THE GLACIAL GEOLOGY OF VERMONT

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INTRODUCTION

The most recent geologic epoch, the Pleistocene, widely known to the non-geologist as the Great Ice Age, is everywhere manifested on the surface in the State of Vermont. During this short interval of a million years, or less, the state, as well as the whole of New England, was subjected to repeated episodes of continental glaciation. The glaciers formed far to the north and as they advanced southward over the area the weathered mantle was removed from the bedrock and the topographic expressions of the mountains, hills and valleys were accordingly modified. These advancing masses of ice also scooped out basins of various sizes that were later to be filled with water, forming lakes. As the ice retreated to the north, a new mantle was spread over the surface of the land by the melting ice and the resulting meltwater streams. After the last glacier retreated, the rock structure was the same as it had been, and the major aspects of the topography were only slightly changed; but the local details of the surface expression were greatly modified and the surface material was completely different. As Vermont is predominantly an agricultural and recreational state, it is quite obvious that the glacially formed surface features, both erosional and depositional, have a direct influence on the present cultural aspects of the region.

The Hitchcock Concept of the Great Ice Age

It seems appropriate, since this report is published as a part of the Vermont Hitchcock Jubilee celebration, that this manuscript should begin with a short discussion regarding Professor Edward Hitchcock's concept of the Glacial Epoch and his classification of the drift of Vermont and New England. These concepts are most important because they were accepted by many of Professor Hitchcock's colleagues and were not abandoned by many New England geologists until after the turn of the present century. It is a fact that many of the misconceptions of his writings hindered the correct interpretation of the glacial material for over forty years.

Jacobs (1946 p. 22) described Hitchcock as a Neptunist because of his belief of a universal ocean. He was, however, more than a Neptunist inasmuch as his ideas concerning the Pleistocene Epoch were a mixture of the old drift theory, marine invasion and valley glaciers. No doubt the fact that he was schooled and ordained for the ministry (1821) had a great bearing on his strict adherence to the scripture in regard to the flooding of the continent. It also seems that it was impossible for Hitchcock, as well as his co-workers, to conceive of a continental ice sheet of such magnitude that it would cover the state of Vermont, much less one large enough to cover New England or the northern half of the North American continent.

Hitchcock believed that it was necessary to have a marine invasion (deluge) to explain the wide distribution and the great variety of recent detritus which he has observed on the surface of this region. Since this surface material was everywhere, even on the tops of the mountains, he proposed that a depression of the land of approximately 5000 feet was the beginning of the geologic epoch that is now known as the Pleistocene. To account for the unsorted glacial till, which he designated "unmodified drift," Hitchcock utilized the drift theory. He believed that ocean currents from the north and northwest carried icebergs into New England. As the icebergs melted, the rock material contained in the ice was released, dropped to the bottom of the ocean and was spread over the submerged land surface to form the "unmodified" deposits. This hypothesis is most interesting since it is apparent that Professor Hitchcock had in mind icebergs similar to those common to the north Atlantic Ocean at the present time. These, of course, are formed by the calving of continental glaciers on the island of Greenland. The writer does not know what the source of the presumed icebergs might have been since a 5000 foot rise in sea level would have submerged so much of the land to the north and northwest that no glaciers could have formed, and therefore no icebergs. It is evident that Hitchcock gave little thought to the actual place of origin of the ice, the cause of the icebergs, or the source of the large quantity of material they supposedly carried.

For outwash (glacio-fluvial) deposits, Hitchcock used the term "modified" drift. The origin of these assorted sands and gravels he explained as resulting from shore erosion and the reworking of the "unmodified" drift by wave action. The modification of the iceberg deposited detritus began when the marine water level had been lowered, presumably by an uplift of the land, approximately 2400 feet and, at this stage, marine beaches began to form along the slopes of the mountains. Undoubtedly Hitchcock envisioned a very slow withdrawal of the

sea in order to account for the widespread and relatively thick deposits of glacio-fluvial materials.

Hitchcock did recognize some valleys that he conceded had been carved by ice erosion. These, he believed, were eroded by valley glaciers that formed on the higher peaks of the Green Mountains. The ocean currents from the north, according to his hypothesis, reduced the temperature in Vermont to the extent that local glaciers could have formed in the higher altitudes of the state.

It is interesting to note that the above concepts, published in the Geology of Vermont (1861), record an about-face in the thinking of Professor Hitchcock from those he had expressed twenty years earlier. In his presidential address to the first anniversary meeting of the Association of American Geologists in 1841, the statements he made suggest an entirely different view. A portion of this address was devoted to a discussion of the glacial theory of Louis Agassiz and on this occasion he gave twelve arguments favoring Agassiz's conclusions concerning New England. He even expressed his satisfaction that he had finally found a valid explanation for the origin of the surface material. It should be pointed out, however, that Hitchcock stated at that time (1841) that he could not determine whether or not ice or water was the most important agent.

Hitchcock's agreement with Agassiz and the glacial theory was apparently short lived, however, inasmuch as a letter which he wrote to C. B. Adams, then State Geologist of Vermont, on September 1, 1846 had a distinctly different tone. In this letter, he told his former student that he was sure he (Adams) would find the facts necessary to "clip the wings" of some of the hypotheses (concerning glacial deposits) and the geologists that were making "much noise on the subject" at that time. (Adams, 1846)

The above mentioned address, as already stated, was made at the first annual meeting of the Association of American Geologists. The fact that Hitchcock was the first president of this organization emphasizes his popularity among the American geologists at that early date and their mutual respect for him. It is not surprising, therefore, that his reluctance to accept the glacial theory had widespread influence on the thinking of other geologists, particularly in New England. Professor Hitchcock was one of the most eminent and most productive geologists of his day and it is not the intent of this report to discredit him in any way. The above comments on his mistaken views, pertaining to the subject of glaciation, are presented here merely to show the impact of his widespread influence and to set the stage for a discussion of later studies.

One of Edward Hitchcock's assistants at the time of his geologic work in Vermont was his son, C. H. Hitchcock, whose strong convictions on the subject of glaciology were in agreement with those of his father. The younger Hitchcock perpetuated these views for almost a half-century following the publication of the *Geology of Vermont*. It was not until 1904 that any report stating a changed view was published. In this latter report, C. H. Hitchcock (1904) admitted that his earlier ideas were incorrect and he accepted, for the first time, the glacial theory. In his own statement he said that "icebergs had given way to glaciers." In 1904, however, he still used the confusing term "modified" drift to denote the various types of glacio-fluvial material. This term was erroneously used so much in the early writings of New England geologists that its usage in reference to any kind of glacial deposit is no longer possible without confusion.

There are many valid and useful aspects of the 1861 report as it relates to the Pleistocene geology. The recorded striations contained in this publication are still the most complete of any that have been made to date. These give a very good general view of the most recent ice directions across the various parts of the state. The descriptions of the deposits made by Hitchcock are quite clear and in most cases quite accurate as far as they go. In many cases, as strange as it may sound, his interpretations are more nearly correct than those of recent dates, and his notes on specific items are quite valid and most helpful.

The Status of Previous Investigations of the Surface Material

Professor Hitchcock (1861) stated in his report on the Geology of Vermont that none of his survey members felt qualified to give sufficient attention to the surface geology. For this reason, the surveys that resulted in this report did not include a study of the glacial deposits. The status of the surface geology remained much the same as Hitchcock had left it for almost one hundred years as no program for the classifying and mapping of the surface deposits was begun until 1956. It therefore seems necessary to report here that in the hundred years since the publication of the Hitchcock report little has been added to the knowledge of the Pleistocene history of the state.

There are many reports on various areas of Vermont and on certain

aspects of the Pleistocene geology, but the size of the areas covered by most of these studies is so limited that their regional application is practically impossible. Much work has been done in a general way on the lake deposits of the Connecticut River valley, for example, but no complete survey of this basin has been made. The results of many studies have not always added to the Pleistocene history, inasmuch as the investigations were too limited in their objectives to be conclusive, and others were too incorrect to be of value.

Many well qualified geologists have made studies in the Champlain Basin and the knowledge here is rather complete and is still progressing. The recent work of Chapman (1937; 1941) is particularly noteworthy except that the glacial sequence has been modified by more recent investigations in the St. Lawrence Valley (MacClintock and Terasmae, 1960 and MacClintock and Stewart, in press).

It is probable that some of the most significant glacial deposits of Vermont cover the mountainous, northeastern section of the state. These deposits, however, have not been studied enough to know their true importance. A few of the larger lakes of this area have been described and their origin has been determined. But the glacial surface, the upland swamps and the lacustrine deposits have been merely noted in former reports.

The published material concerning the glaciation of the Vermont Valley and the Green Mountains is very limited. Nothing, to the writer's knowledge, has been done on the glacial deposits of the Taconic Mountains and, even in areas adjacent to the Connecticut River and the Champlain Basin, the glaciation of the uplands has received only general coverage in specific localities.

The above statements briefly summarize the conclusions of a three-year survey of the literature of the past one hundred years relating to the glaciation of Vermont and the adjacent areas. It goes without saying that the results of this survey of the literature have been disappointing, to say the least, since little was gained in so far as the Pleistocene history is concerned.

The Survey of the Surface Deposits now in Progress

Because of the great need for a study of the surface deposits of Vermont, a survey was started (by the writer) in 1956. The most immediate purpose of these investigations is to produce a surface materials map. When completed, this will be the first systematic study of the deposits in Vermont and the surface map produced will also be the first. In addition to the map, however, the objectives of this project include the collecting of data that will add to the knowledge of the glacial stratigraphy and glacial lakes sequence, thereby allowing a more complete interpretation of the Pleistocene history.

The survey, as already stated, is still in progress and to date surface maps of eleven quadrangles have been made. These quadrangles comprise approximately twenty percent of the total in the state (Figure 1). As can be seen, the survey has not yet covered a large enough area to give a very complete answer to most of the problems that are still unsolved (Plate I).

The Source of the Material for this Report

It would be advantageous if the writer of this report could record the details of the glacial materials and structures in all parts of the state from the point of view of one who had studied each section. But, as stated above, the present survey has been in progress only four years and, as yet, only a small portion of the state has been surveyed. The data collected thus far during the study will, of course, be utilized in this report but it will also be necessary to rely heavily on the scanty literature concerning the areas that have not been recently covered.

As a part of the research for the preparation of this manuscript, a literature survey was made covering studies of the glacial geology of Vermont and adjacent areas of New York and New England since 1861. This literature search resulted in the bibliography that appears at the end of this publication. In spite of the lengthy list of titles in the bibliography, the pertinent information concerning many areas is quite limited. The coverage that this report can give these areas is of necessity quite brief and of only a general nature. One of the objectives of the present survey is to produce a more detailed thesis on the Pleistocene history of Vermont when time has allowed a more complete coverage.

It is therefore the purpose of this report to review the general aspects of the glacial geology as it has been recorded in the literature. These concepts will be amplified and modified when it becomes possible to do so from the significant findings of the present survey. Thus, this paper is to acquaint the reader with the Pleistocene history of the State of Vermont to the extent that information is available at this date.

Acknowledgements

The surface survey, now in progress, is sponsored jointly by the Vermont Geological Survey and the Vermont Highway Department. The

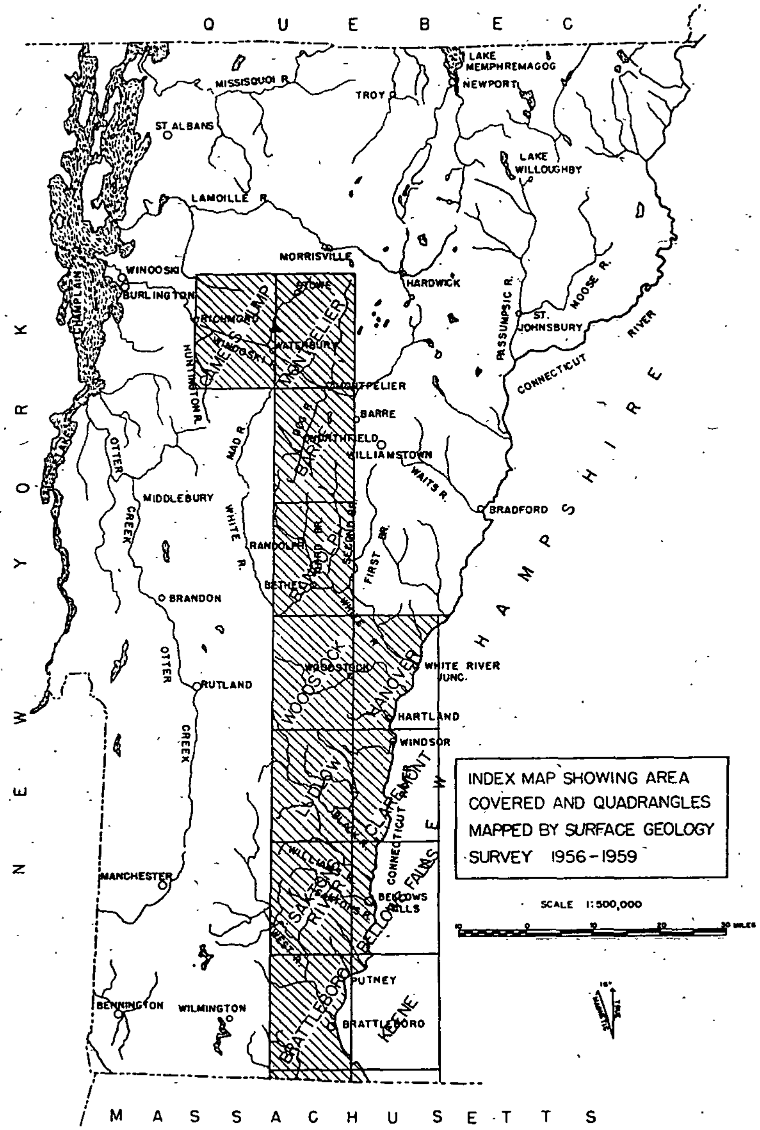

Figure 1. Index map showing area covered and quadrangles mapped by Surface Geology Survey 1956–1959.

work is being carried out under the direct supervision of Dr. Charles G. Doll, State Geologist. The writer was ably assisted in the field during the 1956 and 1957 seasons by Lawrence D. Perry. During the summer of 1958 and a part of 1959, Thomas H. Hawisher was field assistant and Hugo Thomas also assisted during a part of the 1959 field season.

The literature search, mentioned above, was generously supported by Miami University, Oxford, Ohio in the form of graduate research assistantships over a period of two years. J. Douglas Glaeser prepared the bibliography during the 1957–1958 school year. During the next year (1958–1959), Thomas H. Hawisher (who was also a field assistant) assisted in the tedious task of locating, obtaining, reviewing and evaluating the three hundred and fifty titles contained in the original bibliography. Miami University also supplied funds necessary for the drafting of the maps and diagrams contained in this manuscript.

The writer gratefully acknowledges the valuable assistance of Dr. Paul MacClintock of Princeton University and Dr. Raymond E. Janssen of Marshall College. Dr. MacClintock visited the writer in the field and read the manuscript. Dr. Janssen also read the manuscript. Their comments, criticisms and suggestions have been most helpful in the preparation of this report.

THE PLEISTOCENE EPOCH IN NORTH AMERICA

The Pleistocene Epoch of the Quaternary Period (Richards, 1953, p. 351) in North America dates from the climatic extreme that allowed the accumulation of snow and ice in such great proportions as to cause the formation of continental glaciers. Paleontological research and the study of marine sediments may, in the near future, supply a more explicit basis for marking the beginning of this glacial age, but at the present time the climatic factor seems to be the most reliable. In respect to the marking of the end of this epoch, it is not yet possible to ascertain that it has indeed come to an end. The last continental glaciers retreated from North America less than 10,000 years ago, and since the interglacial stages were of much longer duration, it is quite possible that the present is merely an interglacial stage of the Great Ice Age or even an interval of deglaciation of the last glacial episode.

Subdivisions of the Pleistocene Epoch

Basically, the Pleistocene Epoch is a four unit interval in geologic history, represented by four episodes of glaciation. It is estimated that the first ice advance, at the beginning of the Nebraskan Stage, started

about one million years ago. The Nebraskan was followed by three succeeding glacial maxima which have been named the Kansan, Illinoian and Wisconsin (Table I). During each of these stages, the ice formed far to the north, probably in the mountains of northeastern North America (Flint, 1952), and the ice moved outward in all directions. At one time or another during this short epoch, the whole of the northern and portions of the central regions of the North American continent were covered by continental ice sheets. In the eastern and midwestern sections of the United States the southern limits of the ice invasions extended at least as far south as the islands off the coast of New England, into northern New Jersey and Pennsylvania and to the present Ohio and Missouri rivers. Each ice advance was a slow, spreading movement, and after the glaciers had advanced to their terminal positions, the withdrawal was also a generally slow retreat.

After each ice retreat an interval of deglaciation followed during which time temperatures rose again, and the climate approximated that of the present time. These periods of ice retreat, the interglacial stages, have been designated the Aftonian, Yarmouth, Sangamon, and the interval since the retreat of the last glaciers which this report prefers to call the post-Cochrane (Table I). Although it is true that the ice advances and retreats progressed quite slowly in the ordinary concept of time, geologically speaking the Pleistocene Epoch was a very short span and the glacial history is a record of four ice advances and retreats in very rapid succession.

Subdivisions of the Wisconsin Stage

Due to the recentness of the Wisconsin glacial invasions and because the deposits of this stage have not been overridden by subsequent ice sheets, it has been possible to divide it into several subdivisions. For many years it was believed that this stage was represented by four, and only four, distinct till sheets, and it was therefore subdivided into four substages. These were designated (from oldest to youngest) the Iowan, Tazewell, Cary and Mankato. More recent studies, however, have established the fact that there were more than four maxima (substages) and, at the present time, seven substages are recognized in the Wisconsin stratigraphy (Hough, 1959) (Table II).

The above mentioned original four substages were proposed by Leighton in 1933. These were well accepted and were in general use shortly after they were proposed. In 1943, however, Thwaites identified a till sheet, which he called the Valders, that he believed to represent a

Table I.

SUBDIVISIONS OF THE PLEISTOCENE EPOCH

	Post Cochrane
Wisconsin Stage	
	Sangamon Interglacial
Illinoian Stage	
	Yarmouth Interglacial
Kansan Stage	
	Aftonian Interglacial
Nebraskan Stage	

fifth substage. Antevs (1945), however, considered this drift as the same age as Leighton's Mankato. Except for the short period during which there was debate concerning the Mankato and Valders, the Leighton proposal of 1933 was used as the basis for Wisconsin stratigraphic correlation for almost a quarter of a century. In 1950, Leighton and Willman recognized a drift which they believed to be younger than the Iowan and this drift has since been mapped by Shaffer (1956) and is now known as the Farmdale. In 1957, Elson published the results of his studies of the Agassiz Lake basin. In the course of his studies he traced the Valders and the Mankato drifts across the basin and established the fact that they were of different age; the Valders being the younger. The Cochrane was recognized in Canada and described by Karlstrom (1956).

Like the glacial stages, the substages were also separated by times of deglaciation called intervals. These, in general, have not been as well delineated and defined as the interglacial stages. The intervals are usually simply named the Farmdale-Iowan interval for the time between the Farmdale and the Iowan; the Iowan-Tazewell interval between the Iowan and Tazewell substages, etc. (Table II). An exception to this system of naming is the Two Creeks interval between the Mankato and the Valders.

The most significant Wisconsin interstadial, so far as this report is concerned, is that separating the Mankato and the Valders, the Two Creeks interval. This particular interval is important because of the forest beds exposed near Manitowac, Wisconsin, which have been dated by the C_{14} method as 11,400 years before the present. Before the separation of the Mankato and the Valders (Elson, 1957) the Two Creeks interval was believed to separate the Cary and Mankato.

Hough (1959) in his recent compilation of data on the glacial history of the midwest, particularly the Great Lakes region, has proposed three

Table II.

THE SUBDIVISIONS OF THE WISCONSIN STAGE

	Post Cochrane
Cochrane substage	
	Valders-Cochrane interval
Valders substage	
	Two Creeks interval
Mankato substage	
	Cary-Mankato interval
Cary substage	
	Tazewell-Cary interval
Tazewell substage	
	Iowan-Tazewell interval
Iowan substage	
	Farmdale-Iowan interval
Farmdale substage	

new names for the substages of the Wisconsin. These are: the Shelbyville-Iowan for the Iowan, which he believes may be more correctly correlated with the early Tazewell; the Bloomington for the substage that has been called the Tazewell; and the Port Huron for the Mankato. In spite of the fact that the arguments of Hough may be valid in respect to the Shelbyville-Bloomington Morainic System in Illinois, it seems to the writer that less confusion would result by retaining the terminology of Leighton. It is admitted that the recent change in the stratigraphic positions of the Mankato and the Valders has made the new usage of the terms a bit confusing. The fact remains, however, that the Mankato is still a valid designation.

It has recently been suggested that the Iowan in Iowa is actually younger than the Farmdale (Ruhe, Rubin and Scholtes, 1957, p. 671–89; Ruhe and Scholtes, 1959, p. 587–88). This correlation, based on radiocarbon dates, has been severely criticized by Leighton (1958, p. 288–309; 1959, p. 596–98). The writer is not in a position to evaluate these two points of view, but they are mentioned here to point out that the classification, as used in this report, may be in doubt. Further study will be necessary to establish the exact position of the Farmdale interval.

Radiocarbon Dating

Much of the confusion in regard to the Wisconsin stratigraphy, particularly in the classical midwest, has been brought about by the development of a radiocarbon (C_{14}) method of ascertaining the absolute age of organic materials buried in, or associated with, the glacial deposits. This method was developed by W. F. Libby and was first used about 1937 to date archeological materials. The date determination is based on the constant rate at which an organic body loses its radiocarbon (C_{14}) after it ceases to have life. The dating processes and techniques have been developed to the point that the results obtained are now accurate enough (within certain limits) to make this method most useful. The radiocarbon content gives the most accurate dates up to 20,000 years before the present (BP) but is considered by many to be relatively reliable between 20,000 and 35,000 years (BP).

With the development of this dating procedure, it immediately became a tool for use in the establishing of the Pleistocene sequence. In areas where the chronology was already known, for example in the midwest, dates were obtained to establish the absolute, as well as the relative, ages of these deposits in order to set up a basis of correlation by this method. Dates were also made on deposits in areas in which the age relationships were not so well known in an effort to correlate them with the established stratigraphic sequence. As a result, many data have been collected relative to the absolute ages of various deposits containing plant and animal remains associated with the glacial drifts.

As might be expected, confusion resulted from the new method. In the first place, dates obtained in the early days of the application of this method were not always reliable and a wide range of dates was often obtained from the same deposit. With the improvement and further development of the method, and with the improvement of equipment and techniques, these errors have been greatly reduced, and the former criticism based on this factor is no longer valid.

Secondly, dates obtained from deposits in areas where the sequence had been established prior to radiocarbon dating did not always agree with supposed chronology. This meant that either the prior sequence had to be changed or that the dates had to be ignored. Naturally there were some geologists that agreed with the old established stratigraphy whereas others took the radiocarbon dates as fact and immediately set about to formulate a new sequence. This debate is still in progress and at present there are many deposits, the relative ages of which were believed to have been established, that are now in very doubtful positions. It should be pointed out, however, that there are many deposits where the establishing of a date by the radiocarbon method has been most helpful. In these cases, the older stratigraphic units have been given more useful meaning by the dating process.

Probably the most frustrating factor concerning the dating results from the fact that it is impossible to obtain dates unless organic material is available from the deposits. And the discovery of deposits containing buried wood or peat is fortuitous. The organic material, in many instances, occurs in such widely separated areas that the correlation of them is not valid without knowledge of the area between. A more discouraging factor is that in large areas where it has been impossible to do correlation by stratigraphy alone there seem to be no deposits containing organic materials suitable for dating. This is true for the whole St. Lawrence valley area as well as for most of New England. The one exception to this statement is the recently obtained date for the fossil shells of the Champlain Sea deposits.

The writer is not discouraged by the present confusion. It will take much time to assimilate the data of the past few years. And future discoveries will doubtlessly result in a more precise correlation and classification.

THE GLACIAL SEQUENCE IN NEW ENGLAND

New England was undoubtedly covered by ice sheets before the Wisconsin stage, as evidenced by older drifts to the south and southwest. Drifts, probably as old as the Kansan, are recognized in New Jersey and Pennsylvania, and the logical assumption is that New England was also glaciated at that time. The surface materials on the islands off the coast of New York and New England (Nantucket, Martha's Vineyard, Fisher, Block and Long islands) have been, at least tentatively, correlated as early Wisconsin, maybe as old as Iowan. The older drifts, beneath the surface on most of these islands, are believed by many to be pre-Wisconsin in age, perhaps even Kansan or older. The early investigations of Veach (1906), Fuller (1914) and Woodworth and Wigglesworth (1934) lead them to conclude that all four glacial stages were represented by the drifts on the islands. Crosby (1928), Fleming (1934), MacClintock (1934) and Flint (1935b), however, have questioned these correlations and have suggested that there may be drifts of only one pre-Wisconsin stage. In the Boston and Cape Cod regions of Massachusetts, and in central Connecticut and Rhode Island, exposures showing two till sheets, one on top of the other, have been found. The lower of the two tills is generally considered to be pre-Wisconsin because of the great depth of the weathered zone formed at the top of the older drift. The correlation of the older drifts is only speculation (Flint, 1957, p. 359) but they serve to show that there are older glacial deposits in New England.

Except for scattered areas such as those above, the glacial surface in New England was deposited by Wisconsin glaciers, and the drifts of earlier stages, assumed to have been present, were removed by subsequent fluvial or glacial erosion, On Long Island, the Ronkonkoma moraine, trending in a northeast direction through Block Island, Martha's Vineyard and Nantucket Island, and the Harbor Hill moraine, traceable northeastward through Fisher Island to Cape Cod, mark the outer margin of a Wisconsin substage. The exact age of these moraines is still under debate but they are presumed by many to be early Wisconsin, possibly Tazewell or Iowan. This ice invasion covered the whole of New England, and the glacial surface material is either of the same age or younger throughout the whole region. Striations, grooves, boulder trains and eskers indicate that the advance or advances crossed the New England states in a south-southeast to southeast direction.

Correlation of the drifts north of Long Island has been most difficult. It has been impossible to apply the same methods of correlation in the northeast that have been employed successfully in the midwestern states. Chiefly because of the irregular topography and the resulting stagnation of the ice, as is explained later in this report, few frontal moraines were deposited north of the coastal areas. The tills in most upland areas are so thin that a study of their characteristics is almost impossible or at least impractical. The valley drift is predominantly glaciofluvial. The tills that are in the valleys were deposited by ice moving parallel to the valley because the topography had caused a change in the direction of movement.

As yet material recovered suitable for radio-carbon dating north of Hartford, Connecticut has been too scanty to be of much value, and for the same reason pollen analyses have not been conclusive. It is believed, however, that the tills of northwestern New England are younger than those to the south. It also seems that most geologists would agree that the drift in southern Vermont and New Hampshire is of Cary age or younger and that the Mankato is probably present at least in extreme northern sections of these states. Flint (1945) implied that the drift represented by the moraines on Long Island extends as far north as Hartford, Connecticut, and the assumption has been that these were deposited by pre-Cary ice. It seems, therefore, that at least three substages of the Wisconsin are represented in New England.

A few studies of pollen from peat deposits in New England have been made in recent years, and these have given some very interesting results. One of the most recent pollen studies was made by Ogden (1959, p. 366),

and two aspects of his results should be noted here. In the first place, Ogden's location on Martha's Vineyard was beyond the terminal moraine of the Wisconsin ice (Harbor Hill-Charlestown-Buzzards Bay-Sandwich moraines), and the pollen sequence should possibly record the climatic variations of the whole epoch. Secondly, Ogden summarized the previous works of Deevey (1943, 1949, 1951), Leopold (1955, 1956) and Davis (1957) and correlated his results with these earlier studies.

According to Ogden (1959, p. 370–376), three climatic extremes are indicated by the pollen sequence on Martha's Vineyard. This supports the geological point of view stated above. He does not, however, interpret his results as indicative of the same age relationships. Ogden (p. 369) assigned no age to the deposits of the Buzzards Bay (Harbor Hill) moraine, but he dated the drift of the whole of New England north of Durham, Connecticut as Port Huron (Mankato), and he postulated that the Valders terminus is north of New England in Canada. The implication here seems to be that the drift on the off-shore islands and as far north as Durham, Connecticut is Cary or older and the remainder of the New England deposits are Mankato. This correlation, as already implied, does not agree with the current geological point of view since most geologists doubt that the Mankato drift extends so far to the south.

Although the geological evidence is still quite scanty, many geologists believe that the drift north of Connecticut, at least as far north as Vermont, is Cary in age. Flint (1956), for example, suggested that the ice advance reaching Middletown, Connecticut, represents the Cary Maximum. It seems that the chief difference between Ogden and Flint lies in their interpretations of radiocarbon dates from deposits near Middletown and Durham, Connecticut. According to Flint, this re-advance occurred "before, but not long before, 13,000 years ago" and this, he stated, is in "reasonable agreement" with the Cary of the Midwest. It is apparent that Ogden believed that the dates from Durham agree more favorably with the Mankato. There are geologists who are in more agreement with Ogden than with Flint on these interpretations, MacClintock (1954) found two tills in the St. Lawrence valley, and these are correlated tentatively with the Cary and Mankato. At this particular time, it seems inadvisable to try to use the classical midwestern stratigraphy for correlation in New England.

The correlation of the lake episodes of the Connecticut Valley and the Champlain Lowland (and the adjacent areas) is still in a state of confusion. It has been recently established, however, that the marine invasion of the St. Lawrence valley and the Champlain Basin occurred during the Two Creeks interval and not after the Valders as formerly believed. The writer believes, from the findings of the survey now in progress, that the lake episodes of the Champlain Basin (Lake Vermont) followed the last glacial recession in this section and that the lake sequence in the Connecticut Valley may possibly be associated with a different glacial retreat. The results of the work of MacClintock (1958) in the St. Lawrence Valley would definitely support a conclusion that the Valders ice did not extend as far south as New England. These correlations are only tentative, and the evidence will be discussed later in this report.

The glacial drift is generally thin or lacking over the uplands and large areas of bedrock are exposed at the surface. This makes correlation even more difficult since the valley deposits, which are thicker, are only conspicuous as kame terraces banked against the lower slopes of the valley sides.

The discussion of the glacial chronology in North America and New England, in this and the preceding chapters, serves to point out some very significant facts that are helpful to the understanding of the problems of correlation in Vermont. First of all, the glacial stages have been defined in the midwestern area, but the till sheets have not been traced eastward into New England. The correlation of the older drifts in the northwest with the classical stratigraphic sequence has not, as yet, been possible. Secondly, recent radiocarbon dates of deposits in the areas of the classical Wisconsin have been confusing and the proper place of many of the deposits is still in doubt. Thirdly, efforts to correlate the deposits of the last glacial stage with the classical Wisconsin have met with little success in New England. Fourthly, systematic mapping and survey programs have been so limited in extent and detail that data on New England are quite incomplete.

THE GLACIATION OF VERMONT

Like other regions of New England, it seems logical to assume that Vermont must have been glaciated by ice invasions in the pre-Wisconsin stages and the early Wisconsin substages. The manifestations of these earlier episodes, as elsewhere in New England, have been eradicated by later invasions, and drifts of late Wisconsin only are present. The age relationships of the till sheets are facets of the glacial geology that can be only tentatively discussed at this time. The various problems associated with this question will be discussed later.

With few exceptions, there are no terminal or recessional moraines in Vermont or, as already stated, in all of New England. This, of course,

makes the identification of ice margins most difficult. The absence of moraines also complicates the problems of recognizing multiple till sheets since the detritus carried and deposited by an ice invasion may have few characteristics that were not shared by the one that preceded it.

Organic materials, paleosols and interglacial gravels, often valuable markers in the identification of till sheets, are also rare in Vermont. None of these has been seen during the current survey. It is interesting to note that the only records of buried wood that has been found in the state are contained in the writings of Thompson (1853) and Hitchcock (1861). As noted by Hitchcock, some of the wood specimens from the Burlington area were placed in the State Museum in Montpelier. At least a part of these are still in the museum but, as yet, there is insufficient data available concerning the geologic environment in which they were found to warrant having them dated.

Glacial Advance and Retreat

The most common concepts concerning the advance and retreat of continental glaciers have been formulated in regions of low, relatively level to undulating relief such as is exemplified in the Central Lowlands of the midwest. In most of these areas, the ice, with little obstruction, moved more or less in straight lines to a terminus. Here it maintained a relatively stable front for a period of time during which the forward movement of the glacier and wastage due to melting were approximately equal. It was during this interval that the terminal moraine marking the maximum extent of the glacier was deposited. During retreat, the ice melted back en masse (from top to bottom) by the ablation process, and any stationary periods of quiescence were accordingly marked by recessional moraines. The low relief, with no major irregularity, made possible the deposition of features that now record the recession of the ice.

Glacial Advance over Rugged Topography

As has formerly been noted by Flint (1930a) and Goldthwait (1938), the concepts described above cannot be adequately applied in an area of rugged relief such as that found in New England. Many earlier theories concerning the relationships of glacial movement and the terrain proposed that the direction of the ice movement over an area more or less controlled the resulting topography. This misconception was dispelled by Horberg (1955) from his studies on Isle Royale in Lake Superior and in the Rove area of northern Minnesota. Glacial ice, the writer is convinced, is most sensitive to the irregularities of the surface over which it moves and the ice direction at the base of a glacier may be greatly modified by bold, irregular relief without greatly affecting the general, overall direction of movement of the upper portions. As a result of the studies that have been made of the ice directions across Vermont, the writer here submits an hypothesis dealing specifically with the advance and retreat of continental glaciers over areas of rugged topography.

In Vermont, it is believed that at least one of the ice invasions moved over the state in a southeasterly direction. This ice, of necessity, had to pile up in order to overtop the Green Mountains that form a north-south barrier the entire length of the state. The ice that crossed the mountains, however, was spread over an area of valley and highland topography. Striations show that the basal ice of this glacier moved more or less parallel to larger, well established lines of drainage, with little or no relationship to the southeasterly movement of the ice mass. It does not seem to the writer that this fact is difficult to understand inasmuch as the thickness of the ice was great enough for pressure to cause independent movement. The ice in the Dog River valley south of Montpelier (elevations 500 to 700 feet), for example, was in a deep ravine between Northfield Mountain on the west (elevations 2000 to 2500 feet) and the higher ground to the east (elevations 1500 to 2400 feet). The basal ice was not greatly influenced by the southeasterly direction of the upper portion of the glacier inasmuch as it was located on the lee side of the Green Mountains (elevations approximately 4000 feet). The weight of the overlying ice, none the less, was great enough to cause movement of the bottom ice which followed the path of least resistance and moved southward through the valley. The ice direction was thus changed from approximately N40°S to due south. It is apparent that the basal ice followed the course of most of the major lines of drainage in this manner, and for this reason most valleys were scoured and modified into the U-shape so characteristic of those carved also by valley glaciers.

The evidence of this "two-directional" movement of the ice has been on record for over a hundred years. It is the writer's belief that it was this evidence that prompted Hitchcock (1861, p. 66–87) to report two agencies causing striations on the bedrock which he designated a drift direction and a glacier direction. It was not possible, it seems, for him to reconcile such diversified movements of the ice and thus he suggested valley glaciers as a second agency since striae along the streams usually parallel the valley. Dana (1871, p. 233–43) also noted and emphasized the fact that the ice moved more or less parallel to the valleys.

It is probable that independent ice movement in the valleys took place

in the early stages of the advancing phases of glaciation near the ice margins. During this stage of the advance, pressures were not equalized in all directions since the ice thickened in the direction from which the glacier invaded. Movement was therefore in the direction of less pressure, in the general direction of the ice margin. After the upper, main body of the glacier had moved over the area, pressures were stabilized and the devious movement of the basal ice ceased or was greatly reduced.

The writer is convinced that, with certain modifications, this concept of ice advance is compatible with the mechanics of glacial motion described by Demorest (1942, p. 95–100). It can easily be visualized that the surface slopes would be quite irregular as the glacier moved over an area of rugged relief and that obstructions to linear flow would be encountered, particularly by the basal ice, all along the routes of advance.

Deglaciation in Regions of Bold Relief

The effects of irregular surfaces on the advance of a glacier are not as significant, geologically speaking, as the net effect on glacial waning. According to the hypothesis here developed, it is impossible for a moving mass of ice to retreat in an area of rugged relief in the same manner as it would in a plains region.

Flint (1930, p. 56–69) proposed that ice stagnation (en masse) was necessary to explain the glacio-fluvial and glacio-lacustrine deposits in Connecticut. After criticism of certain aspects of his report, and particularly those of Alden (1931, p. 172–74), Flint admitted that he had been somewhat carried away by his new hypothesis. In subsequent writings, however, Flint (1931b, p. 174–76; 1932, p. 152–56) has so revised his original concepts that the emphasis on stagnant ice has been almost completely eliminated. Goldthwait (1938, p. 345–72) strongly supported the stagnant ice hypothesis although he did not suggest how such stagnation might have occurred. He also conceded that the last ice in the Connecticut valley must have retreated northward as Antevs (1922, p. 65–101) had proposed from varve studies.

The writer agrees that the original hypothesis by Flint was in error in the proposed idea of complete stagnation of the glacier and the uncovering of New England from north to south. It seems, however, that the evidences presented to show that much of the ice was stagnant during the deposition of much material by glacio-fluvial action were, and still are, valid. The condition and position of much of the drift in Vermont seems definitely to indicate stagnant ice as do those in New Hampshire described by Goldthwait (1938 p. 345–57).

This report contends that the upper, major portion of a glacier did recede from south to north in progressive steps by ablation thinning, and may have fluctuated or even halted during retreat. The basal ice, however, could not so retreat because it was trapped in the basins and valleys between high divides and here it had to remain. As the upper active ice waned above the trapped basal portion, pressures may again have been so unbalanced that the basal ice was reactivated. The basal ice therefore may have moved again in the direction of lessened pressure, away from the direction of the upper ice retreat. It is even possible that movement in the lower portions may have been increased by the fact that it was nourished by the retreating upper mass. In this case, a valley ice lobe might have been active several miles, or even tens of miles, downstream from the margin of the waning upper ice, provided the mass of the glacier was still in an upstream segment of the valley.

It is the opinion of the writer that it was in this manner that the glaciers that covered Vermont were gradually thinned by ablation and were reduced to a thickness equal to the heights of the higher elements of the topography. It may be deduced that the valleys were the last to be uncovered. In general, this is true, but not necessarily so since temperature variations between high and low altitudes in some areas may have been great enough to allow valley ice to melt first. Rounded cobbles and boulders on the slopes of the higher areas, particularly in the southeastern part of the state, suggest water transportation down the slope and tend to support the latter possibility. The kame terraces on the slopes of the Green Mountains, particularly in the vicinity of Mt. Mansfield, however, would surely show that the ice in this region melted from the top down.

The above hypothesis is not entirely new, for evidence has been cited by several former studies which tends to substantiate the belief that the ice in New England did not retreat in a "normal" manner. The concepts here presented are modifications of, and additions to, those of Flint (1930a, p. 56–69; 1931b, p. 174–76; 1932, p. 152–56), Goldthwait (1938, p. 345–57) and Rich (1943, p. 95–100).

Glacial Erosion

It is difficult to make a discussion of the subject of glacial erosion quantitative, since any estimate of the total removal by glacial activities would merely be supposition. Certain generalizations can be made, however, and this treatment of the subject is designed to point out some of the current concepts, and to clarify certain aspects of glacial erosion that are often misrepresented, particularly by the non-geologist.

PLATE III

Figure 1. Glacial striae on the crystalline bedrock near Lampson School two miles north of Brookfield.

It has already been implied that the major aspects of the topography were not greatly changed by the erosion of the ice during glaciation. If it were possible to construct a map of the physiography of Vermont prior to the Pleistocene Epoch, it is believed that the outcome would be essentially the same as the present map (Figure 2). That is to say, the Green Mountains, the Taconics, the Northeastern Highlands, etc., had essentially the same boundaries in preglacial time. The details of the topography of each of these units was, as would be expected, modified locally but not to the extent, in most cases, that they now represent distinctly different geomorphic forms. The most conspicuous result of glaciation insofar as erosion is concerned, was the removal of the preglacial rock mantle and the planing off of the outer few feet of the bedrock. As a result, the present bedrock cover is completely new. It is definitely presumed that on the uplands and mountains of Vermont, the rock mantle was of greater thickness before glaciation since the glacial detritus is very thin or entirely absent in these regions. Because weather-

PLATE III

Figure 2. Clayey bouldery till bank 100 feet high. Exposure along Bragg Brook one mile west of Norwich.

ing had been uninterrupted for millions of years prior to the ice age, there must have been a relatively thick rock mantle present, and in spite of the fact that erosion had also progressed continuously during the same eras, the accumulation was no doubt greater than that deposited by the retreating ice.

As stated earlier, the stream valleys were reshaped by the scouring of the ice that moved through them. The valleys probably received the greatest changes that were produced by ice erosion. Here again, however, it is not possible to propose quantitatively the amount of erosion that took place. Any such suggestion would have to be based on a conclusion that the preglacial depth, breadth and shape of each valley could be ascertained and this is not a valid assumption. Modification was prominent because the preexisting stream valleys caused the ice to be diverted in a direction parallel to them, thus concentrating the erosive effects along the courses. That is to say, ice erosion was increased by geologic factors, in no way related to glaciation, that existed before the Pleistocene.

The above statement, concerning geologic factors that influence ice erosion is, in the majority of cases, true of all locally-formed features. Erosive forms such as lake basins, linear hills, mountain notches, and bold relief were formed because of previously existing geologic conditions. The geologic phenomena which, like the stream valleys, influenced the reshaping of the local topographic detail include the structure, fabric and relative hardness of the bedrock, the presence of fracture structures as joints and faults, igneous intrusions and linear mountain chains.

In Vermont, the most common causes of local surface irregularities, produced by ice erosion, are the structure, the fabric and the relative hardness of the bedrock. The fact that the shaping of the topography is generally due to the rock structure was proven by Zumberge (1955). The results of Zumberge's research are as applicable to Vermont as they are to Michigan and Minnesota, except that in Vermont rock fabric is probably more important than is the dip of the rocks. Inasmuch as the rocks are highly metamorphosed, mostly schistose, slaty or gneissic, the linear expression of the local relief carved from these rocks parallels the foliations. The variation in the hardness of the rock similarly aids in the sculpturing of local forms, the more resistant rocks yielding less to erosion than those composed of softer constituents.

The basin containing Lake Champlain is an excellent example of glacial erosion controlled by preglacial geologic conditions. The Champlain Lowland lies between two ancient uplifts, the Adirondack and Green Mountains, and is structurally a down-faulted basin floored by soft Paleozoic sediment. Drainage was here established during a much earlier geologic period and in time, no doubt, a wide stream valley, bounded on either side by rock terraces (Fenneman, 1938, p. 217–22), set the stage for the future excavating of the valley by ice. A combination of factors including the high mountains on either side, the stream valley, the faulting associated with the mountain fronts and the bedrock all contributed to the concentration of the ice erosion at this locality. In a similar manner, according to Jacobs (1921), the Lake Willoughby basin was formed by faulting and the differential hardness of the rock in the elongate zone which it occupies.

The trend of the Green Mountains not only influenced the cutting of the Champlain lake basin but also the erosion of the Champlain Lowland and the Vermont Valley. Since the ice that invaded the lowland moved approximately parallel to the range, ice erosion was thus concentrated. In the case of the Vermont Valley, the Taconic Mountains on the west would have similarily influenced the erosive action.

It might be summarized, therefore, that although the exact amount of

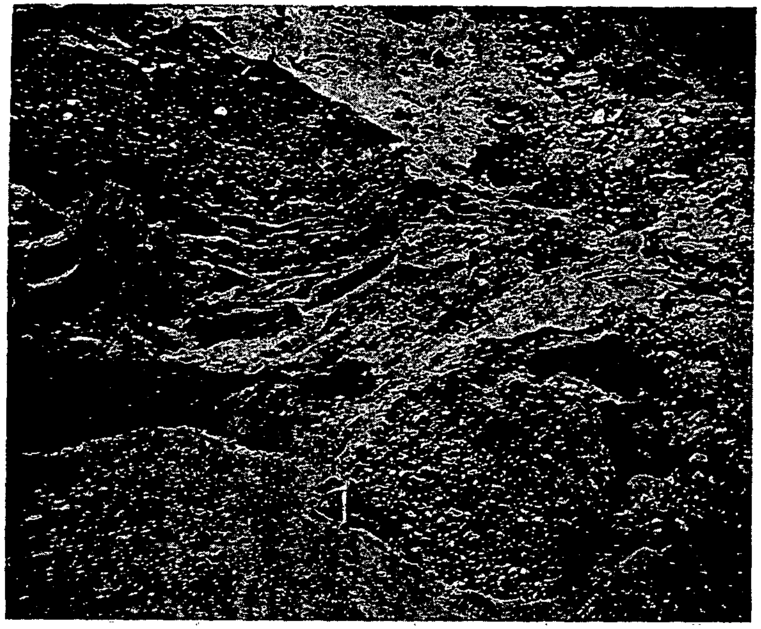

PLATE IV

Figure 1. Kame gravel showing ice-contact structures in gravel pit in kame terrace at White River Junction.

erosion that can be attributed directly to glaciation cannot be computed, all of Vermont was nevertheless subjected to severe erosion by the ice. The major aspects of the topographic features were formed in preglacial times, but the surface was modified locally in varying degree. The reshaping of the land surface by glaciation and the resulting manifestations of ice sculpture were greatly influenced by geologic phenomena that were present before the Pleistocene Epoch.

Glacial Materials

Although glacial deposition is everywhere manifested on the surface, it should be emphasized that in most areas, the drift is not of sufficient thickness to mask completely the topography of the underlying bedrock. Locally of course, particularly in the valleys, the surface expression is that of the glacial deposit, as for example, kame terraces, eskers or spillway gravels. The fact that the drift is too thin to modify greatly the

PLATE IV

Figure 2. Normal fault. Ice-contact structure in kame terrace gravel on Barnes Hill. One and one-half miles north-northeast of Waterbury Center.

bedrock topography, however, does not mean that it does not have geologic significance. The mantle material is indeed important to an understanding of the glacial history but the thinness often makes the interpretation of the deposits difficult.

Glacial drift varies greatly from one area to the next, but there is uniqueness to the tills and outwash of any region, especially in the case of an area as large as Vermont. These materials also have a wide variety of characteristics from one locality to the next. The variations usually result from the composition of the bedrock over which the glacier moved, the mode of deposition, the distance of transport and the amount of weathering since deposition.

Till

Till is the material deposited directly from glacial ice. Since it is spread over the surface, without reworking by water, it is unsorted and contains all sizes of sediment from clay through boulders (Plate III, Figure 2). This fact makes till, in most cases, easily distinguishable from all other sedimentary detritus.

In Vermont till is the most widespread of all of the surface material inasmuch as it covers all upland areas. The surface cover varies in thickness from a very thin veneer to a hundred feet, but it is most commonly found to be less than twenty-five feet. On the uplands the till is generally thin and large areas of the underlying rock are exposed. The composition of the till varies, as is normally expected, with the matrix ranging from clay to sand, and the clayey till being much more widespread than the sandy till. The unweathered till is more commonly grey in color, but brownish tills are often found.

The thicker till deposits are found in the valleys. In most cases these occur in narrow stream valleys that trend in directions that were more or less normal to the movement of the glacier and the valley was filled with debris from the ice. In some localities, however, the trend of the valley does not seem to have been the importnat factor causing deposition, and in these it is difficult to ascribe deposition to any single cause. The best till exposures are in valleys that are situated in regions of higher relief, above the main valley, and this situation no doubt influenced deposition.

It is customary when mapping glacial deposits, to designate the topographic feature containing the till. That is, it is usually possible to note frontal moraines, ground moraines and drumlins, and by definition these are composed of till. It is impossible, however, so to indicate the topographic expression of the till deposits of Vermont. As has already been stated, no significant frontal moraines have been found during the survey now in progress except a very few valley moraines that were too small to map. Thus far it has also been impossible to identify any feature that could definitely be called a drumlin.

The tills of Vermont consequently might be designated ground moraine, for there is little or no surface expression to the deposits, and the detritus was apparently dropped from melting ice during retreat. In a very general interpretation of the term it would be correct to assign the ground moraine designation to all such deposits. In a more restricted definition, however, the term implies till showing evidences of plastering and of having been overridden by the advancing glacier. It is true that there are compacted tills in Vermont. Most of the thicker valley deposits could be so classified. In upland areas, where till is much more widely

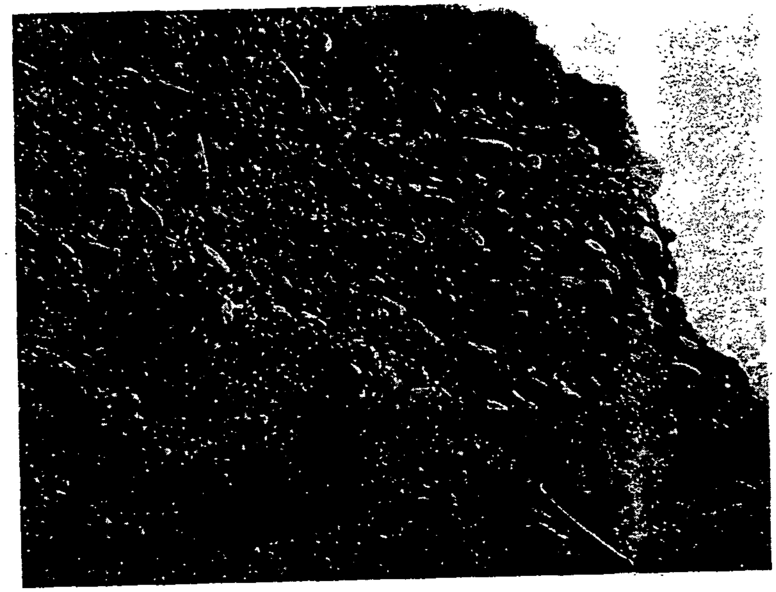

PLATE V

Figure 1. Mantle material showing angular boulders. One mile west of Norwich.

distributed, however, the till is loosely consolidated, and varies greatly in thickness within short distances. In some areas it contains rounded (water worn) boulders and cobbles, and in other areas rounded cobbles and pebbles are strewn over the surface giving a gravelly appearance. The most widespread kind of upland till is composed of a loosely packed, sandy matrix containing a minimum of faceted and striated glacial boulders and a high percentage of angular fragments of local bedrock. These materials do not fit the writer's concept of the term "ground moraine." The only differentiation that the writer feels might have been feasible, would have been a distinction between the compacted clayey valley till and the loosely compacted upland till. It was believed, however, that the time required to make such a separation would have been so great, and the significance so minor, that it did not seem justifiable to do it. The till areas were therefore simply mapped as till, and no other qualifying divisions have been legended on the maps.

PLATE V

Figure 2. Mantle material covering delta built into glacial lake by Bragg Brook. One mile west of Norwich.

Mantle Material

One of the most confusing materials encountered during the surface survey to date is a mantle material that covers much of the surface of the Hanover, Woodstock, Randolph and Barre quadrangles. The till in these areas is usually a bluish-grey, clayey, bouldery material. It is exposed in numerous stream valleys, and particularly in the Hanover and Woodstock quadrangles, outcrops in excess of sixty feet in height are common. The surface of this till, however, is covered with a reddish-brown, sandy, bouldery layer that averages five to ten feet in thickness (Plate V, Figure 1). The rock fragments in this mantle are more angular, more weathered and contain a higher percentage of the local bedrock than those of the underlying till. The mantle modifies the surface expression to form smooth rounded hills or a series of terraces along and usually near the base of the slopes (Plate VI, Figures I and 2).

PLATE VI

Figure 1. Hillside covered by surface mantle giving smooth, gentle slope. North side of Lulls Brook valley, two miles southwest of Hartland Four Corners (Hartland Township).

The surface mantle is distinctly different from the underlying, compacted till. It is evident that movement has occurred and that water was in some way associated with the movement. One of the confusing aspects of this mantle is that its position is not always over till but it has been found covering other deposits. In a gravel pit near the village of Norwich, for example, the mantle covers the foreset beds of a small delta (Plate V, Figure 2). Here the surface material is four to twelve feet thick and masks completely the delta characteristics of the underlying gravel.

The exact mode of deposition of this mantle is not, as yet, completely understood, and it is quite possible that its origin cannot be attributed to any single agency. Three possible origins seem possible. First of all, this material may be ablation till that was deposited by the retreat of the last ice sheet. If it is till, the characteristics of the material suggest that it is either a superglacial till or that it was deposited near the terminus of a thin ice sheet. It may be that it was thus deposited and modified by water during deposition, and if so, it is more or less a transitional deposit between till and outwash. Similar material has been noted in New Hampshire; some writers have described it as till.

PLATE VI

Figure 2. Small terraces formed by surface mantle. Along state Route 12 four miles south of the Ottauquechee River (Hartland Township).

A second possible origin of the mantle is that it was moved by freezing and thawing, probably accompanied by some type of aqueous action. In this case, the material would merely be a modification of the underlying deposit and would thus be unrelated to glacial advance and retreat. The third possible origin is that it was formed by mass movement including creep, slump and solifluction downward from higher ground. It seems to the writer that many of the exposures of this material had characteristics that definitely indicated solifluction. The question that immediately arises then is whether or not this surface marks the periglacial zone of the last ice advance north of this region.

This report concludes that the surface layer may have been caused by two or more of the above mentioned possibilities. Further study, and particularly the mapping of the extent of this mantle, will no doubt eventually prove its full significance if, indeed, it does have geologic importance.

Glacio-fluvial Deposits

Meltwater transportation and deposition during the retreat of the

ice accounts for the most common and best developed features of the state. Although till is more widespread than the glacio-fluvial sands and gravels, the outwash deposits are far more significant inasmuch as it is possible to identify the topographic form of the deposit. In contrast to the till, the fluvial deposits are concentrated in the valleys of the larger streams.

Kame terraces. The kame terrace is, by far, the most common land form made in Vermont during the Pleistocene. These are ice-contact, valley deposits formed along the margins of ice, between the ice and the valley wall, after most of the ice had retreated (Plate IV, Figure 1 and 2). The terraces are common in almost all of the major drainage lines and in some areas they occur high on the slopes of the mountains.

As to texture the composition of the gravel varies greatly. In some areas the gravel is quite coarse whereas in other areas the deposits are predominantly sand. In all areas, the percentage of sand is usually higher than would normally be expected. In the southern part of the state, and particularly in the Connecticut valley, sandy terraces are the most common. The high content of sand makes it difficult to identify definitely the terraces as such. The ice-contact structures contained in the deposit are, the writer believes, the most reliable characteristic for identification. (Plate VII, Figures 1 and 2).

Many geologists have made studies in the Connecticut valley and have designated many of the kame terraces as deltas, lake sediments and even post-glacial fluvial terraces. The present survey has reclassified these as kame terraces because of the ice-contact structures. The writer believes that the fineness of the material in the deposits is evidence of slow moving water rather than lake deposition. It is quite apparent that much of the deposition was in contact with stagnant, residual ice and in water that had a more or less restricted movement. The structures within the deposit, however, indicate flowing rather than stagnant water. In none of the terrace deposits from the Vermont-Connecticut boundary north to Norwich are there structures showing the characteristic long foreset bedding and the lobate frontal margin of deltas built into open waters.

Outwash plains. The outwash plain has not been identified in Vermont by the present or previous surveys. The absence of outwash aprons, of course, is no doubt related to the absence of moraines. As has already been stated, it is believed that the stagnation of the ice below high topographic barriers account for this unique characteristic. The slope of the foothills of the Green Mountains in the southeastern part of the state contain a gravelly surface that in some ways resembles outwash. But in these areas the rounded cobbles are merely scattered over the surface

PLATE VII

Figure 1. Kame terrace at White River Junction.

and are too thin to be measured or mapped. This conspicuous characteristic, the writer concludes, is due to excessive amounts of water produced by the melting down of the ice.

Valley trains and spillway deposits. There is a surprisingly small number of stream valleys that contain gravel deposited by streams flowing out from a melting ice margin. In most of the larger valleys kame deposits are found lining the valley walls high above flat, relatively wide valley floors, but no outwash gravel is present in the bottom. It is true that most of these valleys contained one or more lakes after the retreat of the ice, but there is no evidence that outwash underlies the lacustrine sediment. All evidence has substantiated the belief that there is none.

It is believed that the high content of sand in the kame terraces, and the small amount of outwash (spillway gravel) in the valley bottoms, substantiate the hypothesis that stagnant blocks of ice remained in the valley long after the upper, actively moving part of the ice sheet had receded to the north. The sandy kames were deposited along the margins of the residual ice by meltwater flowing slowly downstream. By the time

PLATE VII

Figure 2. Kame terrace 150 feet high. West side of White River one-half mile south of Bethel.

the stagnant ice in the valley melted, the melting margin of the glacier had receded beyond a divide and the outwash was diverted elsewhere, and did not furnish sediment to the valley floor.

Eskers. Eskers are long, narrow, gravel deposits with a winding or serpentine trend. They are usually considered to be deposits made by subglacial streams or in some cases, crevasse fillings. In general, the trend of an esker is parallel to the direction of ice movement and it is therefore an indication of this directional movement. Except for portions of a large esker in the Connecticut valley, no eskers have been mapped during the survey now in progress. Eskers have been noted by other students as rather common in the northeastern part of the state.

Till Sheets and the Age of the Drift

As discussed earlier, the correlation of the drift sheets in New England remains uncertain. The present survey, however, has revealed data relative to the drift sequence that needs presentation.

Since there are no moraines, buried wood or identifiable till sheets, the only approach to the ice invasions, at this writing, seems to be from a study of the ice directions as indicated by striations (Plate III, Figure 1) and boulder trains. The present survey has mapped striations wherever possible and the recent bedrock studies of Christman (1959 and in press), Dennis (1956 and personal communications) and Hadley (1950) have included studies of the striation directions. Earlier literature, specifically the reports of E. Hitchcock (1861), C. H. Hitchcock (1904, 1908) and Dana (1871), also contains information on the striations with locations that can be plotted on a map. Information on the boulder trains (indicator fans) is available from Flint (1947, 1957), Jacobs (1942) and Hadley (1950). All of the data from these sources were collected and carefully plotted to produce the map showing the ice movements across Vermont (Plate II). There are many areas about which there is little or no information, but the writer believes that the map serves the purpose for which it was intended. It is always possible to question the validity of information found in the older literature. In the construction of the map, however, it was often possible to find striation directions for a single area from more than one of the above sources, and in those cases the different authorities checked with surprising accuracy. The boulder trains, which were described in literature other than that noting the striations, also correlated well with the striations. The map (Plate II) also matches favorably the recent Surficial Geology map of New Hampshire (Goldthwait, Goldthwait and Goldthwait, 1951).

During the summer of 1958, the present survey mapped the Barre and Randolph quadrangles. It was noted at that time that two distinct striation directions were present. On Braintree Hill, between the Third Branch of the White River and Ayers Brook, the striations ranged from S45°W to S20°E. A small boulder train on this hill, containing a pink granite, had a north-south trend. On the divide between the Second and Third Branches of the White River, west of East Bethel, the striations trend S10°E to S15°E. East of East Randolph, on the divide between First and Second branches, the striation directions are S15°E and S25°E.

Due north of the area noted above, on the high ground north of Brookfield, however, striations were found in four different localities trending S25°W, whereas to the northeast of this locality on Northfield Mountain the directions are S5°E and S30°E. The area covered by the two quadrangles, however, was hardly large enough to make any deductions possible. During the following summer (1959), it was discovered (in the Montpelier Quadrangle) that the ice moved southeast on the west side

of Worcester Mountain and south to south-southwest on the east side of this mountain. The Camels Hump Quadrangle was also mapped and the ice direction in this quadrangle is definitely southeast. Christman (1955 and in press) has recently completed bedrock surveys in the Camels Hump and Mt. Mansfield quadrangles. The surface maps published in his reports show striations that leave no doubt that the ice direction across the Green Mountains in this area was S25°E to S40°E.

When the results of the above investigations were studied, it became apparent that two different ice invasions had moved into the Montpelier, Barre and Randolph quadrangles. It is intended that the present survey will continue to collect data on the ice directions but the need for information relative to a discussion in this report, prompted the decision to make the ice direction map (Plate II).

It is apparent from a study of the ice directions on this map that at least two different ice invasions are recorded in Vermont. The most wide-spread of these is that which moved in a southeasterly direction across the state. A more limited invasion, however, did enter the state from the north, in the region of Lake Memphremagog, and moved southward at least as far as the White River and possibly as far south as Brattleboro. It is impossible at this writing to mark the contact between the drifts of the two invasions. It does seem fairly certain that the north-south drift did not cross Worcester Mountain into the Stowe Valley nor did it extend westward beyond the Dog River in the vicinity of Northfield. It also seems that the north-south ice did cross Braintree Hill (in Braintree Township) and that it probably reached the White River in the vicinity of Bethel.

The available data on striations in the Memphremagog Basin and south of it to the Connecticut River are few in number and quite scattered, but the boulder train of the Craftsbury granite attests to the southward movement in that region. It is assumed that the glacier from the north covered most of the Piedmont between the Green Mountains and the Northeastern Highlands at the Canadian border, but in the latitude of Hardwick it extended from Worcester Mountain to the Connecticut River. This glacial advance must have crossed the Connecticut River (in a southerly direction) between Guildhall and Barnet for the New Hampshire surface map shows north-south striations south of the river in that area. Furthermore, the terminus of this ice must have extended as far south as the White River at White River Junction and Sharon and possibly as far west (on the White River) as Stockbridge.

It is difficult to trace the ice direction south of the White River, but the striation map seems to support an hypothesis that the southward advance moved down the Connecticut River valley at least as far as Ascutney Mountain. South of that point, however, the striations showing southward movement are concentrated so near the river that it is impossible to distinguish between striations of the southward moving glacier and those made by ice diverted down the valley by the topography. Certain peculiarities of the deposits in the Connecticut valley seem to suggest an ice readvance to the vicinity of Brattleboro, and these will be noted later in a more detailed discussion of the valley.

At this writing, it is not possible to state definitely that the ice invasion from the north was of a later date than the glaciers from the northwest. C. H. Hitchcock (1908, p. 172–78) noted these ice directions, and he definitely believed that the southerly movement was later than the southeasterly. The fact is, Hitchcock insisted that there were local minor ice movements after the retreat of the ice to the north. He named the ice invasion that moved down the Connecticut valley the Connecticut Glacier because, at that time (1908), he believed that valley glaciers from the White Mountains were the source of the ice. In spite of the mistaken concept of valley glaciers that prompted the name, this report, for the sake of clarity, will refer to the glacier with southward movement as the "Connecticut Glacier".

MacClintock (1958) has mapped two tills in the St. Lawrence valley. The younger of these till sheets, deposited by ice from the northwest, has been named the Fort Covington. The older till, deposited by a glacier from the northeast, has been designated the Malone till. It is now proposed by MacClintock and Terasmae (1960) that the Malone till is of Cary age, and the Fort Covington is Mankato-Port Huron, since both antedate the Champlain Sea deposits that are now dated the same as the Two Creeks interval. MacClintock (1958) traced the outer margin of the Fort Covington drift to the western side of Lake Champlain, just across the lake from Vermont.

It has been assumed by the writer that the southeastward ice invasion into Vermont was the same ice that MacClintock has designated the Fort Covington. At least this must be the case in the northwestern part of the state. The one factor, however, that makes it difficult to correlate all the southeast movement in Vermont with the Fort Covington is the large area which it covers. It is evident from the ice direction map (Plate II) that a southeastward moving glacier crossed the Green Mountains the entire length of the state and completely covered Vermont, at least south of the White River. The map of the Surface Geology of

New Hampshire shows that this was also the direction of the ice across that state south of the White Mountains. The inference here is that the drift of Vermont, except in the region covered by the Connecticut Glacier and the drift south of the White Mountains in New Hampshire, is Fort Covington.

There is one serious problem to such a correlation as that inferred above. In the St. Lawrence valley of New York, the Fort Covington drift, advancing from the northwest, moved only as far south as the north edge of the Adirondack Plateau. This margin is also just a few miles south of the Canadian border and ten miles west of Lake Champlain (MacClintock and Stewart, in press). These facts prompted the writer and MacClintock (personal communications) to assume that the Fort Covington invasion did not extend as far south as the White River in Vermont. The question then is, did the same ice invasion that moved only to the Adirondacks in New York cover a large portion of New England east of the Adirondacks? This report does not suggest that it is impossible, but merely cites the problem.

An ice "readvance" into the Connecticut River valley as far south as the mouth of the Passumpsic, eight miles south of St. Johnsbury, has been noted in the literature by several different authorities. The disturbed varves and the presence of interbedded glacial deposits led Antevs (1922; 1928) to propose a readvance into this area. Crosby (1934) reported two till sheets separated by sand at the site of the Fifteen Mile Falls Dam (Monroe, New Hampshire) and till overlying varved clay just south of Gorham, New Hampshire, twenty-five miles east of Fifteen Mile Falls. Flint (1953, p. 908) suggests that these deposits mark the border of the Mankato-Port Huron drift. Flint also supposes, since no evidence had been found in support, that the border did not extend south of the St. Johnsbury area.

In the course of the present survey, two localities were found where exposures of till over varved clay occur. The most easterly of these is one mile north of the White River along Jericho Road in Hartford Township (two and one-half miles upstream from White River Junction). At this location, sixty feet of till cover the lacustrine sediment. Three exposures of till over varved clay occur along the valley of Lillieville Brook, Stockbridge Township, one and one-tenth, one and four-tenths and one and nine-tenths miles due north of the White River. Gravel over lacustrine sediments in the Connecticut valley in the Bellows Falls-Putney area also seems to be indicative of a readvance.

The writer has not seen the St. Johnsbury deposits but it is assumed that the till covering the lacustrine sediment in the White River valley is similar to that covering the deposits in that area. It is further believed that this ice invasion came from the north (Connecticut Glacier) and was more than a mere readvance of a waning ice sheet. The fact that the direction of the Connecticut Glacier was so different from that of the ice from the northwest (Fort Covington) seems to preclude any suggestion that they are of the same glacial substage. The writer also suggests that this ice advance may have moved down the Connecticut River as far as Brattleboro.

It seems logical to conclude that the southeasterly ice advance immediately preceded the lake in the Connecticut valley (Lake Hitchcock), as well as those of the Winooski valley and the Champlain Lowlands. For this reason, it must have occurred before the Two Creeks interval. According to Flint (1956) this glacial episode would be correlated with the Cary. Ogden (1959), however, interprets much the same data as proving Mankato-Port Huron. The most important aspect of this attempted correlation seems to be the relationship of the Connecticut (glacier) drift to the Champlain Sea deposits. Depending upon these relationships, the invasion from the north might be as young as Valders and probably no older than Mankato.

THE GLACIATION OF THE CONNECTICUT RIVER VALLEY AND ITS TRIBUTARIES

The Connecticut River forms the eastern boundary of Vermont from the Canadian border to Massachusetts. The river flows generally southward from Beecher Falls, where it enters the state, to South Lunenberg. From this locality, the river has a southwesterly course to Barnet and thence southward again, veering south-southwest between Wells River and White River Junction. From White River Junction the river flows almost due south to the Massachusetts line. The south-southwest trend of the river is most conspicuous on both the Vermont and New Hampshire state maps as this accounts for the greater breadth (east-west) of Vermont at the north and the correspondingly greater breadth of New Hampshire to the south.

From the Vermont-Quebec boundary southward to Barnet the Connecticut River is entrenched into the crystalline complex of the Northeast Highlands, a low westward extension of the White Mountains uplift of New Hampshire (Jacobs, 1950, p. 82–4). The knowledge of the geology of this particular region was quite limited until the bedrock surveys of the past few years and the results of these are, as yet,

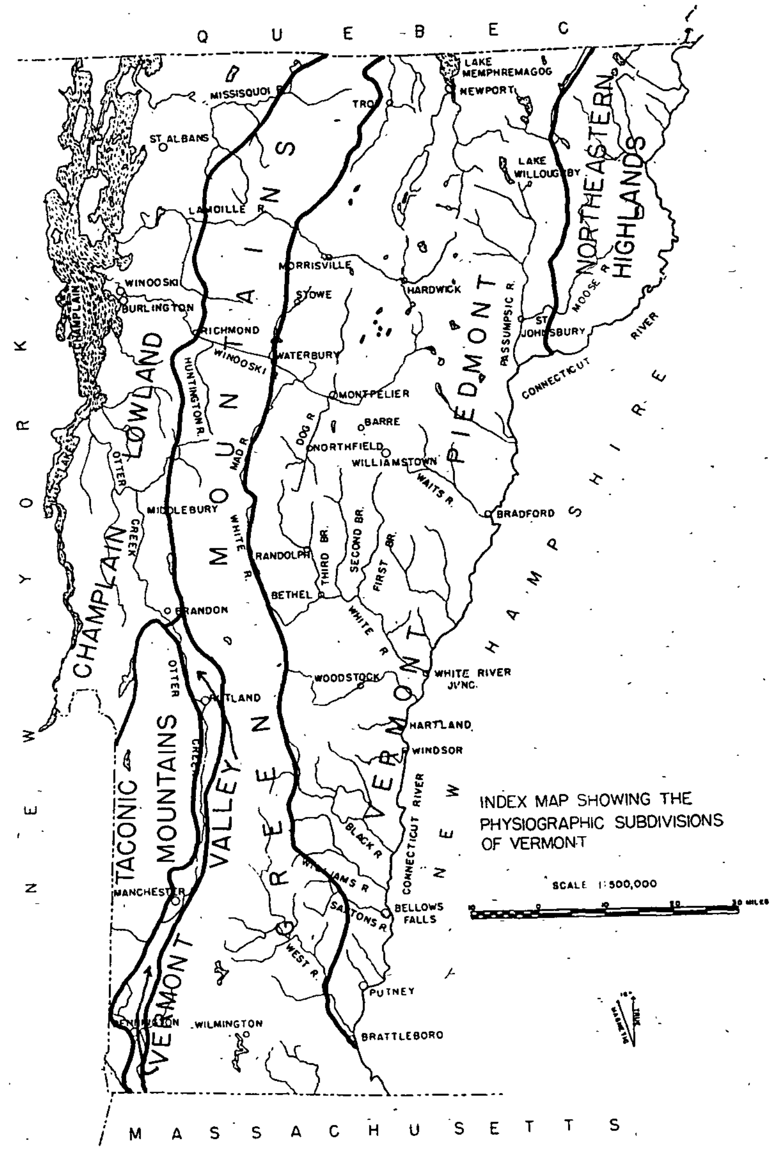

Figure 2. Index map showing the Physiographic Subdivisions of Vermont.

unpublished. To the writer's knowledge, no surface investigations have been made along this segment of the river, and hence the glacial geology must remain a problem for future study. One brief reconnaisance was made into this part of the valley and manifestations of glaciation were found in all localities, but time did not permit a study of their characteristics. The valley here is rather broad, and the valley floor quite level. The stream contains much less water than farther south, but it meanders back and forth across the valley as does an old stream. The overall appearance of this part of the valley gives the impression that it is in an old age stage of development, but it is probable that these characteristics are a result of glacial rather than stream action.

From Barnet southward, the Connecticut River has cut its valley into the metamorphic rock of that plateau-like region designated "The Vermont Piedmont" by Jacobs (1950, p. 79–82). The upland is composed of metamorphosed sediments containing plutonic intrusions, and the structure is complex. The surface is often described as a dissected and glaciated uplifted peneplain. The valley that the Connecticut River has cut into the crystalline complex varies in its characteristics with the rock that it has encountered. The valley is narrow in some reaches, with steep, almost vertical walls, and in other sections the valley widens to a few miles; but even in the widened segments the walls are quite abrupt and steep. It should be noted that the Connecticut Valley Lowland is to the south of Vermont, and that the river valley is in the area about which this report is concerned is better described as a bedrock channel that has been modified by both glacial erosion and deposition. The tributary streams of the Connecticut have also carved their valleys into the crystalline rocks of the Piedmont and their characteristics are much like those of the master stream. In Vermont, these streams flow generally southeastward from the foothills of the Green Mountains.

The Glaciation of the Connecticut Valley Proper

The last ice that invaded the Connecticut valley, as has been proposed earlier, apparently moved into the valley from the direction of Lake Memphremagog (Connecticut Glacier). The striations and boulder trains seem to show that the glacier maintained its course down the valley at least as far as Ascutney Mountain and possibly to Brattleboro. The assumption has been made that the Connecticut glacier may have been a later advance than the ice from the northwest (Fort Covington?) but this hypothesis is yet to be tested. The fact is, it is yet to be established whether or not the southerly ice advance was not a readvance, or a fluctuation, of a preceding glacier, but this report proposes that it was

of a different substage because its striae have a different direction and it covers a large area (Plate II).

Speculations concerning the ice episodes is a most interesting subject relating to the Pleistocene history, but the solution of these problems is not prerequisite to a discussion of the glaciation of the valley. Regardless of the number of glaciers, their direction (or directions) and their relative ages, the whole valley was invaded by ice that moved southward more or less paralleling the present stream. This fact is paramount. Whether or not the general glacial advance was in the direction of the valley, the ice was directed downstream by the valley topography. This is not a new concept nor is the belief that the basal ice moved in a direction independent to that of the glacial advance. Dana first proposed this hypothesis in 1863 (p. 544) from his studies in the Connecticut valley. Later (1871, p. 233–43) he supported his arguments with data collected from surveys of stream valleys all over this section of New England.

As ice moved down the valley, glacial scouring was very active. The valley was widened and deepened, the walls were steepened and the result was a characteristic U-shape similar to that carved by valley glaciation. The amount of widening and deepening that was accomplished was dependent upon the resistance of the rock encountered. Rock resistance varies greatly, and therefore the net results of glacial erosion also varied greatly from place to place. The depth of the bedrock below the sedimentary surface of the valley floor is not too well known and it is very difficult to estimate. In reaches where falls and rapids occur in the river, such as at Bellows Falls, the bedrock is quite near the surface. In other areas, scouring was much deeper, and the thickness of the sediments may be as much as 100 feet or even more. The water wells of the village of Windsor, for example, penetrate the sediment 80 feet without reaching the bedrock. Hadley (1950) reports 50 to 60 feet of drift in the valley at Hanover, and Crosby (1934) records the same depth at the site of the Comerford hydroelectric plant. The course of the valley may have been modified inasmuch as the glacier would have tended to straighten any sharp bends that the course might have contained.

Glacial Waning in the Valley

As has already been inferred, stagnant or quiescent blocks of ice remained in the Connecticut valley after the rest of the glacier had melted from the area. It is believed that the residual ice may have been more or less active after it was uncovered inasmuch as it might have been nourished by thicker ice upstream.

There are many characteristics of the deposits in the Connecticut valley that suggest stagnant ice. These are best known to the writer along the river from Norwich south to the Massachusetts-Vermont boundary inasmuch as this area has been mapped during the present survey. The discussion of these manifestations will therefore be restricted to that segment of the valley.

The most common glacial deposit of the Connecticut valley is the kame terrace. The most conspicuous characteristics of this detritus are the high degree of sorting and ice contact structures. The gravel is in well sorted layers, and the different layers are individually of a wide variety of sizes, ranging from silt to very coarse cobbles or boulders. The sizes of the clastic grains deposited are, of course, proportional to the velocity of the water transporting the sediment. This indicates that the velocity of the water varied from time to time, perhaps from season to season. The kame terrace deposits of the Connecticut valley, however, contain a very high percentage of finer materials, and deposits containing coarse gravel are quite scattered. Considering the great quantity of the kamic material in this area, one would expect to find numerous gravel pits all along the river. This is not the case, however, since most deposits contain too much fine sand and silt to be useful even for highway construction. It is evident that gravel of this texture was deposited by water that did not attain high velocities.

Because of the fine textures of the materials contained in the gravel, these deposits have been described by former writers as many different kinds of topographic structures other than kame terraces. Most commonly they have been designated deltaic, lacustrine, or in some cases fluvial. Such designations do not fit the terraces, however, inasmuch as they contain the ice-contact structures which identify them as glacio-fluvial in origin. This means that the materials were deposited in contact with glacial ice that later melted, and structures were formed as a result of slumping due to melting. The deposits also show bedding structures which do indicate currents down valley. Some glacial geologists have employed the term ice-contact delta for certain similar deposits both here and elsewhere. Ice-contact deltas are known in many places, but they are not found in the Connecticut valley. The deposits do not show the characteristic delta foreset beds of large amplitude and lobate lakeward margin. In contrast, the terraces here have a pitted surface and ice-contact back slopes. In almost all cases where a deltaic classification had been used, the present survey was able to prove that the deposit was kamic. The use of the term delta for the deposits in this area dates back over one hundred years. It was used freely by Edward and C. H.

Hitchcock and the usage is still common with New England geologists. The writer of this report feels that the term ice-contact delta could well be eliminated in this particular region.

A second evidence of ice stagnation in this valley is found in the type of sediment found on the valley floor. If the great thickness of ice that covered this area had receded en masse, the valley should contain large accumulations of till or outwash or both. The writer can think of no reason why this would not be true unless a lake expanded along the retreating ice margin in which case other distinctive characteristics would be present. One of the unique features of the valley deposits, however, is the absence of till and outwash.

Had the glacier in the valley retreated en masse and had the lake waters that later filled the valley expanded with the retreating ice margins there should be evidence of this in the valley deposits. In the first place, a considerable number of large fragments (cobbles and boulders) from the melting glacier should be found included in the lake sediment. In the western half of the St. Lawrence Lowland, where lake waters definitely filled the basin in front of the retreating glacier, the lake sediment contains glacial cobbles and boulders in large numbers. In some of these areas (St. Lawrence Lowland) the lake sediment is actually a stratified till and could be mapped as bouldery lake sediments. Under these conditions the ice calved into the lake, and the boulders were distributed by floating ice (MacClintock and Stewart, in press). Even if the lake water was not deep enough for calving to occur, the drift from the ice margin should either lie below the lake sediment or be included in it. In the Connecticut valley, the lake sediments are usually found directly overlying bedrock with no boulders or cobbles in the lower beds and no detritus between the bedrock and the lacustrine materials.

A second negative factor which does not seem to be compatible with the "normal" glacial retreat theory is the rare occurrence of lake sediment interfingering with the kame gravel. When a lake occupies a basin of glacial retreat and a portion of the lake is bounded by retreating ice, the lake sediment should interfinger with the kame gravel along the ice-contact. The outcrops, where such have been noted in the Connecticut valley, are very few. More often than not, the lake sediment overlaps the kame structure. It is interesting to note that the most pronounced interfingering of lacustrine and glacio-fluvial sediments is just north of the village of Ascutney in a gravel pit in the so-called Connecticut valley esker. This ridge of gravel has been identified as a portion of the esker that extends for 24 miles in this section of the valley, crossing several times from one side of the river to the other (Jacobs, 1942, p. 42). Here is an example of a subglaclal deposit that was in the making when the ice finally melted from the valley, and the lake waters could expand in front of the retreating margin.

Equally important to this discussion of stagnation is the almost complete absence of slumping structures in the lacustrine sediment along the contact with the glacio-fluvial gravel. Had the development of the kame terraces taken place contemporaneously with a massive retreat of active ice from the valley, interfingering would have occurred, as described above, and slumping would have followed, after the ice melted. Such slumping should have affected both the kame gravel and the lacustrine material. But, as has already been stated, this is not a characteristic of the Connecticut valley deposits.

At the mouth of the Saxtons River, immediately to the south of the village of Bellows Falls, is a deposit that should be noted as an exception to the above statements about interfingering and slumping. Just to the south of the river, about one mile south of Bellows Falls, there is, exposed in road cuts along U.S. Route 5, a deposit composed of a variety of sediments of both lacustrine and glacio-fluvial origin. The kame gravel is rather fine texture and the lacustrine sediment is laminated silts and clays. The lake and kame beds do interfinger, and there are structures that indicate slumping (ice-contact) in both materials (Plate VIII, Figures 1 and 2). This deposit, however, is more complex than merely kame and lacustrine deposition. A small gravel opening, below the level of the road (U.S. Route 5) exposes cross-bedded, pebble-sized gravel that resembles deltaic materials below the lacustrine sediments, but the opening was too small to make positive identification. The whole deposit is associated with one of the largest and deepest ice block depressions that has been noted during the present survey. U.S. Route 5 crosses the Saxtons River over the old concrete bridge at the southern limit of Bellows Falls. Just west of the bridge, the river flows through a short, narrow, deep gorge cut through a ridge of resistant meta-igneous rock. Immediately to the west of the gorge and the rock ridge, the river valley opens into a large ice block depression that is four-tenths of a mile long (up and down stream) and six-tenths of a mile wide. The depression has almost vertical walls that rise 80 to 150 feet above the floor of the depression. It is the writer's belief that stagnant ice blocks remained in this basin for an even longer period than the stagnant basal ice in the Connecticut valley. This block, plus other

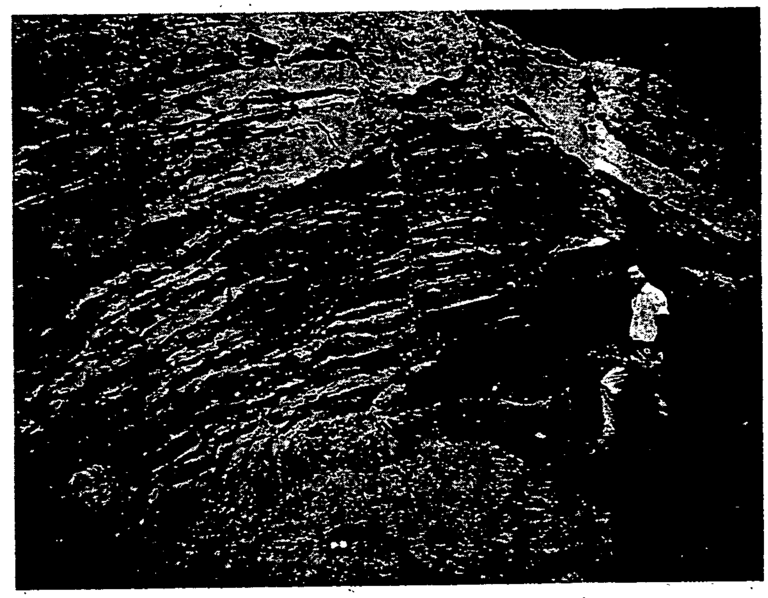

PLATE VIII

Figure 1. Ice-contact and slumping structures in laminated lake sediments. Along U. S. Route 5 one mile southeast of Bellows Falls.

stagnant basal ice in the Saxtons River valley, apparently did not permit the lake waters to enter the valley at all, inasmuch as no lacustrine deposits are found upstream (up the Saxons River) from Bellows Falls. The gravel at the mouth of the Saxtons River is another deposit that other investigators have classified as deltaic. It seems improbable that this could have been a delta since the Saxtons River was blocked with ice and the huge block that formed the depression did not even let the lake waters enter.

Turning now to the evidence against the ice stagnation hypothesis, the most difficult of these to refute is the Antevs (1922) varve count interpretation. According to Antevs, there is an equivalent of 4100 double varves, which he studied and correlated, between Hartford, Connecticut and St. Johnsbury, Vermont. This, said Antevs, proves that the ice in the Connecticut valley retreated from south to north and that it took 4100 years for such retreat to take place. The writer

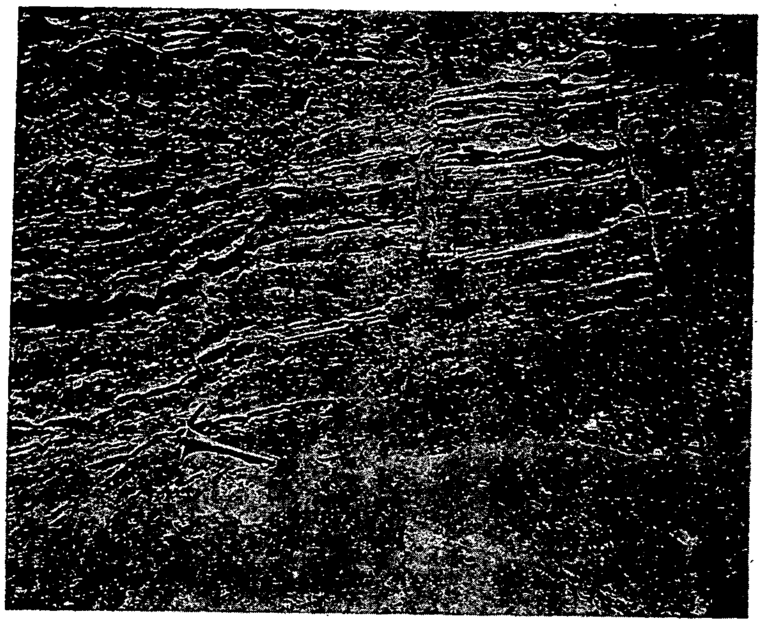

PLATE VIII

Figure 2. Faulting in lake sediments. Along U. S. Route 5 one mile southeast of Bellows Falls.

of this report is not qualified to evaluate these data, nor has any attempt been made during the present survey to correlate the varves. It does not seem, however, that his findings are incompatible with stagnation during ice melting from south to north.

To the writer's knowledge, Antevs was not particularly concerned with the manner in which the ice was removed from the valley or that he even considered the possibility of more than one mode of retreat. There is so much new data and so much recent modification of the glacial concepts that no doubt Antevs would consider a different approach if he were working with the varves at the present time. This report does not contend that the ice remaining in the valley was wholly inactive or that it melted down all over the valley at any one time. No doubt temperatures did moderate from south to north and even the stagnant ice should have been removed in Hartford, Connecticut much earlier than it was

in St. Johnsbury, Vermont, and quite possibly 4100 years earlier. It does not seem that the mechanics of the ice removal would have affected the varve count in any way.

One of the geologists that has refused to accept any interpretation of the varve count other than that submitted by Antevs is Lougee (1940, p. 189–200), who emphatically states that any kind of a stagnation is impossible. Lougee attempts to support his arguments by still insisting that deposits formerly called moraines, but now are believed to be otherwise (Goldthwait, 1938, p. 546–60), are features deposited by actively retreating ice. The writer has not had the opportunity to study the deposits of New Hampshire but he can state that there are no frontal moraines in Vermont, at least in the sections studied during the present survey, and no other structure that can be attributed to actively retreating ice.

The Glaciation of the Connecticut River Tributaries

The glacial advance and retreat in the tributary valleys of the Connecticut River were quite similar to that of the master valley. From the striation map, it is concluded that the ice movement in the area of the lateral streams was in a south-southeast to southeast direction, at least in the southern half of the state. The ice direction was more or less parallel to the course of the streams, and glacial erosion reshaped the valleys as it did in the Connecticut valley. The tributary valleys were, of course, smaller than the valley of the master stream and after glaciation they were still smaller, but the characteristic U-shape attests to glacial erosion.

The deposition in the tributary valleys was chiefly glacio-fluvial sands and gravels, again similar to the master valley. Kame terraces are as common, and in many sections more common, than along the trunk valley. From an economic point of view, the kame gravels in the lateral streams are coarser and more suitable for construction purposes than those commonly found along the Connecticut. It should be pointed out, however, that texture is not the only qualifying property of the gravel used for construction purposes. The rock composition is equally as important, particularly for road construction, and much of the gravel of Vermont contains a very high percentage of metamorphic, schistose rock. When gravel contains a high percentage of foliated rock, such as that containing mica, chlorite, talc, amphibole and/or pyroxene, they must be rejected because of their inability to wear without crumbling.

Valleys that were not eroded by ice moving through them were, generally speaking, filled with glacial deposits. In most cases the material filling the valley is till, but kame sands and gravels are sometimes found in the lower reaches. After the ice receded, streams developed on top of the deposit in the pre-glacial valleys. Also, because the sediments offered little resistance, and the stream was high above baselevel, downcutting by the stream was accelerated. At the same time mass wasting has not kept pace in so far as the shaping of the valley is concerned. As a result, these streams have cut valleys with high, steep, almost vertical walls instead of the expected V-shape. Such valleys one hundred feet deep are not uncommon in southeastern Vermont.

Glacial Lakes of the Connecticut Valley

The existence of glacial lakes in the Connecticut valley has been known to geologists for well over a century, and many studies have been made of the lacustrine deposits. According to the most quoted geologic literature, lakes were formed in the valley as the ice waned and eventually a body of water extended from Middletown, Connecticut northward to Lyme, New Hampshire, a distance of 157 miles, and later it expanded as far north as St. Johnsbury. There are several possible reasons why the lake formed, but the most accepted of these is that the river was dammed at Middletown either by ice or by a glacial moraine. The first and highest of these lakes has been named Lake Hitchcock. A second lake that supposedly developed when the higher water level was abruptly lowered 90 feet has been designated Lake Upham (Jacobs, 1941, p. 37; Goldthwait, Goldthwait and Goldthwait, 1951, p. 47; Lougee, 1935, p. 5–8).

That at least one lake occupied the Connecticut valley in the section that borders Vermont is unquestionable. Varved clays, laminated silts, and lacustrine sands are found all along the valley. Many of the exposures measure 75 to 100 feet in height. In the vicinity of Putney, the small tributary streams of the Connecticut have carved the deposits into steep sided forms, and varves of the "typical textbook variety" are exposed. Because of the high content of clay in the deposits, slumps and slides are common so that fresh exposures are not difficult to find. Such varved sediments are exposed more or less continuously all along the walls of the Connecticut valley from Brattleboro to St. Johnsbury. In some sections the deposits are more correctly described as laminated silts and clays inasmuch as the deposits in these localities are too silty to be classified as varves. In other areas the lacustrine sediment is

predominantly sand, and a few of these are thick, well sorted and fine textured, containing no apparent lamination or bedding.

The most conspicuous and unexpected aspect of the lacustrine materials is the complete absence of beach gravel and phenomena that are normally associated with lake shore erosion and deposition. A few former writers on this subject have stated that they found shore phenomena, and some have implied that the shore line was traceable up and down the valley. Such conclusions have been made by using the beforementioned "delta" deposits as shore indicators, but as has been stated, the present survey classifies these as kame terraces. The reclassification of the deposits, formerly designated as deltaic is more in accordance with that employed by Brown (1931) in his studies in Massachusetts. Said Brown, "the deposits show typical internal structure of kame terraces and differ from them only in areal extent." It seems most unlikely that a lake could have existed in this valley long enough for the accumulation of such a great thickness of lacustrine sediment without the deposition of shore gravel and the formation of some types of shore structures. It is admitted that the lake was quite narrow, averaging from one to two miles, but this would not explain the complete absence of shore phenomena. As will be discussed subsequently in this report, the writer is of the opinion that shore deposits were formed but that they have since been removed by the erosion of a glacial readvance.

After studying the deposits of the Connecticut valley for two field seasons, between the Massachusetts line and Pompanoosuc, a distance of approximately 90 miles, the writer would hesitate to make any definite statements concerning a shoreline of the lake that made the valley deposits. It can only be pointed out here that the top of the higher lake sediments are at elevations of approximately 325 to 350 feet in the Brattleboro-Putney area, 375 to 400 in the Westminster-Bellows Falls section, 450–500 feet in the vicinity of Weathersfield Bow, 500 to 600 feet between Ascutney and Windsor, 600–625 between Hartland and White River Junction and 650–695 in Norwich. This report does not suggest that these figures represent shoreline elevations or that they give a reliable estimate of the amount of uplift that has occurred in this section of the valley. The figures do, however, show progressively higher elevations from south to north. The writer is of the opinion that data is not yet available in Vermont to compute the rate of uplift, and would hesitate to predict that further study would supply the information needed to do so.

The present survey has noted certain deposits that have characteristics similar to beach gravel but the specific structures necessary to classify them definitely as such has been lacking. That is to say, the materials in the deposit are compatible with, but not proof of, shore line deposition. The most noteworthy of these gravel deposits are located in the Wilder-Norwich area and this is the only locality where indications were favorable enough to warrant special study. A tentative beach was marked at an elevation of 695 feet in this area. The gravel is located along the slope of the valley wall, trending approximately north-south, and occurs in scattered deposits for a distance of six miles. The most southerly of this series is located along the Hartford-Norwich road, three-fourths of a mile northeast of Hartford. From this locality the gravel is more or less continuous to Bragg Brook where it joins a small delta that was made into the lake three-fourths of a mile northwest of Norwich Village. Similar gravel is found at the same elevation on both sides of the valley that runs northwest from Norwich into Charles Brown, Bloody and New Boston brooks.

Three very small gravel deposits of similar characteristics are located along U.S. Route 5 south of White River Junction. These occur at the top of the lake terrace between the White and the Ottauquechee rivers. The most northern of these is located on the grounds of the Veterans Hospital, one and one-half miles southwest of White River Junction, and the other two are one and two miles south of the hospital. As has already been stated, the evidence here is not conclusive. It is hoped, however, that when the present survey moves back to the Connecticut River valley to continue the mapping north of Norwich that this probable shore can either be definitely proven or disproven. According to the literature (Lougee, 1935, p. 5–8) two lake levels are distinguishable in the Hanover, New Hampshire area, across the river from Norwich, at elevations of 658 and 570 feet. These are the supposed levels of Lake Hitchcock and Lake Upham respectively. The writer of this report believes that the higher lake level must be above 658 feet in this section.

Another very noteworthy sand deposit of possible importance is found in the vicinity of Windsor. Mill Brook, which follows the northern base of Ascutney Mountain to the vicinity of Mountain School, leaves its south-southeasterly course about one mile east of Mountain School and turns sharply north to flow through a deep, bedrock gorge (The Narrows) before entering Mill Pond at Windsor. Horseback Ridge, a sharp, bedrock remnant occupies the space between The Narrows and the Connecticut River. It is apparent that at some former time Mill Brook followed the base of Ascutney Mountain, more or less parallel to the present road between Ascutney and Brownsville. At any rate,

there is an abandoned valley south of Horseback Ridge that must have been the former course of Mill Brook. The abandoned valley is now filled with lacustrine sand to an elevation of approximately 600 feet, and sand can be traced continuously from here southward to Weathersfield Bow and northward to Hubbard Brook, one and one-half miles northwest of Windsor. The sand, particularly in the abandoned valley, is quite thick, reportedly over 200 feet, fine textured, and, insofar as it was possible to determine, almost structureless. The position of this deposit suggests deltaic deposition, but it is impossible to call such a structureless, fine textured deposit a delta. Just south of the village of Ascutney the sand deposit swings up a second Mill Brook for about a mile. A small portion of the detritus in this valley is gravelly, similar to beach gravel, and also suggests a lake level near the 600 foot contour. This report does not mean to imply that this is the mark of a definite shore level, it is possible that it is, but this has not been proven. It is apparent, however, that the lake level was at least as high as 600 feet in the Windsor area.

The sand deposit at the mouth of the West River at Brattleboro is very similar to the deposit at Windsor except that it is not as thick or as widespread. The higher terraces in Brattleboro are definitely kamic as is evidenced by ice-contact structures and an undrained kettle just north of the high school. The lake sand that laps upon the kame terrace in the southern part of Brattleboro is at an elevation of 350 feet.

All of the deposits mentioned above are believed to have been deposited in the high level lake of the Connecticut valley commonly noted in the literature as Lake Hitchcock. As noted above, manifestations of these waters are seen everywhere along the valley. The existence of a second lake, Lake Upham, that according to former studies, maintained a level ninety feet lower than Lake Hitchcock, is much more difficult to prove. According to Jacobs (1942, p. 37), Lake Upham was in existence for 600 years at Hanover and 1600 years near Haverhill and Woodsville. If this be the case, it is possible that more definite evidence of such a lake may be found between Norwich and St. Johnsbury when the present survey has progressed that far. From the investigations made south of Norwich, however, it is concluded that Lake Upham may not have existed at all, or if it did exist, it was a series of small, shallow lakes occupying disconnected basins in the valley.

Lake Sediment in the Tributary Valleys

One of the most interesting factors concerning the lake sediment of the Connecticut River valley is that fact that no lacustrine sediment can be found in the tributary valleys south of the Ottauquechee River. The present survey has mapped a large portion of the West, Saxons, Williams and Black rivers, and no lacustrine sediment has been noted in these valleys. Lacustrine sediments have formerly been reported in the Black River valley at Cavendish (Jacobs, 1927), but this was a small independent water body, between Plymouth and Cavendish, and was not an extension of the Connecticut valley lake waters. The only explanation that can be given for the absence of lake sediment in these valleys is that ice must have remained in the tributaries much longer than in the master valley. This argument may be valid as far as it goes, but the reason for the ice in these valleys is more difficult to explain, considering that Lake Hitchcock was supposedly in existence for approximately 4000 years. These rivers are short drainage ways from the Green Mountains which are much closer to the Connecticut River in the southern part of the state. It may have been possible that the closeness of the mountains in this region could have been an influence in allowing the ice to remain for a longer time. There is also the possibility that the valleys may have been filled with sediment but no evidence suggesting so much sediment has been found.

One deposit in the Black River valley that may or may not be related to the lake episode in the Connecticut valley is noted here. This is one of the largest and most interesting gravel terraces of southeastern Vermont, and is located at North Springfield. The material composing this terrace varies greatly in texture, and it is believed that it was deposited in contact with ice in the valley. The terrace extends up the Black River from Springfield to Amsden, a distance of nine miles. At North Springfield the valley widens to the west as though a pre-glacial meander had existed here. In the wider section, at North Springfield, the terrace occupies an area two and one-half miles long (east-west) and two miles wide (north-south), and the most interesting aspect is its almost level surface at an elevation of approximately 580 feet. The Springfield airport is located on the north side of the flat. The Black River and Great Brook, that flows onto the deposit from the west, have entrenched into the western and central portions, but there are still large areas that are quite level (Plate IX, Figure 2). It seems possible that the flat top may be a stagnant water level although it has been impossible to identify the deposit as a delta.

From investigations made during the present survey, this report concludes that lake waters extended up the Ottauquechee River only as far as Quechee Gorge. Since 1908, when C. H. Hitchcock wrote a report

PLATE IX

Figure 1. Varved clay containing rounded cobbles. South side of White River one-half mile west of Riverside (Stockbridge Township).

on the Hanover, New Hampshire region, the sand and gravel deposits at Deweys Mills and Quechee have been called a delta (C. H. Hitchcock, 1908a) that was built into an extension of Lake Hitchcock. The evidence to prove that this deposit is a delta has not been found by the present survey, and this deposit was mapped as a kame terrace. The deposit is similar to terraces at North Springfield, and there may be some relationship to the top of this deposit and the lake level, but the gravels are believed to have been made in contact with ice that prevented the lake from actually expanding beyond the gorge.

The waters of Lake Hitchcock, did indeed, occupy the White River valley from White River Junction all the way to the headwaters of the present stream. The survey now in progress followed the White River valley to the western boundary of the Randolph Quadrangle, and lacustrine silts and clays were mapped as far as the survey progress in that direction. Lake silts and clays were also noted in the valley of the First Branch of the White River to Tunbridge, on the boundary of the

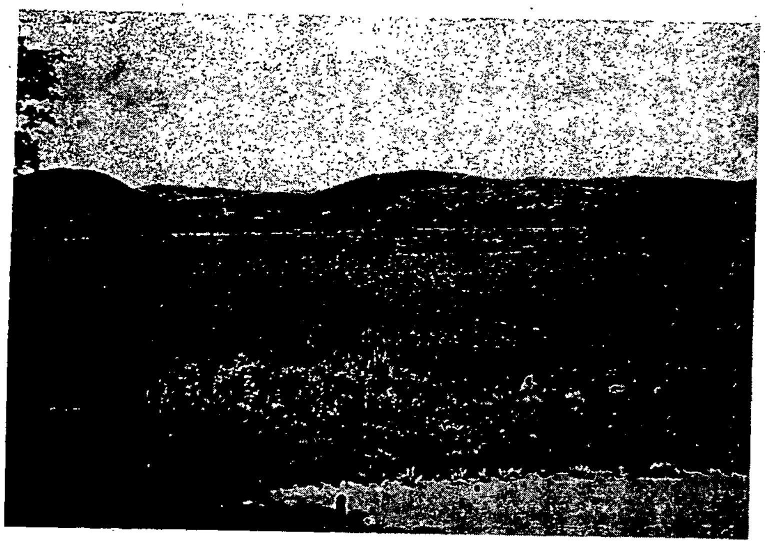

PLATE IX

Figure 2. Level surface of the kame terrace at North Springfield. Picture taken looking south one mile due west of North Springfield.

area mapped. The Second and Third branches were followed to their heads, and the lacustrine materials were found to extend up the Second Branch just north of East Randolph, and the Third Branch to Randolph. The silty clay was found north of Randolph along Ayers Brook, a tributary to Third Branch in the vicinity of East Braintree. The top of the lake sediment in these valleys is at elevations of 610 feet on the west side of the White River just west of West Hartland, 700 feet in the vicinity of East Randolph (Second Branch) and 700 feet along Ayers Brook, north of Randolph.

There is very little information concerning the lacustrine sediments of the tributary streams of the Connecticut River north of White River Junction. It is known that the lake covered the Passumpsic valley as far north as the vicinity of West Burke.

According to Jacobs (1942, p. 37) the lake deposits of the tributary valleys, including the White River, were deposited in water of the Upham Lake episode. As has already been stated, this report does not hold that this lake stage, if it existed, was of such height and magnitude.

The elevations of the lake sediments exposed along the White River and its tributaries are even higher than the highest manifestations in the Norwich area. These are, therefore, at least tentatively, correlated with the Lake Hitchcock deposits.

Evidence of Ice Readvance in the Connecticut Valley

There has been much discussion in other reports of the past quarter century concerning the possibility of an ice readvance into the Connecticut River valley. The evidence of a readvance at Fifteen Mile Falls (8 miles south of St. Johnsbury) is on record (Crosby, 1943). Much has been written about this particular locality where two tills were observed in dam excavations along the Connecticut River. Crosby also noted till overlying lake sediments on the Peabody River, 25 miles east of Fifteen Mile Falls. These, and other manifestations, have lead to the conclusion that an ice readvance occurred in this region. It has been concluded, however, that the advance did not extend to the south, inasmuch as no evidence had been found to support such a conclusion (Flint, 1953). The present survey, however, has collected data which the writer believes supports a conclusion that the readvance of the ice moved down the Connecticut valley at least as far south as the White River and possibly to Brattleboro or beyond.

The most convincing evidence of ice readvance in the Connecticut valley area is found along the White River in two different localities where varved lacustrine sediment is covered by 25 to 50 feet of till. The most eastern of these exposures is located in Hartford Township (Hanover Quadrangle) two and one-half miles west of the Connecticut. The till and varved clay have been exposed by the downcutting of a small, unnamed, tributary stream of the White River. Jericho Road enters this valley from State Route 14 three-quarters of a mile west of the village of Hartford. As one follows Jericho Road upstream from State Route 14, the till and varved clay are easily visible on the right side of the road, one mile north of the White River. At this exposure, 50 feet of clayey, bouldery till is exposed above varved clay. It is apparent that the lacustrine sediment extends to the bottom of the stream valley, but it is covered below the road. At least 25 feet of the clay is visible, however, between the road and the contact of the till. This outcrop had been noted prior to the present survey in the Guidebook of the 46th annual meeting of the New England Intercollegiate Geological Conference (1954), prepared by the Department of Geology of Dartmouth College.

A second locality showing till over varved clay is located in the valley of Lillieville Brook in Stockbridge Township three and one-quarter miles southwest of Bethel (Randolph Quadrangle). Three additional exposures were found on the steep, western slope of Lilliville Brook valley 1.1, 1.4, and 1.9 miles north of the White River. In this section roughly 60 feet of varved and laminated silts and clays are exposed below 25 to 40 feet of clayey bouldery till.

Even before the discovery of the till over lake sediment, late in the second season of the present survey, certain manifestations which seemed to indicate ice readvance had been noted in the Connecticut valley. In the vicinity of Putney, and particularly along Canoe Brook, the relative position of the lacustrine sand, varved clay, and the overlying gravel seems to indicate ice erosion between two episodes of lake deposition (Figure 3). The varved clay here averages 60 to 80 feet in thickness, and is exposed in numerous cuts and valleys. The in-valley side of the clay (toward the Connecticut River), however, seems to have been cut off, and a thickness of lacustrine sand was deposited against the end of the clay. The contact between the sand and the clay is quite sharp in the Canoe Brook valley. The clay and the sand are partially covered by a thin layer of gravel which resembles kame terrace, but the ice contact structures are rare.

It seems logical to assume that the varved sediment was deposited in considerable thickness across the valley inasmuch as good, high exposures are found on both sides of the river. The truncation of the in-valley side of the clay could have been caused by a readvance of a small ice lobe down the valley. The gravel, if this interpretation is true, was deposited along the margins of the ice before it retreated from the valley. The sand, according to this reasoning, would have been deposited in a lower lake that followed the retreat of the later ice. The possibility that both the gravel and the sand resulted from a shallowing phase of a single lake has been considered. The evidence does not, however, seem to be compatible with this process. In the first place, the sand of a lowering lake level should overlay the varved sediment and grade into the gravel toward shore. The gravel should also show wave formed structures. These characteristics are not found in this section.

In other sections of the river, especially in the Putney-Bellows Falls area, the varved clay is covered by gravel which in turn is mantled by sand. The gravel in these deposits more closely resembles kame gravel than any other type and the surface sand is believed to be lacustrine. Occurrences of deposits such as these are rather common, and they are

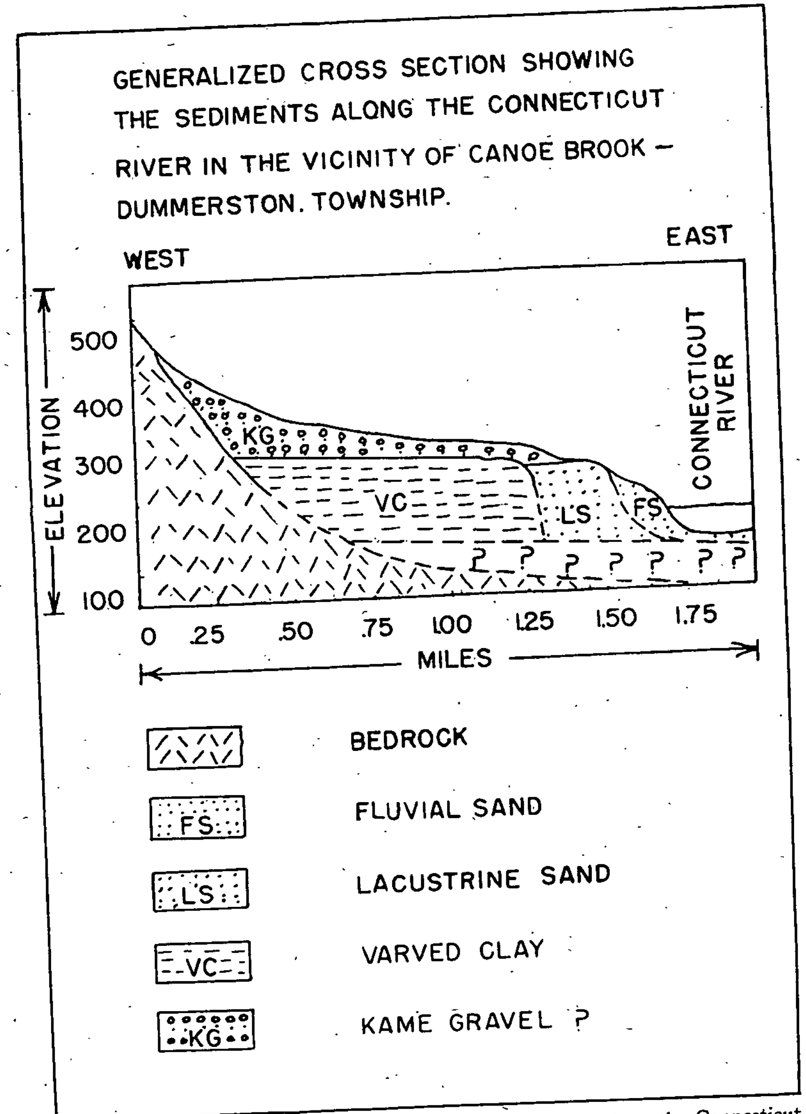

Figure 3. Generalized cross section showing the sediments along the Connecticut River in the vicinity of Canoe Brook—Dummerston Township.

exceptionally well exemplified along the Connecticut River, a mile east of Putney and along the small streams in the vicinity of Westminster.

It has been stated earlier that the manifestations of a second lake stage in the Connecticut valley (Lake Upham) are not too well defined. The sequences of sediments, as described in the preceding paragraphs, are examples of the kind of manifestations that have been mapped during the present survey. If these interpretations are correct, and it is established that a glacier did readvance down the valley, then it follows that Lake Upham was instead a small, disconnected series of lakes that followed Lake Hitchcock, and that the two lakes were separated in time by an ice advance into the valley.

There are also deposits in the Randolph Quadrangle that seem to indicate an ice readvance following the deposition of the varved clay in the valleys of the First, Second and Third branches of the White River. In these areas, the varved and laminated lacustrine clays and silts are overlain by sands and gravels believed to be kame terraces. The definite classification of the overlying deposits is quite difficult because of the great amount of slumping and the absence of openings that would show the structure. The writer believes, however, that these are kame gravels inasmuch as ice-contact structures have been noted in some areas. The greatest thickness of the overlying gravel was mapped on the east side of the Third Branch immediately south of the village of Randolph (Figure 4).

It is admitted that any one of the above deposits used to support the argument of an ice readvance could have been caused by a minor fluctuation of the ice. It is believed, however, that all of them together add up to enough evidence that can no longer be ignored. Support is also found in the literature that notes the occurrence of till over lacustrine and marine sediment in many regions of New England, as is recorded in the writings of White (1947), Currier (1941), L. Goldthwait (1941a), Taylor (1931) and the early reports of C. H. Hitchcock (1879, 1893).

The most timely question relative to the above mentioned ice readvance concerns the time of deposition of the till and kame gravel covering the lacustrine sediment. This question cannot, as yet, be answered. Two hypotheses concerning the age relationship of the upper deposits are now being given major consideration. In the first place, it is possible that the varved sediments were covered by a minor readvance (fluctuation) of the retreating active ice contemporaneously with the

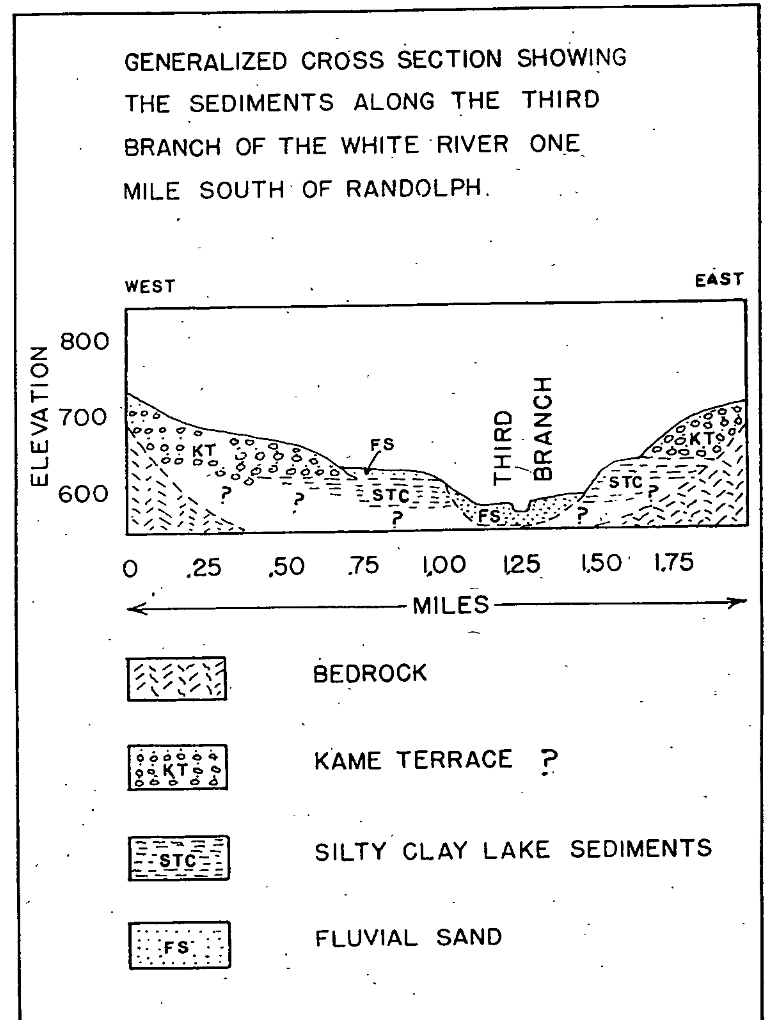

Figure 4. Generalized cross section showing the sediments along the Third Branch of the White River one mile south of Randolph.

lake episode. This possibility, however, is not compatible with the stagnant ice hypothesis. Secondly, it is possible that the overlying till and gravel were deposited by an ice sheet that invaded New England in the substage that followed the deposition of the lake sediments. Although most geologists conclude that a later advance did occur, its southern maximum has been estimated to be in the approximate latitude of St. Johnsbury.

It is the opinion of the writer that the latter of the above mentioned hypotheses may be the more valid. It seems that as this survey continues, the evidence of the later ice advance becomes more and more widespread and that the extent will prove to be too great to be explained by a minor fluctuation of a retreating glacier. These facts, however, do not prove the age of the later deposits, and this emphasizes the importance of future research on the age problem.

It seems appropriate to add a short discussion of the absence of shore phenomena associated with the lacustrine sediment in the Connecticut. That is to say, it might be well to look back at this problem in view of the preceding discussions. As already stated, it does not seem possible that a lake (or lakes) could have existed for so long without the development of shore terraces and the deposition of beach gravel. It is believed, however, that at least a partial explanation can be proposed, and that the absence of shore phenomena bears largely on the following factors:

First of all, the fact that rather definite evidence has been found showing that an ice readvance occurred over this region between the two lake episodes probably best explains the absence of a high (Hitchcock) shore. The ice activity was great enough to obliterate parts of the shore, and kame sands and gravels were deposited over the varved sediments along the sides of the valley in many sections. How much lake sediment was actually covered with till is unknown inasmuch as the only locality where till has been found over the lake sediment is along the White River.

Secondly, as stated earlier in this report, it seems that mass wasting in the form of solifluction might have occurred in some of the areas that have been mapped. Undoubtedly a part of the shoreline in these sections was covered by the flows. The fineness of the material in some of the solifluction mantle, particularly at lower levels, would seem to indicate that they were fed by an abundance of silt and sand that might well have been of lacustrine origin.

A third factor that must be considered is the size of Lake Upham and

the length of time it existed. The sediment that can definitely be attributed to this latter lake is composed of sand and is of shallow water origin. It seems that this stage was of short duration, and it is quite possible that it was a chain of several lakes at different levels rather than a single lake. The hypothesis that the latter lake sediment was deposited in slack waters along the margins of the ice that occupied the valley cannot be discounted.

THE GLACIATION OF THE VERMONT PIEDMONT

The Vermont Piedmont (Jacobs, 1950, p. 79) is actually that part of the New England Upland (Fenneman, 1938, p. 358–68) which lies within the boundaries of this state. The upland is the largest geomorphic subdivision of the New England crystalline belt, and only a small portion of it lies within Vermont. The whole section is best described as an uplifted surface that has been dissected and glaciated. The present topography has been carved from the complexly folded and faulted metasediments of this region. The surface expression is similar to that of an uplifted and dissected peneplain, but the peneplains of New England and elsewhere, have been, in recent years, the subject of much debate. It is not within the scope of this report, however, to discuss the pros and cons of the peneplain problem. The metasediments of the New England Upland have been intruded by both acidic (silicic) and basic (mafic) plutons of various kinds and sizes that are scattered throughout the region.

In Vermont, the western boundary of the piedmont follows the eastern foothills of the Green Mountains from Brattleboro northward to the Canadian border west of Lake Memphremagog, in the vicinity of North Troy. The eastern boundary in Vermont is the Connecticut River from Brattleboro to within a few miles of the village of East Waterford. North of East Waterford, the piedmont is bounded on the east by the Northeastern Highlands, and this contact runs generally north from Concord. The Vermont piedmont is thus a long narrow strip running north and south for a distance of 150 miles, with an average width of from 20 to 40 miles. The rocks composing this section are chiefly gneisses and schists, but the degree of metamorphism and the composition varies greatly from place to place. The plutonic intrusions are scattered over Vermont and are exemplified by the granites at Dummerston, Plymouth, Bethel, Barre, and Craftsbury; the syenites of Ascutney Mountain and Cuttingsville; and the serpentines, formed by the alteration of basic rock, at Roxbury, South Duxbury and Lowell.

There is nothing spectacular about the glacial deposits of the piedmont upland. The surface is veneered by a thin mantle of loosely consolidated till that, as has already been stated, has a structure and surface expression that is essentially nondescript. The composition of the drift varies greatly from place to place inasmuch as the bedrock is of different kinds and of variable resistance. There is a high percentage of material in the till that is derived from the local bedrock, and the larger cobbles and boulders are commonly quite angular since the distance of transport was relatively short. The thickness of the till varies on the summits of the piedmont, but it is usually zero to fifteen feet with large areas of bedrock exposed.

Glacial erosion on the uplands also varied with the resistance of the rock. Except for the scouring of the valleys cut into the piedmont, erosion was chiefly the removal of the rock mantle and the scraping off of the surface of the rock. The glaciers carved channels and grooves in the bedrock over which they moved, but these are of local extent and of little significance. Post-glacial, subaërial erosion has obliterated much of the evidence of glacial erosion, since the till is thin or missing, and the rock is easily weathered. This report concludes that little of the irregular, erosional topography of this section can be attributed directly to glaciation. The rock mantle was removed, the bedrock was planed off, and the valleys were modified, but the over-all aspect of the topography, except for local details, was not greatly altered. It is probable that the pre-glacial topography was sharper, with the valleys more V-shaped and the ridges more irregular and steeper, but the slopes between the valleys and ridges were, no doubt, more gentle before the glaciers.

The uplands are the logical areas in which to study the direction of ice movement because the general direction was not so much influenced by topography on the higher parts of this region. It is believed that the striations found on the flatter summits are the most reliable for use in determining the true direction of glacial movement across the state. It is regrettable, therefore, that the bedrock is not more resistant to weathering and erosion, for there are large areas where no striations can be found because they have been removed since the recession of the last glaciers. The bedrock is here steeply inclined, and the surface generally truncates the gneisses and schists at an angle of 45 to 90 degrees to the foliations. The high content of platy minerals facilitates rapid weathering, and after a period of time striations cannot be distinguished from the characteristic linear expression of the fabrics of the weathered rock. Even the granites do not retain the striations inas-

PLATE X

Figure 1. Glacial striae and polish on crystalline rock of Vermont Piedmont. Near Lampson School two miles north of Brookfield.

much as sheeting is so common and so closely spaced at the surface that it is weathered, chiefly by frost action, into thin angular blocks (Plate X, Figure 2). For this reason reliable striations can not be found over large sections of this subdivision.

Another clue to the ice direction may be found in a few localities where boulder trains or indicator fans of distinguishable bedrock have been found. The most southerly of these is at Ascutney Mountain. At least a part of this mountain is composed of quartz-syenite, and since this is the only outcrop of this particular rock in the immediate area, the direction in which this rock was carried by the glacier is indicative of the direction that the ice moved. The indicator fan is here oriented south-southeast, indicating that the ice moved in that direction. This is one of the largest boulder trains in New England.

A much smaller, but very definite, indicator fan also occurs in Braintree Township on Braintree Hill (Randolph Quadrangle). A pinkish granitic rock, easily distinguishable because of its color, outcrops on the western and southern slopes of Mt. Nevins in the extreme northern part of Braintree Township (southwest corner of Barre Quadrangle).

PLATE X

Figure 2. Sheeting of granite intrusion. Note thinness at the top. Bethel granite two and one-half miles north-northeast of Bethel.

The indicator fan of this outcrop is a concentrated train of boulders trending south to south 10° east for a distance of three to four miles. The boulders are conspicuously concentrated just west of the old Braintree Hill Church, three and one-half miles north-northwest of Randolph village. Also of interest in this vicinity is the "Rolling Rock of Braintree," a huge boulder propped up on a few small cobbles, one mile northwest of Braintree Hill Church (Plate XI, Figure 1). It is reported that this large rock could formerly be rocked by one man but it is a bit more stable at the present time.

Another small, but significant train of boulders is located in Thetford Township. Approximately four-tenths miles due east of Thetford village is a small, now-concealed, olivine gabbro-porphyry intrusion that was eroded by glacial action. The fan of the gabbro was spread out in a southward direction for a distance of one and one-half miles (Hadley, 1950, p. 10). A fourth indicator fan on the Piedmont is formed of Craftsbury Granite (Flint, 1957, p. 126). The granite fan, from exposures in

the vicinity of Craftsbury (Hardwick Quadrangle), shows a definite southward trend to the ice movement. It is interesting to note that the boulder trains correlate well with the available striation data in the sections in which they are located (Plate II).

A rather unique characteristic of the surface of the Piedmont section is the occurrence of rounded pebbles and cobbles (and in some cases boulders) in the surface till. The rounded fragments are so common in the southern part of the region that the till surface has a gravelly appearance. In a few small, isolated spots this material was so concentrated that gravel pits were actually found in the till. It is, of course, true that these pits did not produce a high quality gravel but some of them contained enough sand and rounded material to be suitable for use on the town roads. The gravelly surface was most common in the southern part of the area that has been mapped, and was not so much noted north of the White River. It should also be stated that this surface is not restricted to the Piedmont since it was as common to the foothills of the Green Mountains of the Brattleboro and Saxtons River quadrangles.

This report contends that the sandy, gravelly surface so extensive in the Piedmont is another manifestation of basal ice stagnation below the summits of the Green Mountains. It cannot be questioned that the rounded cobbles and boulders have been transported by waters, but it seems that the water activity was not great enough to deposit outwash or a layer of sorted gravel. It also appears that the water activity was contemporaneous with the deposition of the till. Certain aspects of this material are indicative of a deposition similar to sheet wash that would necessitate the removal of ice from the bottom of the slope toward the top. This might have been possible inasmuch as the difference in elevation between the lower and higher reliefs could have caused a correspondingly large temperature differential with much colder temperatures in the higher mountains. Such a removal (from bottom to top with sheet wash) might explain the fineness of the kame deposits in the valleys since the coarser materials would be deposited on the higher slopes. There has been no detailed work done in the Green Mountains of this area to supply needed proof or disproof of this. The one questionable aspect of the sheet wash concept is whether or not the amount of water available would be sufficient to cause the rounding of the cobbles.

The cobbles and boulders of this surface are compatible, however, with the concept that ice melted down from the mountains from top to bottom. Inasmuch as the basal ice was stagnant and the slopes are steep

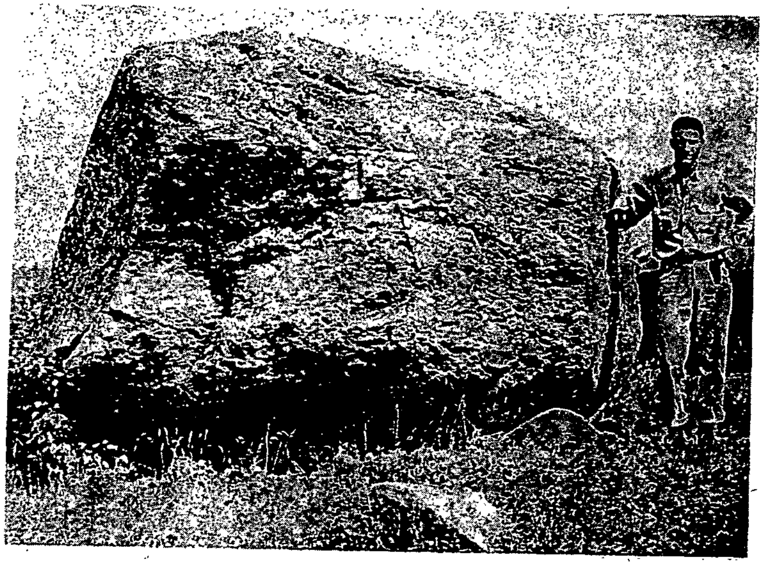

PLATE XI

Figure 1. The rolling rock of Braintree. Crystalline boulder propped up on a few small cobbles. Braintree Hill four miles north-northwest of Randolph.

the water would undoubtedly flow subglacially down the slopes. The fact that the gravelly material is strewn over the surface of large areas, however, would not be indicative of subglacial channels as might be expected. The material may have washed down from higher altitudes where it was included in the deposition of the final stages of the melting of the ice on the Piedmont.

Another mantle material, described earlier in this report as a possible solifluction mantle, is also well developed in the Piedmont upland. This surface was first encountered in the southwestern part of the Hanover Quadrangle, and it was found extensively in this and the Woodstock, Randolph and Barre quadrangles. As has been noted, there are several possible origins of this surface material. The writer is of the opinion that some type or types of mass wasting, probably solifluction and slumping, is (are) the most logical explanation from the present data. There seem to be a terrace development and a separation of coarse and fine textures

PLATE XI

Figure 2. Kame terrace gravel in Gunner's Brook, one mile north of Barre.

in so many of the deposits that solifluction in definitely indicated. That this material could be associated with a periglacial zone, and the possibility that it is a super-glacial till, cannot be ignored. These possibilities can be better explored after further studies have been made on the age of the drift. If, as has been suggested, the material proves to be a till that was deposited in association with a great deal of meltwater activity, then it may become another supporting evidence of stagnation.

The northern one-half of the Vermont Piedmont has not been surveyed during the investigations now in progress, and not enough information is available to make a detailed discussion of this region possible. It has not been actually determined that the above noted characteristics of the southern portion of the Piedmont are also applicable to the north. It is believed that the deposits to the north may reveal some very significant information relative to the glacial history when it is eventually studied. It is apparent, judging from observations made during a brief reconnaissance into that area, that there is much more glacial topography,

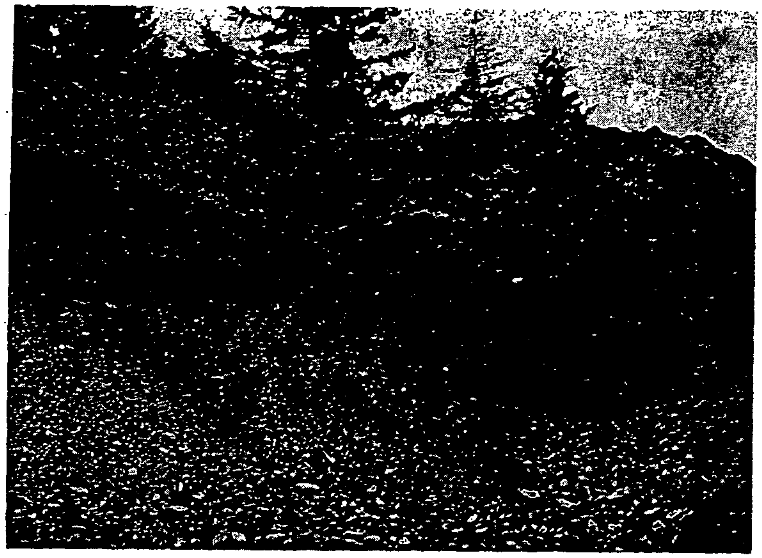

PLATE XII
Figure 1. Gravel in kame terrace along Gold Brook east of Stowe Hollow, one and three-quarters miles southeast of Stowe.

particularily depositional land forms, and much thicker glacial and glacio-fluvial detritus. The true perspective of these deposits, however, will have to await further study.

Piedmont Lakes of Glacial Origin

Probably the most conspicuous evidence of glaciation that can be readily noted in the northern half of the Piedmont is the number and size of the lakes in that region as compared to other parts of the state. One informative publication has recently been published (Mills, 1951) recording the results of a detailed study of fourteen such lakes in Essex and Orleans County. Five of the lakes described in the Mills report are in the Northeastern Highlands section and the other nine are in the extreme northeastern part of the Piedmont. The report, moreover, because of its detail, gives an excellent coverage of the types and origins of the lakes.

The lake area extends northward from Groton State Forest to the Canadian border, and the lakes, in general, become more numerous and

PLATE XII

Figure 2. Ice-contact structure of kame terrace in Gold Brook valley, one and three-quarters miles southeast of Stowe.

larger toward the north. The larger more important of these water bodies have been studied and described by Mills (1951), and it will suffice to give only a short resumé of the subject here. According to Mills (1951, p. 37) the origins of the lake-basins in this region resulted from two types of glacial activity. First of all, some basins were carved out of the bedrock by glacial erosion. Secondly, other lakes occupy kettles that were formed in the glacial drift by the melting of ice blocks that were buried in the sediment.

The important lakes that occupy glacially-gouged rock basins are Big Averill, Little Averill, Seymore, Crystal and Willoughby lakes and Holland Pond (Mills, 1951). The most spectacular of these is, of course, Willoughby Lake that is contained in a basin cut through a mountain of granite. The lake is over 300 feet deep, and the rock cliffs on either side rise 1500 feet above the lake. Lake Seymore is the largest body of water lying wholly within the state but it is not as scenic as Willoughby. It is also suggested that the basin is dammed at the north by glacial material and that the lake is deepest where ice had been confined where the granite crosses at the south end (MacClintock, personal communications).

The kettle lakes of the northern Piedmont include Island, Spectacle, Pensioner, Salem, Clyde and Norton ponds. These are not large, and they are much shallower than the bedrock basins, but their greater number almost makes up for their size. Silver Lake at Barnard and Elmore Lake, near Morrisville, are also kettle lakes, located south of the area here described. The above described lake-types are typical of a glaciated area, and the rugged terraine of Vermont is, no doubt, the only reason that more of them are not located in the state.

A consideration of the lakes of the northern region would not be complete without the mention of Lake Memphremagog which is located wholly within the New England Upland partly in Vermont and partly in Quebec. The lake, at an elevation of 686 feet, occupies a large bedrock trough that is 40 miles long and covers 48,000 acres, of which 9,600 are in Vermont (Richardson, 1908, p. 269). According to Richardson (1908, p. 269–70) the lake basin is cut in Trenton limestones and shales and granite. Surprisingly enough, there is very little more that can be learned from the literature about this lake, the largest in Vermont except Lake Champlain. It is assumed that the trough containing the lake owes its origin to glacial erosion although, to the writer's knowledge, no investigations have been made that prove its exact origin.

Post Glacial Lakes of the Piedmont

The post glacial lakes of the Piedmont are mostly contained in the tributary valleys of the Connecticut River on the east, and the valleys of streams flowing into the Champlain Basin on the west. The lakes of the Connecticut valley have been discussed in a preceding chapter of this report, and the lakes of the Winooski, Lamoille, and Missisquoi valleys will be dealt with in the chapter on the Champlain Basin. The one sequence of interglacial lakes that is not included in either of these was a series of lake episodes that were formed in the Memphremagog Basin during the retreat of the last ice sheet.

C. H. Hitchcock (1894) named this lake Glacial Lake Memphremagog. It is regrettable that this name has been used in the literature inasmuch as it bears the same name as a lake now existing in that area. To avoid confusion, however, no other name is proposed in this report. It is believed that a more appropriate designation may result from more detailed study, and when the extent of this glacial lake episode is determined.

According to C. H. Hitchcock (1906, p. 248) Glacial Lake Memphremagog had three different levels. In the vicinity of the present lake these levels are now found at elevations of 970; 1,060 and 1,270 feet. On the east side of Salem the elevations of the three strands are at 1,060; 1,071 and 1,274 and in the Barton River valley at 1,080; 1,180 and 1,284. Hitchcock proposed that Glacial Lake Memphremagog extended southward into the Lamoille and possibly the Stowe valleys. One outlet of the glacial lake was at Elligo Pond, according to Hitchcock, and the inference is that the two lower levels probably used this channel. The highest level seems to relate to a drainage at Runaway Pond or possibly the divide at Lake Willoughby.

The above facts prove the existence of a post glacial lake sequence in the Memphremagog Basin. The evidence, however, has been studied in so few localities that it would hardly make possible a discussion of any detail pertaining to the extent and drainage of the lake. It is true that a lake shore has been mapped in the Stowe area at about 1200 feet during the survey now in progress, and this is, no doubt, the evidence that prompted Hitchcock to propose that the lake expanded into the Lamoille and Stowe valleys. These strands have not been traced; and the shore manifestations that have been studied are so scattered that the writer does not believe that any definite conclusions can be made at this time.

THE GLACIATION OF THE NORTHEASTERN HIGHLANDS

The Northeastern Highlands (Figure 2) are a western extension into Vermont of the White Mountain uplift (Jacobs, 1950, p. 83). Essex County is covered by this subdivision, and interestingly enough, the mountains are almost wholly restricted to this county. According to Jacobs, (1950, p. 83) there are 130 mountains and high hills in this highland section, and the higher summits range from 2700 to 3400 feet in elevation. The data available on the Pleistocene geology of this region are hardly enough to justify a separate chapter but, because of the regional treatment given the state in this report, a brief consideration is deemed necessary.

The northern lake region of the Piedmont Upland extends eastward into the highlands, The lakes are contained in both erosional rock basins and kettle depressions formed during deposition of the drift. Mills (1951) has studied and described in detail the five larger lakes. These are Big Averill, Little Averill, Maidstone lakes and Island and Spectacle ponds. According to Mills, the first three are located in glacially gouged rock basins, and the latter two in morainic deposits. Upland swamps of the types commonly associated with mountain glaciation seem to be quite widespread on the broader summits and on terraces or benches that flank the mountains.

A brief reconnaissance of the highlands leads the writer to conclude that the region contains more widespread and thicker glacial deposits than most of the mountainous areas of Vermont. The validity of this conclusion, however, and the geologic importance of the deposits is yet to be proven by detailed field investigations. Jacobs (1950, p. 83) reports that the highlands contain "several glacial troughs" and that "huge erratics and glacial till cover large areas of the lower lands and overtop the mountains."

The Northeastern Highlands lie immediately to the east of the Piedmont region that is believed to have been covered by the southward moving glacial ice (Plate II). It is assumed that the Connecticut glacier overlapped the western flank of the mountains and probably covered the whole subdivision south of Maidstone Lake. Southeastern trending striations have been reported in the northeastern part of the section but the location of these is not known. No doubt a detailed study of the manifestations of ice movement would be most helpful in ascertaining the time and the significance of the Connecticut glacier.

THE GLACIATION OF THE GREEN MOUNTAINS

The Green Mountains, since they are the dominant physical feature, form the backbone of the topography of Vermont (Jacobs, 1950, p. 73). The mountain chain crosses the state from north to south, a distance of about 160 miles, and it is 21 miles wide at the northern border and 36 miles wide at the Massachusetts boundary (Figure 2). The elevation of the summits average approximately 2000 feet, but five peaks rise above the 4000 foot contour. The mountains are rugged, the crests are sharp and the slopes, in general, are quite steep.

In spite of the great interest in the Green Mountains over the past century, geologically and otherwise, the details available concerning the glacial geology are very limited. Even the general aspects of this facet of the geology of the range are in many sections incomplete. Many investigations of the surface deposits were made during the half century following the publication of the Hitchcock report (1861). The chief concern of the geologists during that fifty years was whether or not the mountains had been glaciated, and if they had been, whether or not local glaciers had formed on the summits during the ice age. The occurrence of glacial drift and erratic boulders was noted on the crests of the mountains as early as 1846 by C. B. Adams. The knowledge of these deposits, and similar detritus in the White Mountains, prompted Hitchcock (1861) to propose a submergence of 5000 feet for all of New England. Even the striations and grooves, including their correct directions, were recorded in the Second Annual Report of the State Geologist (Adams, 1846) but these, of course, were designated "drift directions" in the *Geology of Vermont*.

Hungerford (1868) was the first to correctly interpret the glacial drift, erratic boulders, striations and grooves. Since that time there has been no confusion concerning the question as to whether or not the mountains had been glaciated. There followed, however, much debate relative to the type of glaciation that had occurred. C. H. Hitchcock (1904), who helped develop the submergence hypothesis, made several studies along the summits of the range and, although he agreed that glaciers had deposited the erratics and the drift, he concluded that local mountain glaciation accounted for the manifestations of ice erosion and deposition. It was not until 1916 that it was definitely established that continental glaciers had crossed the Green Mountains when J. W. Goldthwait emphasized the complete absence of evidence of local glaciers.

It seems that geologists lost interest in the glacial geology of the mountains with the answering of the two questions discussed above. Also, no Pleistocene studies were made in that section for over forty years. Christman (1959 and in press) has recently made bedrock studies in the Mount Mansfield and Camels Hump quadrangles. During the course of his studies he has recorded the striations and the significant glacial deposits in those areas. The surface maps made for publication in his reports have been most useful in the preparation of this report.

It seems definitely established that the last glaciers to move over the northern part of the Green Mountains crossed the range in a southeast direction (Plate II). It has been suggested earlier in this report that this invasion was probably the Fort Covington that MacClintock (1958) has delineated in the St. Lawrence valley. This is undoubtedly true in the northern one-third of the mountains, but the southern limit of this ice has yet to be determined.

Probably one of the most significant influences of the Green Mountain Range was the fact that this north-south barrier caused the ice immediately to the west of it to be diverted to the southward. It was necessary for the ice to pile up on the western flank in order to pass over the summits. As a result, the movement in the Champlain basin and the Vermont valley was changed to a direction paralleling the range. No doubt this fact definitely, but indirectly, accounted for the great amount of erosion in the valleys.

Glacial Drift in the Green Mountains

The summits of the Green Mountains contain only scattered deposits of till and the higher peaks are generally free of any kind of detritus. Even on the barren crests, however, residual boulders of both local and foreign origin attest to the fact that glacial deposition did take place. Many of the large erratics have compositions that prove they were transported great distances before being deposited on the mountains. It was the presence of the erratics, plus erosional features transverse to the trend of the range, that finally convinced geologists that continental glaciers had crossed the summits.

On the slopes of the mountains, particularly below the 3100 foot contour, the drift is much more continuous although it is generally quite thin. The thickness of the drift varies greatly from one place to another depending upon the amount of slope and the amount of removal during deposition. It has been noted earlier in this report that the till on the mountain slopes in the southern part of the state has a gravelly surface resembling outwash. The gravel surface is most conspicuous in the Brattleboro and Ludlow quadrangles and least noticeable or absent in the Camels Hump area.

The most significant deposits on the slopes of the mountains may be the kame terrace gravels that occur in various areas. The extents of the terraces are still unknown but the present survey has noted kame gravels in the Camels Hump and Mt. Mansfield quadrangles. Such terraces are still to be mapped and studied elsewhere.

Glacial Erosion in the Green Mountains

It is probable that the most widespread evidence of glaciation in the Green Mountains is the occurrence of numerous features formed by glacial erosion. These include the notches (gaps) through the ranges, summit lakes, striations, grooves and related features.

The most noted gap in the Green Mountains is Smugglers Notch between Mt. Mansfield and Sterling Mountains. The notch has been studied by Jacobs (1938, p. 41) and Christman (1959, p. 70), and both agree that this is a col between the watersheds of the Brewster and West Branch rivers. Undoubtedly the stream erosion occurred prior to Pleistocene glaciation but, as Jacobs suggests, glaciers must have moved through it and reshaped it to some extent. The writer agrees with these authorities in that the notch does not exhibit "typical" glacial charac-

teristics. Probably the greatest effect of ice erosion was the deepening of the notch and the steepening of the sides. The writer was most impressed by the huge blocks of rock noted along the road into the notch from the east. The blocks are undoubtedly frost-riven and it follows that the steep cliffs on either side of the gap owe their steepness, at least in part, to this process. It is possible that the intense frost action may have been associated with glacial or periglacial climates, but this is not necessarily true since low temperatures are quite common even at the present time. This report concludes, however, that much of the frost-riving took place during or shortly after the last ice episode.

The lake basins occupied by the Lake of the Clouds and Bear Pond on Mt. Mansfield and Sterling Pond on Spruce Peak are, according to Christman (1959, p. 69), the result of glacial erosion. The locations of these on the mountain summits, says Christman, preclude their possible formation by any other erosive agent.

Two valleys located in the Camels Hump Quadrangle have characteristics that are more typical of those formed by ice erosion. The most northerly of these is Bolton Notch one mile south of West Bolton. The valley, two and one-half miles long, has a relatively flat bottom and almost vertical sides that rise over 900 feet on the east side and over 400 feet on the west. It is probable that a col between two streams, flowing in opposite directions, existed here before the Pleistocene. But, unlike Smugglers Notch, the Bolton Notch was definitely shaped by the ice that moved through it. Two miles south-southwest of Jonesville is another glacial valley, that to the writer's knowledge, has never been named or mentioned in the literature. This gap is designated Gillett valley in this report inasmuch as the pond located in it is known by this name. The valley is very narrow at the bottom, only approximately 500 feet wide, and is more V-shaped than Bolton Notch. The valley walls, however, slope approximately sixty degrees and rise over 800 feet above the valley floor. In spite of the V-shape, the writer believes this to be a glacial trough and that the basin containing the pond in the valley floor was formed as an ice block depression inasmuch as it is dammed at each end by glacial drift.

There are numerous other gaps through the Green Mountains including Hazens Notch, Lincoln, Middlebury, Mt. Holly and Brandon gaps and Sherbourne Pass. It is assumed that glacial erosion had some part in the erosion of some or all of these but they have not been investigated by the present survey and, so far as the writer has been able to ascertain, their origin has not been noted in the literature.

The Lamoille and Winooski rivers flow through the Green Mountains from east to west. The gorges of these streams cut the mountain near the highest peak. The Missisquoi River rises in the mountains of Vermont and flows into southern Quebec, through the mountains and thence southeastward across the Champlain lowland of Vermont. Its course through the Green Mountains is therefore not in Vermont. Jacobs (1938, p. 7) proposed that these streams are superposed, or as stated by Christman (1959, p. 11), they were cut down from an earlier erosion surface. The present survey has not noted evidence that would indicate that the streams have been superposed. Inasmuch as the range is narrow in this section, it seems more logical to assume that the headward erosion of streams on either side of the mountain developed a gap that later was cut down to stream level and resulted in stream capture. This is the proposed origin for Smugglers Notch. Stream gorges result from the same activity except that more intense erosion was necessary. There is little evidence that glacial erosion greatly affected the stream valleys. It is possible that these were lowered during the Pleistocene, but the well developed stream pattern east of the mountains would indicate that the drainage was established prior to glaciation. Glaciers did block the valleys and caused the development of glacial lakes, but these will be discussed later in conjunction with the lake episodes of the Champlain Lowland.

Ice Waning in the Green Mountains

Evidence from various sections of Vermont has been cited in this report that indicates the ice stagnated below the higher elements of the topography after the upper ice had waned. That is, the upper part of the glacier still possessed glacial motion, but it melted at its margins faster than it was moving forward. As a result the upper, "active" part of the glacier melted northward leaving in its wake masses of the lower part of the ice sheet as stagnant blocks in the valleys and lowlands. The Green Mountains, because of their great relief should, and indeed do, supply much evidence to support the stagnation hypothesis. The present survey has covered such a small area of the range that it is impossible to record evidence from many areas but certain data are available which should be noted.

If the glaciers actually did stagnate over Vermont, the critical elevations would be those of the Green Mountain summits since these are the highest in the state. That is to say, the ice above the crests of the range waned by ablation thinning but the ice below could only remain as isolated, quiescent masses. Accordingly, the crests of the higher sections

of the mountains should have been the first to have been uncovered, and as a result, ice marginal lakes and glacio-fluvial drainage would result. Hence, high level lake and glacio-fluvial (mostly kame terrace) deposits would now be present to attest to such conditions.

Lake deposits have been noted along the slopes of the Green Mountains. Christman (1959, p. 71) described a lake sequence at an elevation of 1900 feet on the west slope of Mt. Mansfield. On the east slope of Mt. Mansfield lake sediments were noted in the Montpelier Quadrangle along the mountain road to Smugglers Notch. These are at elevations of approximately 900 and 1200 feet. Kame terrace gravel, above the lake sediments, is found on the east slope along the Mountain Road from the entrance of the toll road to the base of Spruce Peak. The writer believes that these deposits were formed along the ice margins as the ice melted down (top to bottom) from the top of the mountain.

Similar isolated patches of lake sediments have been noted during the present survey in the Brattleboro and Woodstock quadrangles. Two miles west of Barnard, in the valley of Locust Brook, an exposure of six to ten feet of brown silty, lacustrine clay is exposed on the west slope of the valley. The elevation of the deposit is approximately 1150 feet. Another high-level, isolated deposit of lacustrine clay is found on the headwaters of Adams Brook (Marlboro Township) three and one-half miles south-southwest of South Newfane. The deposits are so far removed from any interglacial or post-glacial lakes that the only explanation is that they were deposited in small ice marginal lakes.

Jacobs (1938, p. 37) implies that Mt. Mansfield was once covered by glacial till but that it has since been "washed down" from the upper levels. The writer of this report, however, maintains that the almost complete absence of till above the 3100 foot contour is not because of removal after deposition. It seems that drift was never deposited on the higher ridges. The great amount of water associated with the melting of the ice transported the finer materials to lower levels leaving only the boulders that are strewn over the surface of the summits.

THE GLACIATION OF THE CHAMPLAIN LOWLAND

The Champlain Lowland is a structural trough located between the Green Mountain and Adirondack uplifts. In Vermont, this small area contains the lowest elevations as well as the most nearly horizontal and least metamorphosed bedrock. The surface is not flat, however, since hills are scattered all over the lowland. The hills, according to Jacobs

(1950, p. 59), are homoclinal structures some of which are klippes separated from the Green Mountain overthrust, and others that were formed from thrust blocks of the Champlain fault. A large portion of the surface of the lowland, between the hills, is a relatively flat plain composed of lacustrine sediment deposited in the glacial lake that covered it during the waning stages of the last glacial episode.

The lowland includes Lake Champlain that extends 100 miles north and south along the New York-Vermont border. The lake is generally less than one mile wide south of Chimney Point but it widens to ten miles at Burlington and is twelve miles at the Canadian boundary. The lowland in Vermont, designated the Vermont Lowland by Jacobs (1950, p. 57), borders Lake Champlain throughout its entire length. It averages slightly more than twelve miles in width, measuring approximately sixteen miles in the latitude of Brandon, twelve miles at Burlington and St. Albans and twenty-five miles at the Canadian border.

Glacial Erosion and Deposition

The last glacier to invade the Champlain Lowland was the Fort Covington (Mankato-Port Huron) from the northwest. The ice direction, however, was altered to the southward by the abrupt, western slope of the Green Mountains. Striations show that the ice moved south 20° east to south 40° east in the northern part of the lowland, but the direction becomes more and more southerly to the southward. (Plate II) The basin now containing Lake Champlain was, according to Fenneman (1938, p. 220), a preglacial valley. The ice direction was diverted even more southerly in this valley, and ice erosion concentrated in this long narrow strip. The valley must have been scoured by the ice since the bottom of the lake basin is now some three hundred feet below sea level.

In contrast to the other areas of Vermont, where the rocks are steeply dipping, the rocks of the lowland, except for the homoclines, are essentially horizontal. The rocks are composed of Cambrian and Ordovician sediments, chiefly dolomites and limestones, that have been little altered by metamorphism. Because the fault blocks (homoclinal structures) are distributed over an area of flat-lying strata, two types of land forms were carved by glacial erosion. In the areas of the tilted fault blocks, the hills were reshaped by the glaciers, but they still have sharp narrow crests with sloping sides. The side slopes generally reflect the rock structure with a gentle slope in the direction of the dip and a steeper slope on the other. The horizontal strata, in contrast, were carved into low tablelands with flat tops and sharp, almost vertical, sides.

Glacial deposits are known to be quite thin and scattered on the lowland. According to MacClintock and Terasmae (1960, p. 235), the Fort Covington ice edge calved into the lake that expanded over the basin as the ice waned. Most of the detritus contained in the ice was accordingly distributed over the lowland by floating icebergs. The retreating ice did not, therefore, form till; but instead, the clastics were strewn over the lake bottom contemporaneously with the deposition of the lacustrine silts and clays. The evidence for a calving retreat, says MacClintock and Terasmae (1960, p. 237), is found in "the large number of ice-rafted, striated stones embedded in the varved sediment".

Lake Vermont

It is probable that the first surface deposits of Vermont to attract the attention of the early geologists were the lacustrine clays of the Champlain Lowland. The lake sediments were noted on the New York side of the Lake Champlain by Ebenezer Emmons (1838) in his report on the Second Geological District of that state. The manifestations of lake deposition have since been studied and noted by numerous geologists, notably C. H. and Edward Hitchcock (1861), Baldwin (1894), Upham (1877), Peet (1904), Fairchild (1916, 1919a, 1919b) and Merwin (1908). Of these, however, only Merwin did extensive field investigations in the course of his studies. The actual history of the lake episodes, none-the-less, had to await the work of Woodworth (1905) in New York and Chapman (1937) in Vermont.

The Chapman reports on the Champlain valley (1937, 1941) are quite complete. Since the present survey has not studied the area, this report can only summarize the Chapman report and modify the history of the lake episodes in accordance with more recent investigations, chiefly those of MacClintock (1945b, 1958), Terasmae (1959), MacClintock and Terasmae (1960), and MacClintock and Stewart (in press).

The data collected during the survey now in progress have led to the assumption that the Fort Covington ice covered the entire Champlain Lowland. Studies in the St. Lawrence Lowland [MacClintock (1958); MacClintock and Terasmae (1960)] have established that the lake episodes of this basin were associated with the waning stages of the Fort Covington glacier. It has been also established that this glacial maximum is of the same approximate date as the Mankato-Port Huron of the midwest (MacClintock and Terasmae, 1960).

The lakes that developed in the Champlain trough during the Fort Covington recession are collectively designated Lake Vermont. The lake actually included two lake stages that maintained different water levels and utilized different outlets. The first lake stage, the Coveville, was so named because the outlet of the water was through a col at Coveville, New York (Chapman, 1937, p. 95). In its early stages, the Coveville lake was quite small but as the ice waned northward the waters expanded with the calving edge of the Fort Covington glacier. At its maximum extent, the Coveville lake extended from the Coveville outlet northward to Plattsburgh, New York and Milton, Vermont. When the ice stood at this latitude, according to Chapman (1937, p. 101), the lake stage (Coveville) came to an abrupt end because of the development of a lower outlet.

With the drop of the waters of Lake Vermont to a new level, the Fort Ann stage was formed. Chapman (1937, p. 101) accounts for the water level change as resulting from the re-excavation of the Hudson gorge near Schuylersville, New York, which allowed the waters to overflow at a lower level. The outlet of this lake was near Fort Ann, New York, hence the naming of this stage. Soon after the lowering of the waters of Lake Vermont to the Fort Ann stage, according to MacClintock and Terasmae (1960, p. 237), the Fort Covington ice withdrew from Covey Hill, and the waters of the lake in the St. Lawrence valley lowered to the same level and were confluent with the Fort Ann lake in the Champlain Basin. The waters of Lake Vermont during this stage thus covered the whole Champlain Lowland in Vermont and extended northward as the lake in the St. Lawrence valley.

The shore phenomena marking the water levels of the two stages of Lake Vermont are now tilted, due to post-glacial uplift. Chapman (1937, p. 95) records the present elevation of the uppermost water plane of the Coveville lake as approximately 450 feet near Brandon, and rising northward to an elevation of about 700 feet in the vicinity of Milton. The Fort Ann lake beaches are similarly tilted and rise from 390 feet near Bristol to 591 feet at the International Boundary. The tilt of the water plane is not north-south but trends in a southeast-northwest direction (Chapman, 1937, Figure 14), and the Fort Ann shore is at an elevation of 749 feet on the New York side of the lowland at the Canadian border.

The Champlain Sea

After the lake episodes in the Champlain Lowland, the northern part of the basin was covered by a marine invasion. The manifestations of the ocean waters are found in beach and bottom sediments similar to the lacustrine materials except that they contain fossil shells of marine

invertebrates. The marine estuary extended up the St. Lawrence valley at least as far west as Ogdensburg (MacClintock and Stewart, in press) and southward into the Champlain valley to the approximate vicinity of Whitehall (Chapman, 1937, p. 113).

According to Chapman (1937, p. 115–6) the upper marine plane stands at an elevation of approximately 300 feet at the Sherman Falls delta southeast of Sherman Falls, Vermont and rises to 539 feet at the international boundary. There are no good features marking the marine limit south of the Burlington Quadrangle in Vermont. On the west side of the lowland, however, a deposit four miles north of Port Henry, New York, marks the southernmost evidence of the marine invasion noted by Chapman.

The uplift that tilted the water planes of both the lake and the marine episodes occurred after the marine invasion when an isostatic rise of the land elevated the region out of the sea. Since the isostatic rise followed the Champlain Sea, and brought an end to this episode, both the lacustrine and marine shore phenomena are today parallel with each other though tilted to the southeast. Chapman (1937, p. 95) suggests the possibility of a rise of the land prior to the Coveville lake stage, which may have been the reason for the separation of the waters of Lake Vermont from Lake Albany in the Hudson valley. The writer does not believe, however, that there is any evidence to support a conclusion of a pre-Coveville uplift.

Prior to 1958 it was assumed that the Champlain Sea invasion occurred immediately after the withdrawal of the ice from the lowlands. This assumption was based on the belief that the ice had depressed the land to the extent that the St. Lawrence and Champlain valleys were below sea level when they were uncovered by the last glacial retreat. It was therefore assumed that the marine deposition rested conformably upon the lacustrine silts and clays. During the course of his investigations in the St. Lawrence valley, however, MacClintock (1958) has studied areas where the upper foot or two of the varved sediment was "fractured into a sort of a breccia" and oxidized (MacClintock and Terasmae, 1960, p. 238). MacClintock and Terasmae (1960, p. 238) interpret these features as evidence of weathering and the drying out of the lake clay prior to the deposition of the marine sediment. The conclusion is, therefore, that the St. Lawrence valley, and therefore the Champlain basin, were not below the sea level of that time when the Fort Covington ice withdrew. Sea level at this time was 300 feet or more lower than at the present. The valleys were invaded by the marine waters because of eustatic rise of the

PLATE XIII

Figure 1. Gravel in kame terrace on Barnes Hill. Bedding resembles delta but ice-contact structures are abundant. One and one-half miles north-northwest of Waterbury Center.

sea. Later, in post-glacial time, the land rose above sea level by isostatic adjustment (MacClintock and Terasmae, 1960, Figure 6).

Mention has been made several times in this report that the Champlain Sea invasion is now believed definitely to correlate with the Two Creeks. The basis of such a conclusion is a radiocarbon date of 9500 years (BP) for the fossil shells of the sea deposits (Terasmae, 1959, p. 335). This date also prompted the correlation of the Fort Covington drift with the Mankato (Port Huron) since it is overlain by the marine clays (MacClintock and Terasmae, 1960, p. 239).

Lacustrine Deposits of the Winooski, Stowe and Dog River Valleys

The only tributary of the Champlain Basin that has been studied thus far during the present survey is the Winooski valley. The valley was mapped in the Montpelier and Camels Hump quadrangles during the 1959 field season. The Stowe and Dog River valleys, tributaries of the Winooski, have also been mapped. Opportunity to study the detail of the sand, silt and clay of lacustrine origin was offered by road cuts along

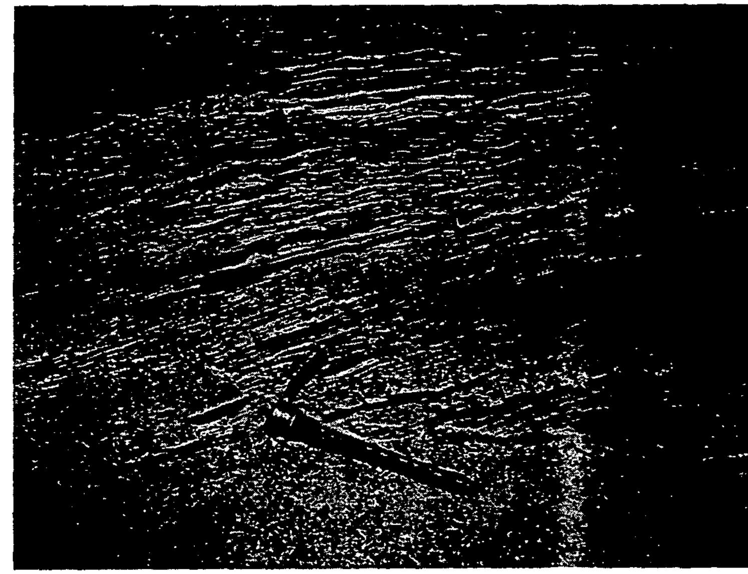

PLATE XIII

Figure 2. Bedding of laminated lacustrine silt. Dog River valley one mile north of Northfield.

the new interstate highway between Montpelier and Waterbury. The area covered by these quadrangles was so small, however, that a tracing of the lake levels was impossible, but enough data was collected to give an idea of the problems of correlation of the lake episodes in these valleys (Plate XIII, Figure II; Plate XV, Figures 1 and 2.)

Some of the strand lines discussed in the following pages may mark ice marginal lakes of very limited extent; in fact the writer is of the opinion that this is true in some areas for deposits at more than one level. Lakes of this kind commonly form along the margins of the ice, and since it is believed that stagnant ice remained in the valley for some time, this is a most likely probability. The following discussion does not assume this to be the case, however, inasmuch as more detailed study will be necessary to delineate the extent of all the shore phenomena.

The interstate highway excavations revealed that the varved sediments in the Montpelier-Waterbury area is covered by a great thickness of lacustrine sand. There appears to be no unconformity at the top of the varved sediment but it has been impossible thus far to ascertain whether or not the sand was deposited by the same lake episode. The varved clay measures as much as 75 to 100 feet in thickness, and the sand covers as much as 75 feet. It is apparent that at least one lake was in existence for a relatively long period of time (Plate XVI, Figure 2)

It was impossible to establish definitely all of the lake levels in this area during a single field season. Since most of the lakes extended beyond the bounds of the area presently mapped, the extent could not be determined. The shore phenomena were mapped, however, and elevations were taken with an aneroid barometer. The results have been tabulated, and it seems, at present, that shore deposits of at least seven levels were mapped with at least one other level being probable. These levels and their possible extent can be summarized as follows:

1. 1150 to 1200 feet—Stowe valley north of Stowe, probably the Lamoille valley
2. 1000 to 1050 feet—North Branch valley north of Montpelier (This is a probable level)
3. 875 to 950 feet—Winooski valley east of Middlesex, probably the Lamoille valley
4. 800 to 850 feet—Winooski, Stowe and Lamoille valleys
5. 750 to 775 feet—Winooski valley
6. 700 to 725 feet—Winooski valley, traced in Champlain valley to Underhill
7. 650 to 675 feet—Winooski valley—Champlain basin
8. 450 to 500 feet—Winooski valley west of Bolton-Champlain basin

Since the lake waters fluctuated and several levels were maintained, the higher shore deposits were reworked during subsequent episodes. No strand line is therefore continuous for any great distance. It is necessary, for this reason, to correlate from fragmental remnants in widely spaced localities. The most difficult problem in the identification of shore phenomena in this region is the finding of exposures of sufficient size and depth to show the structure. A deposit cannot be identified as a shore deposit unless the material and structure can be examined. This is particularly true of deltaic material where the foreset bedding is the most positive characteristic on which to base identification.

The correlation of the various lake levels is complicated by three factors. First of all, to the writer's knowledge, no systematic investigation of these lake levels has been made in which the strand lines have been actually traced from one area to another. The literature, particu-

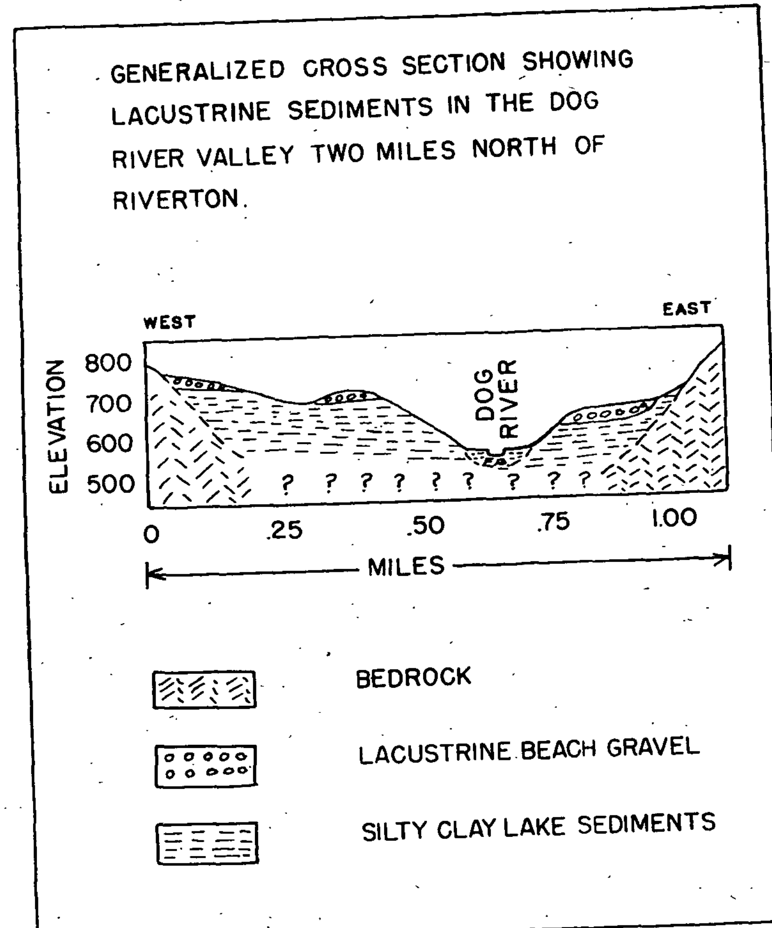

Figure 5. Generalized cross section showing lacustrine sediments in the Dog River valley two miles north of Riverton.

larly the writings of C. H. Hitchcock (1906 and 1908b), Fairchild (1916) and Merwin (1908), has noted the shore phenomena and has attempted a correlation. These surveys did not trace the shore deposits, however, and correlations were based on approximate elevations of widely scattered deposits. Many of the levels noted during the present survey have not been mentioned in former literature. It is the writer's conviction, therefore, that prior studies have not been of sufficient detail or of large enough extent to furnish a basis of correlation.

A second complicating factor, insofar as correlation is concerned, is the fact that the amount of post-glacial uplift has not been accurately determined for this region. Chapman (1941) has produced an isobase map of the Champlain Basin but this does not include the area to the east. Isobase maps have been prepared by other investigators but the accuracy of these is questionable.

The third complicating factor is that the present survey has not, as yet, covered enough area to ascertain the extent of these lake waters. As this work progresses, more detail will be available for calculating uplift and the extent of the strand lines of the various lake levels.

Possible Correlation of the Lake Levels of the Winooski, Dog River and Stowe Valleys

The preceeding discussion has emphasized that an accurate correlation of the various lake stages is impossible from data now available. The results of previous studies should be noted, however, and a discussion of the various lake levels and their possible correlation follows.

(1) *Level; 1150 to 1200 feet:* It is the opinion of the writer that this is the same strand as that which C. H. Hitchcock (1908b, p. 641) correlated with one stage of Glacial Lake Memphremagog. Surveys prior to the present, however, have not found this level to be above the 1100-foot contour. Whether or not this strand line can be traced to the Memphremagog basin remains to be seen, but for the present, at least, it should be tentatively matched with one of the stages of that sequence.

Merwin (1908) and Hitchcock (1906) suggest that this lake had an outlet through the divide between the Dog River and the White River at Roxbury. Fairchild (1906) assumed it drained outward through Williamstown Gulf. The data collected during the present survey, however, show that this lake did not extend into the Winooski valley, but that the water was blocked to the south of Stowe by the Fort Covington (Mankato-Port Huron) glacier. The ice impinged against the west side of Worcester Mountain and held the water at this high level. It is apparent that the water drained, at least for a time, along the ice margin (between the glacier and Worcester Mountain) and probably entered the Winooski valley through Middlesex Notch.

The shore phenomena proving the existence of this lake are quite definite. On West Hill, two miles north of Stowe, a pit in the gravel (one

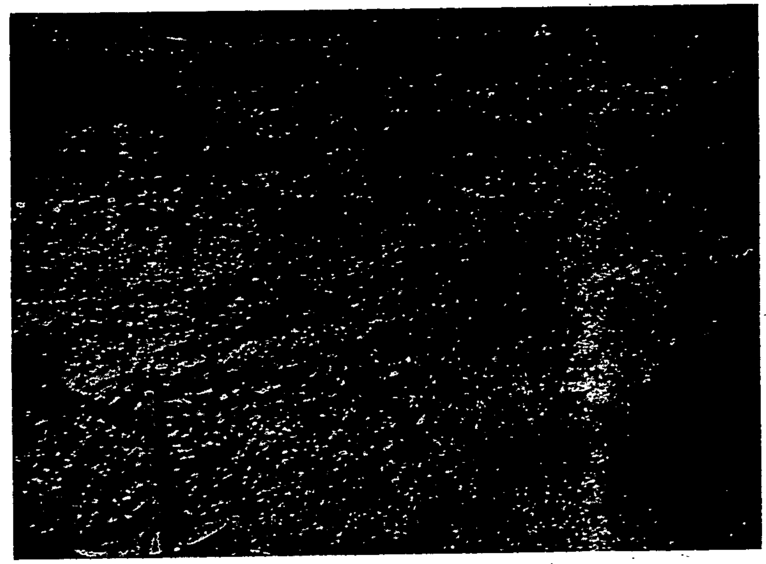

PLATE XIV

Figure 1. Beach gravel at elevation 1110 feet. West Hill two miles north of Stowe.

mile south of the old West Hill School) shows well the composition and structure (Plate XIV, Figure 1). The elevation of the top of the gravel is 1110 feet. One and one-quarter miles west of the pit, shore bars, traceable for over one-tenth mile, are most conspicuous on the slope of the hill (Plate XIV, Figure 2). The highest bar in this area was measured (by aneroid barometer) at an elevation of 1170 feet. On Sunset Hill, two miles west-northwest of Stowe another gravel deposit with surface characteristics similar to a bar was found at an elevation of 1200 feet. No openings were found in this ridge, however, and it was impossible to identify it definitely as a shore feature.

Sandy shore bars also occur on the west slope of Elmore Mountain in the Hyde Park Quadrangle. These are most conspicuous in the vicinity of the old, now abandoned, Delano School where the ridges run parallel to, and six-tenths of a mile west of, the Elmore Mountain Road. The bars here rise above the 1140 foot contour. Shore gravel has also been noted on the mountain slopes to the west of Morrisville. These have been identified at elevations between 1100 and 1200 feet on the eastern slopes of

PLATE XIV

Figure 2. Beach gravel terrace at elevation 1170 feet. Two miles north-northwest of Stowe.

Caper Hill. The southern part of the Hyde Park Quadrangle has not been mapped, and the detail of this area is yet to be studied but preliminary investigations established that the lake did extend into the Lamoille valley.

(2) *Level; 1000 to 1050 feet:* Definite evidence for a lake at this level has not been found and the extent has not been determined. This probable level is noted because of the small gravel deposits on the west side of the North Branch valley three to six miles north of Montpelier. Three deposits are found along tributary streams of the North Branch that crudely resemble deltas. Inasmuch as more study in this valley is anticipated, no definite correlation is proposed at this time. The tops of these do, however, seem to match one of the levels of Hitchcock's Glacial Lake Memphremagog, but there does not seem to be any possible col through which the two areas could have joined. The divide to the north between the head of North Branch and Lake Elmore is too high for any possible merger of the waters in the two valleys and, as has been stated

above, the Stowe valley south of Stowe was filled by ice at that time.

The writer is of the opinion that, if a lake level can be established at this elevation, the drainage must have been southward via Williamstown or Roxbury. It is also believed that the lake was relatively small and restricted in its extent to the eastern end of the Winooski valley. If this can be proven, the lake would no doubt correlate with that named Lake Montpelier by Fairchild (1906) in spite of the fact that both Fairchild (1916) and Merwin (1908) did not believe that the waters of this lake stood above 900 feet.

(3) *Level; 875 to 900 feet:* This level has been recorded in the literature by Hitchcock (1906), Merwin (1908) and Fairchild (1916) but it has not been too well defined by the present survey. The drainage of this lake was supposedly southward and restricted to the Winooski valley. It is probable that this level represents a lowering of the 1000 to 1050 stage.

(4) *Level; 800 to 850 feet:* This is a very common level in the Winooski, Stowe and Lamoille valley. The direction of the outlet for these waters has not, as yet, been ascertained. It seems that this level is too low to have drained southward into the Connecticut River. The elevation is also higher than any here-to-fore recorded for lakes draining into the Champlain Basin. Further study will be necessary to solve this problem but the most plausable hypothesis at this time seems to be that this water drained westward into an early stage of Lake Vermont.

(5) *Level; 750 to 775 feet:* This is probably the most common elevation of the shore phenomena of the area here described. This level definitely records a change in the drainage to the west, no doubt around the ice in the Champlain Basin as described by Chapman (1941). Whether or not this is the lake designated Winooski by Fairchild (1916) and Mansfield by Merwin (1908) is uncertain, but it seems probable that this is the same sequence. Since these waters occupied both the Winooski and Lamoille valleys, the name Lake Mansfield would be more appropriate.

This lake stage is marked by beach gravel deposits in the Dog River valley between Riverton and Montpelier and similar gravels south of the Winooski River at Middlesex. It is also recorded by a delta built into the lake north of the Winooski, four miles west of Montpelier. The lake has also been noted in the writings of Merwin (1908), Hitchcock (1906) and Fairchild (1916) in many areas of the Winooski and Lamoille valleys.

(6) *Level; 700 to 725 feet:* This is also a frequently found strand elevation of the Winooski valley. Beach materials of this lake were also traced into the Champlain Basin north of Richmond to the vicinity of Underhill. It is the opinion of the writer that this level marks a drop in

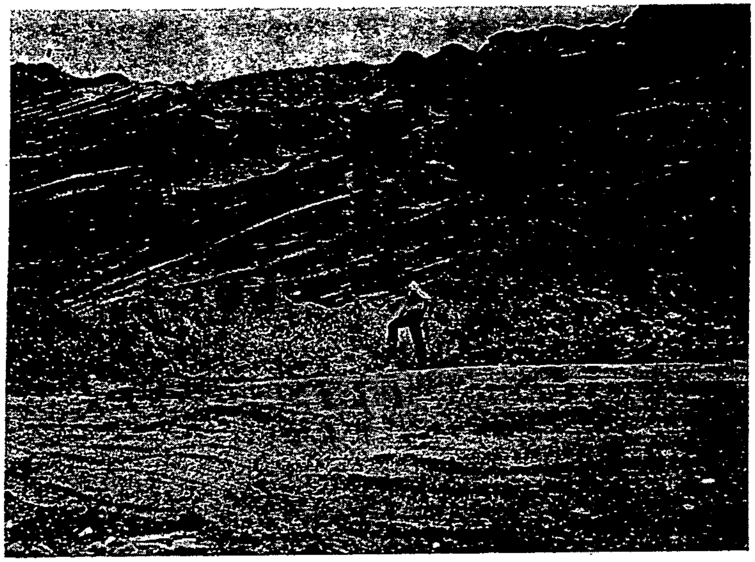

PLATE XV

Figure 1. Cross-bedding in lacustrine sand of the Winooski valley. Interstate highway cut one and one-half miles southeast of Middlesex.

the 750 to 775 lake that was caused by a melting back of ice in the Champlain valley or a gradual lowering because of channel cutting at the outlet. How far into the Champlain Basin it will be possible to trace this lake is not known but it is believed that these waters were held along the ice margin as ice melted down from the western slope of the Green Mountains. The extent of such shore features would therefore be restricted.

The beach of this lake is well manifested in Jericho Township. A pit in the gravel on the south side of the Lee River valley, one-half mile north north-east of Jericho Center shows ten to fifteen feet of gravel. The deposit follows the valley for about two miles, and the higher portions of the deposit are at an elevation of 725 feet. The highway between Jericho and Riverside (State Route 15) follows a rather extensive beach south of the Brown River. Two pits occur in the higher parts of this beach south of the highway, and the elevation of the highest bar is 710 feet.

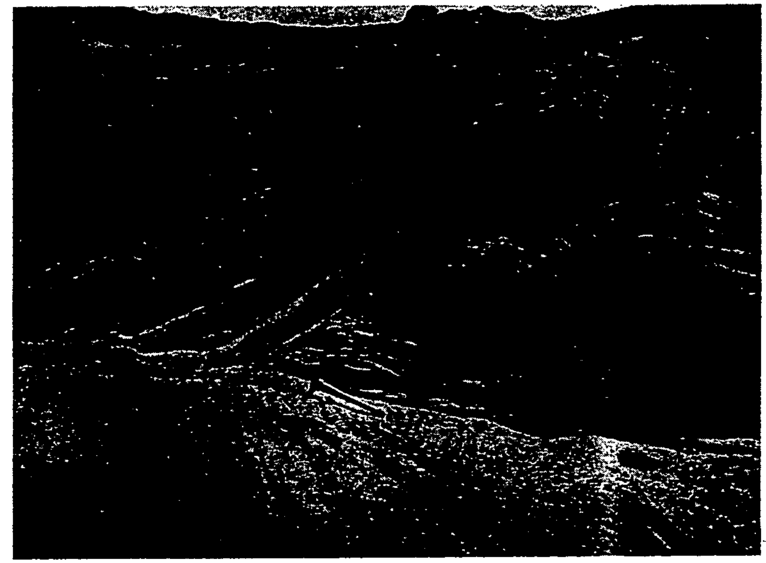

PLATE XV

Figure 2. Cut and fill in sandy lake sediments. Interstate highway cut one and one-half miles southeast of Middlesex.

(7) *Level: 650 to 675 feet:* This, of course, is the level of the Coveville stage of Lake Vermont in the Winooski valley. The strand can be traced up the Winooski to Barre and down the Dog River to Northfield. Several good beach deposits, ten to twenty feet in thickness, have been noted at this level in the two valleys.

(8) *Level: 450 to 500 feet:* This is the level of the Fort Ann stage of Lake Vermont. This lake was too low to extend up the Winooski valley past Bolton. Shore terraces made by this lake are found in the vicinity of Richmond and upstream to the mouth of the Huntington River. Chapman (1937) records the level of this lake as 500 feet at Richmond. The writer, with an aneroid barometer, determined the elevation to be 460 feet south of the Winooski and immediately west of the mouth of the Huntington River.

Proposed Correlation and Naming of the Lake Stages

From the foregoing discussion it can be surmised that the area here

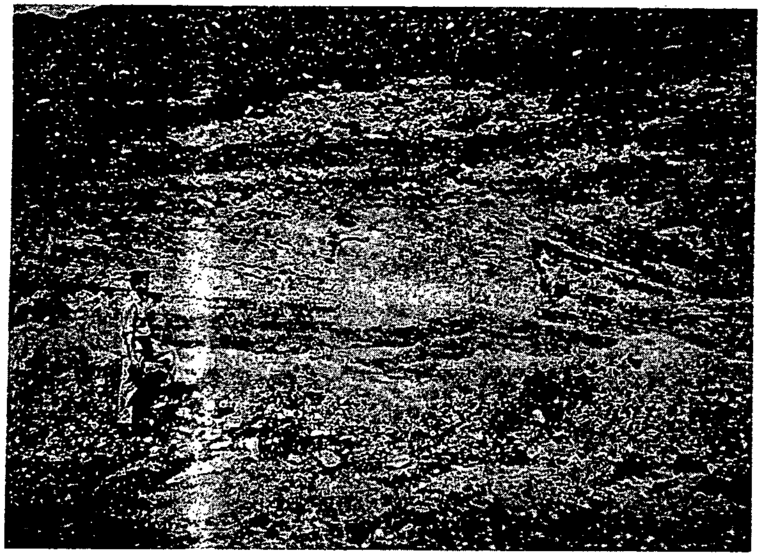

PLATE XVI

Figure 1. Topset and foreset beds of delta built into a glacial lake of the Winooski valley. Two miles west-northwest of Montpelier.

discussed contains a scrambled collection of lake shore phenomena that has been reworked by subsequent lake waters and dissected by later stream action and mass wasting. It would be a bit early, at this time, to make a definite correlation or to propose definite names for all of these lakes. It is the opinion of the writer, however, that as the lake deposits are mapped and the shore phenomena are traced from one area to the next it will be possible to correlate this maze of data with about five major lake episodes. For this reason, the following probable correlation, with names, is proposed at this time as a basis of future studies.

Glacial Lake Memphremagog: It is regrettable that this name, the same as an existing lake, has been so commonly used in the literature. Another name would be more appropriate, but it does not seem advisable to change it inasmuch as this would only add to the confusion. It is proposed here, assuming that the high level lake north of Stowe can be traced into the Memphremagog basin, that this high level be so desig-

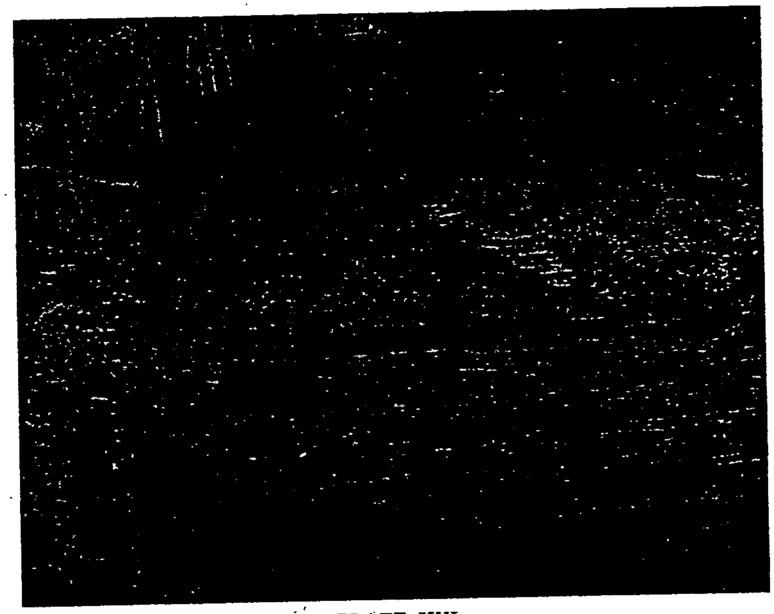

PLATE XVI

Figure 2. Varved clay covered by seventy feet of lacustrine sand. Interstate highway cut immediately north of Waterbury.

nated. Lake stages of this episode, as previously discussed, were held high by the Fort Covington (Mankato-Port Huron) when it blocked the valley south of Stowe.

All of the lake stages of this sequence occupied the Memphremagog basin, and some expanded into adjacent areas. No doubt several stages will evolve and, as supposed by Hitchcock, several outlets will be defined, including the one found south of Stowe.

In the event that the high level lake in the vicinity of Stowe does not extend into the Memphremagog basin, a different designator will have to be used. The extent of the lake, in this case, will no doubt suggest an appropriate name.

Lake Winooski: This name was originally proposed for a lake occupying the Winooski valley when it was assumed that the lacustrine sediments, particularly the varved clays, were deposited by a lake that was restricted in its extent to this valley. It is apparent now that no lake limited to the Winooski valley of the size and extent originally described ever existed. It seems that the lakes restricted to this valley were quite small, occupying only the upper reaches. The lake was contemporaneous in its early stages with Lake Memphremagog (in the Stowe valley), but when the ice blocking the Stowe valley receded, the higher waters dropped to the level of Lake Winooski.

As here proposed, therefore, Glacial Lake Winooski existed as long as drainage was southward into the Connecticut watershed via Williamstown and/or Roxbury. Even this does not restrict the lake to the Winooski valley because the waters expanded into the Lamoille valley as it was freed of ice. It is apparent that the shore deposits of this episode will lie between elevations of 875 and 1050 feet.

Lake Mansfield: The writer suggests that this name be used, as proposed by Merwin (1908), to designate the sequence of lakes that occupied both the Winooski and the Lamoille valley and drained westward into the early stages of Lake Vermont. As explained by Chapman, the outlet for these waters was south along the ice margin between the ice and western foothills of the Green Mountains and in the direction of the southern part of the Champlain Basin. Thus, Lake Mansfield covered the Winooski and Lamoille valleys as long as the narrows in these valleys through the Green Mountains were blocked with Fort Covington ice. The ice had at this time, however, stagnated and melted down in the upper reaches of these valleys. The range in elevation of the stages of this lake should be between 700 and 800 feet, and possibly as high as 850 feet, in the Winooski valley.

Coveville Stage, Lake Vermont: The strand lines of Lake Vermont have already been established, and the correlation is merely a matter of tracing them and measuring elevations. Chapman gives the elevation of this stage as 641 feet at Williston, and it is apparent that it rises to elevations as high as 675 feet in the Winooski valley.

Fort Ann Stage, Lake Vermont: This lake has also been definitely defined in the Champlain Basin. The elevation at Richmond, according to Chapman (1937), is 500 feet, and tracing it will not be a difficult problem.

THE GLACIAL GEOLOGY OF THE BURLINGTON AREA

A Report of Progress—1960

During the field season of 1960, the writer, assisted by Hugo F. Thomas, mapped the Burlington area as a continuation of the survey

begun in 1956. Because of the significant nature of the data collected, and because time did not allow the revision of this report to include these data, it was decided that this material should be included as a chapter in this report. The area mapped during the summer, included the Burlington Quadrangle, that part of the Willsboro Quadrangle within the state of Vermont, the Milton Quadrangle south of the Lamoille River and the southeast corner of the Plattsburgh Quadrangle. All of the elevations noted here were taken by aneroid barometer.

The Glacial Sequence in the Champlain Basin

The 1960 season was most rewarding inasmuch as the Burlington region offered more manifestations of the Pleistocene history than all of the preceding years. It has not been possible, as yet, to project the glacial chronology as interpreted in this area to include all of Vermont, but it seems logical to assume that this will be done as the survey progresses. From the investigations made and the data collected during the summer, the following Pleistocene sequence, in chronological order, can be deduced for the Champlain basin.

I. An interval of glaciation from the northeast with the deposition of dense till.
II. An interstadial lake episode followed, and varved lacustrine clays were deposited over the Champlain lowland.
III. Glacial readvance into the Champlain Valley from the north-northwest and the deposition of till over the varied clay where it was not removed by the advancing ice.
IV. Lake Vermont formed and occupied the Champlain Valley during the retreat of the latter glacier. Three lake stages are recorded in the deposits of this lake episode. From highest to lowest these are:
    a. Quaker Springs Stage
    b. Coveville Stage
    c. Fort Ann Stage
V. With the withdrawal of the ice dam in the St. Lawrence Valley, the waters of Lake Vermont were drained out completely leaving the Champlain lowland dry land subject to weathering and erosions.
VI. Marine waters slowly invaded the Champlain basin due to a eustatic rise of the sea. The sea rose to a maximum and was "stabilized" long enough to form a major beach as sea and land were rising at about the same rates.

VII. The land continued to rise and emerged to its present altitude above present sea level.
VIII. Lake Champlain developed
IX. Post glacial erosion

The Earliest Recorded Glaciation

The earliest known recorded glaciation in the Burlington area was an ice invasion from the northeast. It is probably that this ice covered all of Vermont, although the direction may have been more nearly north-south in the southern part of the state.

The northeast glacial advance was identified in the Burlington Quadrangle in a small stream valley, one and one-quarter miles south-southwest of Shelburne village. The valley walls at this location expose a layer of dark grey till over the bedrock, varying in thickness from three to twelve feet. The grey till is overlain by eight to fifteen feet of reddish brown, sandy till that is covered by four to eight feet of bouldery lacustrine clay. The basal dark grey till has a relatively high content of clay and it is quite compacted. The orientation of the pebbles in the grey till show a fabric with maximum approximately north 30° east. The fabric of the overlying till is north 15° west.

As has been noted earlier in this report, MacClintock (1958, pp. 6–12) described an older till from the northeast in the St. Lawrence Lowland which he designated the Malone till. It is now believed that this drift was deposited during the Cary substage of the Wisconsin. Although a definite correlation of the lower till at Shelburne with the Malone till is not possible at the present time, it does seem logical to assume that they were deposited during the same ice invasion.

Evidence of an Early Interstadial Lake

It has been assumed for many years that the earlier ice advance of the Champlain valley was followed by a lake episode as was the last glaciation. Field data to substantiate this belief however, had not been recorded until the 1960 field season. Evidence for a lake following the retreat of the early glacier was found in a single exposure in the southern part of the Burlington Quadrangle. Varved lacustrine sediment was found in the valley of Lewis Creek, two and three-quarter miles north-east of the village of North Ferrisburg. The lake sediment is compacted showing evidence of having been overriden by ice. The lacustrine deposit is overlain by a massive till twenty feet in thickness, and the till is covered by lacustrine material deposited during the Lake Vermont episodes. The fabric of the till is north-northeast.

PLATE XVII

Figure 1. Gillett Notch. Three miles south-southeast of Jonesville. Picture taken looking north through the gap.

The Last Glaciation of the Burlington Area

Till exposures are not common in the Burlington area. Till fabric studies at scattered outcrops and striae on the more resistant bedrock, however, definitely attest to the fact that the last ice invasion was from the northwest. A fresh, twenty-foot thickness of this till was exposed in the excavation for the foundation of the new men's dormitory at the University of Vermont. This structure is being erected at the intersection of Main and Spear Streets in Burlington. The exposed till is sandy, reddish-brown in color and has a fabric north 15° west. The striae on the bedrock under the till, exposed in the same excavation trend north 15° west to north 30° west. The texture, color and fabric of this till is the same as that overlying the older till at Shelburne and the older lacustrine sediment in Lewis Creek valley.

Not all of the till deposited by the ice from the northwest is the same color. Exposures of till were found that had the same fabric as the reddish-brown but the color varied from greyish brown to dark grey. Grey till with a northwest fabric was found in the valley of Muddy

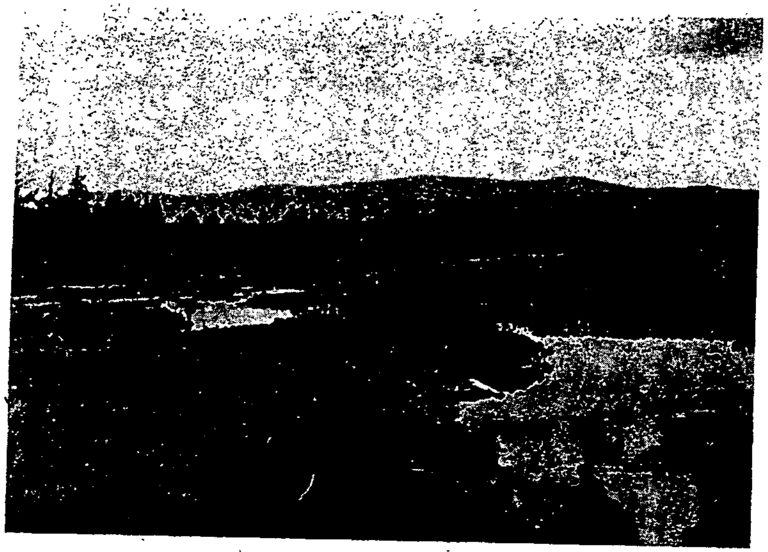

PLATE XVII

Figure 2. Poorly drained upland of the northeast highlands at East Brighton.

Brook on both sides of U. S. Route 2, three miles east of Burlington and along a headwater stream of the LaPlatte River, two miles southeast of Hinesburg. It is the belief of the writer that the variation in the color is due to the bedrock over which the ice moved. The Monkton Quartzite outcrops along the western side of the Burlington Quadrangle. The red color of portions of this formation adds color to the till. The Monkton also increased the sand content of the till, thus increasing the porosity and the depth of oxidation. To the west, in the Burlington Quadrangle, the bedrock is mostly dark shales and limestones and hence a grey color to the till.

It has already been noted in this report that ice from the northwest invaded the Stowe and Dog River valleys. These valleys, it is now believed, mark the eastward extent of the Northwest glaciation. The southern limit of this ice, however, is yet to be determined.

The kame terraces that overlap the foothills of the Green Mountains in Hinesburg Township were no doubt deposited in association with the northwestern ice advances. There have been several different interpretations as to the origin of these deposits (Chapman, 1937, p. 99; Merwin,

1908, p. 119; Fairchild, 1916, p. 24), but the present survey agrees with Chapman that they are kame terraces since they show ice-contact structures. As will be pointed out later in this report, however, these terraces have subsequently been modified by the wave action of two stages of Lake Vermont.

It is too early to propose a definite correlation for the glacial drift with a northwest fabric. But, as already been noted in this report, MacClintock (1958, p. 6) described a younger till from the northwest in the St. Lawrence Valley which be designated the Fort Covington. It is now believed that the Fort Covington is of Mankato (Port Huron) age (MacClintock and Terasnae, 1960, p. 239). This report concludes that the northwestern fabric of the younger till in the Burlington area strongly supports a conclusion that it is a correlative of the Fort Covington.

Lake Vermont

The lake episodes that developed during the retreat of the last glacier in the Champlain Basin have been collectively called "Lake Vermont." A calving retreat of the ice is definitely suggested by the large number of glacially striated boulders embedded in the lacustrine sediments all over the Burlington Quadrangle. One of the best exposures exhibiting the bouldery nature of the lake deposits is located along the Lake Champlain shore at the mouth of Holmes Brook. The calving retreat no doubt explains the absence of till over most of the region. It has been assumed that there were two stages associated with this lake, but the present survey identified a third lake stage during the 1960 season.

The Highest Lake Vermont—The Quaker Springs Stage

Woodworth (1905, p. 193) described shore features in New York which he believed to mark an early (high) stage of Lake Vermont. The higher lake he called the "Quaker Springs Stage." Chapman (1937, p. 97), noted Woodworth's descriptions but discounted them on the belief that they were formed by local, high-level ice-marginal lakes. The present survey in 1959 identified a series of high-level lake-shore phenomena in the Jericho Center-Underhill area and these have already been noted in this report. During the summer of 1960, this shore has been traced from Starksboro Township northward to the Lamoille River and it is concluded that a pre-Coveville lake stage did occur. Since the name "Quaker Springs" has already been used for similar deposits in New York, it seems that this name has priority.

Shore phenomena of the Quaker Springs lake are well developed and it is concluded that the lake was in existence for as long a time as the Coveville stage or perhaps longer. The deposits seem to be larger and more widespread than either those of the Coveville or the Fort Ann episodes in the Burlington region.

The most southerly of the Quaker Springs deposits investigated during the summer was along the east side of State Route 116, one-half mile due east of Rockville. This locality is two-tenths of a mile south of the southern border of the Burlington Quadrangle. Here, a shore terrace was cut into the drift covering the mountain slope and a beach gravel deposit extends from the highway up the slope to an elevation of 695 feet. The terrace at the base of the wave-cut cliff is covered with a concentration of large subangular to rounded boulders. Two or three low, short bars were also found on the terrace. This shore feature is traceable from this point northward for approximately three miles to the Hinesburg kame terraces.

The Quaker Springs shore in this area is at the same level as the flat top of the Hinesburg kame terraces and is the opinion of the writer that the terraces were levelled off by the waves of the Quaker Falls lake. This conclusion is supported by the valleyward slope of the gravel on the outer margins of the terraces. Gravel pits penetrating deeper than the sloping beds show ice-contact structures of kame deposits. It is therefore evident that structures found in these deposits cannot be attributed to a single activity.

Another significant lacustrine deposit above the Coveville level is found along Johnnie Brook, one-half mile south of Fays Corners. Here, a lacustrine deposit of laminated silts and clays is covered by a deposit of gravel. The top of the deposit is at 740 feet. The shore features of this lake stage at Jericho Center and Underhill have already been described in this report.

Although a shore of Quaker Springs stage is not found on North Williston Hill, the evidence of a lake higher than the Coveville is none the less manifested. The summit of this hill is at an elevation of 703 feet, near or somewhat below the level of this lake. The Coveville shore is at 625 feet. Beach gravel can be traced up the southwest side of the hill to an elevation of 685 feet. On the north and northeast sides of the hill is a sloping, wave-washed plane that was carved from the till on the slope. The plane has an elevation of 685 feet at the north summit, and it slopes in a northeast direction. It is concluded that the waves came from a north or north-northwest direction to form the wave-washed plane and deposit the gravel. It is assumed that wave base was below

the summit of the hill. Another small hill immediately north of the village of Williston (one-half mile southeast of North Williston Hill, also has beach gravel deposits on the slopes to an elevation of 685 feet.

The most northerly of the shore deposits thus far mapped is located on Prospect Hill, two miles south-southwest of Fairfax. The shore overlaps the till on the slopes of the hill and a low, but distinct shore cliff is cut into the till. The elevation of the top of the beach gravel at the base of the cliff is 750 feet. Beach gravel is widespread on the hill and is traceable for approximately four miles.

Another deposit that should be noted here is at the head of Abbey Brook, two and one-half miles north of Jericho, where a sand and gravel deposit extends in a northeast direction up to a swampy area at the top of the hill. The present survey tentatively identifies this deposit as a delta built out into the Quaker Springs Lake by a stream from the Northeast. Time did not allow the study of the area to the east (Mt. Mansfield Quadrangle), however, and a definite interpretation of this deposit will have to await further study. The sands and gravels surely must be associated with the Quaker Springs lake episode.

The Coveville Lake Stage

The Coveville Stage of Lake Vermont was given detailed coverage by Chapman (1937, pp. 95–102) and a summary of his findings has already been included in this report. It is necessary, therefore, that this chapter add only the significant findings of the current season's work.

The most southerly and the best-developed Coveville shore terrace is on the sides of the hill just north of Monkton Ridge. The summits of this hill rise to 800 feet, and a well-developed terrace and wave-cut cliff of the Coveville lake are quite conspicuous on both the eastern and western sides. The terrace, at an elevation of 585 feet, was cut into till, and it is veneered with beach gravel and strewn with a large concentration of large boulders.

The Coveville waters also cut into the slopes of the kame terraces at South Hinesburg, forming a bench that is most conspicuous one and one-half miles north of South Hinesburg village. The gravel that was removed by the terrace-cutting processes was carried out over the slopes of the kames forming the valleyward sloping beds that are seen in the gravel pits of that area. The kames have been thus modified by the wave action of two lake episodes, and it is probably that the polygenetic origin of the structures accounts for the several different classifications of them in the past.

The Coveville shore was described by Chapman (1937, p. 99) on Mt.

PLATE XVIII

Figure 1. Two tills exposed in stream valley, one and three-tenths miles south-southwest of Shelburne Village.

Philo (elevation 540) in the southwestern part of the Burlington Quadrangle. The same shore is marked by a large beach gravel deposit on the south side of Pease Mountain. The elevation at the top of the gravel is 545 feet. Scattered gravel and sand accumulations of the shore of this lake were also found in the Brownell Mountain-Mt. Pritchard area, but shore features were absent and no elevations were measured.

At Fay's Corners, the Coveville beach gravel has a top elevation of 620 feet and the gravel at this locality is thicker than those described above. There is also a thick beach gravel deposit on the southeast side of North Williston Hill. The elevation at the base of the wave-cut cliff is 625 feet. Chapman records a Coveville elevation of 641 on this hill.

The Fort Ann Lake Stage

The Shore phenomena of the Fort Ann Stage of Lake Vermont (Chapman, 1937, pp. 103–113) are poorly developed and quite scattered in the Burlington area. Chapman noted the shore manifestations on Mt. Philo, the Winooski delta and the terrace on Cobble Hill.

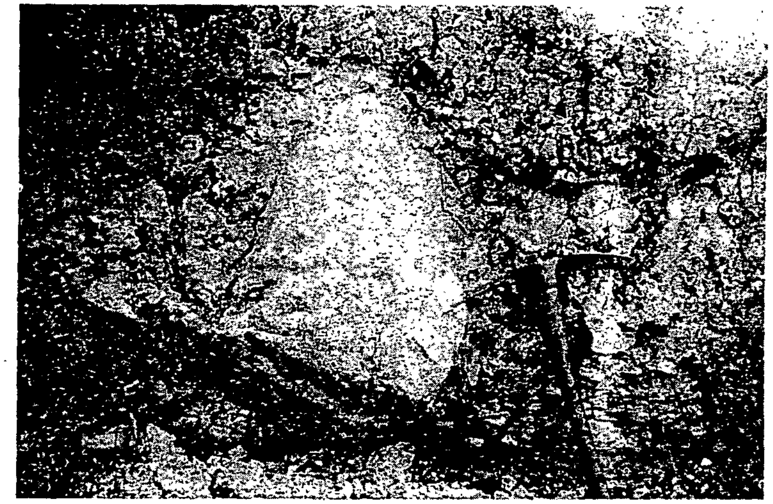

PLATE XVIII

Figure 2. Glacial erratic thirty inches across embedded in varved lacustrine clay. Champlain shore at the mouth of Holmes Brook.

The two most noteworthy localities noted during the present survey were on Pease Mountain and Jones Hill. The Fort Ann terrace on Pease Mountain is cut into the beach deposit of Lake Coveville at an elevation of 460 feet. On Jones Hill, the Fort Ann beach-gravels rise to an elevation of 465 feet east of the Charlotte cemetery and a prominent terrace is cut in till at the same elevation to the west of the cemetery. This shore is better developed than the beach on Mt. Philo at 540 feet, but the elevations are so nearly the same that, together, with the Pease Mountain beach, they make a good marker for the lake level in the southwestern part of the Burlington Quadrangle.

A small patch of beach gravel, at an elevation of 525 feet, one-half mile west of North Williston hill is the only other beach deposit of the Fort Ann lake in the Burlington Quadrangle, if indeed this is a deposit of that lake. In the southern part of the Milton Quadrangle, one-half mile north of Butlers Corners, is a gravel deposit that the writer believes should be a Fort Ann beach. The elevation of the gravel, and the wave-washed till to the north of it, are at an elevation of 580 feet. This is several feet higher than the Fort Ann level has been presumed to be in this area, and the present survey has not progressed far enough north to correlate the deposit with others of that area.

Pre-Champlain Sea Erosion

With the withdrawal of the Fort Covington glacier and the removal of the ice dam in the St. Lawrence valley, the waters of Lake Vermont drained out of the Lake Champlain basin. It is believed that sea level was lower than at the present time, and during the interval the Champlain lowland was dry land. The whole region was therefore subjected to weathering and erosion before the invasion of the Champlain Sea.

It must be noted here, however, that evidence of the erosion interval has not been found in the Burlington area. Manifestations of a pre-Champlain Sea erosion have been reported in the St. Lawrence Valley (MacClintock and Terasmae, 1960, p. 238), however, and the same conditions doubtlessly existed in the Champlain basin. The only known areas where the erosion might be inferred in the Burlington area are along the Winooski River upstream from Essex Junction. A more or less continuous terrace has been traced along this section of the river at elevations of 350 to 400 feet. It has formerly been believed that this terrace was formed during the Champlain Sea episode when the Winooski cut down to the level of the sea. The possibility that the river had already cut down lower than the sea (during the erosion interval) is a question that has not been considered by prior studies. This is a distinction that cannot be easily made, but deposits on the terrace with characteristics closely resembling beach gravel might be a key to the solution of this problem. Future work, none the less, will have to be done, and it is hoped that better evidence will be found to the north of the Burlington Quadrangle.

The Champlain Sea

Following the erosion interval, the eustatic rise of the sea caused a slow invasion of marine waters into the Champlain basin to form what has been named the Champlain Sea. The sea rose to a maximum and remained stationary in relation to the land to form a major shoreline.

The Highest Marine Shore

The highest marine level is well-marked in the Burlington region, and it can be traced with more accuracy than any of the lake shores. The most southerly of the shore phenomena are on a ridge two miles south of the village of Charlotte. The fossiliferous beach gravel covers the top

and western side of the ridge to an elevation of 275 feet. Pebbly beach sand marks the shore at the same elevation between the ridge and Jones Hill just north of the village. From Charlotte, the shore can be traced almost continuously to Burlington. The excavation for the foundation of the new wing of the DeGoesbriand Memorial Hospital was dug in the beach deposits of the sea. Because of man-made changes in this area, it is difficult to obtain a reliable elevation, but is estimated that the top of the gravel is approximately 340 feet. Between Shelburne Village and Burlington, the sands of the Winooski River delta lap upon the sides of the hill, and in most areas the sand is capped with beach gravel that spreads out from a prominent sea cliff. The deltaic structure of the sand was well-exhibited in a sand pit in South Burlington, one and one-half miles due east of Red Rock Point. The marine origin of the deposit was shown by the fossils contained in the foreset beds. The Winooski Delta sands are found all over the South Burlington, Burlington and Winooski sections since the sands were spread out from the Winooski Valley at Essex Junction.

The Withdrawal of the Champlain Sea

After the sea had remained stationary with respect to the land for a short time, the land finally began to rise out of the sea. Marine beach deposits at various levels below the high shore indicate that the withdrawal was relatively slow and more or less continuous. Chapman, (1937, pp., 116–121) proposed three distinct stages in the lowering of the marine water when the water level was again stationary in relation to the land. The findings of the present survey thus far do not support this conclusion inasmuch as no definite water plane can be established from the data collected. It seems more likely that the water withdrew at a slow but constant rate and that shore phenomena developed because of the conditions along the shore. All of the better beach deposits below the high marine level seem to be located in areas that were protected by high rock outcrops or other topographic features.

The isostatic rise of the land continued until the land was elevated to its present altitude. After the marine waters withdrew, Lake Champlain developed in the Champlain basin. Erosion has subsequently modified much of the land surface to form the present topography.

The Monkton Quartzite Boulder Train

The Monkton Quartzite boulder train (indicator fan) has been mentioned earlier and is located on the ice direction map (Plate II) of this report. As has been pointed out, the information concerning the boulder train was from the literature inasmuch as the present survey had not studied the Burlington area until the 1960 field season. After having worked a summer in that region, the writer feels that some clarification of this indication fan should be made.

The Monkton Quartzite, because of its texture, slight metamorphism and color is a very distinctive rock. It is easily identified, and it is conspicuous in the deposits in which it is found. For this reason, it is a good marker. It should be pointed out, however, that the size of the area in which this formation outcrops limits its use as an indicator of ice direction. Whereas the most desirable boulder trains spread out from a distinctive rock outcrop of small areal extent, the Monkton Quartzite is exposed for a distance of approximately sixty miles in a north-south direction on the Champlain lowland. It is, therefore, impractical to use this rock locally as a measure of the direction of ice movement. It should be used only in a general way in regions farther removed from the area of outcrop, as, for example, east of the Green Mountains.

The Essex Esker

The only esker that has been noted in the five years of the current survey was mapped in the Burlington area. For this reason, this feature has special interest in that area, in spite of the fact that it has little geologic significance.

The esker is located along Alder Brook, three-quarters of a mile north-northeast of Essex Center in Essex Township. Thus, it is designated the "Essex Esker" in this report. The gravel ridge runs generally north-south for a distance of two and one-quarter miles, exhibiting the characteristic sinuous trend of the textbook description. The esker is mostly along the eastern side of Alder Brook, but it splits into two ridges at the south end and one of the prongs is on the western side of the stream. The extreme northern end also crossed to the west side of the brook. The esker is higher and the material is coarser at the south than at the north.

THE GLACIATION OF THE VERMONT VALLEY

The Vermont valley is a long, narrow lowland that is located between the Taconic and Green Mountains. In Vermont it trends generally southward from Brandon to East Dorset and thence south-southwest to the Vermont-Massachusetts boundary south of Bennington (Figure 2). The length of the valley is approximately 85 miles (in Vermont), and

the width varies from one to five miles. There is no stream through the valley, the drainage being through portions of Otter Creek, Batten Kill, and the Walloomsac River and their tributaries.

The writer believes that the last glacier to invade the Champlain Lowland, the Fort Covington, moved down the Vermont valley. Because of the narrowness of the valley and the heights of the steep walls on either side, the basal ice moved southward and scoured to considerable depth. To the writer's knowledge there are no published figures as to the depth to bedrock or the thickness of the sediment flooring the valley.

Various types of sediment cover the valley floor. Much of the material is kame gravel, but reports of material encountered in the drilling of water wells indicates that there might be a sequence of till and outwash composed of several different layers. Gordon (1942, p. 1847) stated that his study of the outwash gravel indicates that all of "the various stages of deglaciation" represented were formed in association with stagnant ice. The absence of frontal moraines here, as elsewhere in Vermont, are manifestations of stagnation. Gordon (1942) believed that "much of the till on the mountain slopes also suggests that the ice melted down from the mountain summits."

THE GLACIATION OF THE TACONIC MOUNTAINS

The Taconic Mountains, in the extreme southwestern part of the state, extend for 78 miles from the Massachusetts border to the vicinities of Brandon and Sudbury (Figure 2).

Of all the areas of Vermont, the data on the glacial geology of this region is the least known. Like the Green Mountains, this area was first studied to ascertain whether or not it had been glaciated and later to determine the type of glaciers that had been active there. The summits of the mountains are covered with glacial materials and poorly preserved striations trending in a southeasterly direction (Hitchcock, C. H., 1904, p. 76). This same trend has been noted on the higher areas of the Taconic peaks to the south in Massachusetts.

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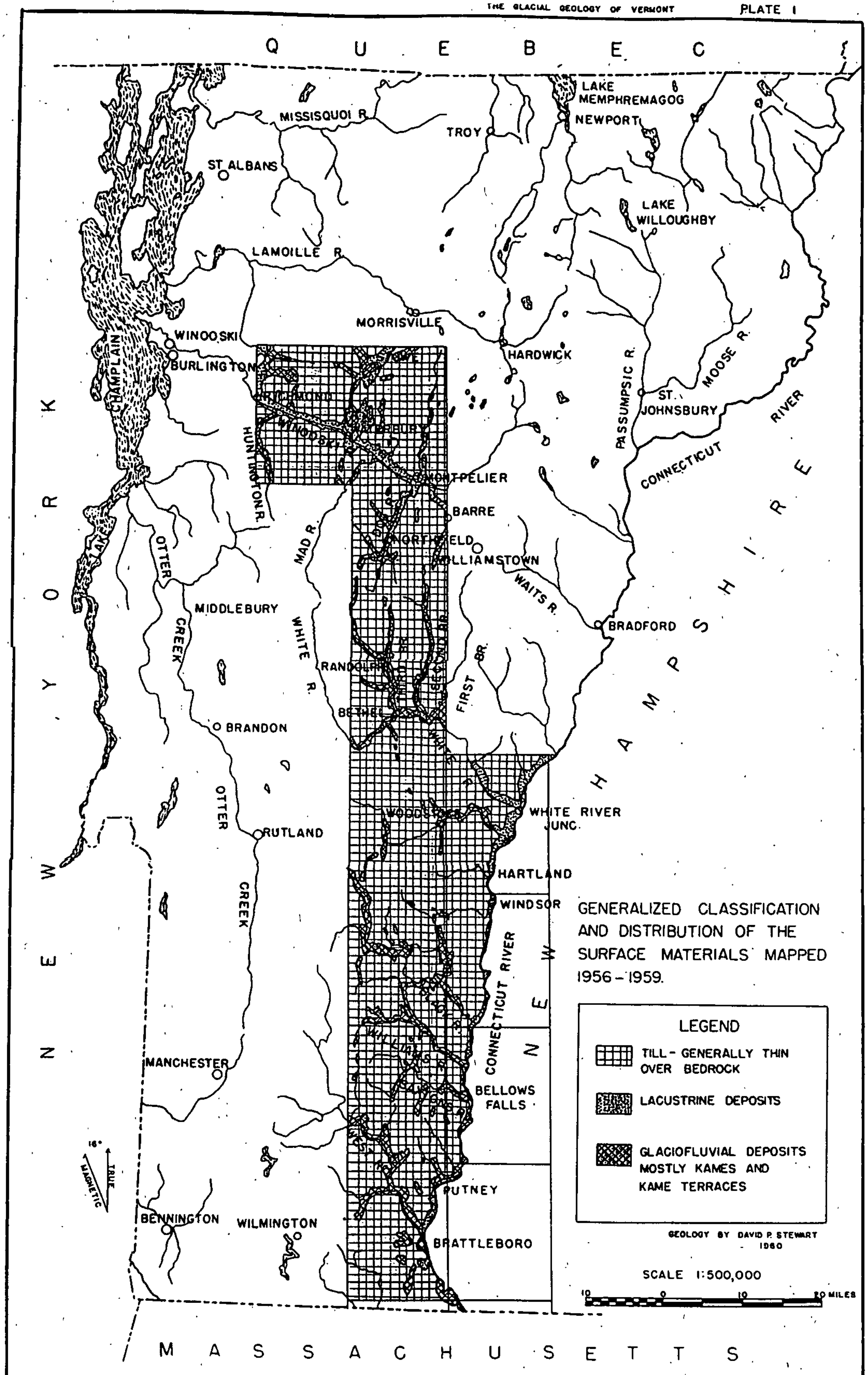
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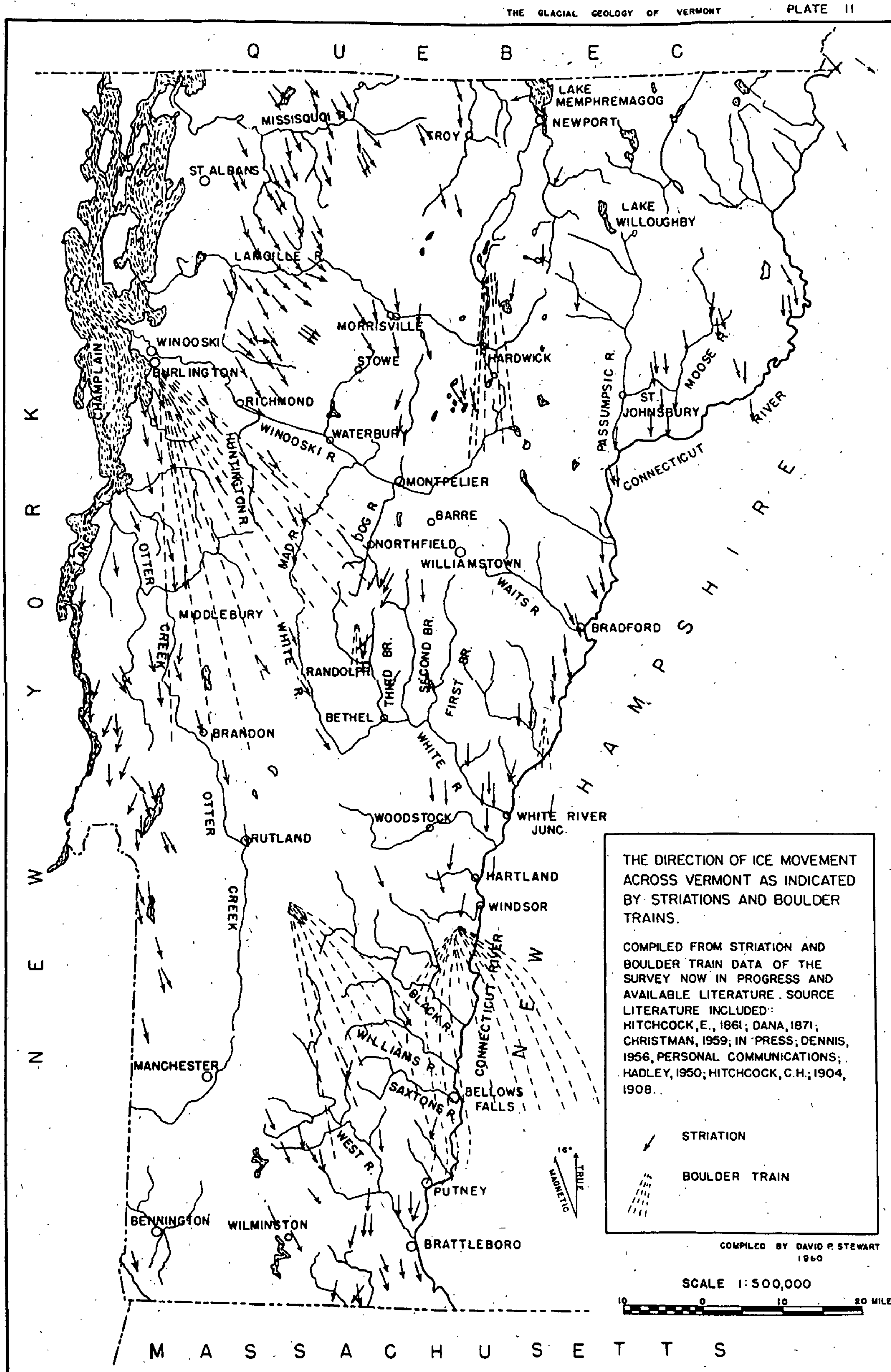
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BRATTLEBORO
THE DIRECTION OF ICE MOVEMENT ACROSS VERMONT AS INDICATED BY STRIATIONS AND BOULDER TRAINS.
COMPILED FROM STRIATION AND BOULDER TRAIN DATA OF THE SURVEY NOW IN PROGRESS AND AVAILABLE LITERATURE. SOURCE LITERATURE INCLUDED: HITCHCOCK, E., 1861; DANA, 1871; CHRISTMAN, 1959; IN PRESS; DENNIS, 1956, PERSONAL COMMUNICATIONS; HADLEY, 1950; HITCHCOCK, C.H.; 1904, 1908.
STRIATION
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