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THE GEOLOGY OF THE LAKE CHAMPLAIN BASIN AND VICINITY
Proceedings of a Symposium

CONTENTS

VERMONT GEOLOGICAL SOCIETY, INC.

Vermont Geological Society
Box 304
Montpelier, Vermont 05602

FORWARD

The Vermont Geological Society was founded in 1974 for the purpose of "1) advancing the science and profession of geology and its related branches by encouraging education, research and service through the holding of meetings, maintaining communications, and providing a common union of its members; 2) contributing to the public education of the geology of Vermont and promoting the proper use and protection of its natural resources; and 3) advancing the professional conduct of those engaged in the collection, interpretation and use of geologic data". To these ends, in its 7 year history, the society has promoted a variety of field trips, an exposition on Vermont geology, presentations of papers by both professional and student researchers, teacher's workshops, a seminar on water quality, a soils workshop and a seismic workshop. The society was also active in lobbying for continuation of the State Survey in 1975 and in selecting a successor to the retiring state geologist at that time. Other position papers have been prepared in response to issues of land use and uranium prospecting. Four yearly meetings have become established: an all day fall field trip, presentation of professional papers in the winter, and in the spring, presentation of student research papers and a teacher's workshop and field trip.

The society's newsletter, the Green Mountain Geologist, containing announcements and short articles, is published quarterly, but it always has been thought that we should publish a bulletin devoted to Vermont geology. This, our first effort, contains six of the papers presented at our third annual winter meeting in February 1980 held at Norwich University in Northfield, Vermont. Each paper published in this issue has also been reviewed by at least two outside professionals.

We intend to continue using papers from the winter meetings as a basis for future volumes of Vermont Geology. Plans already have been made for the 1981 meeting to be held early in February. Papers are being enlisted under two topics: applied geology in Vermont and regional studies in western New England and adjacent Quebec. Anyone interested in presenting a paper about either of these topics should contact Lance Meade, PO Box 133, Pittsford, Vermont 05763 (applied geology) or Rolfe Stanley, Perkins Geology Hall, University of Vermont, Burlington, Vermont 05405 (regional studies). Anyone wishing further information about the time and place of the meeting should write to VGS at the above address.

ACKNOWLEDGMENTS

The Vermont Geological Society is indebted to assistance received from the Academic Computing Center of the University of Vermont and the Department of Geology, also of UVM. Reviewers for this issue were: D.P. Bucke, B.H. Corliss, R. Falkundiny, F.J. Gunther, S. Holdahl, N.M. Ratcliffe, C.A. Ratté, D. Vanecek and D. Wise.

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INTRODUCTION TO THE ENVIRONMENTAL GEOLOGY OF LAKE CHAMPLAIN AND SHORELAND AREAS

R. Montgomery Fischer
Lake Champlain Basin Program
New England River Basins Commission
177 Battery Street, Ice House
Burlington, Vermont 05401

ABSTRACT

Lake Champlain, the dominant physical feature of an 8,234 square mile (21,325 square kilometer) drainage basin, lies at the bottom of the broad, open valley between the Adirondacks in New York and the Green Mountains in Vermont. The lake forms the boundary between the two states for 100 miles, and its total length is 125 miles from its southern end near Whitehall (New York) to its outlet near Ash Island (Quebec) where it discharges into the Richelieu River. The area of the lake is approximately 439 square miles. Tributaries which drain the basin account for 90 percent of the water entering Lake Champlain.

In 1975 it was estimated that 460,000 persons lived in the ten counties which generally correspond to the boundary of the basin. About one half of the total basin population lives in the communities bordering the lake. The waters of the lake support many of the needs of these communities. It is used primarily for water supply, wastewater disposal, swimming, boating and fishing.

Flooding along the shores of the lake is an international problem. Increased eutrophication, growth of nuisance aquatic plants, runoff from agricultural land and presence of toxic substances threaten the future water quality of the lake.

The challenge facing basin residents is to maintain or improve the current quality of the basin environment and at the same time accommodate the demands for continued economic growth within the basin.

Lake Champlain and the rivers that drain into it provide the ingredients that make up the character of the Lake Champlain basin as we know it today (Figure 1). This basin is nestled between the Adirondack Mountains to the west and the Green Mountains to the east. In the middle is Lake Champlain, one of the largest lakes in the U. S., aside from the Great Lakes (Johnson, 1980).

In 1975 close to 460,000 persons lived in the ten U. S. counties which generally correspond to the boundary of the basin. Forty-five percent of the total basin population lives in the communities bordering the lake (Figure 2). Between 1980 and 1990, it has been projected that there will be a 16 percent population increase in these shoreland communities (Vermont State Planning Office, 1977).

Since the climate is mild, as compared with other areas in Vermont, farmers have long since cleared the white pine forests that originally grew here and changed the area into a prime agricultural producing region. The large lake-adjacent marshes are important breeding, spawning, nesting and resting areas for a variety of wildlife.

Located in the northeastern part of the United States and southern Quebec in Canada, the Lake Champlain basin is bounded by the Chateauguay and St. Lawrence River basins in the west and north. To the east it is bounded by the St. Francis and Connecticut River basins, and to the south by the Hudson River basin. The total basin area is 8,234 square miles (21,325 square kilometers)

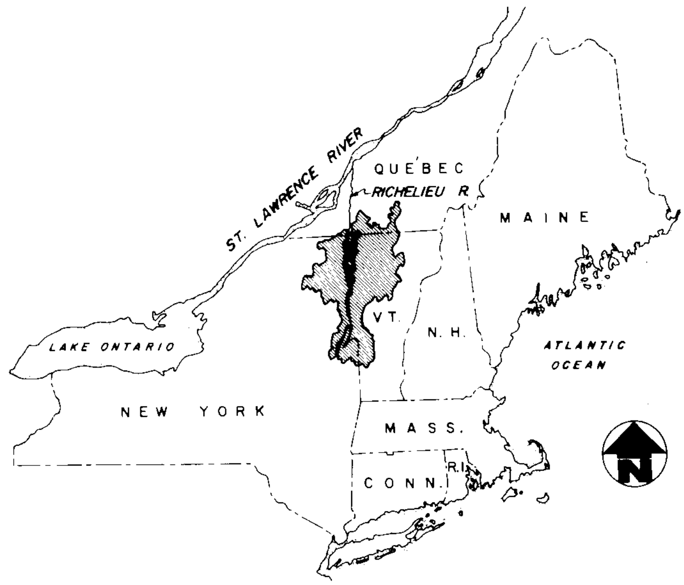

Figure 1. General location of Lake Champlain drainage basin. Source: Lake Champlain Basin Study, 1977.

(Figure 3). It is split by political boundaries, with 56 percent of its land areas in Vermont, 37 percent in New York, and 7 percent in Quebec.

Lake Champlain is one of the largest freshwater lakes in the United States, and unlike most U. S. watercourses it flows northward. It is 125 miles (200 kilometers) long, as measured from its southern end near Whitehall, New York, to its outlet near Ash Island, Quebec, where it discharges into the Richelieu River.

Being a north-south trending lake, it is narrow, measuring only 12 miles at its widest point near Burlington, Vermont, although the broad plain between the Adirondacks on the west and the Green Mountains to the east is almost 20 miles (32 kilometers) wide.

It has a surface area of 439 square miles (1,124 square kilometers), and a maximum depth of nearly 400 feet (122 meters) just off Charlotte, Vermont. The lake is subdivided into at least five distinct morphological units, each with its own individual characteristics (Figure 4). Seventy-one small islands, each smaller than 1,000 acres, dot the surface of the lake.

About 300 tributary streams feed directly into the lake, with 70 percent of the drainage coming from Vermont sources. This is consistent with the fact that two-thirds of the watershed lies to the east of the lake.

The character of the basin and the tributary basins can be directly traced to past geologic events, but these events are not readily revealed without careful analysis and insightful interpretation of the geologic record.

The Lake Champlain Valley is a lowland developed on glacial lake and marine sediments, which

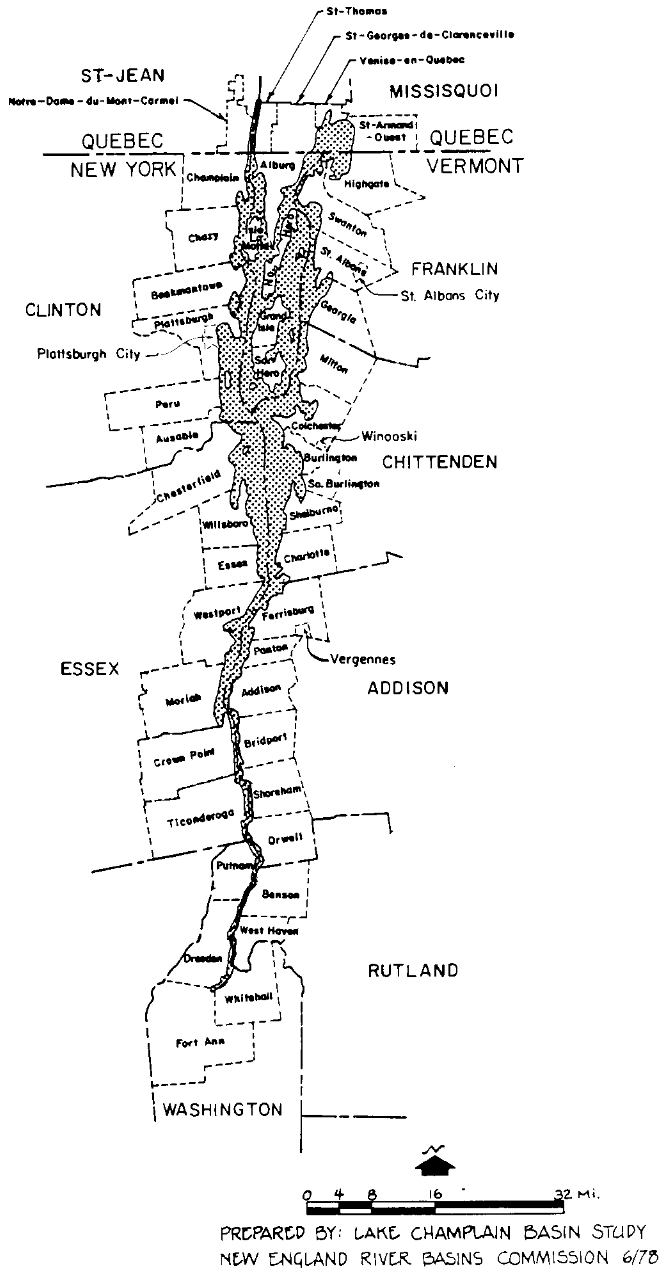

Figure 2. Shoreland towns of Lake Champlain.

in turn overlay eroded beds of Cambrian and Ordovician ages. The valley itself is located between the intensively metamorphosed Precambrian rocks of the Adirondack Mountains on the west and the lesser metamorphosed severely compressed Cambrian strata of the Green Mountain anticlinorium on the east.

The Champlain Valley was a shallow (500-600 feet deep) warm sea during the Early-Middle Ordovician Period, about 550 million years ago. This is when many of the plants and animals that have now become fossils were living organisms. Button Island and Isle La Motte have fine evidence of coral fans, brachiopods, trilobites and sponges.

The recent glacial history of the basin is of importance when unraveling why the region looks as it does today. Only the last glaciation, the Wisconsin, is well known in New York, Vermont and southern Quebec. Earlier glacial advances during the Pleistocene obscured or obliterated evidence of previous ones.

The Ice Age started about one million years ago in this region. Four major periods of ice covered New England. The glacier, as it was grinding away, left one common form of glacial deposit, which is known as glacial till. Then as the glacier receded, several higher level lakes occupied the Champlain Valley. Collectively geologists call it "Lake Vermont"; however, there are several stages of Lake Vermont (Table 1). As the ice was melting, all of its melt water collected in the Champlain basin and could not flow anywhere, so it made large lakes in the Champlain Valley up to the elevation of at least 600 feet. All the way into Montpelier, areas were flooded 10,000 years ago as the ice was leaving.

As is happening now in Lake Champlain, material was carried by the rivers out into Lake Champlain. The heavier material does not travel far into the lake; it gets deposited right at the mouth of the river and builds a delta, like the Winooski River is now building. So there were successively lowered stages of this Lake Vermont as the ice receded northward. Then for a brief period the ice receded north of the St. Lawrence Valley, and salty water from the Atlantic came into the Champlain Valley. There was a marine environment in the basin for 1,500 years, and it existed at an elevation of about 300 feet above sea level. The flat, sandy land that South Burlington is built on is a delta that was deposited by the Winooski River flowing into the Champlain Sea. The Champlain Sea was able to come in because the ice receded, but the crust of the earth had not been able to spring back up from its load of ice. After a relatively short time (about 1,000 years), the earth recovered and started to bounce back after the ice load had been removed. This rebound of the land caused the rock sill at Rouses Point to rise and close off the flow of sea water. As the sea was closed off, all the major rivers dumped fresh water into what was the Champlain Sea. That salt body slowly freshened into what we now know as Lake Champlain. The northern end of the lake is still rising slowly, and if you extrapolate the rate of rise into the future, it appears that in 10,000-15,000 years the lake could drain to the south rather than to the north as it does now.

There are a number of problems facing lake users today. Flooding along the lakeshore is a chronic problem with international implications. Associated with this flooding problem is the severe rate of erosion along more than 200 miles of Lake Champlain's 600 mile shoreline (Table 2). Although shoreline property owners have tried many methods to retard erosion, most have only achieved limited success. Improperly located development has placed increased construction costs on shoreline property owners and contributed to a visually degraded shoreline.

Dense growth of milfoil and other rooted aquatic vegetation covering nearly 300 acres in St. Albans Bay have contributed to making the bay virtually unavailable for swimming and boating during certain times of the year.

Lake Champlain water is still safe to drink, and fish from it are good to eat. The levels of metals and organic compounds pose no immediate threat to public health and safety. However, certain potentially toxic substances may be present. In the face of increasing industrial and municipal discharges, along with the increase in the use of chemicals in manufacturing processes, there is a concern over preserving the current high quality of the lake.

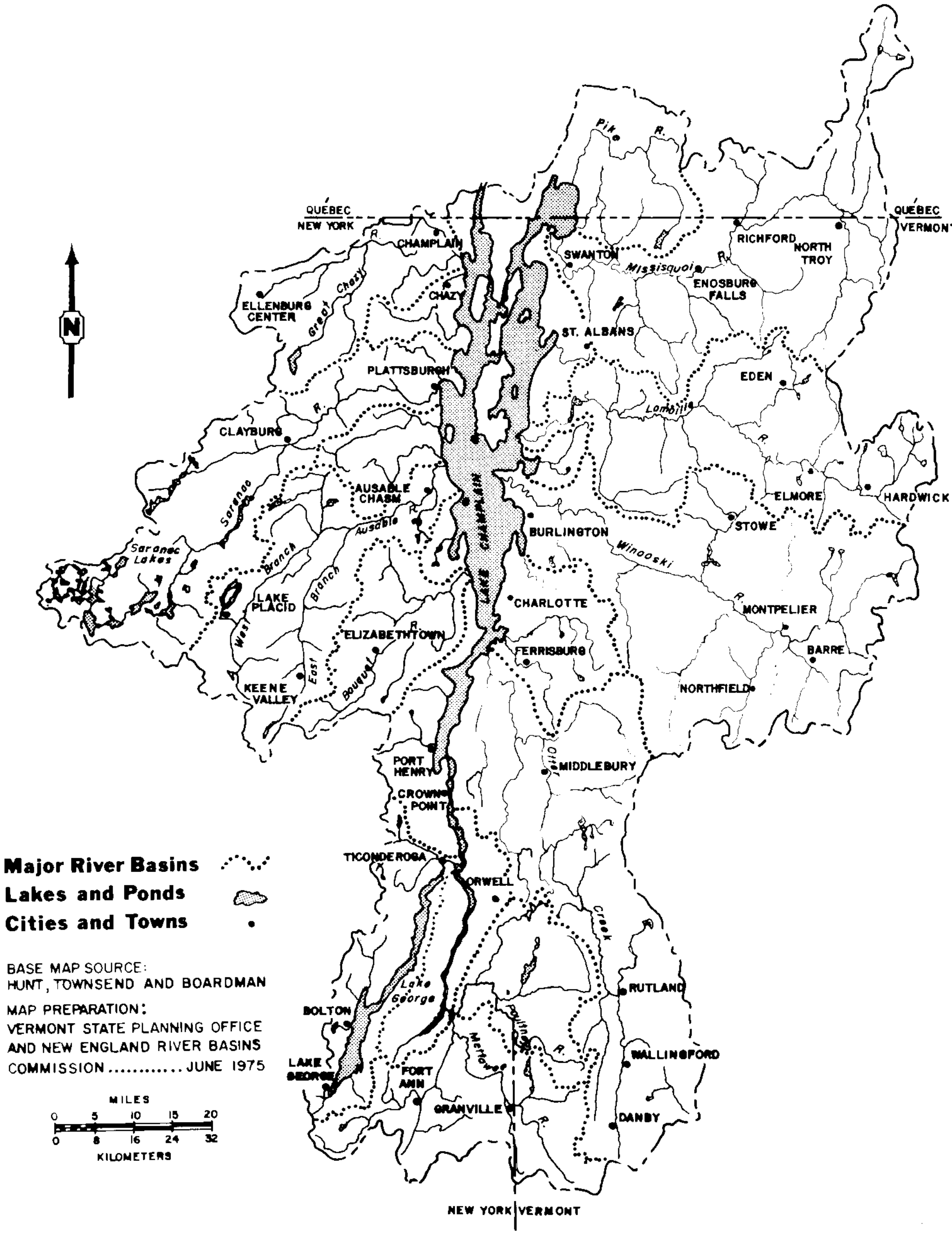

Figure 3. Lake Champlain drainage basin.

TABLE 1. BEACH LEVELS OF MAJOR WATER BODIES IN THE CHAMPLAIN VALLEY

Body of water	Beach levels in Plattsburgh, N.Y. area		Beach levels in Rouses Point, N.Y. area	
	(meters)	(feet)	(meters)	(feet)
Lake Champlain				
Present day	29	95	29	95
Champlain Sea				
Port Henry stage	27	89	?	?
Plattsburgh stage	40- 52	130-170	52- 67	170-220
Port Kent stage	76- 91	250-300	91-110	300-360
Beekmantown stage	91-113	300-370	110-134	360-440
Upper Marine stage	113-140	370-460	140-162	460-530
Lake Vermont				
Fort Ann stage	165-180	540-590	180-198	590-650
Coveville stage	192-207	630-680	207-229	680-750

SOURCE: Myer and Gruendling, 1979.

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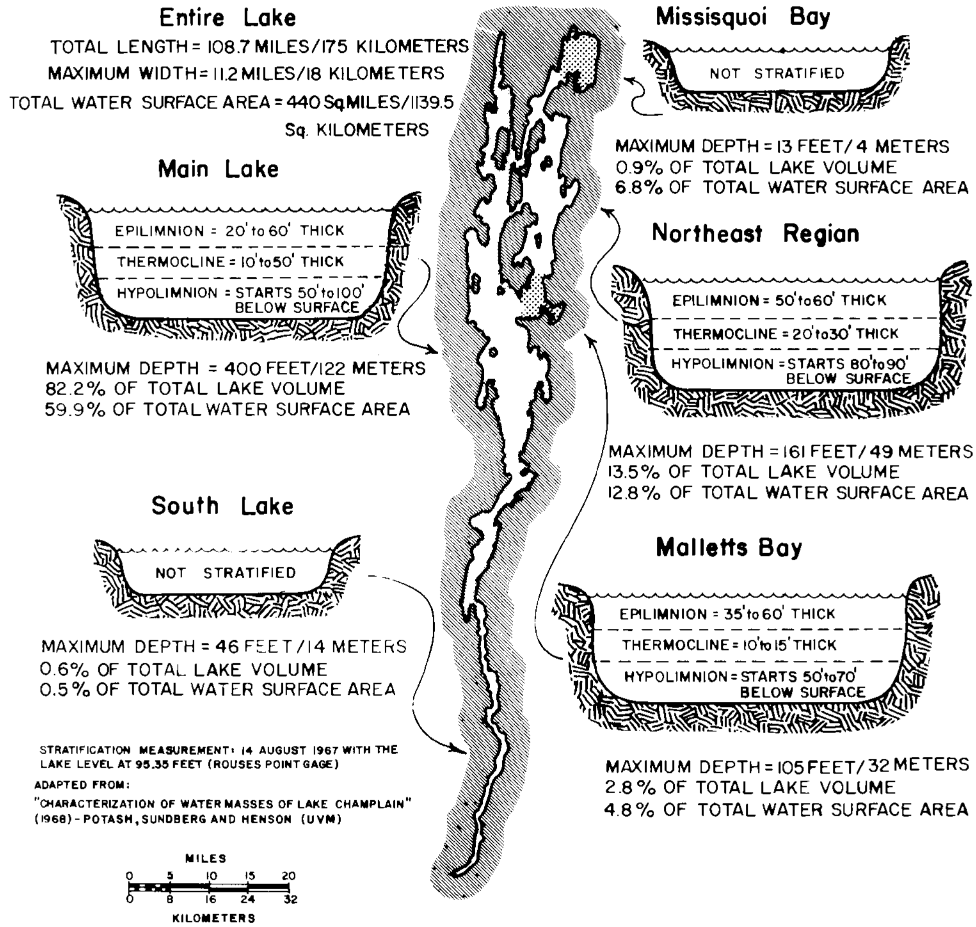

Figure 4. Morphometric characteristics of Lake Champlain. Source: State of New York, State of Vermont, and the New England River Basins Commission, Lake Champlain Planning Guide, 1976.

TABLE 2. PRESENT SHORELINE EROSION AND PROTECTION, LAKE CHAMPLAIN, 1977.

	PRESENT EROSION								PRESENT PROTECTION							
Categories of erosion	Quebec Miles	%	New York Miles	%	Vermont Miles	%	Total Miles	%	Quebec Miles	%	New York Miles	%	Vermont Miles	%	Total Miles	%
None	3	25	46	30	83	27	132	28					0.5	0	0.5	0
Slight			19	13	30	10	49	11			2.1	4	2.2	2	4.3	2
Moderate	3	25	20	13	55	18	78	17	2.6	31	6.4	12	27.5	24	33.9	20
Severe	6	50	41	27	85	28	132	28	57	69	33.9	65	64.0	56	103.6	60
None-slight			13	9	21	7	34	7			1.6	3	1.9	2	3.5	2
Slight-moderate			2	1	9	3	11	2			1.6	3	2.7	2	4.3	2
Moderate-severe			11	7	20	7	31	7			6.3	12	15.7	14	22.0	13
	12	100	152	100	303	100	467	100	8.3	100	51.9	99	114.5	100	172.1	99

SOURCE: Hunt, 1977

THE APPLICATION OF LAKE CHAMPLAIN WATER LEVEL STUDIES TO THE INVESTIGATION OF ADIRONDACK AND LAKE CHAMPLAIN CRUSTAL MOVEMENTS

Stockton G. Barnett
Department of Earth Sciences
State University of New York
Plattsburgh, New York 12901

Yngvar W. Isachsen
New York State Geological Survey
Albany, New York 12230

ABSTRACT

Releveled lines along the eastern flank of the Adirondack Mountains and across the central part of the dome suggest that the region is undergoing contemporary uplift; the line along the eastern flank suggests that the northern end of Lake Champlain is sinking at the rate of 4.0 millimeters per year relative to the southern end. Lake Champlain, which borders the northern three-quarters of this line and has a demonstrably stable lake-level gauge at each end, provides a perfect "natural level" to evaluate the releveling profile and further refine interpretations of the region's neotectonics.

Apparent vertical movement between Whitehall, New York, at the southern end of the lake and Rouses Point, New York, at the northern end was determined by linear regression of the differences between mean lake levels over the eight-month period, April-November, for the time interval 1940-1978. The linear regression slope indicates that the northern end of the lake is rising 0.7 millimeters per year with respect to the southern end. This movement is opposite to that suggested by the releveling survey along the west side of the lake.

The seeming disparity between results of the two methods may be reconciled if one recognizes that the releveling profile indicates only relative movement, and uses the lake level data to "calibrate" the releveling profile. The releveling profile can be superimposed on the water level gauging profile by drawing a line between Whitehall, New York and Rouses Point, New York with an upward slope of 0.7 millimeters per year to the north. The resulting combined profiles suggest that during the 18 year interval 1955-1973 the region from Willsboro, New York north to Rouses Point, New York remained essentially stationary, while that between Whitehall, New York and Willsboro, New York underwent an arching of 2 millimeters per year. It appears noteworthy that this uparched area corresponds exactly with the more elevated and breached part of the Adirondack dome, suggesting that the present arching along the eastern Adirondacks is the continuation of more prolonged neotectonic uplift. On a more regional scale, the comparison of vertical movement rates as determined from water level gauge data indicates that Burlington, Vermont may be subsiding relative to Rouses Point, New York, Phillipsburg, Quebec, and Whitehall, New York. These tentative suggestions of Adirondack movement need independent verification. A continuously recording water level gauge placed in the Ticonderoga-Westport, New York area could provide such verification within 10 years.

INTRODUCTION

Knowledge of crustal movement patterns in the Champlain Valley and Adirondacks is of interest for both basic scientific and practical reasons. From a geological standoint it is important to know to what extent the region may still be undergoing post-glacial rebound, and to what extent it may be responding to contemporary tectonic forces as has been suggested from releveling (Isachsen, 1975, 1976).

From the point of view of basin management, possible navigational problems in the southern part of Lake Champlain could be predicted if the lake outlet in the north were determined to be undergoing significant differential sinking causing the southern end of the lake to shoal. The rate of vertical movement determined in this study, however, is negligible from a practical standpoint, as noted in the conclusions.

PREVIOUS STUDIES

The only previous study pertaining to vertical crustal movements in the Champlain Valley and the immediately adjoining portion of the Adirondacks is that of Isachsen (1975), who compared first order leveling surveys made in 1955 and 1973 along the Delaware and Hudson (D and H) Railway between Saratoga Springs, New York and Rouses Point, New York (Figure 1)*. Differences in elevation between the two surveys, using the 1955 survey as a base line and arbitrarily assuming no elevation change at Saratoga Springs, are reproduced in Figure 2. The releveling profile indicates uplift along the eastern flank of the Adirondack dome, and suggests that the northern end of the Champlain Valley is sinking about 4.0 millimeters per year with respect to Whitehall. This profile led Isachsen (1975) to suggest that the Adirondack dome is undergoing contemporary uplift and that differential post-glacial rebound in the region had been superceded by this arching.

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* See back cover.

Figure 1. General geological setting of Lake Champlain and the releveled line between Saratoga Springs and Rouses Point. The positive and negative values indicate the rate of vertical movement in millimeters per year for water level gauge locations (Whitehall, Burlington, Phillipsburg and Rouses Point) and locations along the releveled line (Ticonderoga, Westport, Willsboro and Plattsburgh).

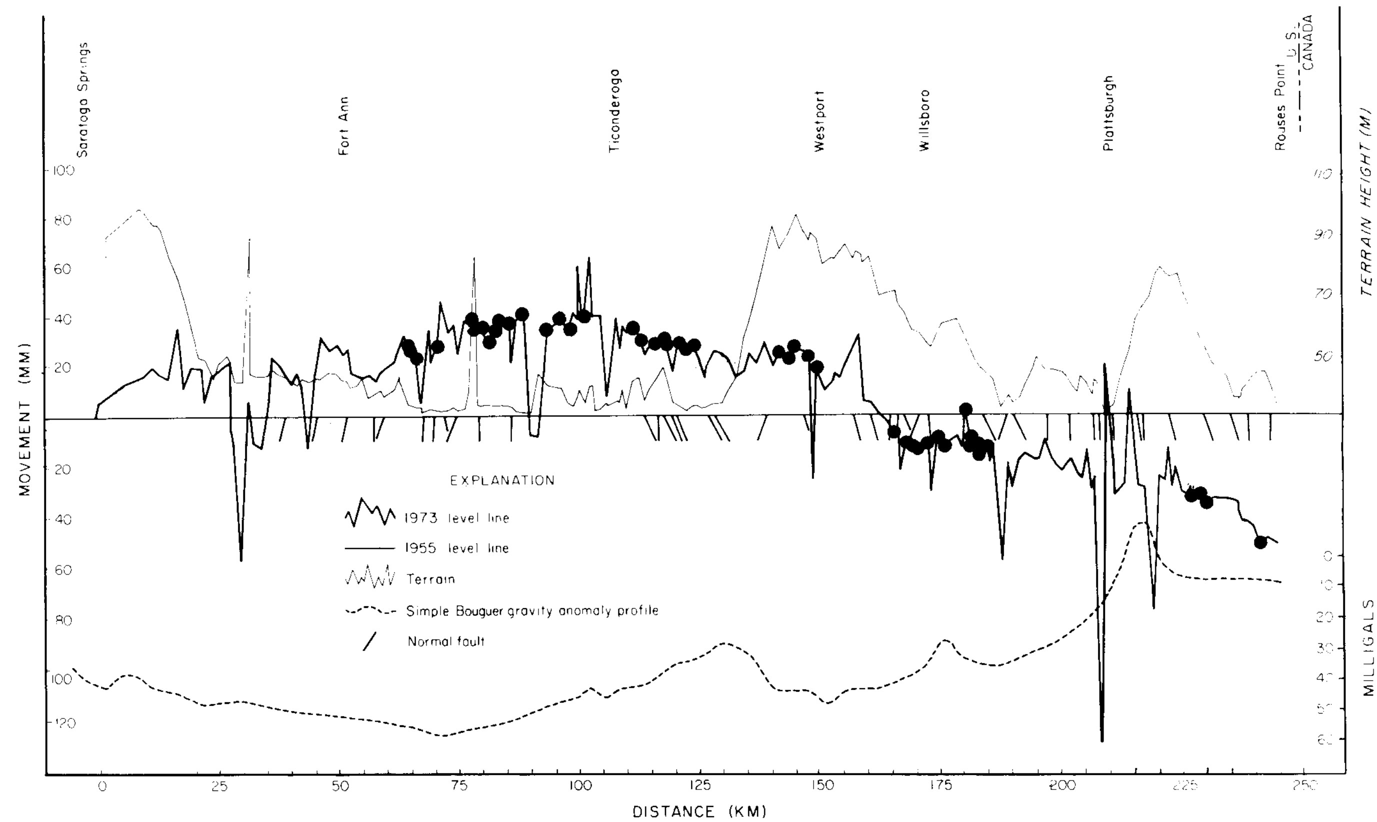

Figure 2. Geodetic releveling profile showing changes in benchmark elevations from 1955-1973, based on arbitrary assumption of no vertical change at Saratoga Springs. Solid dots represent benchmarks anchored in bedrock. From Isachsen (1975). Whitehall is located north of Fort Ann at mile 62.

PROCEDURE AND OBSERVATIONS

This study uses a different method of measuring contemporary crustal movements along the Champlain Valley, namely, that of comparing average annual lake levels over a period of years as determined from data obtained from gauging stations on Lake Champlain. A gauge at Rouses Point, New York at the northern end of the lake, and another long-occupied gauging station at the southern end of the lake in Whitehall, New York make it possible to look at water level differences across a north-south span of 150 kilometers, a distance corresponding to the northern three-quarters of the Saratoga Springs, New York - Rouses Point, New York releveled line (Figure 1.) The difference between mean annual gauge heights was plotted for the time period 1940-1977, and the apparent vertical movement rate calculated from a linear regression of the differences (Figure 3). The slope of the linear equation for this regression expresses the rate of change in elevation between these gauges; it suggests that Rouses Point, New York is rising with respect to Whitehall, New York, at a rate of 0.72 millimeters per year.

DISCUSSION

It is clear from a comparision of Figures 1 and 3 that the releveling profile and the water level data suggest opposite directions of crustal movement at Rouses Point with respect to Whitehall: the former suggests that Rouses Point is sinking, the latter that it is rising.

In evaluating this discrepancy, it is necessary to analyze possible sources of error in each type of measurement. We will first consider the

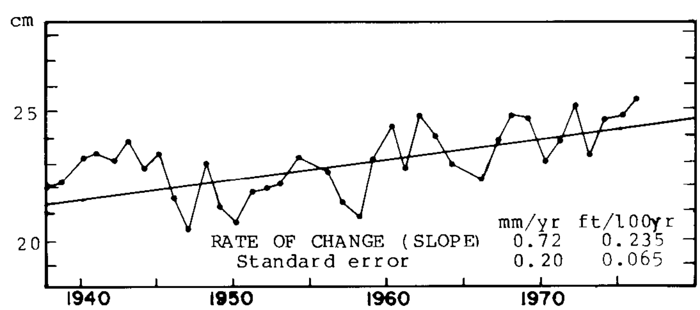

Figure 3. Rate of change of annual water-level differences between Rouses Point and Whitehall. Rouses Point is rising.

most common cause of error in water level measurements, namely gauge instability, because small movements of a gauge can produce large errors in the analysis of crustal movement. Each gauge will be discussed in terms of its vertical stability, after which other factors that can influence water level readings will be considered.

The lake level gauge at Rouses Point is the longest-established gauge on Lake Champlain, with daily readings available from 1871 to the present. During this time period, however, the gauge was relocated twice. From 1871 to 1923 a manual gauge, read once daily, was located at Fort Montgomery (Figure 4). From 1923 to 1938 the gauge was positioned on the Rutland Railroad bridge 27 meters south of the present gauge site. It also was manual

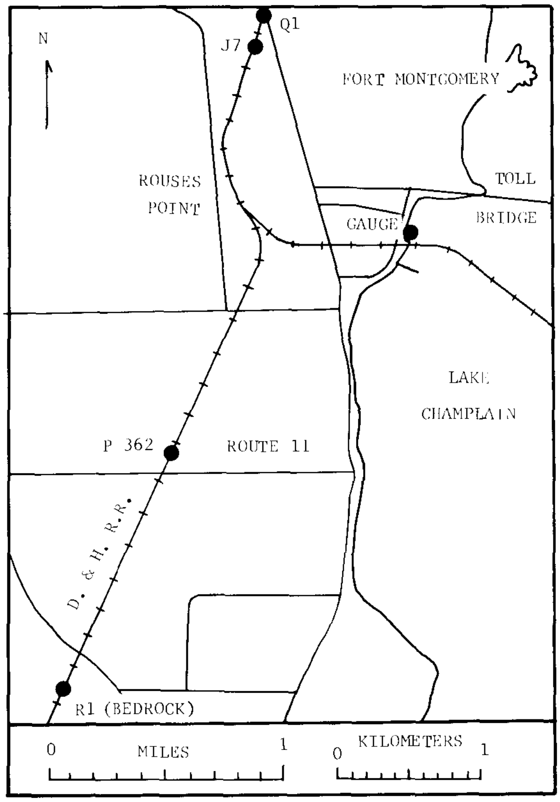

Figure 4. Location of Rouses Point lake level gauge with respect to benchmarks on the releveled line along the D. and H. Railway.

and read once daily. From 1938 to the present, a continuously-recording gauge has been in operation at the present gauge site. To monitor its stability, this gauge has been surveyed to a Coast and Geodetic benchmark network several times since 1938. Results of these surveys are shown in Table 1.

The benchmarks identified in the field as Gauge BM NO. 1, Gauge BM NO. 2,5,and 6 are part of the gauge network and are located within 100 meters of each other and the gauge. The other benchmarks (R1, Q1, and J7) are located along the releveled line which runs 1 km west of the gauge (Figure 4); benchmark R1 is anchored in bedrock. The consistency of elevation differences between benchmarks over the long time period involved clearly demonstrates the stability of the gauge, despite its not being anchored in bedrock. The gauge is also clearly tied to the releveling profile along the D and H Railway. In this study only gauging data beginning with 1938 was used because it is not known if, in the earlier processes of moving the gauge, it was positioned at exactly the same elevation (see Barnett, 1978, p. 72b.)

The Whitehall gauge was established in 1916 on the concrete works of the then recently completed Lock 12 of the Champlain Barge canal. The gauge is stable because the lock was built on bedrock, and the gauge is periodically surveyed to a benchmark on the concrete wall of the lock several feet from the gauge. The gauge is manual and has been read four times daily. It is located only 200 meters east of the releveled line (Figure 5), although it has not been periodically surveyed to this line.

We conclude from the foregoing that the gauges at both Rouses Point and Whitehall are demonstrably stable with respect to bedrock. This leaves another possible source of error to be considered, namely the possibility that one or both gauges are located on ground that is moving locally, along faults, with respect to surrounding terrain. This possibility was tested by carefully examining the releveling data in the general vicinity of the Whitehall and Rouses Point gauges. Special attention was paid to areas where the survey lines crossed geologic faults, such as those shown along the profile in Figure 2. The releveling data showed no local variations in crustal movements within several kilometers of each gauge. We conclude that these gauges are both stable and are representative of the terrain surrounding them.

TABLE 1. ELEVATION DIFFERENCES (IN METERS) BETWEEN BENCHMARKS USED TO CHECK STABILITY OF GAUGE AT ROUSES POINT, NEW YORK

Date of leveling	Benchmarks along releveled line *			Local gauge benchmarks within 100 meters of gauge				
FROM BM:	R1	J7	Q1	5	6	6	5	GAUGE BM NO. 2
TO BM:	Q1	Q1	5	6	GAUGE BM NO. 2	GAUGE	GAUGE	GAUGE BM NO. 1
1921	25.873	+7.303						
1938						+2.926	+2.914	
1945						+2.938	+2.916	
1955	25.873	+7.303						
8/1958								-5.162
4/1961				+0.053	+4.788			-5.134
9/1961		+7.304	-13.259	+0.049	+4.795			-5.162
5/1970				+0.050				
7/1970		+7.306	-13.254	+0.051	+4.796			
10/1970				+0.050				
4/1973				+0.049				
8/1973				+0.048	+4.796			
9/1973	25.867			+0.048				
9/1976		+7.303	-13.257	+0.048	+4.796			
9/1977						+2.929	+2.914	

* Benchmarks R1, Q1 and J7 are located along releveled line; R1 is anchored in bedrock (see Figure 4).

Sources: Open files of National Ocean Survey, National Geodetic Survey, and U.S. Geological Survey

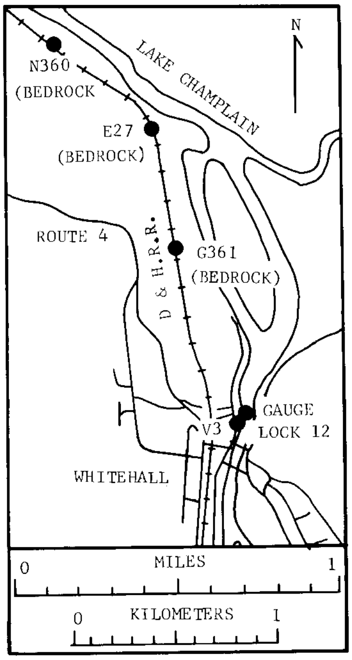

Figure 5. Location of the Whitehall lake level gauge with respect to benchmarks on the releveled line along the D. and H. Railway.

Other factors must also be considered. Day-by-day comparisons between the Rouses Point and Whitehall gauges indicate that water levels at Whitehall are more sensitive to runoff than those at Rouses Point. After a storm, levels rise to their maximum at Whitehall in about two days, as compared to five or six days at Rouses Point. The maximum at Whitehall has, on occasion, been 0.6 meters or more higher than the synchronous level at Rouses Point. This, however, is only a temporary condition, confined primarily to the spring runoff period and the use of the April-November data for annual gauge comparisons dampens out this effect. Winter data is not used for this gauge because of the difficulty of reading a manual gauge when ice conditions are present.

Although measurable tidal conditions are not present in the lake, a final source of possible error is the local fluctuation in lake level introduced by changing patterns of wind and barometric pressure over the lake. The scatter that occurs in lake level data probably largely reflects this effect, the most pronounced of which would be in the form of seiches. Such events, however, occur so frequently during the course of a year, that the averaging of daily records largely dampens their effect.

The above analysis suggests that the Rouses Point and Whitehall gauges are probably providing an accurate picture of crustal movements.

It is now necessary to evaluate the possible sources of error in the geodetic levelings represented in Figure 2. First order leveling is done with a high level of precision. For random errors, the standard deviation of the 1955 first order leveling is approximately equal to 1.0 millimeters (distance in kilometers)$^{1/2}$, and that for the 1973 leveling 0.7 millimeters (distance in kilometers)$^{1/2}$ as estimated by the National Geodetic Survey (see Isachsen, 1975, p. 175). The total deviation for both surveys would be $(\rho_1^2 + \rho_2^2)^{1/2}$ where ρ_1 and ρ_2 are the standard deviations for the two levelings. For a distance of 100 kilometers, for example, this amounts to only 12 millimeters.

A more important source of error is benchmark instability. To focus attention on this factor, benchmarks anchored in bedrock have been separately identified by dots in Figure 2. It is considered significant that these benchmarks show little deviation from a smooth curve. All the wild "spikes", in contrast, occur in less stable monuments. The majority of spikes can readily be explained as due to settling, and all can be eliminated from consideration, as representing movements independent of bedrock.

Numerous systematic errors remain to be considered that can creep into precision leveling and which are difficult or impossible to evaluate after the fact. These are enumerated and discussed at some length by Bomford (1971) who concludes that crustal movements as determined by releveling surveys can safely be considered real, if:

1. water level measurements at a number of gauges agree with the releveling data.

2. observations at three or more levelings show progressive vertical change; Wilson and Wood (1980, see our Figure 6) show a successful application of this criterion.

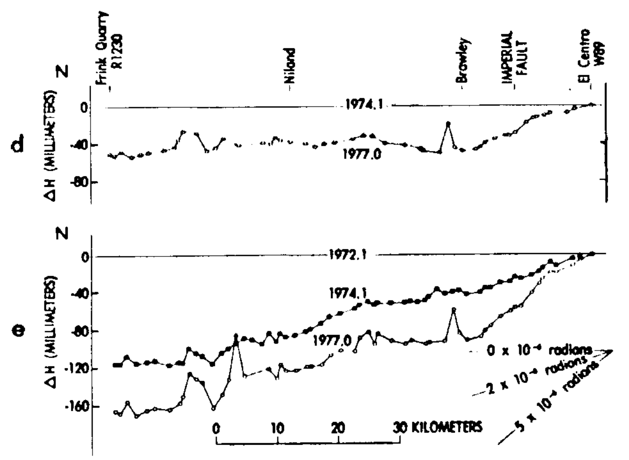

Figure 6. First-order leveling surveys repeated three times near the Salton Sea, California (From Wilson and Wood, 1980, p. 184).

3. geological evidence, such as raised beaches, is in agreement with the releveling profile.

4. systematic error can be determined; circuit closures and fore-and-back-leveling can eliminate some sources of systematic error, but not all sources can be eliminated easily.

We will now use these criteria to evaluate the Champlain Valley releveling surveys following the same numbering as above:

1. As pointed out in the beginning of this section, the water level measurements do not agree with the apparent vertical movements shown by the releveling, and it is this disagreement we seek to explain.

2. Three separate leveling surveys have been made along the railroad bed which follows the west shore of Lake Champlain. These were done in 1921, 1955, and 1973. The 1921-1955 releveling

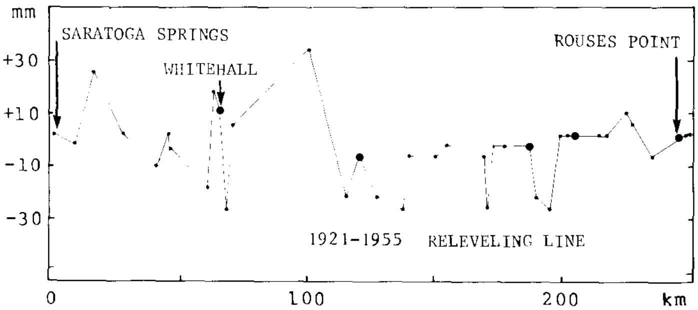

Figure 7. Geodetic releveling profile for the interval 1921-1955 based on arbitrary assumption of no vertical change at Saratoga Springs. Solid dots represent benchmarks anchored in bedrock. The vertical scale has been reduced to 50 percent of the original because the time between the two surveys is twice that of the 1955-1973 releveling line. This allows the two releveling lines to be compared more readily.

profile is shown in Figure 7. It does not depict the pattern of arching shown in the 1955-1973 profile of Figure 2. This disagreement might well be attributable to the larger standard deviation in the 1921 survey which could mask a slight arching. However, one cannot be certain.

3. The raised beach criterion cannot be applied to this study at present. Raised glacial-like beaches do exist, and their profiles indicate post-glacial uplift to the north (e.g. Chapman, 1931). These beaches are thought to be greater than 9,000 years old (Denny, 1974) and, therefore, give no indication of whether the region has more recently reached post-glacial isostatic equilibrium.

4. Systematic error remains as the probable cause of the apparent discrepancy between the geodetic leveling and the lake gauge studies. The 1955 leveling survey of the National Geodetic Survey had closures, but the 1973 survey did not. (S. Holdahl, 1979, personal communication). In addition, information pertaining to possible sources of systematic error, such as those described by Bomford (1971, p. 239-243), are not available for these surveys.

We conclude from the foregoing that the water level data are reliable, and that the most likely cause of disagreement between the two studies of vertical movement is the accumulation of systematic error over long distances (i.e between Saratoga Springs and Rouses Point) in one or more of the leveling surveys. We will, therefore, use the water level gauge profile as a base line to "calibrate" the relative movement profile derived from releveling. We, thus, abandon the original arbitrary assumption of no vertical change between levelings at Saratoga Springs that is inherent in Figure 2, and rotate the releveling profile so that it fits the lake level profile as shown in Figure 8. In this manner the water level gauge profile is used to provide long-distance control, and the releveling profile to define the more local crustal movements. The resulting combined profiles (Figure 8) suggest that the Champlain Valley consists of two segments with differing crustal movement: a northern portion (from Willsboro to Rouses Point) which exhibits a slight relative rise of about 0.7 millimeters per year to the north, and a southern segment (from Whitehall to Willsboro) which is undergoing arching up to 2 millimeters per year.

In summary, it appears that a slight relative rise (post-glacial rebound?) is occurring in the northern part of the basin, and an arching of about 2 millimeters per year is occurring along the Adirondack perimeter between Whitehall and Willsboro. This area of uplift corresponds exactly to the more elevated and breached part of the dome (Figure 1), suggesting that the present arching along the eastern Adirondacks represents the continuation of a more prolonged neotectonic uplift. It is noteworthy that the "calibrated" profile better fits the geologic-topographic data than did the original profile shown in Figure 2.

In view of a renewed appreciation of systematic leveling errors due to refraction (Holdahl, 1980, in press), it would be desirable to recheck the apparent vertical movements reported in this paper using elevation data that have received refraction correction. This will be done when such corrected data are made available by the National Geodetic Survey. It would also be desirable to resurvey the level line or, as an alternative, install a continuously recording water level gauge (anchored in and surveyed to bedrock) in the general area of Ticonderoga, New York to Westport, New York, where the uplift appears to be most rapid. If uplift is, in fact, as rapid as the relevling indicates, about ten years of water level data would display it.

ADDITIONAL VERTICAL MOVEMENT DATA FOR LAKE CHAMPLAIN

For the sake of presenting a slightly more regional picture, the two other water level gauges on Lake Champlain will be discussed, one in Phillipsburg, Quebec, and the other at Burlington, Vermont.

The gauge at Phillipsburg, Quebec is located just a few kilometers north of the international border (Figure 1). It was established in 1968, and is a continuously-recording type. It is located on a concrete and steel pier and, according to S. Huberman of Environment Canada, has been repeatedly surveyed to a benchmark on the foundation of a house about 122 meters away, with consistent results. The gauge and house are stable with respect to each other, but inasmuch as neither is on bedrock we only regard this gauge as probably stable.

The rate of change of annual water level differences between Phillipsburg and Rouses Point shows no statistically significant change over the period measured (Figure 9).

The gauge at Burlington, Vermont is about midway along the lake on the eastern shore (Figure 1). It was originally established in 1907 as a manual gauge and was located on the Lake Champlain Transportation Company dock at the foot of King Street. Comparisons of average gauge heights with those at Rouses Point and Whitehall indicate that between 1916 and 1938 the Burlington gauge was markedly unstable. In 1939 the gauge was moved to a position between the Burlington Electric Building and the Gulf Oil docks and has been of the continuously recording type. It has been surveyed periodically to a benchmark (Rm. #6) which is located 42 meters east and 6 meters north of the gauge house. Results of this surveying are given in Table 2, and show no independent motion of the gauge with respect to the benchmark. However, even though the gauge and its benchmark show no relative movement to each other, neither the gauge nor the benchmark to which it is tied is in bedrock. For this reason we regard this gauge as only probably stable. The rate of change of annual water level differences between Burlington and Rouses Point is shown in Figure 10; Burlington is undergoing relative subsidence at about 0.9 millimeters per year.

Table 3 shows the relative vertical movement rates between the water level gauges on Lake Champlain. They show no significant change between Rouses Point and Phillipsburg, a relative uplift at Rouses Point (and Phillipsburg) with respect to other gauges on the lake, and a relative subsidence at Burlington. In order to provide an areal view, the magnitude of relative crustal movements for various

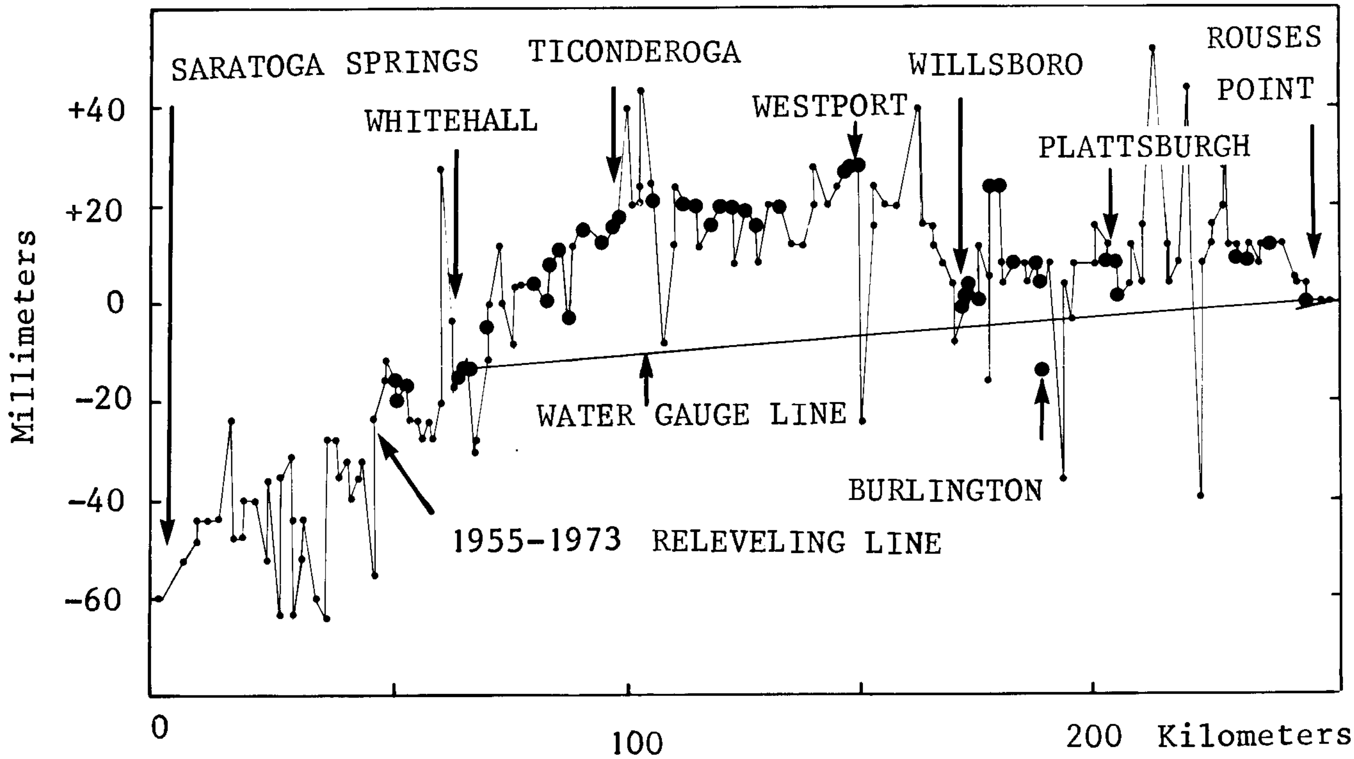

Figure 8. The 1955-1973 releveling line "calibrated" to the water level gauge profile line for the same period. The releveling line has been rotated so that the Whitehall-Rouses Point slope corresponds to that of the water gauge profile.

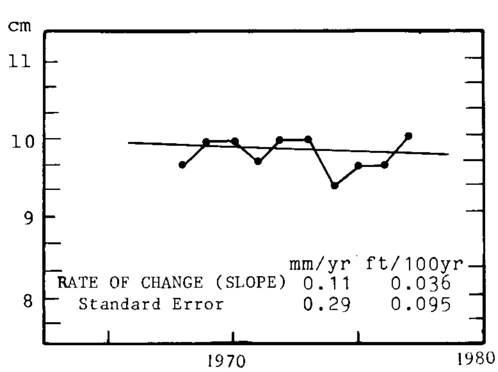

Figure 9. Rate of change of annual water level differences between Phillipsburg, Quebec and Rouses Point, New York, showing no change.

TABLE 2
ELEVATION DIFFERENCES (IN METERS) FROM THE BURLINGTON GAUGE TO BENCHMARK Rm. #6, LOCATED ENE OF GAUGE HOUSE

YEAR	METERS	YEAR	METERS
1946	4.051	1956	4.048
1948	4.050	1957	4.050
1949	4.050	1960	4.048
1950	4.050	1962	4.049
1951	4.049	1965	4.049
1952	4.050	1969	4.049
1953	4.050	1971	4.049
1954	4.049	1977	4.046
1955	4.047		

Source: U.S.G.S. Open Files

TABLE 3. LAKE CHAMPLAIN VERTICAL MOVEMENT RATES AS DETERMINED FROM WATER LEVEL GAUGE DATA

PERIOD	STATION	STATION	RATE* mm/yr	RATE* ft/100yr	STANDARD ERROR mm/yr	STANDARD ERROR ft/100yr	R	SIGNIFICANCE
1968-1977	Rouses Point	Phillipsburg	+0.11	+0.036	0.29	0.095	.157	.668
1940-1978	Rouses Point	Burlington	+0.92	+0.302	0.06	0.020	.906	.999
1940-1978	Rouses Point	Whitehall	+0.72	+0.235	0.20	0.065	.557	.999
1940-1978	Burlington	Whitehall	-0.20	-0.064	0.17	0.055	.201	.869

* Positive values indicate that first gauging station is rising with respect to second station

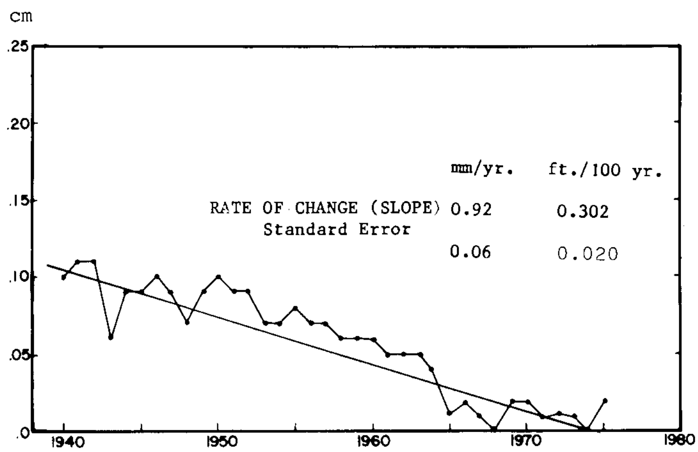

Figure 10. Rate of change of annual water level differences between Rouses Point and Burlington. Burlington is subsiding.

parts of the Champlain Valley, arbitrarily assuming no vertical movement at Whitehall, are summarized on Figure 1. If the relative subsidence at Burlington is real, it can easily be accounted for by slight movements on one or more of the numerous faults between the eastern and western shores of the lake shown on the Geologic Map of New York (Fisher and others, 1971).

Although we have indicated that measurable vertical crustal movements exist in the Champlain Valley, the magnitude of these movements is probably inconsequential from a basin management point of view. The largest differential movement which would likely occur between any two places on Lake Champlain (between Ticonderoga, New York and Burlington, Vermont) would be only about 170 millimeters in 100 years. Therefore, crustal movements will probably not create navigational problems, affect marina harbor depths, affect the siting of docks and piers, or necessitate redesignation of the 100 year flood plain level. Hence, vertical crustal movements will also probably not play a role in the International Joint Commission's decision on proposed regulation of Lake Champlain.

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THE STRATIGRAPHY OF UNCONSOLIDATED SEDIMENTS OF LAKE CHAMPLAIN

Allen S. Hunt
Department of Geology
University of Vermont
Burlington, Vermont 05405

INTRODUCTION

An understanding of the geological history of Lake Champlain has been growing since the early 1900's when field studies of the late glacial history of the Champlain Valley first began. The lake itself, however, received little attention from geologists until the middle 1960's, in spite of the fact that much of its history is recorded in its sediments. In addition to lake history, a host of contemporary geological problems, which range from sedimentation rates to neotectonism within the Champlain basin, can be studied using lake stratigraphy. This paper is a summary of current knowledge of Lake Champlain stratigraphy and biostratigraphic methods available for subdivision of its sedimentary units.

PREVIOUS STUDIES

The earliest contribution to an understanding of Lake Champlain stratigraphy came from bottom-sampling studies which investigated the physical and gross chemical properties of near-surface sediments. Much of this research was undertaken by faculty and students at the University of Vermont and Middlebury College. The work includes that of Millet (1967) on clay mineral compositions, Johnson (1969) on the composition and distribution of ferromanganese concretions, Townsend (1970) on heavy minerals in central Lake Champlain, Clement (1967) on bottom sediments in Malletts Bay, Fillon (1970) on sediments of Mississquoi Bay, Aubrey (1971) on sediments in southern Lake Champlain, Hunt (1971) who conducted a reconnaissance sediment study of the entire lake, the Vermont Department of Highways (1972) on sediments in Ticonderoga Bay, Calabi (1972) on sediments of the Otter Creek delta, Carroll (1972) on sediments in the area of Crown Point, Folger (1972) on sediments in Ticonderoga Bay, Gramsey (1973) on sediments from Button Bay to Chimney Point, Corliss and Hunt (1973) on St. Albans Bay and Mason and others (1977) on Ticonderoga Bay sediments. These investigations documented the distribution and composition of existing sediment types, and were useful in understanding the processes that influenced lake sedimentation.

In the late 1960's, a piston coring program was initiated with the objective of obtaining complete sequences of the unconsolidated sediments resting above bedrock. Early attempts met with limited success because in most areas the sediments were too thick to be penetrated with a piston corer. The problem was resolved by using sub-bottom profiling techniques to locate coring sites where units were sufficiently thin to allow the recovery of complete sequences. Investigations of the lake's physical stratigraphy, combining coring and geophysical methods, include the work of Chase and Hunt (1972) in central Lake Champlain and of Gramsey (1973) and Lynde (1975) in southern Lake Champlain. Other research involving the analysis of piston cores includes work on the paleomagnetic stratigraphy (Freed, Hunt and Fillon, 1975; March, 1975), on diatom biostratigraphy (Sherman, 1972), on benthonic foraminiferal biostratigraphy (Fillon and Hunt, 1974), and on ostracode biostratigraphy (Gunther and Hunt, 1976 and 1977).

High-resolution sub-bottom profiling techniques were first used to map the thickness and distribution of sedimentary strata in the early 1970's. Using a 3.5 khz "pinger", Chase and Hunt (1972) studied the aerial distribution of sedimentary units in central Lake Champlain and Lynde (1975) mapped sediments in southern Lake Champlain with an EGandG Uniboom system. The total thickness of sediments to bedrock was determined by Hunt (1976), Turner (1976) and Hunt (1977) for the central and northern portion of the lake using a Bolt pneumatic sound source.

LAKE STRATIGRAPHY

Chapman (1937) recognized three major episodes in the post-glacial history of Lake Champlain. In the earliest episode (13,200 to 12,800 YBP), glacial ice blocked the northern Champlain and St. Lawrence Valleys permitting a large proglacial lake to form which has been named Lake Vermont. As ice retreated farther north, marine waters entered the St. Lawrence and Champlain Valleys, forming an estuary called the Champlain Sea (12,800 to 10,200 YBP). A third and final episode (10,200 YBP to present) was initiated when a sill developed north of the international border cutting off marine waters which allowed the lake to freshen gradually, producing the present Lake Champlain.

The high-resolution sub-bottom profiling of Chase and Hunt (1972) and Lynde (1975) demonstrated the presence of three stratigraphic units separated by distinct acoustical boundaries. Biostratigraphic study revealed that the three units correspond to the three episodes recognized by Chapman. Much of the stratigraphic work done on cores since the discovery of these units has been directed to their further subdivision. Although numerous methods have been attempted, including the use of grain size, texture, color, moisture content, and pore water composition, to date only the microfossils have proven consistently useful. The lithologic and biostratigraphic characteristics of the three stratigraphic units are discussed below and a summary is presented in Table 1.

Lake Champlain unit

The Lake Champlain unit is quite variable in thickness, ranging from zero to a measured maximum of 24 meters. This unit is exposed on the lake bottom today where many of its facies can be readily studied. Lithologically, it contains four sediment types: gravels, sands, manganese nodules and organic muds (Hunt and Henson, 1969). Gravels make up less than 4 percent by area of the surface bottom sediments. Sands cover 22 percent of the lake bottom. Both gravel and sand occur primarily in shallow nearshore environments surrounding islands and at the mouths of rivers. These sediments are rarely encountered in piston cores, presumably because they represent shallow water facies.

The third sediment type, iron-manganese concretions, has been identified in seven areas of the lake (Johnson, 1969). Although they do occur in a pure state, the concretions typically are mixed with a terrigenous matrix constituting up to 90 percent of the sample. Individual concretions

Years B.P.	Major Sedimentary Unit	Forams	Ostracodes	Diatoms	Pollen	Other	LITHOLOGIC CHARACTERISTICS
		BIOSTRATIGRAPHIC CONTENT					
80 120 170	LAKE CHAMPLAIN			*Fragilaria capucina* and similar forms common *Cyclotella bodanica* and similar forms common	Hardwoods and Evergreens- *Ambrosia* common Hardwoods and Evergreens- *Ambrosia* rare	Anthro- pogenic debris Bryozoan statoblasts, spore cases, arthropod fragments, plant fibers	Olive gray to grayish brown organic muds which may contain sandy layers or black mottling
10,200	CHAMPLAIN SEA	*Elphidium clavatum* and *Protelphidium orbiculare* dominant *Islandella islandica* and *Islandella teretis* dominant	*Cytheropteron arcuatum* and *Cytheropteron vespertilio* common *Candona rawsoni*			pelecypods, gastropods, sponge spicules, otoliths; unidentified spines, egg cases, and fragments	Dark gray mud
11,800 13,000	LAKE VER- MONT				Evergreens- pollen rare	Plant debris	Top contains dark gray clay alternating with brown and gray layers

Table 1. Lithologic and biostratigraphic characteristics of post-glacial sediments in Lake Champlain. Dates for the major episodes based upon Connally and Sirkin (1971), Elson (1969) and Mott (1968).

vary in size from a few millimeters to ten centimeters and in shape from spherical to reniform or discoidal. The nodules occur primarily on shallow water platforms in water depths of less than 15 meters. Because the manganese concretions form in an oxidizing environment, they do not occur at sediment depths below about half a meter.

Organic muds are the deep-water facies of the sands and gravels. The sands make up 75 percent of the surface sediments and are the dominant sediment type of the Lake Champlain unit. Mineralogically, they are composed primarily of detrital quartz and clay minerals, but locally over 50 percent of the sediment consists of diatom frustules. The water-sediment interface of the organic muds is a grayish to reddish-brown hydrosal. Beneath the surface, the muds are dark gray to grayish-brown in color. The muds have a high organic content (10-20 percent). Although they are quite uniform in appearance, mottling and lamination does occur in some areas which is revealed by X-ray photographs of piston cores.

Biostratigraphic subdivision of the most-recently deposited sediments of the Lake Champlain unit has been achieved using pollen (Haworth, 1980; Hunt, 1979) and diatoms (Sherman, 1972). Based upon historical records of lumbering in the Champlain Valley, the increase in ragweed pollen (Ambrosia and related forms), which is a consequence of clear-cutting by man, was placed at 1810 by Haworth (1980). We are presently attempting to establish a pollen chronology that might be useful in subdividing earlier intervals of the Lake Champlain unit.

Sherman (1972), in a study of Lake Champlain diatom stratigraphy, recognized the increase of Fragilaria caprucina and other eutrophic indicator species to be a result of industrial development and population expansion which occurred about 1900. Although diatoms have been used primarily as a 1900 marker horizon, they are present in abundance throughout most of the Lake Champlain unit, and additional study may prove them more useful in biostratigraphic subdivision. Near the base, the Lake Champlain unit is devoid of diatoms. This "barren" zone rests above the first appearance of benthonic foraminifera in the Champlain Sea unit. Parker and Edgington (1976) noted a similar disappearance of diatoms with depth in freshwater Lake Michigan cores and attributed it to solution of the frustules. The disappearance in Lake Champlain sediments may be explained in a similar manner. It is also possible that the transitional marine-to-freshwater environment prevented immediate diatom repopulation.

In addition to pollen and diatoms, other fossils occur exclusively in the Lake Champlain unit which further study may prove to be useful in correlation. These include bryozoan statoblasts, arthropod fragments, spores and agglutinating protozoans. Anthropogenic materials, including wood chips, sawdust and steamboat cinders, which occur locally at the mouths of rivers and in harbors, have also been used for dating.

Champlain Sea unit

The boundary between the Lake Champlain unit and the Champlain Sea unit can be recognized acoustically as well as biostratigraphically. Acoustically, the boundary is characterized by a distinctive three-banded layering which appears on "pinger" profiles (Chase and Hunt, 1972). Biostratigraphically, the boundary can be recognized by the first appearance of benthonic foraminifera, ostracodes and other marine fossils. Lithologically, the unit consists of unlaminated dark gray muds of low organic content (less than five percent). Based upon seismic data, the unit is known to have a maximum thickness of about 15 meters and typically is between 7 and 14 meters thick.

Although over 50 benthonic foraminifera and 25 ostracode species have been recognized in Champlain Valley sediments of New York and Quebec (Cronin, 1976), only a few forms are abundant in the Champlain basin. Our work on several dozen cores from central and northern Lake Champlain indicates that a complex pattern of biogeographic and biostratigraphic variation exists, which presumably is a consequence of the many local differences inherent in an estuarine environment. A few generalizations, however, can be made. The youngest fauna encountered is characterized by low species diversity and is dominated by the foraminifera Elphidium clavatum and Protelphidium orbiculare as well as species of the ostracode genus Cytheropteron, suggesting shallow water, low salinity conditions (Walton, 1964). In older Champlain Sea sediments, species diversity increases and the foraminifera Islandella islandica and Islandella teretris, become the dominant forms indicating that the water was colder, deeper and of more nearly normal salinity (Fillon and Hunt, 1974).

The freshwater ostracode species, Candona rawsoni and Candona subtriangulata are present only near the base of the Champlain Sea unit. These species occur in separate horizons and are not found with foraminifera, indicating that freshwater influxes interrupted the early marine invasion of Lake Vermont as was suggested by Cronin (1976). The freshwater horizons vary in number from none to several with no apparent geographic consistency, suggesing that influxes were local in extent.

Lake Vermont unit

The maximum thickness of the Lake Vermont unit is not known because seismic data does not, in all instances, permit us to discriminate between these sediments and glacial till. The unit has a known thickness in excess of 100 meters and its thickness may exceed 200 meters. Because of this great thickness and depth of burial beneath the younger Lake Champlain and Champlin Sea sediments, only the top portion of the unit has been recovered in piston cores. The youngest Lake Vermont deposits typically are composed of dark gray or brown clay layers a few centimeters in thickness, alternating with light gray layers. These layers are probably varves and may represent annual sediment deposition. Elsewhere, the Lake Vermont clays are homogenous or without evident bedding and may contain a small fraction of gravel or sand-sized material. The Lake Vermont clays are low in organic content, typically, containing no more than a few percent.

The Lake Vermont unit has not yielded appreciable organic remains useful for biostratigraphic subdivision. This paucity of fossils may be explained by the high sedimentation rate and low nutrient level associated with proglacial lakes. Evergreen pollen does occur sparsely and plant debris is relatively common. Although diatoms have been reported from Lake Vermont sediments (Gramsey, 1973), we have not encountered them and, if present, they are rare.

CONCLUSIONS

Seismic work using a 3.5 khz transducer has permitted subdivision of the sedimentary units deposited during the three major episodes recognized by Chapman (viz. Lake Champlain, Champlain Sea and Lake Vermont). Coring has demonstrated that the acoustical boundaries show close correspondence to biostratigraphic boundaries. Additional subdivision of the lake sediments deposited during the three major episodes is possible using diatoms, pollen, foaminifera and ostracodes. Future study of pollen, diatoms and other organic remains should allow even finer definition of the time interval spanned and provide further information on the history of Lake Champlain.

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ALKALIC DIKES OF THE LAKE CHAMPLAIN VALLEY

J. Gregory McHone
Department of Geology
Indiana University - Purdue University
at Indianapolis
Indianapolis, Indiana 46202

E. Stanley Corneille
4 Terrace Street
Randolph, Vermont 05060

ABSTRACT

Over 250 Early Cretaceous lamprophyre and trachyte dikes have been mapped in the central Lake Champlain Valley of Vermont and New York. Two regional subdivisions are recognized: a smaller swarm of about 80 lamprophyre (mostly monchiquite) dikes in the Plattsburgh-South Hero-Milton area, and a larger swarm of over 150 monchiquite, camptonite and trachyte (mostly bostonite) dikes in the Willsboro-Shelburne-Charlotte area. Most of the dikes trend approximately east-west, but more diverse orientations are found near the syenitic Barber Hill stock in Charlotte, Vermont. The lamprophyres have basanitic compositions and mineralogies which include augite, kaersutite, olivine, phlogopite, andesine and analcime. The trachytes are mainly anorthoclase, with minor quartz, biotite and microcline. A new Rb-Sr age for seven trachyte dikes of 125 ± 5 m.y. enhances a cogenetic model for all the dikes. The Champlain magmas are part of the Monteregian-White Mountain igneous province, and may represent a mafic-felsic pair formed by an immiscible liquid mechanism from mantle-derived camptonitic magma.

INTRODUCTION

Mafic and felsic intrusives in the central Lake Champlain Valley of Vermont and New York have attracted attention for many years. Previous work includes papers by Thompson (1861), Kemp and Marsters (1893), Alling (1927), Hudson and Cushing (1931), Hawley (1956), Migliori (1959), Dimon (1962), Gillespie (1970) and Laurent and Pierson (1973). Our thesis work at the University of Vermont (Corneille, 1975; McHone, 1975) concentrated on petrographic and geochemical descriptions of the dike rocks. Our studies have been integrated with related work on the Mesozoic magmas of New England and Quebec (McHone and Butler, 1978). In this paper we shall describe the physical and chemical characteristics of the Champlain dikes, and follow with a discussion of petrogenetic models.

DESCRIPTION OF DIKES

Appearance

The lamprophyre dikes are fine-grained, dark gray rocks, with rusty-brown weathered surfaces. Fracture sets within the rocks are usually developed both perpendicular and parallel to the dike walls. Glassy chill margins are ubiquitous. Elongate black phenocrysts of amphibole and stubby pyroxene prisms are often visible, even without a hand lens. Most lamprophyres contain white amygdules in flow streams parallel to the dike contacts. Pink to gray ellipsoidal bodies 0.5 to 1 cm across, called ocelli, are found in many of the lamprophyres. Xenoliths have been noted in 22 examples, usually as small, rounded pieces of Paleozoic shales and carbonates, and Precambrian gneisses, marbles and anorthosite from the underlying Grenville basement. The trachyte dikes are buff to reddish-tan colored rocks, most with visible phenocrysts of alkali feldspar. Fifteen trachyte dikes contain xenoliths, usually larger and more angular than the inclusions found in lamprophyres but with similar lithologies. Several trachyte "breccia dikes" exposed on Shelburne Point have more than 90 percent xenoliths by volume (Hawley, 1956).

Distribution

Figure 1 shows the locations for most of the Mesozoic intrusions known in the Champlain Valley. The majority of the dikes are exposed along the bedrock shorelines of Lake Champlain, but many are also found in roadcuts and quarries away from the lake. A few sills are also known, including a large felsic intrusion in Willsboro, New York called the Cannon Point sill (Fig. 1). Few dikes can be traced far because of surficial cover, but Fisher (1968, p. 33) projects lengths up to a mile for dikes in the Plattsburgh area. Lamprophyre dikes may generally have width-to-length ratios of over 1:1,000 (McHone, 1978b).

The group of dikes southeast of Plattsburgh (Fig. 1) is separated from the dikes south of Burlington by barren shorelines in New York, and by a lack of dikes at North Burlington and Colchester Point, Vermont. Some dikes may be covered by the Winooski River delta. The Plattsburgh anomaly (Fig. 1) is most likely a magnetic and gravity expression of a shallow, unexposed mafic stock similar to several of the Monteregian plutons in Quebec (Diment, 1968). The northern dikes do not encompass the anomaly. No dikes have been reported along the rocky shores of Lake Champlain north of Grand Isle and Milton, Vermont, and dikes are very rare south of Vergennes, Vermont.

Two-thirds or more of the northern subgroup of dikes are monchiquite. Trachyte dikes are confined to the southern subgroup, being most abundant in the Shelburne and Charlotte areas of Vermont. Charlotte also contains the Barber Hill stock, a small syenitic complex (Fig. 1). Camptonite dikes are present throughout both subgroups, and are also found westward in the Adirondacks (Jaffe, 1953) and eastward through Vermont and much of New England (McHone, 1978b).

Orientation and size

Over 90 percent of the lamprophyre dikes trend within 20 degrees of east-west (Fig. 1), intruding a major fracture set in the Champlain Valley (Stanley, 1974). Whether the fractures are coeval with the dikes or predate them, this trend preference implies a north-south "least compression", or extensional orientation of crustal stresses in Early Cretaceous time. Tectonic implications of dike-stress relationships in New England have been discussed elsewhere (McHone, 1978a and 1978b). The trachyte dikes also tend to be east-west, but in addition many have orientations radial to the Barber Hill stock (Fig. 1), as noted by Gillespie (1970).

The great majority of the lamprophyre dikes are between 20 cm and 150 cm wide, while 1 to 3 m is more typical of the trachyte dikes. Apparently the mafic magmas, being less viscous, were able to flow more readily than the felsic magmas and are thus more likely to occur as narrow fracture fillings.

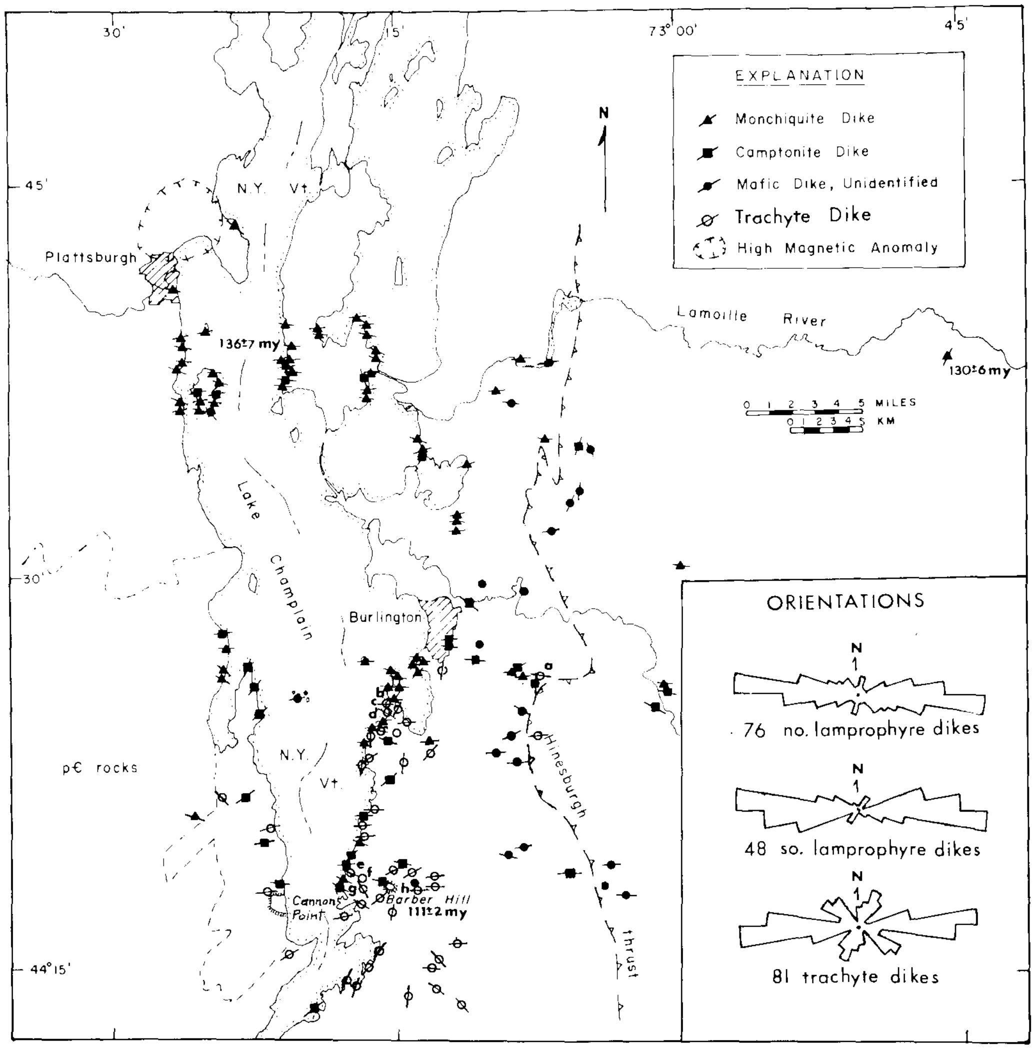

Figure 1. Alkalic dikes of the Lake Champlain Valley, New York and Vermont.

AGES OF DIKES

Zartman and others (1967) reported a K-Ar date of 136 ± 7 m.y. for biotite separated from a monchiquite dike on the western shore of South Hero, Vermont (Fig. 1). To the east, kaersutite (a basaltic hornblende) separated from a monchiquite dike near the Johnson-Cambridge town line (Fig. 1) yielded a similar date of 130 ± 6 m.y. (McHone, 1975). These dates are slightly younger than the 141 m.y. time boundary for the beginning of the Cretaceous Period (Van Eysinga, 1975). In southwestern Vermont, Zen (1972) reported an age of 105 ± 4 m.y. for a camptonite dike in West Rutland, Vermont (K-Ar on hornblende). Although younger than the Champlain dates, the Zen date falls within the 100 to 120 m.y. range of the Monteregian plutons (Philpotts, 1974) and of the younger group of alkalic plutons of the White Mountain magma series (Foland and Faul, 1977). Gillespie (1970) separated "slightly chloritized" biotite from syenite of the Barber Hill stock in Charlotte, Vermont (Fig. 1) which was dated by Armstrong and Stump (1971) at 111 ± 2 m.y., similar to the Monteregian ages.

Two monchiquite dikes are crosscut by trachyte dikes on the Shelburne lakeshore. If the trachyte dikes are cogenetic with the Barber Hill stock as suggested by Gillespie (1970), it would seem that the trachytes postdate the Champlain lamprophyres by around 20 m.y. However, both Welby (1961, p. 189) and Dimon (1962) report that a lamprophyre dike crosscuts the western side of Barber Hill. We could not locate this site during our field studies.

In order to help clarify the age relationships, eight trachyte dikes were selected for a whole-rock Rb-Sr isotopic age determination (sites a through h, Fig. 1). Four dikes were chosen from the Shelburne-South Burlington area, and four others were from the Barber Hill area of Charlotte. The results, listed in Table 1, form a reasonably good isochron except for sample "e" (Corneille's SPS-63) which has an anomalously high Sr^{87}/Sr^{86} value relative to its Rb^{87}/Sr^{86} (Fig. 2). This dike contains a variety of xenoliths that probably contaminated its chemistry, and was therefore disregarded in the isochron calculations.

The Rb-Sr date of 125 ± 5 m.y. for the trachyte is closer to the age of the lamprophyres than to Barber Hill. It seems likely to us that the 111 m.y. date for Barber Hill is erroneously young, possibly because of argon loss from the mica due either to alteration or to a prolonged cooling of the stock. If so, all of the lamprophyres, trachytes and syenites of the Champlain Valley may actually have intruded about 130 m.y. ago, satisfying both the geological relationships found by Gillespie (1970), and the crosscutting data.

TABLE 1. RB-SR ISOTOPIC DATA FOR CHAMPLAIN TRACHYTE DIKES.

Sample no.	Rb (ppm)	Sr (ppm)	Rb/Sr	Rb87/Sr86	(Sr87/Sr86)N
a (MWB)	229	47	4.896	14.2080	0.7326
b (SP41)	206	30	6.781	19.6990	0.7441
c (SP45)	103	1189	0.087	0.2508	0.7088
d (SP43)	226	33	6.868	19.9480	0.7410
e (SPS63)	127	73	1.741	5.0477	0.7235
f (SPS61)	95	438	0.217	0.6273	0.7090
g (SPS71)	251	358	0.701	2.0290	0.7108
h (MPMB)	43	205	0.210	0.6070	0.7078

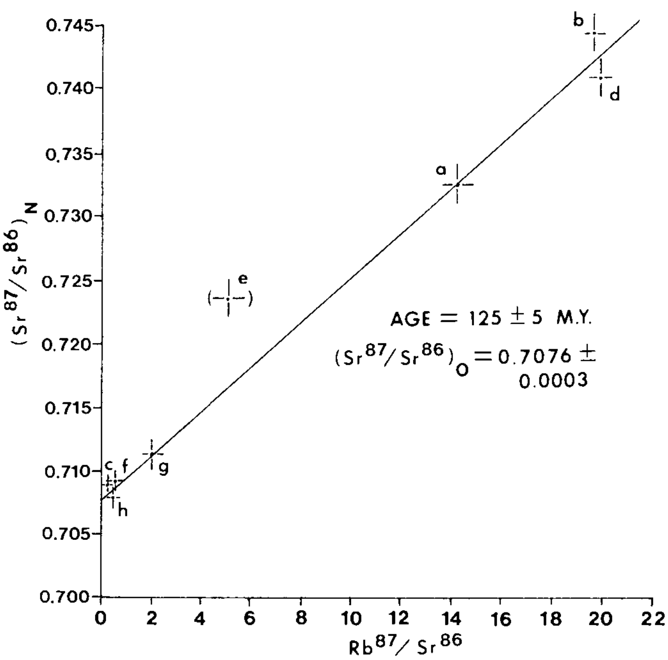

Figure 2. Rb-Sr isochron for Champlain Valley trachyte dikes.

PETROGRAPHY

Thin-section studies of the lamprophyres reveal microporphyritic textures in which both phenocrysts and matrix minerals tend to be euhedral (panidiomorphic). Both camptonite and monchiquite have similar mineralogies, except that monchiquite contains little or no feldspar, while camptonite has abundant andesine or labradorite in its groundmass. In extreme examples of monchiquite, mafic minerals (augite, olivine and minor calcite) constitute over 90 percent of the rock (dikes MT-1 and MT-4, Table 3). All phenocrysts are mafic minerals, usually augite or kaersutite (titaniferous hornblende). Optical data for the phenocrysts are listed in Table 2, and mineral chemistries will be discussed in the next section. Zoning is pronounced in the augites, with reddish Ti-rich rims, and rarely, green Na-rich cores. Augite, kaersutite and phlogopite are common groundmass minerals, in addition to lesser amounts of plagioclase, analcime, calcite, titaniferous magnetite and alkali feldspars. The ocelli have lobate boundaries which are crosscut by some kaersutite phenocrysts, and are rich in alkali feldspars, plagioclase, kaersutite, apatite, calcite and zeolites. Studies by Philpotts (1976) and others contend that ocelli are separated as immisible liquids in the lamprophyric magmas, based

TABLE 2. OPTICAL DATA FOR DIKE PHENOCRYSTS

Dike type	Optical Data* (degrees)	Mineral
Trachyte	2Vx = 46 - 58	Anorthoclase
	2Vx = 58 - 60	Albite
	2Vx = 74 - 88	Microperthite
Camptonite	2Vz = 84 - 88	Olivine
	2Vz = 54 - 70	Augite
	Z∧c = 32 - 52	Augite
	2Vz = 44 - 48	Titanaugite
	2Vz = 70 - 85	Aegirine-augite
	2Vx = 66 - 82	Kaersutite
	Z∧c = 9 - 19	Kaersutite
	Z∧c = 17 - 36	Plagioclase
Monchiquite	2Vz = 84 - 88	Olivine
	2Vz = 50 - 70	Augite
	Z∧c = 35 - 48	Augite
	2Vz = 30 - 55	Titanaugite
	2Vx = 66 - 85	Kaersutite
	Z∧c = 4 - 15	Kaersutite

*accuracy ± 2%

TABLE 3. GEOCHEMICAL DATA FOR CHAMPLAIN DIKE ROCKS AND MINERALS.

	1	2	3	4	5	6	7	8	9	10
SiO_2	38.0	35.4	36.85	36.20	46.0	44.57	46.9	41.2	62.1	69.0
TiO_2	2.97	3.80	1.78	2.25	1.83	2.16	2.40	2.01	0.45	0.24
Al_2O_3	11.0	11.7	9.78	11.87	13.6	13.71	14.2	12.1	19.5	16.5
Fe_2O_3	6.41	12.0*	11.75*	13.44*	5.67	2.30	4.44	1.62	2.4*	2.5*
FeO	6.92	-	-	-	6.88	7.18	4.87	8.71	-	-
MnO	0.22	0.26	0.18	0.26	0.19	0.18	0.13	0.16	0.21	0.20
MgO	11.6	5.97	11.59	8.24	7.28	8.39	4.58	8.71	0.16	0.06
CaO	12.8	18.0	15.57	15.11	10.7	10.44	8.00	11.4	0.18	0.50
Na_2O	1.81	3.31	2.59	2.04	2.81	2.94	3.68	2.38	6.5	4.3
K_2O	2.00	1.49	1.96	2.42	1.51	2.67	2.79	0.76	6.6	4.0
P_2O_5	0.94	1.19	1.26	2.68	0.56	0.55	0.60	0.45	0.07	0.10
H_2O^+	2.45	1.85	0.98	2.00	0.86	0.90	1.41	4.45	1.2	3.0
CO_2	1.84	4.84	6.61	2.62	1.35	3.41	4.89	7.70	-	-
H_2O^-	0.50	0.49	0.45	0.94	1.19	0.25	0.80	1.31	-	-
Total	99.46	100.28	100.46	100.07	99.43	99.65	99.69	100.27	99.37	100.40
Rb	54	50	67.6	77.4	41	79.6	61	11	229	43
Sr	1257	1443	1480	2030	830	1130	857	733	47	205
Y	-	-	34.2	53.7	-	25.2	-	-	-	-
Zr	335	492	272	254	279	230	471	192	-	-
V	-	-	225	254	-	189	-	-	-	-
Cr	-	-	331	111	-	444	-	-	-	-
Ni	-	-	258	101	-	168	-	-	-	-
Ba	1060	710	1950	1980	800	1150	943	665	-	-
Sr87/86	0.7040	0.7042	-	-	-	0.7046	-	0.7047	0.7326	0.7078

	11	12	13	14	15	16	17	18	19	20
SiO_2	41.24	39.21	35.57	43.12	45.87	39.66	50.18	52.51	34.39	55.32
TiO_2	0.22	5.36	4.54	2.30	2.63	2.52	0.96	0.02	0.11	0.15
Al_2O_3	0.00	12.70	16.38	16.77	6.90	14.15	7.66	24.91	15.87	28.08
FeO*	10.14	11.93	8.74	16.37	7.03	13.58	5.42	0.03	13.88	0.70
MnO	0.26	0.23	0.18	0.30	0.17	0.25	0.04	0.12	0.17	0.01
MgO	47.20	11.87	18.18	5.58	12.55	4.93	12.63	0.00	22.43	0.06
CaO	0.20	11.81	0.23	0.91	23.84	23.71	21.63	0.72	0.51	10.48
Na_2O	0.06	2.53	1.06	0.31	0.43	0.73	1.60	12.78	0.11	4.66
K_2O	0.00	1.38	8.45	6.48	0.00	0.02	0.00	0.18	0.32	0.38
Total	99.32	97.02	93.34	91.15	99.41	99.56	100.12	90.54	87.79	99.84

*total iron

Note: volatiles not analyzed in microprobe analyses.

1 - Dike BU-2 (Monchiquite)
2 - " MM-1 "
3 - " MT-1 "
4 - " MT-4 "
5 - " BU-4 (Camptonite)
6 - " BU-8 "
7 - " BU-15 "
8 - PL-2 (Camptonite)
9 - " MWB (Trachyte)
10 - " MPMB "
11 - Olivine (MT-2)
12 - Kaersutite (BU-1)
13 - Phlogopite (BU-2)
14 - Biotite (MT-2)
15 - Augite (BU-2)
16 - " (MT-2)
17 - " (BU-2)
18 - Analcime (BU-1)
19 - Chlorite (BU-2)
20 - Feldspar (WR-8)

on their high-temperature mineralogy and forms indicative of liquid-liquid interfaces. Amygdules often coexist with the ocelli but are clearly distinguishable, being filled with calcite, zeolites, and rarely quartz; all are low-temperature infillings of former gas bubbles.

The trachyte dikes have traditionally been called bostonite, a term which was originally applied to non-porphyritic rocks with alkali feldspars texturally grouped into oriented "clumps" of grains. Many Champlain examples are porphyritic rocks with trachytic flow textures, and so the broader term trachyte is more appropriate for all of the felsitic dikes. Anorthoclase predominates both as phenocrysts and matrix grains, but quartz, biotite and microcline can also be present in minor amounts. The trachyte matrix is often partly altered, with carbonates and clay minerals obscuring the textures. Optical data for the trachyte minerals are listed in Table 2.

GEOCHEMISTRY

Selected whole-rock and mineral analyses of Champlain dikes are presented in Table 3. Analytical procedures are described elsewhere (McHone, 1978a). In general, the lamprophyres display very high volatile (CO_2, H_2O, P_2O), strontium and barium contents. The monchiquites reflect their feldspar-poor and augite-rich mineralogies with low silica and aluminum and high calcium values. The lamprophyres also tend to have high ferric/ferrous ratios, due to high oxygen fugacities during crystallization. Despite the high volatile contents, the lamprophyres show little alteration in thin section, except that olivine is generally replaced by calcite or serpentine minerals. The CO_2 and H_2O are present mainly in the primary micas and amphiboles, and in late-stage zeolites and calcite in the groundmass and amygdules. The trachyte analyses (nos. 9 and 10, Table 3), although incomplete, must be closely similar to the chemistries of their principal mineral component, alkali feldspar. Barber Hill syenites which were analyzed by Laurent and Pierson (1973) are chemically indistinguishable from the trachytes.

The forsteritic olivine analysis (no. 11, Table 3) is typical for most of the original lamprophyric olivines. Kaersutite (no. 12, Table 3) is distinguished from other pargasitic amphiboles by its high titanium content; most alkalic lamprophyres, including the type camptonite at Livermore Falls, New Hampshire, contain at least minor amounts of this amphibole. The chemistry of kaersutite is remarkably like many lamprophyre whole-rock analyses (compare nos. 1 and 12, Table 3). Augite chemistries (nos. 15, 16 and 17, Table 3) vary considerably in Champlain lamprophyre rocks, mainly a result of concentric zoning in the phenocrysts. Formula calculations indicate high Fe^{+3} contents, and the presence of ferric iron and titanium is coupled with charge-balancing substitutions in the crystal sites; mainly Al for Si, Fe^{+2} and Mg; and Na for Ca. The low silica contents of the magmas may promote such substitutions.

The lamprophyre and mineral analyses of Table 3 are similar to rock and mineral analyses of many alkalic basalts, in particular basanites and nephelinites, from both oceanic and continental volcanic suites (see summary values by Carmichael and others, 1974, and LeMaitre, 1976). However, most chemical analyses published for alkalic basalts show very low CO_2 and H_2O contents, in contrast with the lamprophyres. Much of this difference could be due to low-pressure degassing of magmas in the volcanic extrusions, the most commonly studied regime of basaltic rocks. Unlike volcanics, the dikes preserved volatiles because they crystallized under confinement, sealed by chilled, glassy margins. The effect of these primary volatiles was to delay the crystallization of feldspar while encouraging the growth of H_2O and CO_2-rich minerals. Lamprophyres are therefore members or at least equivalents of the alkali olivine basalt series, a fact not emphasized by the literature.

PETROGENETIC MODELS

The trachyte dikes of the Champlain Valley are closely associated with the Barber Hill stock, and are chemically similar. In general, trachytes world-wide are believed to be offshoots from syenitic magmas, either from volcanic complexes or from alkalic plutons (often radiating from the parent body). The origin of trachyte magmas, then, is only a small step from the origin of syenites in alkalic complexes.

Lamprophyre dikes are most commonly identified in continental areas, and are enriched in lithophilic elements like alkalies and volatiles. Along with a supposed association with granites (see Hyndman, 1972, p. 192), this relationship has led many geologists to speculate that lamprophyres were originally "normal" magmas which became contaminated through assimilation or leaching of granitic rocks. In fact, however, only the felsic varieties of lamprophyres (spessartite, kersantite, minette) are often associated with granites. The alkalic, more mafic lamprophyre called camptonite is found either in independent regional dike swarms or with some general overlap of dikes with alkalic pluton distributions. It is significant that monchiquite, despite petrologic characteristics which appear to be gradational with camptonite, is never abundant except when near a gabbro-rich pluton. As discussed earlier, the alkalic lamprophyres are mineralogically, chemically and isotopically similar to many oceanic intraplate basalts (basanite and nephelinite). We believe their origins are also similar; that is, derivation by partial melting of relatively "undepleted" mantle rocks followed by magmatic differentiation.

In petrology, the "Daly gap" refers to the paucity of rock types intermediate between rhyolite and basalt in many alkalic volcanic suites. This bimodal distribution is not easily explained using schemes of differentiation by crystal fractionation. The Daly gap is evident in the Cretaceous rocks of the Monteregian and White Mountan magma series (McHone and Butler, 1978), and is reflected in the Champlain region by the presence of only trachyte and lamprophyre dike types. Yoder (1973) suggested a model in which trachyte is fractionally melted from a parental material (the mantle?) at one "invariant point", followed by basalt at a later invariant stage. One problem with his model is the sequence of magmas, for many volcanic complexes have the contrary order of basaltic before trachytic eruptions.

Currie (1972) and Philpotts (1974) have proposed that liquid immiscibility is a major mechanism for the Daly gap in the Monteregian plutonic rocks. The production of ocelli in lamprophyres is critical evidence for this process (Philpotts, 1976). According to the scheme, a lamprophyric magma would split much more completely while cooling slowly in a large chamber than it can in a rapidly cooled dike. The felsic phase may then separate as an independent pluton, or remain associated with its mafic complement. Fractional crystallization and/or wall-rock assimilation within each magma body could produce the other varieties of observed rock types. Chemical studies and other evaluations of this model are still in progress, but we favor it as a reasonable if unorthodox explanation.

In the Champlain Valley and elsewhere, ocelli are much more common in camptonite than in monchiquite. Camptonite is also compositionally intermediate between trachyte and monchiquite, which we therefore suggest are derived from the immiscible separates. If camptonite represents a "primitive" magma as generated in the upper mantle, the occurrence of camptonite in wide-ranging dike swarms independent of plutons makes an origin as plutonic offshoots unlikely. Monchiquite dikes may be expected to be adjacent to a cogenetic alkali gabbro pluton, possibly indicated in our area by the Plattsburgh anomaly, since they represent a plutonic differentiate. Finally, trachyte dikes, formed by magmas more viscous than those which make up lamprophyre dikes, can never be far from a syenite body. In this area, the Barber Hill stock forms one source, while perhaps another may be expected to be present beneath Shelburne Point.

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MESOZOIC FAULTS AND THEIR ENVIRONMENTAL SIGNIFICANCE IN WESTERN VERMONT

Rolfe S. Stanley
Department of Geology
University of Vermont
Burlington, Vermont 05405

ABSTRACT

A 50 km long system of north-trending, remarkably linear faults of small to moderate displacement cut the folded Paleozoic carbonates of the Champlain plate into horsts and grabens. Similar structures are probably expressed in bedrock beneath Lake Champlain.

The St. George fault, the longest fault in the system, extends through kaolin mines and goethite-manganite deposits several km south of East Monkton, offsets major folds in an apparent sense by as much as 1 km in the eastern part of the Hinesburg synclinorium, and cuts the western edge of the Hinesburg thrust near St. George and Colchester Pond. Associated minor faults, slickensides and right-lateral offset of stratigraphic units indicate down-to-the-east displacement of 500 m.

East of Colchester Pond, the Indian Brook fault is also downthrown to the east and may well extend 30 km southward through Essex Junction to Lake Iroquois.

Four km north of Monkton Ridge, the St. George fault branches to the southwest forming the west-dipping Monkton fault which was originally interpreted by Cady (1945) as a low-angle thrust. Stratigraphic offset, fracture fabrics, and dip slickensides on the exposed fault surface indicate 850 m displacement. Numerous small cross faults of normal and possible strike-slip movement of less than 100 m cut the north-trending system and probably inhibited subsequent movement.

This regional system of high-angle faults developed during early Mesozoic extension since they cut compressional structures of western Vermont and are transected in several places by lamprophyre dikes of presumed Early Cretaceous age. Post-dike fracture systems are well documented in the Burlington area but definitive evidence for younger faulting has not been recognized yet. The fault system is important in controlling mineral deposits and groundwater resources. The potential for earthquakes on the high-angle system appears low because the inferred resolved shear stress is small and it is locked by discontinuous cross faults. Seismic potential on the Champlain thrust beneath the Green Mountains could well be a far more significant hazard if the east-west compressive stress characteristic of eastern New York extends into Vermont.

INTRODUCTION

Low- and high-angle faults have long been recognized as an important structure in the Champlain basin (Doll and others, 1961; Isachsen and McKendree, 1977; Fig. 1). The Champlain and Hinesburg thrusts and those of the Taconic allochthons are important regional low-angle thrusts. Regionally, extensive high-angle faults are well known west of the Champlain thrust where they strike northeasterly and generally are downthrown to the east forming a step-like sequence from the core of the Adirondacks to the shores of Lake Champlain. In fact, they may be present in the bedrock beneath Lake Champlain as they are to the south in Lake George (Hunt, 1979, personal commun.)

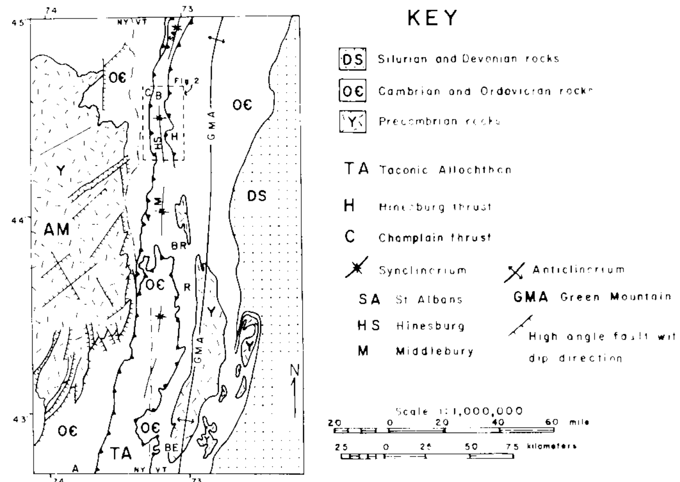

Figure 1. Simplified geological map of Vermont and eastern New York showing the distribution of the major areas of Precambrian (Y) rocks, Cambrian and Ordovician (OЄ) rocks and Silurian and Devonian (DS). Also shown are major high-angle faults of the eastern Adirondack massif (AM), the Champlain thrust, the Hinesburg thrust and the Taconic allochthons. Figure 2, the Hinesburg synclinorium, is located by dashed rectangle. Burlington B, Brandon Br, Rutland R, Bennington Be, and Albany A are shown.

In the last 10 years, seismicity in the northeastern United States has attracted considerable attention due to current and proposed construction of nuclear power plants, dams, and underground structures for storage or disposal of waste material (Fletcher and others, 1978). The extensive network of sensitive instruments in New York and New England that has been active during this time can now detect and locate earthquakes of magnitude 4 or less. These data combined with older information indicate considerable seismic activity in the northeast particularly in the St. Lawrence lowlands, the Adirondacks of northern New York and the Boston area of Massachusetts. Sykes and Sbar (1973) have suggested that these areas are part of a northwest-trending belt of earthquakes extending from Boston to Ottawa, Canada (Boston-Ottawa Seismic Zone). This belt passes directly through the Hinesburg synclinorium where approximately 25 percent of Vermont's population resides. Despite subdued seismic earthquake activity in west-central New Hampshire and Vermont (Fletcher and others, 1978), it is appropriate to examine the earthquake hazard of western Vermont in light of a recent discovery of a regionally extensive, high-angle fault system along the trace of the Hinesburg thrust in west-central Vermont (Figs. 1 and 2).

This north-trending system is here designated the St. George-Indian Brook-Monkton fault system (GIM). It has been mapped from Milton southward to Bristol for over a distance of 50 km. It is remarkably straight throughout much of its extent but

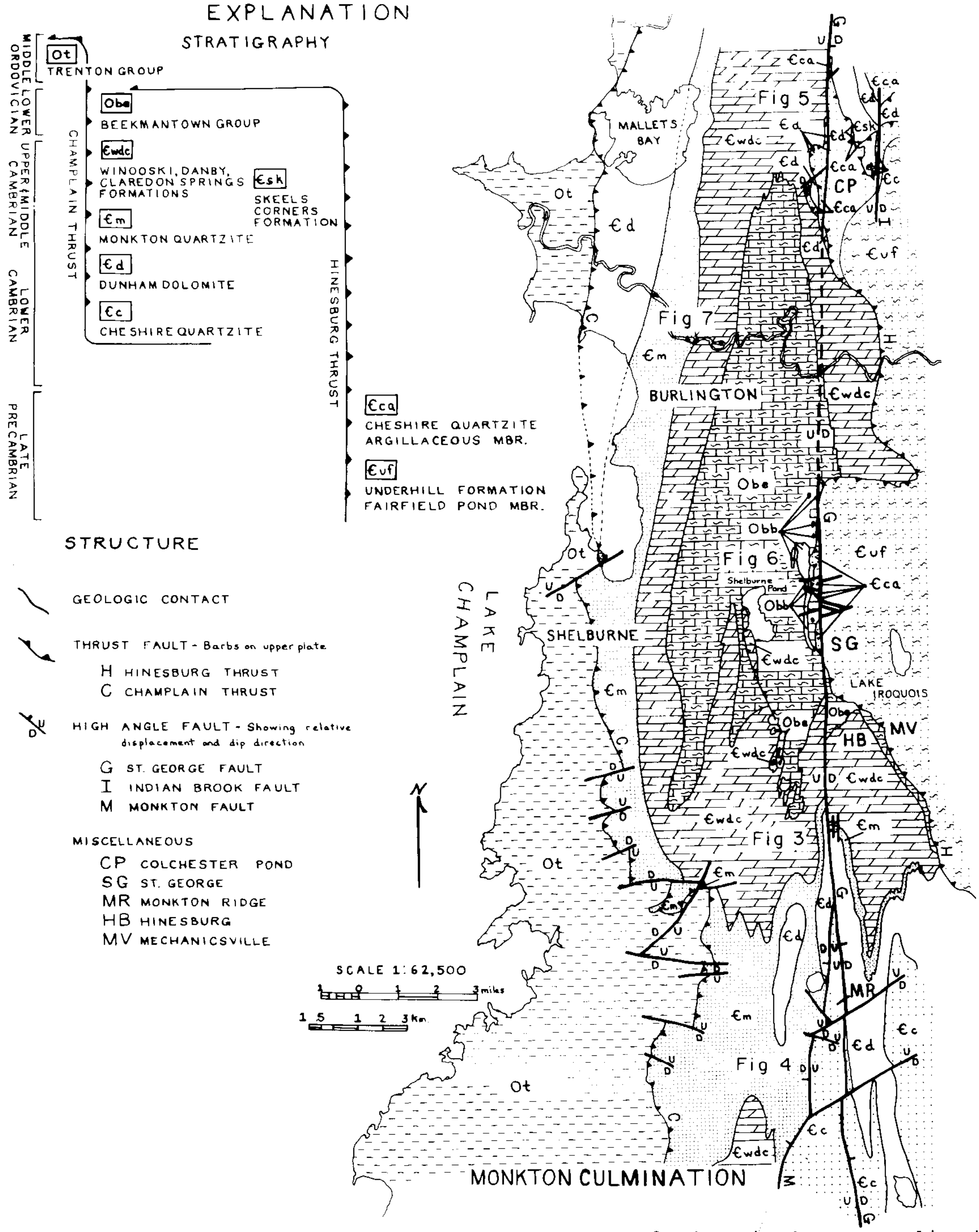

Figure 2. Generalized geologic map (1:62,500) of the Hinesburg synclinorium, west-central Vermont showing the St. George (G)-Indian Brook (I)-Monkton (M) fault system with associated cross faults in the Monkton Ridge area (MR). Champlain thrust (C) places Dunham Dolomite (€d) and Monkton Quartzite (€m) on shales of the Trenton Group (Ot) along Lake Champlain. Hinesburg thrust (H) to the east places argillaceous Cheshire Quartzite (€ca) and Underhill Formation (€uf) on carbonate rocks and quartzites of the Cheshire through Beekmantown interval. Obb are calcereous phyllite of the Brownell Mountain Member of the Bascom Formation in the Beekmantown Group. Details of the GIM system are shown in Figures 3-6.

it is offset by small cross faults in the latitude of the Monkton culmination. Dip-slip movement varies from 100 m for the Indian Brook fault to 500 m for the St. George fault to 850 m for the Monkton fault. The short, east-west faults which offset the GIM system are normal faults with 50 to 70 m displacement.

Although mapping has been confined to the Hinesburg synclinorium, there is evidence that it may be part of a larger system that extends just west of the Green Mountains to the Massachusetts state line (Fig. 1). South of Rutland, Vermont, a 6 km-long normal fault borders the eastern side of a ridge of Precambrian rocks in the Cambrian and Ordovician rocks of the Vermont valley (Doll and others, 1961). Its age is largely unknown. A similar fault of unknown extent cuts a lamprophyre dike dated at 105+m.y. in West Rutland (Zen, 1972, p. 2583). Both faults are downthrown to the east. Eleven km south of Bennington, Vermont, a 10 km long, west-dipping normal fault extends southward into Massachusetts

(Fig. 1). This fault, called the Reservoir Brook fault by MacFayden (1956, p. 48) is very similar to those of the St. George-Monkton system in that the adjacent rocks are brecciated and highly polished. Slickensided surfaces are numerous, and high-angle fracture fabrics are common. Displacement on this fault is estimated to be 750-900 m in the dip direction. Harwood and Zietz (1974, p. 182), however, estimate displacement in the order of 1900 m based on offset estimates of regional magnetic anomalies in the area. The kaolin deposits of the Bennington region are oriented along an elongate belt that overlies or is adjacent to the Cheshire Formation (Burt, 1927, Fig. 9). The two southern deposits coincide with the Reservoir Brook fault. Burt (1927, p. 66-67) shows that other deposits are present near the "Green Mountain Front Fault". Recent excavation along the Route 7 bypass north of Bennington uncovered a number of north-trending normal faults all with displacement less than 50 m. Two kaolin deposits occur just to the northeast. This fault zone lines up nicely with the Reservoir Brook fault. The association of kaolin deposits and iron-manganese deposits with the St. George-Monkton fault system suggest a similar structural control for the Bennington region as well as the other reported deposits in Brandon, Vermont. It appears, therefore, that the St. George-Monkton fault system is part of a larger system of normal faults that extends throughout the length of western Vermont.

The first part of this paper describes the structural geology of the GIM system in 4 recently mapped areas and the second part addresses the environmental significance of the system, with specific reference to mineral and water resources and the potential for seismic activity on this and other major faults in western Vermont.

STRUCTURAL GEOLOGY OF THE GIM FAULT SYSTEM

The evidence for the St. George-Indian Brook-Monkton fault system falls into three major catagories; stratigraphic (offsets and abrupt "pinch-outs"), structural (offset folds, fracture fabrics, breccia zones and exposed fault surfaces), and geomorphic (valleys, swamps and notches not controlled by a change in rock type). The evidence is best displayed where the rocks are quartzitic - it is most cryptic where the rocks are marble and shale.

The first area is located at the hinge of a major north-trending anticline about 1.6 km south of the town of Hinesburg (Figs. 2 and 3). Here a large swamp occupies the hinge region of the anticline. It is continuous with a major valley that extends northward to the fault-controlled valley at St. George to the north (Fig. 2). Excellent exposures of the Monkton Quartzite in the hills on either side of the swamp indicate a cumulative strike-slip separation of 1450 m of which 1100 is attributed to displacement on the master fault (Fig. 3). Parallel minor faults offset the eastern limb and indicate that the St. George fault zone is at least 500 m wide. Exposed dip-slip slickensides on the minor faults plunge 80 degrees due east. Assuming an inclination of 80 degrees due east for the major fault, a calculated net slip of 690 m is obtained. Five hundred meters of this total displacement is assigned to the master fault.

The second area is located south of the first zone near Monkton (Fig. 2). Four km north of Monkton Ridge, the Monkton fault branches southwestward from the St. George fault, isolating Monkton Ridge and the area to the south as a horst cored by Cheshire Quartzite (Figs. 2 and 4). Here the St. George-Monkton fault system enters the main region of the Monkton culmination where resistant quartzites of the Monkton and Cheshire formations are prevalent. The Monkton fault was first recognized by Keith (1932) as a thrust because of its irregular trace and the fact that the Dunham Dolomite "pinches out" between the stratigraphically older Cheshire Quartzite and the younger Monkton Quartzite just west of Monkton Ridge near Cedar Lake (Figs. 2 and 4).

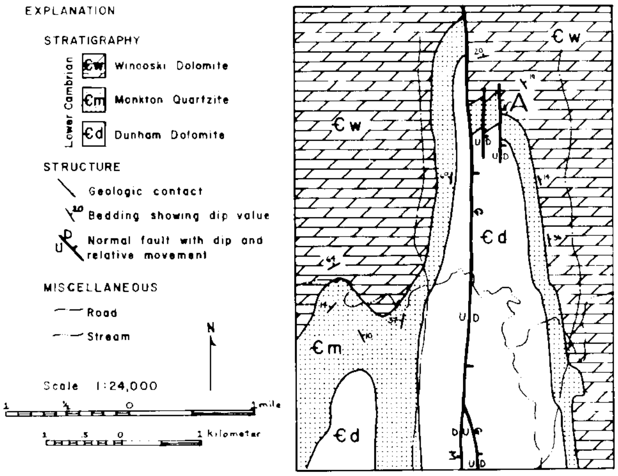

Figure 3. Geologic map of the Monkton anticline just north of Monkton Ridge (Fig. 1) showing the St. George fault (G) and the Monkton fault (M). Locality A is a lamprophyre dike cutting across high-angle fault. Based on unpublished work by P. Moreau, M. Crane and R. Stanley, 1974.

Cady (1945, p. 574) estimated the stratigraphic throw in the order of 760 m. Detailed mapping has shown that the fault actually dips 60-70 degrees westward and is offset by several short cross faults, thus explaining its apparent irregular trace. Actual exposures of the fault surface in three places display polished slip surfaces with slickensides in the dip direction. Steep fracture fabrics compatible with the normal fault geometry are abundant near the Monkton fault, but decrease away from it. Total dip-slip movement is calculated at 850 m, assuming the faults dip west at 65 degrees and the stratigraphic throw cited by Cady of 760 m is correct.

Stratigraphic separation on the St. George fault is missing south of Monkton Ridge but its position is determined by actual exposures of the fault zone and breccia deposits, some of which are mineralized by goethite and manganite. A pronounced valley in the quartzite also follows the trace of the fault. Furthermore, the kaolin deposits south of Monkton Ridge occur along this fault zone and appear to be specifically localized by the intersection of small cross faults with the St. George fault. These intersections formed vertical channels of highly fractured rocks that trapped residual alumina in the Cheshire Quartzite and formed commercial kaolin in this area.

It is likely that the St. George and Monkton faults curve at depth because the folds that have been mapped in the Cheshire Quartzite of the central horst trend slightly east and plunge steeper than the minor and mapped folds of the same age in the adjacent downthrown blocks (Monrad, 1976).

At least four major cross faults offset the Monkton fault and the St. George fault although the exact geometry of these offsets is not defined by closely-spaced outcrops (Fig. 4). Cross faults offset mappable folds in the Cheshire Quartzites, and are otherwise marked by fracture zones, fault breccias, and actual slip surfaces with slickensides. Most of the notches in the ridges underlain by the Cheshire Quartzite south and east of Monkton Ridge are fault controlled. Some of these contain kaolin.

The third area is located in the northern part of the GIM fault system east of Mallets Bay near Colchester Pond (Figs. 2 and 5). Here south-plunging minor folds gently deform the penetrative, regional cleavage. It would appear from the map (Fig. 5) that

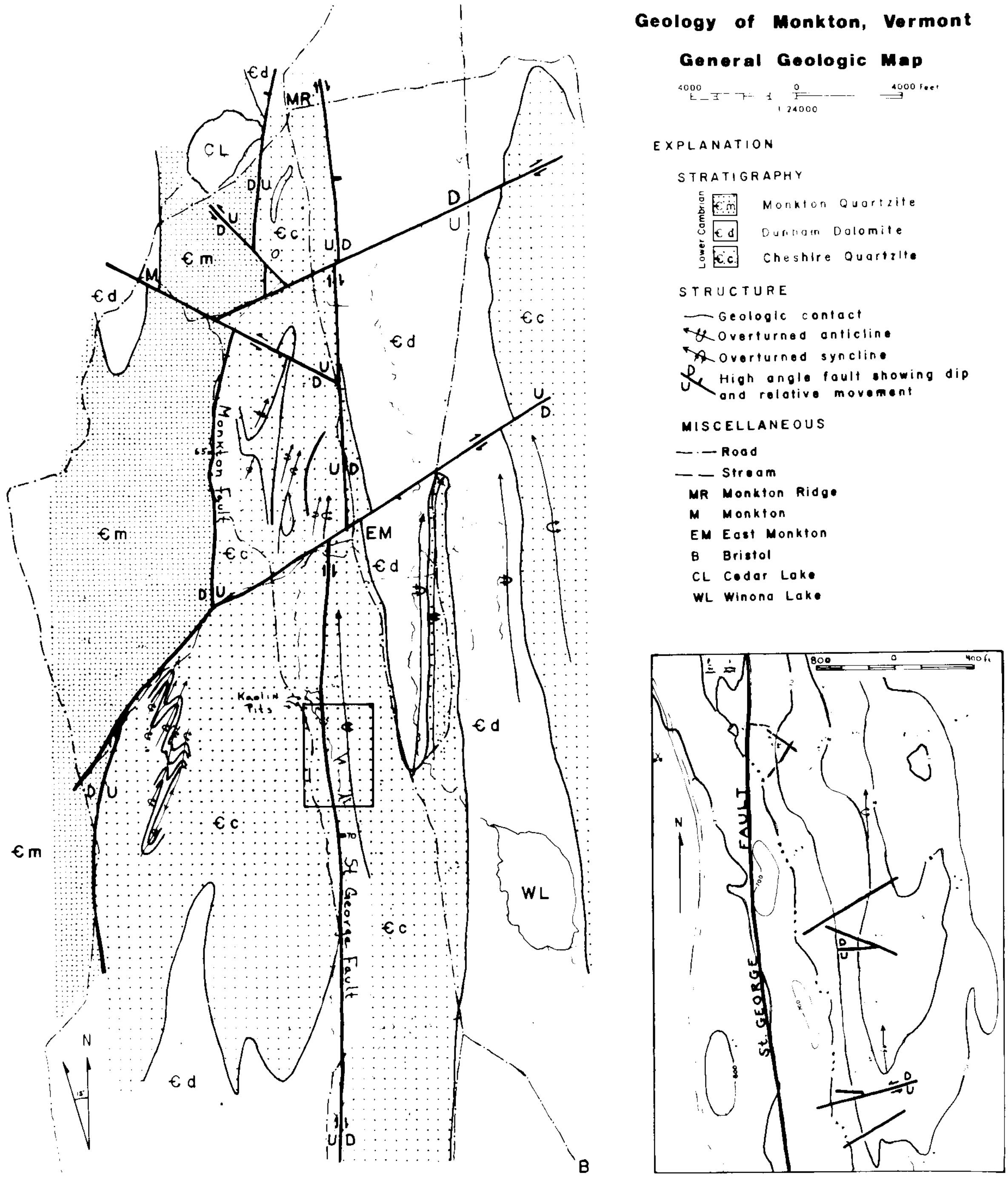

Figure 4. Geologic map of Monkton, Vermont, showing the St. George and Monkton faults. Black phyllite is folded tightly in the horst of Cheshire Quartzite. Geology by Black (1975) and Monrad (1976). Inset map shows the St. George fault and east-west faults and fracture zones near the kaolin pits.

the folds are large enough to control the outcrop pattern of the thrust zone and produce a major south-plunging anticline at Brigham Hill and a major syncline at Colchester Pond. This is basically the interpretation suggested by Stone and Dennis (1964, pl. 1). Detailed mapping by Rosencrantz (1975) and Agnew (1977) in this area has shown that the thrust surface and the bedding and regional cleavage in both plates do not change dip systematically from Brigham Hill to the west edge of the Hinesburg thrust as required by the syncline interpretation (Fig. 5). The key area is just north of Colchester Pond where the western edge of the Hinesburg upper plate is remarkably straight. If this were the western limb of a major syncline, the bedding or older layering should be vertical. Instead, it dips at very gentle angles to the west. Rosencrantz (1975) suggested that the western margin of the Hinesburg thrust has been cut by the northern extension of the St. George fault. Upthrown movement on the western block has partly removed the upper plate resulting in 2 km of left-lateral strike separation on the map (Fig. 5).

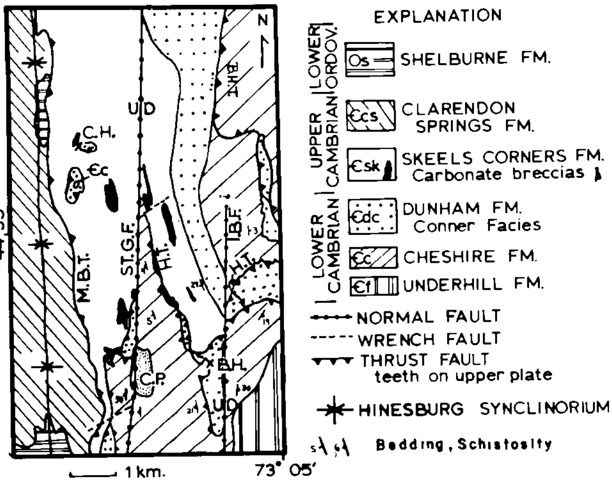

Figure 5. Generalized geologic map of the Colchester Pond area, modified from the Geologic Map of Vermont (Doll and others, 1961), by Rosencrantz (1975) and Agnew (1977) showing the location of the Muddy Brook thrust (M.B.T.), St. George fault (St.G.F.), Hinesburg thrust (H.T.), Indian Brook fault (I.B.F.) and Bald Hill thrust (B.H.T.), Colchester Pond (C.P.), Brigham Hill (B.H.), Cobble Hill klippe (C.H.) are also shown.

This interpretation is supported by a pronounced valley which passes through Colchester Pond. North-trending high-angle fractures in the Cheshire Quartzite are abundant in outcrops near the fault trace.

Approximately 2 km east of the St. George fault, just east of Brigham Hill, (Fig. 5), the Hinesburg thrust is again offset 200 m in an apparent left-lateral sense by the Indian Brook fault (Agnew, 1977). This fault is based on the abrupt change in the elevation of the Hinesburg thrust that occurs along the line marked "IBF" in Figure 5. West of this line, the Hinesburg thrust steps up 30-35 m exposing carbonate rocks and shale of the lower plate. Minor folds, which deform the regional schistosity and are coeval with flexing of the Hinesburg thrust to the east of the Indian Brook fault, are open and are not abundant in the vicinity of Brigham Hill as they should be if the map pattern of the Hinesburg thrust were due soley to late folding. Thus, the south-plunging anticline is also ruled out here.

The St. George fault and the Indian Brook fault, therefore, form two down-to-the-east step faults that offset the map pattern of the Hinesburg thrust in a left-lateral sense. The Indian Brook fault has not been traced southward, but Lake Iroquois may be controlled by it.

South of the Winooski River, the trace of the Hinesburg thrust swings westward forming a prominent "flap" on the eastern limb of the Hinesburg synclinorium. The St. George fault occupies a minor valley that extends from Williston southward through St. George to Monkton Ridge (Fig. 2). In the vicinity or St. George, the fourth area, the fault is marked by the sharp contact between the Cheshire Quartzite (Gilman facies) and the underlying Underhill Formation (Fig. 6). This contact is normally a very gradational boundary marked by a progressive increase in quartz in pelitic rocks, beds of quartzite and the simultaneous decrease in pelite as the section becomes younger. In this area, the Cheshire forms isolated blocks that terminate abruptly against the Underhill along the St. George fault. Throughout this area, small east-west vertical faults offset the Hinesburg thrust and a syncline in the Brownell Mountain Phyllite of the Bascom Formation. They do not offset the St. George fault, and therefore, must be older. In the Underhill Formation, they are represented by fracture zones (Fig. 6). Displacement on the cross faults is dip-slip because the apparent strike-slip offset varies according to the dip of the offset feature.

SUMMARY AND AGE OF THE GIM FAULT SYSTEM

Based on the above discussion, the GIM fault system is a remarkably continuous structure of small to moderate displacement. This fact probably explains why it has not been recognized earlier by workers in western Vermont. Although the evidence for its existence was taken from the Hinesburg synclinorium, the information cited in the Introduction indicates the GIM fault system may extend southward to Massachusetts. Furthermore, it may be a landward expression of the north-trending faults that appear to control the major topographic features beneath Lake Champlain (Hunt, 1979, personal commun.). Together, these faults may represent east-west expansion during the early stages of the opening of the present-day North Atlantic as suggested by Burke (1976).

What, then, are the age constraints on the GIM system? Clearly, it is younger than the Hinesburg thrust and the major folds of the Hinesburg synclinorium (Fig. 2). These major folds extend southward into the Middlebury synclinorium and the Taconic allochthon where they are considered Middle Devonian (Acadian orogeny) in age (Zen, 1972, p. 2578; Crosby, 1963, p. 128; Voight, 1972), although a Middle-Late Ordovician (Taconic orogeny) age is not completely ruled out to the north (Stanley, 1972, Fig. 17). Late movement of the Champlain thrust folded the penetrative slaty cleavage in the lower plate and gently flexed the western edge of the upper plate. Although radiometric age control is lacking, these events probably represent a late event in the Acadian orogeny. Alleghenyan deformation of late Paleozoic age, however, cannot be ruled out in western New England (Geiser, 1980).

The lamprophyre and trachyte dikes could provide the chronological key for better establishing the age of faulting in the region. Dated dikes are Late Jurassic-Early Cretaceous in age (125-160 m.y.; McHone, 1978, Fig. 5; Zaartman and others, 1967). The alkaline stocks of Charlotte and Cuttingsville in western Vermont and Ascutney in eastern Vermont cover the same age range. Unfortunately, the highly fractured character of the fault zones and the lack of continuity of the dikes across fault zones hampers the application of these dikes to the GIM fault system. One lamprophyre dike does cut a minor normal fault associated with the St. George fault near the hinge of the anticline north of Monkton Ridge (point A, Fig. 3). Another well-exposed dike cuts north-trending normal faults of minor displacement at Winooski Falls, Winnoski, Vermont (point B, Fig. 7). These faults are considered coeval with the St. George system because they cut compressional fracture fabric correlated with the Champlain thrust. Faulted dikes are reported from several localities in western Vermont (Zen, 1972, p. 2584; McHone, 1978, Table 2). Thus, the GIM fault system and its possible extensions appear to have developed before the intrusion of the dikes. I suggest a Late Triassic or Early Jurassic age because other major grabens were developing in the Appalachians and the North Atlantic at this time (deBoer, 1967; Burke, 1976). An older middle to late Paleozoic age for the system, however, cannot be ruled out.

ENVIRONMENTAL SIGNIFICANCE

The Mesozoic high-angle faults are important in terms of mineral deposits and groundwater resources. They may also form a potential earthquake hazard. Although the available information is far from complete, important conclusions can be suggested for each of these three.

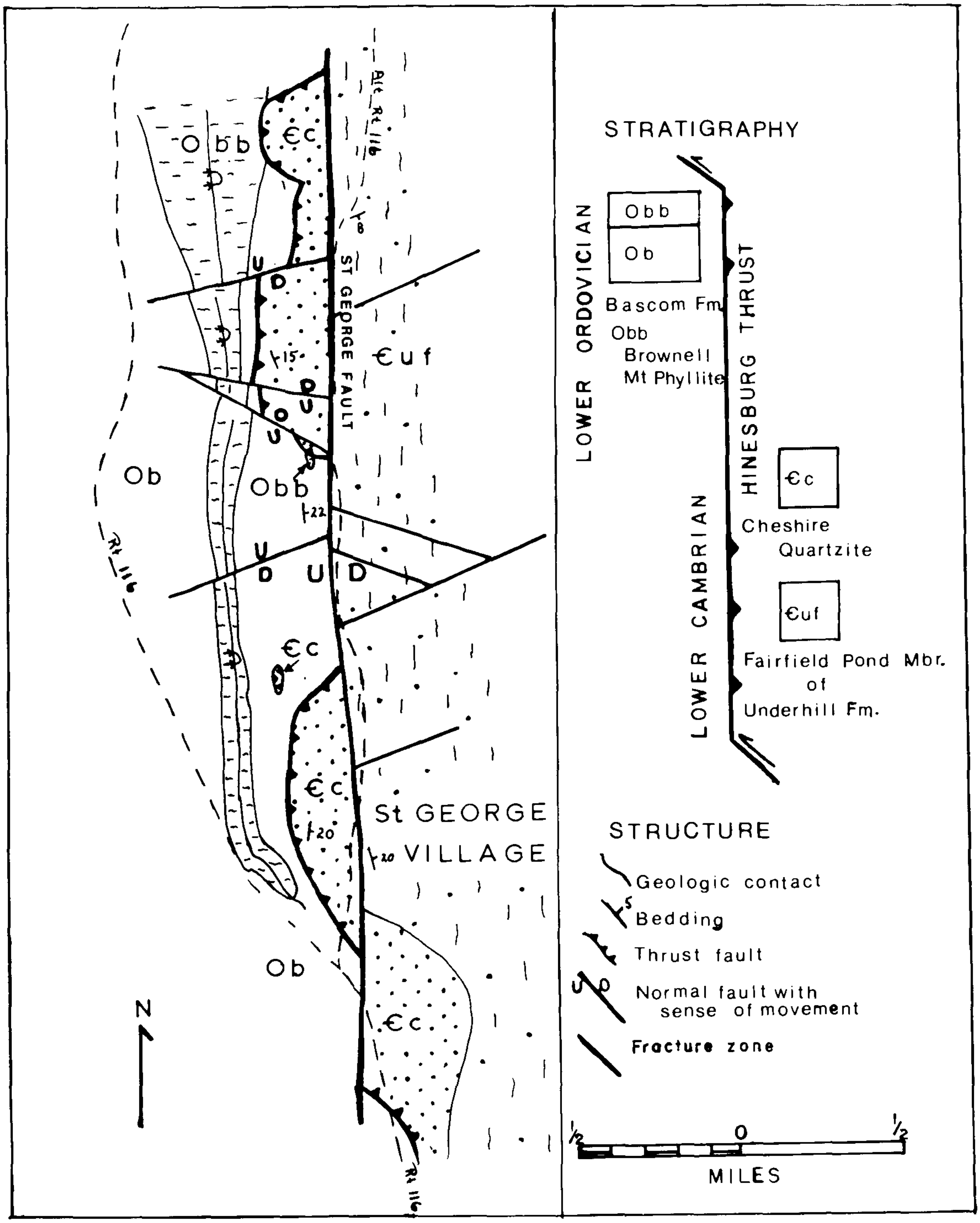

Figure 6. Geological map of the St. George area showing the Hinesburg thrust, short east-west faults to the west, and fracture zones to the east truncated by the St. George fault. Cross faults and fracture zones are vertical. St. George fault is assumed to dip east. Map modified from unpublished work of M. Black and M. Bergeron for Field Geology, University of Vermont, 1975.

Most of the kaolin deposits in western Vermont are found in surficial material underlying glacial drift (Burt, 1927). They commonly overlie the Cheshire Quartzite and occur very close to the steep western front of the Green Mountains where faults are abundant. Many of the low grade iron-manganese deposits in western Vermont are associated with or occur near kaolin deposits. At Brandon, lignite is also present (Burt, 1927, p. 80). Similar deposits are reported through the central and southern Appalachians (for example, Foose, 1945; King, 1950; and King and Ferguson, 1960). They are commonly thought to originate as residual products from the normal weathering of iron- and manganese-bearing carbonate rocks (King, 1950,p. 68, to cite only one example) or of feldspar-bearing rocks, in the case of kaolin. The residual material was deposited in topographic depressions. It is suggested that differential weathering of fractured rock may control the depressions in western Vermont.

The iron-manganese deposits at East Monkton certainly fit this interpretation because the ore is deposited as a cement in brecciated quartzite along the St. George fault. Burt (1927, p. 79-80) suggests a late Mesozoic to early Cenozoic age, perhaps Miocene, for these deposits based on the fossils in the lignite at Brandon. This age would

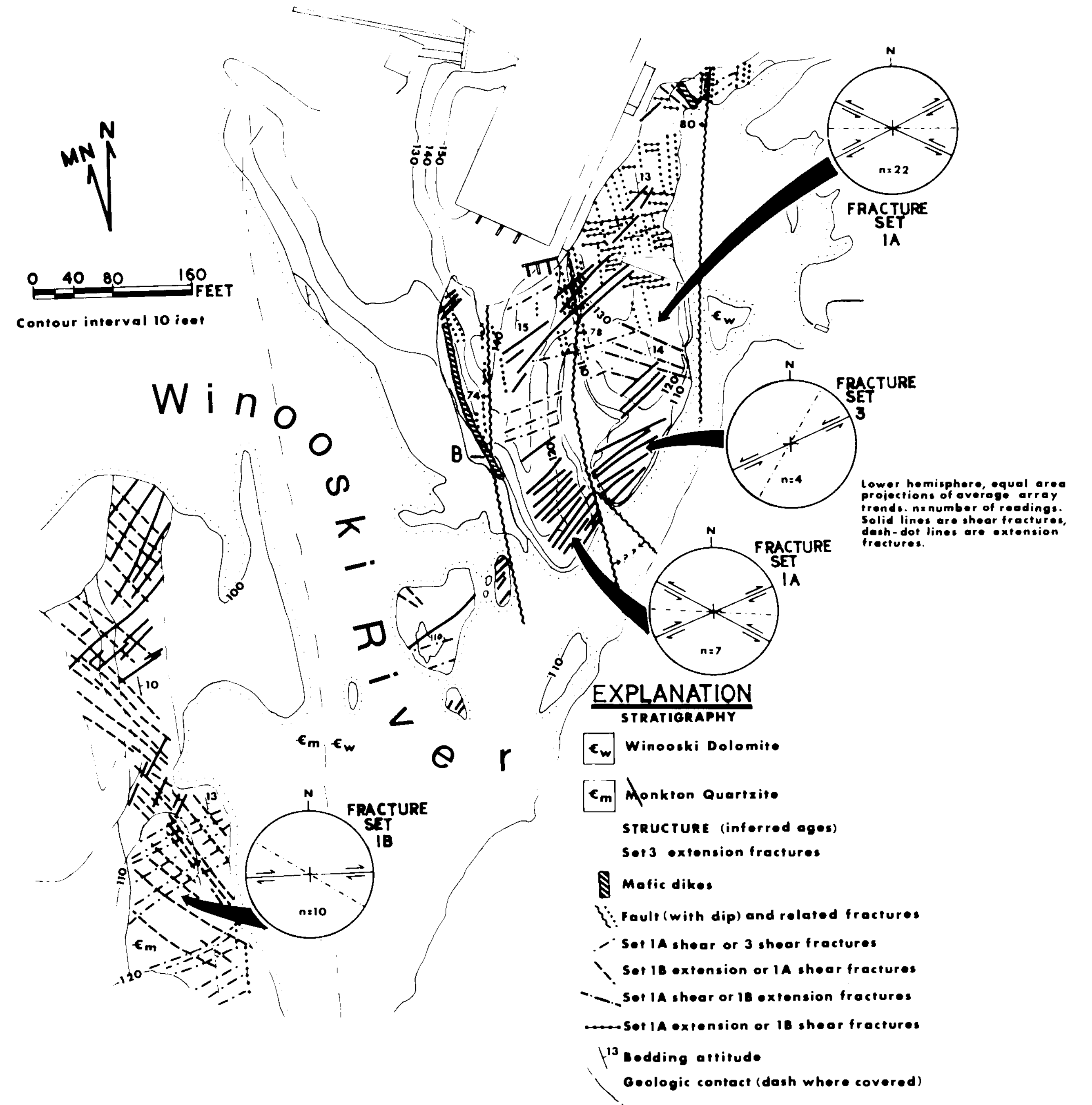

Figure 7. Geologic map of the Salmon Hole along the Winooski River, Burlington Vermont (Fig. 2), showing fractures, faults and en echelon fracture arrays. Based on field work and analyses by P. Winner and G. Smith, University of Vermont, 1979. Point B locates site where north-trending lamprophyre dike cuts north-trending normal fault.

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also apply to the surficial kaolin deposits associated with the iron and manganese.

The kaolin deposit at East Monkton, however, is quite different from the other kaolin deposits in Vermont in that it occurs in fractures and distinct layers within the Cheshire Quartzite rather than a surficial deposit on bedrock. According to Ogden (1969, Fig. 5), the deposit is at least 35 m deep and consists largely of kaolinitic quartzite interlayered with massive quartzite. The kaolin has replaced silica cement in porous quartzite. Fresh, authigenetic feldspar is common in voids between fractured quartz grains. Partly kaolinized feldspar is absent. Feldspathic quartzites are not reported from the Cheshire Quartzite in western Vermont, thus precluding formation of the deposit by deep weathering of indigenous material by groundwater solutions. Thus, irrespective of the origin of the kaolin, it does appear that many of the kaolin, iron-manganese deposits of western Vermont are controlled by faults. Those to the south of the Hinesburg synclinorium may or may not be part of the GIM system although their trend seems to suggest they are.

Perhaps one of the more important practical aspects of the high-angle faults is their relationship to the bedrock groundwater resources in western Vermont. The clastic rocks are devoid of inherent intergranular porosity because they are throughly recrystallized. Instead, the porosity and permeability are controlled by postmetamorphic fractures. Fracture density and extent (horizontal-vertical) are a function of rock type, strain intensity and relationship to such major structures as faults and folds. Whereas fold-related fractures have been documented elsewhere (Stearns, 1968; Friedman and Stearns, 1971), none of the fracture studies in western Vermont have shown this relationship (Sarkisian, 1970; Marcotte, 1975). Instead, the fractures cut across the major folds of the Hinesburg synclinorium and presumably show the same relationship to the major folds to the south and north. Many of the fracture fabrics increase towards major faults and infer stress configuration compatible with known displacement on the faults. Fracture density is more abundant in shale and dolostone and less abundant in marble and quartz. Fractures in shale are limited in extent and are tight due to the weak nature of the rock. In dolostone, they are commonly healed, whereas they are more open and extensive in marble due to ground water solution. Although fractures are less abundant in quartzite, they are far more continuous both horizontally and vertically. Furthermore, they appear to be open possibly due to larger stored elastic strain compared to carbonates and shale. The fracture characteristics in quartzite and marble, therefore, are important in providing significant porosity and permeability for bedrock aquifers in western Vermont. Because fractures are more abundant near faults, particularly where they intersect each other, the geology of the high-angle faults where they cross marble and quartzite takes on added importance. Areas underlain by the Cheshire Quartzite and cut by the high-angle faults are topographically high and thus form potentially important recharge areas. The Monkton area is a good example. Knowledge of the three dimensional geometry of the thick marbles (Shelburne Marble) and quartzites (Cheshire and Monkton) will also provide a valuable basis for evaluating bedrock reservoir potential.

Considerable attention has recently been paid to seismicity in the northeastern part of the United States (Sbar and Sykes, 1973; Fletcher and others, 1978, for example). Western Vermont is a part of a suggested northwest-trending belt of seismic activity from Boston to Ottawa. The activity, however, in Vermont is sparce compared to the Adirondacks where earthquakes of magnitude 4 or less are abundant (Sbar and others, 1972). The question here, then, is whether or not the fault system described in this paper constitutes a potential hazard in light of the seismic activity to our east and west. Despite its relative inactivity, this hazard cannot be disregarded. The length and remarkable planarity of the St. George-Indian Brook-Monkton system makes it an ideal candidate for reactivation. Furthermore, the Champlain and Hinesburg thrusts should be considered, although seismic displacement on the Hinesburg thrust may be unlikely because it is folded by Acadian deformation and offset by the high-angle faults described here. Displacement by stable sliding as represented by minor quakes would be expected if the differential stress were of sufficient value, and the direction of σ_1 oriented in such a position to generate sufficient finite shear stress along the fault surface. No such displacement has been reported in the historical past in western Vermont. The geology of the north-trending fault system is offset by a number of cross faults. These may well lock the north-south system allowing the deviatoric stress to build up until the strength of the locks is exceeded or until the frictional resistance is reduced by abnormal pore pressure. Thus, the lack of seismicity in Vermont may be deceptive.

The degree to which the faults of western Vermont are a hazard not only depends upon their geometry, but also the orientation of the present-day direction and magnitude of maximum compression, σ_1. If, for example, this direction is oriented at an angle to the north-trending faults, then a finite shear stress would exist. The severity of the resulting quake then depends upon the frictional resistance to movement. Thus, the degree to which the cross faults offset the planarity of the north-trending system becomes increasingly important since they effectively prevent strike slip movement. This aspect of the problem can be evaluated by straightforward geological mapping of the fault system. The other parameter, the orientation of the principal axis of compression, can be determined by _in situ_ stress analysis, hydrofracture, or analysis of seismic first motion studies. The latter two are far more reliable in evaluating the present day regional stress in rocks, compared to _in situ_ stress analysis which is complicated by retention of residual stress.

Sbar and Sykes (1973, Fig. 5) show that the direction of maximum compression trends east-northeast in a nearly horizontal position for a considerable part of the northeastern United States, stretching from eastern New York through Illinois. This information is based on hydrofracture, pop-up features, and seismic-first-motion studies. One of the most complete analyses comes from an earthquake swarm in 1971 at Blue Mountain Lake directly west in the Adirondacks (Sbar and others, 1972). Here the fault-plane solutions suggest that σ_1 plunges S 78 W at 5 degrees for shallow quakes (2 km or less) and S 61 E at 30 degrees for deeper events (2-3.5 km) (Sbar and others, 1972, Figs. 9 and 10). Information of this quantity is not available for Vermont to the author's knowledge. Oliver and others (1970, Fig. 1) report numerous small post-glacial faults in the Paleozoic shale and slate extending along the east side of the Hudson and Champlain valleys northeasterly into Quebec. Nearly all of these faults parallel the dominant cleavage (essentially north or northeasterly strike) with the east or southeast side upthrown. This indicates that the westerly orientation of σ_1 may persist into at least western Vermont.

In contrast, strain relief _in situ_ stress measurements from Proctor (dolostone) and Barre (granite), Vermont, indicate a north-trending direction (N 14 W-N 4 W) almost 90 degrees to the orientations to the west. According to Sbar and Sykes (1973, Fig. 5) this same general orientation persists to the east and south into Pennsylvania. Therefore, western Vermont appears to be on the threshold of a major change in the orientation of σ_1, assuming that all the data reflects the present-day state of stress. Interestingly enough, both orientations are reflected in the fracture fabrics, minor faults and deformation lamellae in quartz and dolomite from west-central Vermont, although many of these features appear to have developed in Paleozoic and Mesozoic time (Stanley, 1974; Sarkisian, 1970; Marcotte, 1975, for example). At Shelburne Access area near Burlington, Vermont, σ_1 has switched from east-west to north-south, although the former was an orientation of long duration and is thought to be associated with Acadian movement on the Champlain thrust (Stanley, 1974). The second configuration is also interpreted as Acadian, although a younger date certainly cannot be ruled out. Better chronological control is found, however, directly north at Winooski Falls where four stress configurations have been worked out primarily on en echelon fracture arrays, north-trending normal faults, and a north-trending lamprophyre dike (Fig. 7). A west-trending σ_1 position and northwest-trending σ_1 position (σ_2 vertical in both cases) are older than the normal faults (vertical σ_1 direction) which are in turn cut by the dike of presumed Early Cretaceous age (point B, Fig. 7). Northeast-trending fractures and coeval en echelon arrays of both left- and right-handed patterns cut the dike and indicate a subhorizontal σ_1 configuration trending northeasterly and essentially parallel to the orientation for shallow earthquakes from the 1971 Blue Mountain Lake earthquake swarm (Sbar and others, 1972).

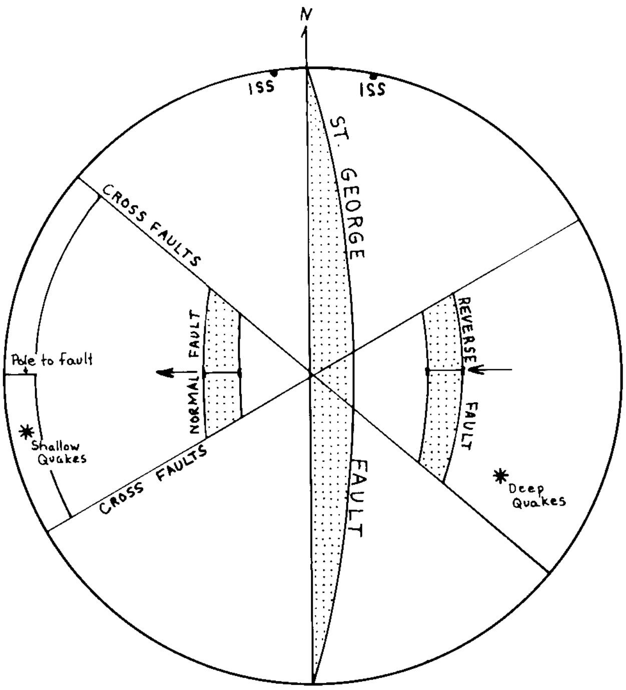

Figure 8. Lower hemisphere, equal-area projection of the St. George fault and associated cross faults in the Monkton area (Fig. 5). Asterisks (*) locate σ_1 positions for the Blue Mountain quakes for 1971 and 1972. In situ stress measurements are labeled ISS. Stippled area labeled "St. George fault" shows range of dips (80E - 90). Stippled area labeled "Normal fault", "reverse fault" are optimum σ_1 positions assuming a coefficient of friction of 30 degrees according to the Coulomb-Mohr fracture criterion. For pure dip-slip motion, the position is marked by short line with arrow. Other σ_1 positions within the stippled area result in oblique motion. The σ_1 position for pure strike-slip motion on the St. George fault is indicated by western arc near the circumference of the net. Resolved shear stress on the fault decreases to zero when σ_1 falls on the short line labeled "pole to fault". Arcs are restricted by locking cross faults that inhibit oblique slip with a major strike-slip component. Plane of the projection is horizontal with N marking north.

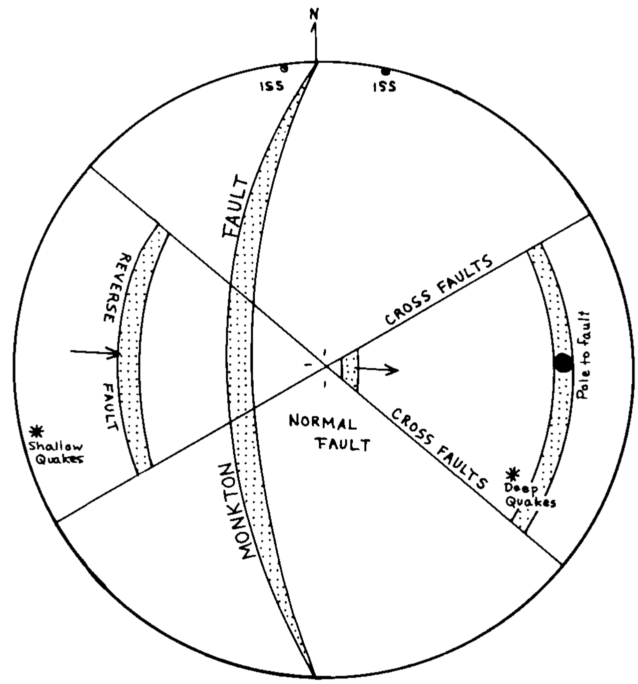

Figure 9. Lower hemisphere equal area projection of the Monkton fault and associated cross faults in the Monkton area (Fig. 5). Stippled area labeled "Monkton fault" shows range of dips (65 - 70W). Refer to Figure 8 for all other features on net.

With this information as a background, I will attempt to evaluate the potential for seismic activity of the St. George fault, the Monkton fault and the Champlain thrust. The Hinesburg thrust is not considered because it is folded and offset by subsequent faults, although all the evidence has not been considered in this paper. The cross faults are also not considered hazards since they are discontinuous and have only been displaced by less than 100 m. They are, however, important in retarding strike-slip movement.

If the in situ stress measurements for Proctor and Barre do represent the regional stress in western Vermont, then the seismic hazard of these three faults is very low since the inferred σ_1 direction would essentially parallel their strike (Figs. 8 and 9). The east-west σ_1 direction based on a variety of data from northeastern New York requires careful consideration. In Figures 8 and 9, I have plotted the St. George fault, the Monkton fault, the cross faults of the Monkton Ridge area, and the inferred stress positions from the Blue Mountain earthquake swarms of 1971 and 1972 (Sbar and others, 1972). I have selected this solution because it is probably the most representative measurement of the current regional stress system. Two fault plane solutions were given for the Blue Mountain swarm, one for shallow quakes (less than 2 km) and one for deeper quakes (2-3.5 km). Although both positions are plotted in Figures 8 and 9, Sbar and others (1972, p. 1314) believe the shallower σ_1 position probably represents the regional system.

If we compare the relationship of the shallow σ_1 position to all three faults, it is obvious from Figure 10 that the resolved shear stress is largest for the Champlain thrust, and smallest for the St. George fault. The lack of recognizable activity on the Champlain thrust may mean a reorientation of the σ_1 direction to a position where the resolved shear stress is greatly reduced, i.e., the north-south direction indicated by the in situ stress measurements. Alternatively, the thick section of shale below the thrust could very well dissipate the differential stress for east-west stress systems by distributed shear throughout the section. Earthquakes, however, should still develop to the east where the thrust cuts through the carbonate rocks and quartzite and finally into the crystalline rocks of the Precambrian (Fig. 10).

The St. George and Monkton faults appear to constitute low potential hazards since the resolved shear stress is moderate to low. The inferred shallow σ_1 position of Sbar and others (1972) would produce right-lateral motion on both faults with a reverse component for the Monkton fault (Figs. 8 and 9). The cross faults in the Monkton Ridge area obviously prevent this type of movement, particularly in the case of the Monkton fault. Interestingly enough, many of the markers are offset in a right-lateral sense by the St. George fault, although the actual movement is thought to be dip-slip. If the inferred deep σ_1 position of Sbar and others (1972) represents the stress system in western Vermont, then left-lateral movement would occur on both faults with a definite reverse component for the St. George fault.

What happens to the St. George and Monkton faults at depth? In Figure 10, the faults are shown as planar surfaces. The data by Monrad (1976) on

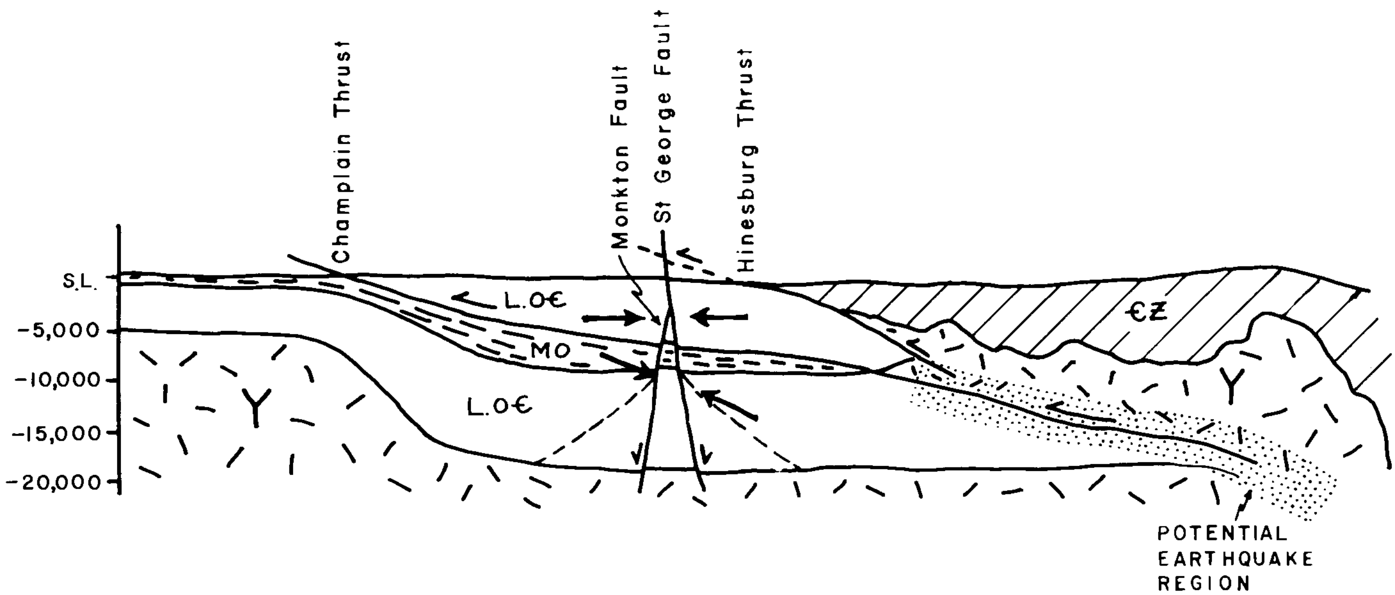

Figure 10. Cross section through west-central Vermont simplified from section B-B' (Doll and others, 1961) showing the earthquake potential of the Champlain thrust, Hinesburg thrust, Monkton fault, and St. George fault. Arrows on either side of the St. George-Monkton fault are based on the Blue Mountain data for 1971 and 1972 and represent the inferred shallow and deep σ_1 positions projected into the line of section. Solid lines represent planar surfaces and dashed lines represent curved surfaces for the St. George and Monkton faults at depth.

coeval fold fabrics in the Monkton horst and the adjacent blocks clearly shows that the faults must be curved at depth. If they curve away from the horst as shown by the dashed lines in Figure 10, the St. George fault would become more parallel to the Champlain thrust at depth. Therefore, the resolved shear stress would become larger for both the shallow and deep σ_1 positions of Sbar and others (1972). The critical relation, however, is the σ_1 orientation at the level where the high-angle faults begin to flatten out.

In summary, then, it appears that the Champlain thrust forms the most likely seismic hazard in western Vermont if preexisting faults are only considered. Along this surface, earthquakes are more likely to form beneath the Green Mountains than the Champlain Valley (Fig. 10). The GIM fault system may form a lock on the Champlain thrust surface if its form is anything like that shown in Figure 10. The significance of the lock in preventing westward upper-plate movement, however, may not be important since the down-to-the-east movement is only estimated at 500 m and, thus, trivial on a regional scale. The Monkton fault is a low seismic hazard because it is relatively short and the resolved shear stress appears to be small. The St. George fault may be more of a hazard because it extends for a minimum strike-distance of 50 km and may flatten toward the Green Mountains at depth. Both the St. George and Monkton faults are locked by cross faults in the Monkton culmination.

Thus, the lack of seismic activity in western Vermont appears to be due to a combination of factors: the lack of continuity of fault surfaces (Hinesburg thrust, cross faults) and low resolved shear stress on the GIM fault system. The thick shale section on the lower plate of the Champlain thrust may absorb elastic strain, thus minimizing seismicity. Alternatively, all of western Vermont could be decoupled on a still lower ductile surface as suggested by Dimant (1980).

ACKNOWLEDGMENTS

Much of the field work on the GIM fault system has been done by students of mine at the University of Vermont. Peter Moreau first discovered the St. George fault south of Hinesburg, Vermont. He and Meredith Crane revised the geology along the Cambrian and Ordovician contact in the southern part of the Hinesburg synclinorium (Fig. 2). Glen Monrad, Jake Overton, Don Skiba, Tad Pfeffer, Jim Lantz and Gail Manning mapped the Monkton area in a summer field geology course. Monrad and Marty Black continued this work as part of their senior honors theses. The fine work of Eric Rosencrantz and Paul Agnew in the Colchester Pond region extended the St. George fault northward. Agnew discovered the Indian Brook fault. Alfred Friedman extended their work southward. Black, Marcel Bergeron, and Richard Gillespie provided the new information in the St. George area. The work at Winooski Falls was done by Philip Winner and Jerry Smith. Quenby Dunlap and Wendy Seymour did the drafting. To each of these students, I extend thanks for their diligence, intellectual stimulation, and patience with their professor.

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ESTIMATING RECHARGE TO BEDROCK GROUNDWATER IN A SMALL WATERSHED IN THE LAKE CHAMPLAIN DRAINAGE BASIN

Craig D. Heindel
Wagner, Heindel, and Noyes, Inc.
Burlington, Vermont 05401

INTRODUCTION

The groundwater in the bedrock of a region can be replenished by precipitation percolating through overlying soils, by percolation from surface waters in streams and ponds, or by bedrock groundwater entering from outside the region. The identification of the relative contributions to bedrock groundwater from various areas of a region, and from various segments of the hydrologic cycle, is an important step in determining its hydrogeologic regime.

The study area, shown in Figure 1, is comprised of the adjacent watersheds of Muddy Brook and Allen Brook, located in west-central Chittenden County of northwestern Vermont, encompassing approximately 100 square kilometers. Both streams originate in the low hills in eastern Shelburne, St. George, and southern Williston, and flow northward into the Winooski River. Shelburne Pond, a 160-plus hectare pond with associated wetlands, is a major hydrologic feature of the Muddy Brook watershed.

The watersheds are located on the eastern margin of the Champlain lowlands, at the center of the Green Mountain uplands area. The western portion of the study area is relatively flat-lying and low in elevation, with surficial materials being lacustrine sediments and large deltaic deposits, underlain by the carbonates of the Lower Ordovician Beekmantown Group. The eastern portion of the study area has greater relief and elevations, with predominantly till soils underlain by the Cambrian schists and phyllites of the Camel's Hump Group.

Previous studies of bedrock groundwater recharge areas in Vermont have been primarily of a reconnaissance nature. Wagner and Dean (1972) described recharge to bedrock as tending to occur in the uplands. The soil associations of Farmington-Nellis-Stockbridge, Palatine-Vergennes, and Stockbridge-Nellis-Palatine were given as the most suitable to allow bedrock recharge. In an evaluation of groundwater resources in Chittenden County, the Vermont State Planning Office (1972) identified the uplands in the south-central and southeastern portions of the watersheds and the rocky knobs east of Shelburne Pond as possible or probable bedrock groundwater recharge areas. Acomb (1977) described groundwater contributions to the hydrologic cycle of a watershed in Mendon, Vermont. He found that groundwater was recharged primarily in the till-covered upper flanks of the hills, with smaller contributions coming from the lower elevations, due to the existence of a fragipan there. Absolute amounts of recharge to groundwater were estimated by two methods, one using Darcy's Law to calculate infiltration capacity.

METHODS

Of the three possible sources of bedrock groundwater recharge described in the opening paragraph, the latter two were determined to be insignificant in the watershed being studied. Elevations of static water levels in bedrock wells just outside the watershed boundaries are lower than in wells just inside the boundaries. This is taken as an indication that relatively little flow is entering in the bedrock from outside the watersheds. Stream gauging indicates that surface waters do not contribute significantly to recharge. Therefore, effort was concentrated on the evaluation of recharge to bedrock groundwater by percolation of precipitation through overlying soils.

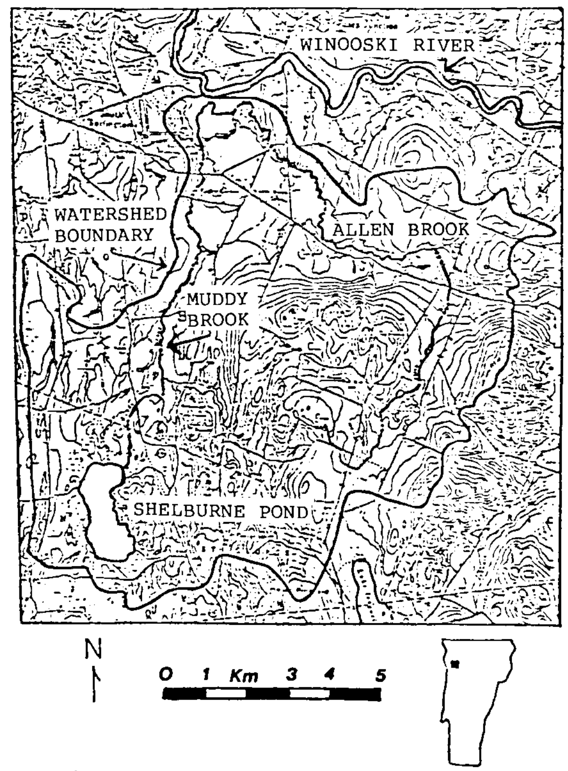

Figure 1. Location map of study area, showing boundary of two watersheds, major streams and the Winooski River.

A three-step process is used here to identify the particular portions within the watersheds that contribute significant recharge to bedrock groundwater:

1. STEP 1: determine the total soil permeability of the various soils in the watersheds,
2. STEP 2: determine the areal extent of the soils of similar total permeability, by measurements on a map generated by STEP 1,
3. STEP 3: evaluate the relative contributions to bedrock groundwater recharge from various portions of the watersheds, using information from STEPS 1 and 2 in conjunction with theory of flow through porous media.

RESULTS

STEP 1 - Total Soil Permeability

The determination of areas of similar total permeability was made on the basis of information from three sources:

1. an understanding of the surficial deposits and glacial history of the watersheds,

2. detailed soils information from the Soil Survey of Chittenden County (U.S. Department of Agriculture, 1974),

3. field observation made to locate areas with extensive bedrock outcropping.

Sources (1) and (2) above give information on the regions capable of allowing similar amounts of percolation of precipitation or snowmelt into bedrock. Source (3) locates areas where precipitation or snowmelt can enter directly into bedrock. Because regions of high total permeability are especially important to groundwater recharge, this analysis emphasizes the determination of the locations of such regions.

Source (1) - Surficial deposits. An understanding of the surficial deposits and glacial history of the watersheds can give an initial determination of regions of high total permeability. Lacustrine sediments have extremely low permeabilities and high water bearing capacities, and so would prohibit major recharge to bedrock. Lacustrine sediments are found in much of the western portion of the study area, and are also assumed to underlie the deltaic sands and gravels in the northern portion. Ice-contact deposits consist of moderately to highly permeable material for the most part, and are in some locations deposited directly on bedrock; these locations might be regions of high total permeability. Because ice-contact features in the watersheds are of very small areal extent and are scattered, isolated deposits, major recharge to bedrock groundwater is not likely to come solely from these features. The other predominant surficial material, till, varies greatly in permeability, thickness, density and degree of development of fragipan. Some tills in the watersheds are permeable enough to allow significant recharge to bedrock groundwater, while other tills are very impermeable and are presumed to prevent much recharge. Detailed mapping of different till types and properties was not undertaken in this study. As a first approximation in this study, all tills were considered to be regions of high total soil permeability. As shown in Figure 2, regions of high total permeability are located in the till-covered uplands in the south-central and southern and eastern margins of the area, as well as in the larger hills in the northeastern section, and in the low hills surrounding Shelburne Pond. The lowlying areas in the watersheds are underlain primarily by impermeable surficial deposits with low total soil permeability.

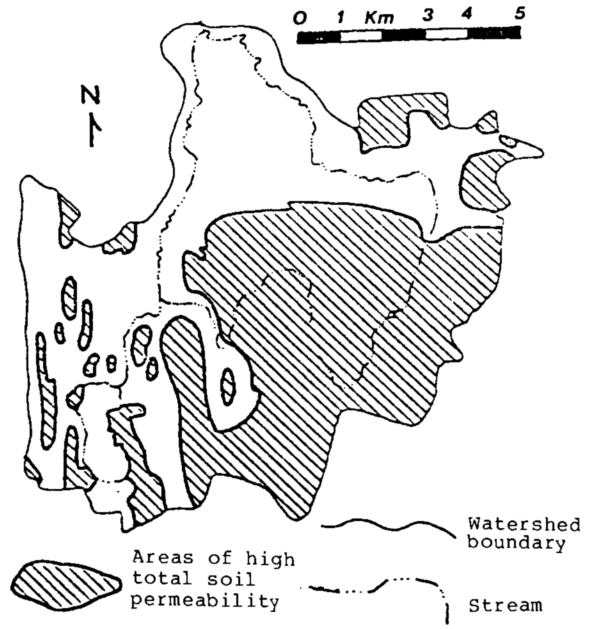

Figure 2. Map showing areas of high total soil permeability, based on surficial deposits information.

Source (2) - Soil Survey. The Soil Survey of Chittenden County (U.S. Department of Agriculture, 1974) has mapped in great detail the soil types in the study area, primarily from an agricultural and engineering perspective. Practically every soil type described in the Soil Survey can be found in the watersheds studied. For the purposes of determining total soil permeability, the soil characteristics described in the Soil Survey that are important to consider are (1) permeability, (2) depth to bedrock, (3) the nature of underlying unconsolidated materials, and (4) the existence of a fragipan. Table 1 lists these properties for each soil type, as given by the Soil Survey. This table also lists the total soil permeability of each soil, determined by evaluating the above four characteristics for each soil type. This evaluation is not information included in the Soil Survey. Soils characterized by the author as having high total permeability are those soils not underlain by lacustrine sediments or fragipan, are shallow to bedrock, and have a moderate or high permeability. Soils of moderate total permeability have only one or two of these favorable characteristics. Soils of low total permeability are those derived from or underlain by lacustrine sediments or recent floodplain deposits, or are thick tills with fragipans.

Figure 3 shows the locations of soils with high total permeability as determined by this author from Soil Survey information. This figure indicates that areas of high total permeability occur in the till-covered uplands in the shallow Farmington, Lyman and Palatine soils, and in the waterlain and ice-contact deposits in Colton, Deerfield, Duane, Stetson and Agawam Soils. The area of the uplands and hilltops that have thick or impermeable covers of till have a moderate or low total permeability. Regions of low permeability occur in the lowlands of the western and northern portions of the study area, in the highly impermeable soils.

The information on which this determination of total permeability is based is much more detailed than the information used to identify the areas of high total permeability shown in Figure 2, based on surficial geologic mapping. Comparing Figure 3 with Figure 2 indicates that the total area of regions of high total permeability according to Soil Survey information is less than the total area suggested by the more general surficial deposits information. This difference is due to the estimation, using surficial deposits information alone, that all areas underlain by till could potentially allow significant amounts of recharge to bedrock groundwater. Soil Survey information allows the elimination of certain tills that are very thick or dense or have a well-developed fragipan, both of which are presumed to hinder significant amounts of recharge.

Source (3) - Areas of bedrock outcrops. Field observations of areas containing extensive bedrock outcrops can also contribute to the location of recharge areas. No formal mapping scheme was used in this study to locate areas of extensive outcrops, but several portions of the watersheds can be crudely characterized as containing significant amounts of bedrock at ground surface. These areas were located by field observation and are shown in Figure 4, and include the several rocky knobs surrounding Shelburne Pond and the entire top and west flank of the height of land along the south-central boundary of the study

TABLE 1.

Soil type	Lacustrine Sediments	Depth to Bedrock (feet)	Perme- ability	Presence of Fragipan	Total Soil Perme- ability
Adams	?	5+	H		L,M
Agawam		5+	M,H		M,H
Au Gres	?	5+	M,H		L,M
Belgrade	X	5+	L		L
Cabot		5+	L	X	L
Colton	?	4+	H		L,M,H
Covington	X	5+	L		L
Deerfield	?	4+	M,H		L,M,H
Duane		5+	H		H
Eldridge	X	5+	L,M,H		L
Enosburg	X	5+	M,H		L
Farmington		1-1.5	L		M,H
Georgia		5+	L		L,M
Groton	?	4+	M,H		L,M
Hadley		5+	L		L
Hartland	X	5+	L		L
Hinesburg	X	5+	L,M,H		L
Limerick		5+	L		L
Livingston	?	5+	L		L,M
Lyman		1-1.5	L,M		M,H
Marlow		5+	L	X	L
Muck and peat		5+	-		L
Munson	X	5+	L		L
Nellis		5+	L		L,M
Palatine		1.5-3.5	L		M,H
Peacham		5+	L	X	L
Peru		5+	L	X	L
Raynham	X	5+	L		L
Scantic	X	5+	L		L
Scarboro	?	5+	L,M,H		L,M,H
Stetson		4+	M,H		M,H
Stockbridge		5+	L		L,M
Vergennes	X	5+	L		L
Whately	X	5+	L,M		L
Windsor	?	5+	L		L
Winooski		5+	L		L

Table 1. Forty-one soil types found in the watersheds studied, with listings of pertinent soils properties, as determined by the Soil Conservation Service (U.S. Department of Agriculture, 1974). The last column is an assessment of the total soil permeability made by this author, based on the four soil characteristics listed, and is not information supplied by the Soil Conservation Service. Codes are also developed by the author: X = present; ? = presence unclear from S.C.S. information; L = low permeability, 0 to 2 in/hr (1.41×10^{-3} cm/sec) or low total soil permeability; M = moderate permeability, 2 to 6.3 in/hr (1.41×10^{-3} to 4.44×10^{-3} cm/sec) or moderate total soil permeability; H = high permeability, greater than 6.3 in/hr (4.44×10^{-3} cm/sec) or high total soil permeability.

area. Color infra-red aerial photography would be useful in locating areas of extensive bedrock outcrops although that method was not used in this study.

Synthesis - STEP 1, Total soil permeability. Information from the three sources described above is synthesized into one map, locating areas of high and moderate-to-low total soil permeability. Regions of high total permeability occur in the uplands underlain by thin, permeable tills, and in isolated pockets of waterlain and ice-contact deposits. These regions are located in the higher elevations of the central portion of the watersheds, in scattered heights along the eastern margin, and in the several low hills scattered throughout the southwestern and northeastern portions of the study area. Figure 5 is

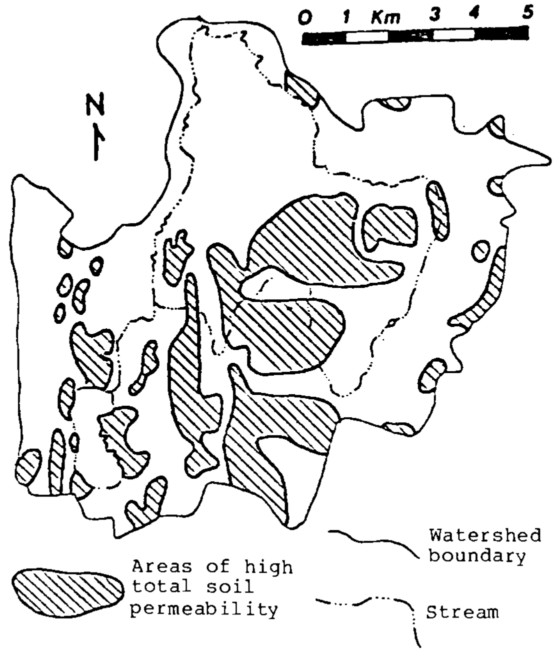

Figure 3. Map showing areas of high total soil permeability based on Soil Conservation Service information.

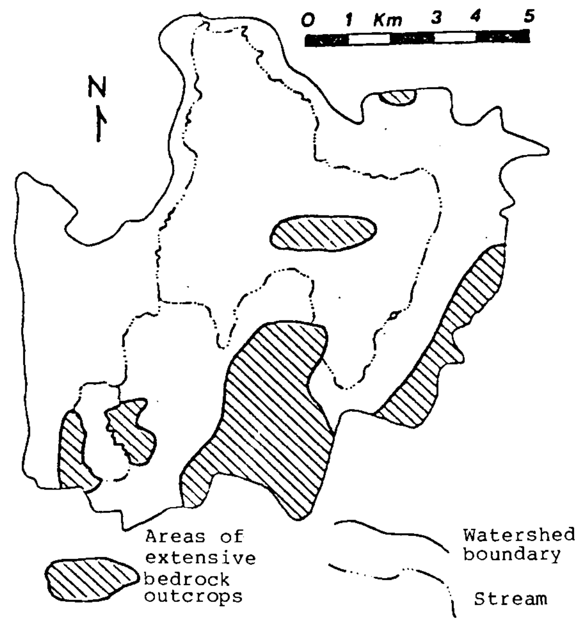

Figure 4. Map showing areas of extensive bedrock outcrops.

a map showing regions of high total soil permeability, determined by combining the information on Figures 2, 3, and 4. This map is not a simple summation of areas shown on those figures, since the information providing the basis for those figures has different degrees of precision. For that reason, the Soil Survey information (Fig. 3) is used as the primary locator of regions of high total permeability, and Figure 5 includes all areas shown on Figure 3 except those areas adjacent to streams. Gauging of streams in the watershed suggest that they do not contribute significantly to bedrock groundwater recharge, so they should be excluded from a map showing such recharge areas. A few additional areas are added based on information from Figures 2 and 4. Because of the varying degrees of precision in the base information, because the resulting map is not a simple summation of this information, and because natural systems are difficult to categorize precisely, the areas shown on Figure 5 should not be considered as absolute "black - or - white" determinations, but rather as areas defined within the limits of the precision of the base map.

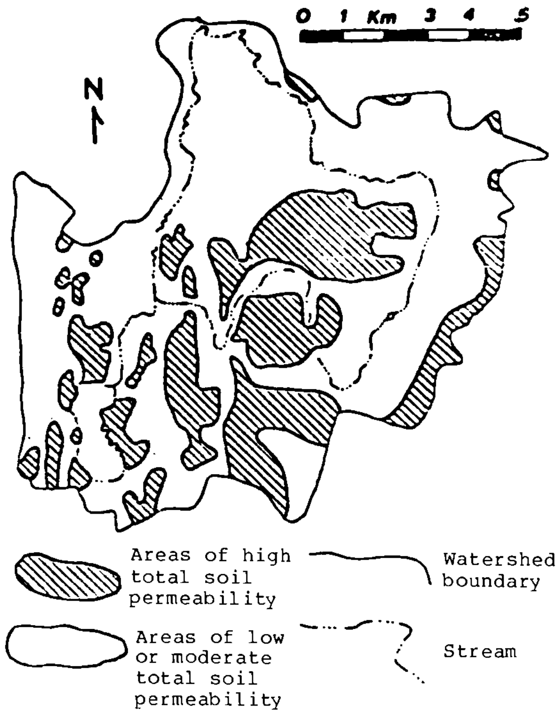

Figure 5. Map showing areas of high and low-to-moderate total soil permeability, based on a synthesis of information shown on Figures 2, 3 and 4.

STEP 2 - Areal extent

Regions of high total soil permeability as mapped on Figure 5 occupy approximately 32 percent of the total areas of the watersheds. The remaining 68 percent of the watersheds has low or moderate soil permeability.

STEP 3 - Relative contributions to recharge

The determination of those regions that contribute significantly to the recharge of bedrock groundwater is here done by combining information from Figure 5, calculations of area from STEP 2, and estimates of soil properties from the Soil Survey.

Current understanding of fluid flow through porous media (for example, see U.S. Department of Interior, 1977) is based on the classic work by French hydrologist Henri Darcy (1856), who discovered that velocity of flow is proportional to the hydraulic gradient, and who identified the constant of proportionality as a characteristic of the medium. Numerous refinements to this basic concept permit mathematical calculations to be made on many aspects of porous media flow. Pertinent to this analysis is the use of Darcy´s relationship expressed in the following form (U.S. Department of Interior, 1977):

$$Q = K \times i \times A$$

where Q is the rate of flow (dimensions L^3/t); K is hydraulic conductivity (L/t), the constant of proportionality relating velocity to hydraulic gradient; i is the hydraulic gradient (dimensionless), the change in elevation head per distance of flow; and A is the area normal to the direction of flow (L^2).

Applying this formula to recharge information on the watersheds under study allows the determination of the relative contributions to groundwater recharge from various regions. The following assumptions are made in this analysis: (1) flow is assumed to be vertical through overlying soils, which is likely to be true in most cases but not universally true, and (2) saturated hydraulic conductivity values are used, even though some flow is likely to be unsaturated. Although these assumptions do not hold true throughout the watersheds, a rough indication of recharge patterns can be gained from such an application of Darcy´s Law.

Thirty-two percent of the study area has high total permeability. According to the Soil Survey, the permeabilities of these regions range from 6 to 20 inches per hour (4.23×10^{-3} cm/sec to 1.41×10^{-2} cm/sec), with 12 inches per hour ($8{,}47 \times 10^{-3}$ cm/sec) being a reasonable representative value (chosen by this author). Sixty-eight percent of the watersheds has low or moderate total permeability, with permeabilities ranging from 0 to 6 inches per hour (4.23×10^{-3} cm/sec), and with 2 inches per hour (1.41×10^{-3} cm/sec) being representative (chosen by this author). This analysis assumes predominantly vertical flow through soils into the bedrock, so a unit hydraulic gradient is used.

Using these assumptions and representative values for permeability, the relative contributions to bedrock groundwater recharge from the two types of regions can be estimated by Darcy´s Law as follows:

areas of high total permeability:

K = 12 inches per hour

i = 1

A = 0.32

Q = (12) x (1) x (0.32)

Q = 3.8 relative flow units

areas of low and moderate total permeability:

K = 2 inches per hour

i = 1

A = 0.68

Q = (2) x (1) x (0.68)

Q = 1.3 relative flow units

ratio:

high total permeability : low or moderate total permeability
approximately 3 : 1

These calculations indicate that regions of high total permeability, occupying 32 percent of the total area, contribute about 75 percent of the total recharge to bedrock groundwater. Regions of low and moderate total permeability contribute about 25 percent of the total bedrock groundwater recharge.

SYNTHESIS OF RESULTS

The synthesis of all information presented above on recharge areas allows the development of a map of recharge areas. In the particular watersheds studied, areas of high total permeability have been calculated to supply approximately three-quarters of the recharge to bedrock groundwater, and contributions from sources outside the watersheds or from streams or ponds have been determined to be insignificant. Therefore, areas of high total permeability also constitute the areas of major contribution to bedrock groundwater recharge. The areas shown on Figure 5 can be considered the major recharge areas of the two watersheds studied, as shown on Figure 6.

It must be cautioned here that recharge areas cannot be located in a precise "black-or-white" manner without more detailed investigations than were possible in this study. Moreover, geologic factors that allow significant recharge to bedrock groundwater in some cases tend to vary gradually rather than abruptly, so any lines on a map such as Figure 6 must be viewed within the limits of rather imprecise base information about complex natural systems.

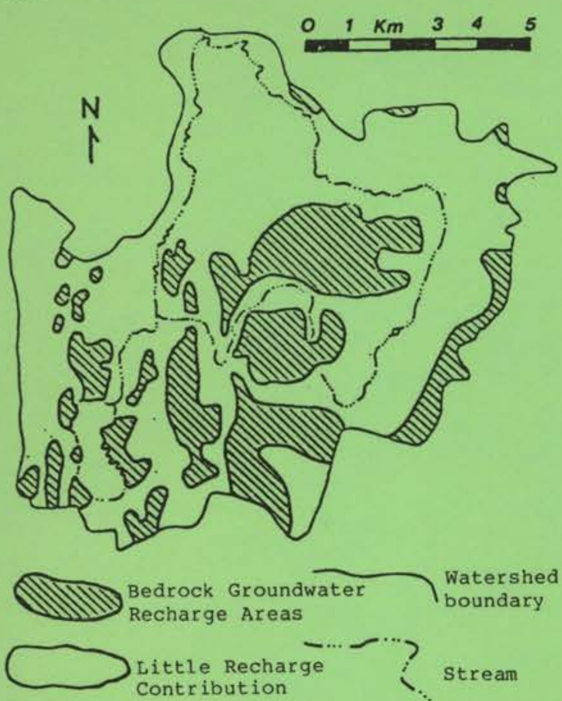

Figure 6. Map showing areas of significant bedrock groundwater recharge, based on information in Figure 5 and on a quantitative analysis using Darcy's Law, explained in the text. This map is identical to Figure 5 because the quantitative analysis indicates that areas of high total soil permeability contribute about 75 percent of the bedrock groundwater recharge, suggesting that these areas are the significant bedrock groundwater recharge areas.

CONCLUSIONS

Recharge to bedrock aquifers in the watersheds studied occurs primarily by the percolation of precipitation through overlying surficial deposits into the fractures in the bedrock. Areas of significant recharge are those whose soils are shallow and moderate to highly permeable throughout their entire depth. Significant recharge areas occupy approximately one-third of the total area of the region, but account for roughly three-quarters of the bedrock groundwater recharge. These areas are located in the uplands along the south-central and southeastern margins of the watersheds, in the upper elevations of major hills in the north-central and northeastern portions of the watersheds, and in the rocky knobs surrounding Shelburne Pond. The remaining two-thirds of the region contributes about one-quarter of the recharge to bedrock groundwater, presumably by slow leakage through low-permeability soils.

DISCUSSION

A method has been developed, using information available to and interpretable by surficial geologists and soil scientists, to estimate the relative contributions from specified areas to bedrock groundwater recharge by percolation of precipitation through overlying soils. This method should not be used to obtain absolute numerical values, because of the several assumptions and generalizations used in this application. It does provide useful relative numerical values. Results of the application of this method could be part of a larger study of the entire hydrogeologic regime of an area. When combined with a contour map of bedrock groundwater elevations and other information, a clear picture of the flow patterns of bedrock groundwater might emerge. This method could also be used in a groundwater protection and management program. Because recharge areas often provide water for large areas of an aquifer, and because soils in recharge areas are often thin, relatively coarse, and less able to attenuate pollutants, the protection of these areas from pollution is especially critical.

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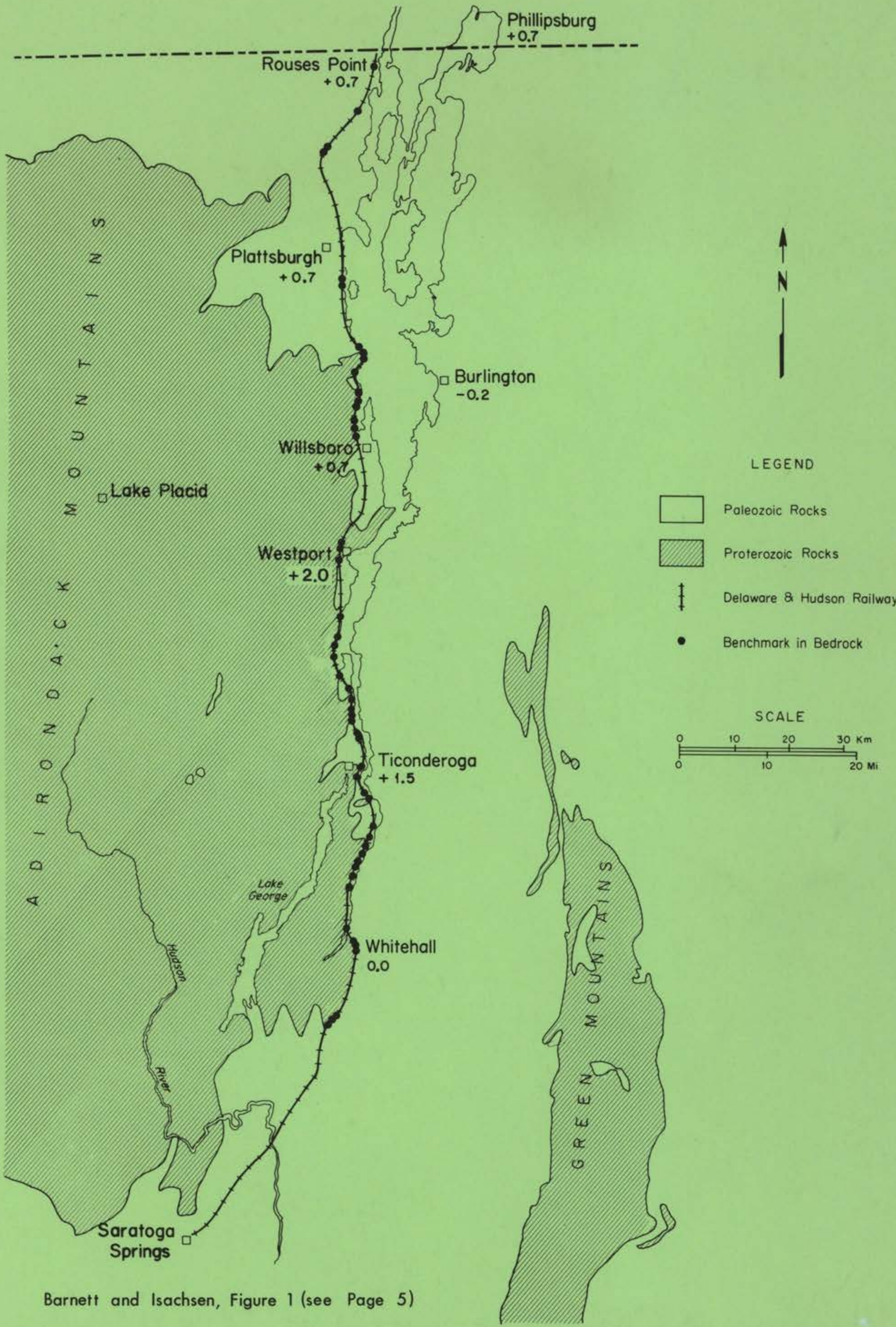

Barnett and Isachsen, Figure 1 (see Page 5)