz boulders on that moraine, which probably came from one of the silicified zones in the local bedrock.

Figure 4. Google Earth view looking west along U.S. Route 2 (left) in Bowman area of Randolph. Lowe’s Store is seen in foreground. Labels indicate moraine ridges and a meltwater channel in vicinity of Stop 1.

4.4 Return to Route 2 and turn R.
8.9 Route 2 crosses one of several hillside meltwater channels (just before large yellow house).
9.4 Turn L on Route 115.
12.6 Turn L on Cherry Mountain Road.
14.7 Turn L on unmarked National Forest dirt road.
15.0 Turn around and park at side of road.
STOP 2. MILL BROOK VALLEY DELTA (A BOWMAN STAGE LAKE ISRAEL DELTA) AND ASSOCIATED ALLUVIAL FEATURES (Jefferson).

According to W. Thompson’s work in the region, the Mill Brook Valley should contain evidence of the earliest stage of ice-dammed glacial Lake Israel (Thompson et al., 2017). Based upon observation of delta topset/foreset beds in a former pit on Route 2 at the junction with Valley Road (Fig. 5), a glacial lake was present in the eastern portion of the Israel River valley. This lake discharged eastward across a saddle near present day Bowman.

Following deglaciation and resultant emptying of the lake, the whole Israel River system reversed course and drained west as it does today. Fowler and Barker’s (2015) mapping of the surficial geology in the Mt. Dartmouth quadrangle documents the Mill Brook Valley Delta, which built northward into Lake Israel. However, it wasn’t until LiDAR was available that the full extent of this delta and other features were revealed. The LiDAR will provide the basis for modifying the current features on the Mt. Dartmouth surficial geologic map.

Figure 5. Ice-margin positions and lake levels as presented by Thompson et al. (2017) with LiDAR lake level projection by Barker. Base map is a slope map derived from LiDAR Digital Elevation Map (DEM) with an overlay of colorized DEM.

Mill Brook is the primary stream in a northwest-facing watershed which starts at the saddle between Cherry Mountain and Mount Dartmouth. This steep terrain creates a high-energy fluvial system. Much of the watershed is covered by glacial till. Glacial-fluvial deposits are also present, in valleys now occupied by modern streams. Additionally, a large alluvial fan deposit has developed along the western flank of this basin, on the lower part of Streator Brook (Fig. 6). This fan was not delineated in the current mapping but will be part of the revisions to this
map. The road leading to Stop 2 (Cherry Mountain Road) provides ample viewing of Holocene stream terrace deposits. Gentle undulations of former stream channels can be seen in the LiDAR (Fig. 6) but are also apparent in the mowed fields along the road. Looking at Figure 6, you will see an impressive alluvial fan along the southwestern flank of the valley.

From where we stop along Mill Brook Road, we will walk downhill toward the confluence of Mill Brook and Appleby Brook (Fig. 6). As the terrain begins to flatten we are entering a portion of the Mill Brook Delta that built out into Lake Israel (Fig. 7) as sediments were deposited into the Bowman Stage of the lake by the late Pleistocene Mill Brook. The delta was subsequently bisected by present day Mill Brook.

There is very limited exposure of the delta but sampling along traverses up a couple of the terraces showed a basal till overlain by sand and sub-rounded gravel and cobble deposits. Projection of the Bowman spillway elevation (457 meters/1499 feet) using the LiDAR DEM provides excellent agreement with the flat upper surface of this deposit being a submerged delta top. Looking closely at Figures 7 and 8 shows that as the lake receded, the delta top became exposed and Mill Brook began incising channels into it. At least three stream channel positions are shown in the LiDAR. Subsequently, Mill Brook established its present course and cut down through the center of the delta.

Figure 6. Surficial geologic map of Mill Brook Valley by Fowler and Barker (2015). Base map consists of LiDAR slope map, overlain by colorized DEM and 1:24k-scale topographic map features and contours.
Figure 7. LiDAR slope map and colorized DEM overlay map of the Mill Brook Delta and surroundings. Surficial map units are denoted by labels and their outer boundaries to allow full viewing of LiDAR slope map and colorized DEM overlay.
Figures 2 and 3 show the relationship of the Carroll Delta to inferred ice-margin positions.

Aerial photographs taken in 1955 (prior to the pit operation) show a network of subparallel channels extending from north to south across the full width of the delta plain. These photos also show a steep ice-contact slope that formerly existed on the north edge of the delta. According to Goldthwait (1916), the delta plain (since removed in the pit area) was “strongly pitted” in the proximal part but smooth and non-kettled in the central and distal portions. He also observed that the fluvial gravel on the delta top was less bouldery toward the front of the delta.

The original delta plain has an elevation of approximately 450 m (1475 ft). It was graded to the Gale River 2 stage of glacial Lake Ammonoosuc, which had a spillway at about 445 m (1460 ft) (Fig. 3). This spillway is
prominently visible along U. S. Route 3 southwest of Twin Mountain village, especially when leaves are off the trees. The Carroll Delta received meltwater and sediment from a succession of lateral channels on the hillside to the northeast. These channels are evident on the topographic map but are much clearer on the LiDAR image (Fig. 9). However, it seems unlikely that the volume of sediment removed from the channels would have been sufficient to build the large delta. Most ice-contact deltas in New England were fed mainly by subglacial tunnel drainage. A possible feeder channel for the Carroll Delta is marked by a short esker segment near the railroad track in the woods north of the delta.

![Image](image_url)

**Figure 9.** Hillshade LiDAR image (“sun” angle from NW) with color DEM overlay, covering part of the Bethlehem quadrangle. Image shows gravel pits in the E and W parts of the Carroll Delta (center and lower-L), meltwater feeder channels (upper-R), and other features discussed in text. Pit on Beech Hill is a rock quarry.

The coarse gravel forming the delta topset beds can be seen along the upper east wall of the pit, and sandy foreset beds have been exposed in places during recent excavations. In 2000, Carol Hildreth observed some remarkable features at the proximal margin of the delta in the northern part of the pit area. She noted thrust faults and overturned folds in the deltaic sediments, along with north-dipping gravel and till layers plastered against the
delta. Hildreth described these features in detail during the 2002 Friends of the Pleistocene trip (Thompson et al., 2002). She attributed them to ice shove during a minor glacial readvance. She also observed younger, undeformed lacustrine sediments banked against the readvance till, which most likely were deposited in the earliest phase of glacial Lake Carroll to the north.

Incision of the original Carroll Delta resulted from a drop in lake level when recession of an ice tongue farther down the Ammonoosuc Valley (toward Bethlehem) opened up lower spillways (Fig. 3). A deep meltwater channel cut into the west side of the delta as the lake fell is seen along the railroad track. The elevation of the channel at this point is approximately 429 m. Chris Dorion obtained a sediment core from the channel floor (Thompson et al., 2009a, 2017). He located a favorable site that accumulated organic material once the channel ceased to carry meltwater and found sands on the channel floor containing a late-glacial flora. The basal radiocarbon age from Dorion’s core showed that the vegetation grew on the channel margins by 11,430 14C yr BP (~13.3 cal ka BP). The channel itself could have formed several hundred years prior to this time. It may have initially carried meltwater directly from the ice margin, and then briefly served as an outlet for glacial Lake Carroll just before the opening of Lake Israel in the Cherry Pond area, where we have ages as old as 11,800 14C yr BP (~13.6 cal ka BP).

STOP 4. LITTLETON-BETHLEHEM READVANCE SITE (Bethlehem Earth Materials Pit, Bethlehem).

Good stratigraphic exposures are uncommon in the White Mountain Moraine System. Most of the moraines are composed of bouldery till, which is seldom mined for its own sake. However, a few long-lasting pits have been worked near the distal margin of the WMMS in the Ammonoosuc and Israel Valleys, where the Older Dryas Littleton-Bethlehem readvance built moraines and other till deposits on top of earlier deltaic and outwash sediments. One such area is the vicinity of Trudeau Road in Bethlehem, including Stop 4 and the active gravel pit and other excavations associated with the North Country Environment Services (NCES) landfill SE of here.

The Bethlehem Earth Materials pit has been operated in two principal levels over the past three years, exposing up to ~30 m of total section. The lower level, in the northern part of the pit area, formerly showed ~9 m of pebble to boulder gravel and sand on bedrock. The rock is coarse, massive to weakly foliated, pinkish granite with xenoliths. It is part of an Ordovician gneiss dome belonging to the Oliverian Plutonic Suite (Lyons et al., 1997). This rock is now being quarried to produce crushed stone.

Most of the current pit operation is in the higher level to the south, where up to 20 m of glacial sediment are exposed in the long E-W face. The appearance of the pit face changes as excavations continue and the position of the working face migrates southward, but it generally shows till overlying poorly-sorted pebble to boulder gravel with lesser sand and pebbly sand (Fig. 10). The gravel reaches the surface in the west end of the pit, where the face curves around to the NW. In July 2017, a fresh cut in the E end of the pit showed an excellent exposure of compact, fissile, silty-sandy till overlying coarse, poorly sorted gravel. Well-rounded stones that are present in the till throughout the pit face most likely were incorporated from the gravel during the ice readvance. Shear structures including deformed sand lenses are also common in the till.

Large masses of glacially disturbed bedrock underlie gravel in the western part of the pit, and solid ledge was recently encountered in the pit floor. A considerable amount of bedrock may be exposed by the time of our NEIGC trip. The original ledge surface will be examined for glacial striations when safe to do so.
Figure 10. View looking SE at section in the Bethlehem Earth Materials Pit, showing gray readvance till (in upper face) overlying glacial gravel. The till has abundant shear structures, including deformed sand lenses, and rounded stones incorporated from the gravel. Photo taken in December 2015; from Thompson et al. (2017).

This pit and most other pits along Trudeau Road are in ridges that trend ENE-WSW. These ridges are best seen on the old topographic map of the Whitefield 15-minute quadrangle. We interpret them as moraines, though they are not as sharp-crested as many others in the WMMS. The Trudeau Road moraines exhibit a readvance stratigraphy that varies in detail from place to place. Recent exposures at Stop 4 have shown just one principal till unit overlying gravel. On the other extreme, many years of excavations and test borings at what is now the NCES landfill area have revealed a complex and variable interlayering of till with silt-sand-gravel units (see above description of Bethlehem moraine complex for details).

The depositional history of the waterlaid glacial sediments in this area is not well understood. The puzzle is complicated by a complex sequence of events during deglaciation, coupled with the jumble of descriptions from various people investigating numerous sites over a long time. We see glaciofluvial gravels, like those in the Bethlehem Earth Materials Pit, while pits and borings around the NCES site have shown abundant glaciolacustrine silt and sand. A basic question is whether the waterlaid sediments occur as local packages, or do they form a widespread continuous stratigraphy of fluvial and lacustrine deposits? Either way, these deposits were overridden by the Littleton-Bethlehem readvance and in places became part of the moraine ridges.

We propose a series of events as follows. As the ice margin receded from the southern Trudeau Road area, outwash sand and gravel was deposited by meltwater streams flowing southward into the upper Gale River basin (probably to glacial Lake Gale, proposed and named by Hildreth (2002)). The ice soon retreated slightly north into the Ammonoosuc basin, resulting in local ponding of deltaic lake sediments in the NCES landfill area. Gravel and sand in the vicinity of Stop 4 may have been deposited by meltwater flowing into this water body. The lake in turn drained southward into the Gale River valley. Then the ice sheet readvanced across the northern Trudeau Road area, deforming the fluvial/lacustrine sediments and building moraines as it experienced oscillatory retreat into the Ammonoosuc Valley. Meanwhile, the upper part of Lake Ammonoosuc (in the Twin Mountain village area) was cutting successive spillway channels SW between the moraines, including the G5 spillway (Fig. 3) at the NCES site. Water flow along these channels reworked earlier glacial sediments and modified the terrain to some degree.
STOP 5. GLACIAL LAKE AMMONOOSUC DRAINAGE CHANNELS (Bretzfelder Park, Bethlehem).

If time permits, we will stop at Bretzfelder Park in Bethlehem to consider the multitude of glacial meltwater channels in this area (Fig. 11). A few channels, like the one seen here along Barrett Brook, were direct outlets (spillways) from the Bethlehem stages of glacial Lake Ammonoosuc (Thompson et al., 1999). Many other channels in this area carried the lake outflow farther west and south to the lower Gale and Ammonoosuc Rivers. The latter channels record a complex drainage sequence along the retreating ice margin, while small recessional moraines were also forming here (part of the Bethlehem moraine complex described above).

The general succession of lake spillways and other meltwater drainages in the Bethlehem area has been recognized for a long time, but the number and intricate pattern of these channels (Fig. 11) was apparent only when LiDAR imagery became available in the past year! The succession of channels and their spatial interrelationships support an incremental lowering of the ice-dammed lake and a rapidly changing series of drainage paths as the ice margin retreated. This scenario is consistent with the progressive lowering of glacial Lake Ammonoosuc recorded by the Carroll Delta and younger deltas/terraces around Twin Mountain village. Moreover, it provides further evidence that the ice margin was a tight dam for the lake, rather than “leaking” randomly through englacial drainage.

Figure 11. Hillshade LiDAR image (“sun” angle from NW) with color DEM overlay, showing meltwater channels near Bethlehem village.
37.3 Exit parking lot and turn R on Prospect St.
38.4 Turn L on Route 302.
46.8 Turn L at light on Route 3.
48.7 Turn R on Route 115.
55.2 Turn L on Route 115-A.
56.9 Keep R at fork, staying on 115-A.
57.6 Turn R onto unmarked private driveway. Continue past house and park in clearing at end of woods road.

STOP 6. MORAINES IN THE ISRAEL RIVER VALLEY, JEFFERSON (Kluckie Pit, Jefferson).

This stop is on the NE side of the Israel River valley, midway between Bowman (Stop 1) and Lancaster. Mapping of the Jefferson quadrangle (Thompson, 2016) delineated a few moraine ridges here, along with curious round to irregular hummocks. In the 1930s, James Goldthwait conducted a gravel survey for the State, and his unpublished map of the area likewise showed the latter features, which he labeled as “till hummocks with blocks [boulders]”. At Stop 6 we can see some of the distinct moraine ridges. These and most other such ridges in Jefferson trend E-W to NE-SW and are associated with the more numerous till hummocks. LiDAR imagery (Fig. 12) shows that the area of hummocky moraine extends farther up the Israel Valley than previously known. The part of the valley floor between the “Old Railroad Grade” and the southeast corner of Figure 12 contains this newly recognized part of the hummocky moraine complex, which had been concealed by forest cover, swampy terrain, and the nondescript contours of modern topographic maps. A SE-trending esker ridge appears in the NW to central part of Figure 12. This discontinuous gravel ridge marks the path of a subglacial meltwater drainage tunnel that probably conveyed sediment to Lake Israel deltas and subaqueous fans in the upper Israel Valley.

Figure 12. Slope map with color DEM overlay, showing extensive complex of hummocky moraine and short moraine ridges in Jefferson.
Figure 13 shows an excavation in the area where we will park. It recently exposed a nice longitudinal section through a moraine ridge that trends E-W, with a crest elevation of 342+ m. (This section may have slumped considerably by the time of our visit.) The moraine consists entirely of till, with an exposed thickness up to ~5 m. The till is light olive-gray, massive, loose to compact, sandy, and very stony. A few clasts show glacial striations. Various rock types occur here, including many fragments of local bedrock from an Oliverian gneiss dome of Ordovician age (Lyons et al., 1997; Baker et al., 2016). None of the beige or pinkish flow-banded Jefferson rhyolite was seen in this pit, despite its abundance in float boulders and artifacts at Paleoindian sites just NW of here, on the other side of Route 115-A. Other moraine ridges occur in the clearing S of the parking area. We will examine one of these if time permit.

The stony till at this locality resembles that which forms the moraines of the WMMS. It likewise may have been deposited at the margin of the receding ice sheet, but there is little evidence of depositional processes in the presently exposed section. The depositional environment of the nearby hummocky deposits is even more uncertain due to the lack of exposures. Their location is slightly down the Israel valley from the inferred proximal limit of the WMMS, and the hummocky topography is different from that of the moraine belt. We infer that the hummocks record local stagnation of the Israel Valley ice lobe following the Older Dryas readvance that built the Bowman Moraines. The decaying ice probably developed depressions into which flow till slumped from the debris-rich glacier surface. Some of the hummocks lie below the inferred contemporary level (~339 m) of the Baileys Stage of glacial Lake Israel. We expect that excavations of the mounds would reveal mixtures of diamict with fluvial and lacustrine sediments showing collapse structures, of the sort described by Benn and Evans (2010).

Figure 13. Pit face showing longitudinal section along distal (S) side of a moraine in Jefferson. Photo looks north and was taken in 2015.

58.2 Return to Route 115-A and turn R.
59.8 Turn L on Route 2 in Jefferson village.
60.5 Turn R on North Road.
62.7 Turn R on Gore Road.
64.0 Sharp bend to L, on what is now Garland Road.
64.5 Turn R on Pleasant Valley Road. This road climbs up the gently sloping lower part of the Garland Brook alluvial fan.
65.2 Jct. with Arthur White Road. Keep L on Pleasant Valley Road.
65.5  Turn R on Community Camp Road. Follow leaders beyond end of town road and onto the Crane family’s private road.
66.9  Park in upper part of large shallow gravel pit.

STOP 7. BUNNELL BROOK ALLUVIAL FAN (P&R Excavating Pit, Lancaster) & SAPROLITE EXPOSURE.

At this stop we will first examine one of the many alluvial fans that have developed on the lower slopes of the White Mountains. The Bunnell Brook fan is a composite feature consisting of an earlier, higher, and poorly understood fan deposit (Qfb1 on the map by Thompson, 2016), which was incised and followed by the lower fan adjacent to modern Bunnell Brook (Qfb2). The Qfb2 fan merges downstream with a large, more gently sloping fan along neighboring Garland Brook to the south. We have not studied these fans in detail, so our observations are preliminary and will hopefully encourage further studies of fan deposits in the White Mountains.

Stop 7 is a broad shallow gravel pit in the lower Bunnell Brook fan (Fig. 14). The pit exposes about 3 m of very coarse, poorly sorted pebble-to-boulder gravel. Many of the gravel stones are somewhat angular, suggesting they have not been transported far by water. Boulders ≥ 1 m in diameter are common here, and a few are up to 3 m and larger. The LiDAR image shows a braided stream channel pattern on the fan surface, supporting a fluvial origin for this deposit.

Farther up the valley, Figure 14 reveals a cluster of arcuate head scarps where landslides have occurred next to Bunnell Brook. A reconnaissance of the slide zone did not reveal fresh sections, but a small recent slide was found southeast of the brook an elevation of ~ 570 m, just north of the former (now closed) Mt. Cabot Trail. The fresh section exposed by this slide showed sandy, stony diamict (probably colluvium) and poorly sorted gravel sharply overlying stratified, water-saturated silt-sand interpreted as a lacustrine unit. This stratigraphy suggests that the large NW-draining basin of Garland Brook and its tributaries may have been occupied by an ephemeral ice-dammed glacial lake during deglaciation of the area. Loading of the lake sediments by glacial or postglacial deposits may be responsible for triggering the Bunnell Brook landslides.

Figure 14. Hillshade LiDAR image (“sun” angle from NW) with color DEM overlay, showing the Bunnell Brook fan and headscarps of landslides along upper part of the brook.
The coarse gravel at Stop 7 resembles many other fan sediments seen in stream banks and pit exposures around the White Mountains. We found very little published research on these fans, unlike the debris avalanches that have received considerable attention in northern New Hampshire. The stratigraphy, mode(s) of deposition, and depositional chronology of the fans are still poorly understood. Gravel exposures, observations of modern brooks, and LiDAR channel patterns on fan surfaces suggest that transport by steep mountain streams was the dominant process, but debris flows may have played a significant role in fan construction. The criteria used by Jackson et al. (1987, p. 116) to identify debris flows in the Canadian Rocky Mountains may prove helpful in the White Mountains:

“Fans were classified as having a debris flow component on the basis of the following field geomorphic and/or sedimentologic criteria: (1) exposures of debris flow diamictons characterized by weak stratification, poor sorting and matrix-supported angular clasts… (2) the presence of debris flow levees or debris flow lobes on fan surfaces…and (3) the presence of oversize (>1m) lone boulders on the fan surface…”

Diamicts have not been observed in the lower Bunnell Brook fan (Qfb2), although pit owner Paul Crane said that he encountered a buried clay layer in one part of the pit at Stop 7. Landslide scars in the upper Bunnell Brook valley (described above) indicate that slide events likely supplied a lot of sediment to the fan on one or more occasions and over relatively short times, possibly including mud flows and/or coarse debris flows. Shallow roadside exposures in the older fan unit (Qfb1) did reveal sediments of this type. Along White Rd., Thompson found a cut showing ~1 m of pebbly sand overlying silty-pebbly diamict, and other ditch exposures showed weakly stratified pebbly diamict, silt, and sand. On the northern part of the Qfb1 fan surface, a ditch near the lower end of Community Camp Rd. (aka Mt. Cabot Rd.) exposes stratified clay-silt, locally overlain by sandy pebble gravel in one place and containing a sand lens in another. The best exposure of debris flow facies that we have seen in the region was at the Drouin Pit, located in a composite landslide/fan that resulted from a huge prehistoric slope failure on the hillside just south of the village. The guide for Trip C3 in the 2009 NEIGC guidebook describes and illustrates the stratigraphy at this locality (Thompson et al., 2009b).

The greatest uncertainty surrounds the depositional history of this and other fans in the White Mountains. Most of them are forested, commonly with roads and houses, and show few signs of major aggradation in historical time. It is tempting to propose that these extensive coarse-gravely fans formed in large part just after deglaciation, when glacially derived sediments on the mountainsides remained unstable and were not yet anchored by tree cover. This theory is consistent with a study in British Columbia whose authors concluded that “rapid sedimentation during the paraglacial period contrasts sharply with present-day conditions” (Church and Ryder, 1972, p. 3059). However, Jennings et al. (2003) carried out a detailed study of fans in Vermont (a geographically more comparable area!) and documented a long history of episodic deposition spanning the Holocene. The fans described by the latter authors are much smaller, finer-grained, and better stratified than those in the northern White Mountains. Thus they preserve datable soil horizons that yielded a chronology of sedimentation events alternating with periods of stability. Calibrated radiocarbon ages from wood samples near the base of the Eden Mills fan in the northern Green Mountains indicate rapid sedimentation immediately after deglaciation, but also the presence of woody vegetation less than 1000 yr following ice retreat (Jennings et al., 2003, p. 195).

In contrast, Wells and Harvey (1987) described a series of alluvial fans in northwest England that were deposited in response to a single very intense rain storm. The duration of the storm was only about 2.5 hours, yet it generated 13 alluvial fans ranging from debris flows to fluvial deposits. Estimated volumes of these fans range from 50 to 2,380 m³ – smaller than those in Vermont but nevertheless impressive for one brief storm. The affected region is mountainous but lacks tree cover, so the degree of sediment mobilization may have been greater than would occur from a similar storm in the White Mountains.

Stop 7-A. After leaving the gravel pit, we will stop at an interesting road cut on the access road. The cut shows a massive to weakly layered, varicolored, pasty sediment that appears to be a mixture of clay, silt, and sand-size particles. The clay may be very local here, because stony diamict is exposed along the roadside a short distance to the southeast. A sample of this clay-rich sediment was analyzed by Alexander Falster, Research Technologist at the Maine Mineral and Gem Museum in Bethel, Maine. Its chemical composition determined via energy dispersive spectral analysis reveals a high-Al silicate bulk composition with minor Fe, Na and Mg. The high Al content suggests the presence of kaolinite, associated with members of the smectite group (montmorillonite). Fe appears to
be mainly a staining agent such as late goethite and/or hematite. X-ray diffraction analysis is needed to confirm these minerals (A. Falster, pers. comm., July 2017).

Figure 15. Clay exposure at Stop 7-A. The stones on upper part of section probably were introduced by road work.

The clay minerals suggested by Falster’s analysis commonly occur in saprolites formed by weathering of bedrock over long time intervals (Carroll, 1970). Saprolite is uncommon in the glacially-scoured terrain of New Hampshire, though remnants have occasionally been found that survived glacial erosion (Fig. 16). Clayey sediment like we see in this road cut was also encountered in digging a house foundation ~290 m downslope (SW) from here, so it covers a sizable area in that direction. While most of the clay at this locality has an orangey to brick-red color, some of it is black, white, or even pale greenish. These same contrasting colors were noted when digging the foundation hole down the street.

Assuming this is a saprolite, its mode of emplacement is uncertain. The possibilities are: development by in-situ rock weathering, glacial transport, or mass wasting from higher on the mountain. Considering the lack of remnant parent rock or evidence of glacial deposition at this site, we propose the clay was deposited as a mudflow from the mountainside to the northeast. LiDAR imagery reveals now-dry alluvial channels, possible slump scars including a long N-S scarp just north of here, and many curious lobate features suggesting mass-wasting processes. Mobilization of saprolite from the deep Bone Brook ravine seems like a good working hypothesis. There is a scarp at the head of the brook that may indicate landslide activity. Also, the upper part of the Bone Brook drainage would have been more protected from glacial erosion than the area of Stop 7-A, and thus would have been a favorable place to preserve remnants of preglacial saprolite. The bedrock in that area is Jurassic granite (Lyons et al., 1997; D. Eusden, pers. comm., 2017), which is consistent with Falster’s preliminary analysis of the clay.

We have not thoroughly researched saprolite landslides, but some insights were found in a paper by Durgin (1977) on the relationship between landslides and the weathering of granitic rocks. This author notes that granite saprolites are prone to slumps (rotational slides), and he observes that “Ground water drains through the failure surface and relict joints, precipitating iron and manganese” (Durgin, 1977, p. 130). The black streaks seen in the clay at this site are likely Mn oxides. And finally, are there any unweathered stones within the clay unit that were deposited contemporaneously with the clay? If so, they might indicate the source and mode of transport of the saprolitic sediment. Maybe we will find some on today’s trip!
Figure 16. Section in excavated area above rock quarry on the east end of Beech Hill, Carroll (seen in Fig. 9). This remarkable exposure showed in-situ saprolite (freshly scraped area in lower-central part of photo) developed in metamorphic rock. At the top of the orange saprolite, the relict vertical rock foliation was dragged to the left by glacial ice when the thin cover of lodgement till was deposited. Note oriented (and striated) rock clast in the gray till just above shovel blade. Photo by W. Thompson, 14 October 2005. View looking SSW.

END OF TRIP. Return to Bethel, Maine via U. S. Route 2.

REFERENCES CITED


______1878b, Atlas accompanying the report on the geology of New Hampshire: Julius Bien, New York, large folio, 17 sheets.
THOMPSON AND BARKER


Lougee, R. J., n.d. (ca. 1930), The origin and occurrence of glacial washed deposits in the White Mountain region: Unpub. manuscript based on field work for New Hampshire Highway Department, 26 p. (Baker Library Special Collections, Dartmouth College).


_______1904, Moraines and eskers of the last glaciation in the White Mountains: The American Geologist, v. 33, p. 7-14.

PALEO-DUNES AND OTHER POST-GLACIAL ODDITIES IN THE WOODS AND FIELDS OF NEW SHARON AND CHESTERVILLE, ME

Patricia M. Millette, Mt. Blue High School, ME 04938 patti.millette@maine.edu
Benjamin Andrews, Mt. Blue High School, Farmington, ME 04938
Anna Glass, Dept. of Geology, Bates College, Lewiston, ME 04240
Thaddeus Gunther, Dept. of Geology, Bates College, Lewiston, ME 04240
Roshan Luick, College of the Atlantic, Bar Harbor, ME 04609

INTRODUCTION

This field trip is a relatively easy set of hill-walks through select areas in the towns of New Sharon and Chesterville Maine (fig 1). Although many field researchers would not consider the towns of New Sharon and Chesterville Maine to be the most prestigious or awe-inspiring geologic research locations, the participation of high school students and other locals has allowed the hidden glacial and post-glacial treasures of these two towns to come to light (and there is at least one spectacular view along the way). The simple question,”Can you tell me what this strange mound of sand out in my back pasture is,” and the astute observations of a high school student behind his house, launched an ongoing quest and a multiyear collaboration between the University of Maine and Mt. Blue High School earth science students to discover the story behind a collection of glacial and post-glacial deposits in this area. Although the trip principally highlights various post-glacial aeolian features, it also includes a crag-and-tail deposit, and glacial-marine deposits. If time permits, a short foray into the local uses for Cape Cod Hill Devonian granite is included as well.

TRIP REQUIREMENTS

Casual walking shoes/sneakers will probably be adequate for this trip, but since trip participants will be walking in woods and fields, light-colored, long pants as tick deterrent would be advisable. Bring a lunch and water bottle. Refills of water bottles will be likely, although bathroom stops will be spotty. That being said, several landowners have agreed to allow participants to use the facilities if necessary. Also, there is a general store (Douin’s Market, at the intersection of Rte’s 27 and 2 in New Sharon, 420624.80m E 4943142.95m N) with good sandwiches and other lunch supplies, very close to the starting point (fig 1). Since the trip will end back on Cape Cod Hill at Stop 1, it will be possible to consolidate vehicles, and leave some at that location.

LOCATION AND GENERAL SETTING

The two general locations that are the focus of this field trip are in the foothills of west-central Maine. One location contains several sites along Cape Cod Hill in the town of New Sharon. At its highest point, the hill stands approximately 890 ft (271m) asl, and the initial study site is at approximately 460 ft (140m) asl.

At the Chesterville location, the first site, is on the northern border of the town along Chesterville Hill between 350 and 480 ft (106-146m) asl. Chesterville Hill is a localized section of Old Bluff Hill/Blabon Hill. (Some publications refer to it as Chesterville Hill and some as Old Bluff Hill). The second Chesterville site is on the northeast flank of Zion Hill at approximately 380 feet asl, just west of Chesterville Hill. The third Chesterville site is between the Chesterville Hill and Zion Hill. The state of Maine bedrock geology map shows that all three hills are Devonian granite (Osberg et al.,1985) which appears as local outcroppings. Both locations are south of the Sandy River Valley, which changes course at Farmington Falls from a southeast flow upstream from the Falls, to a northeast flow downstream from the Falls (fig 1). It eventually empties into the Kennebec River at the town border between Starks and Norridgewock.

In addition to the value of the sediment itself for commercial purposes, many of these deposits are also important as a local groundwater supply. Groundwater is close to the surface and springs are visible at two of the locations. Besides their scientific importance, commercial uses, and groundwater potential, these mounded sediment deposits are a noticeable contrast to the topography of the adjacent hayfields and woods, and consequently, they are somewhat of a local curiosity.
Deglaciation in Central/Western Maine

Dorion (1997), Thompson (2001), and Retelle and Weddle (2001) date Laurentide deglaciation in central and western Maine between 13.5 kya, and 11.5 kya. Glacial retreat from Maine was primarily to the northwest, with recession of the thinning ice somewhat controlled by local topography (Jager, 1996; Thompson, 2001; Retelle and Weddle, 2001; Tary et al., 2001; Greve and Syverson, 2003). By 12.5 ka, eustatic sea level was well below its current position (at least 70 m or 230 ft), but because of isostatic depression of the crust, the marine limit extended at least 132 m (433 ft) in elevation from current sea level (Belknap et al., 1987), and resulted in the deposition of sediments that outline the extent of that marine limit (Thompson and Borns, 1985; Neil, 2007). Following this stillstand, rapid regression took place, and at approximately 11-10.5 ka, a relative sea-level lowstand occurred at -55 m (180 ft) (Barnhardt et al., 1995). As rebound slowed to about the same rate as eustatic sea-level rise, various coastal

Figure 1: All of the field trip sites in New Sharon and Chesterville, ME (USGS, 1968)
features were formed at that time, until eustatic rise subsequently became dominant (Belknap et al., 1987).

**Glacial and Glacial Lacustrine Deposits**

As one of the first modern researchers to study sediments specifically in the New Sharon/Chesterville area, Caldwell (1959) made many observations about sedimentary deposits there. Among these, he noted thick layers of till in the New Sharon gorge. Although he and other researchers do not necessarily agree with the timing of the till’s deposition (Caldwell, 1959; Borns and Calkin, 1977; Caldwell, et al., 1985; Weddle and Caldwell, 1986; Weddle, 1989; Weddle, 1992; and Weddle et al., 2006), they do agree that many of the glacigenic sediments in the New Sharon section of the Sandy River Valley were ice-contact or ice proximal. Researchers have proposed that the entire sequence of sediments was deposited during the early Late Wisconsin (Weddle and Caldwell, 1986; Weddle et al., 1989; Weddle, 1992). (Weddle, 1992).

Deformation in the entire sequence occurred as glacial lobes from two different directions, the Kennebec Valley from the northeast and the Sandy River Valley from the northwest, oscillated during glacial advance (Weddle, 1992). These till deposits are potentially very important contributors to the formation of the deposits being studied because the advance of the Kennebec River Valley lobe was partially responsible for damming the Sandy River Valley, creating a pro-glacial lake there (Weddle, 1989, 1992).

As a result of the New Sharon dam, glacial lacustrine deposits can be found as a series of delta remnants in several tributary valleys near New Sharon and Chesterville. These tributaries were conduits through which sediment was transported and eventually graded to lake level. They can be found along with rhythmically bedded sediments composed of sand, silt, and clay (Weddle and Caldwell, 1986).

**Glacial Marine Deposits**

Belknap et al. (1987) also showed that by 12.5 ka, because of isostatic depression of the crust, the marine limit was at least 132 m (433 ft) higher than the current sea level. For reference, the Sandy River between Chesterville and New Sharon is at approximately 320 feet (97 m) asl (USGS, 1968). Jager (1996) discovered deposits of glacial marine clay at elevations of approximately 470 feet (143 m), as far north as Lexington Township (approximately 18 miles (30 km) north of New Sharon and Chesterville). These marine sediments are covered with braidplain deposits, deltaic sediments, and nearshore deposits (Jager, 1996) which indicate that alluvial processes took over following uplift and marine regression. Even though Jager’s description is for the Carrabassett River Valley, not the Sandy River Valley, Neil (2007) mapped marine sediments in the Sandy River Valley as far north as the town of Strong at 470 feet asl (143 m), approximately 15 miles (25 km) northwest of these study sites. Therefore it makes sense that the till dam in New Sharon was breached while in proximity to the ocean and local relative sea level reached elevations of 470 feet asl (143 m) through the New Sharon gorge as well. This is especially likely if the highest outlet from the lake was only at 385 feet (117 m) asl (Weddle and Caldwell, 1986).

**Eolian Deposits**

Eolian deposits have been identified in various parts of Maine, and are abundant along the Kennebec River near Madison (McKeon, 1989). These dunes, described in detail by McKeon (1989), were identified as either longitudinal or wind-shadow deposits, and the dune axes of forty-seven of the sixty-four longitudinal dunes in McKeon’s study are oriented with a strong northwest/southeast trend. McKeon (1989) hypothesized that they formed along the Kennebec as the post-glacial isostatic uplift outpaced the eustatic rise in sea level. In this case, sediments coarser than the glacial-marine clay were first deposited, and then sandier sediments (Embden Formation) (Borns and Hagar, 1965) eventually transported and prograded onto, and interfingered with, fine sediments of the Presumpscot Formation, providing ample material for eolian transport processes McKeon, 1989).

The nature of eolian bedforms depends on both the availability of sand and the nature of the wind regime. Because longitudinal dunes commonly forming approximately parallel to wind direction and have two slip faces (or varying slip faces), this indicates a variation in prevailing winds. These contrast with transverse dunes which form perpendicular to the prevailing wind direction and have only a single slip face (Reffet, et al., 2010). These further contrast with parabolic style dunes which develop vegetated sections in isolated areas, and un-vegetated areas allow wind to scour parts of the dune causing its typical shape of up-wind “arms” (fig 4).

Thorson and Shile (1995) show that the northwest trend of parabolic dunes on the floor of Glacial Lake Hitchcock in western Massachusetts indicates a shift from anticyclonic wind patterns that were present over the Laurentide
ice sheet to predominantly cyclonic patterns (northwesterly wind) by the time the western New England lake drained. They speculate that this might have occurred at approximately 12.4 ka. Sediments in these conditions would have been mobile, especially before stabilization by vegetation. It makes sense that these conditions would have likely been present over Maine as well.

Because of the presence of both glacial-fluvial, lacustrine, and marine deposits in this area, and because the sediments would have been exposed upwind of the western slopes of both Cape Cod Hill and Chesterville Hill with a relatively long fetch along the Sandy River Valley, it is possible that the combination of abundant sediment and relatively steady wind was adequate to build dune ridges along the three hills (Millette, 2014).

**Glacially Streamlined Features**

Other mounded linear sediment features that are found in Maine also include drumlins and crag and tail features. Both are glacially streamlined features and occur during the advance of a glacier, generally by reshaping subglacial till deposits into elongated, more or less, tear-shaped deposits (Munro-Stasiuk et al., 2013). They differ in that the sediment tail of a crag and tail feature is positioned in the lee side of a more resistant bedrock protrusion, and drumlins often occur in large fields, such as the thousands of drumlins in the Finger Lakes area of New York State (Kerr, 2007), and in the lee of the Mt. Agamenticus uplands in Maine (Nelson, 2003).

**INTERPRETATION OF NEW SHARON AND CHESTERVILLE DEPOSITS**

Mounded sediment features along the western flank of Cape Cod Hill in New Sharon, and Chesterville Hill in Chesterville, Maine are most likely eolian coastal dunes, and the large streamlined feature at the southern end of Cape Cod Hill is likely a till-tail (crag and tail). During glacial advance, it progressed south over the top of Cape Cod Hill, scratching striations into the surface of the Devonian granite, and carrying englacial and subglacial till. As it crested the hill, it may have consolidated and then deposited till on the south side of the hill (fig 2). As it continued to the south, it streamlined parts of the till tail and possibly deformed its east/west profile by a somewhat asymmetrical shoving of the sediments (Millette, 2014).

During deglaciation, if the Sandy River ponded behind the till dam at New Sharon, and subsequent flooding from the ocean occurred when post-glacial sea levels rose above the dam, deltaic sediments from multiple feeder streams would have been deposited in large quantities into these low lying, somewhat protected floodwaters. Additionally, areas between the hills (then islands) would have been draped with marine sediments, such as is the case at

![Formation of Mounds on Cape Cod Hill and Chesterville Hill](image)

Figure 2: General Illustration of mound structures on Cape Cod Hill in New Sharon and Chesterville Hill, Chesterville. Includes dune, drunlin, and till tail formation. (modified from Millette, 2014)
Following the breach of the dam in New Sharon, the river would have returned to its normal course and left significant quantities of fine sediments vulnerable to the wind. With a strong prevailing wind from the northwest (Thorson and Shile, 1995), these unprotected sediments would have been transported to the flanks of Cape Cod and Chesterville, and Zion Hills and deposited there when the strength of the wind could no longer carry them any farther up onto the side of the hill (fig 2). Possibly the wind had a more direct approach on Chesterville Hill, resulting in the sediments being carried to a higher elevation than on Cape Cod Hill. Because of this sustained northwest wind, sediment deposits would have been elongated into more or less linear forms with longer, less steep windward faces, and shorter, somewhat steeper leeward slip-faces (Millette, 2014).

Because of large scale changes in post-glacial climate, notably around 10 ka, and possibly between 13.5 ka and 12.3 ka, vegetation changes were rapid (Jacobson, et al., 1987). Whole assemblages and abundance of specific vegetation were dramatically altered within each 2,000 year time frame, and between 14 and 12 ka, shorter (not trees) boreal vegetation types were abundant (Jacobson, et al., 1987). These smaller species of vegetation would have been available to partially stabilize some of these dunes, which may have facilitated the formation of parabolic blowouts on the upwind side. Eventually all of the dunes were stabilized and show vegetative communities unique to their drier, sandy, well-drained soils (White Pine, Red Oak, Eastern Hemlock, Sugar Maple, and White Ash) (Millette, 2014). In more recent historical times, the sand in these dunes has been excavated for local commercial purposes, and there is current local debate over their mining for commercial uses versus their protection as groundwater storage.

ACKNOWLEDGMENTS

Thanks to all of the landowners who not only generously volunteered access to their properties, but who also volunteered to help out with water and bathrooms Paul Mills, Karen Yingling and Helmut Bitterauf, Margaret Cox and Eric Johnson, Kristen Plummer and Deon Olmstead, Anstiss Morrill, Jason Sawyer, Rob Rogers, Dave Fuller and Shirley Hager, Jim and Deb Farley and all the horses, and the Heikenin family. Thanks to Jim Lisius for arranging access to the Farmington Falls water district pump house, and to Rob Rogers who donated his cool north arrow for all the maps. Thanks to Dan Belknap from the University of Maine for access to the GPR unit, and finally thanks to all of the Mt Blue High School students who carefully researched all of these sites.

ROAD LOG

ORIGINAL MEETING PLACE, 419669.92m E 4943034.02m N: New Sharon Town Office, located on the corner of US Rte 2 and Cape Cod Hill Rd (formerly the New Sharon Elementary School) AT 8:30AM (fig 1). Cumulative mileage is given from the original meeting place, and indicates the destination at each leg of the trip.

0 miles Turn right out of the Town Office Parking lot onto Cape Cod Hill Rd.
1.1 miles Pull into the long driveway at 249 Cape Cod Hill Rd, between the house on the right, and carriage house on the left.

STOP 1: AEOLIAN DEPOSITS, (FIG 3), 418615.56M E 4941767.26M N.

With this spectacular view of the western mountains is the site of our initial mystery mound (fig 1). There are some excellent examples of longitudinal and parabolic dunes in this line of features down the west side of Cape Cod Hill, as well as an example of human use for the sediment in the mound in the lower pasture of this farm. Looking to the southwest, Chesterville Hill can be seen as the next ridge over, and sections of the Sandy River Valley are just visible to the northwest. Traveling south along Cape Cod Hill Rd, large parabolic dunes are seen across from the corner of the Dyer Brown Rd, as well as some simpler dunes. Parabolic-shaped features on the west ends of many of the features suggest that during their formation, the dunes were variably stabilized by vegetation (fig 4). The concave features on the east end of a few mounds may suggest that the dunes were non-vegetated long enough at those locations to form barchan-type “wings” on the lee side. Conversely, they may also indicate the removal of sediments
for domestic or commercial uses, as is documented for the first dune at this stop (Mills, pers.comm., 2008), and in sand pits currently being excavated on Chesterville Hill.

Figure 3: Location and topography of stops 1, 2, 3, and 4 (Map imagery from USGS, 1968)
A ground-penetrating radar trace of one of these dunes can be seen in fig 5. Internal reflectors of this feature tested with the GPR showed characteristics that are consistent with simple dune formation.

1.7 miles  From Stop 1, turn right (south) onto Cape Cod Hill Rd, and travel .6 miles. Turn left (southeast) onto the Dyer Brown Rd, and pull off the road to the right in front of the first house after the corner.
STOP 2: LONGITUDINAL AND TRANSVERSE DUNES, 418353.09M E 4940824.04M N.

This location includes an excellent example of a large longitudinal dune oriented northwest to southeast (Shown as M-104 in fig 6). It is adjacent to another mound with a perpendicular orientation to most (Shown as M-103 in fig 6). It is an obvious exception and its orientation and successive profiles (fig 7) suggest that it is a transverse dune. Its location next to a very large longitudinal dune suggests the possibility that wind was funneled in primarily one direction alongside that larger dune, resulting in its transverse orientation.

Figure 6: Dune locations north end of Dyer Brown Rd., Cape Cod Hill, New Sharon, (Base imagery Google Earth, 20113) (Millette, 2014)

STOP 3: DEVONIAN GRANITE CENTER-CHIMNEY FOUNDATION, 418353.09M E 4940824.04M N: (across the Dyer Brown Rd. from stop 2).

Stop 3 does not include any specific sedimentary deposits, although the house here does sit on top of one of the simpler dunes. It is included to show local usage of the local Devonian granite that came from, according to the owners, a quarry on the Smith Rd (the eastern side of Cape Cod Hill). This house was built circa 1820 and has a wonderfully crafted chimney foundation made of Cape Cod Hill granite. The owners were generous in their offer of its viewing. It is a quick stop, but worth the look (also the owners have volunteered their bathroom and water supply as well).

3.1 miles From Stops 2&3, continue southeast 1.3 miles along the Dyer Brown Rd to its end. Park along the roadside at Stop 4.

Traveling towards the east end of Dyer Brown Rd, it becomes clear when sediment on the roadside starts to show cobbles and gravel. This is a harbinger of the next feature. Notice that we have gone southeast of the highest elevation of Cape Cod Hill at this point.
STOP 4: CRAG AND TAIL/DRUMLIN FEATURE, 419866.02M E 4939438.06M N.

Stop 4 marks the end of the Dyer Brown Rd, and its connection with the eastern end of the Smith Rd. (It is not recommended that anyone drive through to the Smith Rd from here). When compared to the dunes already seen on Cape Cod Hill, the large feature at the extreme southern end of the hill has a similar orientation, and an elongated shape, but its grain size distribution suggests an origin very different from all the other sediment features in the area and suggests that it is till. Its grain-size distribution, along with its streamlined shape, orientation parallel to striations on Cape Cod Hill (Thompson and Borns, 1985) and position on the south side of Cape Cod Hill, suggest it is a drumlin-like or a crag-and-tail feature, or more likely a combination of both (Millette, 2014).

4.5 miles Turn around and drive 1.3 miles back to the junction of Dyer Brown Rd., and Cape Cod Hill Rd.

4.8 miles Turn left at Cape Cod Hill Rd and travel west for .2 mi. Bear right onto the George Thomas Rd. across the (partially dirt) intervale. Along the way, notice the relatively flat terrain across the intervale.

5.6 miles Notice an oxbow lake cut off from the Sandy River in the Flood of 1987 visible on the right.

5.9 miles Take care crossing the narrow stone bridge over MvGurdy Stream. At the visible confluence of McGurdy Stream and the Sandy River are some excellent outcappings of Presumpscot clay.

6.9 miles Stop at the junction of George Thomas Rd and Rte 41. At the end of the George Thomas Rd., turn north briefly onto Rte 41).

7.2 miles At the junction of Rtes 41 and 156, turn west onto Rte 156.

7.3 miles Go approximately .1 mile and stop at the Morrill Homestead. It is the first house on the right-hand side of the road. Pull as far off the road as possible, since this is a busy section of the road.
STOP 5: ABANDONED WELLS, (414652.47m E 4941184.23m N) ADDITIONAL DUNES, (414627.71m E 4940358.80m N, and 414920.07m E 4939787.69m N) (FIG 9).

The first part of this stop is to locate the three abandoned water wells at the northern-most base of Chesterville Hill. Although the dunes on Chesterville Hill are designated as lower groundwater flow areas (less than 10gal/minute) (Neil, 2000) they provide a contribution to the municipal supply for the village of Farmington Falls (Tolman, 1999). Notice the pump station for the Farmington Falls water company at the bottom of the hill. If time permits, a tour of the pump house by the former head of the water company is possible.

Figure 9: Location of stops 5, 6, and 7. (Imagery from USGS, 1968)

7.8 miles From the Morrill homestead, drive .5 miles to the Rogers/Fuller/Hagar properties, beginning at 110 Chesterville Hill. We will park here and walk into the woods and loop back around again to the vehicles.
At this location, a regular pattern of longitudinal dunes is apparent throughout the woods. The identification of additional dunes on the west side of Chesterville Hill signifies that dune formation in the area around the Sandy River Valley was not unique to Cape Cod Hill, and their origin was likely from similar conditions. Because of its placement where the Sandy River turns to the east, a long fetch was possible before hitting the north end of Chesterville Hill. This may help explain the relatively higher elevation dunes on the hill.

8.1 miles Drive .4 miles up Chesterville Hill to the junction of Chesterville Hill Rd., and the Stinchfield Hill Rd. At the fork, bear left onto Stinchfield Hill Rd. Although there a few more higher elevation dunes on the Chesterville Hill Rd. (520+ft asl. (Haslam et al, 2017), landowners there are sensitive to having groups of people on the property.

8.3 miles Go slightly past the first house/farm on the right and park along the Chesterville Hill Cemetery.

At the Farley Farm, there are obvious dunes seen around the house and surrounding horse paddocks, and at the Chesterville Hill Cemetery. Their elevation is just over 500 ft asl. (Millette et al., 2016a) This placement of the old cemetery underscores the ease in which deceased persons were buried in the dunes, since the sand here is extremely easy to move with only a shovel. (Feel free to talk to the horses and chickens here).

9.2 miles Turn around and travel back down the Chesterville Hill Rd to Rte 156.

11.8 miles Go west on Rte 156 for 3.5 mi, and stop at the stop-sign in N. Chesterville.

12 miles Turn south briefly on 156 and go across a small bridge to a 3-way intersection of 156, Zion Hill Rd, and Valley Rd.

12.7 miles At the intersection, bear left onto the Valley Rd. to Stop 6. Park carefully along the roadside at 181 Valley Rd.

STOP 6: VALLEY ROAD DUNE, 412884.33m E 4938179.07m N:

The placement of this significant dune on the northeast end of Zion Hill, (approximately 400 ft asl) suggests that it was formed by winds coming from more than one direction in the flatter terrain to the west of Chesterville Hill and to the north of Zion Hill, causing an elongated mound with a step crest on the southeastern end (Millette et al 2016b). A significant part of the dune is beneath the house, and has been modified somewhat (according to the landowners). However, the crest and dune ridge are fairly evident south of the house itself.

13.6 miles From Stop 6, continue south on the Valley Road .8 miles to the stop-sign at the intersection with the Pope Rd.

14.1 miles Continue onto the Pope Rd and drive south for .4 additional miles to the Lowell Cemetery.

STOP 7 LOWELL CEMETERY: GLACIAL MARINE SEDIMENTS DRAPED OVER AN UNKNOWN SUBSTRATE, 413488.19M E 4936243.88M N.

Like many of the others in the area, this mound’s axis is oriented from west to east suggesting that the mound could also be aeolian. Additionally, the profiles show that the mound has a gradual slope on its northwest side and a very steep slope on the south and southeast side. Due to the prevailing northwesterly wind (Thorson and Schile, 1995), this also suggests that mound might be a longitudinal dune. However, the steepness of the southern side of the mound continues along the perimeter of the adjacent field and becomes the bank of the stream there.

Fine-grained sediment (sand and mud), and its location between Zion and Blazon Hill indicate that the mound is not aeolian. The shape of the mound could indicate that the mound is a landscape anomaly draped in glacial marine sediments. Evidence of this is found in the path of the adjacent stream. The steep southern side of the mound continues east and becomes the current stream bank. This indicates that the southern side of the mound is a former stream
bank and that the stream may have migrated. Furthermore, the path of the stream south of the mound runs parallel to the road in several places, which implies that the building of the road caused the stream to migrate. This suggests that stream might have previously flowed farther to the west (fig 10 and 11). More evidence of artificial stream migration is found farther upstream where the stream cuts across the road instead of following the contour line indicated in the topographic map (see fig.3). This contour line suggests that without any artificial stream migration, the stream would have moved along the south side of the mound and created a steep bank along some landscape anomaly (Luick and Millette, 2015).

Figure 10: Illustration, interpretation showing sediment layers draped over granite pluton/till deposit irregularity at Lowell Cemetery (modified from Luick and Millette, 2015)

Figure 11: Illustration, interpretation showing adjacent stream creating steep bank at Lowell Cemetery (modified from Luick and Millette, 2015)
The origin of the underlying landscape anomaly is currently unknown. It could be a small granite pluton or a small irregularity in the surrounding granite plutons such as Zion and Blabon Hills. The mound could also be draped glacial till. The use of the mound as a graveyard supports that hypothesis (Luick and Millette, 2015).

16.9 miles From this stop, drive north on the Pope Rd. back to Rte 156.
17.3 miles Go east onto 156 to the Farmington Falls Bridge.
17.4 miles Turn left and go north across the bridge, and then turn immediately east onto Philbrick St.
17.6 miles At the end of Philbrick St., turn right (east) onto Rte 2/27.
21.7 miles Travel to New Sharon and cross the bridge over the Sandy River. Notice this bridge’s height above the water compared to the Farmington Falls Bridge.
22.9 miles Take a right onto Cape Cod Hill Rd. back to Stop 1 to retrieve cars.

REFERENCES CITED


BEDROCK GEOLOGY OF MT. WASHINGTON, PRESIDENTIAL RANGE, NH

By

J. Dykstra Eusden, Brigit Anderson, Carlos Castro, Patrick Gardner, Chris Guiterman,
Stephanie Higgins, Kelley Kugel, Adam Reid, Charles Rodda, and Caitlin Tamposi
All Department of Geology, Bates College, Lewiston, Maine 04240
Email: deusden@bates.edu

INTRODUCTION AND PREVIOUS WORK

This field guide outlines the bedrock geology of the Presidential Range, New Hampshire and describes some of the significant and readily accessible outcrops along the Auto Road of Mt. Washington. Nearly all aspects of the bedrock geology in the Presidential Range are explored. The Paleozoic stratigraphy, deformation, metamorphism, and plutonism is presented first, followed by the history of Mesozoic fractures and basalt dikes, and then finally the Cenozoic exhumation story as told by apatite fission track ages. During the fieldtrip we will see bits of all of these things at each stop. All of the Bates Geology students listed as co-authors for this field guide played an integral part in this work. Each did a full field season of fieldwork followed by a two-semester thesis project on some aspect of the bedrock history. Eusden is extremely grateful to all of them for their friendship and hard work.

The decades-long bedrock mapping campaign culminated in our finished bedrock geologic map (Figure 1) and report (Eusden, 2010). Other publications by our group include Eusden et al. (1996, 2000) and Roden-Tice et al. (2012). Eusden et al. (1996) has maps of several structural fabrics but the bedrock map does not include the lower wooded elevations. Roden-Tice et al. (2012) focuses on the apatite fission track exhumation history of the Presidential Range. In 2013 we published a layperson's book, "The Geology of New Hampshire's White Mountains", which covers many aspects of the geologic history in the Presidential Range (Eusden et al., 2013)

There has been excellent NEIGC fieldtrip representation for the Presidential Range over the past 21 years! An NEIGC field trip similar but not identical to this was first run by Eusden et al. (1996b). Eusden et al. (2006) led another NEIGC trip focusing on the geology from the Bronson Hill Anticlinorium east into the Central Maine Terrane south of U.S. Rte. 2 in Randolph, NH. The most recent NEIGC trip was led by Eusden et al. (2009) and covered aspects of the Ordovician to Carboniferous bedrock geology and apatite fission track cooling history of the West Branch of the Peabody River in Pinkham Notch and again the Randolph, NH valley. Added to this list are the four excellent NEIGC fieldtrips led by Tim Allen on the stratigraphy, structure, and migmatites of the Pinkham Notch and adjacent Carter-Moriah Range (Allen, 1996, 1996b, 2017 and 2017b (this volume)) and a pioneering trip led by Hatch and Wall (1986) who first proposed the Silurian age assignment for much of the geology and made the initial correlation to the Rangeley stratigraphy in Maine.

The Presidential Range bedrock geology was first mapped in any detail by the Billings in the late 1930’s and early 1940’s. Their efforts culminated in several papers and the 1979 Geology of the Mount Washington Quadrangle report published by the New Hampshire Department of Resources and Economic Development (Billings et al., 1979). Hatch and Moench (1984) made many important contributions to the region, most notably the extension and correlation of the Rangeley, Maine, stratigraphy across the border into New Hampshire, through the Presidential Range, and beyond to the south. The New Hampshire bedrock geologic map (Lyons et al., 1997) updated these previous efforts and more or less retained the contacts shown by the Billings’ and the stratigraphic assignments of Hatch and Moench (1984). Tim Allen did his Ph.D. in Pinkham Notch and the difficult terrain of the Carter-Moriah region mapping the stratigraphy, structures, and migmatization (Allen 1996, 1996b, 2017, and 2017b; Allen et al., 2001). Wes Groome and his University of Maine at Orono colleagues made many important contributions towards our understanding of the strain history during porphyroblast growth in the Littleton Formation as well as the development and strain patterns of the migmatites in the Rangeley Formation (Groome et al., 2006 and Groome and Johnson, 2006). Gary Solar and students have also made meaningful contributions to our understanding of the migmatites, melts, diatexites, and deformation in the Pinkham Notch area (Fletch and Solar, 2014; Wisor and Solar, 2014).
IGNEOUS ROCKS

Basalt dikes presumably Triassic or Jurassic in age. Normally 1-2 m in thickness and up to several hundred meters in length.

Trvb - Triassic (?) vent breccia, predominately felsic in composition.

DCrmg - Two mica granite
A medium- to coarse-grained, light gray to white, two mica granite. The granite is generally not foliated but is places zones of foliation exist. Zones of coarse-grained granite, pegmatite, and fine-grained spilitic are also included in this lithology. Principally forms numerous small dikes, veins, and random lenses, to small to map, intruding into geosynclinal metasedimentary rocks. Also occurs as scattered, small, sill-like, plutons up to 2 kilometers in length. Lensos of well-foliated schist and or gneiss are commonly found within these plutons. Plus signs (+) are exposures of granite, pegmatite and/or spilitic.

Drmg - Two mica granite
A medium- to coarse-grained, gray to rusty weathering two mica granite. Schists and phyllites are common. In places a complete gradation with adjacent metasedimentary units is observed. Fine-grained spilitic and pegmatite are common. Principally forms numerous small concordant plutons many of which are too small to map. Plus signs (+) are exposures of granite, pegmatite and/or spilitic.

Dwd - Wamssuta Diorite
A medium-grained, dark gray diorite. Zones of well foliated diorite are common. The one occurrence of this diorite is in the Great Gulf region at Wamssuta Falls.

Explanation of symbols

- Overturn syncline
- Strike and dip of foliation, strike unknown
- Strike and dip of foliation, strike unknown
- Strike and dip of foliation, strike unknown
- Strike and dip of foliation, strike unknown
- Strike and dip of foliation, strike unknown
- Strike and dip of foliation, strike unknown
- Strike and dip of foliation, strike unknown
- Strike and dip of foliation, strike unknown
- Strike and dip of foliation, strike unknown

Figure 1. Bedrock geologic map of Mt. Washington modified from Eusden (2010). Field stops and parking area shown as black circles with white numbers. Explanation of metasedimentary rocks continues on the next page.
METASEDIMENTARY ROCKS

C. Littleton Formation - Dark gray schists commonly with interbedded quartzite layers of varying thickness and abundance. The schists are composed of quartz, muscovite, biotite, plagioclase, chlorite, sericite, illite, montmorillonite, tourmaline, staurolite, garnet, and andalusite. The quartz is generally completely pseudomorphed by muscovite, illite, and sericite, and is common in the schists forming lumps, approximately 1-3 cm in diameter, and elongate aggregates, from 1-5 mm in length, with rare relic cores of fuchsite and stilpnomelane. The schists are well developed as a parallel or slightly parallel bedding. In F1 fold hinges, bedding and foliation become oblique to each other. The quartzites are fine-grained, light gray, granoblastic, and composed of quartz, plagioclase, muscovite, and biotite. Graded beds, reversely in grain size by high grade metamorphism, are common throughout the formation. All contacts between the members of the Littleton are gradational, Littleton Formation Members. Members in the Northern Revolution are listed first followed by correlative members in the Southern Revolution.

D. Spaulding Lake Member - Dlr - Frog Rock Member
Massive schist with coarse, lumpy pseudoschistosity. Rare quartzite beds up to 10 cm thick, are occasionally present. Poorly bedded with very rare graded beds.

E. Alpine Garden Member
Massive quartzite commonly 5 to 8 meter in thickness with rare, 1 to 8 cm thick quartz interbeds, often well graded. Most quartzites show a light gray, brown weathering near the base of the bed. Thickness, 0 to 175 cm.

F. Sphere Dome Member - Dlr - Tuckerman Ravine Member
Well bedded schist and quartzite, the couplet being approximately 10 - 50 cm in thickness and equally divided between the two lithologies. Well developed, aligned pseudoschistosity and graded beds.

G. Mt. Jefferson Member - Dlr - Cow Pasture Member
Massive schist with coarse lumps to well aligned pseudoschistosity, and up to approximately 10% quartzite that are 10 to 30 cm in thickness, in places graded bedding is common.

H. Mt. Madison Member - Dlr - Huntington Ravine Member
A rhythmically bedded schist and quartzite, each couplet 3 - 10 cm thick. Rare garnet coticules are found in the thinner bedded horizons. Aligned pseudoschistosity in the schist with poorly developed bedding. Maximum occurrences of 1 to 1.5 meters thick, well bedded schist and quartzite and massive quartzites.

I. Grand Rapid Path Member - Dlr - Great Gulf Member
A well bedded quartzite and schist, the couplet being approximately 20 - 100 cm in thickness and equally divided between the two lithologies. Well developed, aligned pseudoschistosity and graded beds. Rare occurrences of massive quartzite between 1 and 2 meters in thickness, with 5 to 10 cm interbeds of schist.

J. Edmonds Col Member - Dlr - Bigelow Lawen Member
Massive schist with randomly spaced, thin quartzites approximately 1 to 5 cm in thickness. The quartzites make up approximately 5 to 10% of the unit. Graded bedding is rare. Garnet coticules are occasionally found and pseudoschistosity are less coarse.

K. John Quincy Member - Dlr - Oakes Gulf Member
A well bedded quartzite and schist, the couplet being approximately 10 - 150 cm in thickness and equally divided between the two lithologies. Well developed, aligned pseudoschistosity and graded beds. One 30 cm thick, 1.5 m long quartz pebble conglomerate lens was found in Oakes Gulf.

L. Pine Link Member - Dlr - Abandoned Bridge Member
This member consists of massive, slightly nubby schist with thin quartzite of the schist layers are commonly 1 to 1.5 m in thickness with rare occurrences of 2 to 3 mm thick. The quartzite beds are 1 to 5 cm in thickness. Pseudoschistosity, are common and both graded beds and garnet coticules are rare.

M. Old Jackson Road Member
This member consists of layers of quartzite alternating with thinner layers of schist. The quartzite layers are up to 1 m thickness with rare beds up to 2 m. The schist layers are commonly 15 to 50 cm in thickness. Rare graded beds are found in this member and coticules and pseudoschistosity are not observed.

N. New River Member
This member consists of mostly massive schist with thin quartzite beds. The quartzite beds are commonly 3 to 4 cm in thickness. Graded beds and pseudoschistosity are not observed. Garnet coticules are observed.

O. Mignette - Malignant Littleton Formation
A light gray migmatic gneiss with alternating layers of quartz + feldspar and biotite-rich schist. Layering is planar to slightly wavy, rare, overall 20 to 100 cm long (granulite class I) are found. Very similar to Sec below but without rusty weathering or calc-silicate pods.

---

Sm - Madrid Formation - The Madrid Formation is a fine-grained, thinly laminated, granofels with well-defined alternating layers of dark biotite-rich, schistose granofels and greenish-pale, calc-silicate-rich granofels. The individual layers of granofels are from 1 to 5 cm thick. No graded beds are found. The formations are composed of a dark greenish-gray and consists of actinolite, quartz, biotite, plagioclase, chlorite, staurolite, garnet, and trace amounts of chlorite and epidote. Total thickness of the formation is approximately 10 - 50 meters.

Sfd - Small Falls Formation - The Small Falls Formation is a well foliated schist with distinct red-brown rusty weathering. The formation has a dark gray to black fresh surface and is highly susceptible to weathering. The mineral assemblage includes muscovite, quartz, biotite, plagioclase, chlorite, staurolite, garnet, biotite, plagioclase, chlorite, pyrrhotite, and ilmenite. Quartzite makes up less than 5% of the unit, with layers up to 1 cm in thickness. No graded beds are found. Total thickness of the formation is 10 - 50 meters.

Spn - Perry Mountain Formation - Dark gray schist with interbedded light gray to white quartzites that are commonly 10 - 15 cm in thickness. Quartzites make up 30 to 40% of the unit and can range up to 60 cm in thickness. The mineral assemblage includes quartz, plagioclase, biotite, muscovite, chlorite, sericite, and trace garnet, ilmenite, tourmaline, monazite, zircon, and staurolite. The unit is discontinuous ranging between 0 and 75 in thickness.

Sr - Rangeley Formation - A gray migmatic orthogneiss with abundant calc-silicate lenses. Beds of schist and quartz are sometimes preserved, having eroded migmatization, and in these locations, rare graded bedding is found. The mineralogy of the gneiss is composed of quartz, plagioclase feldspar, biotite, muscovite, chlorite, sericite, staurolite, garnet, ilmenite, and trace monazite, zircon, and tourmaline. The calc-silicate lenses are most often aligned parallel to schistosity planes, but some are at slight angles or, in the extremes, perpendicular to schistosity. Minerals in these lenses include quartz, actinolite, biotite, plagioclase, diopside, biotite, diopside, epidote, garnet, magnetite, and trace monazite, zircon, and ilmenite. Within the gneisses are mappable zones of rusty gneiss, rusty schist, calc-silicate granofels, and amphibolite. The descriptions of the Rangeley Formation members given below focus on the lithologic variations in the gneiss and the mappable subordinate units mentioned above. Stratigraphic order is based on the uninterrupted juxtaposition of the younger Small Falls, Madrid, and Littleton Formations.

Sec - Crawford Member
A migmatic gneiss with alternating layers of white quartz + feldspar and black biotite-rich schist. Layering is planar to slightly wavy. Angular quartz and/or feldspar segregations (class II) between 2 and 8 cm in diameter are common. Elongate, rectangular lenses (class II), 5 to 20 meters in length, of well-bedded calc-silicate granofels and epidote biotite biotite (class II), 10 to 50 cm long, of concentrically mineralized zones of calc-silicate granofels without bedding are common throughout this unit. In places the gneiss is extensively injected by two micapegmatites, plagioclase, and granites.

Sr - Granofels
Within this member are sections of fine-grained, well layered granofels with alternating layers of dark gray biotite + plagioclase schistose granofels and gray to dark purple-green, garnet + actinolite + biotite calc-silicate granofels. Bedding is discontinuous with large blocks that range between approximately 50 and 700 m in length and 10 to 30 meters in width. Lithologically similar to Sec and Sm.

Snc - Rusty Schist
A thin unit composed of platy schist with extreme red-brown rusty weathering. Minerals included are quartz, biotite, plagioclase, muscovite, chlorite, sericite, ilmenite, and trace amounts of monazite, zircon, garnet, pargasite, and pyrrhotite. Bedding is indistinct due to the lack of quartzites. The unit is approximately 5 to 25 meters wide and has only been mapped in small belts, each approximately 500 m long. Lithologically similar to Sfd.

Sve - Eisenhower Member
A moderately well-beded rusty weathering, medium- to coarse-grained, migmatic gneiss with abundant calc-silicate lense. The distinguishing feature between the Crawford and Eisenhower Members is the prominent rusty weathering that the Eisenhower Member exhibits; otherwise these members are identical.

Sve - Granofels
Within the Eisenhower Member is a section of fine-grained, well layered granofels with alternating layers of dark gray biotite + plagioclase schistose granofels and gray to dark purple-green, garnet + actinolite + biotite calc-silicate granofels. Bedding is discontinuous with large blocks that range between approximately 800 m in length and 30 to 100 meters in width. Lithologically similar to Snc and Sm.

Sve - Amphibolites
This is an area within the Eisenhower Member migmatite where rare blocks of amphibolite are found. The blocks are dark brown to black in color, show0 indistinct layering and range in size from 1 to 20 m in length. Minerals include hornblende, quartz, plagioclase, biotite, magnetite, garnet, and trace amounts of monazite, and zircon.

Sr - Rangeley Formation unmetamorphosed
The unmetamorphosed Rangeley weathering rusty gray to brown in color and is composed of layers of dark schist and gray quartzite variable thicknesses. The schist layers are generally well foliated and rusty and usually 30 cm to 1 m thick with rare layers of 1.5 cm. The quartzite layers range in thickness from 2 to 20 cm with rare beds over 30 cm. The average ratio of schist to quartzite bedding thicknesses is 60:40. Graded beds are observed but rare and grading is not as defined as the Littleton schists and quartzites. Small lenses and larger layers of calc-silicate lenses are observed up to 1.5 m in length. Injections of pegmatite are found throughout the entire unmetamorphosed Rangeley Formation and make up approximately 19% to 20% of the outcrops. Also present in this member is a massive quartz pebble conglomerate approximately 38 m in thickness containing 1-2 cm clasts of quartz in a well foliated muscovite, biotite, quartz matrix.

---

Figure 1 continued
PALEOZOIC STRATIGRAPHY, DEFORMATION, METAMORPHISM AND INTRUSIONS

Overview
Here is a brief synopsis of the Paleozoic stratigraphy, deformation, metamorphism and geochronology in the Presidential Range. The stratigraphy (no fossils found yet!) is an on strike extension of the Rangeley, Maine sequence consisting of Silurian and Devonian turbidites and includes 7 members of the Early Silurian Rangeley Formation, the Middle to Late Silurian Perry Mtn., Smalls Falls, and Madrid Formations, and 11 members of the Early Devonian Littleton Formation.

The deformation has seven phases with D₀ pre-metamorphic normal faults, D₁ east vergent, isoclinal nappes, D₂ thrust faulting, D₃-D₄-D₅ folding, and doming of the Oliverian (only D₁ and D₄ are seen nearly everywhere in the Presidents).

The igneous activity consists of the early, syn-kinematic 408 Ma Wamsutta diorite suite, the intermediate, circa 390-400 Ma, Wildcat granite suite, and the youngest suite of Carboniferous (circa 360-350 Ma) two mica granites (e.g. Peabody River stock, Breton Woods, and Bickford granites).

The metamorphism started with the growth of large andalusite grains that define the L₁ lineation. Sillimanite metamorphism (synchronous with the end of D₁) overprinted this and culminated in areas of stromatic migmatite almost exclusively restricted to the Rangeley Formation. The field gradient is in the sillimanite zone (sill+bio+gar+-staur+quartz+musc) and lastly retrograde metamorphism(s) occurred characterized by chlorite and coarse muscovite replacements. Monazite from Littleton and Rangeley formation schists and migmatites, and the syn-metamorphic Bigelow Lawn and Slide Peak Granite gives 207Pb/206U ages of circa 397–405 Ma, suggesting the peak of Acadian metamorphism and intrusion of early two-mica granites occurred then.

Stratigraphy
One of the hallmarks of the stratigraphy of the Presidential Range is the incredible control on topping given by graded bedding that has been reversed in grain size by metamorphism. A quick glance at the bedrock map (Eusden, 2010) shows hundreds of both upright and inverted graded bed strike and dip symbols across the Range (Figure 1). This afforded us excellent control on stratigraphic order and even more so on structural position of D₁ nappe folds (discussed below).

One interesting aspect of the Central Maine Rangeley sequence exposed in the Presidential Range is the discontinuous nature of the Perry Mountain Formation due to non-deposition. In most sections, we see instead a stratigraphy, without breaks, from the Rangeley, Smalls Falls, and Madrid up to the Littleton Formations. Age constraints on the stratigraphy are all by correlation to distant fossiliferous sections.

Bradley and O'Sullivan (2016) recently published a detrital zircon age spectrum for a sample of Littleton Formation from a quartzite at the 4,000 ft. elevation level on the Auto Road. This was part of a regional detrital zircon age study of the Central Maine Belt in New Hampshire and Maine. The detrital zircon data show a Ganderian signature with peaks at 923, 576, 524 and 494, and 388 Ma, with a lesser peak at 1176 Ma (Bradley and O'Sullivan, 2016).

The Littleton Formation in the Great Gulf of the Presidential Range is cross cut by the 408.4 +/- 1.9 Ma Wamsutta diorite pluton (Eusden et al., 2000) part of the syn-tectonic Piscataquis Volcanic Arc (Bradley et al., 2000; Bradley and Tucker, 2002). Given this, the youngest detrital zircon age peak in the Littleton formation should be older than 408 Ma and that is why Bradley and O'Sullivan (2016) attributed the 388-Ma detrital zircon peak to a metamorphic over-print. However, given that peak's error of +/- 11 Ma and that we know from Bradley et al. (2000) and Bradley and Tucker (2002) that Littleton sedimentation was syn-collisional and syn-intrusion with the Piscataquis Volcanic arc, it is possible that deposition continued to as young as circa 400 Ma. and the detrital zircon data is representative of a maximum depositional age, not a metamorphic overprint. Regardless, a somewhat younger than accepted age of Middle to Early Devonian age for the Littleton Formation seems a reasonable conclusion given this new information.

It is interesting to note that three formations from the type Rangeley section in Maine sampled for detrital zircon analysis by Bradley and O'Sullivan (2016) gave maximum youngest age peaks much younger than the traditionally
accepted stratigraphic ages. For example, Part A of the Rangeley Formation may be mid-Silurian rather than Early Silurian, the Perry Mountain Formation may be Early Devonian rather than mid-Silurian, and the Smalls Falls Formation may be Early Devonian rather than Late Silurian (Bradley and O'Sullivan, 2016). This trend toward younger than expected ages warrants a flurry of new mapping linked with detrital zircon geochronology.

D1 Deformation

In terms of Paleozoic ductile structures, regional D1 and D4 folding and localized D5 folding are highlighted on this trip. The D1 folds are characterized by macroscopic east-facing nappes that are well characterized above treeline but less so in the wooded lower elevation regions. The topping direction of graded bedding has revealed a systematic geometry of broad upright limb regions, transitioning into narrow but mappable hinge zones, and in turn into broad inverted limb regions. This pattern coupled with the 1.5 kilometer of vertical relief and abundant rock exposure makes these early nappes some of the best-controlled fold structures in the Appalachians and as such potentially useful in reconstructing plate tectonic geometries. We interpret these east-facing folds to also have been east vergent as no evidence exists to suggest they have been backfolded. One of the problems this introduces is how east vergent early Acadian nappes fit neatly with the overall westward migrating Acadian orogenic front model of Bradley et al. (2000) and Bradley and Tucker (2002). Our favored tectonic model suggests that these D1 nappes verged east over a west, but shallow dipping Acadian subduction zone, probably on Ganderian continental crust in a retroarc setting. The east vergence can be explained as the west migrating orogenic front stalled upon contact with the Silurian "tectonic hinge" (Hatch and Moench, 1984) and Bronson Hill anticlinorium that likely acted as a buttress prohibiting any westward vergence of structures.

The pseudoandalusite porphyroblasts in the schists of the Presidential Range are very well developed and formed during syn-D1 metamorphism. The andalusites are typically 1 to 2 cm in diameter and between 6 and 20 cm long. The andalusite was variably replaced by sillimanite during the peak of metamorphism, when conditions reached the upper sillimanite zone. Further pseudomorphing of this assemblage is characterized by coarse muscovite +/- staurolite during a retrograde event most likely caused by the intrusion of one or more late tectonic granitoids. Late stage chlorite alteration is also occasionally present. When replacement by muscovite is complete, it is common to see relict chisotolite crosses now completely composed of muscovite. Inclusion trails and microstructures have been obliterated by the retrograde metamorphism (Groome and Johnson, 2006), making the meso-scale of observation the best to use when quantifying strain patterns of this lineation fabric.

The porphyroblasts of the L1 lineation lie mostly within the S1 schistosity plane that in places exhibit reverse S1 cleavage refraction with adjacent quartzites (Eusden et al. 1996, 1996b). Groome et al.’s (2006) study of the reverse refraction relationship showed that the stiffening effects of the andalusite porphyroblasts may have rendered the pelite more viscous than the adjacent psammites during foliation development. When aligned, porphyroblasts are also parallel to F1 hinge lines. The hinge-parallel nature of the pseudoandalusite L1 lineations has been extremely helpful in determining the geometries of F1 nappe-scale folds. F1 folds are non-cylindrical, as seen by fold geometries progressively changing across kilometer-scale hinge zone regions. F1 folds with gently plunging hinge lines and moderately inclined axial surfaces change to folds with steeply plunging hinge lines and moderately inclined axial surfaces and finally to reclined folds. Overall, the F1 geometries are probably best classified as sheath-like structures (Alsop & Carreras, 2007).

An interesting aspect of this lineation is its variable degree of alignment (Figure 2). Outcrops where the andalusites are not aligned are often within meters of those with a strongly developed L1 lineation. In other places, the transition from non-aligned to aligned andalusite occurs gradually over hundreds of meters (Eusden et al., 1996). We assume that lineation’s current orientation reflects strain conditions at the time of its development and alignment during D1 - that is, we believe that the andalusite porphyroblasts were variably aligned during D1 deformation and that alignment has not been changed by D2 or by any other subsequent phase of deformation. This is consistent with the general appearance of the schist-quartzite couplets that have well-preserved primary structures and no exposed mylonitic fabrics. Groome et al. (2006) also concluded that the porphyroblasts of the L1 lineation had not been rotated after D1.
Figure 2. Stereonets of L, lineations (dots), S, foliation (great circle), and mean angular separation value (θ in degrees) at individual outcrops along Chandler Ridge. Upper center inset shows the correlation between θ and percentage of quartzite in Littleton members; lower θ (better aligned) in units with lowest % quartzite. Inset in lower right show synoptic stereonet of contoured L, with average S, foliation (dashed great circle).
We've studied the variable development of the L₁ lineation throughout the Presidential Range (Guiterman and Eusden, 2004; Rodda and Eusden, 2005; Higgins and Eusden, 2008). In doing so we developed a quantitative metric (mean angular separation or "Θ" method) to measure the degree of lineation development. The mean angular separation method calculates the average acute angle between measured trends and plunges of a group of linear features and was developed by Bates senior thesis student Stephanie Higgins (2008). In this method, values of Θ lie between 0° at perfectly aligned sites, and <90° at randomly aligned sites. Values for mean angular separation ranged from a minimum of 0° near Mt. Washington, the most highly aligned site, to a maximum of 53° near Mt. Madison, the site with the lowest porphyroblast alignment. There was no statistical difference between andalusite lineation development or Θ on the upright limbs, inverted limbs, or the hinge zone regions of the D₁ folds. There was a slight but not statistically significant increase in lineation development in the inverted limbs. This conforms to expectations, with slightly more alignment on inverted limbs that may have experienced greater shear strain. Regardless, structural position appears to not be the primary control on lineation development, or, if it was to a certain degree (e.g. on inverted limbs) that was greatly overprinted by another force.

There is a statistically significant correlation however between the proportion of quartzite at any outcrop and the degree of porphyroblast alignment at that site. In short, the more quartzite that surrounded a schist bed, the less aligned its porphyroblasts were. The data also show a clear connection between bed thickness and porphyroblast alignment: thicker beds yielded poorer alignment. We also examined the alignment variability on a larger scale than outcrop by studying the overall development of the Littleton formation units: Edmond’s Col Member (Dlec); Mt. Jefferson Member (Dljm); Mt. Madison Member (Dlmm); and the Israel Ridge Path Member (Dlirp). There is a perfect positive correlation between percentage quartzite in the four bedrock members and porphyroblast non-alignment (high angles of Θ) in that member. Dlec (10% quartzite) shows the strongest alignment, with Dlirp (50% quartzite) showing the poorest alignment (Figure 2). Thus, the macroscale rheological variations of the different Littleton Formation stratigraphic members also reveal clear control of the lineation’s development. The positive correlation between the development of this mineral lineation and the sedimentary features defined by the turbidite bedding in the Littleton Formation continues to argue for linking detailed mapping with any structural analysis.

D₄ Deformation

D₄ deformation is the most common and abundant phase of folding seen in the Presidential Range. F₄ folds are characterized as mesoscopic, moderately inclined to overturned, generally gently but often moderately plunging, asymmetric F₄ folds. The F₄ folds have no, or only a weakly developed, axial planar S₄ cleavage. F₄ folding is not uniformly distributed throughout the study area. All areas show some F₄ folding, but in places it is distinctly more pervasive than in others. Mesoscale folds have double amplitudes ranging from 10 cm up to several meters and wavelengths from 10 cm to 10 meters. Interim angles classify all mesoscale folds as open. Axial surfaces generally strike N and dip moderately to steeply W. Hinge lines trend N or S with shallow plunges. The typical shape of an F₄ fold consists of long, moderately dipping west limbs and short, steeply dipping east limbs. In some instances, the east limbs are overturned. This asymmetry gives a west over east sense of rotation, which we interpret to be related to east tectonic vergence of these folds. Pegmatite and aplite veins, dikes and sills seen throughout the alpine zone are folded by F₄. This deformation was syn-metamorphic but occurred after the peak of metamorphism in the region, likely around 380-365 Ma or in the late Acadian to Neoacadian timeframe.

Shortening calculations of well-exposed D₄ fold trains in the Presidential Range of New Hampshire were done at the mesoscale to quantify the strain and to evaluate the strain partitioning (Figure 3). Transects along Osgood Ridge on Mt. Madison, Chandler Ridge on Mt. Washington, and Caps Ridge on Mt. Jefferson, all containing well-exposed D₁ nappes refolded by D₄ fold trains were examined (Kugel and Eusden, 2004; Rodda and Eusden 2005; Reid and Eusden, 2005; Tamposi and Eusden, 2008). Shortening was calculated using the equation e = (l₀-lₚ)/lₚ * 100, where lₚ = the hinge to hinge straight length of the fold train and l₀ = the length of the folded layer.

For Chandler Ridge, meso-scale fold train shortening varied systematically over a macro-scale D₄ Dome, ranging from 1.1% on the flank to 47.8% on the apex with a mean value of 14.2%. Osgood Ridge showed mesoscale shortening ranging from 5.9% and 32.3% on the limbs and crests respectively of macro-scale D₁ anticlinal synforms and synformal anticlines with a mean value of 16.4%. Caps Ridge mesoscale shortening ranged from 4% to 34% with a mean value of 12.4%. Collectively, these results suggest that mesoscale D₄ shortening is on average about 14% across the Presidential Range and this strain is preferentially partitioned on the crests and troughs of
Figure 3. Stereonets of $S_n$ cleavage (great circle) and $F_n$ hinge lines (red dots) at outcrops where shortening was measured on $D_n$ fold trains. Mean mesoscopic shortening (in %) by structural domain shown in red. Overall mean macroscale shortening on Chandler Ridge (entire map area shown) was 14.2%. Inset photo shows an example of a fold train on Cragway Corner of the Auto Road. Other inset shows overall shortening of 28.7% when micro-, macro- and mesoscale shortening calculations are cumulatively multiplied.
macroscopic folds by parasitic mesoscale folding. The total D₄ shortening (micro+macro+meso) is approximately 30%. Depending on the duration of the D₄ event (10 my - 1 my), shortening or convergence rates would range from .1cm/yr to .01 cm/yr during the Late Acadian and Neoacadian orogenies.

**D₃ Deformation and Doming**

Uplift and doming of the Ordovician Oliverian Jefferson Dome was restricted to the Randolph valley region of the Presidential Range. A region of east-southeast dips occurs in a two-mile wide zone immediately southeast of the contact between the dome and cover rocks. Beyond this zone the regional west dip prevails. We are unsure of the timing relationship between this doming and other late Acadian folding events (e.g. D₄ folding) but suspect that doming developed prior to the intrusion of the several Late Carboniferous plutons in the region. Regardless, it is likely Neoacadian and deforms all other fabrics in the rock except, of course, brittle structures created during Mesozoic rifting.

D₃ folding is a Neoacadian event, restricted to the Pinkham Notch area, and manifest as microscale crenulations and macroscopic folds that trend E-W and plunge quite shallow. D₃ is probably related to the Early Carboniferous intrusion of the 355 Ma Peabody River Stock granite because D₃ folds are found in proximity to the contact and die out away from it.

**MESOZOIC FRACTURES AND BASALTS**

Mesozoic extensional structures were mapped in Great Gulf, Tuckerman Ravine, and Huntington Ravine in the Presidential Range, NH, to determine their relative ages, distribution, and paleostress fields (Castro and Eusden, 2010; Gardner and Eusden, 2010; Kindley and Eusden, 2011; Eusden et al., 2011). Over 3000 joints and 7 major dikes were measured within the folded Devonian Littleton Formation schist and quartzite and massive migmatites of the Silurian Rangeley Formation (Figure 4).

Two dikes were found in Tuckerman Ravine, three in Huntington Ravine (Pinnacle Gully, Diagonal Gully, and Escape Hatch dikes), and three in Great Gulf (Pipeline Gully, Airplane Gully, and Oblique Gully dikes). All dikes are 1-3 meters wide, extend for several 100's of meters, show composite dike textures, are often vesicular, are not significantly influenced by bedding plane anisotropies, and mostly classified as alkaline dolerites. The dikes all exhibit preferential erosion as compared to the schist country rock. As a result the dikes are found in the base of topographically prominent, narrow, steep walled gullies in the ravines.

Only common joints and few if any shear or conjugate joints were observed suggesting near surface conditions (<3 km?) under low lithostatic pressure. Four joint sets were identified and assigned relative ages based on field relations. From oldest to youngest the sets are: 1) NE striking vertical set; 2) E striking vertical set; 3) NNW striking vertical set; and 4) sub-horizontal to west dipping sheeted set. Key cross cutting relations were: 1) Pinnacle Gully where the Pinnacle NE striking dike is cut by the E-W joints; and 2) the south face of Mt Clay where the NNW joint set terminates in a "T" intersection against open E-W joints.

The oldest and most abundant joints have a NE strike and sub-vertical dip, are most abundant near and sub-parallel to the principal NE striking dikes in each ravine (the Tuckerman Ravine, Pinnacle, Pipeline, Airplane, and Oblique Gully dikes), and likely formed during regional NW-SE extension associated with the Late Triassic and Jurassic rifting of Pangea (Faure et al. 2006). The second oldest joint set has a E-W strike, is sub-parallel with the Diagonal Gully and Escape Hatch dikes in Huntington Ravine, and interpreted to be part of regional N-S extension associated with the Middle Cretaceous New England-Quebec province (McHone and Butler, 1984; Faure et al., 1996a and 1996b). The third youngest fracture set strikes NNW -SSE, has no correlative basalt dikes, and may correlate to the regional trend of the White Mountain Magma Series (McHone and Butler, 1984). The youngest joint set has shallow W-NW dipping sheeted joints thought to have formed from Quaternary glacial and Late Cretaceous tectonic unloading.
Figure 4. A. Stereonets of joints and dikes in the Presidential Range, arrows show extension direction; Rose plots of joints in Great Gulf (B), Tuckerman (C) and Huntington Ravine (D). E. Mt Clay joints color coded by set; steeply dipping sets from oldest to youngest: yellow NE, blue E, and purple N; and green sheeted set.
CRETACEOUS EXHUMATION

Apatite fission track (AFT) ages in samples collected along the 5000 foot relief (1500 m) exposed at Mt. Washington in the Presidential Range of New Hampshire have been used to constrain the Cretaceous cooling history of the area (Roden-Tice et al., 2011; Anderson et al., 2012). Nine AFT ages along the Mt Washington Auto Road and thirteen AFT ages along the Cog Railroad were collected for samples of the Littleton and Rangeley formations (Figure 5).

AFT ages range in from ~150 Ma at the highest elevations (~1900 m; ~6000 ft) to ~90 Ma at the base (~500 m; ~1,500 ft). These values yield an exhumation rate of 0.024 mm/yr between approximately 150 Ma and 80 Ma for both the Cog Railroad and Auto Road transects.

A 2D cross sectional view of the AFT age surfaces (surfaces at 140, 130, 120, 110, 100, and 90 Ma.) shows that the 140-120 Ma surfaces are tilted easterly between 3.2° and 10° with a mean of 6.8°. The 110-90 Ma surfaces are sub-horizontal with easterly dips between .5° and 2.9° with a mean of 1.8°. We suggest that the more tilted older surfaces were exhumed in a syntectonic setting and perhaps tilted by active faults in the area such as the local Ammonoosuc or Glen Ellis Faults, or perhaps even the distant Norumbega Fault. The sub-horizontal younger surfaces were exhumed during a period of post-tectonic erosion and were not deformed in any significant manner. One possibility is erosion due to incipient stream incision of the ancestral Androscoggin, Saco, and Connecticut River drainages. AFT track length modeling discussed below seems to support this hypothesis.

AFT track lengths along the two transects were nearly identical with longer tracks at the base of the mountain and shorter tracks at the summit. AFT track modeling suggests that there was rapid cooling (to produce the longer tracks) through the apatite partial annealing zone of 60° to 90°C at the base of the mountain during the time period from circa 125-60 Ma. This suggests that rapid exhumation occurred in the valleys of Pinkham and Crawford Notches during this time. Conversely the slower cooling through the apatite partial annealing zone (producing shorter tracks) at the summit suggests much slower exhumation for the same time period.

These heterogeneous trends in exhumation between the summit and the lower elevation notches may be due to late Cretaceous regional magmatic events associated with local asthenospheric upwelling (Matton and Jebrak, 2009) that reactivated zones of crustal weakness and possibly triggered the initiation of the ancestral drainage systems in the Presidential range. This would lead to rapid incision during the time period 125-60 Ma. in the notches while the summit eroded much more slowly.

From 60 Ma. to the present the AFT track modeling suggests a uniform exhumation rate across the topographic relief of the Presidential Range. We conclude that the most likely explanation for the common exhumation history across >1 km of relief is that the relief was established by the end of the Cretaceous and has persisted with steady-state topography through the Tertiary to the present (Roden-Tice et al., 2012).

The AFT results are consistent with an earlier relief method study employing ⁴⁰Ar/³⁹Ar muscovite cooling ages (Eusden and Lux, 1994). Geothermal gradients calculated from the results of both studies yield ~ 40 °C/km suggesting that this gradient persisted throughout Permian and Mesozoic times.
Figure 5. A. Apatite fission track ages (in Ma) on Mt. Washington. B. Proposed erosion and resulting topography over time. C. Track length modeling showing rapid exhumation in notches from 125-60 Ma. D. Exhumation rates calculated using relief method. E. Cross section of AFT surfaces through Mt. Washington. 140-120 surfaces are tilted by faulting and horizontal 110-90 surfaces are post-tectonic.
DRIVING AND WALKING LOG FOR STOPS

Time, Place, Logistics
Saturday September 30th, 7:30 AM in the gravel parking area on the west side of NH 16 to the immediate south of the Auto Road entrance (322363.00 m E, 4906302.00 m N). The base of the Auto Road is about 30 miles west of Bethel, ME and takes about 40 minutes to drive. From Bethel follow Rte. U.S. 2 west to Gorham, NH then take Rte. NH 16 south to the Auto Road entrance. Due to the fragile nature of the alpine ecosystem, please always walk on trails or rocks. Warning: Expect extremely cold and unpredictable weather. Be prepared with proper clothing and good hiking boots for very rocky, uneven terrain, winter–like conditions and extremely high winds. Bring your lunch and water as there will be absolutely no chance to pick up anything once the trip begins. There are no bathrooms facilities. Vehicles must be consolidated. The folks at the Mt. Washington Auto Road have kindly agreed to charge us only $29 per vehicle and waive the passenger fee. Half-ton vans are permitted up the road but with a maximum passenger and luggage weight of 900 lbs (that's approximately 6 people, including driver). If you are planning on driving your vehicle, please carefully check the vehicle restrictions on the Auto Road web site that can be found here.

Mileage
0.0 Base Station. Drive from the base of the Auto Road to the summit of Mt. Washington
8.0 Lower parking lot (elev. 6,200 ft.) of Mt Washington Summit. Park vehicles and walk to STOPS 1A-B

STOP 1A. In blasted outcrop along the NW side of the road just below parking lot, inverted graded bedding of Huntington Ravine member of Littleton Formation within inverted limb of the Mt. Washington Nappe, small sills of pegmatite as well. AFT ages in the summit region are between 147 and 151 Ma.

Walk down the road about .1 mile just below the parking lot with some oil tanks on the west side of the road at about elevation 6,100 ft.

STOP 1B. On the east side of (downhill side) the road are blocks and outcrops of schists and thin quartzites of the Huntington Ravine member of the Littleton Formation with excellent L1 pseudoandalusite lineations. Where aligned, the L1 orientation is parallel to the F1 fold axes throughout the Presidential Range. Mean angular separation or "\(\Theta\)" values for the andalusite lineation in this region are 15° to 25°, meaning strong alignment.

Return to vehicles and drive down the Auto Road.
8.6 Park vehicles in lot on west side of road which is just above the 6,000 ft elevation post. STOP 2 is a 2 mile walking loop to several outcrops.

STOP 2A. Walk .1mi. east and slightly uphill to Ball Crag (elev. 6,112 ft.) and examine inverted graded beds of the Huntington Ravine member of the Littleton Formation with well developed F4 folds and poorly developed S4 axial plane cleavage.

Walk back to the parking lot with the vehicles and walk .1mi west to the intersection of the Great Gulf and Gulfside Trails. Follow the Gulfside Trail WNW paralleling the Cog Railroad, passing the junction of the Westside Trail, to the Clay Col and junction of the Gulfside and Clay Loop trails, about .9 mi.

STOP 2B. In the Col the geology passes from schists of the Bigelow Lawn member of the Littleton Formation to the SE to gneisses and abundant pegmatites of the Mt. Clay member of the Rangeley Formation to the NW. The Madrid, Smalls Falls, and Perry Mountain Formations are cut out here. The contact between the Rangeley and Littleton is not exposed, but you can get within about 20 meters of it; look for the last outcrop of schist and the first outcrop of gneiss with calc-silicate lenses. This contact we have interpreted as the D2 Greenough Spring Thrust Fault. The NE striking Airplane gully dike is exposed in the steep gully along the headwall of Great Gulf and blocks of basalt can be seen in the saddle.

Proceed N along and up the Clay Loop Trail for about .1 mi.
STOP 2C. In the trail and just to the W on the first pitch up from the Col, are outcrops of the Mt. Clay member of the Rangeley Formation gneisses with abundant quartz segregations (clasts?) and rare, oval, concentrically zoned, calc-silicate lenses (clasts?). Proceed up the first steep pitch on the Clay Loop Trail another .05 mi., which is halfway up the first or S summit of Clay. On the E side of the trail are exposed meter-long, bedded clasts of calc-silicate with bedding truncated by the gneissic matrix. We have interpreted this type of Rangeley to be a metamorphosed olistostromal mélangé. This steep pitch also shows excellent fractures and all four joint sets (NE, E, N and sheeted) are well exposed. In a few places the N set exhibits "T" with the E and NE sets, demonstrating that the N joints are the youngest steeply dipping set.

Return to vehicles along Clay Loop and Gulfside Trails, about 1 mi., and drive down the Auto Road.

9.5 Park at the lot on the NW side of the road at elevation 5,700 ft.; the relatively flat area of the so-called "Cow Pasture." STOP 3 is a 1 mile walking loop to several outcrops

Walk to the junction of the Nelson Crag and Huntington Ravine Trails and proceed up the Nelson Crag Trail about .1 mi to about elevation 5,800 ft.; this would be between the 14th and 15th cairns on the Nelson Crag Trail from the junction. Turn off the trail to the south, paralleling the slope and contours for about .1 mi to prominent outcrops marked by a small cairn.

STOP 3A. F1 hinge zone exposures of the Tuckerman Ravine Syncline with thick bedded quartzites and thin schists of the Alpine Garden member of the Littleton Formation. S0, bedding, and S1, axial planar schistosity, are perpendicular here. Graded beds in the hinge zone suggest northeasterly facing directions for the F1 folds.

"Rockwack" downhill, southeast, for about .2 mi to the headwall of Huntington Ravine and the junction of the Huntington Ravine and Alpine Garden Trails. Please only walk on rocks and keep off the vegetation!

STOP 3B. Pinnacle Gully and Diagonal dikes intruding upright schists and thick micaceous quartzites of the Huntington Ravine member of the Littleton Formation. The Pinnacle dike belongs to the NE striking earliest joint/dike set while the Diagonal dike is part of the E striking joint set. E joints cut the Pinnacle dike suggesting it is younger.

Follow the Alpine Garden trail .2 mi northeast to junction with the Nelson Crag Trail.

STOP 3C. Exposed is an outcrop of a F4 warping fold about 15 meters southeast of the big cairn at the junction. This fold happens to be along the crest of the most significant macroscopic F4 structure, the Chandler Ridge Dome. The F4 folds exhibit maximum shortening of 47.8% on the crest of the Dome and D4 folds are the most geometrically diverse in this region.

Follow the Nelson Crag trail about .1 mi. uphill to the unnamed crag at elevation 5,735'.

STOP 3D. On the north and northeast flanks of this crag are upright beds of the Cow Pasture member of the Littleton Formation. S0 and S1 are now parallel here as the F1 hinge zone of the Tuckerman Ravine syncline is slightly above us.

Proceed .1 mi. to the junction with the Huntington Ravine Trail and then back to the parking lot with the vehicles. Drive down the Auto Road.

10.7 Park in the vicinity of Cragway Spring (elev. 4,800 ft.) in the small parking lots above and just at the hairpin turn. STOP 4 is a 1.5 mile walking loop to several locations.

Hike down the Nelson Crag Trail .6 km to treeline at approximately 4,300 ft. elev.

STOP 4A. Outcrops of the Great Gulf member of the Littleton Formation. Exposed here are schists with 10 cm thick quartzite interbeds. Bedding, S0, and schistosity, S1, are not parallel indicating that we are in the F1 hinge zone of the Horn Nappe. S1 schistosity is refracted through the inverted bedding defined by the quartzites. Fresh, pink andalusite and andalusite rimmed by fibrolitic sillimanite is common. There is a nice float block with a complete F1
fold exposed. There are good glacial striations on a quartz vein outcrop that also shows some unusual staurolite-bearing pegmatites.

Proceed back up the Nelson Crag Trail about .2 mi through the first patch of scrub spruce and into the next higher treeless area. Head off the trail to the south about .05 mi to the prominent outcrops.

STOP 4B. Exposed here are the predominately massive schists of the Cow Pasture member of the Littleton. L₁ lineations are folded by abundant F₄ folds with circa 11% macroscopic shortening which define the topography here. Bedding and S₁ schistosity are again parallel as we are now above the Horn Nappe F₁ hinge zone.

Proceed back up the Nelson Crag Trail about .1mi through another patch of scrub spruce and to the higher treeless area that includes the Auto Road at Cragway Spring. Head off the trail to the south about .05 mi.

STOP 4C. Exposed are schists and thin quartzites of the Huntington Ravine member. Well developed F₄ structures fold thin aplite sills here. These small granitic apophyses are the earliest phase of granitic intrusion and are pre-F₄ in age. Some of the larger granite plutons of this generation impart local staurolite grade metamorphism to the schists. Mean angular separation or $\Theta^\prime$ values for the andalusite lineation in this region are highly variable over short distances and range between 22° (well aligned) to 48° (poorly aligned)

Return to the vehicles examine the road cuts and outcrops on the "inside" of the hairpin turn.

STOP 4D. Exposed are schists and thin quartzites of the Huntington Ravine member. Upright graded beds are common. Pseudoandalusite is replaced by fibrolitic sillimanite in the core with coarse-grained muscovite and staurolite in the rim. A great place to see this texture is a 1m diameter lichen-free zone of bed rock marked by a small patch of concrete (4 cm diam) with a small metal pin (1 cm diam) embedded in it. F₄ folds are everywhere and as with all locations in STOP 4 are consistently oriented.

Return to vehicles and drive down the Auto Road.

STOP 5A. Outcrops of thick schists and 10-20 cm thick quartzites of the Great Gulf member of the Littleton. These outcrops show the exact hinge location for the Horn Nappe as bedding and schistosity are perpendicular. S₁ is again refracted through the quartzites. Mean angular separation or $\Theta^\prime$ values for the andalusite lineation in this region are around 45° meaning poorly aligned. We think this is due to the thick quartzites of the Great Gulf member.

Walk about .2 mi up the ridge between the Winter Cutoff and Auto road to about elevation 4,450 and then bushwhack down to the Auto Road. Walk down the road about .1 mi.

STOP 5B. Nice exposure of a rare F₁ mesoscopic fold hinge in the outcrops on the west side of the road. The fold exhibits reverse refraction where the steeper angle is in the schist, made more viscous by the growth of andalusite (Groom and Johnson, 2006; Groome et al., 2006). AFT ages range from 137 to 123 Ma in this vicinity.

STOP 5C. Garnet-bearing granite sill exposed in a road outcrop. We've tried to get zircon and monazite out of this for an age date but no luck. An age determination would nail down the timing of F₄ folding and associated contact metamorphism in the Range.

Return to vehicles and drive down the Auto Road.

STOP 5. Park in the lot at elevation 4,200 ft. just below the junction with the Winter Cutoff Road. STOP 5 is a short walking loop of about 1 mile.

Walk back up the Auto Road to the Winter Cutoff road and proceed up that about .1 mi to elev. 4,400 ft.

STOP 5A. Outcrops of thick schists and 10-20 cm thick quartzites of the Great Gulf member of the Littleton. These outcrops show the exact hinge location for the Horn Nappe as bedding and schistosity are perpendicular. S₁ is again refracted through the quartzites. Mean angular separation or $\Theta^\prime$ values for the andalusite lineation in this region are around 45° meaning poorly aligned. We think this is due to the thick quartzites of the Great Gulf member.

Walk about .2 mi up the ridge between the Winter Cutoff and Auto road to about elevation 4,450 and then bushwhack down to the Auto Road. Walk down the road about .1 mi.

STOP 5B. Nice exposure of a rare F₁ mesoscopic fold hinge in the outcrops on the west side of the road. The fold exhibits reverse refraction where the steeper angle is in the schist, made more viscous by the growth of andalusite (Groom and Johnson, 2006; Groome et al., 2006). AFT ages range from 137 to 123 Ma in this vicinity.

STOP 5C. Garnet-bearing granite sill exposed in a road outcrop. We've tried to get zircon and monazite out of this for an age date but no luck. An age determination would nail down the timing of F₄ folding and associated contact metamorphism in the Range.

Return to vehicles and drive down the Auto Road.

STOP 5. Park in the lot on the east side of the road just above the 4,000 ft. elevation post.

Walk down the road about .2 mi. to just below the 4-mile post.
STOP 6. Examine the large outcrop of thick quartzites of the Oakes Gulf member of the Littleton. There is one enormous bed, 5 meters in thickness, which is wonderfully graded and inverted. Bedding and schistosity are not parallel, meaning we are still within an F1 hinge zone. This outcrop was where Bradley and O'Sullivan reported a 388 Ma youngest zircon peak age, attributing that to a metamorphic age.

Return to vehicles and drive down the Auto Road

13.8 Park in the lot where the Madison Gulf Trail crosses the road.

Walk .5 miles up the trail to the junction with the Lowe's Bald Spot cutoff and hike up the trail to the top.

STOP 7. Lowe's Baldspot. Beginning from the trail junction and continuing to the top you will see the schists with thinly beds of quartzite or just massive schist outcrops of the Bigelow Lawn Member of the Littleton Formation. One of its characteristics is the presence of garnet +/- quartz coticule. Structurally there are two sets of crenulations one N-S trending and the other E-W trending. These are D4 and D5 folds respectively. D5 spatially surrounds the Peabody River two mica granite intrusion which has been dated to be 355 Ma.

Return to vehicles and drive down the Auto Road .

14.3 Park in tight parking lot along the road.

Walk across the road to the south side and bushwhack .1 miles down a little drainage to outcrops of Silurian Smalls Falls Formation.

STOP 8. The Smalls Formation well bedded, rusty schists are exposed in the stream bed. Above this outcrop is found the Abandoned Bridge member of the Littleton Formation with no intervening Madrid Formation. We interpret this stratigraphic gap to have been caused by a pre-metamorphic fault and have named it the Pinkham Notch Fault. AFT ages range from 99 to 101 Ma in this vicinity. Proceed back to the road and examine the outcrops of Triassic (?) vent breccia. Besides the basalt dikes we saw earlier this is the only other mappable Mesozoic unit in the Presidential Range. A separate section of this unit can be seen at Crystal cascade on the Tuckerman Ravine trail.

Return to vehicles and proceed down the bottom of the Auto Road

16 Parking lot that we started from.

End of Trip

REFERENCES


Groome, W. and Johnson S., 2006, Constraining the relative strengths of high grade metamorphic rocks: an example from the northern New England Appalachians: Journal of Structural Geology.


INTRODUCTION

Purpose

The East Andover quadrangle was selected for surficial mapping during the 2016 field season as part of the Maine Geological Survey’s STATEMAP program. The area was coarsely cataloged by Stone (1899) and Leavitt and Perkins (1935), but the surficial geology had not been officially mapped at a scale finer than the state level of 1:500,000 (Thompson and Borns, 1985), and thus became a priority for the Survey. Bedrock geology of the area was mapped at the 1:62,500 scale (Rumford quadrangle) by Moench and Hildreth (1976). The recent surficial mapping efforts (Locke and others, 2017; Weddle and others, 2017) provided a reminder of how much the area has to offer the visiting geologist, with everything from pegmatites and a rich mining history to eskers and stream terraces. This field trip will provide participants with the opportunity to view a subsample of the geologic features in the East Andover quadrangle and to consider their formative processes. Figure 1 offers an overview of the trip route with stop locations. We will begin in the southeast part of the quadrangle, work our way up the Ellis River Valley, then loop back down to the central portion of the quadrangle to end the day at Rumford Whitecap Mountain.

Physiography Overview

On a broad scale, the East Andover quadrangle lies within the Central Highlands of New England, which contains a wide range of topography and the highest peaks in the region (Denny, 1982). Hanson and Caldwell (1989) focused in on Maine and refined Denny’s classification, but this did not significantly change the description of the East Andover area, which was placed in the Central Maine Highlands. This geomorphic region is comprised of several mountain ranges, including the Mahoosuc Range and Blue Mountains, which lie to the west/southwest and north/northeast of the quadrangle, respectively. The foothills of these ranges and lesser peaks make up the topography within the quadrangle and to the south/southeast. Many locals simply refer to this region as the Western Maine Mountains. Tributaries that eventually form the West Branch of the Ellis River flow from the Dunn Notch and Sawyer Notch areas to the west and northwest of the quadrangle. Tributaries of the East Branch of the Ellis River make their way from the Black Brook Notch, Little Ellis Pond, and Ellis (a.k.a Roxbury) Pond areas to join the West Branch just south of Andover village. The area to the north and northeast of this confluence is a relatively large and flat basin formed in the Mooselookmeguntic Pluton. Hanson and Caldwell (1989) attributed broad lowlands that occur within the Central Maine Highlands to underlying plutonic rocks, which are (in most cases) less resistant to weathering and glacial erosion than surrounding metamorphic rocks. The Ellis River flows south from the basin, entering a deep valley that is bounded by Plumbago Mountain and Mount Dimmock to the west and Rumford Whitecap Mountain to the east, eventually reaching the Androscoggin River at Rumford Point.

BEDROCK GEOLOGY

The bedrock in the East Andover quadrangle comprises layered clastic sedimentary rock of Devonian and Silurian age that was later altered by metamorphism to develop aluminous schists, granofels, and calcareous units. These units include the biotite granofels of the Devonian Hildreths Formation; the quartzose biotite granofels, biotite schist, and pale calc-silicate rocks of the Madrid Formation; the rusty-weathering graphitic schist and sulfidic granofels of the Smalls Falls Formation; the cyclically-bedded muscovite-rich schist and granofels of the Perry Mountain Formation; and the lead-gray irregularly-interbedded mica schist and quartzose biotite granofels of the Rangeley Formation (Moench and Hildreth, 1976). These units have experienced at least three distinct metamorphic and deformational events (Guidotti, 1970; Guidotti, 1989), the first two associated with regional folding and faulting...
Figure 1. East Andover quadrangle location map. Trip stop locations are labeled with dots and bold numbers. Rumford Whitecap Mountain hiking trails are marked with yellow and red lines. Locations of areas shown in guide figures, but are not stops on this trip, are outlined with dashed boxes and labeled with figure numbers. Cartography by Amber Whittaker.
that imparted foliation with overall near-vertical dip and northeasterly strike, and the third likely associated with the emplacement of several plutonic bodies that are aligned in a northwesterly direction, perpendicular to the regional structures. The northern third of the quadrangle is dominated by the Devonian-age Mooselookmeguntic Pluton, which is a composite igneous complex dominated by two-mica granite with smaller amounts of granodiorite, tonalite, and quartz diorite (Moench and Hildreth, 1976; Tomascak and others, 2005). The string of mountains that cross from west to east in the center of the quadrangle (Plumbago, Rumford Whitecap, and Black Mountains) are characterized by igneous rocks. Plumbago Mountain consists of mafic to ultramafic metamorphosed intrusive igneous rocks cut by beryl-bearing and rare-mineral pegmatite dikes that host the Newry Mines. Rumford Whitecap Mountain, the final stop on this field trip, is named after the white granitic pegmatite and aplite exposed at the top. Black Mountain, although predominantly metasedimentary, is intruded by several pegmatite dikes, some of which have been quarried. Jurassic/Triassic basaltic dikes are scattered throughout the quadrangle and cut all units.

Several economic mineral deposits have been worked in the pegmatites of the East Andover quadrangle. Pegmatites in Maine have been prospected for gem minerals like tourmaline as well as materials used in the manufacture of war equipment such as: sheet and flake mica for use in heat-resistant windows or in capacitors for its dielectric properties; feldspar for use as a flux in ceramic and glass production; beryl for creating beryllium-metal alloys; lithium-bearing minerals such as lepidolite or spodumene for use in ceramics, glass, and metal alloys; and cesium-bearing minerals like pollucite for use as a getter in vacuum tubes (cesium is liquid at room temperature and easily combines with gases like oxygen). In the early 1900s, the most well-known quarries at the Newry Mines on Plumbago Mountain in the western part of the quadrangle produced gem-quality mineral specimens, as well as high-quality feldspar and spodumene with mica and beryl byproducts (Cameron and others, 1954; King, 2000). Dunton Quarry (part of the Newry Mines) yielded gem-quality colored tourmalines in the 1970s. The area is occasionally open for collecting (Thompson, 2013). Colored tourmaline and purple lepidolite were found at Black Mountain in the late 1800s, although there are few reports of gem-quality specimens. In the early 1900s, the quarry was worked for scrap mica, beryl, potash and soda feldspar, and spodumene. Mining ended sometime before 1950 (Cameron and others, 1954; King, 2000). Five pegmatites at the Red Hill quarries in the eastern part of the quadrangle were explored for beryl in the late 1940s (Shainin, 1949).

The pegmatite at Rumford Whitecap was explored for beryl and feldspar based on the success of the deposits at Plumbago and Black Mountains. In 1961, the Whitecap Mountain Syndicate drilled two exploration holes at the summit with a combined length of 994 feet under the direction of Moyd (1961). Moyd’s report indicates that the exploration cores were retained by the mining consultant Robert W. Bridgeman, and that the boreholes were marked with brass-sheathed pipes. Neither the cores nor the borehole locations could be found by MGS staff during preparation for this field trip. Core logs are available in the 1961 report. The cores were sampled by Moench and Zartman for whole-rock rubidium-strontium dating (Moench and Zartman, 1976). Moyd observed that the Rumford Whitecap pegmatite was not concentrically zoned and lacked any of the rare pegmatite minerals that were sought at the time, such as beryl or colored tourmaline; it was also determined that the deposit could not be economically mined for feldspar. The body is primarily granitic pegmatite and aplite, rhythmically layered, with blocks of the surrounding metasedimentary rocks and granite included in the main mass. Mineralogically, the pegmatite contains large crystals of quartz, feldspar, and muscovite, with accessory black tourmaline and garnet. Feldspar in the pegmatite is 40% potassium feldspar and 60% sodium-rich plagioclase (Moyd, 1961).

SURFICIAL GEOLOGY

Previous Observations and Research

Stone (1899) made his way up the Ellis River Valley in the late 1800s as part of his efforts to catalog the glacial deposits of Maine, and even though he was often without accurate maps, his observations were very insightful. He noted the broad Andover basin containing “sedimentary plains of gravel, sand, and silt” and the steep-sided U-shaped valley that leaves the basin at South Andover. Stone also described “a plain of well-rounded glacial gravel” in the valley from South Andover to Rumford Point, with a “plexus of reticulated ridges inclosing kettleholes and a lakelet” near Rumford Point. He argued that glacial processes must be responsible for these deposits since the modern Ellis River was “so gentle that it is impossible to accept such coarse, well-rounded matter as ordinary stream wash.” The possibility of a glacial lake was also mentioned due to the presence of clay in some areas of the valley. Stone presented an elaborate argument for an ice dam in the Androscoggin River Valley near Rumford that would have created a proglacial lake extending up the Ellis River Valley to Andover, and in the Barkers Brook Valley to the south of the Androscoggin River, with a spillway elevation of about 730 feet in North Woodstock (Bryant Pond

The ice or debris dam near Rumford must have persisted for some time to allow a continuous lake to form from the Androscoggin Valley up to Andover as described by Stone (1899), and other theories were put forth by subsequent researchers. Leavitt and Perkins (1935) visited the Ellis River Valley during their efforts to catalog the quality of Maine’s glacial deposits for use in road construction. Many of their observations were comparable to Stone’s with some additions. They noted that the Andover basin, with “an unknown thickness of sand” must have formed in granitic bedrock, similar to other inter-mountain basins in New England. A lake was hypothesized to exist in the upper portion of the valley, including the Andover basin, with an ice or debris dam blocking the Ellis River Valley between Plumbago Mountain and Howe Hill (the western arm of Rumford Whitecap Mountain). Lastly, kame terraces were observed to line the valley, which in the context of their publication refers to a wide variety of stratified sediments.

From 1963 to 1965, Maine Department of Transportation (MDOT) geologist Maurice Fournier cataloged aggregate sources in the Rumford region. Most deposits were mapped from pre-existing borrow pits and their geologic origins were loosely categorized in several MDOT soils reports (e.g. Baker, 1967). Fournier wrote a short synopsis of his thoughts on the geology of the Ellis River Valley, building on the ideas of Leavitt and Perkins: “…damming of the glacial waters probably occurred at a construction [sic] within the valley between Howe Hill and the northerly end of Mt. Dimmock. This allowed the glacial meltwater to rise until another channel developed. Apparently the Ellis River once traversed a more easterly course from North Rumford and flowed into the Androscoggin River about five miles downstream from where it now terminates at Rumford Point.” It is interesting that neither Stone (1899) nor Leavitt and Perkins (1935) mentioned Meadow Brook and Split Brook, which now occupy the valley Fournier described. Both Meadow and Split Brook are underfit for this valley, and it is certainly possible that the Ellis River, or even the Androscoggin River, once took this path. However, this valley likely existed prior to glaciation instead of being carved out by glacial meltwaters due to its size and traverse through metamorphic bedrock.

Aquifer mapping field work for the East Andover quadrangle was completed in 1989 (Nichols and others, 1995), although the physical maps for the area have been updated over the years. Seismic profiling was a part of this field work and revealed bedrock depths of approximately 288 feet (88 meters) in the central portion of the valley between Howe Hill and Plumbago Mountain, providing further evidence of the glacially carved valley. Data from this mapping was compiled into a surficial materials map, which has also been updated over the years and as a result of the most recent mapping efforts. More detailed aquifer information will be presented in the hydrogeology section of this guide.

In the more recent decades, research in the Western Maine Mountains and nearby White Mountains of New Hampshire has focused on the manner and timing of regional deglaciation. Geologists have long debated the state of the Laurentide Ice Sheet during its retreat from the state: Was it actively flowing or stagnant? Stone (1899) believed that the ice was active due to his observations of moraines in the Upper Androscoggin Valley. Leavitt and Perkins (1935) thought that the ice sheet thinned to stagnant tongues within the mountain valleys. Koteff and Pessl (1981) put forth the idea of systematic stagnation zone retreat, in which portions of the ice margin were stagnant. (Borns (1989) and Thompson (2001) can be referenced for a more detailed summary of this debate.) Thompson (2001) mentions the East Andover quadrangle in his account of deglaciation in western Maine, citing striation observations from lower elevations to the southwest of Plumbago Mountain that show a localized change in ice flow direction. Striations, gouges, and streamlined topography throughout the quadrangle are generally oriented to the southeast, averaging about 150°. In the area described by Thompson, ice flow shifts to the south/southwest, indicating that there may have been an active ice tongue flowing towards the Stony Brook drainage in the Puzzle Mountain quadrangle to the west. In areas that lack abundant moraines such as the Western Maine Mountains, the direction of ice retreat may also be tracked by the location of meltwater channels (Thompson, 2001) and sediments deposited by glacial meltwater (morphosequences) (Koteff and Pessl, 1981). Basal radiocarbon ages from ponds in western Maine indicate that the East Andover quadrangle was deglaciated between 13,200 and 12,300 radiocarbon years BP (Borns and others, 2004). New work on cosmogenic dating of glacially transported boulders has the potential to further improve the deglacial chronology of the region (Bromley and others, 2015), but similar research has not yet been conducted in the East Andover quadrangle or in the immediately surrounding areas.
Overview of Landforms and Sediments

The landforms and associated sediments of the Ellis River valley can be subdivided into erosional and depositional categories. LiDAR topographic data for the region became available towards the end of the mapping process in spring of 2017, and has been useful in delineating landforms. On the broadest scale, the most noticeable erosional feature is the valley itself, which exhibits the classic steep sides of a glacially formed U-shaped trough in many reaches (Figure 2). Views of the valley shape are possible from portions of the hike up Rumford Whitecap Mountain during this trip (Stop 7).

Figure 2. The top image shows LiDAR hillshade imagery over aerial photography for a section of the Ellis River Valley between the eastern edge of Little Puzzle Mountain (Puzzle Mountain quadrangle) and Farmers Hill. The graph shows a topographic cross section through the valley (extracted from the 2-meter LiDAR DEM) illustrating the steep sides of a U-shaped valley. Depth to bedrock in this section of the valley approaches an additional 200 feet (61 meters) below the valley floor (Locke and others, 2017).

Rôche moutonées are classic erosional landforms, created by glacial abrasion on the up-ice side of a bedrock obstruction and plucking on the down-ice side, forming an asymmetrical hill. North Twin Mountain is the best example of a rôche moutonée in the area, but lies just to the east in the Rumford quadrangle (Figure 3a). Fortunately,
North Twin Mountain will be visible from the summit of Rumford Whitecap Mountain during Stop 7. LiDAR hillshade imagery also reveals areas of glacially streamlined topography in great detail (Figure 3b).

Figure 3a and 3b. LiDAR hillshade of North Twin mountain (just off the northeast corner of the quadrangle) is shown on the left (3a). The north/northwest side is gently sloping, while the southeast side is steep from glacial plucking. LiDAR hillshade imagery of glacially streamlined topography in the south-central portion of the quadrangle is shown on the right (3b). Arrows indicate approximate ice flow direction.

There are many examples of micro-erosional features in the quadrangle such as glacial striations, grooves, and crescentic gouges. Many of these features will be visible on the hike up Rumford Whitecap Mountain at Stop 7 (Figure 4).

Figure 4. Crescentic gouges along the Rumford Whitecap Mountain red trail. Rock or boulders lodged in the ice were forced against the underlying bedrock, chipping pieces from the bedrock to create the gouges. These gouges are usually concave in the up-ice direction, so the ice was flowing from the bottom to the top of the photo in this example. Photo: Lindsay Spigel.
Meltwater channels and stream terraces round out the major erosional landforms in the quadrangle, transitioning from late glacial to modern stream processes. LiDAR topographic imagery has made it easier to pick out meltwater channel routes, especially in forested areas. Channels formed by glacial meltwater can be distinguished from subaerial streams because these channels often have irregular locations and paths, and are commonly occupied by underfit modern streams. A good meltwater channel example in the East Andover quadrangle is the channel currently occupied by the headwaters of Coburn Brook (Figure 5). The channel is quite deep for the size of the modern stream, especially in the uppermost reach where it is about 50 feet (20 meters) deep. It is unlikely that modern flow in this section of Coburn Brook would be enough to carve this channel. The ice margin was likely in this location long enough for a marginal channel to develop in the saddle between Rumford Whitecap Mountain and Farmers Hill.

![Figure 5. LiDAR hillshade imagery of the Coburn Brook meltwater channel (black arrow). Flow was right (northeast) to left (southwest). Note the small, irregular channels that end abruptly in the left-center of the image (red arrow) – it is possible that these are small meltwater channels. Contrast these to the more orderly channels in the southeast corner of the photo (blue arrow) that most likely formed by sub-aerial processes in late-glacial to early Holocene times, before thick vegetation cover became established.](image)

During late-glacial and early post-glacial times, the young Ellis River responded to changes in base level and began to incise through glaciofluvial and glaciolacustrine deposits, terracing the valley over time (Figure 6). The modern channel continues to slowly meander across the valley floor, leaving many well-preserved oxbow channels – each a snapshot of the river at that point in time. See the Stop 4 Road Log for additional imagery.

Depositional features in the East Andover quadrangle include glacial till, glaciofluvial/lacustrine deposits, quaternary alluvial fans, and eolian deposits. Till in the quadrangle is usually stony with a sandy matrix, but deposits have not been analyzed in any detail beyond basic unit mapping. Esker deposits, kettles, and other miscellaneous ice-contact sand and gravels near the confluence of the modern Ellis and Androscoggin Rivers could be considered a “head of outwash” area (Thompson, 2001), indicating that the ice margin was paused in that area long enough for the materials to accumulate. (See Stop 2 Road Log for more landform descriptions.) There is a similar area further up the valley in South Andover, indicating another possible pause during retreat. There may have been more similar evidence in the valley between these areas that has subsequently been altered or covered by other processes and deposits. Overall, the depositional features also point towards a southeast to northwest ice retreat.

The head of outwash deposits may have blocked drainage of the Ellis River Valley at the southern end, but there was another outlet through the Meadow Brook and Split Brook Valleys as was suggested by Fournier. Any ponded water in this area would not have been very deep, unless this outlet was blocked where it meets the Androscoggin Valley as well, making it part of Glacial Lake Hanover. As the ice retreated up the valley, various glaciolacustrine
and glaciofluvial deposits were laid down, filling the valley with sediments as meltwater flowed from the margin and the surrounding hills. Ponded areas likely filled with sediments, eventually becoming outwash plains. The highest elevation sediments in the valley (usually along the margins) slope from just over 700 feet (213 meters) in the northern portion to about 650 feet (198 meters) in the southern portion, but this slope is not continuous, with breaks at South Andover and Howe Hill, again indicating possible pauses in ice retreat. Good exposures of glaciofluvial and glaciolacustrine sediments in the valley are not common, as many gravel pits in the area are no longer active. Figure 7 shows an example from a fresh exposure that has since been closed. LiDAR imagery has also revealed braided channel patterns on some of these higher surfaces (see Road Log for Stop 5.)

Figure 6. Photo of subtle terraces on the west side of the Ellis River Valley near Stop 3. A black arrow points to the most obvious terrace. Photo is looking roughly northwest. Ryan Gordon for scale. Photo: Lindsay Spigel.

Figure 7. Outwash sand and gravel from an exposure in the northern portion of the Ellis River Valley. The photo was taken facing west. Dipping beds in the center of the photo indicate water flow to the south. Scale bar increments are inches on right, centimeters on left. Photo: Lindsay Spigel.
As ice or till dams dissipated, meltwater streams and eventually modern streams began to flow freely and modify the valley deposits. Wind likely reworked these sediments, too, creating small eolian deposits. Streams draining the barren, freshly deglaciated landscape eroded glacial deposits from the uplands. When these steep mountain streams met the relatively flat Ellis River Valley floor, they lost energy and their ability to transport sediments, depositing these sediments to form alluvial fans. Many alluvial fans in the area are easy to pick out since they are preferable for farming, the cultivated fields popping out from the mostly forested landscape. Fans tend to have finer sediments that were winnowed from the hillside (much preferred over stony till), but are not as threatened by flooding since they are slightly above the main valley floor (Figure 8). (See also Road Log for Stop 1.)

![Figure 8. LiDAR hillshade and aerial imagery of an alluvial fan (roughly outlined in red) in the southeast portion of the quadrangle. The fan is preferable for farming and thus is easy to identify in the otherwise forested landscape. The many channels that fed the fan are also visible.](image)

**HYDROGEOLOGY**

**Water Resources in the Lower Ellis River Valley**

The Lower Ellis River Valley is rich in groundwater and surface water resources. The mean annual precipitation in the watershed is 44 inches. The Ellis River itself is gauged by the USGS at a point in South Andover (USGS gauge 01054300), where the annual mean discharge is 300 cubic feet per second from a watershed area of 130 square miles. Monthly mean flows range from 103 cubic feet per second in September to 880 cubic feet per second in April.

Surficial materials that fill much of the valley are saturated with fresh water and make good aquifers, that is, the saturated sediments are transmissive enough to yield groundwater to a properly constructed well at a rate greater than 10 gallons per minute. The Maine Geological Survey maps most of the valley bottoms in the East Andover quadrangle as significant sand and gravel aquifer, including the entire Ellis River Valley, plus the valley bottoms in the Split Brook and Meadow Brook watersheds (Foster and others, 2016). High-yielding aquifers are mapped on both sides of the Ellis River along most of its length from the confluence of the East and West Branches in Andover down to the Androscoggin River. High-yielding aquifers are defined as those where saturated surficial deposits have the potential to yield greater than 50 gallons per minute to a properly constructed well.

The bed of the Ellis River is very sandy and is likely well connected to the aquifers below and on both sides, so that the surface water and groundwater in the valley can be viewed as a single resource. The Ellis River likely receives net groundwater inputs along most of its length, which increase its discharge in addition to tributary inputs of surface water. On top of the net gain to the river from groundwater, groundwater and surface water probably exchange frequently in both directions in what is called hyporheic exchange, the bi-directional exchange of water through the shallow riverbed and riverbank sediments.
Although farmers may withdraw surface water from the Ellis River and tributaries from time-to-time for irrigation, and residents throughout the valley use springs, bedrock wells, and dug wells for domestic uses, there are no especially large withdrawals of groundwater in the valley, except for the Rumford Water District (see below).

**Milligan Farm Esker**

The esker system that runs roughly parallel to the Ellis River consists of thick sand and gravel deposits (see Stops 2 and 3). The section of esker that is utilized for its water resources is about two miles north of where the Ellis River flows into the Androscoggin, and is called the Milligan Farm esker after the family that owned and farmed the surrounding land. Here the esker ridge runs almost due north-south along the western bank of the river. The ridge is approximately 250 feet wide, and 20 to 30 feet higher than the land to the west, which is mostly river terrace and wetland. The Ellis River has eroded into the esker sediments at several locations where meander bends arch westward, and has cut entirely through the esker to the north (See Stop 3, Figure 14). Exposed sediment near the crest of the esker shows medium to coarse sand and rounded gravel and cobbles (Figure 9). At locations near the southern end of the esker, similar rounded gravel and cobbles can be seen in the river bed and bank where the stream impinges on the steep esker side (Figure 10). In these locations, concentrated spring flow is clearly visible entering the river through the bank close to the water line. Beneath the esker, sand and gravel deposits have been found to at least 90 feet below the ground surface (Locke and others, 2017).

![Figure 9. Exposed sediment near the crest of the esker, showing medium to coarse sand, rounded gravel and cobbles. Photo: Ryan P. Gordon.](image)

**Rumford Water District Wells**

Beginning in 1913, the Mount Zircon Reservoir, managed by the Rumford Water District (RWD), was the primary supply of drinking water for the town of Rumford. When the Safe Drinking Water Act was amended in 1986, the new federal law added additional regulated contaminants and mandated that surface water systems be filtered, among other strengthened drinking water standards. These changes prompted RWD to make improvements in their water supply infrastructure, including a move away from surface reservoirs. A study by the district indicated that the system needed significant improvements, including a new source of groundwater. At that time, RWD operated two wells on the Swift River, installed in 1953 and 1963, but the yield was insufficient, and they were only used when drought or water quality impacted the reservoir (A.E. Hodsdon, 1989a).

In 1988, preliminary reconnaissance by Bradford Caswell noted several locations where the gravel esker is in contact with the adjacent river on property owned by Robert Milligan (Milligan Farm Esker), near where RWD had
done limited test drilling in 1961 (A.E. Hodsdon, 1989a). The more southerly area, where rounded gravel was noted in the river bed, was identified as the highest priority site to investigate and to install additional test wells. In 1989, 14 test wells were installed in the esker area, finding as much as 100 feet of coarse sand, gravel, cobbles, and boulders (A.E. Hodsdon, 1989b). Prior to pumping, groundwater flow through the esker was to the south and east. Analysis of a 5-day pump test at 400 gallons per minute determined high transmissivity values of between 300,000 to 400,000 gallons per day per foot, with an average saturated thickness of 50 feet, leading to an estimated hydraulic conductivity of around 900 feet per day (A.E. Hodsdon, 1989b). In 1990, a production well was drilled that provided up to 700 gallons per minute. The well was brought online in 1991, but the well pumped sand and yields soon dropped. It was determined during investigations in 1995 that the well had been improperly constructed and was faulty (A.E. Hodsdon, 1998).

A replacement well was installed in 1997, to the north of the first well, and continues to supply water of excellent quality to the Rumford area today. This well is 95 feet deep with a 20-foot screen, and has a safe yield in excess of 1000 gallons per minute. During the testing of this well, a combined 1700 gallons per minute were pumped from both production wells for two days, which successfully induced recharge from the Ellis River (A.E. Hodsdon, 1998). This pump test demonstrated that a connection between the river and esker aquifer does exist, and that the river can provide water to the wells in addition to the natural recharge and storage of the aquifer.

ACKNOWLEDGEMENTS

Entry to many sites listed here is at the discretion of the landowner. The authors wish to thank the many landowners for granting access to private property and permission to park for the purposes of this trip on September 30, 2017. Thank you to the Mahoosuc Land Trust for hosting us on their trails to Rumford Whitecap Mountain. Thank you to Woody Thompson for providing photos and additional information about the area.
ROAD LOG

MEETING POINT: Saturday, September 30th, 8:30 AM in the Maine DOT Riverside Rest Area on U.S. Route 2 (356262.00 m E, 4923630.71 m N) in Bethel, ME. From the intersection of State Route 35 and U.S. Route 2 in Bethel, head east on U.S. Route 2 towards Rumford; the rest area will be on the right side in about 3.5 miles. We will consolidate vehicles as much as possible and head east to the first stop from here. Those who will not be hiking Whitecap Mountain should share a vehicle so they don’t get stuck at the last stop without a ride back to the meeting point. The mileage starts over at each stop to avoid any confusion that might occur from parking and turning around in slightly different areas.

Mileage

0.0 From the Maine DOT Riverside Rest Area, turn right onto U.S. Route 2.
13.3 Turn left onto Andover Road. This road follows the Split Brook Valley. Note the size of the valley in comparison to the size of Split Brook, which we will cross at approximately mile 14.7 (there is a small bridge with sign).
15.3 Turn right onto Kimball Road.
16.3 Drive to the end of Kimball Road (passing the big red barn and farmhouse). Turn around in the small cul-de-sac at the end of Kimball Road and drive back down towards the farmhouse. Park on the side of the road just before reaching the farmhouse.

STOP 1. KIMBALL FARM (3694763.11 m E, 4933470.70 m N).

The Kimball Farm has a very interesting landscape, which can be viewed from Kimball and Red Hill Roads. Please stay on the roads/edge of roads and refer to the LiDAR image below to guide you during the stop (Figure 11). Walk to the north end of Kimball Road or view the area as we turn around at the end of the road. Observe the small ridges to the west. It is possible that there was a small ice tongue flowing down the east side of Rumford Whitecap Mountain, and these ridges are likely very small esker segments that are part of some associated miscellaneous ice-contact deposits. View a till exposure on the east side of Kimball Road, north of the farmhouse. Walk up Red Hill Road and look down into the hillside channels; consider how and when these channels formed. These channels were also viewed during a previous NEIGC trip (Thompson, 1989). At the time, much of this hillside was pasture and the channels were easily viewed from the road and accessed on foot. It was postulated that these channels were glacial meltwater channels. What do you think? Walk south on Kimball Road to the edge of the farm and view the alluvial fan and overall landscape. In 1990, two sediment cores were taken in the fan roughly south of the main barn as part of aquifer mapping in the area. Sediments in the cores varied widely, but generally indicated mixed fluvial sediments (silt to pebble gravel) over possible lacustrine sediments and/or till (Locke and others, 2017).

Figure 11. Stop 1 LiDAR hillshade and aerial imagery. Star at approximate parking location.
0.0 From the Kimball Farm House, head south on Kimball Road.
0.6 Turn right onto Andover Road.
1.8 Turn left onto Whippoorwill Road.
3.7 Turn right onto U.S. Route 2.
3.9 Turn right onto Maine Route 5 (Ellis River Road).
5.0 Turn right into gravel pit.

STOP 2. BERNARD PIT (366529.67 m E, 4931204.44 m N).

This gravel pit is located in a portion of the Ellis River esker system (Figure 12). Eskers form at ice margins as sub-glacial meltwater carves conduits up into the base of the ice, releasing sediment from the ice which accumulates over time to build the ridge-like landform. Fluvial transport of materials within the esker system also tend to round and sort the sediments. An excellent synopsis of esker formation can be found in Thompson and Hooke (2016). The pit has not been excavated in a few years, but one can still get an idea of the esker formation. Please do not climb up on the pit face. Observe the cobbles present in the deposit. No provenance studies have been done at this location, but rock types represented in the esker are similar to the bedrock units described to the north and west of this location including: Several different kinds of igneous rocks like granitic pegmatite, granite, granodiorite, or tonalite with included blocks of related igneous rocks that may be from the Mooselookmeguntic Pluton and dominate the deposit; coarse or fine-grained mafic cobbles that may be from the Plumbago mafic or ultramafic intrusions, or from younger basaltic dikes; and smaller amounts of metasedimentary rocks like schist or granofels that match descriptions of the regional bedrock units. Observe the sand on the east side of the pit; consider how this deposit was formed. Figure 13 is an older photo of this section of the pit with a relatively fresh exposure, but note that material has been removed from this area since the photo was taken.

Figure 12. Stop 2 LiDAR hillshade and aerial imagery. Star marks approximate parking location.
From the pit drive, turn right onto Maine Route 5 (Ellis River Road). Pass Davis Pond Kettle on right.
0.9 Turn right on farm road, just past large white barn. Park on either side of the farm road before the gated bridge. Please do not block the gate.

STOP 3. RUMFORD WATER DISTRICT (366829.95 m E, 4932509.98 m N).

Here we will walk into the water district property and over the esker, viewing an exposure of esker sediments and the well pump houses (Figure 14). We will then walk toward the southern end of the esker to see the river bank springs and eroded esker cobbles.

Figure 14. Stop 3 LiDAR hillshade and aerial imagery. Star marks approximate parking location.
At the intersection of the farm road and Maine Route 5, turn right onto Maine Route 5.

7.7 Turn right into the third entrance to Woodlawn Cemetery. Follow the road to the back of the property where it turns to loop back around and park, staying on road. Please use caution – some graves are very close to the road so please do not stray from the road tracks! If the gates are closed, we will have to park along Maine Route 5 and walk to the back of the cemetery.

STOP 4. WOODLAWN CEMETERY (361554.13 m E, 4942498.84 m N).

At this brief stop we will walk down a woods road at the back of the cemetery to view an impressive terrace scarp and large oxbows (Figure 15).

Figure 15. Stop 4 LiDAR hillshade and aerial imagery. The small excavation in this image (bottom left, above the scale bar) is the location of the exposure shown in Figure 7. Star marks approximate parking location.

At the intersection of the cemetery road and Maine Route 5, turn right.

0.9 Turn right onto Maine Route 120 in Andover Village.

1.7 Stay straight onto East Andover/Rumford Center Road.

STOP 5. CHADBOURNE PIT (363478.00 m E, 4942388.84 m N).

This gravel pit has not been recently active, but has exposures of coarser glaciofluvial sediments. Consider the sediments along with the imagery in Figure 16. The area includes an unnamed kettle lake (west), evidence of former braidplain (east and southwest), and a small esker remnant (southwest). The braidplain to the east lies at about 705 feet (215 meters) and is oriented northeast to southwest. There is also evidence of braidplain on this higher surface to the northwest in the Andover and Ellis Pond quadrangles - this plain lies at about 702 feet (214 meters) and is oriented northwest to southeast. (We drove over this surface as we passed through Andover Village.) Meltwater was likely diverted around a streamlined hill that runs northwest to southeast just north of the area displayed in Figure 16, and met near the modern confluence of the East and West Branches of the Ellis River. Sediment was still coarse enough to warrant a braided channel system even as the stream began to incise, as there is still remnant braidplain at about 656 feet (200 meters) just south of the kettle lake. The modern floodplain lies at about 636 feet (194 meters), so that’s about 69 feet (21 meters) of stream incision since deglaciation. Along with Stops 4 and 5a, we will have experienced the transition of fluvial systems from late-glacial to modern times, all in one corner of the quadrangle.
From parking area in the pit, follow logging road to the southwest corner of pit.

Turn around in log yard and park on side of road. From the log yard, we will walk down a woods road about ½ mile to:

**OPTIONAL STOP 5A. MEETING OF THE WATERS (362999.85 m E, 4941895.23 m N).**

Time permitting, we will view an impressive cut bank on the West Branch of the Ellis River just upstream of its confluence with the East Branch (hence the “Meeting of the Waters”). The view is best if you wade across the river here, so bring your rubber boots if you packed them. One can really get a feel for the type and volume of glaciofluvial sediments in this area (Figure 17).
0.0  From the gravel pit parking area, turn right onto East Andover/Rumford Center Road.
2.5  Bear right onto Covered Bridge Road.
3.0  Park on side of road. There is a small pull-out just before the bridge that will fit about three cars, but others must park on the side. Do not cross the bridge – the road narrows after the bridge, making it difficult to park. Do not park in the entrance to Covered Bridge Campground on the west side of the bridge.

**STOP 6: LUNCH AT LOVEJOY COVERED BRIDGE (362453.80 m E, 4939229.61 m N).**

Just upstream of the covered bridge is the USGS continuous gauging station on the Ellis River (station #01054300). This station uses a nitrogen gas bubbler to measure the river stage (height of water). The stage is related to discharge by creating a rating from discharge measurements that are periodically made downstream of the bridge by USGS personnel.

0.0  Turn around in the road and head east on Covered Bridge Road.
0.5  Turn right onto East Andover/Rumford Center Road.
4.1  Parking lot for Rumford Whitecap Preserve. There is only room for about five cars in this lot. Please do not park in the lot if you will not be hiking. Those at the end of the caravan should park on the more open west side of the road before reaching the parking lot.

**STOP 7: RUMFORD WHITECAP PRESERVE (366236.73 m E, 4934338.15 m N).**

A separate handout will be distributed for this stop so hikers don’t have to lug their guidebooks along and can move at their own pace. We will hike up the yellow trail, meeting the orange/red trail for the last stretch to the summit, and will descend on the orange/red trail. The hike is about five miles round trip, so bring whatever you think you will need for that distance. Time permitting, we will begin with a visit to some inactive pits just off the yellow trail to the Rumford Whitecap summit. These pits expose a wide variety of materials in a relatively short transect. The group will stay together while viewing these pits and then those that do not wish to hike can depart.

Outcrops of the metasedimentary units that are crossed on this hike – the rusty-weathering graphitic schist of the Smalls Falls Formation on the lower slopes, and the light-colored quartz-rich biotite granofels and quartzite of the Perry Mountain Formation near the tree line – are small and poorly exposed. Look for relatively fresh clasts of the country rock in the surficial deposits (float) at the start of the trail – these angular and friable pieces of rusty-weathering, dark-colored schist obviously did not travel far from their source.

The pegmatite contact nearly coincides with the tree line as you hike up the mountain. The rock is very white due to the abundance of quartz, feldspar, and muscovite – hence the name “Whitecap” – and there is nearly 100% exposure of the rock. The large size of the minerals make them easy to identify. Numerous other igneous features are easily observable – pegmatite dikes cutting fine-grained granite, xenoliths of granite or metasedimentary rock included in the pegmatite, and graphic intergrowths of feldspar and quartz or tourmaline and quartz are common (Marvinney, 2012).

Glacial striations and crescentic marks are easily observed above the tree line, with many examples occurring right along the trail (Figure 4). Glacially transported boulders are also plentiful, but are not erratics since they are of the same rock type as the bedrock in the vicinity (Marvinney, 2012). Rumford Whitecap’s summit offers 360° views of the surrounding glaciated landscape – hopefully the weather will be clear so we can take advantage. The remainder of the ascent and descent mostly traverses stony till.

**End of trip.** To return to the meeting point, head south on East Andover/Rumford Center Road from the Rumford Whitecap Preserve parking lot. Drive 0.2 mi and turn right onto Andover Road (esker ridge follows the left side of the road for a bit). Drive 0.3 mi and turn left onto Maine Route 5/Ellis River Road. Drive 2.9 mi and turn right onto U.S. Route 2. Drive 8.8 miles to Maine DOT Riverside Rest Area on left.
REFERENCES

A.E. Hodsdon, 1989b, Hydrogeologic evaluation of 8-inch test well at Site 1 on the Ellis River Esker Aquifer in Rumford, Maine, report dated December 14, 28 p.


Thompson, Woodrow B., and Borns, Harold W., Jr. (editors), 1985, Surficial geologic map of Maine: Maine Geological Survey, 42" x 52" color map, scale 1:500,000.


DEVONIAN GRANITE MELT TRANSFER IN WESTERN MAINE: RELATIONS BETWEEN DEFORMATION, METAMORPHISM, MELTING AND PLUTON EMPLACEMENT AT THE MIGMATITE FRONT

By

Gary S. Solar
Paul B. Tomascak
Michael Brown

Email addresses: solargs@buffalostate.edu, tomascak@oswego.edu, mbrown@geol.umd.edu

INTRODUCTION

This trip is designed to examine evidence for syn-tectonic metamorphism, melting and granite magma ascent during transpressive deformation, as recorded by structures, petrology, geochemistry and ages of metasedimentary rocks, migmatites and granites in the Rangeley-Rumford area of western Maine (Figure 1). The trip is intended as a close examination of the variation of mineral fabrics, shapes and sizes of granite bodies (from leucosomes to plutons), and the relation of these to the regional structure, by visiting rocks that were deformed, metamorphosed, and partially melted during the Devonian Acadian orogeny. Data that form the foundation of models developed to explain the evolution of the region are summarized below and are presented by our group in the literature published in the last 20 years (Brown and Solar, 1998a, 1998b, 1999; Brown and Pressley, 1999; Pressley and Brown, 1999; Solar and Brown, 1999, 2000, 2001a, 2001b; Johnson et al., 2003; Tomascak et al., 2005). The field work of Solar (1999) that is the base of these papers owes a debt of gratitude to the career studies of C.V. Guidotti and R.H. Moench, and influence of E-an Zen. The migmatites and the surrounding rocks of this trip are a subset of a 2001 Geological Society of America field excursion (Solar et al., 2001). The rocks of this trip are closely related to non-migmatitic rocks that are the focus of both the 2001 GSA field guide and Trips A2, B1, and B4 of the present NEIGC guidebook. The migmatites are related to the rocks on Trip C6, although our interpretations are dissimilar. The granites on Whitecap Mtn. that are included in part of Trip B5 are part of the granite suite that is a focus of this trip.

The trip is in three parts, corresponding to rock type and timing. The first part consists of Stops 1 and 2 at Coos Canyon and north, the non-migmatitic rocks immediately north of the migmatite front (Figure 2). Rocks at Stops 1 and 2 (and optional stops) illustrate the regional NE-SW-striking structural geology as it is recorded both by folds of the stratigraphic sequence and by the orientation and intensity of metamorphic mineral fabrics. Structural, petrographic, geochemical and geochronological data support the idea that these rocks and structures are characteristic of the migmatite protolith before anatexis. The second part of the trip is devoted to examination of the petrological and structural record of migmatite formation in rocks of the stratigraphic sequence, and the record of granite ascent as illustrated by associated granites in the migmatites (Figure 2; Stops 3 to 8). The third part is the plutons in the system, the Mooselookmeguntic Igneous Complex (MIC) in particular (Figure 2; Stops 9 and 10).

GEOLOGICAL SETTING

The northern Appalachians are divided into NNE-SSW-trending tectonostratigraphic units (Figure 1). The Central Maine belt (CMB), continuous with the Central Mobile belt in Maritime Canada, is the principal unit that occupies most of the eastern part of New England and New Brunswick. The CMB is composed of a Lower Paleozoic sedimentary succession, deformed and metamorphosed at greenschist to upper amphibolite facies conditions, and intruded by plutons of Devonian age (e.g. Moench et al., 1995; Bradley et al., 1998; Solar et al., 1998). The CMB is located between Ordovician rocks of the Bronson Hill belt (BHB) to the W and NW that were deformed and metamorphosed during the Ordovician Taconian orogeny, prior to deposition of the CMB sedimentary sequence. To the SE of the CMB, rocks are Neoproterozoic to Silurian age of the Avalon Composite terrane (ACT).

Regional deformation was partitioned heterogeneously during dextral transpression in response to Early Devonian oblique convergence (van Staal and de Roo, 1995; van Staal et al., 1998). Dextral– SE-side-up
Figure 1. Geological map of Maine illustrating principal tectonic belts, structures, metamorphic zones and plutons (modified after Solar and Brown, 1999; metamorphic zones modified after Guidetti (1989). The Migmatite-Granite Complex and Sebago pluton (S) are subjects of Solar and Tomaskak (2016 NEIGC). Pluton distribution is apparently independent of metamorphic zones. Metamorphic zones increase grade NE to SW along regional strike, whereas plutons are apparently arranged into belts that parallel regional strike. Metamorphic isograd patterns are apparently perturbed at structural zone boundaries (Solar and Brown, 2001a). D is Deblois, K is Katahdin, L is Lucerne, Lx is Lexington, M is Mooselookmeguntic, CLB is Chain Lakes Massif.
displacement was accommodated within the CMB shear zone system (Brown and Solar, 1998a; Solar and Brown, 2001a) while dextral-transcurrent displacement was accommodated within the Norumbega shear zone system (Figure 1; West and Hubbard, 1997; West, 1999) along the southeastern side of the CMB. CMB deformation had ceased by the Carboniferous, and strain localized into the Norumbega shear zone system (West, 1999).

The Rangeley stratigraphic sequence

The CMB is composed of a Silurian to Early Devonian sedimentary succession called the "Rangeley stratigraphic sequence" (Figure 2) that was deformed and metamorphosed during the Devonian (Acadian orogeny; Bradley et al., 1998). The sequence is estimated to be as much as 10 km in thickness, made up of ~5 km each of Silurian and Devonian rocks (Moench and Boudette, 1970). Stratigraphic units are defined in the western Maine area (Figure 2; Moench and Boudette, 1970), and have been extended across most of the New England Appalachians (e.g. Hatch et al., 1983).

The stratigraphy of the sequence is preserved through metamorphism, and begins in the northwest with a proximal coarse conglomerate in the lower part of the Rangeley Formation (Stop 1 of Solar et al., 2001), interpreted to mark the beginning of the Silurian (e.g., Moench, 1970), grades upward (to the SE) into a progressively distal Silurian turbidite sequence and finishes in the central part of the area (Figure 2) with a distal Devonian unit (see Moench et al., 1995, for a complete summary). The sequence has been separated by Moench (1970) into seven apparently conformable formations, based largely on the relative thickness and frequency of alternating centimeter-to decimeter-scale psammite v. pelite layers (inferred relict bedding; Moench, 1970; Moench and Boudette, 1970), coupled with variations in the proportion of metamorphic minerals in the pelite layers (R.H. Moench, 1998, personal communication). Locally, cross-stratification is preserved in psammite layers [particularly in the Perry Mountain Formation (STOP 1)]. The summary stratigraphic succession listed in Figure 2 follows the compilation of Moench et al. (1995), modified in the eastern part of the study area after mapping by Solar (1999; Solar and Brown, 2001a).

Regional metamorphism

High-\(T\) – low-\(P\) polymetamorphism of the Rangeley stratigraphic sequence in western Maine is well documented, particularly regarding the rocks to the northwest and east of the migmatite domains (Guidotti, 1970, 1974, 1989; Holdaway et al., 1982, 1997; See Figure 2). The trip area is located in the part of Maine where greenschist facies rocks to the northeast increase in grade to upper amphibolite facies within 20 km along strike to the southwest (Figure 3b; see summary in Guidotti, 1989). Across the area, the amphibolite facies rocks are characterized by porphyroblasts of garnet, staurolite and locally pseudomorphed andalusite enclosed within a matrix dominated by muscovite, biotite, quartz, plagioclase, and opaque phases (ilmenite, graphite and pyrite). Fibrolite is an important fabric-forming matrix phase at upper amphibolite facies, especially in migmatitic rocks. The peak of regional metamorphism in western Maine was reached at c. 404 Ma during the waning stages of transpressional deformation (Solar and Brown, 1999, 2001a, 2001b). Regional isotherms are inferred to have been shallowly inclined at lower grades and closely spaced around synmetamorphic granites and at the migmatite front, consistent with advection-controlled intracrustal redistribution of heat (‘pluton-driven metamorphism’) within the regionally extensive thermal high (Johnson et al., 2003).

Thermodynamic modeling by Johnson et al. (2003) in the MnNCKFMASH subsystem is consistent with field data and implies a metamorphic field gradient from ~3.5-4.0 kbar at lower grades (500-520 °C) to >4.5 kbar at suprasolidus temperatures (> 700 °C). Because peak pressures vary both along and across the strike of the CMB, Brown and Solar (1999) and Johnson et al. (2003) interpreted differential thickening during syntectonic metamorphism. Contact metamorphism associated with the Mooselookmeguntic igneous complex occurred ca. 35 million years after the regional metamorphic peak (Tomascak et al., 2005), and records slightly higher pressure
Figure 3. (a) Form line map of foliation (from Solar and Brown, 2001a). More closely spaced lines denote steeper dips, more widely spaced lines denote moderate dips. Stereograms and lower-hemisphere, equal-area projections. (b) Simplified structural and metamorphic zone map of (a). Section A-A' shows the inferred listric geometry of the structural system where ‘straight’ belts (AFZs; shaded) converge into a root zone at depth (Brown and Solar, 1999). The section below A-A' is the area within the dotted box showing the inferred geometry of folds of the stratigraphic sequence. Section B-B' is drawn to show the higher-grade rocks, particularly the distribution of migmatites. Line segments on both sections are intersections of foliation with the plane of the sections.
conditions than the regional event (Johnson et al., 2005). The final increment of late Acadian thickening beyond ca. 404 Ma accounts for the pressure increase, consistent with the inferred regional cooling prior to the emplacement of the Mooselookmeguntic igneous complex. An overall counter-clockwise P-T-t evolution is implied in the CMB (Johnson et al., 2003), consistent with that proposed for Acadian metamorphism in western New Hampshire.

**Plutonism**

Plutons are kilometer-scale (Figures 1 and 2). In contrast to country rocks, the plutonic rocks record only local evidence of solid-state deformation internally, although foliation is apparently deflected in rocks around the Redington pluton (R; Figures 2 and 3). Some larger-volume plutons cut across the regional structures without either significant deflection of structural trends or formation of a significant deformation aureole (Figure 3a), suggesting displacement of rock out of the map plane. The close association between smaller-volume plutons, such as the Phillips pluton, the Lexington pluton in its northern part, and heterogeneous migmatite in similar structural zones (Figure 3), has been used to suggest a relation between structure, granite ascent and emplacement (Brown and Solar, 1998a, 1998b, 1999; Pressley and Brown, 1999; Solar and Brown, 2001b).

**Timing of orogenesis**

U-Pb monazite ages from samples of pelite collected from staurolite zone rocks demonstrate two distinct concentrations of metamorphic ages that are interpreted to reflect regional metamorphism at 405-399 ± 2 Ma and contact metamorphism related to the Mooselookmeguntic Igneous Complex (MIC; “M” in Figure 2) at 369-363 ± 2 Ma (Smith and Barreiro, 1990). U-Pb zircon and monazite ages (interpreted to date crystallization) from samples of granite sheets and lenses in stromatic migmatite, and plutons (including the Phillips pluton) are concordant, and are similar in the range c. 408-404 Ma, except the younger MIC, which yielded ages of ca. 389 and ca. 370 Ma from two discrete granite types (Solar et al. 1998) and younger northern lobe of the Lexington pluton which yielded an age of ca. 365 Ma whereas the central and southern lobes yielded ca. 404 Ma. These data support a model of contemporaneous deformation, metamorphism, and granite ascent (Brown and Solar, 1999).

**STRUCTURAL GEOLOGY OF THE RANGELEY-RUMFORD AREA**

**Two types of structural zone: ‘straight’ belts and intervening zones**

Solar and Brown (2001a) describe the principal structures of western Maine as: (1) the kilometer-scale open to tight folds of the stratigraphic succession, defined by the orientation of centimeter to decimeter scale psammite–pelite compositional layers (see Figure 3b, section A-A’); and, (2) the kilometer scale alternation of structural zones, defined by the NE–SW-striking and steeply SE dipping domainal structure of the CMB shear zone system and the characteristic fabrics of the alternating zones (Figures 1, 2 and 3; Solar and Brown, 1999, 2001a). The regional structure defined by the former is illustrated by the geological map of Fig. 2, whereas the structure defined by the latter is illustrated by the foliation form line map of Figure 3a, in which a pattern emerges of zones of straight and sub-parallel form lines envelope zones where the form lines are more variable in strike. These two types of structural zone coincide with differences in the dip of compositional layers; the layers are more steeply dipping in the zones of straight foliation form lines. The structural style and intensity of fabrics in each of these zones vary with compositional layer and metamorphic grade (Solar and Brown, 1999, 2001a).

Stereograms of the attitudes of mineral fabrics and compositional layers (Figure 3a) show the fundamental difference between each zone. A strong NE-plunging penetrative mineral elongation lineation (amount of plunge variable) is present in metasedimentary rocks in both types of zone, and is defined by the same metamorphic minerals at the same metamorphic grade. In contrast, the intensity and orientation of foliation vary by zone. Where foliation form lines are sub-parallel, foliation is intense and sub-parallel to contacts between compositional layers (Figure 3; see STOP 1). This structural style occurs in 'straight' to arcuate belts at outcrop and map scales (Figure 3). In contrast, rocks in the intervening zones between these 'straight' belts have conspicuously less intense foliation with variable orientation (Figure 3; see STOPS 2, 2a and 2b). Further, compositional layers in the intervening zones vary in attitude, and are not parallel to foliation in the same outcrop, which transects the layers. At map-scale, the pattern of structural zones shows an alternation of these two types such that 'straight' belts of consistently-oriented NNE-striking planar structures, some of which anastomose, separated by intervening zones, some of which are lens-shaped, in which planar structures vary in orientation (Figure 3). Boundaries between the structural zones are
gradational in outcrop over meter-scale transition zones. Across these transitions, traversing away from 'straight' belts into the intervening zone rocks, compositional layers and foliation is progressively more variable in strike and more moderate in dip, concurrent with lower foliation intensity. However, the orientations of both mineral lineations and hinge lines of folded compositional layers generally do not vary significantly across these transitions. Lineation is equally well developed in both zones, with a progressively more variable attitude across the transition (Figure 3a, stereograms). Contacts between stratigraphic units generally occur within these structural transition zones, as do the transitions between stromatic and heterogeneous migmatite types, but at map-scale unit contacts are transected at a shallow angle by structural zone boundaries (cf. Figures 2 and 3). Also, styles of migmatite and shapes of granite bodies within the migmatite domains (TAD and WAD; Brown and Solar, 1999; Solar and Brown, 2001b) vary consistently with structural zone, where stromatic migmatite and sheets of granite are largely within 'straight' belts, whereas heterogeneous migmatite and cylinders of granite occur exclusively within intervening zones (Figure 3).

Structural data collected from 'straight' belt rocks at all grades of metamorphism, including migmatite, show a consistency in orientation of all structural elements that led Solar and Brown (2001a) to suggest nearly complete transposition of compositional layers into sub-parallelism with the tectonic fabric as defined by the matrix minerals. The consistency of orientation of both foliation and lineation across the strike of 'straight' belts suggests that the foliation is sub-parallel to the \( \lambda_1 - \lambda_2 \) plane, and the lineation is sub-parallel to the direction of maximum principal finite stretch (\( \lambda_1 \)).

Apparently coeval symmetric structures (e.g. biotite 'fish' and elongate strain shadow tails around porphyroblasts) and en echelon structures (e.g. stacked ramps of compositional layers, tension gashes, pytigmatic folds and shear fractures (all seen at STOPs 1 and 3)) all suggest consistent kinematics: oblique (SE-side-up and dextral) translation (Solar and Brown, 2001a). The obliquity between inclusion trails (S) and matrix foliation (S\(_{m}\)), as measured in lineation-parallel thin sections (Figure 4) suggests porphyroblast nucleation and growth occurred before final recrystallization of the matrix minerals (Solar and Brown, 1999, 2001a). Textural zones within porphyroblasts with a successively smaller rake between S and S\(_{m}\) suggest progressive rotational reorientation of foliation relative to the porphyroblasts during punctuated porphyroblast growth (Solar and Brown, 2001a). Thus, porphyroblast growth was syn- or inter-kinematic. Solar and Brown (1999, 2001a) proposed that variations in S\(_{m}\)-S rake reflect nucleation and growth of porphyroblasts during progressive regional fold tightening of the stratigraphic succession (Figure 5).

**Alternating zones of finite strain: AFZs v. ACZs**

Solar and Brown (2001a) concluded that if the mineral fabrics define the state of finite strain, the ellipsoid defined by grain shapes in the 'straight' belts has an oblate to plane-strain shape, consistent with the similarity in orientation between compositional layers and the foliation across the zone. In migmatite in the 'straight' belts, the mineral fabrics define a triaxial to uniaxial oblate ellipsoid. Nonetheless, all rocks within 'straight' belts have fabrics that define oblate shapes. Therefore, Solar and Brown (2001a) inferred that 'straight' belts are zones of \( S > L \) tectonite where rocks accommodated apparent flattening-to-plane strain, and refer to 'straight' belts as "zones of apparent flattening" (AFZs; Figure 3). The general parallelism of compositional layers and tectonic fabric suggests high strain, consistent with folding and formation of fabrics in response to finite flattening deformation.

In contrast, in rocks of the intervening zones, compositional layers are not rotated into parallelism with the tectonic fabric, but there is a consistency in orientation of all linear structures. This led Solar and Brown (2001a) to interpret a different state of finite strain in these rocks in comparison with rocks in the 'straight' belts. The variable orientation and weak definition of the foliation suggests that it did not form parallel to the \( \lambda_1 - \lambda_2 \) principal plane, or that this plane has a variable orientation across zones; however, consistency in orientation of the lineations across the regional strike suggests they formed sub-parallel to the maximum principal finite stretch (\( \lambda_1 \)). Structures are symmetrical along the lineation (e.g., strain shadow tails, biotite-quartz pull-aparts along the lineation; STOPs 2, 2a and 2b). This led Solar and Brown (2001a) to suggest dominantly coaxial deformation, with a principal finite stretch along the lineation. As in 'straight' belts, fabrics wrap around the porphyroblasts suggesting that nucleation and growth of porphyroblasts have occurred before final recrystallization of the matrix minerals (Solar and Brown, 1999). Accordingly, Solar and Brown (2001a) concluded that the strain ellipsoid in the intervening zones is a prolate shape, consistent with the well-developed mineral elongation lineation, and the poorly developed foliation that is not consistently parallel to compositional layers. Solar and Brown (2001a) inferred from this that intervening zones are zones of \( L >> S \) tectonite, and refer to intervening zones as "zones of apparent constriction" (ACZs; Figure 3b).
Figure 4. Rose diagrams of Si-Se rakes measured in thin sections cut perpendicular to foliation and parallel to mineral lineation ($\lambda_1$-$\lambda_2$ plane) from rocks in the central central AFZ, collected at locations across strike (from Solar and Brown, 2001a). d.-h. are from rocks collected at Coos Canyon (STOP 1) at approximately 30 m spacing across strike of compositional layers. The vertical arrow on each diagram shows both the intersection of the foliation with the diagram, and the plunge direction of the lineation in the foliation plane.

Figure 5. Schematic model sections to show the evolution of the CMB shear zone system in western Maine (from Solar and Brown, 2001a). The shaded fields are the same unit at t0 to t4, and represent the folds of the Rangeley stratigraphic sequence (cf. Fig 3b, section A-A') during the development of the structure. At t0 (a), the CMB rocks lay undeformed above basement rocks (stippled field). As the crust shortened, and was contracted inboard (t1, b), the sequence began to buckle asymmetrically to form a ramp-like anticline in the middle crust, detached at the CMB-basement contact. As the system evolved, deformation was localized to tighten folds and strain-harden the sequence (t2, c). Shortening was accommodated more easily within the flattening zones to for zones of flattening strain (AFZs) separated by intervening zones of apparent constrictional strain (ACZs). The alternation between zones corresponds to the areas of tight and open folds, respectively (d and e).
Three-dimensional structure of the CMB shear zone system

Interpretation of the three-dimensional structure of western Maine is based upon the mapped structural pattern and interpretations of various types of geophysical data on the subsurface structure (see Brown and Solar, 1998b, for a discussion). Figure 6 is a lineation-down-NE-plunge perspective block diagram from Solar et al. (1998a) that serves as a summary of the structural system. From the arc described by the ‘straight’ belts (AFZs) in the map plane (concave to the SE) and their strike length (Figure 3), Solar and Brown (2001a) inferred that the AFZs continue to depth. These zones likely converge into a sub-horizontal root zone at approximately 13 km depth (see Figure 3b, section B-B’). Intervening zones (ACZs) are interpreted to narrow and pinch out with depth. Although ACZs likely thicken upward as the ‘straight’ belts narrow to maintain strain compatibility, they also change in three-dimensional geometry from lens-shaped to planar, as reflected in the spatial change from SE to NW from deeper to shallower structural levels. Thus, Solar and Brown (2001a) interpreted the three-dimensional shape of the structure to be listric where dips shallow with increasing depth. The steeper plunge of lineations in the central part of the area of Figure 2, where the foliation is rotated from NE- to N-striking (Figure 3a), may reflect deformation in a restraining bend within the structure where steeper lineations may record a larger component of dip-parallel displacement, which may account for the occurrence of migmatite in this area, reflecting exhumation of deeper crust.

Figure 6. Perspective block diagram of the field area, combining elements of the geological map and sections in Figs. 2 and 3 (after Solar et al., 1998) to show relations between granite plutons and the structural system as described in the text. AFZs are indicated by dashed formlines of fabrics, and ACZs are unornamented. The stratigraphic units are filled with cooler colors whereas the anatectic rocks and basement terranes are ornamented as in Figs. 2 and 3.
MIGMATITE AND GRANITE

The area of excursion is the northern limit of migmatite in the Appalachians, in the Tumbledown and Weld anatectic domains (TAD and WAD, respectively; Figure 2). In western Maine, Solar and Brown (2001b) separate migmatites into stromatic (layered) and heterogeneous (variably-layered) varieties (cf. STOPS 4 and 5; see also Brown and Solar, 1999). These two types of migmatite map into discrete zones, with transition zones between, that alternate across the regional strike, roughly corresponding to ‘straight’ belts (stromatic) and intervening zones (heterogeneous) (Figure 2). Brown and Solar (1999) interpreted both types to have formed at similar structural levels. The structural relations led Brown and Solar (1999) to interpret heterogenous migmatite to be within the cores of regional thermal antiforms (in ACZs; see Figures 2, 3, and 6), flanked by stromatic migmatites (in AFZs). The protolith of the migmatites is interpreted to be rocks of the Rangeley stratigraphic sequence based upon compositional layers that reflect relict bedding, and because the mineral assemblage of sillimanite (fibrolite) + garnet + biotite + quartz + plagioclase + opaque phases ± muscovite ± clinozoisite is consistent with metamorphism of psammitic and pelite at upper amphibolite facies conditions (Solar and Brown, 2001b; Johnson et al., 2003). This is supported by the continuation of the regional structure across the migmatite front (Figure 2, 3, and 6).

The migmatites are separated based upon areal distrubution and regional structure (Solar and Brown, 2001a). Migmatite varies from strongly-foliated metasedimentary rock with a few mm-scale leucosomes per m², in which relict primary structures are preserved, to rocks structurally disrupted by the migmatite formation process (progressive metatexis and increasing volume of leucosome, disruption by apparent flow; diatexis) and schlieric granite. Leucosome density and disruption of relict primary structures both increase across strike from the migmatite front (Solar and Brown, 2001b). In both migmatite types, the common matrix hosting the leucosome is sillimanite (mostly fibrolite), biotite, garnet, quartz, plagioclase, opaque phases (usually ilmenite), and coarse, skeletal muscovite books that cut the fabric, interpreted by Solar and Brown (2001b) to be retrograde. Fibrolite and biotite are the main fabric-forming phases, with fibrolite apparently grown at the expense of primary fabric-forming muscovite to suggest it was produced after muscovite breakdown (Solar and Brown, 2001b). Migmatite leucosomes are discrete to diffuse with a common mineralogy of plagioclase, quartz, muscovite, and locally biotite. The microstructure of leucosomes shows crystal faces and mineral films along grain boundaries that suggest some crystallization from melt, and melt-present formation (Solar and Brown, 2001b).

Based on the petrography, Solar and Brown (2001b) interpret the melt-producing reaction as:

\[ \text{Ms + Pl + Qtz + water} \rightarrow \text{melt + Sil + Bt}, \]

followed closely by:

\[ \text{Ms + Pl + Qtz} \rightarrow \text{melt + Sil + Kfs + Bt}, \]

because these two reactions are closely spaced at low \( P \) (Thompson and Tracy, 1976). At a depth of \( \sim 15 \) km these reactions indicate \( T \) of \( > 700°C \). Given the limited amount of water-rich metamorphic volatile phases that can be stored in rocks at upper amphibolite facies conditions, melting is dominated by the muscovite-breakdown melting reaction (Johnson et al., 2003). In Maine migmatites, the absence of primary muscovite in migmatites, in comparison with the metasedimentary rocks outside the migmatite front where muscovite averages \( \sim 25 \) vol. % in the mode (Solar and Brown, 2001b), and the universal occurrence of sillimanite as a fabric-forming phase with biotite in the migmatites, suggest that the material that hosts the leucosome is depleted of melt, leading Solar and Brown (2001b) to refer to darker host rock that does not form a distinct melanosome as “melt-depleted host rock”. Further, the generally K-feldspar poor nature of leucosomes led Solar and Brown (2001b) to suggest melt has been lost from the migmatites as a whole. These observations are consistent with syntectonic migmatite formation, and consequent syntectonic melt extraction from the anatectic domains. Solar and Brown (2001b) and Johnson et al. (2003) evaluated this postulate in the light of the contemporaneous deformation.

Stromatic-structured metatexite migmatite and associated granites

Approximately half of the exposed migmatite in western Maine is stromatic-structured (layered) metatexite, characterized by a planar structure in which each layer is mineralogically and texturally distinct. This type of migmatite, found mostly in AFZs (Figure 3), is composed of discrete mm- or cm-thick discontinuous sheet-like
bodies of granite (leucosome) separated from medium-colored high-grade metamorphic host rock by dark-colored selvedges (melanosome). The orientation of metatexite layers and mineral fabrics are concordant, as reflected by the consistent steeply dipping orientation of these structures at all scales within each structural zone (Figure 3). At regional scale, the layers are parallel to those of metasedimentary rocks in the same structural zone (Figure 3a, see stereograms).

Leucosomes are trondhjemitic, making up ~3 vol.% of the metatexite at outcrop (Solar, 1999). Millimeter-scale leucosomes range from ~ 1 to 25 cm in length, whereas cm-scale leucosomes are typically ~ 1 to 2 m long; both have low width-to-length ratios. Melanosomes range from 0.1 to 0.6 mm wide, rarely up to 1 mm, and are most conspicuous where in contact with leucosomes. They are composed of > 80 vol.% biotite, accompanied by fibrolite, minor quartz, plagioclase and retrograde chlorite. Biotite, generally ~1 mm long, is clustered with a strongly preferred orientation that defines a foliation parallel to leucosome edges. The intervening host rock layers are 2 to 10 cm thick, being composed of biotite, quartz, sillimanite (mostly fibrolite), garnet, pyrrhotite and/or ilmenite, muscovite (skeletal), and locally plagioclase, tourmaline and clinozoisite. Typically, fibrolite has grown at the expense of primary muscovite in the foliation; the fibrolite forms a fabric in addition to the penetrative biotite foliation, and these sillimanite-biotite folia alternate with quartz-feldspar folia. Fabrics are oriented sub-parallel to fabrics in the metasedimentary rocks outside the migmatite domains down temperature from the migmatite front (Figure 3, see stereograms). Elongate fibrolite aggregates define a steeply plunging lineation visible in the field, and elongate quartz aggregates define a weak sub-horizontal lineation seen only in cut hand specimens and suitably oriented thin sections.

Leucosomes show a 3-D ‘pinch-and-swell’ structure (Solar and Brown, 2001b). A longer wavelength in the sub-horizontal dimension suggests the maximum apparent ‘pinch’ is sub-vertical and down-dip, consistent with kinematic indicators in the metasedimentary rocks (Solar and Brown, 2001a). This triaxial, oblate shape is similar to that defined by the mineral grains in the host rock layers. Centimeter- to m-scale sub-vertical tabular bodies of granite have cut stromatic-structured migmatite at concordant to weakly discordant angles to the planar structures. Many of these granite sheets are composite (Brown & Solar, 1999), and most have a ‘pinch-and-swell’ structure with a longer wavelength in the sub-horizontal direction. Within 1 km along strike in the area of STOP 4, a progressive increase occurs in the proportion of m-scale composite granite sheet to host stromatic-structured migmatite such that the metatexite migmatite becomes disrupted (STOP 4) ultimately to occur only as isolated schollen in granites (STOP 4, south part) to make up a sheeted granite complex. Specimens of these granites (from STOP 4 and north) yielded U-Pb zircon ages of 408 ± 2 Ma and 404 ± 2 Ma, respectively (Solar et al., 1998).

Heterogeneous metatexite and diatexite migmatite and associated granites

A regular planar structure is absent in the remainder of the exposed migmatite in western Maine, an observation that led Solar and Brown (2001b) to refer to these migmatites as “heterogeneous”, found exclusively in ACZs. The orientation of grain-shape fabrics and geometry of leucosomes vary more in these rocks than in the stromatic-structured metatexite migmatite (Figure 3a, see stereograms). Weak foliations and lineations in heterogeneous migmatite are defined by sillimanite (mostly fibrolite) and biotite. There are two types of heterogeneous migmatite, vein-structured metatexite migmatite in the S and SW and diatexite migmatite in the N and NE. Contacts between the two types are gradational over tens of meters in transition zones (Solar and Brown, 2001b). Simply, vein-structured metatexite migmatite (STOP 3) has phlebitic, or veined leucosomes, and this type typically displays meter-scale compositional layers interpreted to be relict from the protolith (Solar and Brown, 2001b). Diatexite migmatite (STOPS 5, 6 and 7), in contrast, is a rock in which the protolith structures are not observed, suggesting destruction by diatexis. Vein-structured metatexite migmatite shows sharp leucosome contacts. In contrast, contacts between leucocratic and melanocratic domains in diatexite migmatite are diffuse and gradational at the cm-scale.

Vein stromatic-structured metatexite migmatite. Centimeter-scale pod- or lens-shaped trondhjemitic leucosomes up to 20 cm long are separated by cm-scale anastomosing darker host layers similar to the melanosomes of stromatic-structured metatexite migmatite. Leucosomes make up ~15 vol.% on outcrop surfaces, and display ‘pinch-and-swell’ structure (Solar and Brown, 2001b). The melanosome host rock contains sillimanite (mostly fibrolite), biotite, muscovite, plagioclase, quartz, pyrrhotite and/or ilmenite, garnet and locally K-feldspar. Sillimanite is found as clots within both muscovite and plagioclase. Biotite (1-3 mm in length), muscovite, and elongate untwinned plagioclase (up to 6 mm long) all show a distinct grain shape fabric, defining a strong moderately dipping foliation (variable dip direction) and weak, moderately-plunging, down-dip lineation.
**Diatexite migmatite.** Diatexite migmatite varies at outcrop from ‘patchy’ leucosome-dominated to biotite-sillimanite-dominated rock, and outcrop to outcrop from schlieren-rich diatexite migmatite to schlieric granite with schollen of vein-structured migmatite and non-migmatitic calc-silicate-rich psammite (Solar and Brown, 2001b). Most types are characterized by a discontinuous, weakly-defined foliation of variable attitude. Leucosomes and leucocratic domains make up ~9 to 15% of diatexite migmatite outcrop surfaces, and are generally uniformly distributed. Discrete leucosomes appear as cm-scale quartzo-feldspathic mineral domains that vary from diffuse to sharp at their margins, and have sub-equant shapes in pavement outcrops, 1 to 2 cm in diameter, but are elongate down-dip of the sillimanite fabric in the host rock, with lengths up to 20 cm (Solar and Brown, 2001b; see STOP 5). Thus leucosomes tend to be rod-shaped, plunging moderately-to-steeply ENE, sub-parallel to both the weakly-defined mineral lineation in the host rock and the strongly-defined mineral lineation in the metasedimentary rocks outside the migmatite front (Solar and Brown, 2001b). These fabrics define a triaxial, strongly prolate shape similar to that of the metasedimentary rocks in the same structural zones (ACZs; Figure 3).

Diffuse domains of the host rock interfinger with the leucosome at the mm-scale, and consist of biotite, sillimanite (mostly fibrolite) and garnet, with accessory muscovite, plagioclase, quartz, pyrrhotite and/or ilmenite, and K-feldspar (Solar and Brown, 2001b). Fibrolite has grown at the expense of muscovite and untwinned plagioclase to form a younger generation of foliation-forming minerals. Different proportions of these minerals account for the gradual variation of diatexite migmatite from more quartzo-feldspathic (leucocratic) to more ferromagnesian (melanocratic) types (Solar and Brown, 2001b), both seen at STOP 7. In the extreme case, either schlieric granite is formed (leucocratic diatexite migmatite; STOP 8), or the mineral assemblage is dominated by biotite, sillimanite and garnet with less than 10 vol.% (plagioclase and quartz) to give the rock a melanocratic appearance (STOP 7). Biotite, muscovite and elongate untwinned plagioclase show a preferred grain-shape fabric that defines a weak, moderately-to-steeply dipping foliation and strong, moderately ENE-plunging, down-dip lineation.

Most outcrops of diatexite migmatite are cut by meter-scale, cylindrical granite bodies (STOP 7). These granite bodies are elongate subparallel to the mineral lineation in diatexite migmatite, and to the rod-shaped leucosomes (see Brown and Solar, 1999, for a discussion). Entrained blocks of strongly foliated biotite-garnet schist are found in the interior of the granite cylinders. The granite cylinders lack a fabric, except proximal to the margin of these blocks. An example of the granite cylinders (from STOP 7) yielded an age of ca. 400 Ma (Solar, Brown and Tucker, unpublished data).

**The Phillips pluton**

Immediately NE of the WAD (Figure 2) is the coeval Phillips pluton (ca. 404 Ma; Pressley and Brown, 1999), that is interpreted to be hemi-ellipsoidal with long dimension parallel to the regional moderately NE-plunging lineation (Brown & Solar, 1998b; 1999; Pressley & Brown, 1999). It is located in an ACZ, similar to the diatexite migmatites, and it has a similar geometry to the smaller-volume cylinders of granite found in heterogeneous migmatites (STOP 7). These observations have been used to suggest a relationship between structure, granite ascent and emplacement (Brown and Solar, 1999). The geochemistry of common leucogranite (~95% by area) from the Phillips pluton has been interpreted to reflect an origin by melting after muscovite dehydration of a source with geochemical characteristics similar to the metasedimentary rocks of the CMB (Pressley and Brown, 1999). The remaining ~5% by area is granodiorite interpreted to reflect an origin by biotite dehydration melting of a source geochemically similar to “Avalon-like” rocks (Figure 7a; see Fig. 1). The bodies of granite found in the migmatites do not possess this latter component (Figure 7a). For these reasons, Solar and Brown (2001b) evaluated what relation exists between the migmatites, the smaller-volume granites in the migmatites and the common leucogranite of the Phillips pluton.

**The Mooselookmeguntic Igneous Complex (MIC)**

The MIC was previously mapped as three petrographically distinct plutonic bodies: the Mooselookmeguntic, Umbagog and Adamstown plutons (Figure 2; Moench et al., 1995). Considering the information currently available, Tomascak et al. (2005) grouped the MIC into two principal types of rock: biotite granodiorite to quartz monzodiorite (the monzodiorite suite), and biotite and two-mica granite to granodiorite (granite). Biotite monzodiorite togranodiorite enclaves, petrographically similar to rocks of the monzodiorite suite, occur in the granite. The
monzodiorite suite dominates the southwestern portion of the complex (previously the Umbagog pluton). The enclaves in the MIC granite occur as multi-meter-scale blocks with petrographic character similar to the monzodiorite suite. Portions of both the monzodiorite suite and the granite lie to the north and south of the BHB-CMB contact (Figure 2). The Adamstown pluton in the north (“A” in Figures 2 and 3) is split from the MIC based upon penetrative solid-state fabrics not found in the rest of the body. Solar et al. (1998) reported crystallization ages for two MIC rocks: a U-Pb zircon age of 389±2 Ma for a granodiorite enclave (referred to as “biotite granite”), and a
concordant U-Pb monazite age of 370±1 Ma for a granite (“two-mica leucogranite”). Tomascak et al. (2005) reports two samples from distinct parts of the MIC monzodiorite suite yield identical ages, slightly older than the granite (one is at STOP 10, see Figure 2 for location; ca. 377 Ma on four concordant to 3% discordant U-Pb zircon fractions). This is equivalent to the U-Pb concordia upper intercept age of 378±2 Ma published by Moench and Aleinikoff (2002) for an alkali gabbro “border facies” of the monzodiorite suite.

Other plutonic rocks

The Redington pluton in the CMB, is dominantly a porphyritic biotite granite that has its northern boundary coincident with the tectonite zone that separates the BHB and CMB (Figure 2). Tomascak et al. (2005) report the upper U/Pb intercept age of four zircon fractions is 408±5 Ma (from Solar et al., 1998), and two concordant zircon fractions that suggest an age of ca. 406 Ma is more accurate. The Sugarloaf pluton (2 specimens; Figure 2; Tomascak et al., 2005) occupies a similar position to the Redington pluton, but is apparently mingled felsic-mafic rocks that bear strong resemblance to large portions of the more mafic part of the adjacent Flagstaff Lake intrusive complex. Rocks are dominantly gabbroic with subordinate felsic rock pods. The Lexington composite pluton comprises northern, central and southern portions, based partly on interpretation of the 3-dimensional character of the pluton (Brown and Solar, 1998b, 1999), occurring east along strike from the main body of the MIC (Figure 2). Rock types are primarily granite and granodiorite, based on normative composition, with gabbro enclaves. All granite samples contain biotite and some contain two igneous micas. One U-Pb zircon age determined by Solar et al. (1998; 404±2 Ma) is from the central portion of the Lexington pluton, and an identical concordant zircon age from the southern lobe is reported by Tomascak et al. (2005; ca. 403 Ma). Tomascak et al. (2005) also report a younger age from the northern lobe (ca. 365 Ma, two concordant U-Pb zircon fractions).

GEOCHEMISTRY

Major- and trace-element and isotope geochemistry of the granitic rocks is summarized in Figures 7 and 8, and in Pressley and Brown (1999; Figure 7a), Solar and Brown (2001b) and Tomascak et al. (2005; Figures 7b and 7c). See Brown and Solar (2001b) and Johnson et al. (2003) for summaries of the geochemistry of the migmatites and surrounding metasedimentary rocks (Figure 8). In regard to the migmatite geochemistry (Figure 8), textures show that biotite was apparently stable on a regional basis, indicating that the biotite dehydration melting reaction was not crossed during regional metamorphism. Solar and Brown (2001b) constrained the processes involved in leucosome and pluton formation by comparing whole-rock and migmatite component geochemistry versus experimental melt and residuum compositions (Figure 8). Based on the structural evolution of western Maine (Solar and Brown, 2001a), and the field relations and geochemistry of the metatexite and diatexite migmatites, and granites (Figure 8), Solar and Brown (2001b) proposed a model of progressive separation of melt and residue during deformation of the CMB metasedimentary succession. Based upon field relations and geochemistry (Figure 8), metasedimentary rocks similar in composition to those of the CMB are inferred to be the protolith for the migmatites. The depleted nature of metatexite and diatexite migmatites is not balanced by the smaller-volume granites alone (Figure 8). The Harker plots in Figure 8 (a. to d.), and the K₂O v. Na₂O plot (Figure 8e) indicate that differential separation of melt from residual solid material was not the sole petrogenetic process involved in producing the variation observed. If mass balance is preserved at all scales during melting, melt segregation and transfer, and crystallization of the melt, then the processes involved may be tracked in the ternary plot K – (Fe* + Mg + Ti) – (Na + Ca) (Figure 7f). In such a plot, biotite lies along the edge (Fe* + Mg + Ti) – K; it represents the major residual phase. The feldspar join is represented by the edge (Na +Ca) – K, close to which lie melts produced from crustal protoliths. Residual compositions will be displaced from the field of metasedimentary rocks toward (Fe* + Mg + Ti) – K, whereas leucosome and granite compositions will trend toward the feldspar join. In Figure 8f, migmatite compositions are displaced toward the (Fe* + Mg + Ti) – K edge in comparison with the CMB metasedimentary rock field, whereas granites and leucosomes are weakly clustered toward the feldspar join, closer to the (Na + Ca) apex than the K apex. The three specimens from the block of biotite-garnet schist (from STOP 7) plot between biotite and plagioclase, but much closer to biotite, and are displaced from the CMB metasedimentary rock field toward garnet, reflecting the dominance of these phases in the rock (Solar and Brown, 2001b). The whole rock chemistry of the schlieric granite specimen of STOP 8 plots between the field of CMB metasedimentary rocks and the feldspar join, suggesting the specimen is enriched in the feldspathic components compared to the protolith composition. The migmatite leucosomes are seen to define an array of compositions from melt-dominated, plotting close to the MBS melt compositions in the experiments, to cumulate-dominated, plotting close to a cumulate composed of ~80% plagioclase and ~20% biotite. Variable loss of a K-rich liquid is implied (Solar and Brown, 2001b). In contrast, the
Figure 8. a.-e. Oxide compositions of CMB metasedimentary rocks, migmatites and granites (from Solar and Brown, 2001b). Lines connect migmatite components and their whole rock compositions. MBS is muscovite-biotite schist. f. Major-element modeling of melt-residuum separation during migmatization of CMB rocks (from Solar and Brown, 2001b). Tick marks on the line connecting plagioclase and biotite are biotite content.
smaller-volume granites define a triangular field between the MBS melts and the cumulate join between plagioclase and biotite, with the leucosome array as the bottom edge and extending along the cumulate join from ~20 to ~35% biotite. The common leucogranite of the Phillips pluton and one other granite specimen crystallized from a K-enriched (evolved) liquid in comparison with the MBS melts. Based on data in Figure 8f, Solar and Brown (2001b) inferred that none of the smaller-volume granites has a melt composition, but instead they have cumulate compositions, each with a variable amount of cumulate material and residual, fractionated liquid, possibly with some residual biotite and plagioclase; a K-rich liquid has been partially lost from these rocks.

**Rare earth elements.** Assuming the CMB metasedimentary rocks were the protolith for the TAD and WAD migmatites, and the source for the granites, Solar and Brown (2001b) normalized to REE contents to one representative metasedimentary rock (Figure 9; see Solar and Brown, 2001b, for discussion). Examination of the protolith normalized REE patterns shows that whole rock compositions of the migmatites resemble the protolith closely, but are generally enriched in total REE, likely reflecting melt loss at the scale of the hand specimen relative to the protolith. All granites have REE patterns that lie well below the protolith composition, and all but granite 95-121 have positive Eu anomalies. One granite specimen from the diatexite shows LREE-enrichment relative to the protolith. The melt-depleted host rock in the migmatites is REE-enriched, with the exception of one specimen that shows MREE depletion. Leucosomes show variable total REE depletion. Schlieric granites are slightly depleted in the REE, although 95-215 (STOP 8) shows slight HREE enrichment. Consistent with the cumulate hypothesis,

---

**Figure 9.** REE concentrations of leucogranite, migmatite and migmatite components normalized to a representative metasedimentary rock (represented by the heavy dashed line) (Fig. 12 in Solar and Brown, 2001b).
migmatite leucosomes and smaller-volume granite commonly exhibit positive Eu anomalies. The common leucogranites from the Phillips pluton are REE depleted in a similar fashion to the smaller-volume granites, and are closely similar to the pattern of granite 95-121. Complementing the suite of granites, the three analyses from the block of biotite-garnet schist (from STOP 7) show strong total REE enrichment, with the exception of the LREE of one specimen (b) that may reflect the higher proportion of garnet in that specimen (Solar and Brown, 2001b).

DISCUSSION: TRANSPRESSIVE DEFORMATION AND GRANITE ASCENT

Solar and Brown (2001a) interpreted the CMB shear zone system to have developed as a thrust system during dextral-transpressive deformation in response to Early Devonian (Acadian) oblique convergence (Figure 5 and 6). The deformation was partitioned between zones of apparent flattening strain (AFZs, the ‘straight’ belts) and zones of apparent constrictional strain (ACZs, the intervening zones). This partitioning was a consequence of the serial development and subsequent progressive modification of thrust ramps in the Silurian to Early Devonian stratigraphic succession above the Avalon-like basement (Figure 3b, section B-B’), a model that is consistent with the tectonic model for the Acadian orogeny of Bradley et al. (1998), which primarily uses the crystallization age of plutons to suggest that the Acadian orogenic front migrated inboard during the Devonian. A relative increase in volume in the ACZs as the AFZs encroached on them with progressive accommodation of strain is suggested as a mechanism to maintain the prolate fabrics (Solar and Brown, 2001a). The model is consistent with models of transpression zones, where strain is distributed heterogeneously into belts of contrasting finite deformation (Robin and Cruden, 1994), and in which the transpression zone is stretched along strike (Dias and Ribeiro, 1994).

Regarding the relation of melt flow in this structural system, Solar and Brown (2001b) noted that melt loss from metatexite and diatexite migmatites is implied by the K-feldspar poor nature of the leucosomes, an idea that is supported by the residual chemistry of the host rock, and the lack of mass balance of trace elements between migmatite components to suggest open-system behavior at the scale of hand specimens. The ‘pinch and swell’ structure of granite sheets (e.g., at STOP 4) is consistent with melt flow during deformation and weak strain during or after emplacement. The strong correlation of regional fabrics across the migmatite front (field ‘solidus’) in the same structural zone (Figure 3a) supports the interpretation that migmatite formation occurred while the rocks were accommodating strain. Solar and Brown (2001b) postulated that leucosomes and smaller-volume granites record evidence of syntectonic melt flow within and through the migmatite, and that granite in plutons apparently outside the migmatites at the level exposed represent evolved melt that escaped syntectonically from a similar source to the migmatites exposed currently in the TAD and WAD.

Given the steep fabric orientation, Solar and Brown (2001b) interpreted the retrograde muscovite and chlorite to be a consequence of buoyancy-aided fluid flow parallel to the rock fabrics. This fluid is likely to have been derived from crystallizing melts in the migmatites (Figure 10). As melt crystallizes in the deforming rocks, liberated H2O may promote melting in adjacent units at suprasolidus conditions or retrogression at subsolidus conditions. No influx of water-rich volatile phase is necessarily implied or required. Solar and Brown (2001b) postulated that this is the principal cause of regional syntectonic retrogression of staurolite and andalusite in rocks outside the migmatite front.

SUMMARY AND CONCLUDING STATEMENT

In Maine, metamorphism and granite crystallization are contemporaneous; and syntectonic magma ascent was controlled by deformation and development of strain fabrics. Mineral assemblages and geochemical data are consistent with melting after muscovite-dehydration reactions, and suggest that migmatite leucosomes and smaller volume granites represent cumulate rocks (+/- residual material and some retained fractionated melt) that complement common leucogranite in the Phillips pluton. Metatexite and diatexite migmatite residual geochemistry relative to the metasedimentary rocks, and the ‘pinch and swell’ structure of granite sheets, are consistent with melt loss from those rocks, perhaps driven by the accommodation of deformation. Schlieric granites suggest some melt redistribution prior to melt loss, in this case leaving a more felsic cumulate than the residual migmatites.

Migmatites preserve evidence in cumulate leucosomes of the flow network that drained these rocks of an evolved melt. We postulate that migmatites similar to those exposed represent the source of common leucogranite in the Phillips pluton. Thus, the Phillips pluton may be connected at depth to granites similar to those found in the
migmatites of the TAD and WAD, in the manner described by Brown and Solar (1999) where the heterogeneous migmatites and granites in ACZs formed in the cores of thermal antiforms developed during regional contraction (Figures 2, 3, 6 and 10). However, the relation between migmatite leucosomes, smaller-volume granites and leucogranite in plutons is not straightforward and involves multiple processes. It is naïve to expect that migmatite leucosomes and leucogranite in plutons should show simple melt compositions without entrained residue or modification by fractional crystallization. This is likely given the common association of migmatite leucosomes and smaller-volume granites with cumulate compositions, whereas the leucogranite of adjacent plutons is consistent with the putative liquid lost from these rocks. The classically popular notion that migmatites represent ‘failed’ granites has been reconsidered in the light of multiple syntectonic processes (see Sawyer, 2008, and references therein).

ROAD LOG

Mileage

0.0 Begin at the Coos Canyon Rest Area parking lot at the intersection of Maine Rt. 17 and Weld Rd. in the town of Byron adjacent to the Coos Canyon Campground, ~ 6 miles north of Roxbury, Maine.

STOP 1: PERRY MOUNTAIN FORMATION (SP), COOS CANYON, BYRON, MAINE
(UTM 4842161N, 368525W)

See Figure 11 for location. Plastically deformed meta-turbidite (Silurian; centimeter-scale alternating layers of pelite and psammitic) similar to that at STOP 2 and 3. Layers and mica foliation are both steeply ESE-dipping. There is a moderately to steeply NNE-plunging penetrative bladed muscovite and acicular biotite lineation. Fabrics are viewed best using mutually perpendicular outcrop surfaces. Pelite layers are staurolite bearing throughout, locally andalusite bearing (pseudomorphs), and show distinct centimeter-scale ‘P’- and ‘Q’-domains defined by matrix minerals (Solar and Brown, 2001a). The long-axis orientations of staurolite porphyroblasts were measured at two locations here, and those data are discussed in Solar and Brown (1999; see their Figures 4b and 4c).

This outcrop is within the central zone of apparent flattening strain (Brown and Solar, 1999; Solar and Brown, 2001a). Because of the extent of this outcrop, most structures common to rock within the zones of apparent flattening strain are found here, including the restricted variation of foliation and lineation.
Figure 11. Topographic, simplified geologic and location map for the area of STOPs 1 and 2, and optional stops 2a and 2b. Structural data are taken from Solar (1999). Symbols are consistent with those on Fig. 2; Oq is Quimby Formation, Sr is Rangeley Formation, Sp is Perry Mountain Formation, Ss is Smalls Falls Formation. g is granite and gd is granodiorite of the Mooselookmeguntic pluton in the SW. The location of Maine Rt. 17 is highlighted across the map (along Swift River. See road log for details of STOP locations and geology).
Figure 12. $S_1$-$S_3$ for garnet from a 133 m traverse of Coos Canyon (after Steely, 2001). The plane of the diagram is parallel to the thin sections, all perpendicular to foliation, the intersection of which is the vertical line. The bimodal distribution defines two distinct populations of garnet relative to ‘P’ domains and ‘Q’ domains. The different populations suggest differences in the competency between the two matrix domains (Steely, 2001).

attitudes (see Figure 3). Other structures include meter-scale domains of meter-scale tight folds of the compositional layers with transecting foliation. Biotite fish are apparent locally along surfaces eroded sub-parallel to the mica lineations. Brittle structures are also present including left-stepping en echelon quartz-filled gashes (restricted to the psammite layers), and dextrally offset quartz-filled shear fractures with larger apparent offset in the pelite layers. Foliation-parallel surfaces are best to view the apparent preferred fabrics, including the apparent alignment of staurolite and andalusite. Staurolite is replaced within meter-scale domains across strike at this location. Some andalusite has pink cores remaining after partial replacement.

Three of the Rose diagrams in Figure 4 (d, f and h; from Solar and Brown, 2001a) are Si-Se rakes illustrated by porphyroblasts in rocks at this locality. Rakes were measured for each porphyroblast in the three thin sections cut from rocks collected across strike. These data show a distinct and consistent obliquity, the breadth of which Solar (1999) attributed to the diachronous growth histories of the main porphyroblasts, garnet vs. staurolite, garnet having more strongly discordant inclusion trails relative to the matrix foliations in all thin section views. Solar and Brown (2001a) interpreted these data to reflect a record of punctuated (inter-kinematic) porphyroblast growth during progressive deformation/reorientation and recrystallization of matrix fabrics. Steely (2001) confirmed a difference in obliquities between porphyroblast phases, but also found a difference between porphyroblasts located in ‘P’- and ‘Q’-domains (Figure 12), and between porphyroblasts viewed in lineation-parallel vs. lineation perpendicular thin sections.

Mileage

0.0 Turn right to go north on Rt. 17.
1.2 Intersection of Rt. 17 and Garland Pond Road (road to Little Ellis Pond; Figure 11).
2.0 Dirt access road on left along West Branch of Swift River.
2.3 Park on right side of road at the turnout where the Swift River is joined by the East Branch of Swift River (Figure 11). The outcrop for STOP 2 is in the Swift River, west bank, accessed via the path to the river.

STOP 2: RANGELEY FORMATION (SR), SWIFT RIVER, UPSTREAM FROM (NORTH OF) COOS CANYON, BYRON, MAIN (UTM 4955439N, 368599W)

See Figure 11 for location. Plastically deformed meta-turbidite (Silurian; centimeter-scale alternating layers of pelite and psammite) similar to that at STOP 1 with the notable exception that the variation of compositional layer attitudes is large, not ‘straight’ by comparison with layers at Coos Canyon (see Figure 3). This STOP, along with Optional Stops 2a and 2b (Figure 11), are listed here to demonstrate the structural variation of the intervening zones between zones of apparent flattening (e.g., STOP 1; Coos Canyon is an extensive outcrop, but STOPS 2 and 2a are only small exposures). Compositional layers are in open folds that penetrate the outcrop. Layers and mica foliation are not sub-parallel, and foliation here is only weakly developed (cf. STOP 1). However, as at Coos Canyon, there is a moderately NNE-plunging penetrative bladed muscovite lineation, but unlike at Coos Canyon, there is a biotite-pull-apart lineation.
that is parallel to the muscovite lineation. Pelite layers are staurolite-bearing (partially replaced). These fabrics illustrate the apparent constriction recorded here, and in the intervening zone in general (i.e. zone of apparent constrictional strain). This rock is just NW of the central zone of apparent flattening strain (Brown and Solar, 1999; Solar and Brown, 2001a), within the zone of apparent constriction, but is transitional to the two structural zones in that foliation is better developed here relative to typical rocks in the zones of apparent constrictional strain (just north, see Figure 11 for example optional stops 2a and 2b).

*----------------- Optional Stops 2a and 2b ----------------*

** STOPS 2a and 2b are other examples of rocks similar to those at STOP 2, typical of rocks in the intervening zone to the west of the central 'straight' belt (Figures 2 and 3). These stops have their own road log starting at STOP 2. Return to STOP 2 to resume the trip road log.

Optional stop mileage

0.0  Continue north on Rt. 17.
0.6  Park on side of Rt. 17 just north of three houses located on the left (west) side of the road (Figure 11). The outcrop for STOP 2a is pavement in the Swift River, upstream of the houses.

STOP 2a (optional): RANGELEY FORMATION (SR), SWIFT RIVER, HOUGHTON, MAINE
(UTM 4956299N, 368836N)

See Figure 11 for location. Plastically deformed meta-turbidite (Silurian; centimeter-scale alternating layers of pelite and psammite) similar to that at STOP 2. Note the mesoscopic biotite-quartz pull-aparts with long dimension parallel to the muscovite elongation lineation (moderately NNE-plunging). Compositional layers define an open fold, and mineral lineations are sub-parallel to the fold hinge line, but are more steeply NNE-plunging. Again, foliation here is only weakly developed (cf. STOP 1).

Optional stop mileage

0.6  Continue north on Rt. 17.
1.0  Bridge over West Branch Swift River (Figure 11).
1.3  Bridge over Swift River (Figure 11).
1.8  Intersection with road to Angel Falls (on left; Figure 11). Continue north on Rt. 17.
3.0  Park in dirt lot on the east side of the road just before (south of) the bridge over Swift River (Figure 11). Walk upstream (east, right) through the woods via the dirt road, and along the path to the powerline. Continue on the path across the powerline, and descend along the path afterward, down to the river for STOP 2b.

STOP 2b (optional): RANGELEY FORMATION (SR), SWIFT RIVER GORGE, HOUGHTON, MAINE
(UTM 4960186N, 370476W)

See Figure 11 for location. Plastically deformed meta-turbidite (Silurian; centimeter-scale alternating layers of pelite and psammite) similar to that at STOPS 2 and 2a. Compositional layers are variable in moderately NE dip. Pelite layers are staurolite bearing throughout, and the replacement of the staurolite is also extensive. Note the “new” staurolite growth within the staurolite pseudomorphs. This “new” growth is due to thermal aureole effects of the nearby Mooselookmeguntic pluton (west of this location; Figures 2 and 11, and STOPS 9 and 10). The long-axis orientations of the pseudomorphs were measured here, and those data are discussed in Solar and Brown (1999; see their Figure 4h).

*----------------- End of Optional Stops ------------------*

Road log continues from STOP 2.
Milestone

2.3 Turn around and return south on Rt. 17.
4.6 Coos Canyon.
7.8 Meter-scale sheets of biotite granite along east bank of the Swift River.
8.7 Roxbury town center.
9.3 Outcrops of diatexite migmatite in Swift River (STOP 7).
9.5 Intersection with Walker Brook Road on left (to STOP 8).
10.9 Outcrops of diatexite migmatite in Swift River (STOP 6).
12.0 Outcrops of diatexite migmatite in Swift River (STOP 5).
12.3 Intersection with Rt. 120 access road at Frye, ME (the road to STOPS 4, 9 and 10).
13.7 Roadcuts of diatexite migmatite.
14.8 Black Bridge Road. (Outcrops in the Swift River under the bridge are stromatic-structured metatexite migmatite.)
15.8 Turn right into “Mexico Recreation Park” and continue through the park, past the athletic fields to the parking area at the far northwest end (close to the Swift River which flows along the west end of the park). The outcrops for STOP 3 are in the river as pavement exposures.

STOP 3: VEIN MIGMATITE (HETEROGENEOUS MIGMATITE), SWIFT RIVER, MEXICO, MAINE
(UTM 4937207N, 375839W)

Pod-shaped (vein) leucosomes in a stromatic-structured metatexite migmatite that is plastically deformed. We interpret this as the former metasedimentary rock similar to that of STOPS 1 and 2. This exposure is large enough to see the similarities between this rock and the non-migmatite stratigraphic sequence (the centimeter-scale alternating layers of pelite and psammite in the Perry Mountain Formation, at Coos Canyon; STOP 1). Vein leucosomes and melt-depleted host rock (interpretation based on geochemical data discussed in Solar and Brown, 2001b) are restricted to the meta-pelite layers, except where they are discordant to layers and formed within right-stepping en echelon shear fractures across the foliation, and where they define left-stepping en echelon ptygmatic folds, also across the foliation (both suggest right-lateral shear along the compositional layers). Compositional layers and foliation are sub-parallel and dip moderately ENE. Foliation in the meta-pelite domain is defined by penetrative fibrous sillimanite and biotite. This rock is within the domain of heterogeneous migmatites in the southern part of the TAD (Figure 2). Structurally, this rock is typical of vein-structured metatexite migmatite within the intervening zone between the western and eastern limbs of the central zone of apparent flattening, but is located near the transition zone between the heterogeneous and stromatic-structured migmatite (see STOP 4) (Solar and Brown, 2001a, 2001b). This outcrop is used by Solar and Brown (2001b) as evidence that the Silurian metasedimentary rocks are the protolith for the migmatites because of the structural and geochemical similarities between the rock types.

Milestone

15.8 Return to Rt. 17.
16.3 Left turn on Rt. 17 to return north to Roxbury.
17.8 Pass intersection with Black Bridge Road on left. Continue north on Rt. 17.
20.4 Turn left on access road to Rt. 120 in Roxbury (Frye).
20.5 Bridge over Swift River.
20.6 Turn right onto dirt logging road that goes north into the woods (on left side of a house). This logging road is visible on the satellite photograph in Figure 13. Pass the driveway for the house. This is an active logging road. Limit your speed to 25 mph, and yield to logging vehicles.
20.7 Pass through logging road gate and continue north on the logging road as it follows Swift River upstream.
22.3 Left-hand curve in road (away from Swift River at STOP 6; Figure 13).
22.4 Bridge over Philbrick Brook (Figure 13).
22.9 Bridge #2.
23.1 Pass intersection with dirt road to left (Figure 13). Continue straight.
23.3 Bridge #3.
23.5 Powerline (Figure 13).
Figure 13. Topographic, aerial and locations maps with simplified geology for the area of STOPs 4 through 8. Structural data are taken from Solar (1999). Age of the Bt granite from Solar et al. (1998). Symbols are consistent with those on Figs. 2 and 11; g is granite. See the road log for details of STOP locations/geology.
24.3 Bear left and intersection with road to Bunker Pond (Figure 13).
24.9 Bridge #4.
25.3 Bridge #5.
25.5 Bridge #6
25.6 Bear right to go east on main logging road (Figure 13), and cross the powerline for a second time (Fig. 13).
26.1 Turn right onto dirt logging road (Figure 13) and cross over bridge immediately.
26.4 Continue straight onto narrower logging road (do not turn left to go down to the river).
26.7 Park in clearing on right (Figure 13). Walk across the road and walk into the woods towards the Swift River. You will find a trail before reaching the river. Follow this trail to the right. There are two outcrops for this stop, a steep portion almost immediately accessible after finding the trail, and an outcrop of stream pavement farther along the trail, just before the right-hand curve in the river and trail.

STOP 4: STROMATIC-STRUCTURED METATEXITE MIGMATITE AND SUB-CONCORDANT METER-SCALE LEUCOCRANITE SHEETS, SWIFT RIVER, ROXBURY, MAINE (UTM 4947726N, 373045W)

See Figure 13 for location. Stromatic-structured metatexite migmatite formed during plastic deformation of metasedimentary rock. Locally, alternating layers of pelite and psammite similar to that in the non-migmatite stratigraphic succession are apparent, to suggest the protolith of the metatexite migmatite was rocks similar to the Rangeley stratigraphic sequence. Mica foliation, leucosomes and melt-depleted host rock domains [interpretation based on geochemical data shown in Figure 9 and discussed in Solar and Brown (2001b)] are all sub-parallel, all steeply E-dipping. Granite sheets are concordant to sub-concordant, and display no record of internal plastic strain, but are pinched-and-swelled to suggest deformation in the host rock continued after intrusion. The sheets have a biotite foliation parallel to the structure of the sheets. This rock is within the central zone of apparent flattening strain (Figures 2, 3 and 13), along strike from Coos Canyon (STOP 1; Figure 11). The thickest granite sheet yielded a concordant U-Pb zircon crystallization age of 408 ± 2 Ma (Solar et al., 1998). Granite-migmatite relations in the steep (south) part of the outcrop suggest that progressive sheeting of granite constructed kilometer-scale lens-shaped plutons in the migmatite domain (see Figures 2, 3 and 13). Local outcrops are meter-scale granite sheets only, consistent with this interpretation (see Figure 13, top for example location). The outcrop of stream pavement was mapped at the 1:24 scale by Chmura (2001).

Mileage

26.7 Return to vehicles and return to Rt. 17. 33.0 Turn left to go north toward Roxbury (Figure 13). Look for Pineview Cemetery on the right.
33.3 Pull over on the right and park just north (past) of the cemetery. The outcrops for STOP 5 are in the Swift River across the street (west of Rt. 17, Figure 13). Descend down the steep grade from the road to the stream pavement.

STOP 5: DIATEXITE (HETEROGENEOUS) MIGMATITE AND SUB-CONCORDANT CENTIMETER-SCALE LEUCOCRANITE CYLINDERS, SWIFT RIVER, SOUTH OF ROXBURY (FRYE), MAINE (UTM 4942666N, 374467W)

See Figure 13 for location. The diatexite migmatite is residual in composition similar to that of the stromatic-structured migmatite (STOP 4), but in contrast, leucosome structures are generally rod-shaped in the diatexite migmatite with long dimensions plunging steeply E, similar to the mineral fabric. This rock is typical of all rocks in the northern TAD where it is within a zone of apparent constrictional strain (Figure 2; Solar and Brown, 2001a, 2001b). West-dipping surfaces display leucosome rod-shapes end-on (cross section), whereas steeply south-dipping surfaces are sub-parallel to the long dimension.

Mileage

33.3 Continue north on Rt. 17 toward Roxbury.
34.6 Park on the right side of the road (opposite side of the road from the river) just across from a private dirt parking lot located along pavement outcrops in Swift River. The outcrops for STOP 6 are in the river here, but there is an outcrop for this stop on the east side of the road.

STOP 6: DIATEXITE (HETEROGENEOUS) MIGMATITE, THREE-POOLS (A.K.A. SWIFT RIVER FALLS), SWIFT RIVER, SOUTH OF ROXBURY, MAINE (UTM 4945237N, 374010W)

See Figure 13 for location. Diatexite migmatite that is residual in composition [interpretation based on geochemical data shown in Figure 8 and discussed in Solar and Brown (2001b)] similar to that of STOP 5. Again, leucosome structures are generally rod-shaped with long dimensions plunging steeply E similar to the mineral fabric. This rock is within the west margin of the intervening zone between the two limbs of the central zone of apparent flattening strain. The outcrop in the River is transitional to stromatic-structured metatexite migmatite located to the west (just across the river in the woods; Figure 13) and homogeneous diatexite migmatite located on the east side of Rt. 17 (poorly exposed across from the parking lot). The diatexite migmatite on the east side of Rt. 17 is typical of all rocks in the northern TAD within the intervening zone of apparent constrictional strain (Figure 2; Solar and Brown, 2001a, 2001b).

Mileage

34.6 Continue north toward Roxbury (Figure 13).
36.3 At just before the antique shop at the south end of Roxbury town, pull over on the left carefully where the Swift River returns to the roadside, at a metal gate on a dirt road that follows the river (or pass the gate and turn around and park at the gate). The outcrop for STOP 7 is in the river.

STOP 7: DIATEXITE (HETEROGENEOUS) MIGMATITE AND SUB-CONCORDANT METER-SCALE CYLINDERS OF LEUCOGRANITE, SWIFT RIVER, ROXBURY, MAINE (UTM 4946963N, 374154W)

See Figure 13 for location. Meter-scale cylindrical granite bodies within diatexite migmatite. The diatexite migmatite is residual in composition (interpretation based on geochemical data discussed in Solar and Brown, 2001b) suggesting melt extraction. A meter-scale residual block of diatexite just north of center of the exposure is the “block of Bt-Grt schist” of Figure 8 and is taken to represent extreme melt depletion in the migmatite domain. Meter-scale structures are pipe-shaped with their long dimensions plunging steeply E similar to the mineral fabric. This rock is within the intervening zone between the two limbs of the central zone of apparent flattening. Brown and Solar (1999) suggest the cylinders represent transfer structures for melt that supplied plutons above (Solar and Brown, 2001b). The outcrop is mapped at 1:24 scale (Chmura, 2001).

Mileage

36.3 Return south on Rt. 17 (south of Roxbury town).
36.6 Turn left onto a dirt logging road (Figure 13) called locally “Walker Brook Rd.”, and head uphill into the woods on the east side of Rt. 17. Please exercise caution on this road; it is active.
36.7 Turn left at the intersection to go north on a logging road (Figure 13).
38.3 At just after a sharp right turn, pull over at the pavement outcrop along the right side (Figure 13).

STOP 8: SCHLIERIC GRANITE IN THE TRANSITION ZONE BETWEEN STROMATIC-STRUCTURED METATEXITE MIGMATITE AND DIATEXITE MIGMATITE, NORTHEAST OF ROXBURY, MAINE (UTM 4948786N, 374079W)

See Figure 13 for location. Cumulate granite with a penetrative steeply E dipping schlieric fabric. The fabric is concordant with the stromatic migmatite structure, but this rock is located within the transition zone between stromatic migmatite to the west (downhill) and diatexite to the east (uphill) (Solar and Brown, 2001b). This rock includes schollen of calc-silicate rich metasedimentary rock that have apparently boudined and rotated during what was probably en masse melt-accommodated granular flow of this crystal-rich magma. See Figure 8 for geochemical data from rocks from this outcrop.
Mileage

38.3 Return back to Rt. 17.
40.0 Turn left on Rt. 17 to go south.
42.8 Turn right onto the access road for Rt. 120.
43.0 Pass intersection with dirt logging road to STOP 4.
43.1 Right turn to go west on Rt. 120.
48.6 Intersection of Rt. 120 and Roxbury Notch Road.
50.1 Ellis River outcrops of granite of the Mooselookmeguntic Igneous Complex.
50.4 Black Brook.
51.9 Right turn at T-intersection to continue on Rt. 120.
52.1 Turn right (north) onto S. Arm Road.
54.5 Bridge over Black Brook and intersection with Lohnes Road on right.
54.9 Outcrop of aplite of the Mooselookmeguntic Igneous Complex.
56.6 Turn left onto dirt road and drive down 0.1 miles to Black Brook and STOP 9.

STOP 9: TWO-MICA GRANITE OF THE SOUTHERN MOOSELOOKMEGUNTIC IGNEOUS
COMPLEX (MIC), BLACK BROOK ALONG SOUTH ARM RD. AT “SILVER RIPPLE CASCADE,”
ANDOVER SURPLUS, MAINE (UTM 4949556N, 362070W)

Medium- to coarse-grained Bt granodiorite (local quartz) in ~1-10m blocks in fine-grained Bt granite
that is typical of the southern lobe of the MIC. This stop is “96-60” of Tomascak et al. (2005). Magmatic Bt
foliation in the granite is steeply dipping, and pegmatite dikes cut both types of rocks. Whole rock
geochemistry and age data from this location are discussed in Tomascak et al. (2005).

Mileage

56.7 Return to S. Arm Road, and turn right to backtrack to Rt. 120.
61.4 Turn right onto Rt. 120 and continue to the town of Andover.
61.9 Continue straight through the intersection with Rt. 5 in the Andover town center.
63.6 Intersection with Right Cross Road. Rocks of Stop 10 are in the stream.

STOP 10. MONZODIORITE OF THE MOOSELOOKMEGUNTIC IGNEOUS COMPLEX, ANDOVER,
MAINE (UTM 4925999N, 357943W)

Medium-grained Bt granodiorite with a weak N-S foliation and schlieren typical of the western part of
the MIC. This stop is “J98-40” of Tomascak et al. (2005). This granodiorite is not as coarsely foliated as
similar rocks at STOP 9 (in the blocks). Whole rock geochemistry and age data from this location are
discussed in Tomascak et al. (2005).

END OF TRIP.

ACKNOWLEDGEMENTS

We are pleased to offer this trip as part of the 2017 NEIGC. The trip leaders remain grateful to Chuck Guidotti
and E-an Zen for getting us started in Maine, and in the Roxbury area in general. We thank the 2017 NEIGC field
trip coordinators for asking us to participate, and we thank all those who have helped us during the course of the
work that has produced the data and interpretations presented as this field trip guide. Naturally any misconceptions
and errors are our own. We remain grateful for the field support of Fred and Cheryl England formerly of Weld, ME,
Anna Solar, and Heather McCarthy. We are thankful for partial support of this work from the NSF (EAR-9705858),
the GSA Research Grants Program, the Department of Geology, University of Maryland, the Department of Earth
Sciences, SUNY College at Buffalo, and the Department of Atmospheric and Geological Sciences, SUNY Oswego.
REFERENCES CITED


APPLYING THE COSMOGENIC NUCLIDE DIPSTICK MODEL FOR
DEGLACIATION OF MT. WASHINGTON

by

P. Thompson Davis¹, Dept. Natural & Applied Sciences, Bentley Univ., Waltham, MA 02454-4705
Alexandria J. Koester² and Jeremy D. Shakun³, Dept. of Earth and Environmental Sciences, Boston
College, Chestnut Hill, MA 02467
Paul R. Bierman⁴ and Lee B. Corbett⁵, Dept of Geology, University of Vermont, Burlington, VT 05405
Email addresses: ¹pdavis@bentley.edu, ²koestera@bc.edu, ³jeremy.shakun@bc.edu, ⁴pbierman@uvm.edu,
⁵Ashley.Corbett@uvm.edu

INTRODUCTION

The purpose of this field trip includes the following: 1) describe the motivation and methods being used for a
NSF-funded project to understand deglaciation of the mountains in northern New England and adjacent areas, 2)
review past glacial history studies in the Presidential Range, 3) examine field sampling sites and initial results from
Mt. Washington, one of the key mountainous areas in our study, and 4) discuss implications for global sea-level
change during the last deglaciation in New England.

The late Pleistocene ice sheets were important agents of land surface and climate change during the Ice Ages of
the last 2.8 million years. Research over the past several decades has generated ever more precise reconstructions of
ice sheet extent histories based on organic ¹⁴C (e.g., Dyke, 2004), cosmogenic nuclides (e.g., Balco et al., 2002;
Carlson et al., 2007; Rinterknecht et al., 2006), and varve dating (e.g., Ridge et al., 2012). In contrast, the thickness
evolution of the ice sheets has been far more difficult to constrain, largely because the ice sheets generally covered
flat regions. In the rare instances where ice sheets covered mountainous areas such as northern New England, the lag
time was highly variable between deglaciation and the deposition of organic matter for ¹⁴C dating at higher
elevations (Davis and Davis, 1980). While a robust relationship between ice sheet area and volume exists for
glaciers in equilibrium (Paterson, 1994), this need not be the case for a deglaciating ice sheet due to, for instance,
changes in basal temperature and subglacial meltwater, ice streaming, ice-ocean interactions, interruption of ice flow
in regions of complex topography, and elevation-temperature and precipitation feedbacks (Fyke et al., 2014).

Accurate ice sheet thickness reconstructions are important for several reasons. (1) Ice sheet orography is a
critical boundary condition for modeling paleoclimate during the last deglaciation (e.g., Liu et al., 2009). For
instance, the height of ice sheets has a direct effect on surface temperature through lapse rate cooling, but also has
downstream effects related to atmospheric planetary waves that control heat flow and storm tracks, and the strength
of Atlantic Meridional Overturning Circulation (AMOC) (Ullman et al., 2014). (2) Ice sheet thickness must be
known in order to quantify ice volume changes, and thus the contribution of individual ice sheets to global sea-level
rise and the attendant freshwater forcing to the ocean (Carlson and Clark, 2012). (3) The timing and rate of ice sheet
thinning sheds light on how ice sheets responded to the overall global warming and abrupt climate changes of the
last deglaciation (Gregoire et al., 2012). (4) Ice thickness reconstructions can help validate numerical ice sheet
models, which are important for understanding the processes of deglaciation as well as improving projections of ice
sheet responses to future global warming (Stokes et al., 2015).

The Greenland ice core record reveals abrupt Northern Hemisphere warming and cooling events during the last
deglaciation (Clark et al., 1999; Andersen et al., 2004) that have been linked to variations in the strength of the
AMOC and its associated northward heat transport (Clark et al., 1996; McManus et al., 2004). Heinrich Stadial 1, a
cooling event between ~19 and 14.6 ka, is thought to have occurred due to freshwater forcing from the Northern
Hemisphere ice sheets weakening the AMOC. Recovery of the AMOC at ~14.6 ka then produced an abrupt
warming into the Bølling-Allerød interstadial (Liu et al., 2009). Meltwater Pulse 1a (MWP-1A), a sea-level rise event of 14-18 m in 350 years (Deschamps et al., 2012; Carlson et al., 2012), occurred synchronously with the Bølling interstadial. However, it is unclear from which ice sheet MWP-1A was sourced (Liu et al., 2015), though two scenarios have been proposed. The first is the ‘Northern scenario’ where melting Northern Hemisphere ice sheets caused MWP-1A, leading to a weakening of the AMOC and Older Dryas cooling after the Bølling warming (Fairbanks, 1989; Manabe and Stouffer, 1995; Peltier, 2005; Peltier and Fairbanks, 2006). The second is the ‘Southern scenario’, which suggests the melting of the Antarctic Ice Sheet caused MWP-1A, triggering a reactivation of the AMOC and the Bølling warming (Clark et al., 1996; Weaver et al., 2003; Bassett et al., 2005).

Modern studies of the Greenland Ice Sheet are reporting increasingly negative surface mass balance trends (Veliconga et al., 2014; McMillan et al., 2016), possibly due to accelerated thinning from surface water drainage to the bed (Zwally, 2002). Contributions from the Greenland Ice Sheet to global mean sea level rise is larger than Antarctica and has increased from 0.09 mm yr⁻¹ over 1992-2002 to 0.59 mm yr⁻¹ over 2002-2011 (Vaughan et al., 2013), and has more recently reached 0.74 ± 0.14 mm yr⁻¹ (McMillan et al., 2016). Paleo-constraints on inland ice sheets can provide valuable information about ice sheet thinning dynamics during periods of abrupt climate change, which can be used to improve models predicting future ice sheet decay (Hansen et al., 2015; Shakun et al., 2015; Winkelmann et al., 2015).

THE PROBLEM AND OUR APPROACH

Sea-level reconstructions reveal global ice volume variations through time, but provide little information on how that volume was partitioned among ice sheets (Clark et al., 2009). Although geophysical models attempt to invert isostatic rebound patterns into ice sheet thickness reconstructions, they exhibit considerable disagreement (Clark and Tarasov, 2014; Peltier et al., 2015). Similarly, numerical models attempt to simulate past ice sheet evolution in response to climate change (e.g., Abe-Ouchi et al., 2013; Gregoire et al., 2012), but they are subject to a wide variety of uncertainties such as the climate forcing and ice-sheet dynamics (Stokes et al., 2015). All such models ultimately require ground-truthing – but stronger geologic constraints are needed to advance this data-model dialogue, as highlighted by a recent community-wide workshop (Whitehouse and Tarasov, 2014).

A potential, albeit geographically-limited, solution to the problem of reconstructing ice sheet thickness through time is cosmogenic exposure dating along vertical transects, also known as ice sheet “dipsticks.” This technique measures the build-up of cosmogenic nuclides in a series of glacial boulders and/or outcrops down a mountainside to determine when each was exposed to cosmic radiation as the ice sheet surface lowered during deglaciation (see commentary by Bierman, 2007). Glacial dipsticks have been instrumental in constraining the thinning history of ice sheets in Scandinavia (Brook et al., 1996; Goehring et al., 2008), Antarctica (Stone et al., 2003; Ackert et al., 2007; Mackintosh et al., 2007, 2011; Johnson et al., 2014), and Greenland (Corbett et al., 2011; Nelson et al., 2014).

Strikingly, other than our own recent study at Acadia National Park in Maine (Koester et al., 2017), no major glacial dipsticks have been measured for the Laurentide Ice Sheet (LIS), which was the largest body of ice at the Last Glacial Maximum (LGM), accounting for ~65-90 of the 130 m LGM sea level lowstand (Clark and Mix, 2002). There are only two regions where substantial topographic relief (>1000 m) was uncovered by LIS retreat during the last deglaciation – the mountains of New England and southern Quebec (Fig. 1), and much more remote parts of the eastern Canadian Arctic, including Baffin Island and northern Labrador. Samples from either of these regions could directly constrain the thinning history of the large, but now vanished LIS, although, for reasons related to ice sheet basal thermal conditions in the Canadian Arctic (Marsella et al., 2000; Corbett et al., 2016; Margreth et al. 2016), the mountains of New England are much more likely to provide accurate deglacial ages.
GLOBAL CLIMATE AND SEA LEVEL DURING THE LAST DEGLACIATION

The last deglaciation provides an outstanding opportunity to understand the complex interplay between ice sheets, ocean circulation, and climate. We provide below a summary of the last deglaciation to highlight the relevance of these questions and to flag uncertainties that our proposed research could help address.

The Oldest Dryas (19-14.6 ka)

The last deglaciation commenced with an abrupt 5-10 m sea-level rise over a few centuries at ~19 ka (Carlson and Clark, 2012), which has been associated with initial pullback of Northern Hemisphere ice sheets due to summer insolation forcing (Fig. 2a) (Clark et al., 2009), although emerging marine data suggest that the Antarctic Ice Sheet also began retreating at this time for unknown reasons (Weber et al., 2011, 2014). The resulting freshwater forcing to the North Atlantic may have weakened the AMOC causing a bipolar seesaw climate response with hyper-cold conditions centered around the North Atlantic and warming in the Southern Hemisphere (Fig. 2b,c) (He et al., 2013). Cold stadial conditions in the Northern Hemisphere and a sluggish AMOC persisted for the next four millennia of the Oldest Dryas interval (McManus et al., 2004; Shakun et al., 2012), but the Northern Hemisphere ice sheets continued retreating and the LIS underwent a major iceberg discharge episode during Heinrich event 1 at ~16 ka (Hemming, 2004). Significant ice loss during the cold Oldest Dryas may seem somewhat surprising. A possible explanation is that the Oldest Dryas was characterized by extreme seasonality, with cooling predominantly during winter as the weakened AMOC promoted sea-ice expansion while summers continued warming due to rising insolation and atmospheric CO₂ (Denton et al., 2005). Coral-based estimates of global sea-level rise during the Oldest Dryas range widely from 8 to 21 m, and the contribution of individual ice sheets is especially uncertain (Carlson and Clark, 2012). Of particular interest, sea-level rise seems to have outpaced LIS area retreat (Fig. 2a). Marshall et al. (2002) highlight a similar disparity, noting that, “the isostatic record demands substantial ice thinning subsequent to LGM, at a time (14-20 ka) when there is no strong signal of ice sheet retreat (Dyke and Prest, 1987). Model results suggest that this is possible via the increasing role of fast basal flow in this period, as

Figure 1. The locations of 12 ice-sheet mountain dipsticks (red dots) to constrain the thinning history of the southeastern LIS. As of July 2017, we have sampled along vertical transects from all sites except for the Catskills and Mt. Kearsarge. The smaller map shows the LIS outline at the LGM (yellow), and highlights the study region (white box). Note that this region contains the only large mountains underlying interior portions of the LIS, and thus provides a unique opportunity to reconstruct the vertical collapse of the ice sheet.
more of the Laurentide ice sheet becomes warm-based and basal melt water accumulates at the bed. This essentially argues for a transition from thicker, largely cold-based ice sheets at the LGM to a thin and mobile, more West Antarctic Ice Sheet-like ice sheet through the deglaciation.” As a review on deglacial sea level by Carlson and Clark (2012) recently concluded though, “the volume contributions of individual ice sheets to sea level change between 19.5 ka and 14.6 ka, which are required to specify freshwater fluxes and their entry points to the ocean, need to be better determined.” Determination of the volume contribution for the southeastern part of the LIS is a major goal of this project.

**Figure 2.** Deglacial ice melt, climate, and ocean circulation. (a) Global sea level (green) (Lambeck et al., 2014) LIS areal extent (blue) (Dyke, 2004), and northward LIS retreat in central New England based on varves (red) (Ridge et al., 2012). (b) Greenland δ¹⁸O, a proxy for temperature (NGRIP members, 2004). (c) Protactinium/thorium ratios in a North Atlantic sediment core, a proxy for AMOC strength (McManus et al., 2004). Note the differences between LIS extent and global sea level, the increased rate of LIS retreat in New England coincident with MWP-1A, and the general associations between changes in ice-sheet retreat/sea-level rise, ocean circulation strength, and temperature.

**Meltwater Pulse 1a and Bolling warming (14.6-14.3 ka)**

An abrupt warming of the Northern Hemisphere occurred at the onset of the Bolling interstadial at 14.6 ka as the AMOC resumed (Liu et al., 2009), and coincided with the largest jump in deglacial sea level –MWP-1A (Fig. 2). This 14-18-m sea-level rise occurred in no more than 350 years, implying rates of sea-level rise in excess of 40 mm/yr (Deschamps et al., 2012), or more than an order of magnitude faster than sea-level rise today (3 mm/yr; Church and White, 2011). While MWP-1A was first assumed to have originated exclusively from the LIS (Fairbanks, 1989; Peltier, 1994), given its large size, sea-level fingerprinting and Southern Ocean marine evidence suggest a significant though uncertain Antarctic contribution (Weaver et al., 2003; Deschamps et al., 2012; Weber et al., 2014). Planktonic δ¹⁸O runoff records from the Gulf of Mexico (Wickert et al., 2013), the Arctic (Carlson, 2009), and the Labrador Sea (Obbink et al., 2010) detect only minor contributions from various sectors of the LIS to MWP-1A. Furthermore, LIS areal retreat was no greater during MWP-1A than before or after the event (Fig. 2a). Therefore, any major LIS sea-level contributions could only have come from rapid ice sheet thinning. Just to provide a sense of scale, if MWP-1A were sourced evenly from across the entire LIS, it would lower the ice sheet surface by
~600 m, a thinning easily detectable using the dipstick method we are currently employing. Sourcing MWP-1A from only a part of the ice sheet would obviously increase this surface lowering estimate further. Gregoire et al. (2012) simulate a 9-m sea-level rise in 500 years in a numerical ice sheet model as the LIS and Cordilleran Ice Sheet separated due to saddle collapse and suggest that this process may account for MWP-1A, though $^{14}$C ages suggest that these ice sheets actually separated well before MWP-1A (Clague and James, 2002; Dyke, 2004). The Eurasian Ice Sheet complex was much smaller than the LIS, and therefore a less likely candidate to explain MWP-1A. Adding to this puzzle, recent glacioisostatic modeling suggests that Antarctica only contained ~8 m sea-level equivalent of additional ice at the LGM (Whitehouse et al., 2012), limiting its potential contribution to MWP-1A, and available Antarctic dipsticks indicate only modest thinning (Mackintosh et al., 2007, 2011). An accounting of the sources of sea-level rise during this singular event (MWP-1A) is thus far from complete (Fig. 3).

The Allerød, Younger Dryas, and Holocene (14.3-6.5 ka)

Sea-level rise returned to pre-MWP-1A rates after 14.3 ka and sea level increased ~7-10 m during the Allerød period over the next millennium (Fig. 2a) (Edwards et al., 1993; Bard et al., 1996; Peltier and Fairbanks, 2006), likely dominated by Northern Hemisphere sources (Carlson and Clark, 2012). The abrupt Younger Dryas cold event in the Northern Hemisphere commenced 12.9 ka as the AMOC weakened again, perhaps due to southern LIS retreat into Canada and routing of freshwater runoff from the Mississippi to the St. Lawrence drainage (Broecker et al., 1989; Broecker, 2006). LIS retreat and sea-level rise slowed during the Younger Dryas, before again picking up pace as the AMOC resumed at the onset of the Holocene at 11.6 ka (Bard et al., 2010) and summer insolation reached a maximum. The Eurasian Ice Sheet disappeared by 10 ka (Boulton et al., 2001), and the remaining Holocene sea-level rise came from LIS retreat in Canada (Dyke, 2004; Carlson et al., 2007) and the Antarctic Ice Sheet.

The central message that emerges from this summary is that while the broad pattern of climate change, ocean circulation, and sea-level rise during the last deglaciation are reasonably well constrained, the contributions of individual ice sheets to these processes, which is critical to understanding the internal dynamics of the climate system, are not. In particular, well-documented ice margin retreat histories are not complemented by similarly strong vertical thinning constraints, and thus ice volume uncertainties remain substantial.
SOUTHEASTERN LAURENTIDE DEGLACIATION

With the global summary above as context, we detail here the deglaciation of the southeastern LIS, the focus of our current research. Central New England has one of the best-constrained ice margin histories in the world (Fig. 4), owing to considerable $^{14}$C dating of lakes and bogs, and in particular, extensive varve sequences, which have been tied to the $^{14}$C timescale and the Greenland ice core record (Ridge et al., 2004, 2012). Furthermore, $^{10}$Be ages from the Connecticut and Champlain Valley lowlands have also been linked to these other chronometers, resulting in a precise regional production rate calibration (Balco et al., 2009). This well-dated margin record makes the region especially conducive to constraining ice volume changes and understanding ice sheet behavior, if, and only if, the thinning history can be well determined.

Figure 4. Laurentide Ice Sheet lateral extent through time. Isochrones show the North American Varve Chronology of deglaciation (from Fig. 11 in Ridge et al. (2012)). The black dots are moraines dated using $^{10}$Be and calibrated with the northeastern North American production rate. (Martha’s Vineyard and Buzzard’s Bay – Balco et al., 2002; Old Saybrook and Ledyard – Balco and Schaefer, 2006; Ashuelot Valley, Perry Mountain and Litleton-Bethlehem – Balco et al., 2009; Androscoggin – Bromley et al., 2015; Basin Pond – Davis et al., 2015; Pineo Ridge and Acadia National Park – Koester et al., 2017). Inset figure shows the extent of the LIS at 21 ka and 10 ka (Dyke, 2003).
Ice Extent

The LGM and subsequent deglaciation chronology throughout the New England area has been well researched and includes minimum-limiting radiocarbon ages (e.g. Dyke et al., 2004), glacial varves (Ridge, 2004; Ridge et al., 2012), and cosmogenic nuclide dating (Balco et al., 2002; Balco and Schaefer, 2006; Bierman et al., 2015; Bromley et al., 2015; Davis et al., 2015; Corbett et al., 2017; Koester et al., 2017; Hall et al., 2017; Fig. 4). The LIS reached its maximum extent (Fig. 4) before ~25 ka as indicated by recalculated 10Be exposure ages on terminal moraines from Martha’s Vineyard, MA (27.5 ± 1.6 ka, Balco et al., 2002) and northern New Jersey (25.2 ± 1.3 ka, Corbett et al., 2017). After the LIS began pulling back from its terminal moraines, northward retreat was gradual through several millennia at ~50 m/year (Ridge et al., 2012), then increased after ~20 ka likely due to increased Northern Hemisphere insolation (Clark et al., 2009). There is little indication of Heinrich event 1 at the southern margin, except perhaps indirectly, with the modest Chicopee readvance in Massachusetts at 17.3 ka, possibly occurring in response to North Atlantic cooling (Ridge et al., 2012). Coastal Maine, mostly below the marine limit, rapidly deglaciated 15-16 ka (Davis et al., 2015; Koester et al., 2017; Hall et al., 2017), and the North Charlestown moraines were deposited across central New England just prior to the Bølling warming. Thereafter, and synchronous with MWP-1A, the rate of retreat increased dramatically to ~300 m/year (Fig. 2a, 4) (Ridge et al., 2012). Except for the Littleton-Bethlehem Readvance north of the White Mountains of New Hampshire at 13-14 ka (Balco et al., 2009; Ridge et al., 2012) (Fig. 2b), this rapid retreat continued until the LIS margin exited New England into southern Quebec during the late Allerød (Fig. 4) (Dyke, 2004).

The North American Varve chronology from the Connecticut River valley provides insight into the retreat of the LIS from Massachusetts to Vermont and indicates the retreat rate began between 50 and 100 m/yr (Ridge, 2004; Ridge et al., 2012). Thereafter, the retreat rate increased to ~300 m/yr during the Bolling Interstadial and passed Mt. Washington around ~14.2 ka (Ridge, 2004; Ridge et al., 2012; Fig. 4). Glacial varves were also thicker during the Bølling, implying more intense summer melt. In addition, a model of the LIS at the end of the Oldest Dryas (~15 ka) and during the Bølling-Allerød (~14.2 ka) found the surface balance increased during the Bølling (Carlson et al., 2012), supporting the varve thickness record. As the ice sheet retreated further inland the Pineo Ridge moraine complex in coastal Maine was abandoned around the Bølling Interstadial (14.5 ± 0.7 ka; Koester et al., 2017) followed by the Littleton-Bethlehem moraine, just north of the Presidential Range in northern New Hampshire (13.8 ± 0.7 ka, n = 4; Balco et al., 2009; Thompson et al., 2017), and the Androscoggin moraine in northeastern New Hampshire and western Maine (13.2 ± 0.7 ka, n = 7; Bromley et al., 2015) before retreating further north into Canada.

Ice Thickness

In contrast to the dozens of 10Be ages, hundreds of organic 14C ages, and thousands of varve counts constraining ice retreat at lower elevations in New England, the ice thickness history is largely uncertain. Existing 14C ages at higher elevations are scant, they come almost exclusively from lake and bog basal sediments and are thus only minimum-limiting ages, and they are very noisy, spanning several millennia and showing no coherent trends with elevation (Table 1). It is thus not possible to say whether ice sheet thinning occurred predominantly during the cold Oldest Dryas or the warm Bolling/MWP-1A interval or the still-warm but slower-sea-level-rise Allerød interval. It is similarly an open question whether ice sheet drawdown was very rapid (centuries) or much more gradual (millennia).

Two recent studies from our team present initial cosmogenic ages on a small set of samples from Katahdin, ME, and Mt. Washington, NH, and raise intriguing questions in these regards. Samples from the Katahdin highlands have statistically indistinguishable deglaciation ages (n=6, 15.3±2.1 ka, 1σ) from boulders on the Basins Pond moraine halfway up the mountain (n=5, 16.1±1.2 ka, 1σ), as well as a lone boulder in the nearby lowlands (14.5±0.8 ka, 1σ) (Davis et al., 2015) (Fig. 5). These ages thus imply rapid ice surface lowering at ~16-15 ka, though more gradual...
thinning cannot be excluded due to the small sample number and age scatter, which may reflect measurement imprecision as AMS was done in the 1990s. In addition, $^{10}$Be and $^{26}$Al nuclide concentrations in several summit bedrock samples from both mountains are 2-10 times higher than would be expected due to ~15 kyr of postglacial exposure (and confirmed in corresponding in situ $^{14}$C ages) (Fig. 5), suggesting that the summits were covered by non-erosive, cold-based ice at the LGM (Bierman et al., 2015). Also important, nuclides inherited from prior periods of exposure are found only in summit samples, while lower on the mountains and throughout the rest of New England, such inheritance is rare in boulders and bedrock. The lack of nuclides inherited from prior periods of exposure indicates that in New England ice was largely erosive (warm-based) and that the cosmogenic clock was reset during or after the LGM, as our initial project results at Acadia National Park in Maine demonstrate (Koester et al., 2017), although the relief there is only about 300 meters, substantially less than many other mountains in our study.

<table>
<thead>
<tr>
<th>Site</th>
<th>Material</th>
<th>Elev (m)</th>
<th>$^{14}$C age</th>
<th>cal yr BP</th>
<th>Reference</th>
</tr>
</thead>
<tbody>
<tr>
<td>Deer Lake bog</td>
<td>bulk conv.</td>
<td>1325</td>
<td>13,000±400</td>
<td>14,195-16,820</td>
<td>Spear (1989)</td>
</tr>
<tr>
<td>Mirror Lake</td>
<td>bulk/macros conv.</td>
<td>213</td>
<td>13,800±560</td>
<td>15,720-17,415</td>
<td>Davis and Davis (1980)</td>
</tr>
<tr>
<td>Lonesome Lake</td>
<td>bulk conv.</td>
<td>831</td>
<td>10,535±495</td>
<td>11,065-13,355</td>
<td>Spear et al. (1994)</td>
</tr>
<tr>
<td>Profile Lake</td>
<td>wood conv.</td>
<td>593</td>
<td>10,660±40</td>
<td>12,772-12,885</td>
<td>Rogers (2003)</td>
</tr>
<tr>
<td>Lakes of Clouds</td>
<td>bulk conv.</td>
<td>1538</td>
<td>11,530±165</td>
<td>13,200-13,500</td>
<td>Spear (1989)</td>
</tr>
<tr>
<td>Lost Pond</td>
<td>bulk conv.</td>
<td>625</td>
<td>12,870±370</td>
<td>14,700-16,000</td>
<td>Spear et al. (1994)</td>
</tr>
</tbody>
</table>

Table 1. Low versus high-elevation $^{14}$C ages from the White Mountains, NH. Note that the age pairs tend to be fairly similar, suggesting that ice-sheet drawdown may have been rapid. On the other hand, these ages are only minimum-limiting, and the close correspondence in ages may reflect the timing of revegetation and the first occurrence of datable organic material. Indeed, this complication may explain why higher-elevation ages tend to be younger than lower-elevation ages, opposite the pattern expected from top-down deglaciation.

Figure 5. $^{10}$Be ages from the Katahdin, ME, and Mt. Washington, NH, areas. Panel on top shows zoom-in of panel on bottom. Ages are from Balco et al. (2009), Davis et al. (2015), Bromley et al. (2015), and Bierman et al. (2015), and were determined using the CRONUS calculator using the northeast North America production rate (Balco et al., 2009). A few bedrock samples from the summits have older exposure ages than expected, indicating nuclides inherited from prior periods of exposure. The data are too sparse and do not include boulder ages higher on the mountains to provide robust constraints on thinning, but they highlight the potential of such work.
The Presidential Range is located in the White Mountains of New Hampshire and consists of 13 peaks ranging from 1,235 m (Mt. Jackson) to 1,917 m (Mt. Washington). The large cirque basins throughout the Presidential Range (i.e. Huntington Ravine, Tuckerman Ravine, and the Great Gulf on Mt. Washington) have been a topic of debate since they were first studied by J.W. Goldthwait (1913, 1916). J.W. Goldthwait concluded that the cirques were carved by alpine glaciers before the most recent continental glaciation and cited evidence that included till on the cirque floors from northern provenances, absence of end moraines on cirque floors, and asymmetric cirque cross-valley profiles. Despite the strong evidence, Antevs (1932) and Johnson (1917, 1933) opposed J.W. Goldthwait’s conclusions about the sequence of cirque and ice sheet glaciation. R.P. Goldthwait (1970) later provided evidence from pebble lithologies in till on the uplands and in the north-facing cirques and concluded that the till in the Presidential Range was deposited by continental ice, supporting the idea that the cirques were carved before continental glaciation. Further, unlike Wagner (1970) in northern Vermont, Waitt and Davis (1988) and Davis (1999) found no evidence of cirque glaciation following continental ice overriding all mountainous areas of northern New England. However, Fowler et al. (2012) suggest that deposits at the mouth and along the sidewalls of the Great Gulf may provide evidence for alpine (or continental ice) in that cirque post-dating continental ice overriding the Presidential Range.

Of particular interest to our current work are the relatively level high-elevation areas known as “lawns” (ex. Bigelow Lawn, the Alpine Garden, and Monticello Lawn), which together make up a topographic feature called the Presidential Upland (Fig. 6) that represents an old Tertiary erosion surface formed during a prolonged lull in tectonic uplift (Goldthwait, 1940), whereas Thompson (1960a, 1960b, 1961) believed that the features were the result of freeze-thaw processes during the Quaternary. Eusden and Fowler (2013) sided with R.P. Goldthwait.

There is clear evidence from glacial erratics (Fig. 7), glacially molded surfaces, and the thinness of soils on summits that the LIS overrode New England mountains (Tarr, 1900; J.W. Goldthwait, 1916; R.P. Goldthwait, 1940; Davis, 1976, 1989; Fowler et al., 2013), as opposed to the mountains being centers of radial outflow of ice during the late Wisconsinan (Flint et al., 1942; Flint, 1951). However, the timing of initial LIS advance into New England
during the last glaciation is poorly constrained. The glacial stratigraphic record of most of New England includes an upper till and a lower till. The upper till is part of a widespread drift sheet in New England and is interpreted to be late Wisconsinan based on minimum-limiting radiocarbon ages. The more weathered lower till deposits underlying late Wisconsinan (Marine Isotope Stage (MIS) 2) deposits is less common in New England, but occurrences have been described from the Boston Harbor drumlins, Massachusetts (Kaye, 1961), Nash Stream, New Hampshire (Koteff and Pessl, 1985), and New Sharon, Maine (Caldwell, 1986; Weddle, 1989), but its age remains unknown. The lower till’s age was originally assigned to the early Wisconsinan (MIS 4) (Borns and Calkin, 1977; Stone and Borns, 1986; Vincent and Prest, 1987), but others have suggested the lower till is more likely Illinoian (MIS 6) or older due to radiometric dating and amino acid racemization age estimates on detrital coral from the upper part of the Sankaty Head on Nantucket Island, MA (Oldale, 1982; Oldale and Colman, 1992). In addition, ice volume estimates from oxygen isotope and sea level records indicate ice was less extensive during the early Wisconsinan than the late Wisconsinan, possibly suggesting that the LIS did not extend as far south as New England then (Oldale and Colman, 1992; Lambeck et al., 2014).

Figure 7. View looking south at Mt. Washington from about 30 m below summit of Mt. Jefferson with two large, sub-rounded, granitic erratics in foreground (rucksack and trekking pole for scale). Angular and sub-angular blocks are frost-riven from the schist bedrock, common on all of the summit cones of the northern Presidential Range. Great Gulf cirque headwall and Mt. Clay on north side and Mt. Monroe to southwest of Mt. Washington.

Deglacial constraints on LIS retreat at higher elevations in New England include minimum-limiting basal radiocarbon ages from alpine lakes (Table 1) and a few cosmogenic nuclide exposure ages on Katahdin and Mt. Washington. For instance, a basal bulk radiocarbon age from the lower Lakes of the Clouds (1,534 m), just below the summit cone of Mt. Washington, dates to 13.35 ± 0.2 cal. ka (Table 1; 11,530 ± 165 14C yrs BP; I-10684; Spear, 1989, Cwynar et al., 2001), although this age probably substantially post-dates deglaciation since the ice margin had already retreated to the Canadian border by this time (Ridge et al., 2012). Radiocarbon ages from alpine lakes on Katahdin are also several thousand years younger than continental ice retreat, perhaps due to the lag in vegetation colonization following deglaciation (Davis and Davis, 1980). Davis et al. (2015) measured surface exposure ages at multiple sites from the top to base of Katahdin and concluded that the LIS thinned rapidly between 16 and 15 ka in central Maine. Cosmogenic nuclide exposure ages from the summits of Katahdin and Mt. Washington were 2-10 times higher than expected indicating that cold-based ice likely covered the summits of New England mountains (Bierman et al., 2015).
MOTIVATING QUESTIONS

The background given in the previous two sections highlights several long-standing problems related to ice sheet deglaciation, sea-level rise, climate, and ocean circulation to which we can contribute greatly through the dipstick approach. Our research specifically addresses the following five questions concerning the southeastern LIS.

1) Did the LIS thin during the Oldest Dryas cold interval, or specifically during Heinrich event 1?

   Such thinning might suggest that Oldest Dryas cooling was mostly during winter and related to sea-ice expansion (Denton et al., 2005), or that Heinrich 1 dynamical discharge caused significant ice-sheet drawdown over the southeastern LIS (Shaw et al., 2006).

2) Did the LIS thin synchronously with MWP-1A and Bølling warming?

   The southern LIS would have presumably been one of the most vulnerable ice masses in the world to ablate during the Bølling given its southerly location and likely sensitivity to climate change in the nearby North Atlantic. The Midwest ice lobes, however, were thin due to underlying deformable beds (Clark, 1992) and contributed little to sea-level rise based on Gulf of Mexico runoff records (Wickert et al., 2013). The North American varve record indicates that ice margin retreat there increased dramatically during MWP-1A, perhaps implying a significant contribution, though this margin retreat was no faster than during the following thousand years after the sea-level event ended (Fig. 2a, 4). There are currently no ice sheet dipsticks anywhere in the world that date pronounced thinning to the MWP-1A interval, although scatter in vertical data from Norway permit this possibility (Goehring et al., 2008).

3) How fast did the ice surface lower?

   Rapid drawdown might point to active ice dynamics, such as meltwater delivery to the bed and increased sliding, and would provide constraints on how fast a vulnerable subpolar ice sheet can collapse in a warming climate, perhaps broadly analogous to the southern Greenland Ice Sheet today. Gradual thinning would imply weaker climate forcing and/or less dynamic discharge.

4) How did southeastern LIS melt relate to changes in the AMOC?

   In addition to the general correspondence between New England LIS margin retreat and North Atlantic temperature evolution (Fig. 2a,b), New England varve thickness records also suggest a direct correlation between summer melt of the LIS and Greenland temperature (Ridge et al., 2012). On the other hand, a δ¹⁸Osw record from the nearby Laurentian Fan suggests that southeastern LIS meltwater production was inversely related to North Atlantic temperature, being higher during the Oldest Dryas and three cold intervals within the Bølling/Allerød than when the North Atlantic was warmer (Obbink et al., 2010). Likewise, mass-balance modeling suggests that the impact of Bølling warming on the southeastern LIS may have been offset by an associated increase in precipitation (Carlson et al., 2012). These conflicting views highlight the causal uncertainty in how southeastern LIS evolution and North Atlantic climate are linked – for instance, to what extent does LIS melt cause North Atlantic cooling through freshwater forcing of the AMOC versus North Atlantic warming causes LIS melt?

5) How well do ice sheet models simulate LIS deglaciation?

   Ultimately, one of the central goals of Earth science is to accurately model the Earth system, and this enterprise hinges crucially on ground truthing models (Stokes et al., 2015). While several models of ice-sheet retreat exist, based on either isostatic inversion techniques or glaciological modeling, the models are still poorly constrained by geologic data and exhibit substantial differences, such as estimates of the LIS contribution to MWP-1A (Fig. 3). These model discrepancies are apparent for the southeastern LIS, with pronounced differences in ice thickness at the LGM, the ice-sheet profile, and the timing and rate of thinning through the deglaciation (Fig. 8). Our data provide a novel test for these models and serve as a target for future modeling efforts.
RESEARCH STRATEGY

Sampling

We constrain the timing and rate of southeastern LIS thinning by measuring cosmogenic nuclide dipsticks on a dozen of the highest peaks in New England and southern Quebec (Fig. 1). We target boulders as much as possible, given that they are less likely to contain inheritance from prior periods of exposure (Hallet and Putkonen, 1994; Putkonen and Swanson, 2003; Putkonen and O’Neal, 2006; Balco, 2011; Heyman et al., 2011); bedrock is only sampled when boulders are not available because nuclides created during prior periods of exposure can be preserved in bedrock beneath cold-based ice (e.g., Bierman et al., 1999, 2015; Colgan et al., 2002; Briner et al., 2003; Goehring et al., 2008). Our goal is production of a 3D-model of deglaciation of the mountains of the northern New England area to assess ice volume changes over time.

Figure 8. Ice sheet models of deglaciation. Southeastern LIS profiles along 71°W (the longitude of Mt. Washington, shown as the black line in the inset map) during various time steps of the last deglaciation (see legend for ages) based on the ICE-5G model (Peltier, 2004), a model from the Australian National University (ANU; Lambeck et al., 2002), and a dynamical model of the North American ice sheet system (G12; Gregoire et al., 2012). The summit location of Mt. Washington is represented by the solid black triangle. Simulated ice thinning during the Bølling warm interval (14.5-14.0 ka) is highlighted in pink on each panel. Note the dramatic differences in ice sheet thicknesses at 21 ka, ice sheet profiles, and timing and rates of thinning. Our dipsticks are well-positioned to distinguish between these, and other, models of deglaciation.

Sample sites (Fig. 1) include the highest points along the Green Mountains of Vermont (Jay, 1177 m; Mansfield, 1330m; Killington, 1289 m); high peaks of New Hampshire’s White Mountains (Washington, 1917 m; Lafayette, 1600 m; Kearsarge, 895 m; we note that Joe Licciardi and one of his graduate students at UNH exposure dated the highest mountain in southern New Hampshire, Mt. Monadnock (Hodgdon, 2016)); the highest peaks in Maine (Katahdin, 1606 m; recently added Mt. Bigelow, 1247 m; and Cadillac, 466 m, on the coast); western (Greylock, 1064 m) and eastern (Wachusett, 611 m) Massachusetts high points; the Catskill Mountains of New York (up to 1277 m); and the tallest mountain in southeastern Canada (Jacques-Cartier, 1268 m). These mountains were chosen because they: (i) provide the maximum relief available and so were exposed to much of the ice sheet thickness, (ii) span 7° of latitude and 9° of longitude, (iii) are composed mostly of quartz-bearing rocks (granite, schist, quartzite) ideal for cosmogenic exposure dating, (iv) will be relatively inexpensive to study since they are easily accessible and close to our home institutions, and (v) the six northern mountains were all within the LIS margin at the time of MWP-1A (Fig. 4).

Nuclides

In situ $^{10}$Be and $^{14}$C are primarily formed by cosmic ray spallation of oxygen in quartz-bearing rock and soil surfaces where nuclide concentrations build up over time (Gosse and Phillips, 2001). Cosmic ray flux attenuates within a few meters of surface, but the highest concentration is at the surface due to neutron attenuation with depth (Gosse and Phillips, 2001). In glacierized areas, erosive, warm-based ice typically erodes many meters into the
underlying bedrock to remove nuclides from prior periods of exposure. The concentration of cosmogenic nuclides can be converted into exposure ages with a production rate to measure when ice retreated and exposed the area. However, non-erosive, cold-based ice that is frozen to the bed can leave behind nuclides inherited from previous exposure periods leading to older than expected ages (Gosse and Phillips, 2001). On the other hand, post-glacial cover by snow or soil can shield the surface from cosmic rays and lead to an artificially young exposure age (Schieldgen et al., 2005). A regional $^{10}$Be production rate has been calibrated for northeast North American from independently dated moraines within New England reducing our uncertainty on exposure ages (Balco et al., 2009).

Based on our initial cosmogenic exposure ages on bedrock from the summits of Katahdin and Mt. Washington (Fig. 5; Bierman et al., 2015; Davis et al., 2015), we should not have been surprised that a few $^{10}$Be ages from higher summits are older than expected considering the existing varve and radiocarbon age control (Fig. 4, Table 1). Similarly, existing data sets (Balco et al., 2002; Balco and Schaefer, 2006; Davis et al., 2015) suggest that a few boulders carry inherited nuclides. For samples with higher than expected ages, we measure in situ $^{14}$C (Bierman et al., 2015), which removes the confounding variable of inheritance from exposure prior to the LGM (with its short half-life, most $^{14}$C produced during prior interglacials decays away during 20-30 ky of burial by ice (Lifton et al., 2001).

On the basis of existing paired $^{26}$Al/$^{10}$Be data for Katahdin and Mt. Washington (e.g., Davis et al., 2015; Bierman et al., 2015), which provide statistically similar exposure ages for both nuclides, we believe that measuring $^{26}$Al in dipstick samples likely does not provide useful additional information about sample history. Although $^{26}$Al measurements can be useful in other situations, measuring $^{26}$Al is redundant in New England samples because during a 100 ky glacial cycle, total exposure is several times longer than burial (~80 ky of exposure, 20 ky of burial); therefore, interglacial exposure quickly raises $^{26}$Al/$^{10}$Be ratios to production values following each short period of burial by ice. As a result, $^{26}$Al analysis mirrors $^{10}$Be ages, thus funds are better spent measuring in situ $^{14}$C or more samples for $^{10}$Be.

**Thinning rates**

We calculate thinning rates following Johnson et al.’s (2014) approach, who measured Holocene cosmogenic dipsticks at Pine Island Glacier, Antarctica. This approach involves fitting error-weighted least-squares regressions through the dipstick profiles. Uncertainties are quantified through Monte Carlo simulations, in which the cosmogenic ages are allowed to randomly vary within their Gaussian uncertainties and the regression is recalculated (Fig. 9). Negative or zero slopes are rejected as physically untenable.

![Figure 9. Synthetic dipsticks for Mt. Washington, NH, testing sensitivity to deglaciation duration, sampling density, and $^{10}$Be geologic uncertainty typical for cosmogenic datasets. The thin colored lines behind each dipstick show 500 Monte Carlo-generated regressions in which $^{10}$Be ages were allowed to randomly vary within their Gaussian uncertainties, and dotted lines give 68% confidence interval.](image)
A New England area grand synthesis

The data we generate will be incorporated into a comprehensive database detailing all extant chronological data from the region – radiocarbon ages (e.g., Dyke, 2004), cosmogenic exposure ages (e.g., Balco et al., 2002; Balco and Schaefer, 2006; Balco et al., 2009; Bromley et al., 2015; Davis et al., 2015; Bierman et al., 2015; Koester et al., 2017; Hall et al., 2017), and varve constraints (Ridge et al., 2012). We envision this reconstruction representing the culmination of decades of glacial geologic work in this data-rich region, with ice volume calculations now possible using the vertical constraints that our cosmogenic exposure age data will provide. This reconstruction will then be compared to offshore marine records, climate records (Fig. 2b), and models (Fig. 8) to understand the nature of southeastern LIS deglaciation, and infer its possible causes and consequences. Our reconstruction will allow us to estimate New England ice volume losses, which together with a recent reconstruction of the south-central LIS’s contribution to deglacial sea-level rise based on ice-sheet models and Gulf of Mexico runoff records (5.5 ± 2.1 m; Wickert et al., 2013), will better constrain the sea-level contribution history of the entire southern part of the LIS.

INITIAL RESULTS

Twenty \(^{10}\text{Be}\) ages from boulders and bedrock along a vertical transect on the east side of Mt. Washington, the highest peak in New England (1917 m), constrain the timing and rate of LIS thinning during the last deglaciation (Fig. 10). Also, six new in situ \(^{14}\text{C}\) ages for bedrock and boulders from the upper reaches of the mountain provide additional age constraints on deglaciation. With our data, we also seek to better explain the distinctive topography of the Presidential Range in New Hampshire, specifically the relatively level landscape at higher elevations versus deeply incised mountainsides below.

Our \(^{10}\text{Be}\) exposure ages range from 12.5 ± 0.6 to 81.6 ± 4.5 ka (Fig. 10), and show a strong ordering with elevation; they are similar and agree with the ~14 ka timing of regional deglaciation (except for MW-13 (34.3 ± 0.6 ka)) up to ~1,600 m asl, but then curve to increasingly older ages toward the summit, reaching values that are 3-6 times higher than the regional deglaciation age (Fig. 11). These anomalously old ages suggest that there was minimal glacial erosion higher on the mountain, consistent with a transition from warm- to cold-based ice at about 5300 ft (1,600 m) elevation, which accords with the gradual topography of the Presidential “lawn” above this elevation.

The \(^{10}\text{Be}\) ages between 1520 and 730 m a.s.l. are indistinguishable from one another at 1σ, and have a mean exposure age of 15.1 ± 0.8 ka (n = 7; 1SD). The two \(^{10}\text{Be}\) exposure ages on boulders from Pinkham Notch at ~670 m a.s.l. (13.0 ± 0.4, 12.7 ± 0.2 ka) are substantially younger than the 15.35 ± 0.6 cal \(^{14}\text{C}\) age (12,870 ± 370 \(^{14}\text{C}\) yrs BP, Spear et al., 1994) from near-basal organic sediments in nearby Lost Pond (Table 1), slightly younger than the well-dated Androscoggin moraine north of the Presidential Range (13.2 ± 0.4 ka; Bromley et al., 2015), and much younger than suggested by ice retreat in the Connecticut Valley varve chronology to the west (14.1 ka; Ridge et al., 2012). Thus, we consider the two exposure ages on boulders downslope from Square Ledge in Pinkham Notch to be outliers, perhaps due to the boulders falling into place following deglaciation; therefore, we consider the mean Androscoggin moraine exposure age to be a better constraint for the base of our glacial dipstick.

DISCUSSION

Comparison of our exposure dating results from Mt. Washington suggest that continental ice in the area lowered rapidly during the Bølling-Allerød, which accords well with an increase in ice margin retreat rates in the Connecticut River valley to the west based on the North American varve chronology and with the NGRIP \(\delta^{18}\text{O}\) ice core record from Greenland (Fig. 12). Our Mt. Washington glacial dipstick with its 1100 meters of relief provides a better opportunity to assess ice surface lowering than did our 300-meters of elevation range at Acadia National Park in Maine (Koester et al., 2017).
Although none will have quite as much relief, we are developing several other glacial dipsticks from mountains in the New England area, including the Chic Choc Mountains of Quebec (Fig. 1), for comparison with our Mt. Washington and Acadia National Park studies.

Figure 10. A LiDAR digital elevation model of Mt. Washington showing \(^{10}\)Be ages (white circles; black outlined boxes) with 1σ internal uncertainties (in ka). Bierman et al. (2015) uncertainty-weighted \(^{10}\)Be–\(^{26}\)Al ages are shown in boxes without outlines. Bedrock ages are grey italicized, boulder ages are black, and frost-riven block ages are blue italicized. One in situ \(^{14}\)C age for a summit frost-riven bedrock sample is shown, but six new in situ \(^{14}\)C ages along the elevational transect are not. The inset shows the location of Mt. Washington in New Hampshire.
Figure 11. $^{10}$Be ages of frost-riven blocks (blue circles), boulders (black circles), and bedrock (grey squares) taken from Mt. Washington, with horizontal bars showing internal error. The inset diagram shows all of the data, including samples with inherited $^{10}$Be nuclides.

Figure 12. Paleoclimate records from the Northern Hemisphere compared to our $^{10}$Be ages from Mt. Washington. (a) 44° N June insolation curve (Laskar et al., 2004), (b) Greenland ice core $\delta^{18}$O (NGRIP dating group, 2006), a proxy for North Atlantic temperature, (c) The NEVC modified from Ridge et al., 2012 (d) $^{10}$Be dipstick ages from Mt. Washington with $^{14}$C summit age (Bierman et al., 2015), and Androscoggin moraine (Bromley et al., 2015).

We are also sampling bedrock from 20 additional summits in the White Mountains as a check for $^{10}$Be inheritance to determine the elevational and spatial distribution of non-erosive, cold-based continental ice. Although $^{10}$Be inheritance is common in polar landscapes (Bierman et al., 1999, 2014, 2016; Davis et al., 1999, 2006, Marsella et al., 2000; Briner et al., 2006; Miller et al., 2006; Corbett et al., 2013, 2016; Margreth et al., 2016), our study is one of the first to suggest that non-erosive, cold-based ice sheets are a factor to be considered in temperate mountainous regions. As suggested in Bierman et al. (2015), variable glacial erosion rates between summits and valleys may play a strong role in development and maintenance of northern Appalachian topography through the Quaternary.
Bierman et al. (2015) also compared exposure ages from the summits of Katahdin and Mt. Washington to a global sea-level of ice volume record (Lambeck et al., 2014), as shown here in reduced form (Fig. 13). Our 20 new $^{10}$Be exposure ages also include several from the summit area of Mt. Washington that require multiple exposure periods prior to the LGM. However, recent work suggests that the St. Lawrence Lowland was free of Laurentide ice before about 31 ka (Parent and Dubé-Loubert, 2017), which along with the post-glacial incursion of the Champlain Sea as early as 14 ka (Lamothe, 1989; Parent and Occhietti, 1989, 1999), leaves less time for decay of in situ $^{14}$C in rock on the Mt. Washington summit areas created prior to overrunning by continental ice. Perhaps a local Appalachian ice sheet or ice cap covered the mountainous areas of northern New England during parts of the late Wisconsinan, an idea invoked long ago by Flint (1951). However, the moraine record in the lowlands adjacent to the Presidential Range to the north and the North American varve record from glacial Lake Hitchcock to the west suggest continental ice recession toward the north during deglaciation (Thompson et al., 1999, 2017; Ridge; 2004; Ridge et al., 2012; Bromley et al., 2015).

Figure 13. Schematic history of exposure samples from Mt. Washington reported in Bierman et al. (2015). Benthic $^{18}$O record proxy for global ice volume (Lambeck et al., 2014). Dark bars are uncertainty-weighted average ($^{10}$Be, $^{26}$Al) exposure ages. Twenty new $^{10}$Be exposure ages reported in this chapter show a similar distribution (see Fig. 11). White arrow: one in-situ $^{14}$C exposure age; six others are being analyzed. Gray shaded area represents five half-lives of $^{14}$C (~29 k.y.) required to decay $^{14}$C created prior to overrunning by Laurentide Ice Sheet. Regional deglacial age (16–14 ka) is shown by dotted line. LGM: Last Glacial Maximum.

ACKNOWLEDGMENTS

We thank the Mount Washington Auto Road and the Mount Washington Observatory, who have graciously agreed to assist us with this field trip. We also thank Brian Higginbotham, who assisted in the field, along with Chris Halsted and Anthony Vickers, who are now involved with the project as graduate students at Boston College.

ROAD AND TRAIL LOGS

Time, Place, Logistics

START TIME AND LOCATION: Trip begins on Sunday, October 1st, 7:30 AM, in the gravel parking area on the west side of NH Rte 16 to the immediate south of the Auto Road entrance (322363.00 m E, 4906302.00 m N). The base of the Auto Road is about 30 miles west of Bethel, Maine, and takes about 40 minutes to drive. From Bethel follow U.S. Rte 2 west to Gorham, NH, and then take NH Rte 16 south to the Auto Road entrance.
DESCRIPTION: This field trip will include a drive up Mount Washington’s Auto Road to the summit for examination of sample sites for cosmogenic nuclide exposure dating that are part of a NSF-funded research project to construct glacial dipsticks for the deglaciation history of northeastern United States and surrounding areas. Depending on the weather, those interested are invited to hike down the mountain about 7 km (about 4.5 miles) and 1200 vertical meters (about 4000 vertical feet) via the Nelson Crag Trail to visit additional sampling sites to those on the summit and along the Auto Road. The Nelson Crag Trail is steep and difficult, especially if wet, so those wishing to participate in this part of the field trip must be prepared for adversity. There also will be an opportunity for a guided tour of Mount Washington Observatory on the summit, where hot lunches and drinks also may be purchased. We will drive as a caravan to the summit, making two or three stops on the way up to observe sampling sites and the alpine landscape, including the Great Gulf cirque and the northern peaks of the Presidential Range. For those not hiking down, the field trip will end by 2 pm, allowing additional time for the drive home. For those hiking down, the field trip should end by about 5 pm.

Warning: Due to the fragile nature of the alpine ecosystem, please always walk on trails or rocks. Expect the possibility of extremely cold and unpredictable weather. Be prepared with proper clothing and good hiking boots for very rocky, uneven terrain, winter–like conditions, and perhaps extremely high winds. There are no bathroom facilities for those electing to hike down the Nelson Crag Trail. Vehicles must be consolidated.

Road Mileage

0.0 Begin at Mount Washington Auto Road gate, opposite NH Rte 16 from Glen House.

4.1 STOP 1. Park in small lot on the east side of the road just above the switchback and 4000 ft elevation post, and carefully walk across road for views of the Great Gulf cirque; the northern Presidential peaks, right to left, Madison, Adams, Jefferson; and the hanging cirques Madison, Gulf, Jefferson Ravine, and Sphinx Basin. Cosmogenic exposure age samples were not collected here.

5.3 STOP 2. Park in the vicinity of Cragway Spring at about 4800 ft elevation on the outside of the right hairpin turn and just above on the west side of the road. Be really careful crossing the road here! Both bedrock and boulders were collected near here for cosmogenic exposure dating. The bedrock exhibits beautiful glacial polish quartz veins and blebs, but unfortunately our one bedrock sample for exposure dating consisted of mostly feldspar rather than quartz. Those hiking down the Nelson Crag Trail in the afternoon will have the opportunity to examine boulders sampled downslope to the east from here.

6.5 STOP 3. Park in the large lot on the northwest side of the road at about 5700 ft elevation in an area known as the “Cow Pasture.” This elevation lies above the “lawns” in the Presidential Range, which are believed to be part of a pre-Quaternary surface known as the Presidential Upland, formed during prolonged periods of fluvial erosion beginning about 60 million years ago. The landscape above this elevation, including the cone of Mt. Washington (known at “The Rock Pile”), was not eroded by streams during the Tertiary or ice sheets during the Quaternary to the same degree as the Presidential Upland. A short 0.1-mile walk east to the junction of the Nelson Crag and Huntington Ravine Trails will allow examination of some large frost-riven blocks that were sampled for cosmogenic exposure dating.

8.0 STOP 4. Park in the lower summit parking lot at about 6200 ft elevation. A short walk up wood steps brings one to the summit complex, including the Mount Washington State Park building, which provides rest rooms, a cafeteria, and an optional tour of the Mount Washington Observatory, which has maintained continuous weather records on the summit since 1932. The summit sign at 6288 ft (1917 m) elevation marks the most prominent peak east of the Mississippi River. A few meters northwest from the summit...
sign, just north of the Tip Top House (a historic hotel, now museum, built in 1853 and the oldest surviving building on the summit), lie several frost-riven blocks of bedrock that we sampled for surface exposure dating. A short 0.1-mile walk southwest from the summit sign to “Goofer Point,” where several bedrock samples were collected for surface exposure dating, provides a fine view of Lakes of the Clouds, the adjacent A.M.C. hut with the same name, and Mt. Monroe behind, with its striking stoss-lee glacial erosional topography (gentle side to N20W, steep side facing S20E).

For those not hiking down the Nelson Crag Trail, the summit marks the end of trip. Please drive down the Auto Road carefully!

**Trail Mileage for hike down Nelson Crag Trail**

<table>
<thead>
<tr>
<th>Mileage</th>
<th>Description</th>
</tr>
</thead>
<tbody>
<tr>
<td>0.0</td>
<td>Trail log from the lower summit parking lot for hike down Nelson Crag Trail.</td>
</tr>
<tr>
<td>1.0</td>
<td><strong>STOP 1.</strong> Examine boulders sampled for exposure age along the Nelson Crag Trail where it lies about 0.1 mile east of the 6.5-mile mark at the “Cow Pasture” on the Auto Road, same area as Stop 3 on the drive up.</td>
</tr>
<tr>
<td>1.8</td>
<td><strong>STOP 2.</strong> Examine boulders sampled for exposure age along Nelson Crag Trail near the bump known as Nelson Crag.</td>
</tr>
<tr>
<td>2.9</td>
<td><strong>STOP 3.</strong> Examine boulders and bedrock sampled for exposure age along Nelson Crag Trail just below Cragway turn on Auto Road, same area as Stop 2 on the drive up.</td>
</tr>
<tr>
<td>4.6</td>
<td><strong>STOP 4.</strong> Continue down Nelson Crag Trail to intersection with Old Jackson Road (trail) near 2-mile mark on Auto Road and Lowe’s Bald Spot. Unfortunately, we were not able to collect exposure age samples near here, but collected samples from similar elevations from Square Ledge and the Glen Boulder Trail in Pinkham Notch.</td>
</tr>
<tr>
<td>6.5</td>
<td><strong>STOP 5.</strong> Follow Old Jackson Road south to A.M.C. Pinkham Notch Camp where vehicle shuttles will return hikers to the base of the Auto Road, about 3 miles north on NH Rte. 16.</td>
</tr>
</tbody>
</table>

**End of trip; thank you**

**REFERENCES CITED**


Miller, G.H., Briner, J.P., Lifton, N.A., Finkel, R.C., 2006. Limited ice-sheet erosion and complex exposure histories derived from in situ cosmogenic \(^{10}\)Be, \(^{26}\)Al, and \(^{14}\)C on Baffin Island, Arctic Canada. Quaternary Geochronology 1, 74-85.


NGRIP Dating Group, 2006, Greenland Ice Core Chronology 2005 (GICC05), IGBP PAGES/World Data Center for Paleoclimatology Data Contribution Series # 2006-118, NOAA/NCDC Paleoclimatology Program, Boulder CO, USA.


Shakun, J.D., Clark, P.U., He, F., Marcott, S.A., Mix, A.C., Liu, Z., Otto-Bliesner, B., Schmittner, A., Bard, E., 2012. Global warming preceded by increasing carbon dioxide concentrations during the last deglaciation. Nature 484, 49-54.


FIELD RELATIONS, PETROGRAPHY AND PROVENANCE OF MAFIC DIKES, WESTERN MAINE.

David Gibson dgibson@maine.edu, Donald Osthoff and Chase Rerrick.
Department of Geology, University of Maine at Farmington, Farmington, Maine 04938.

INTRODUCTION

In his 1965 report on the Geology of the Bryant Pond quadrangle, Maine, C.V. Guidotti mapped some 65 mafic dikes within this field area. However, his rather enigmatic map (see Figure 2 of his report) served only to inspire a group of undergraduates and their advisor to further investigate these intrusions. Guidotti (1965) assigned a preliminary Triassic age to these intrusions but to date no accurate age determinations have been performed on these rocks. Since publication of Guidotti’s report many other researchers have reported and mapped petrographically similar dikes that intrude further to the south into the Sebago Lake region.

The mafic dikes of the study area intrude a variety of rock types that are found in the general area of high-grade metamorphism in western Maine (Figure 1). Specifically, these dikes intrude the Devonian Songo granodioritic pluton (Stop1 on this field trip), the 293±2 Ma Sebago (sensu latto) two-mica granite (as displayed at Stops 3 and 4) and the metamorphic country rock envelope that surrounds both these intrusions. Mafic dikes are also observed intruding several of the pegmatities that form a distinct zone in western Maine, such as the Bumpus, Tamminen and Havey quarries.

In general the spatial and temporal occurrence of numerous dikes (forming a dike swarm) is a strong indication of extensional tectonics. Many other examples of dikes swarms are observed along the Maine coast such as the exceptional exposures around the Schoodic peninsula, but others are also observed throughout New England and in Maritime Canada. Therefore there are a number of possible sources for the western Maine mafic dike swarm. They could be related to the Eastern North American (ENA) magma series, which are remnants in N. America of the Central American Magmatic Province (CAMP). These are assumed to be ~ 200 Ma and are predominantly tholeiitic in composition. However, they might also be related to the older (~ 225 – 230Ma) Coastal New England (CNE) magma series, which have a definite alkaline affinity and maybe related to a deeper “plume” source. There are distinct geochemical parameters (as discussed below) that can be utilized to distinguish between these magma series. In addition, the western Maine dike (WMD) suite could be related to the White Mountain Magma Series, a group of mostly plutonic rocks of alkaline affinity. The latter are represented in western Maine by small stocks, e.g. Rattlesnake Mountain, that intrude the Sebago granite and have described by Creasy in the Jackson volumes (1989).

Our field trip today is a north to south traverse that will enable examination of the range of dikes that constitute the WMD series. The main objectives are to examine -

a) the petrography of these dikes in an extended area from Bethel in the north, to Sebago Lake in the south (see Figure 9),
b) their field relationships, specifically their orientations and any cross cutting evidence that may pertain to one or more sets of dikes(?),
c) any relationships between the host rock and dike petrography,
d) the provenance of the WMD swarm, specifically their major and trace element composition to assess if they have a geochemical affinity with other dike swarms that occur regionally (as outlined above).

GENERAL GEOLOGY OF THE STUDY AREA

The igneous and metamorphic geology of western Maine has been the focus of many NEIGC field trips over the years and therefore we will highlight only the salient details in this section. Readers are referred to the following NE GSA and NEIGC field trip guides for more comprehensive information – Guidotti et al. (1986 and 1996), Solar et al. (2001, 2009 and 2016). The main rock units that host the mafic dikes are the Songo pluton, the pegmatitic intrusions and the Sebago “batholith” (sensu latto).

The Songo pluton is a post-Acadian intrusion that covers approximately 350 km². It is a medium grained, equigranular granodiorite that displays some textural and compositional variation. Initial age dates for the pluton...
suggest a crystallization age of ~ 380 Ma. However, new LA ICP-MS zircon data reveal a much younger age of 364 ± 1.3 Ma for the Songo (pers comm. van Rooyen, 2017). For the most part the pluton is an unfoliated granodiorite with biotite ± hornblende ± sphene, a typical I-type granite mineralogy (Chappell and White, 2001). However, in other (mappable) areas (see Figure 2 of Gibson and Lux, 1989) the Songo pluton lacks hornblende, the biotite has a distinctive red/brown sheen often indicative of high TiO₂ contents and can be strongly foliated (Stop 1). The latter areas are coincident with the proximity of the Sebago granite and/or pegmatitic bodies that also often contain sheet-like bodies of leucocratic granite not dissimilar to the Sebago itself. In addition, ⁴⁰Ar/³⁹Ar dating of hornblende from the Songo pluton (Lux and Aleinikoff, 1985 and Lux et al. 1989) yielded significantly younger ages than the crystallization age of the pluton. These younger, reset hornblende ages led Guidotti et al. (1985) to suggest that this metamorphic event was related in time and space to the emplacement of the Sebago pluton. However, subsequent research has cast doubt on this idea.

**Pegmatitic intrusions** are ubiquitous throughout western and south-western Maine and are often associated with the Sebago granite in the southern part of the Oxford pegmatite field. Wise and Francis (1992) classified them as LCT REE-granite pegmatites and they fall into three categories – 1) simple pegmatites that are composed of K-feldspar, quartz, muscovite and bitoite and often display classic graphic intergrowths, 2) beryl-type pegmatites containing beryl along with albite (cleavlandite), and 3) Li-enriched pegmatites of the spodumene subtype. In many of these pegmatitic intrusions mafic dikes are observed, e.g. Songo Pond, Bumpus, Tamminen, and Havey.

**FIGURE 1.** Bedrock map of western Maine as displayed on the 1985 State map (Osberg et al. 1985) compared to current map of the major rock units (after Solar and Tomascak 2009).
The Sebago batholith/pluton et al. Although originally mapped as a single intrusion, as shown on the 1985 bedrock map of Maine (Osberg et al. 1985), the Sebago “batholith” has been shrinking in size ever since. The detailed research of Gary Solar and Paul Tomascak (2001, 2009 and 2016), and their students, have shown that the Sebago granite pluton covers an area of only 400 km² which roughly encircles Sebago Lake. They subdivided the previously mapped batholith into a more homogeneous core of two-mica, fine to medium-grained granite, the Sebago granite pluton, and a heterogeneous area surrounding it. The latter area is highly variable composed of two-mica granite and granodiorite, scheiiric granite, pegmatite dikes, metapelitic pelites and diatexites. This rock “smorgasbord” surrounds the Sebago granite and is referred to as the Sebago Migmatite Domain or more recently the Migmatite-Granite complex (Solar and Tomascak, 2016). We will see some of these rocks at our final stop on the field trip.

FIELD RELATIONSHIPS AND PETROGRAPHY OF THE MAFIC DIKES

The mafic dikes that intrude all the above country rocks are variable in size, petrography and orientation. They can be broadly subdivided into three main types – 1) Thin, < 1m wide, basaltic dikes, 2) Larger, > 1m wide, basaltic to basaltic andesites, and 3) Trachy basalts which have a distinctive glomeroporphrytic texture.

TYPE 1 DIKES – These basaltic dikes are usually between 60 cm to a few cms in width and have narrow chilled margins (< 1cm) with the host rocks. They are most commonly aphyric in texture, but some examples are porphyritic with small, ~ 1cm long, plagioclase lathes. Photomicrographs showing the textures of these basaltic rocks are shown in Figures 2, 3 and 4, which also show the abundance of plagioclase lathes often arranged in a pseudo-radiating texture indicating quenching. In addition, abundant pyroxene is present with the distinctive reddish/brown color of titanaugite. Future mineral chemistry studies should help confirm their detailed mineralogy. They are vertically inclined and trend from NNE – SSW to NE – SW. We will see examples of the Type 1 dikes at Stops 1 and along the large road cut of Stop 5. At the latter locality some of the thin vertical dikes are displaced by minor sub-horizontal faults although they display ample evidence to confirm that they are post deformation.

FIGURE 2. Photomicrographs of WMD-2 showing aphyric texture with semi-radiating plagioclase lathes and possible titanaugite. PPL on the left; CPL right. Field of view 3.5mm.

---

1 Which in fact is a good thing as the Sebago “batholith” was considered as a potential repository for nuclear waste!!
FIGURE 3. Photomicrographs of WMD-4 showing aphyric texture with abundant possible titanaugite and skeletal opaques indicative of quenching. PPL on the left; CPL right. Field of view 3.5mm.

FIGURE 4. Photomicrographs of WMD-6 showing plagioclase-phyric texture. PPL on the left; CPL right. Field of view 3.5mm.

TYPE 2 DIKES – The larger basaltic to basaltic andesite dikes have a distinct chilled margin with the host rocks, as shown in Figure 5. Away from their margins they have a coarser-grained texture appearing more diabasic and can display ophitic and subophitic textures, Figure 5. The larger Type 2 dikes commonly contain xenoliths of their host rock and also larger selvages of granite, as we will observe at Stop 5. These dikes vary in inclination and orientation. Most of them are vertical but some notable exceptions dip steeply, generally to the SE. They also have a more consistent trend to the NE – SW.

TYPE 3 DIKES – These dikes are distinct from the other two types as they are lighter in color with clusters of minerals forming a glomeroporphyritic texture. These areas are 2 – 3cm across and possibly contain Na-rich pyroxenes, Figure 6. The groundmass is fine-grained with abundant plagioclase, pyroxene and some olivine. These dikes are vertically inclined but have a more definite E – W trend.
FIGURE 5. Photomicrographs of WMD-7. Top set show the chilled margin of this dike in contrast with the coarser-grained diabase interior, bottom set. PPL on the left; CPL right. Field of view 3.5mm.

FIGURE 6. Photomicrographs of WMD-9a, the glomeroporphyritic trachy basalt. On left the groundmass texture of this dike; on the right one of the glomerocrysts. PPL on the left; CPL right. Field of view 3.5
GEOCHEMISTRY

As this is a field trip we will not delve too much into the geochemistry of these dike rocks. However, given their fine-grained textures it is necessary to utilize geochemical data in order to classify them. Geochemical discrimination diagrams (Pearce and Cann, 1973) can aid in examining the tectonic setting of these intrusions. In addition, several geochemical parameters can be employed to discriminate between dikes of potential magma sources, i.e. the Coastal New England (CNE) suite and the Eastern North America (ENA) suite (Dorais et al., 2005 and McHone, 1992). In particular the levels of TiO$_2$ and Zr are higher in the CNE series compared to the ENA, and the CNE dikes are enriched in Nb whereas the ENA series have Nb and Ti anomalies.

25 samples have been collected from western Maine and analyzed for 18 major, minor and trace elements. Analyses were performed in the UMF XRF laboratory using standard procedures of sample preparation and analysis, including USGS standard materials. Only those samples that are definitely post deformation have been plotted on the following diagrams.

Figure 7 demonstrates that the WMD are predominantly alkaline although some subalkaline compositions are evident. The other bivariate diagrams show the WM dike compositions in comparison with the fields for dikes of the CNE and ENA series. It would appear that the WM dikes are more closely related to the CNE suite than the ENA dikes. The discrimination diagrams, Figure 8, not surprisingly demonstrate that the WM dikes are within plate basalts with a spread of alkali and tholeiitic compositions.

FIGURE 7. Bivariate diagrams showing the range of compositions for the WM dikes. (A) Plot of total alkalis V SiO$_2$ wt.% ; Plots (B), (C), and (D) plot geochemical parameters, suggested by McHone (1992) and Dorais (2005), to discriminate between the CNE (red field) and ENA (blue field) dike suites.
FIGURE 8. Basaltic tectonic discrimination diagrams (Pearce and Cann, 1973) showing the within plate signature and alkalic nature of the WMD series.

SUMMARY

The WMD series are predominantly aphyric basalts and diabases although some have higher amounts of the alkali elements and are trachy basalts. These are most likely related to the alkaline Mesozoic intrusive series as described by Creasy (op cit). They are predominantly post deformation although some may have experienced late stage faulting. Many examples may contain titaugite though this needs to be confirmed by electron microprobe analysis. They have higher TiO₂ contents than the ENA series and display greater affinity to the CNE suite of dikes. Their trace element concentrations place them in the within-plate fields on the standard basaltic discrimination diagrams. In conclusion, the majority of western Maine dikes have a close geochemical affinity with the Coastal New England magma series and this extends the geographic area over which this series is observed.

ACKNOWLEDGEMENTS

We would like to thank Dan Mason for extensive help with sample preparation and fieldwork. UMF students Dylan Moreau, Tom Alexander, Sumaya Hamdi also helped with fieldwork. We also thank Layne Nason for his skill and expertise in drafting some of the figures.

ROAD LOG

MEETING POINT: Meet at the Casablanca Cinema 4 parking lot on Cross Street in Bethel (44° 24' 40.73" N; 70° 47' 30.64" W) at 8:30am on Sunday, October 1st. This parking lot is beside the Bethel Area Chamber of Commerce. There will be limited opportunities to get food so you might consider bringing lunch with you. The trip will end close to access roads for I 95 north and south. The road map for this field trip, from Bethel to near Gray, is displayed in Figure 9.
FIGURE 9. Map of the route taken on this field trip.
MILEAGE

0.0 Turn right out of the parking lot onto Cross Street and then take next right onto Mechanic Street. Take another right after crossing the railway tracks onto Railroad Street. The road then goes under Rte 2, and merges in 0.7 miles with Rte. 2 W. In 0.1 miles turn left onto Rtes 5 & 35, Songo Pond Road. Continue S passing Songo Pond on your left.

8.0 Park on shoulder. (NOTE: this is a solid hard shoulder but others on the field trip are not. Please try to ensure you always keep two wheels on more solid surfaces)

STOP 1. DOUGLASS QUARRY. PETROGRAPHY OF THE SONGO GRANODIORITE, PEGMATITE AND LEUCOGRAINITE SHEETS ALL INTRUDED BY MAFIC DIKES.

Access to the quarry is by permission only and if you wish to revisit this locality you must get permission from the owners.

At this locality we can observe a couple of thin, < 1m, mafic dikes intruding highly foliated Songo granodiorite, a leucogranite sheet and pegmatite. The Songo here is typical of the deformed petrographic variant and the red/brown sheen of the biotites can be clearly observed. The mafic dikes are aphyric basalts with narrow chilled margins. The field relationships clearly show that the mafic dikes are younger than the other rock types in this quarry, Figure 10.

FIGURE 10. Stop 1 Douglass quarry showing thin basaltic intruding the Songo granodiorite (at back wall of quarry) and cutting leucocratic sheets and pegmatite. Insert shows close up of contact with chilled margin.

Continue S on Rtes 5 & 35.

13.0 Turn left onto Rte 35 S. (NOTE: Interesting signpost at this intersection, a Maine peculiarity!)
14.0 Jct. with Rte. 118 stay on Rte. 35.
18.0 Jct of Rte. 35 with Rte. 37, continue on Rte. 35.
21.0 Right hand turn onto combined Rtes. 37 and 35 to Bridgton. Stay on Rte. 37 after Bear Pond on right
Reduced speed zone going through N. Bridgton passing Bridgton Academy on right.
25.0 Pull off onto shoulder just beyond the junction with Rte. 117.

STOP 2. MAFIC DIKES INTRUDING THE SEBAGO GRANITE.

At this locality two mafic dikes are observed intruding Sebago granite like country rock. Even though the main area
of the Sebago pluton is observed further to the south isolated areas of similar rocks are found within the Migmatite-
Granite complex. Chilled margins to the dikes and granite xenoliths within them clearly demonstrate the intrusive
nature of the dikes.

FIGURE 11. Stop 2 - examples of basaltic dikes with chilled margins, xenoliths of country rock and apophyses.

Continue heading S on Rte. 117

28.0 Jct. with Rte. 302 turn right and continue on Rte. 302.
29.0 Major intersection in Bridgton. Stay on Rte. 302 E to Naples (Maine that is!).
43.0 Intersection with Rtes. 35 and 11 south of Naples. Turn left.
47.0 Turn L onto Rte. 121.
48.0 Park at second layby on right hand side and walk up to the outcrop opposite Birch Terrace.
STOP 3. GLOMEROPORPHYRITIC TRACHY BASALT DIKE.

At this locality there is an excellent example of a trachy basalt dike intruding the Sebago granite. Notice the crystal clusters that dominate the texture of this rock. Are these glomerocrysts or could they be small xenoliths? Further along the outcrop look for a thin basaltic dike like other examples we have observed on this field trip.

Return to vehicles and reverse direction going S on Rte. 121.

49.0 Junction with Rtes. 35 and 11, turn right.

53.0 Turn left onto Rte. 302. Take care this is a very busy junction.

56.0 Park on shoulder just beyond Rabbit Lane and before Varney Hill Road.

STOP 4. A VARIETY OF DIKES INTRUDING SEBAGO GRANITE.

At this locality there are three dikes that intrude the Sebago granite. Our objectives at this outcrop are to locate the dikes at this locality; secondly examine their petrography. Are they all the same? Thirdly, are there any cross cutting relationships that could give relevant information regarding the timing of these intrusive events?

![FIGURE 12. Two types of “porphyritic textures observed in the dikes at Stop 4.](image)
STOP 5a. RTE. 26 SECTION - N. NUMEROUS MAFIC DIKES AT THE EASTERN MARGIN OF THE MIGMATITE-GRANITE COMPLEX.

CAUTION: Rte. 26 is a very busy road and we must exercise extreme caution here. Please cross the road in the designated area and only when you are told it is clear to do so.

At this locality numerous mafic dikes intrude the leuco- and melano-somes of the migmatitic envelope of the Sebago pluton as defined by Solar and Tomascak (2016). Firstly examine the dikes on the west side of the road and then the eastern side. Several questions need to be addressed here - the variations of dike petrography, number of intrusive events and the disappearance of dikes from one side of the road to the other?

Return to vehicles, turn right onto Rte. 26 and continue south. As we drive south passengers note the numerous mafic dikes intruding this exceptionally well exposed migmatitic zone. Drivers keep your eyes on the road!

STOP 5b. RTE. 26 SECTION S. MULTIPLE GENERATIONS OF MAFIC DIKES INTRUDING THE EASTERN MARGIN OF THE SEBAGO/MIGMATITE DOMAIN.

At this locality there are different generations of dike intrusion. Cross cutting relationships can be definitely inferred if not observed directly. Again examine the petrographic variations of the dikes, their orientations and relationships with the migmatitic country rocks, see Figures 13 and 14 below.

FIGURE 13. Large mafic dike intruding the leucosome and melanosome of the migmatitic margin to this complex.
This concludes our fieldtrip. Turn left out Sabbathday Road onto Rte. 26 and 26a to access I 95 N and S.

REFERENCES CITED


TRANSECT FROM THE MIGMATIZED CENTRAL MAINE BELT TO THE BRONSON HILL ANTICLINORIUM

By

J. Dykstra Eusden, Dept. of Geology, Bates College, Lewiston, Maine 04240
Sarah Baker, Dept. of Geology, Bates College, Lewiston, Maine 04240
Jordan Cargill, Dept. of Geology, Bates College, Lewiston, Maine 04240
Eric Divan, Dept. of Geology, Bates College, Lewiston, Maine 04240
Ian Hillenbrand, Dept. of Geology, Bates College, Lewiston, Maine 04240
Paul O'Sullivan, GeoSep Services, 1521 Pine Cone Road, Moscow, ID 87872
Audrey Wheatcroft, Dept. of Geology, Bates College, Lewiston, Maine 04240
email: deusden@bates.edu  cell: (207) 240-9150

INTRODUCTION

This trip will visit outcrops along a transect from the migmatized Central Maine Belt in Maine to the Bronson Hill Belt near the New Hampshire-Vermont border. We will roughly follow U.S. Rte. 2 starting in West Bethel, ME, with the migmatized Silurian Rangeley Formation and intrusions of two mica granite and pegmatite, then proceed west to Randolph, NH, and the Ordovician Ammonoosuc Volcanics and Oliverian Jefferson Dome. The trip ends in Lancaster, NH with the Cambrian Albee Formation and Ordovician Lost Nation Pluton (Figure 1). This is a good trip for those who are heading west or southwest after the conference.

The bulk of the research presented here was generously supported by the Maine and New Hampshire Geological Surveys through the U.S.G.S. StateMap program for bedrock mapping in the Bethel and Gilead, Maine and Jefferson and Mt Washington East, NH 7.5' quadrangles. Dr. Paul O'Sullivan of GeoSeps Services, Moscow, ID has provided the very important absolute age control with LA-ICP-MS ages on detrital and crystallization zircons. Much of the work was done by Bates College Geology seniors working on their theses and they are all co-authors of this fieldguide. Eusden is extremely grateful to all of them for their friendship and hard work.

BETHEL-GILEAD, MAINE; MIGMATIZED CENTRAL MAINE BELT

Previous Work
In the Maine portion of the fieldtrip, the bedrock geology has been summarized in regional maps by both Osberg et al. (1985) and Moench et al. (1999). The tectonic setting of the Bethel-Gilead region is within the Central Maine Belt, at the north end of the Sebago batholith, within the Migmatite-Granite Complex of Solar and Tomascak (2016), and along the Piscataquis Volcanic Arc (Bradley and Tucker, 2002, and Bradley et al., 2000). In the Gilead-Bethel Maine region the metasedimentary rocks shown on Osberg et al. (1985) and Moench et al. (1999) are dominated by the Devonian Littleton Formation with narrow belts of the Silurian Madrid, Smalls Falls, and to a lesser extent the Rangeley Formation. Brady (1991) mapped the eastern portion of the Bethel quad as part of his M.S. thesis at Orono, breaking out many units that were variably correlated to portions of the Siluro-Ddevonian Rangeley stratigraphy of Moench and Boudette (1987). The Rangeley stratigraphy was extended through this area and into adjacent New Hampshire by Hatch et al. (1983) and those correlations formed the basis of the most recent lithotectonic compilation of the Appalachians by Hibbard et al. (2006).

Gibson et al. (2017) have recently dated the Songo granodiorite to 364 +/- 1.3Ma.. The Songo pluton and numerous small quartz diorite-tonalite plutons we've mapped in the region (Eusden et al. 2017a) are part of the Piscataquis Magmatic Arc (Bradley and Tucker, 2002, and Bradley et al., 2000) that developed syntectonically on the leading edge of the migrating Acadian orogenic front in the Devonian.

Recent work done by Solar and Tomascak (2016) on the Sebago Pluton and the Migmatite-Granite Complex of western Maine suggests the Sebago is much diminished in map size, is 288 ± 13 to 297 ± 14 Ma, and is surrounded by rocks showing regional migmatization that occurred around 376 ± 14 Ma. Thus migmatization occurred well before the intrusion of the Sebago Pluton. Bradley et al. (2016) have dated many of the small pegmatite bodies in this region and some (e.g. Mt. Mica pegmatite) are as young as Permian (264-260 Ma), likely related to the Alleghanian orogeny.
Current Work
Twelve new metasedimentary units have been mapped in the region and define a stratigraphy that overall best correlates to the Silurian Rangeley Formation (Figure 2). The new stratigraphy has terrible topping control with only a few scattered graded beds found in the region and it is pervaded by migmatite obscuring primary sedimentary features. Hence, correlation by lithologic similarity and sequence is the most useful, but the least certain, method to determine stratigraphic age. To provide absolute age control we have done detrital zircon age dating of two samples from one of the metasedimentary units, the Bog Brook Granofels, as well as crystallization zircon ages for two different cross cutting two-mica granites from Wheeler Mine (Wheatcroft, 2017). The presence of belts of rusty weathering schists and quartzites, biotite and calc-silicate granofels, and ubiquitous calc-silicate pods, blocks, and lenses are hallmarks of the Rangeley Formation in nearby New Hampshire (Eusden et al., 1996) and strengthen the arguments for the correlation to the Rangeley Formation. The new stratigraphic assignments will require significant changes to regional maps in western Maine and adjacent New Hampshire.

Detrital zircon age population density histograms for both Bog Brook granofels samples yield maximum depositional ages of 419.5 ± 3.9 Ma and 427.6 ± 3.9 Ma (Wheatcroft, 2017; Wheatcroft et al., 2017). Taking both ages together, the mean maximum depositional detrital zircon age of circa 422 Ma places this unit within the traditional age assignment for the Late Silurian Madrid Formation (Hatch et al., 1983). If the Bog Brook Granofels was the Madrid Formation, as correlated by traditional age assignments, one would expect it to be directly in contact with rusty schist, equivalent to the Smalls Falls Formation. That is not the case, instead the Bog Brook granofels is surrounded by the gray schist and quartzite with calc-silicate pods that are all characteristic of the Early Silurian Rangeley Formation to which we have correlated the Bog Brook Unit.

Interestingly, our younger detrital zircon ages are consistent with new detrital zircon ages done by Bradley and O’Sullivan (2016), which also yielded consistently younger than expected ages for many of the type sections in the Rangeley stratigraphy. Our new detrital zircon age spectra match incredibly well the first, inboard-derived cycle of the Ganderian-derived sediments (composite detrital zircon spectrum of Quimby, Rangeley, Perry Mountain and Smalls Falls Formations) within the Central Maine Basin (Bradley and O’Sullivan, 2016). The age discordance between traditional stratigraphic ages and newer detrital zircon-based ages does suggest that much more detrital zircon geochronology is needed for the Rangeley Stratigraphy and indeed remapping and reassessment of the type section may be in order. At a minimum, a middle Silurian rather than earliest Silurian age for the Rangeley Formation seems a logical conclusion from our work and that of Bradley and O’Sullivan (2016) (Figure 3).

Crystallization ages for two, two-mica granite sills from Wheeler Mine yielded ages of 349.2 ± 2.1 Ma and 355.3 ± 2.3 Ma respectively. This supports their intrusion as small mappable granites that post-date the Late Devonian-Carboniferous Migmatite-Granite Complex (376 ± 14 Ma) of Solar and Tomascak (2016) but pre-date the younger Sebago pluton (288-297 ± 14 Ma). These granites cut stromatic layering in the migmatized metasedimentary rocks as well as the macroscale Neoacadian D3 open, reclined folds (Divan et al., 2017) defined by the map pattern. The detrital zircon and crystallization zircon ages bracket the deposition, deformation, and metamorphism of this area to within a 65-79 million year time period. The numerous pegmatites exposed at Wheeler Mine cut the two mica granites as well as the Bog Brook Granofels. These have not been dated but are likely as young as Permian based on the recent pegmatite ages determined by Bradley et al. (2016).

In summary, the major geologic events in the Gilead and Bethel, Maine region are: (1) end of Rangeley Formation deposition at 422 Ma with minor D1 pre-metamorphic faulting; (2) Acadian D2 nappe-stage folding and the onset of regional metamorphism; (3) widespread migmatization occurring around 376 Ma (Solar and Tomascak, 2016); (4) Neoacadian D3 open, reclined folding; (5) Neoacadian emplacement of the Songo Pluton at 364 ± 1.3 Ma (Gibson et al., 2017), (6) Neoacadian two-mica granite emplacement at 349.2 ± 2.1 and 355.3 ± 2.3 Ma; and (7) Alleghanian emplacement of the Sebago-type plutons and even younger pegmatites within the region between 294-260 Ma (Solar and Tomascak, 2016; Bradley et al., 2016).
Figure 2. Bedrock geologic map of the Gilead, Maine region showing metasedimentary and igneous units. STOP 1 at Wheeler Mine is located by the white ellipse. Crystallization zircon ages (CZ) and detrital zircon ages (DZ) in Ma are indicated by the green and red dots respectively. Bedrock map made by Audrey Wheatcroft and cross section made by Eric Divan, both Bates ’17.
Figure 3. Detrital zircon ages for Bog Brook Granofels at (A) Bog Brook (427.6 Ma) and (B) Wheeler Mine (419.5 Ma). C: Bog Brook Granofels interpreted depositional age (422 Ma) compared to other detrital ages from Bradley and O’Sullivan (2016) all showing younger than expected ages. D and E: Crystallization ages for two mica granite plutons (349 and 355 ma) from the Wheeler Mine region. Credit: Audrey Wheatcroft.
Previous Work
The tectonic setting of the Randolph Valley region is along the very east edge of the Bronson Hill Belt where it is in contact with the Central Maine Belt and where local dips are to the E-SE. The Presidential Range bedrock geology was first mapped in any detail by the Billings in the late 1930’s and early 1940’s. Their efforts culminated in several papers and the 1979 Geology of the Mount Washington Quadrangle report published by the New Hampshire Department of Resources and Economic Development (Billings et al., 1979). Hatch et al. (1983) made many important contributions to the region, most notably the extension and correlation of the Rangeley, Maine, stratigraphy across the border into New Hampshire, through the Presidential Range, and beyond to the south. The New Hampshire bedrock geologic map (Lyons et al., 1997) updated these previous efforts and more or less retained the contacts shown by the Billings’ and the stratigraphic assignments of Hatch and Moench (1984).

Our decades-long bedrock mapping campaign in the Presidential Range culminated in our finished bedrock geologic map (Figure 4) and report (Eusden, 2010). Other publications by our group include Eusden et al. (1996, 2000) and Roden-Tice et al. (2012). Eusden et al. (1996) has maps of several structural fabrics but the bedrock map does not include the lower wooded elevations. Roden-Tice et al. (2012) focuses on the apatite fission track exhumation history of the Presidential Range. In 2013 we published a layperson’s book, “The Geology of New Hampshire's White Mountains”, which covers many aspects of the geologic history in the Presidential Range (Eusden et al., 2013). The outcrops visited in the Randolph valley for this trip were also visited on two earlier NEIGC’s (Eusden et al., 2006 and 2009).

Current Work
The unusual aspect of the Bronson Hill Anticlinorium in the Randolph Valley is that the region between it and the Central Maine Belt lacks the mantling shelf facies of the Silurian Clough and Fitch Formations. Instead, there is an abrupt boundary with shear fabrics separating migmatized Rangeley Formation from the Ammonoosuc Volcanics, which we have interpreted as the Mahoosuc Fault. Internal to the Bronson Hill Belt, there is another discontinuity separating the Ammonoosuc Volcanics from the Oliverian Jefferson Dome again with ductile shear fabrics that we have interpreted as the Moose River Fault.

The most significant result of our mapping has been to better define the contact between the Ammonoosuc and Oliverian Jefferson Dome. Over the years, the nature of the Ammonoosuc-Oliverian contact has been debated with some researchers supporting an intrusive contact (Billings et al., 1979; Moench and Aleinikoff, 2003) and others a fault contact (Hollocher, 1993; Kohn and Spear, 1999). In the Randolph valley, Billings et al. (1979) show the contact to be delineated by silicified pods that line up to form a late (presumably Mesozoic) brittle fault. Dupee et al. (2002) and Foley and Eusden (2009) were able to subdivide the Ammonoosuc Volcanics into a variety of different lithologic units including amphibolite, rusty gneiss and gray gneiss. They have not been able to easily correlate this new subdivision to the upper and lower Ammonoosuc designations on the state map (Lyons et al., 1997). The maps produced by Dupee et al. (2002) and Foley and Eusden (2009) show that the Ammonoosuc stratigraphy is cut off against the main contact with the Jefferson Oliverian Dome by the Moose River fault. Foley and Eusden (2009) have also mapped meter-scale intercalations, presumably fault slivers or horses, of Ammonoosuc Volcanics and Jefferson Dome rocks. In addition, several exposures of S-C mylonites that exhibit variable intensity have been mapped within the Ammonoosuc and Jefferson Dome rocks adjacent to the main contact (Dupee et al., 2002; Foley and Eusden, 2009). Shear sense for the mylonites is complex with both normal and reverse motions found in the Ammonoosuc Volcanics and reverse motion only in the Jefferson Dome rocks. The foliations and lineations in the mylonites eventually die out away from the main Ammonoosuc-Jefferson Dome contact. It is tempting to interpret the normal motions as doming-related, having developed as simple shear dominated along the flanks of the rising dome. The principal boundary between the Ammonoosuc and Jefferson Dome is probably a structure of this type and thus we interpret the Moose River fault as a late Acadian, normal sense, shear zone. The reverse motion indicators in the both the Ammonoosuc Volcanics and Jefferson Dome are not what one would expect during doming, and since doming is likely the last phase of Acadian deformation in the area, it is possible that the reverse motion indicators represent a pre-doming shear history. However, there are no age constraints on the different mylonites.
Figure 4. Bedrock geologic map of the Randolph, NH valley modified from Eusden (2010). Location of STOP 2 indicated by white oval. Jefferson Dome in blue and Ammonoosuc Volcanics in purple and brown colors. Monazite geochronology shown by red dots with ages in Ma listed next to the dot.
Foley and Eusden (2009) performed monazite microprobe age determinations on two samples, one from a Jefferson Dome mylonitized sample in Cold Brook (sample MFC02, Stop 2a on this trip) and the other from an Ammonoosuc Volcanics mylonitized rusty gneiss (sample MFS01) in Snyder Brook, just above the main contact with the Jefferson Dome. The oldest monazite age population gives a late Silurian date of 415 +/- 7.2 Ma in MFS01 from the Ammonoosuc Volcanics. Both samples contain ages in the early to middle Devonian with MFC02 showing ages of 392 +/- 3 and 380 +/- 3 Ma and sample MFS01 giving an age of 382 +/- 4.4 Ma. The samples also show middle to late Devonian ages in the 370 Ma range with MFC02 yielding an age of 367 +/- 3.2 Ma and MFS01 an age of 370 +/- 6.2 Ma. Finally both samples show late Devonian to Early Carboniferous ages with MFC02 yielding an age of 346 +/- 3.6 Ma and MFS01 an age of 357 +/- 6.4 Ma. The Silurian age may be an early, pre-Acadian, shearing event when the Ammonoosuc and Jefferson Dome initially came into contact with each other. The early Devonian ages are consistent with other ages we have determined for the peak of Acadian metamorphism in the Presidential Range (circa 400-380 Ma). The middle Devonian 370 ages may represent the age of doming. The Devonian-Carboniferous ages are probably related to the intrusion of the local Bickford Granite that has been dated to 363 Ma.

LANCASTER, NH: ALBEE FORMATION AND LOST NATION PLUTON

Previous Work
The tectonic setting of the Lancaster region is on the west dipping edge of the Bronson Hill and west flank of the Jefferson Oliverian Dome. The earliest detailed mapping was done by Chapman (1942) and later compiled and republished as part of the Mt Washington, NH 15' quadrangle bedrock map and bulletin by Billings et al. (1979). These pioneering geologists established the stratigraphic framework for the region identifying the classic New Hampshire units of the Cambrian Albee Formation and the Ordovician Ammonoosuc Volcanics, Oliverian Jefferson Dome, and Lost Nation Pluton. The Ammonoosuc Volcanics are found mantling the surrounding flanks of the Oliverian domes (Leo, 1991). The emplacement of the Ammonoosuc Volcanics (461± 8 Ma) over the Oliverian plutons (Jefferson Dome: 454±5 Ma) (Moench and Aleinikoff, 2003) is still debated. Dorais et al. (2008) identified continental arc signatures in the Oliverian, Highlandcroft, and Lost Nation suites and suggested that these are all part of the same arc complex. Dorais et al. (2012) further identified the Ammonoosuc Volcanics as having a peri-Gondwanan isotopic signature while the Oliverian, Highlandcroft and Lost Nation rocks have Laurentian signature. Dorais et al. (2012) proposed that obduction of the Ammonoosuc Volcanics along a west directed thrust onto composite Laurentia emplaced it above the Oliverian rocks. Presumably this assembly occurred at the end of the Taconic Orogeny (Maclnnes et al. 2014; Taconic III of van Staal et al., 2009). Working just west of the Jefferson, NH 7.5' Quadrangle, Rankin et al. (2013) established that the Albee Formation was separated from the Ammonoosuc Volcanics by the Penobscottian unconformity caused by the Cambrian to Early Ordovician Penobsottt Orogeny. Rankin et al. (2013) also endorsed the correlation of the Albee Formation with the Dead River Formation in Maine and subdivided the Albee Formation into several members. Karabinos et al. (2017) recently proposed that the Bronson Hill arc formed on the newly baptized Moretown Terrane and that the western edge of Ganderia is buried somewhere east of the Bronson Hill. Our recent work has culminated in new bedrock maps of the Mt Dartmouth and Jefferson, NH 7.5’ quadrangles, is the subject of this part of the trip, and discussed below (Eusden et al., 2015 and 2017b).

Current Work
New mapping and detrital zircon geochronology in the northern part of the Jefferson 7.5’ Quadrangle has revealed a previously unknown region of Cambrian Albee Formation along with the previously recognized Ordovician Ammonoosuc Volcanics, intrusive rocks of the Ordovician Oliverian Dome and Lost Nation-Highlandcroft Pluton, and Jurassic igneous cone sheets of the Pliny Caldera Complex (Figure 5) (Hillenbrand, 2017). Detrital zircon geochronology was conducted on a total of four samples of thinly bedded to pin-striped quartz-rich rock all but one previously thought to be the Albee Formation (Figure 6) (Hillenbrand, 2017; Hillenbrand et al., 2017). Two samples came from lower elevations near Tug Mtn. with one previously mapped as Albee and the other as Ammonoosuc Volcanics (Lyons et al., 1997). The other two samples are roof pendants in Jurassic granite from the higher peaks of Terrace Mtn. and both previously mapped as Albee Formation xenoliths (Chapman, 1942).
Figure 5. Bedrock geologic map and cross sections of the Jefferson, NH 7.5' quadrangle. Location of STOPs 3 and 4 are indicated by the white ovals. Detrital zircon ages shown by red dots with ages in Ma listed next to the dot.
Figure 6 Detrital zircon ages from the Jefferson quadrangle, NH. Cambrian Albee Formation samples (A and B) yielding ages of 522 and 545 Ma. Devonian Tarantola Formation (C and D) yielding ages of 415 and 406 Ma.
The two lower elevation samples on Tug Mtn. yielded youngest detrital zircon age peaks of 522 ± 3.8 Ma and 545 ± 17 Ma respectively, supporting their designation as Cambrian and part of the Albee Formation. Therefore, the mapped region of the Albee Formation in the Lancaster region has expanded considerably. Population density plots of detrital zircon ages for both samples show excellent similarities to other Cambrian Ganderian units (e.g. Dead River, Ellsworth, and Moretown Formations).

By contrast, the roof pendants on Terrace Mtn. yielded youngest zircon ages of 406 ± 11 Ma and 415 ± 17 Ma suggesting a Devonian age. Population density plots of zircon detrital ages and very similar to that of the Tarratine Formation in Western Maine and may correlate to the 3rd Acadian, outboard-derived, detrital zircon cycle of Bradley and O’Sullivan (2016). Therefore, these rocks are certainly not Albee Formation and represent instead Devonian sedimentary rocks that are stopped blocks from a higher structural level not typically seen on the current erosional surface.

Though the contact between the Albee Formation and overlying Ammonoosuc Volcanics is not exposed, we speculate, based on their age difference of at least 50 million years and stark contrast in deformation style, that these rocks are in unconformable contact with each other. This unconformity is likely the Early Ordovician (?) Penobscot unconformity. The Albee Formation is multiply deformed showing classic pin-stripping, transposition, injection of quartz veins, and a multitude of micro-, macro-, and macro-scale folds. Conversely, the Ammonoosuc Volcanics shows primary features in the form of lapilli and interbedded mafic and felsic units and only a single phase of folding at just the macro-scale. This latter deformation defines a series of NE plunging, map-scale, reclined folds of Acadian or Taconian age. The diapiric doming of the Oliverian Jefferson Dome in the Neoacadian in turn deformed these fabrics.

In the Lancaster area, Chapman (1942), Billings et al. (1979) and most recently Lyons et al. (1997) all show the Ammonoosuc Fault juxtaposing the Lost Nation pluton and a sliver of Albee Formation to the north against the Ammonoosuc Volcanics to the south. These maps show the fault extending east through the Pliny Caldera Complex and back into the Ordovician section east of the Pliny Range. We do not support the existence of the Ammonoosuc Fault in this part of the Lancaster area. This is based on the new extensions of the Albee Formation across the proposed fault, the lack of any evidence of a fault (such as silicified zones or zones of crenulation) in both the Ordovician and Jurassic rocks, and the existence of a chilled intrusive contact, without fault disruption, between the complex mafic intrusive rocks of the Lost Nation Pluton and older rocks.

**ROAD LOG**

**Time, Place, Logistics**

Sunday October 1st, 8:00 AM meet at the first STOP parking location which is the junction of Fleming and Mine Roads in West Bethel, Maine (350813.44 m E, 4917688.46 m N). To get to the meeting place from the Maine Mineral and Gem Museum in Bethel drive west on U.S. Rte. 2 about 5 miles and turn left (S) on to Fleming Road proceed for .2 miles and park at the junction with Mine Road. Bring your lunch and water. We will have a quick pit stop at the Subway sandwich shop in Gorham, NH. Ticks are really bad in the region so get those socks up over your pant cuffs!

**Mileage**

0.0 Junction of Fleming and Mine Roads in West Bethel.

**STOP 1** Hike up the well maintained mine access road that approaches from the north and leads in about .5 miles to Wheeler Mine. Along the road up we pass swirly migmatized outcrops, probably a good candidate for diatexite, of the gray schists and quartzites of the Bog Brook unit and many two-mica granite intrusions and cross cutting pegmatites. Two of these granites were dated to be 349.2 ± 2.1 Ma and 355.3 ± 2.3 Ma. Up at the main entrance to the mine there is abundant outcrop of the Bog Brook Granofels, which is where our detrital zircon age sample was taken that yielded a maximum age of 419.5 ± 3.9 Ma. For safety reasons, we've been asked to NOT enter the mine so only look from a distance. Inside the mine are excellent 3-D exposures of granofels, cross cutting two mica granites, and even younger cross cutting pegmatite. The fine-grained granofels and fine-grained varieties of the two mica granite can be difficult to tell apart! The cross cutting pegmatites have some very long (circa 1 m) biotite crystals in them.
Return to the vehicles and proceed back to U.S. Rte. 2 heading west. Follow Rte. U.S. 2 17 miles to Gorham, NH.

17 Brief pit stop at the Subway Sandwich shop (66 Maine Street).

Continue on U.S. Rte. 2 west for 1.3 miles.

18.3 Turn left at the stoplight to continue on Rte. U.S. 2 while Rte. NH 16 continues straight to Berlin.

Proceed 4.5 miles to Durand Rd.

22.8 At bottom of big hill, turn right onto Durand Rd and proceed west for 1.2 miles to Coldbrook Rd. (private).

24 Turn left on to Coldbrook Rd., immediately cross the bridge over the Moose River, and park near Dwight and Lauren Bradley's farm. Say "Hi" to their one-eyed corgi Moxie.

Begin hiking mileage for stop 2

0.0 Go up the dirt road toward U.S. Rte. 2.
0.1 Pass by the Billings cabin where Marland Billings and Katherine Fowler-Billings stayed while they mapped the Presidential in the ’30’s and 40’s. Continue across Rte. 2, crossing the old railroad bed, now a recreation path, to the power lines.
0.2 At power lines, bear right and follow trail into the woods in about 100 yards.
0.3 Woods road merges with Randolph Mountain Club Pine Link Trail.
0.35 Junction with the Beechwood Trail, bear right on the Pine Link Trail.
0.5 Memorial Bridge. This bridge was built in 1924 as a “Memorial to J.R. Edmands and E.B. Cook and those other pioneer pathmakers” in Randolph. The monument at the east end of the bridge is made of mylonitized Oama, the Ammonoosuc Volcanics Amphibolite member. This is a good rock in which to see the foliation and lineation so that you can determine the shear sense plane.

STOP 2a. Jefferson Dome. Just under and downstream of the bridge is an outcrop of Obqm, the foliated pinkish biotite quartz monzonite of the Oliverian Plutonic Series, part of the Jefferson Gneiss Dome of the Bronson Hill.

The foliation formed during either the doming stage in the Neoacadian or during faulting along the Moose River Fault. The lineation is harder to see and trends moderately southeast. Also seen is a green, probably pre-Acadian, metamorphosed basalt dike and some pegmatite of the 363 Ma. Bickford granite. Maura Foley (Foley and Eusden, 2009) examined the shear sense in thin sections from this outcrop and found S-C fabrics with tops to the northwest (reverse slip). Shear sense for all of the mylonites in the Jefferson Dome rocks shows reverse motion. She also collected some monazite microprobe ages here with the following age populations: early to middle Devonian ages of 392 +/- 3 and 380 +/- 3 Ma (peak Acadian metamorphism); middle to late Devonian ages of 367 +/- 3.2 Ma (doming?); and Carboniferous ages of 346 +/- 3.6 Ma (intrusion of local granites).

From the bridge continue up the west side of Cold Brook on the Sylvan Way trail to Cold Brook Falls.

.55 Cold Brook Fall (very slippery rocks here).

STOP 2b. Moose River fault, Ammonoosuc Volcanics, and mylonites. At the falls are exposed Oama, the Ammonoosuc Volcanics Amphibolite member, and a few intercalated layers of Oamg and Obqm. The regional strike and dip (70°, 48 SE) is seen well in outcrop on both sides of the falls. The lineation (125°, 38°) is defined by fine-grained biotite and can be seen in good light. This dip is a result of doming by the Obqm when it diapirically rose up through the more dense cover rocks in the Neoacadian. Mylonitic fabrics are seen throughout the outcrops and particularly well in the amphibolite and gneiss horizons. Oriented thin sections from this locality show tops to the northwest sense (reverse slip) (Dupee et al., 2002). Based on the lineation orientation, there is some minor strike slip motion that is dextral, but the dominant motion appears to be dip slip. Collectively these mylonites make up the Moose River fault separating the Oliverian Jefferson Dome from the Ammonoosuc Volcanics. Farther upstream
Foley and Eusden (2009) have found normal shear sense indicators in mylonites. With both normal and reverse motions found in the Ammonoosuc Volcanics it is difficult to determine the overall motion of the Moose River fault. Are there two generations of faulting with early reverse followed by later normal motion? More work is needed. Regardless, the evidence shown by the mylonitic fabrics along the Moose River Fault and truncation of Ammonoosuc sub units strongly suggest that the contact between the Ammonoosuc Volcanics and Oliverian Jefferson Dome is a fault and the intrusive contact, if there ever was one, is obscured.

Continue upstream on the east side of Cold Brook along an old abandoned trail.

STOP 2c. Ammonoosuc Volcanics and sill of Bickford granite. Exposed in the river are a few more outcrops of less mylonitized amphibolite and interbedded gneiss. One amphibolite horizon that has an abundance of sub-centimeter size plagioclase grains in a dark matrix suggests it was once a crystal lapilli mafic tuff. Above these rocks is a sill of medium to fine-grained 363 Ma. Bickford Granite. Sills of medium to fine grained two mica and biotite granite and coarse pegmatite continue upstream for a couple of miles. The granite is not deformed by doming or shearing, and appears to mark the end of Neoacadian tectonism sometime at the end of the Devonian or the earliest Carboniferous. The age of the Bickford is coincident with the timing of the Neoacadian to the south (Mass.) but here no deformation or extensive metamorphism of that age is seen here. There is also no evidence of any Alleghenian tectonism in this part of the Appalachians.

Proceed back to the vehicles and drive west on Rte. 2.

Continue Road mileage

Drive uphill on the dirt road past the Bradley's Farm and carefully turn right on to U.S. Rte. 2 and proceed west 15 miles to Jefferson, NH and turn right on the North Rd.

STOP 3 On both sides of Garland Brook are outcrops of what we've interpreted as Cambrian Albee Formation. The two detrital zircon samples came from about .5 and 1 miles to the west and northwest in the woods. The maximum depositional ages of these samples are 522 ± 3.8 Ma and 545 ± 17 Ma. The Albee Formation here has the classic pin stripping with dark schist and quartzite layers and lighter veins of quartz. The outcrop is complexly folded and shows some evidence for transposition. We attribute the intensity of meso-scale folding to this outcrop's position on the hinge of a macro-scale reclined overturned, anticline. That fold has a hinge line plunging north at about 45°, an axial surface striking west and dipping north also at about 45°, and an interlimb angle of about 60°. This region had previously been mapped as Ordovician Ammonoosuc Volcanics (Chapman, 1942; Billings et al., 1979). Our new mapping extends the Albee across what used to be the Ammonoosuc Fault. Based on the lack of fault rocks and structural evidence and the continuous extent of the Albee and Ammonoosucs across the previously mapped fault, we now believe the fault doesn't exist in this region and dies out somewhere to the southwest.

Walk back to vehicles

Retrace the drive .6 miles on Brook Rd. to North Rd.

Turn right on North Rd and proceed west for 2.5 miles to Grange Rd.
45.2 Turn right on Grange Rd. and follow it for 2 miles to Pleasant Valley Rd.

47.2 Continue straight on Pleasant Valley Rd. while Grange Rd. veers off to the left. There is a large outcrop of the Lost Nation pluton on the left side of the road at the intersection.

Proceed on Pleasant Valley Rd. for .2 miles to 23 Paradise Valley Road.

47.4 Park at the bottom of the driveway to 23 Paradise Valley Road.

Walk up the driveway to examine the washed outcrop next to the house.

**STOP 4** Exposed in this awesome washed outcrop in the yard of the house is the Ordovician Lost Nation pluton. This is a fairly obscure pluton in Appalachian lore and linked temporally and magmatically to the Highlandcroft Plutonic Series (Lyons et al., 1997; Dorais et al., 2008). Complicating the story of the Lost Nation is its 442 ± 4 Ma age by Moench and Aleinikoff (2003), which makes it roughly coeval with the Oliverian Jefferson Dome. Dorais et al. (2008) summarizes much of the geochemical and petrologic work done on the Lost Nation and suggest the Highlandcroft and Oliverian Plutonic Suites are magmatically related arc magmas derived from a Laurentian source.

The basic characteristics in the field for the Lost Nation are lack of a well-developed foliation, typically a more mafic lithology ranging from metagabbro to metagranodiorite, and an irregular discordant pluton shape which is west of the neatly domed pattern of the Oliverian Jefferson Dome.

Exposed here is a greenish-black, salt and pepper texture, metadiorite to metagabbro. The massive rock has little foliation in it, shows a few streaks of schlieren of older metasedimentary units (Ammonoosuc?), and is cross cut by numerous thin, white to pink veins of granite that are presumably offshoots from the Jurassic Pliny Complex to the east.

**End of trip. Safe journeys back to your destination!**

**REFERENCES**


THE NEW HAMPSHIRE SPHERULITIC RHYOLITES: ROCKS OF IMPORTANCE TO PREHISTORIC NATIVE AMERICANS

By
Sarah Baker, New Hampshire Geological Survey, 29 Hazen Drive, Concord, NH 03301 (sarah.baker@des.nh.gov)
Richard Boisvert, New Hampshire Division of Historical Resources, Concord, NH 03301
J. Dykstra Eusden, Department of Geology, Bates College, Lewiston, ME, 04240
Nathan Hamilton, Department of Geography and Anthropology, University of Southern Maine, Gorham ME, 04038
Stephen Pollock, 145 Ferry Road, Saco, ME 04072

INTRODUCTION

New Hampshire’s characteristic granite bedrock brings in thousands of visitors each summer as a major attraction for the casual geologist or climber, but over 10,000 years ago the first inhabitants of North America found themselves drawn to New Hampshire for its high-quality volcanic rocks. The fine grained varieties of materials necessary to create a tool with a sharp edge are found only in very isolated pockets of geology in the northeast. Therefore, the handful of locations where these materials are naturally occurring appear to have been well known and frequently visited by the earliest people of New Hampshire and its neighbors.

Outcrops of rhyolite, a volcanic rock of granitic composition, can be found in the outskirts of New Hampshire’s White Mountains. Obsidian, or entirely glassy rhyolite, is ideal for tool making but not available locally. While not all rhyolite is glassy in texture, some varieties are glassy enough to be useful for tool making. There are two known archaeological sources of flow banded, and sometimes spherule bearing rhyolites in New Hampshire, located in the towns of Berlin and Jefferson. A location map showing where these rhyolite sources are within the state of New Hampshire is shown in figure 1, and fluted points made from both rhyolite sources are photographed in figure 2.

Figure 1. Index map showing the location of selected archaeologically significant quarry areas in northeastern North America. Yellow circles show the locations of archaeological sites from which spherulitic rhyolite has been recovered.
Figure 2. Photographs of each of the sources of New Hampshire flow banded spherulitic rhyolites found as artifacts.
MOUNT JASPER

Previous Work

The earliest mention of the Mount Jasper mine appears in an 1869 publication titled “Mineralogy Among the Aborigines of Maine,” where the source is described as “a variety of ribbon jasper, found in Berlin, [that] was extensively employed by the Androscoggin Indians.” The author also noted that “this locality has been rediscovered within a few years, where the chips were found which they had left,” suggesting it was still a place of importance to residents in the 19th century (True, 1869). Those living in the 19th century also seem to have recognized the widespread distribution of Mount Jasper material beyond the immediate area of the dike. H.W. Haynes reported an instance where flakes of spherulitic rhyolite of archaeological significance were found along the Androscoggin River approximately 11 km north of the Mount Jasper (Haynes, 1888).

Nearly a century passed before the Mount Jasper mine was explored any further in terms of its archaeological importance. Interest in the mine, however, resumed in the early to mid 1980s, when there were several workshop sites excavated at the base of Mount Jasper, as well as an excavation of the dike at the summit and the mine itself (Gramly 1980, 1984). In the early 1990s, the lithic source was placed on the National Register of Historic Places (Boisvert, 1992).

Archaeology

The dike here was mined during the late Paleoindian through Archaic period before gradually falling into disuse during the Woodland period. While researchers believe that this source was exploited predominately between 6000 and 7000 years ago, inhabitants were visiting this source for a period spanning over 11,500 years (Boisvert, 2009). Mount Jasper is a small hill that overlooks the confluence of the Androscoggin River and the Dead River, and therefore situated on a major path that links the Androscoggin and Connecticut Rivers. The location of the source between these two major throughways, providing easy access through much of the northeast, is likely one reason the site was exploited for such an extensive period of time (Pollock et al, 2008a). The small but significant hill also likely acted as a useful landmark for navigation purposes.

Interestingly, no tools of Paleoindian tradition have actually been recovered at the Mount Jasper site itself (Boisvert, 2012). Therefore dating of the site relied upon the discovery of several diagnostic projectile points manufactured from Mount Jasper material found at sites across the northeast (Boisvert, 1992). Some projectile points made from other regional materials that were recovered by Gramly were also useful in dating the site (Gramly, 1984) It appears that occupations at Mount Jasper focused primarily on tool manufacture, without many other aspects of life carried out at this location.

During the Archaic and Woodland periods the distribution of artifacts mined from Berlin was largely confined to the Androscoggin drainage. Paleoindian sites in Maine and Massachusetts, and at least one site each in New York and Quebec have recovered artifacts of spherulitic rhyolite (Pollock et al, 2008a). Mount Jasper is a very unique example of mining by Native Americans to acquire stone material for tool making. While at other sites throughout the northeast stones may have been quarried, Mount Jasper is the only instance where there is an actual mine.

Geology

The Mount Jasper lithic source is a 0.75 – 1.3 m wide dike of flow banded rhyolite (Boisvert, 2009). The mine here is by definition an adit, cutting straight into the vein of desirable material and creating the famous “Jasper Cave” as H.W. Haynes described it. Billings and Fowler - Billings (1975) identified four types of dikes that intrude the Ordovician Ammonoosuc Volcanics at Mount Jasper: 1) very coarse-grained to pegmatitic granitic dikes, 2) a biotite granofels dike, 3) basalt dikes and 4) the flow-banded spherulitic dikes of archaeological importance (Pollock et al, 2008 b). Starbuck describes Mount Jasper material as belonging to the Moat Volcanics (2006).
JEFFERSON RHYOLITE

Previous Work

R. W. Chapman made the first geologic map of the Pliny Range, home to the Jefferson Rhyolite, in 1942. Chapman built upon this work in 1946 with Marland Billings, who later mapped the Mount Washington Quadrangle at the scale 1:24,000 in 1956. The Jurassic ring dike complex was also the subject of a master’s thesis by MIT student Eichelberger in 1970.

The Jefferson rhyolite was first recognized as a distinct source in the 1990s. When the base of a Paleoindian fluted point was recovered from the Jefferson I site in 1996, archaeologists believed the point was made from Mount Jasper rhyolite due to visual similarities. The point was in fact from the visually similar, and likely closely related source in Jefferson that had yet to be observed in outcrop (Boisvert 1996; 1999). Researchers in geology and archeology have since collaborated in several attempts to distinguish Mount Jasper rhyolite from the Jefferson rhyolite, both visually and geochemically (Pollock, Hamilton, and Boisvert, 2007, 2008). Previous studies have examined artifacts and flakes made from these rhyolites, as well as blocks of till that were found nearby to archaeology sites and therefore characterized as “source” material for the artifacts. While an archaeological source in the form of large blocks transported by glaciation has been previously identified and studied, this study aims to identify the original geological source outcrop location that the blocks of till would have come from.

Current Work

A handful of small dikes of rhyolite were found in the Pliny Range during fieldwork in 2015, located on the slopes overlooking the Israel River valley. Samples were taken from these outcrops and compared using x-ray fluorescence (XRF), which confirmed that the Jefferson source was distinct from the Mount Jasper source, and that it was indeed geochemically related to artifacts that were visually and macroscopically characterized as Jefferson material by archaeologists and geologists (Baker, 2016).

Although an archaeological source in the form of large blocks transported by glaciation has been identified and studied, the 2016 study aimed to identify the original geological source outcrop location that the blocks of till would have come from. While researchers have no reason to believe the newly discovered rhyolite dikes were exploited by Native Americans directly at the outcrops in Randolph and Jefferson, these are the first discovered outcrops of the distinct variety of spherulitic rhyolite found as artifacts at local archaeology sites. Because these outcrops are so small, just a few meters wide, it is possible and even probable that more rhyolite dikes remain undiscovered in the Pliny Range and nearby areas.

Archaeology

The Israel River Complex consists of six sites (named Jefferson I – VI), all within an area less than one square kilometer. The Israel River is a tributary to the Connecticut River, and the two converge about 15 kilometers to the northwest of the Israel River Complex. It was here that the Jefferson source was first discovered as artifacts, and as blocks of till that littered slopes of the Pliny Range after the last glaciation. The Jefferson Source has also been found at the Potter site in nearby Randolph, New Hampshire (figure 3). The Jefferson source was used for a short time relative to the lengthy exploitation of the Mount Jasper

Figure 3. Relative locations of the Israel River Complex (red dots to west) and Potter site (red dot to the east).
source. The Jefferson source was used exclusively in the Paleoindian period between approximately 12,000 and 9,500 years ago (Pollock et al., 2008).

Both the Mt. Jasper Rhyolite and Jefferson Rhyolite were used by some of the first inhabitants of present day North America. People living during the Paleoindian period used these sources and many others to make characteristic fluted points for hunting big game megafauna, likely caribou. These people likely came to the northeast as the glaciers retreated after the last glacial maximum during the late Wisconsin, following herds along their migratory paths which often coincided with river valleys and bodies of water (Starbuck, 2006). The location of the Israel River Complex was very conveniently placed on the landscape for observing herds of migrating Caribou. It has recently been proposed by Boisvert that the sites actually consist of several specific “vantage point” sites that were used to looking out across the landscape, as well as habitation areas on lower lying parts of the landscape, as shown in figure 4.

Figure 4. Viewshed analysis of one of the habitation sites (above) and one of the vantage points (below) located within the Israel River Complex.
Geology

The major topographic feature in the Jefferson quadrangle is the Pliny Range, whose arcuate shape is the result of Jurassic cone sheet intrusions through the underlying Ordovician Oliverian dome rocks. The more resistant mafic units that intruded along the ring shaped fracture after caldera collapse remained intact, while older rocks eroded away easily over time. The Jefferson rhyolite likely represents a final surge in magmatism fairly close to the surface before volcanic activity ceased completely in this area. This activity could be related to the stocks of Conway granite that outcrop in the Jefferson 7.5° quadrangle. The most recent bedrock geologic mapping effort of the Jefferson 7.5° quadrangle by Hillenbrand is shown in figure 5.
GEOCHEMISTRY OF THE JEFFERSON SOURCE

We must be careful when using the term “source” as Harbottle (1982) explains on the matter of provenance research, as

“... with a very few exceptions, you cannot unequivocally source anything. What you can do is characterize the object, or better, groups of similar objects found in a site or archaeological zone... A careful job of chemical characterisation, plus a little numerical taxonomy and some auxiliary archaeological and/or stylistic information, will often do something almost as useful: it will produce groupings of artifacts that make archaeological sense. This, rather than absolute proof of origin, will often necessarily be the goal.”

This is important to keep in mind in relation to studies of rhyolites, because these volcanic rocks can be highly variable in both appearance and geochemistry, especially in flow banded and/or spherulitic varieties like those from Berlin and Jefferson. Because rhyolites are not as homogenous as some units typically observed in lithic sourcing studies, such as obsidian, rhyolites can not always be easily “fingerprinted” using compositional data alone (Fraser-Shapiro, 2007).

For reasons described above, results of this study were compared to those of previous studies of other volcanic rocks from New England. An earlier study comparing the Mt. Jasper Rhyolite to the Jefferson Rhyolite compared the outcome of non-destructive analysis to destructive analysis of artifacts and source materials (Pollock and Hermes, 2000). This study was the only additional study that used XRF to geochemically classify the Jefferson source for archaeological purposes. The Jefferson rhyolite source and artifact material was “fingerprinted” in the 2016 study, and then compared to results from the 2000 and 2013 studies, consisting only of artifact and loose “source” material from the valley. While only four artifacts and four source samples were analyzed in this 2016 study due to limited availability, several samples were processed in previous studies, which were very useful for comparison to new data.

The XRF data for major elemental geochemistry plotted on a TAS diagram suggests that all samples collected in the field and in archaeology are indeed rhyolites. The rhyolite is fairly chemically homogenous, with SiO2 values ranging from approximately 71% to 76%, values nearly identical to those observed in the Mt. Jasper Rhyolite (Boisvert and Pollock, 2009). Comparison of relative amounts of certain immobile trace elements and rare earth elements confirmed that the Jefferson rhyolite was indeed geochemically distinguishable from the Mount Jasper rhyolite. Plots of some of the results of geochemical analysis, compared to those of several other sources, are shown below in figure 6.

Some discrepancies in the new data when compared to previous studies could be due to variable weathering, or human/instrument error, but this is not likely the major cause of large discrepancies. More likely, the Jefferson source, which is highly visually variable, is also highly variable in composition, possibly even derived from multiple magmatic sources. This notion is strengthened by XRF performed by Williams (2013), which concluded that multiple compositions exist within the samples determined to be of Jefferson origin.

While these data do not definitively show that the outcrops of Jefferson rhyolites are the same as those used by Paleoindians to make stone tools, they do suggest that this outcrop is geochemically distinct from other local sources, and very similar to those determined “Jefferson” by previous studies. It is reasonable to conclude that the Jefferson Rhyolite observed in the Jefferson study area is indeed closely related to the blocks of till quarried during the Paleoindian period. This interpretation is supported by the proximity of the rhyolite dikes to the Paleoindian sites.
Figure 6. Comparison of geochemical compositional data for the source from Mount Jasper, the newly discovered source from Jefferson, and artifact material previously determined as derived from the Jefferson source.

ACKNOWLEDGMENTS

I owe countless thanks to the Bates College Geology department, especially my senior thesis advisor Dyk Eusden, and the late John Creasy, with whom I took my first ever geology class at Bates.
ROAD LOG

MEETING POINT. The trip will meet at the Meet Burger King Restaurant along Route 2 and 16 in Gorham, NH (UTM 0325844W 4917498N) at 8:30 AM on Sunday, October 1st.

Mileage. All UTM Coordinates are in NAD 1927 CONUS

0.0  Meet Burger King Restaurant along Route 2 and 16 in Gorham, NH (UTM 0325844W 4917498N). Turn left onto NH Routes 2 & 16. Drive west on NH Route 2 & 16

0.7  Turn right onto NH Route 16 north at the intersection. Proceed north on NH 16 to Berlin. In Berlin, NH Route 16 is labeled Glen Avenue. This is a one way street northbound.

5.00  Turn left onto NH 110.  NOTE: Roads have been recently reconstructed and there are some curves in the road in this area. Some maps show that Green Street turns into First Avenue which then becomes Wright Street.

0.3  Turn right onto Hillside Avenue. Follow Hillside Avenue north to Mount Calvaire Cemetery.

1.9  Turn left at the first entrance into Mount Calvaire Cemetery. This area of the cemetery currently has no monuments. Proceed to the back of the cemetery.

0.1  **STOP 1. (2 HOURS) Park and proceed to the access path.**

From here we will hike to the top of Mount Jasper, a distance of approximately 1.0 kilometer. The hike involves a gradual 35 meter vertical ascent. Portions of the path are primitive and stout foot wear is recommended. (25 minutes)

Path entrance UTM is 19 T 326351 4928056. The access path is currently blocked with large boulders. In June 2017, this is an ATV track. Walk along the path approximately 290 meters to a gate at UTM 19 T 0326067 4928111 (approximate). Proceed past the gate. Continue another meters to a small cairn on the left (southwest) side of the ATV path at UTM 19 T 0325818 4928227. The cairn which was present in June 2017 is small and easily missed. It marks the entrance to a small foot path to the top of Mount Jasper. If you reach a large clear cut area on the right of the ATV path you have gone too far.

Turn south onto this path and follow it to the top of Mount Jasper. The path is marked by yellow blazes in June 2017. At waypoint 19 T 0325824 4928133 there is a wooden foot bridge across a small brook. Another waypoint is 19 T 0325616 4928104. This is the intersection with an orange blazed path (as of June 2017). Please see end note regarding this trail. Continue following the yellow blazes to the top of Mount Jasper.

Retrace route back to vehicles. Retrace route back to Intersection of Route 2 and 16 in Gorham, NH.

8.0  At the Intersection of Routes 2 and 16 turn right onto U.S. Route 2.

12.8  Turn right onto Ingerson Road. Six Gun City will be on your immediate left.

1.9  Intersection of Ingerson Road and Pond Safety Road. Proceed on Pond Safety Road

0.9  GATE

0.8  Logging Road on left

0.6  Bridge. Hunting camp with tar paper siding

0.1  Parking at the intersection of Pond Safety Road and snow mobile path (19 T 309975 4918188).
STOP 2 (2.5 HOURS) JEFFERSON RHYOLITE DIKE

This is an approximate 1.2 kilometer one-way walk from where we are parked to the exposure. This is a woods hike, NOT a path. We will be bushwacking! Stay together and do not leave the group! Outcrop coordinates are 19 T 0310576 4919241.

End of trip! Pending permission and if time allows, the members from the group can join on a hike to a clear cut area if they wish to see panoramic views of Mount Jasper and the Israel River Valley.

END NOTE: The top of Mount Jasper can also be reached via two trails that are marked with orange and blue blazes. These trails are part of those maintained by the town of Berlin. A trailhead is located to the rear of Berlin High School. Park in the lot closest to the track field. There is a descriptive board showing the trail set. A trail marked by blue blazes is immediately in back of the board. Additionally there is a nearby ATV trail that links to the ATV trail behind Mount Calvaire Cemetery. There are several other unmarked trail sets in the area.

The trail behind the high school to the top of Mount Jasper is approximately 1.2 kilometers long with an elevation gain of approximately 470 feet. This trail is steep in parts and should be attempted by persons in reasonably good physical condition. Stout boots are recommended. The blue trail splits at 19 T 0325876 4927997. If ascending the unmarked trail will be on your left. This unmarked spur is a short cut and rejoins the trail with orange blazes at 19 T 0325842 4928057. If you remain on the trail with blue blazes you will reach an intersection with a trail marked by orange blazes at 19 T 325901 492802. Blue blazes cease at this point. The orange blazed trail to your left will take you to the top of Mount Jasper. The trail with orange blazes joins the trail from Mount Calvaire Cemetery at 19 T 0325616 4928104. If you reach this point turn left to complete the climb to the top of Mount Jasper.

REFERENCES CITED


Creasy, J. W., and Eby, G. N., 1993, Ring dikes and plutons: A deeper view of calderas as illustrated by the White Mountain igneous province, New Hampshire: Field Trip Guidebook for the Northeastern United States:
Amherst, Massachusetts, Department of Geology and Geography, University of Massachusetts, Contribution, no. 67, p. N1-N25.


THE SANDY RIVER RE-VISITED

By

Julia Daly, Dept. of Geology, University of Maine at Farmington, Farmington, ME 04938
Thomas Eastler, Dept. of Geology, University of Maine at Farmington, Farmington, ME 04938
Daniel Locke, Maine Geological Survey, Augusta, ME 04330
Email: dalyj@maine.edu

INTRODUCTION

Investigating recent fluvial changes is exciting to geologists because of the opportunity to observe and better understand geologic processes on a human timescale. These changes also have significant consequences for the humans living and working near rivers and responding to shifting channel position, rates of erosion, and seasonal high-water events. Erosion resulting from meander migration poses a particular threat to farms, buildings, and roads adjacent to rivers. Extending from the Saddleback Mountain (Rangeley area) to the Kennebec River in Mercer (central Maine), the Sandy River watershed measures just over 500 square miles. In 2006, we led an NEIGC field trip visiting several large point bars on the river to look at changes in depositional patterns and channel migration recorded by detailed annual surveys from 2002-2006 (Daly and Eastler, 2006). In the decade since our 2006 trip, erosion at three locations prompted mitigation efforts at three of our trip stops. Each location features a different strategy for addressing erosion, prompting us to re-visit these sites and learn more about the benefits and costs of different treatments.

Geologic Setting

This trip visits reaches in the middle of the Sandy River watershed, seeing point bars that reflect the transition from shallow, faster water in the upper part of the watershed to...
slower, deeper channels in the lower section. Figure 2 shows the generalized geologic context of the four field stops. At the upstream field site (stop 1), the channel sits in a relatively narrow valley bounded by till covered hills (Syverson and Greve, 2003). Cobble, gravel, and sand derived from till and valley-parallel eskers yield abundant sediment. The upper limit of post-glacial marine inundation is likely somewhere between Strong and Phillips, ME. Small nearshore marine deposits (mostly sands and muds) are mapped along the margins of the floodplain from Farmington to Strong (Neil, 2007, and Weddle, 2003) and also contribute some sediment when they intersect an active channel. Downstream of Farmington, the gradient decreases and the floodplain broadens. In this section, the channel incises into the post-glacial marine mud and sand of the Presumpscot Formation overlain by post-glacial stream terraces and floodplain deposits (Weddle, 2003).

Local channel morphology & dynamics

Figure 3. Examples of typical views across point bars on the Sandy River, looking downstream. A) View at Stop 1, typical of the mid-section reaches of the watershed with abundant rounded cobbles on the point bar surface and a relatively shallow channel. B) View at Stop 4, typical of a sandy point bar in the lower reaches of the river adjacent to a wider, deeper channel.

The point bars visited during the trip represent a range of morphologies and grain size that vary predictably moving from upstream to downstream reaches. Farther upstream, the channel is slightly narrower (<50m), shallower (<1m), and water velocities are faster. The surface of many upstream bars is an armor of rounded cobbles and gravel with some interstitial coarse sand. The cobbles are imbricated, and their diverse lithologies (granite, schist, slate, phyllite, chert) reflects their glacial material source. Moving downstream, more and more sand is present on the surface of the bar indicating lower velocities and gentler gradients even as the channel is wider (>50m) and deeper (>1-2m) to accommodate more discharge. The meanders are migrating predictably at each location, slowly increasing curvature and moving downstream simultaneously. When they reach their maximum length, the channel avulses and develops a shorter path. Evidence of old channel positions is found at numerous places along the Sandy River as oxbow ponds or simply abandoned channels beginning to re-vegetate. Our 2006 field guide used high-resolution topographic survey data collected over a period of five years to characterize volumetric change on the bars each year, and concluded that the bars were roughly in equilibrium (Daly and Eastler, 2006).

HISTORICAL BACKGROUND

Farmland and private property loss

Land loss as a result of cutbank erosion is a persistent and challenging issue along the Sandy River, bounded closely by roads and homes along its upstream reaches where the valley is relatively narrow and by agricultural fields downstream where the valley broadens significantly. Two of the three sites we will visit have decades of interventions to mitigate the impacts of erosion: Voter Bar (Avon) and Meader Bar (Farmington Falls).
The recent (twenty year) history of Voter Bar has been covered in a series of “Geologic Site of the Month” descriptions by Dan Locke (Locke 2001, 2006, and 2013). Beginning in 1998, property owners on either side of the channel have applied for a variety of permits to alleviate erosional pressure on cutbanks. As seen in Figure 4a, as the middle meander developed in the late 1990’s it began to threaten the home located on Rt. 4 (indicated by the orange box). At the time, state agencies suggested enlarging an old channel (dotted arrow) to capture flow from the main channel. However, before that work started, the river began to naturally re-occupy the old channel. In this image from 2003, the majority of the discharge is still in the main channel, but a significant volume has started to use the shorter path.

Even fifteen and twenty years ago, erosion along the downstream meander was resulting in loss of agricultural land on the Voter Vale Farm, indicated by the solid arrow. As the avulsion progressed rapidly upstream, this cutbank has retreated even more significantly as seen in Figure 4b. In 2012, the farm owners applied and were granted a permit to remove sand and gravel from both active point bars at this location. The rationale cited in the permit was that lowering the points bars would relieve erosional pressure on the opposing cutbanks by allowing the river to occupy a larger cross-sectional area during high water events (Locke, 2013).
Threats to public infrastructure

Downstream, a similar conflict has arisen between cutbank erosion and human infrastructure. For decades, the Sandy River has been threatening to undercut Rt. 156 and one end of Whittier Road in Farmington Falls. Over time, riprap emplaced along the cutbank has shunted erosion away from the road but has required extensions of the riprap as adjacent areas continue to erode. In the 2006 NEIGC field guide, we described the long-term evolution of the site from the 1950’s to the present. Early cutbank erosion toward Rt. 156 prompted the placement of riprap at the apex of the meander where it was closest to the road. Over time, erosion persisted upstream and downstream of the riprap resulting in removal of a threatened house in the 1980’s and application of several riprap extensions. As the riprap was extended, erosion migrated to either end of the hardened surface resulting in two active cutbanks separated by riprap. The upstream cutbank eroded toward Rt. 156 and a local road; following erosion during Hurricane Irene, the high cutbank came within 35’ of the road, leading to traffic restrictions and some temporary closures (Hanstein, 2013).

The fundamental question underlying this trip is: given these problems associated with fluvial erosion, what are the actions to take to mitigate the damage? How can a geologic approach to these scenarios help inform decision-making?

EROSION MITIGATION

This trip will highlight examples of three mitigation strategies: 1) sand and gravel removal from the point bar, 2) riprap & hard berm emplacement, and 3) rootball revetment emplacement.

Sand and gravel removal

There is a long history of sand and gravel removal from the point bars along the Sandy River. However, between 2000 to 2012 no permits were issued for this activity and the point bars at our study sites (most of which had been skimmed in the past) accumulated sediment at faster rate. When we visited these sites in 2006, it had been less than a decade since this process ended. After over a decade without removal, permits were again issued starting in 2012 and sand and gravel have been removed from Voter Bar (Avon, stop 1) and Meader Bar (Farmington Falls, stop 4). An informational forum hosted by the county Soil & Water district in early 2012 drew over fifty people to learn about a new permitting process for gravel removal. In the intervening years, efforts to restore salmon to the Sandy River watershed were initiated and the new permitting process takes those into consideration (Hanstein, 2012).

Figure 5. Gravel and cobble removal from a point bar in Avon. Students (~1.5 m height) at left for scale. Active channel is to the left of picture, view is downstream. Former height of point bar indicated by dashed line.

Riprap / hardened surfaces

Hardened structures, including riprap, are located sporadically along the Sandy River. Most riprap is installed along cutbanks or other unstable banks and has been effective in maintaining channel position at these locations. In mature riprap installations of rounded boulders, vegetation has taken root between some of the blocks or boulders and partially obscures the rocks. While the bank beneath the riprap has been stabilized, erosion continues at the margins of the riprap. The placement of riprap is controlled by Maine’s Shoreland Zoning Act and is overseen by a permitting process.
Rootball revetment structures

“Soft” cutbank stabilization strategies include construction of a rootball (or rootwad) revetment. In this process, the slope is excavated and re-graded to be less steep, then tree trunks are buried in various orientations to slow water and improve slope stability. At the base, large trunks are anchored with their rootballs pointing upstream to help disperse energy during highwater events. These trunks are locked in place with buried boulders, and other tree trunks are partially buried in more vertical positions higher on the revetment. Vegetation is encourage to grow on the surface, further promoting stability.

Table 1. Summary of erosion mitigation strategies

<table>
<thead>
<tr>
<th>STRATEGY</th>
<th>Pros</th>
<th>Cons</th>
</tr>
</thead>
<tbody>
<tr>
<td>Sand gravel removal</td>
<td>• Inexpensive/yields some material with market value</td>
<td>• Needs to be repeated on an annual basis</td>
</tr>
<tr>
<td></td>
<td>• Removal increases cross-sectional area of channel</td>
<td>• Does not fix position of opposing cutbank</td>
</tr>
<tr>
<td>Riprap</td>
<td>• Maintains channel position</td>
<td>• Cost</td>
</tr>
<tr>
<td></td>
<td>• Long-term solution, may be low maintenance if properly designed and installed</td>
<td>• Diverts erosion to adjacent areas</td>
</tr>
<tr>
<td>Rootball revetment</td>
<td>• Diminishes erosion, doesn’t shunt fast water to other areas</td>
<td>• Cost</td>
</tr>
<tr>
<td></td>
<td>• Long-term solution if properly designed and installed</td>
<td>• Requires significant excavation/re-building, especially if bank is steep and tall</td>
</tr>
</tbody>
</table>

ACKNOWLEDGEMENTS

We gratefully thank the property owners for access: Darren and Angel Allen, Jim Meader, Kevin and Judy Vining, and Busse York. Significant early research on these bars was completed by Romany Shanti.
ROAD LOG

STOP 1. VOTER BAR, AVON, ME (19T 398033.40 m E 4961587.34 m N). This trip will begin at Voter Bar; please turn into the hayfield and park in the indicated area. We will be walking on a worn track across the hayfield to access the bar.

Figure 7. Time series of Voter Bar, highlighting migration of the main channel since 2003 (Google Earth, 2017). After the middle point bar was abandoned, the upstream and downstream bars continued to accumulate sediment and migrate slowly downstream. Sand and gravel removal in 2015 from the upstream bar altered flow in the main channel.

Continued land loss on the north side of the river, especially due to migration of the downstream bar into the agricultural field, prompted property owners to approach Maine DEP for permits to mitigate erosion in the early 2000’s. Initially approved to install rootball revetment on the downstream cutbank to protect their field, the owners couldn’t afford the quarter-million dollar cost of construction (Hanstein, 2012). In 2012, the owners submitted an application to remove sand and gravel from both the upstream and downstream bar; the state approved this project as a two-phase plan (phase I: upstream bar, phase II: downstream bar) (Locke, 2013).

Figure 8. The most recent Google Earth image of Voter Bar, acquired in 2016 (Google Earth, 2017). The main channel now follows only the dashed arrow. Prior to sand and gravel removal, the main channel was adjacent to the cutbank on the north side. A prominent chute, used during highwater flows, is shown by the dotted arrow. This path would have been the presumed new channel had the river migrated naturally.

Following sand and gravel removal, the upstream part of the channel re-located to the excavated area, abandoning the upstream cut bank and effectively slowing erosion along that surface. It is likely that the old channel will still be occupied during high water events, leading to some continued erosion in that area. Since 2003, the
channel upstream of the third (downstream) bar has shortened significantly, potentially increasing the rate of erosion at this location.

**Mileage**

0.0 From the field, turn left onto Rt. 4 South and continue for 10.8 miles.

10.8 Turn right onto Town Farm Rd.

11.8 Turn left into E.L. Vining & Son Construction. Continue 0.3 miles down a dirt road to a large clearing at the south end of the property and park.

**STOP 2. Vining Bar, Farmington, ME (19T 407559.90 m E 4949215.74 m N)**

A well-established flooded borrow pit was breached during a highwater event in March, 2009. A small ice dam formed on a downstream point bar, backing up water locally. The narrow berm left at the north end of the point bar was breached, causing sediment to wash into the ponds and diverting flow from the main channel (Figure 10). The property owners value the wildlife habitat provided by the ponds, and constructed a large new berm, see inset.

Figure 9. The most recent Google Earth image of Vining Bar, acquired in spring 2016. The new berm structure, closing the breach created during highwater in 2009, is indicated. A detail of this structure is shown in the bottom left (Google Earth, 2017). River flow at this location is from north to south.

Figure 10. Pre- and post-breach images of Vining bar. Note lower water levels and additional sediment in 2011 (Google Earth, 2017).
berm with a riprap surface to fill the gap eroded during the breach. Since construction, a modest amount of sand has accumulated in front of the berm, providing some buffer from erosion during high flows.

14.6 After returning to Town Farm Rd and turning left to continue south, turn left at the 4-way stop onto Bridge St.

14.8 At the stoplight turn left onto Rt. 4, staying in the right-hand lane to cross the bridge over the Sandy River.

14.9 Follow signs and turn right onto Rt. 2 East / Rt. 27 South. Continue on Rt. 2.

** If you need to purchase gas/fast food/bathrooms, these are most readily accessible by continuing STRAIGHT through this light and using services at one of the establishments within the next quarter mile. **

17.8 Turn right into the Corn Maze, continue for 0.85 miles to a small clearing near the river and park.

18.8 STOP 3. LINDBERGH BAR, FARMINGTON, ME (19 T 412599.03 m E 4942557.17 m N)

Figure 11a. The most recent Google Earth image of Lindbergh Bar (indicated by red arrow), acquired in spring 2016. This relatively small point bar is associated with a cutbank that is actively eroding agricultural land on the opposite bank. The property owner is planning to submit an application for sand and gravel removal. Flow is from northwest to southeast (upper left to lower right).

The Lindbergh bar (Figure 11a) is an example of a point bar that has not recently been altered for erosion control. In contrast to the larger grain size seen on the surface of some of the upstream locations, this bar is located along a shallower gradient of the river with slower velocities, resulting in finer sediments on the

Figure 11b. View across the river at the Lindbergh bar. Note sand accumulated next to tree trunk in the foreground, and the thick section of floodplain sands and silts exposed in the cutbank. The river downstream of Farmington is incising the cohesive, clay-rich Presumpscot Fm.; this is exposed during low flows.
surface of the bar. The opposing cutbank (Figure 11b) shows a thick section of floodplain sands and muds overlying the clay-rich Presumpscot Formation that hosts the channel in the downstream reaches of the river.

18.8 Re-trace route to Rt. 2.
19.7 Turn right onto Rt. 141 and continue briefly through Farmington Falls and over the bridge.
20.2 Turn right onto Rt. 156 and continue.
20.6 Bear right to remain on Rt. 156 (also named Lucy Knowles Rd here), and cross a narrow bridge at 19.7 miles.
21.0 Turn right onto Whittier Road, continuing for 0.1 miles to a cleared area on the right side of the road. Please park as far to the right as possible, beware of poison ivy.

21.1 STOP 4. MEADER BAR, FARMINGTON FALLS, ME (19 T 413495.44 m E 4941010.66 m N)

As a result of rapid erosion during Hurricane Irene (August, 2011), the Sandy River threatened to undercut Whittier Road. A two-year permitting and construction process followed, resulting in the implementation of two mitigation strategies at this location (Hanstein, 2012, 2013, 2016). The first step was to remove sand from the opposite point bars for the first time in a decade (Figure 13). The town was permitted to remove 12,000 cubic yards of sand in 2012 (Hanstein, 2012b), leaving a low berm around the margin of point bar. Beginning in 2013, a large rootball revetment was constructed at a cost of over $450,000 (Hanstein, 2013). The final steps in stabilizing this bank were completed in subsequent years as vegetation grew over the surface of the revetment. Two rounds of plantings failed, but a recent invasive, Japanese knotweed, colonized the bank and provided the necessary cover. The rootball revetment has successfully survived three years with the major logs in place.
REFERENCES CITED


Thompson, Woodrow B., and Borns, Harold W., Jr. (editors), 1985, Surficial geologic map of Maine: Maine Geological Survey, 42' x 52' color map, scale 1:500,000.

MIGMATITES IN PINKHAM NOTCH, NEW HAMPSHIRE

Tim Allen, Department of Environmental Studies, Keene State College, Keene, NH 03435-2001

INTRODUCTION

Migmatite is a textural term used to describe very heterogeneous metamorphic rocks consisting of intermingled light-colored material (leucosomes), dark-colored material (melanosomes) and intermediate material (mesosomes). The obvious question is how do these rocks form, and under what conditions? Since Tuttle & Bowen’s (1958) experiments in the Granite System, the conventional wisdom has been that migmatites form as a result of anatectic partial melting processes (e.g., Winkler, 1979). However, leucosomes from many migmatites—including those in New Hampshire—do not have granitic minimum melt composition, containing quartz and albitic plagioclase but no K-feldspar (although muscovite is usually present; Dougan, 1979; Eusden, 1988; Allen, 1992). It is now generally recognized that other processes besides anatectic partial melting can play a role in forming migmatites, including sub-solidus differentiation, metasomatism, and injection of melt derived elsewhere (Ashworth, 1985; Olsen, 1985).

Dougan (1979, 1981, 1983) conducted detailed mass balance, compositional, and textural studies of leucosomes, melanosomes and mesosomes in migmatites samples from several localities in New Hampshire, including Pinkham Notch. He determined that these migmatites formed by “closed-system” partial melting, possibly driven by the infiltration of fluids. In geologic mapping of the Pinkham Notch area, Hatch & Wall (1986) identified a “migmatite front,” separating un-migmatized schists from migmatitic gneisses both of the Silurian Rangeley Formation (Fig. 1). Such a front certainly suggests infiltration of fluids, and provides an opportunity to directly compare the migmatite with its protolith. In conjunction with geologic mapping in the area (trip A6, this volume; Allen, 1996a, Allen, 1992), I have undertaken detailed stable isotope, petrologic and geochemical studies of these schists and migmatites to determine the migmatization reaction and the role of fluids. Complete analytical details and results are given in Allen (1992).

COMPARISONS ACROSS THE MIGMATITES FRONT

The mineral assemblage in the schists (outcrop #010, Stop 1b) is quartz + sillimanite + muscovite + biotite + quartz + garnet + plagioclase + ilmenite. Accessory minerals include pyrrhotite and tourmaline, and minor graphite, apatite, monazite and zircon. In these schists, sillimanite + muscovite + quartz pseudomorph andalusite, suggesting a polymetamorphic history for this region (Wall & Guidotti, 1986; Hatch & Wall, 1986). There is no potassium feldspar present in these schists, nor is there any evidence for potassium feldspar ever having been a component of the high-grade metamorphic assemblage.

The migmatites (outcrop #011, Stop 1c) contain exactly the same mineral assemblage as the schists, and belong to the same metamorphic zone, except that if the migmatization is due to partial melting then a melt phase was present in addition to the minerals listed above. This melt phase might now be represented by the leucosomes, which are composed of quartz, plagioclase and muscovite, with only minor amounts of the other minerals. Again, there is no evidence of potassium feldspar ever having been present in these rocks, not even in the leucosomes. Tourmaline, apatite, monazite, and zircon are important accessory minerals.

Comparison of Fig. 2a to 2b and Fig. 3a to 3b shows qualitatively a great deal of similarity between the compositions of the bulk migmatites and the schists, which suggests that the schists indeed represent the migmatite’s protolith, and that the migmatization occurred without significant gain or loss of non-volatile components (e.g. silicate melt)—in a “closed system” (isotopic evidence presented below shows that the migmatization was an open system for hydrous fluids). However, the degree of heterogeneity in these rocks, both of the schists as well as the migmatites, makes it difficult to make quantitative comparisons given a limited number of samples. Other evidence for a non-volatile closed system comes from a modal analysis of the migmatite outcrops (Allen, 1994), which showed that the leucosomes and melanosomes occurred in equivalent proportions, each making up about 20 to 30 % of the rock; and from comparison of Figs. 2c & 2d to 2a & 2b and Figs. 3c & 3d to 3a & 3b, which suggests that the leucosome and melanosome compositions are roughly complementary, consistent with textural observations at the outcrops. Dougan’s (1979) mass balance studies comparing leucosome and melanosome compositions to mesosomes (Olsen, 1983; 1985) quantitatively demonstrated “closed system” migmatization in his Pinkham Notch samples.
Figure 1: Geologic map of the Pinkham Notch, NH, study area. Field trip stops indicated by circled numbers (6–9 not shown on this map). Boxes show outcrop sample numbers, metamorphic temperatures from the garnet-biotite thermometer, and whole-rock oxygen isotope values for selected samples. Numerous samples were analyzed from outcrops #010 and #011 — the oxygen isotope value reported is the average; also reported is the range of oxygen isotope values from the outcrop. The map also shows the staurolite-sillimanite isograd of Hatch & Wall (1986) and the migmatite front. Metamorphic pressures across the region are about 3.5–4.0 kilobars.
Figure 2: Analyses by XRF of schists (a), migmatites (b), leucosomes (c), melanosomes (d), and granites (e). The compositions of all samples have been normalized by the average of the unmigmatized schist compositions (a). The oxides are ordered by relative concentration in this average schist, from highest to lowest.

Figure 3: Rare Earth Element (REE) patterns determined by INAA for schists (a), migmatites (b), leucosomes (c), melanosomes (d), and granites (e). The compositions of all samples have been normalized by the North American Shale Composite of Gromet et al., (1984).
METAMORPHIC CONDITIONS

Figure 1 shows metamorphic temperatures calculated using the garnet-biotite Fe–Mg exchange geothermometer of Ferry & Spear (1978), as calibrated by Indares & Martignole (1985, model B). Metamorphic pressures were calculated at between 3.5 and 4.0 kilobars across the region, using the garnet-plagioclase-Al2SiO5-quartz calibration of Newton & Haselton (1981). (Full details of the temperature and pressure calculations are given in Allen, 1992.) The uncertainties of these calculations is on the order of ±50°C and ± 2 Kbar, but these uncertainties do not strongly affect the precision of the method (Hodges & McKenna, 1987)—thus relatively small differences between samples may be real. These results indicate a fairly steep temperature gradient across the migmatite front, consistent with the isograds mapped by Wall & Guidotti (1986) and Hatch & Wall (1986) (Fig. 1), and similar to the metamorphic “hot spots” studied by Chamberlain & Lyons (1983) and Chamberlain & Rumble (1988).

It is not clear, however, that the temperatures obtained in the migmatites are sufficiently high to cause partial melting. Water-saturated granitic melting requires temperatures of about 650°C at 4 Kbar (Johannes, 1984, 1985; Tuttle & Bowen, 1958; Luth et al.,1964), and water-saturated “trondhjemitoid” melting in the sodic-plagioclase + quartz system does not begin until almost 700°C at 4 Kbar (Johannes, 1978, 1985). However, in addition to water, the presence of P, F or B, particularly in combination with water, has been shown to significantly reduce the solidus temperature in these systems (Wyllie & Tuttle, 1964; Manning & Pichavant, 1983; Manning, 1981; Pichavant, 1981), even to temperatures below 550°C. The addition of these components also moves the minimum melt from “granite” towards “trondhjemite” compositions (Manning, 1981). Some evidence suggests that P and F may have been important in these rocks (Fig. 2 and Fig. 5), such that melting may have occurred at the recorded temperatures. It is also possible that the peak temperatures of these rocks were higher than recorded by the garnet-biotite thermometer.

ROLE OF FLUIDS

Figure 1 also shows the distribution of whole rock oxygen isotope values (δ18OWR) in the region near the migmatite front. In addition, numerous samples were analyzed from the schist (outcrop #010, Stop 1b) and migmatite (outcrop #011, Stop 1c) outcrops immediately adjacent to the migmatite front, and the average and range of these data are reported on Fig. 1. No clear overall trends are apparent in the regional data, because of the variety of rock types sampled, but there is a measurable lowering of the δ18OWR from the schist to the migmatite outcrops immediately adjacent to the migmatite front.

If the migmatization process were completely closed, then the “bulk” migmatites and the original schists should have identical whole-rock oxygen isotope values, regardless of the mineral changes that may have occurred. Without introducing lighter oxygen by injecting melt or removing heavier oxygen by extracting melt, the lowering of δ18OWR observed in the migmatites could be achieved by de-volatilization of fluids with heavier oxygen or by isotopic exchange with an infiltrating fluid of lower δ18O. Among the migmatites, the lowest δ18OWR values are from samples spatially associated with pegmatites, while the highest δ18OWR values are from the “least migmatized” samples, suggesting a relationship between the migmatization process and infiltrating fluids.

Whole-rock isotope measurements tell only part of the story. The quartz δ18O values from three samples of the schist outcrop (outcrop #010, Stop 1b) are all nearly identical, as are the biotite δ18O values (Fig. 4). These three samples have very different δ18OWR, from 13.2‰ to 14.1‰, directly attributable to different modal mineralogies. That the minerals from all three of these samples have identical δ18O values suggests these samples were in equilibrium with a common fluid, at least on the outcrop scale. Using the fractionation factors of Bottinga & Javoy (1973, 1975), this fluid is determined to have δ18Owater of about 13.9‰, and the equilibrium temperature is determined to be about 546°C, in reasonable agreement with that determined by garnet-biotite thermometry (520°C, Fig. 1). The homogeneity of mineral values within schist outcrop #010 is suggestive of the kind of isotopic homogenization that is associated with large scale fluid infiltration (Valley, 1986), however in comparison with mineral values from adjacent migmatite outcrops, there is no large scale homogenization. The small variation in these schist samples is in the direction expected for closed-system behavior (Fig. 4, Gregory & Criss, 1986).
Figure 4: Biotite and quartz oxygen isotope values for schist and migmatite melanosome samples. Isotherms determined from the fractionation factors of Bottinga & Javoy (1973, 1979). Shaded triangles show region of closed-system behavior for schist samples (Gregory & Criss, 1986). Sample #036–1 is considered anomalous since biotites in this rock contain a large number of included minerals (monazite, etc.), which may contaminate its $\delta^{18}O_{\text{Bt}}$ value. The uncertainty associated with these analyses is ±0.2‰ and is shown by the labeled box.

It should be noted that in the migmatites, biotite is present only in trace quantities in the leucosomes. Thus only biotite-quartz pairs from melanosomes are shown in Fig. 4. With the exception of sample #036–1, the biotite-quartz values from the migmatites define an isothermal array, at a temperature of about 500°C. If the quartz and biotite $\delta^{18}$O values of the schist samples are taken as a starting point for the material that formed the migmatites then, as shown in Fig. 4, the values for the migmatites lie outside the area expected for closed system behavior (Gregory & Criss, 1986), further confirming the role of infiltrating fluids in driving migmatization. Both biotite and quartz from melanosomes have $\delta^{18}$O values that are lighter than the original schist, consistent with exchange with an isotopically light fluid. From the migmatite samples with the lowest $\delta^{18}$O values, furthest down and to the left in the array (those presumably most exchanged with the fluid), I estimate $\delta^{18}$O$_{\text{water}}$ for the fluid to be 12‰ or less, using the fractionation factors of Bottinga & Javoy (1973, 1975), at the temperature determined by the array (500°C). This value is consistent with a $\delta^{18}$O$_{\text{water}}$ value estimated to be in equilibrium with a pegmatite (#011–2) cutting the outcrop.

The temperature of this isothermal array is not consistent with the temperature obtained by garnet-biotite thermometry in the migmatites, and in fact is lower than the temperature obtained in the schists by either garnet-biotite or oxygen isotope thermometry. It is not clear why this is, but one possibility may be that continued isotopic exchange of minerals in migmatites occurred during cooling after migmatization. Due to the fluid infiltration-driven migmatization, these rocks may have had a higher water content than the schists, perhaps facilitating such exchange during cooling.

Figure 5: Fluorine content determined by microprobe analyses of biotites versus biotite oxygen isotope value for schist and melanosome samples. Sample #036–1 is considered anomalous since biotites in this rock contain a large number of included minerals (monazite, etc.), which may contaminate its $\delta^{18}O_{\text{Bt}}$ value.

In Figure 5 we can clearly see that migmatite biotites are lighter than those in the schists (with the exception of #036–1), and that the migmatite biotites are somewhat enriched in F relative to the schists. Measurements by Hamza & Epstein (1980) have suggested that there is a large oxygen isotope fractionation between oxygen in the hydroxyl site and oxygen in the rest of the biotite, and that as F substitution displaces OH in biotite, the overall $\delta^{18}$O$_{\text{Bt}}$ should...
become higher. The trend observed in our migmatite samples is in the opposite direction, which clearly indicates that the $\delta^{18}$O_Bt is not a function of F–OH composition, but may be influenced by infiltrating fluids. As the biotites with the lightest $\delta^{18}$Ob values are the most enriched in F, it is possible that the infiltrating fluid is also the source of F. Hydrogen isotope measurements from these biotites show no trends, as all samples have $\delta^D$Bt values of about -75‰ (Allen, 1992), within the “normal” range for biotites from metamorphic rocks (Taylor & Sheppard, 1986).

Figure 6 shows $\delta^{18}$O_Ms versus $\delta^{18}$O_Qt for schist, leucosome and melanosome samples. Muscovite is abundant in both melanosomes and leucosomes, so muscovite-quartz points from both leucosomes and melanosomes. Leucosome and melanosome pairs are connected with a line segment, resulting in the black and white “dumbbells.” In contrast to the biotite-quartz system (Fig. 4), the muscovite-quartz points produce a disequilibrium array. This has also been observed in muscovite-quartz pairs from migmatites studied by Mazurek (1992). Certainly some of the muscovite in these samples is retrograde, as indicated by the mesoscopic muscovite spangle texture prevalent in these rocks. It is not clear what factors are responsible for the wide distribution of $\delta^{18}$OMs values, although there may be a trend of increasing $\delta^{18}$OMs away from the Wildcat Granite (trip A6, this volume; Allen, 1992, 1996a) on a regional scale, and away from pegmatites on an outcrop scale (Allen, 1992).

Figure 6: Muscovite and quartz oxygen isotope values for schist and migmatite leucosome and melanosome samples. Isotherms determined from the fractionation factors of Bottinga & Javoy (1973, 1979). Lines connect leucosome (open circle) and melanosome (filled circle) pairs from the same migmatite sample. The uncertainty associated with these analyses is $\pm 0.2‰$ and is shown by the labeled box.

MIGMATIZATION REACTIONS

James B. Thompson, Jr., (1982) developed methods for derivation of a sufficient set of linearly independent net transfer and exchange reactions that describe all possible reactions in a given mineral assemblage. Ferry (1984) adapted this method to calculate the overall reaction at an isograd. In this study I have applied these methods to determine the reactions that occurred across the migmatite front, including possible melting reactions. The additive phase components, exchange phase components, and system components for the assemblage studied here (quartz + biotite + muscovite + garnet + plagioclase + sillimanite + ilmenite) are listed in Table 1. Only net transfer reactions are important in controlling the texture and mineral modes in metamorphic rocks. Exchange reactions only change the compositions of minerals, for example the Fe–Mg ratio in garnet and biotite, and do not affect the modal abundance of the minerals. The first set of net transfer reactions (1–4, Table 1) are determined for a completely closed system with the above mineral assemblage (where potassium and hydroxyl are coupled). Opening the system to Na or K exchange with a hydrous fluid introduces an additional net transfer reaction (5, Table 1). The enrichment of fluorine in migmatite biotites (Fig. 5) is accommodated by the introduction of HF as a system component and a phase component, and F(OH)$_{-1}$ as an exchange component, which yields another independent net transfer reaction (6, Table 1).
CLOSED SYSTEM ASSEMBLAGE

<table>
<thead>
<tr>
<th>Mineral</th>
<th>Additive Component</th>
<th>Exchange Components</th>
</tr>
</thead>
<tbody>
<tr>
<td>Quartz (Qtz)</td>
<td>SiO$_2$</td>
<td>—</td>
</tr>
<tr>
<td>Sillimanite (Sil)</td>
<td>Al$_2$SiO$_5$</td>
<td>—</td>
</tr>
<tr>
<td>Muscovite (Ms)</td>
<td>KAl$_3$Si$<em>3$O$</em>{10}$O$_2$(OH)$_2$ NaK$_1$ (nk), Al$_2$Fe$_2$Si$_1$ (tk), FeTiAl$_2$ (tx), MgFe$_1$, FeTiAl$_2$ (tx), MgFe$_1$, MnFe$_1$</td>
<td></td>
</tr>
<tr>
<td>Biotite (Bt)</td>
<td>KFe$_3$Al$_5$Si$<em>3$O$</em>{15}$(OH)$_2$ NaK$_1$ (nk), Al$_2$Fe$_2$Si$_1$ (tk), FeTiAl$_2$ (tx), MgFe$_1$, MnFe$_1$</td>
<td></td>
</tr>
<tr>
<td>Garnet (Grt)</td>
<td>Fe$_2$Al$_5$Si$<em>3$O$</em>{12}$ CaFe$_1$ (cf), MgFe$_1$, MnFe$_1$</td>
<td></td>
</tr>
<tr>
<td>Plagioclase (Ab)</td>
<td>NaAlSi$_3$O$_8$ CaAlNa$_1$.Si$_1$ (pl)</td>
<td></td>
</tr>
<tr>
<td>Ilmenite (Ilm)</td>
<td>FeTiO$_3$ MgFe$_1$, MnFe$_1$</td>
<td></td>
</tr>
</tbody>
</table>

**System Components:** SiO$_2$, TiO$_2$, Al$_2$O$_3$, CaO, FeO, MgO, MnO, Na$_2$O, K$_2$O, H$_2$O, HF

Linearly Independent Net Transfer Reactions:

1. $\xi$ Ilm + Qtz = Sil + tx (Ilm)
2. $\xi$ Bt + 5 Qtz +3 tk = 2 Sil + Ms (Bt)
3. $\xi$ Ab + Qtz + tk + pl = 2 Sil +cf (Ab)
4. $\xi$ Grt + 4 Qtz + 3 tk = 4 Sil (Grt)

ALLOW NaK$_1$ and F(OH)$_1$ EXCHANGE WITH ENVIRONMENT

Additional Phase Components:

Muscovite

Biotite

Water

**System Components:** SiO$_2$, TiO$_2$, Al$_2$O$_3$, CaO, FeO, MgO, MnO, Na$_2$O, K$_2$O, H$_2$O, HF

Additional Independent Net Transfer Reactions: $\xi$

1. $\xi$ nk + Ms + 2 Qtz + tk + pl = 3 Sil + cf + H$_2$O (nk)
2. $\xi$ HF = F(OH)$_1$ + H$_2$O (F(OH)$_1$)

PARTIAL MELTING

Additional Phase Components:

MELT (sample 011–1L) Si$_{383}$Al$_{116}$Na$_{15}$Ca$_4$(Fe+Mg+Mn)$_{29}$Ti$_2$O$_{1000}$(H$_2$O)$_X$

MELT (sample 011–3L) Si$_{406}$Al$_{99}$Na$_{35}$K$_{14}$Ca$_7$(Fe+Mg+Mn)$_{6}$Ti$_1$O$_{1000}$(H$_2$O)$_X$

MELT (sample 036–2L) Si$_{397}$Al$_{104}$Na$_{51}$K$_{11}$Ca$_9$(Fe+Mg+Mn)$_{8}$Ti$_1$O$_{1000}$(H$_2$O)$_X$

X might be between 20 and 30

Exchange Components:

NaK$_1$ (nk), CaFe$_1$ (cf), F(OH)$_1$, MgFe$_1$, MnFe$_1$, CaAlNa$_1$.Si$_1$ (pl), Al$_2$Fe$_1$.Si$_1$ (tk), FeTiAl$_2$ (tx)

**System Components:** SiO$_2$, TiO$_2$, Al$_2$O$_3$, CaO, FeO, MgO, MnO, Na$_2$O, K$_2$O, H$_2$O, HF

Additional Independent Net Transfer Reactions: $\xi$ measured by melt

1. $\xi$ 011–1L: 228 Qtz + 9 Sil + 9 Bt + 22 Ms + 19 Ab + 4 pl + 2 Ilm + X H$_2$O = MELT (011–1L)
2. $\xi$ 011–3L: 239 Qtz + 4 Sil + 1 Bt + 13 Ms + 42 Ab + 7 pl + 1 Ilm + X H$_2$O = MELT (011–3L) + 2 tk
3. $\xi$ 036–2L: 278 Qtz + 4 Sil + 2 Bt + 9 Ms + 60 Ab + 9 pl + 1 Ilm + X H$_2$O = MELT (036–2L) + 1 tk

X might be between 20 and 30

Table 1: Mineral assemblage, phase and system components, and linearly independently variable reactions.
# Extents of Reaction, if no Melting

<table>
<thead>
<tr>
<th>Reaction</th>
<th>Extent</th>
<th>#011–1</th>
<th>#011–3</th>
<th>#036–2</th>
</tr>
</thead>
<tbody>
<tr>
<td>1 Ilm</td>
<td></td>
<td>0.14</td>
<td>-0.17</td>
<td>0.42</td>
</tr>
<tr>
<td>2 Bt</td>
<td></td>
<td>-0.87</td>
<td>-4.58</td>
<td>-1.39</td>
</tr>
<tr>
<td>3 Ab</td>
<td></td>
<td>4.74</td>
<td>1.61</td>
<td>-2.61</td>
</tr>
<tr>
<td>4 Grt</td>
<td></td>
<td>0.59</td>
<td>0.22</td>
<td>3.77</td>
</tr>
<tr>
<td>5 nk</td>
<td></td>
<td>0.90</td>
<td>1.36</td>
<td>1.46</td>
</tr>
<tr>
<td>6 F(OH)_1</td>
<td></td>
<td>1.03</td>
<td>0.01</td>
<td>0.13</td>
</tr>
</tbody>
</table>

**Net Reactions, assuming no melt was present:**

**011–1H:**
\[
12.94 \text{ Sil} + 0.87 \text{ Bt} + 5.64 \text{ cf} + 0.14 \text{ tx} =
\text{ 4.69 Qtz} + 0.59 \text{ Grt} + 1.77 \text{ Ms} + 4.74 \text{ Ab} + 0.14 \text{ Ilm} + 0.90 \text{ nk} + 5.64 \text{ pl} + 4.80 \text{ tk} + 0.13 \text{ H}_2\text{O}
\]

**011–3H:**
\[
17.86 \text{ Qtz} + 4.58 \text{ Bt} + 0.17 \text{ Ilm} + 2.97 \text{ cf} + 10.11 \text{ tk} + 1.35 \text{ H}_2\text{O} =
\text{ 1.15 Sil} + 0.22 \text{ Grt} + 5.94 \text{ Ms} + 1.61 \text{ Ab} + 1.36 \text{ nk} + 2.97 \text{ pl} + 0.17 \text{ tx}
\]

**036–2H:**
\[
11.88 \text{ Sil} + 1.39 \text{ Bt} + 2.61 \text{ Ab} + 1.15 \text{ pl} + 0.42 \text{ tx} + 1.33 \text{ H}_2\text{O} =
\text{ 8.86 Qtz} + 3.77 \text{ Grt} + 2.85 \text{ Ms} + 0.42 \text{ Ilm} + 1.46 \text{ nk} + 1.15 \text{ cf} + 5.99 \text{ tk}
\]

# Extents of Reaction, with Melting

<table>
<thead>
<tr>
<th>Reaction</th>
<th>Extent</th>
<th>#011–1</th>
<th>#011–3</th>
<th>#036–2</th>
</tr>
</thead>
<tbody>
<tr>
<td>1 Ilm</td>
<td></td>
<td>-0.08</td>
<td>-0.20</td>
<td>0.05</td>
</tr>
<tr>
<td>2 Bt</td>
<td></td>
<td>-3.23</td>
<td>-4.69</td>
<td>3.17</td>
</tr>
<tr>
<td>3 Ab</td>
<td></td>
<td>1.22</td>
<td>-6.41</td>
<td>-9.55</td>
</tr>
<tr>
<td>4 Grt</td>
<td></td>
<td>-0.15</td>
<td>-0.07</td>
<td>0.07</td>
</tr>
<tr>
<td>5 nk</td>
<td></td>
<td>-0.36</td>
<td>0.80</td>
<td>0.71</td>
</tr>
<tr>
<td>6 F(OH)_1</td>
<td></td>
<td>0.70</td>
<td>0.00</td>
<td>0.53</td>
</tr>
<tr>
<td>7 MELT</td>
<td></td>
<td>0.30</td>
<td>0.25</td>
<td>0.15</td>
</tr>
</tbody>
</table>

**Net Reactions, assuming melt present:**

**011–1H:**
\[
84.08 \text{ Qtz} + 0.15 \text{ Grt} + 5.93 \text{ Bt} + 3.41 \text{ Ms} + 4.48 \text{ Ab} + 0.68 \text{ Ilm} + 0.04 \text{ nk} + 1.18 \text{ cf} + 0.02 \text{ pl}
+ 8.96 \text{ tk} + \text{ X H}_2\text{O} =
\text{ 2.11 Sil} + 0.08 \text{ tx} + 0.30 \text{ Melt (011–1L)}
\]

**011–3H:**
\[
88.49 \text{ Qtz} + 0.07 \text{ Grt} + 4.94 \text{ Bt} + 16.91 \text{ Ab} + 0.45 \text{ Ilm} + 7.36 \text{ pl} + 19.39 \text{ tk} + \text{ X H}_2\text{O} =
\text{ 19.28 Sil} + 2.24 \text{ Ms} + 0.80 \text{ nk} + 5.61 \text{ cf} + 0.20 \text{ tx} + 0.25 \text{ Melt (011–3L)}
\]

**036–2H:**
\[
33.65 \text{ Qtz} + 3.81 \text{ Ms} + 18.55 \text{ Ab} + 0.10 \text{ Ilm} + 10.19 \text{ pl} + 0.5 \text{ tx} + \text{ X H}_2\text{O} =
\text{ 9.70 Sil} + 0.07 \text{ Grt} + 2.87 \text{ Bt} + 0.71 \text{ nk} + 8.84 \text{ cf} + 1.03 \text{ tk} + 0.15 \text{ Melt (036–2L)}
\]

X might be 5 or 6

---

**Table 2:** Extents of reactions (ξ) for the reactions specified in Table 1, and resulting overall net-transfer reactions, without and with melting.
If the migmatization occurred without partial melting, then these six reactions completely describe the possible reaction space of the migmatization. If, on the other hand, partial melting is involved in the migmatization process, then we have to introduce yet another phase component to our assemblage, the melt phase. Silicate liquids probably have a wide range of possible compositions, with no enforced stoichiometry that would enable writing carefully balanced reactions. For the partial melting scenario, it is assumed that the leucosomes represent the melt phase (and that no melt was extracted from or injected into the rocks). Estimated melt phase “stoichiometry” was determined from the chemical compositions of the leucosomes (Table 1), on the basis of 1000 anhydrous oxygen atoms per formula unit. With this we can determine a linearly-independent net-transfer “melting” reaction for each of the leucosome samples (7, Table 1).

If we allow the water content of the melt phase to vary independently of the melt phase stoichiometry, then no specific amount of water is needed to balance reactions involving melt, and these reactions may equally represent either water-saturated melting or dehydration melting (A. B. Thompson, 1982; Vielzeuf & Holloway, 1988; LeBreton & Thompson, 1988). However, it is probable that a significant amount of water is involved as a reactant in the melting reactions (7), or at least that the melt phase is capable of dissolving a relatively large amount of excess water.

These reactions (7) may not be the melting reactions, just as none of the other six linearly independent net transfer reactions are necessarily the reactions occurring within the rock. Rather, all possible reactions within this assemblage, including melting reactions, can be described by some linear combination of these seven independent reactions. The overall reaction responsible for the migmatization of these rocks can be determined by measuring the extent of progress for each of the linearly independent net-transfer reactions. Since each of these reactions involves changes in the modal abundance of minerals, in a system closed to gain or loss of non-volatile components, the progress of each reaction ($\xi_r$, Table 1) can be determined by comparing modes of minerals before (schists) and after (migmatites) migmatization.

Due to the heterogeneity in these rocks, particularly within the migmatites, “bulk” mineral modes were calculated from whole rock chemical compositions and the compositions of the constituent mineral phases, following the techniques of Ferry (1984). The details of these calculations are provided in Allen (1992). The average of the schist compositions was used as the best approximation of the protolith material. The calculations were done on an oxygen basis, rather than the volume basis used by Ferry (1984), because the molar volume of the “melt” phase is unknown and the stoichiometry of the melt phase was defined on an oxygen basis (Table 1). An oxygen basis is considered to be approximately equivalent to a volume basis (Chamberlain, 1986). The modes of minerals present in the “bulk” migmatites, and hence the overall migmatization reactions, were calculated both with and without a melt phase. In the melting scenario, the proportion of leucosome determined by a modal analysis of the migmatite outcrops (Allen, 1994) was taken to be the mode of the melt phase. The amount of components Ab, nk and F(OH)$_{-1}$, which measure the $\xi_r$ for reactions (3), (5) and (6), were determined from the calculated modes of the minerals containing these components together with the measured composition of the minerals (Allen, 1992). The $\xi_r$ for the six (or seven, if melt is present) linearly independent net transfer reactions possible in the assemblage (Table 1), and the overall net-transfer reactions determined for the migmatization of these samples (with and without melting) are given in Table 2.

Regardless of whether melting is involved or not, water is generally consumed in the migmatization reactions presented in Table 2. This suggests that infiltration of water from some external source is required to drive the migmatization process. In the case of melting, it is possible that much larger amounts of water may be consumed than if there is no melt. The loss on ignition during XRF analysis, expected to be mostly water, averaged 2.4% for the migmatite whole rock samples, while the loss on ignition from the schist samples averaged only 1.9%. This difference suggests an increase in the water content of the migmatites relative to the schists, consistent with consumption of water by the migmatization reactions.

**LEUCOSOMES AS MELTS?**

Leucosome compositions shown in Figs. 2c and 3c are similar in character to the compositions of the Wildcat Granite (trip A6, this volume; Allen, 1992, 1996a) samples (Figs. 2e and 3e), but—unlike the “G” phase of the Wildcat Granite (sample #105-1)—they do not appear to be of granitic minimum melt composition (Fig. 7). The “G”
phase of the Wildcat Granite is the only rock in the region that actually contains potassium feldspar. The other samples contain quartz, albite plagioclase and muscovite as the major components. Thus these leucosomes may fit Ashworth’s (1976, 1985) definition of “trondhjemitoid”, as distinguished from K–feldspar bearing “granitoid”, leucosomes. Trondhjemite melts are possible in water-saturated, micaceous but K–feldspar absent rocks, particularly with a more sodic plagioclase (Ashworth, 1985; Johannes, 1985, 1983, 1978), as is present in these rocks. Also, the addition of F and increases in P_{H2O} both move the minimum melt in the granite system (Qtz–Ab–Or) towards a more trondhjemitic composition (Qtz–Ab) (Manning, 1981; Luth et al., 1964).

Figure 7: Normative Qtz-Ab-Or mineralogy of migmatite leucosomes (filled circles) and granite samples (open squares). Gray field shows area of commonly observed granite compositions (LeMaitre c.f. Best, 1982, p. 114), which corresponds with the granitic “minimum melt” composition (Tuttle & Bowen, 1958). Arrow shows direction that “minimum melt” composition moves with both increasing water pressure (Luth et al., 1964; Johannes, 1978) and increasing fluorine content (Manning, 1981). Samples #011–1L and #105–2 “R” had total normative Qtz + Ab + Or < 80%, indicating that their compositions are not well represented by this diagram.

Data presented in Figs. 3c & 3d and Fig. 8 suggest that the Rare Earth Elements (REE) and U, Th and Zr are preferentially retained in the melanosomes, to become more concentrated than in the parent schist, and do not enter the leucosomes. The distribution of the REE and U, Th and Zr in high-grade pelitic rocks and in granitic systems is strongly controlled by the behavior of accessory mineral phases such as monazite and zircon (Watson & Harrison, 1984). Melanosome biotites observed in thin section tend to have a high abundance of pleochroic radiation damage halos due to small inclusions of monazite and/or zircon. These halos are much less abundant in biotites from the schists. Studies showing that these minerals may have very low solubility in granitic melts (Rapp & Watson, 1986; Rapp et al., 1987; Harrison & Watson, 1983) further suggest that the leucosomes may indeed be melts. Similar REE and U, Th, and Zr distributions have been observed in migmatites studied by Weber et al. (1985) and by Barbey & others (1989, 1990). Petrochemical modelling by these authors suggests that their observations are best explained by complex melt-fluid interactions (Weber et al., 1985) or disequilibrium partial melting (Barbey & others, 1989, 1990).

Figure 8: Uranium (a), Thorium (b), and Zirconium (c) concentrations in migmatite leucosomes and melanosomes, normalized by the concentration in the average schist (U*, Th*, Zr*; Allen, 1992). The diagonal lines represent the loci of points for which leucosome and melanosome concentrations are equal. The horizontal and vertical lines at unity represent the loci of points where the leucosome or melanosome concentrations are the same as in the average schist.
Figure 6 shows that for several of the migmatite samples studied, melanosome $\delta^{18}O_{Qtz}$ is lower than leucosome $\delta^{18}O_{Qtz}$ by up to 1‰, a significant and measurable difference (Allen, 1992). Similar results were obtained by Mazurek (1992) studying migmatite samples from drill cores of crystalline basement in Switzerland. This difference is better depicted in Fig. 9, constructed in a similar manner to Fig. 8, comparing leucosome to melanosome.

Possible mechanisms for producing such a shift include: (1) re-equilibration of the melanosome mineral assemblage during cooling; (2) selective post-migmatization alteration, in which late fluids are able to infiltrate only melanosome or leucosome portions; (3) oxygen isotope fractionation between melt and solids; and (4) disequilibrium melting, in which the melt incorporates infiltrated fluids, thereby changing its isotopic composition.

Closed-system isotopic re-equilibration of the mineral assemblages in leucosome and melanosome during post-migmatization cooling could have produced differences in the isotopic composition of minerals in the leucosome and melanosome, because of the very different modal mineralogies. However, given the mineral assemblages in these rocks, one would expect melanosome quartz to take on higher $\delta^{18}O_{Qtz}$ values compared to leucosome $\delta^{18}O_{Qtz}$, rather than lower (Allen, 1992).

Continued isotopic exchange with an external fluid after migmatization could have produced the observed differences, if the exchange occurred only in the melanosomes or the leucosomes. While it is possible that some difference in permeability may have existed, the presence of retrograde muscovite in both leucosome and melanosome argues against this. Fig. 6 shows that the $\delta^{18}O$ of muscovites from leucosome and melanosome are essentially identical given the uncertainties.

Incorporation of infiltrating fluids into the melt phase could have produce the differences, if the fluid had an isotopic composition that was relatively heavy compared to the rocks undergoing melting, as Mazurek (1992) argued. In Pinkham Notch, however, the infiltrating fluids driving the migmatization were isotopically light relative to the rocks. Incorporation of this fluid into molten leucosomes should have lowered the $\delta^{18}O$ values of the leucosomes relative to the melanosomes, rather than raised them.

Of the four possibilities for producing the observed shift in leucosome $\delta^{18}O_{Qtz}$, we are left with oxygen isotope fractionation between melt and solids during partial melting and crystallization. Taylor & Sheppard (1986) have considered the possibility of such fractionation and concluded that any fractionation between rocks and melts of
identical composition is negligible. Garlick (1966) has also demonstrated that oxygen isotope fractionation is independent of physical state—molten or solid. Mazurek (1992) ruled out fractionation between minerals and melt as an explanation of the observed difference in $\delta^{18}O_{qtz}$ from leucosome and melanosome on the basis of these studies. However, these studies neglect the fact that in partial melting of pelitic rocks, the initial melt produced can be of very different composition than the residual solids.

Using the techniques of Chamberlain et al. (1990), which provide a means of calculating the effect of net transfer reactions on the isotopic composition of minerals, I have modelled the isotopic evolution of quartz in leucosomes and melanosomes in these rocks using the migmatization reactions presented in Table 2 (Allen, 1992). I believe that the observed shift in the $\delta^{18}O_{qtz}$ of migmatite leucosome relative to melanosome may be best explained by oxygen isotope fractionation during partial melting. Such fractionation has here-to-fore not been recognized. The fact that not all migmatite leucosome-melanosome quartz pairs have a difference in $\delta^{18}O$ does not invalidate the model, but suggests that either more than one migmatization process may be operating, or that the fractionation due to melting may vary depending on details of the melt composition.

**DISCUSSION AND CONCLUSIONS**

The schists at Stop 1b represent the parent material of the migmatites at Stop 1c, and the migmatization process appears to have been largely closed to gain or loss of non-volatile components. There is a steep metamorphic and thermal gradient across the migmatite front, and the migmatization reactions involved infiltration of hydrous fluids, resulting in a lowering of $\delta^{18}O_{WR}$ in the migmatites and changes in $\delta^{18}O_{mineral}$ systematics from closed to open system behavior. Trace element and oxygen isotope partitioning suggest an anatectic partial melting migmatization process, producing trondhjemitoid melts, driven by the infiltrating fluids. The presumed oxygen isotope composition of the fluids, its relationship to the isotopic composition of pegmatites and granites in the area, and the role of F and P as fluxes for melting, suggests that these fluids may have been of magmatic origin.

There are abundant pegmatites and scattered small bodies of granite throughout the migmatite zone. It is not clear, however, that these small bodies could have supplied the volume of fluids necessary to cause the isotopic and thermal changes observed. Large volumes of fluid must have exchanged oxygen isotopes with the rock in order to produce the observed lowering of $\delta^{18}O_{WR}$ (Allen, 1992).

Since they are not exposed at the surface, the plutons from which these large volumes of fluids might have originated must either be buried below the current erosion surface, or they passed through this zone and were emplaced in rocks above those now exposed and have been eroded away (Fig. 5 of trip A6, this volume, see also Allen, 1996a). While there is no direct evidence to rule out the first of these two possibilities, evidence from geologic and structural mapping (trip A6, this volume; Allen, 1992, 1996a) may support upward pluton migration through the area. This model may also explain the variations in metamorphic P–T paths observed along strike in the Central Maine Terrane. In Maine, P–T paths show isobaric heating (Lux et al., 1986; DeYoreo et al., 1989), while in central New Hampshire, P–T paths show increasing pressure during heating, and in Massachusetts, P–T paths are “counterclockwise” loops showing continued increasing pressure during cooling (Tracy & Robinson, 1980; Robinson et al., 1989). These P–T paths can be explained by the transfer of heat and mass by pluton migration upward through the crust during metamorphism. Granite emplacement in the upper levels of the crust could produce the heating observed in Maine and the tectonic loading observed in New Hampshire and Massachusetts.

**ACKNOWLEDGEMENTS**

This research was supported by National Science Foundation Grant EAR–8957703 awarded to C. Page Chamberlain, by National Science Foundation Grant EAR–9104553 awarded to Joel Blum and C. Page Chamberlain, and by Geological Society of America Research Grant 4357-90 awarded to Tim Allen. Geochemical trace element analyses by INAA were performed by Michael Glascock at the Missouri University Reactor Research Facility, and were supported in part by US Department of Energy grant DE-FG07-80ER10725 and other DOE Reactor Sharing grants to the Missouri University Research Reactor. Thanks to the following for their advice, comments on previous versions of related manuscripts and/or for discussions on earlier field trips: Page Chamberlain, Joel Blum, Leslie Sonder, Doug Rumble, Mark Conrad, Jinny Sisson, Gray Bebout, Kip Hodges, Dyk Eusden, John Lyons, and Bob Moench, and students and colleagues at Dartmouth and Keene State.
ROAD LOG

This trip was run as trip C2 of the 1996 NEIGC (Allen, 1996b), and many of these stops were also visited on a GSA field trip in 2001 (Allen et al., 2001). The stratigraphy and structure in the region is discussed in trip A6, this volume (see also Allen, 1996a)—participants are urged to read the chapter for that trip as well. All of these stops can be located on Washburn’s 1:20,000 map of “Mount Washington and the Heart of the Presidential Range” (1988).

MEETING POINT: Great Gulf Wilderness Parking Area on the west side of NH 16, about 6 miles south of its intersection with US 2 in Gorham, or 3 miles north of Wildcat Ski Area. Our first stops are here, on foot. But set your odometer to 0.0 as we’ll be driving on to subsequent stops.

STOP 1. THE MIGMATITE FRONT (~1.5 to 2 hours): Walk down the trail and across footbridge over the Peabody River; we will return to these outcrops later (Stop 1C). About 150 feet beyond the footbridge, the old Great Gulf Link trail enters sharply from the left; follow this abandoned trail north past some rock ledges on the west bank of the Peabody River (Stop 1B) until it joins the re-located Great Gulf Link trail, and then about 300 feet further to rock ledges on the west bank of the Peabody River.

STOP 1A. SCHISTS and QUARTZITES of the PERRY MOUNTAIN FORMATION (Stop 1A/2A of Hatch and Wall, 1986) (15 minutes): Interbedded quartzites and mica schists, with well preserved bedding, locally graded, with some boudined calc-silicate beds forming “pods.”

STOP 1B. SCHISTS of the RANGELEY FORMATION (Stop 2C of Hatch and Wall, 1986, Outcrop #010 of Allen, 1992) (30 minutes): Coarse-grained schists with some interbedded quartzites and granofels. Calc-silicate “pods” occur in groups along bedding planes. Sillimanite nodules (pseudomorphs after andalusite) and un-oriented spangles of retrograde muscovite give the bedding/foliation surfaces a knobby texture. While there are some bed-to-bed variations, the major element, trace element, and stable isotope compositions are all consistent with a metamorphosed shale. There is no evidence that these rocks reached the sillimanite-potassium feldspar (muscovite absent) grade.

STOP 1C. MIGMATITES of the RANGELEY FORMATION (Stop 2D of Hatch and Wall, 1986; Outcrop #011 of Allen, 1992) (45 minutes): The migmatite leucosomes are usually elongate blebs or stringers, rather than continuous layers and the intensity of migmatization is highly variable. Calc-silicate “pods” are abundant, and usually lie parallel to the migmatite layering, although sometimes they are at an angle to it. These rocks are obviously quite heterogeneous, but in bulk they are geochemically similar to the protolith schists at Stop 1B (Figs. 2 & 3). This suggests that the migmatization of these rocks is the result of metamorphism in an iso-chemical or “closed” system. The migmatites have a lower oxygen isotope value than the schists (Fig. 1) and exhibit open-system isotopic behavior (Fig. 4), which suggests infiltrating fluids may have been important in driving the migmatization process. Isotopic fractionation between the melanosome and leucosome components of the migmatites may be the result of partial melting (Fig. 9). These migmatites have an identical metamorphic mineral assemblage to that at Stop 1B (sillimanite-muscovite), but garnet-biotite geothermometry indicates temperatures were higher. Two generations of pegmatite cut across the outcrop.

0.0 Return to vehicles and drive south on NH 16, towards Wildcat Ski Area and Jackson, NH.
0.3 Turn left into 19 Mile Brook trailhead parking area. Walk across the highway bridge over 19 Mile Brook, cross route 16 (watch for traffic!), and head down a gated access trail along the brook.

STOP 2. RANGELEY MIGMATITES (20 minutes): Extensive washed and pot-holed outcrops of gray to rusty orange migmatitic gneiss. The rocks are generally gray adjacent to granite/pegmatite intrusions, rusty elsewhere. Locally the gneiss appears to “grade” into granite. There are “diffuse” pegmatite bodies, as well as pegmatites with distinct sharp contacts. Pods with moats appear to occur in horizons, generally parallel to the layering. A nice polished surface on the floor of a pool in 19 Mile Brook, approximately 100 feet upstream from its confluence with the Peabody River, shows leucosome/melanosome layering in the migmatite gneiss and the internal structure of pods and their relationship to one another (a family of pods within a larger pod!).

Return to the cars and continue south on NH 16
1.3  pass entrance to Mount Washington Auto Road on the right
1.6  paved parking area on the right. The road cuts opposite are the destination of a traverse made in Stop 3B.
2.0  Pull off into a paved parking area on the right hand side of the road. Park near the upper end. Make your
way down to Emerald Pool and then right over ledges and through woods to a small beach just upstream of
the pool.

STOP 3A. EMERALD POOL  (Stop 3A of Hatch & Wall, 1986) (30 minutes): Here we can examine almost
the entire section in one spot, including orange colored Rangeley migmatite gneiss, rusty and flaggy Smalls Falls
quartzites and schists, well banded light- and dark-green calc-silicate granofels of the Madrid, and aluminous gray
Littleton gneiss.

Return to the parking area and carefully cross the highway to roadcuts on the east side.

STOP 3B. LITTLETON to RANGELEY SECTION  (Stop 3B of Hatch & Wall, 1986) (45 minutes): Walk
north along this half-mile long series of outcrops on the east side of NH 16 and find the transition from the gray
“sinewy” migmatites of the lower Littleton rocks back to the rusty of the Rangeley Formation, with distinctive pods
and exotic cobbles. An outcrop of clean quartzite and schist just off the road in the woods between these roadcuts
may be Perry Mountain. The rocks at Emerald Pool and along these roadcuts appear to form an isolated septum of
Smalls Falls, Madrid and Littleton rocks within the migmatites of the Rangeley Formation (Fig. 1).

Return to the cars and continue south on NH 16.

2.3  Pull off into a paved parking area on the right hand side of the road. Park near the lower (northern) end,
where access to pavement outcrops in the Peabody River is obvious.

STOP 4. WILDCAT GRANITE  (Stop 4 of Hatch and Wall, 1986; Outcrop #105 of Allen, 1992): For those
familiar with the controversy, Granitization rears its ugly head! This is the type locality for the rock I have called
Wildcat Granite ( trip A6, this volume; Allen, 1992, 1996a). There are clearly two different phases: “G” consists of
medium grained, whitish-weathering two-mica granite; “R” is much coarser grained, orangish-weathering granitoid,
also bearing both muscovite and biotite, but with much more abundant biotite. “R” contains abundant calc-silicate
pods, identical to those found in the metasediments, rimmed by strong reaction zones, and its textures and
mineralogy suggest that it may be formed from completely melted and recrystallized Rangeley schists. Both the “G”
and “R” phases are extensively intermingled in a complex fashion, with wispy biotite-rich schlieren observed
throughout the outcrop. Geochemically, the “G” phase is characteristic of an “S” type granite (Figs 1, 2, 3 & 7); the
“R” phase has a composition intermediate between that of the “G” phase and those of the schists and migmatites.
Second-generation white, tourmaline-bearing pegmatites cross-cut both phases

Return to the cars and continue south on NH 16.

3.1  Turn left into Wildcat Ski Area.

STOP 5. (OPTIONAL) THOMPSON’S FALLS  (45 minutes): Follow the Thompson Falls Trail 0.8 miles to
its end at ledges of Wildcat Granite grading into Rangeley migmatites. The contact between the Wildcat Granite and
the surrounding migmatitic metasedimentary rocks is gradational—not a sharp intrusive contact. Excellent views of
Mt. Washington and its ravines.

Return to the cars and continue south on NH 16.

4.0  Turn right into the AMC’s Pinkham Notch Base Camp. Parking may be difficult to find, in which case
continue to the auxiliary parking lot on the south side of the Cutler River. Several optional stops may be
made from here.

STOP 6A. (OPTIONAL) CRYSTAL CASCADE  (30 minutes): Follow the Tuckerman Ravine Trail 0.3 miles
to the Crystal Cascade of the Cutler River. The rocks at the Cascade are Mesozoic volcanic vent agglomerates
(Billings, et al., 1979). Downstream are exposures of granite and pegmatite, and laminated rusty schists of the
Smalls Falls Formation.
STOP 6B. (OPTIONAL) OLD JACKSON ROAD to LOWE’S BALD SPOT (2 hours): Follow the Old Jackson Road (trail) 1.8 miles to the Mount Washington Auto Road, roughly following the contact between the Smalls Falls and Madrid or Littleton Formations. This belt of Smalls Falls and Madrid is distinct from that at Emerald Pool (Stop 3A) and they do not connect. Continue 0.2 miles on the Madison Gulf Trail to Lowe’s Bald Spot, a nice exposure of Littleton schists.

STOP 6C. (OPTIONAL) SQUARE LEDGE (1 hour): Cross to the east side of NH 16 carefully. Follow the Square Ledge Trail 0.5 miles to the top of Square Ledge, an outcropping of non-descript gray Rangeley migmatites with an excellent overlook of Pinkham Notch and view of Mt. Washington and its ravines.

Return to the cars and continue south on NH 16.

4.5 Pull off on right shoulder of NH 16, at a small waterfall opposite a turnout on the northbound side and just before a turnout on the southbound side.

STOP 7. SMALLS FALLS, MADRID & LITTLETON FORMATIONS (15 minutes): The rocks at the base of the falls are laminated rusty schists of the Smalls Falls Formation. These are overlain by layered green calc-silicate granofels of the Madrid Formation near the top of the falls, and above that by well bedded aluminous schists of the Littleton Formation.

Return to the cars and continue south on NH 16.

4.7 Turn right into Glenn Ellis Falls Scenic Area. Follow path under NH 16 and down to falls.

STOP 8. (OPTIONAL) GLENN ELLIS FALLS (30 minutes): Abundant pegmatite and pod-bearing Wildcat Granite. Extremely punky weathering and possibly brecciated.

5.2 (add 0.4 if you went in to Glenn Ellis Falls) Turn left into turnout on east side of NH 16 overlooking the Ellis River valley and Jackson, NH. Carefully cross the highway to roadcuts on west side.

STOP 9. OVERLOOK, more WILDCAT GRANITE (15 minutes): A coarse grained granular rock, with abundant calc-silicate pods and dark rusty red stains. Small weathering pits on the surface may represent cordierite. I have not studied this rock in detail, but would assign it to the Wildcat Granite, representing melted and recrystallized Rangeley and possibly Smalls Falls Formations.

End of trip, safe travels to your next destination.

REFERENCES CITED


