Pleistocene Geology of the Northeast Adirondack Region, New York

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By CHARLES S. DENNY

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A description and interpretation of ice fronts and water bodies in parts of the St. Lawrence and Champlain basins

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PLEISTOCENE GEOLOGY OF THE NORTHEAST ADIRONDACK REGION, NEW YORK

By Charles S. Denny

ABSTRACT

The northeast Adirondack region includes Clinton County and parts of Franklin and Essex Counties in the northeast corner of New York State. Lowlands west of Lake Champlain and southeast of the St. Lawrence River are underlain chiefly by Paleozoic carbonate rocks and are bordered by a dissected plateau held up by a massive quartzitic to arkosic sandstone (Potsdam Sandstone of Cambrian age). The adjacent uplands and mountains are underlain by Precambrian rocks, chiefly granitic and other gneisses, anorthosite, metagabbro, and metasedimentary rocks; in many valleys are thick masses of sand and gravel. In the lowlands the till is a calcareous silt loam containing a few clasts of sedimentary rocks; in the plateau areas underlain by Potsdam Sandstone, the till is a sandy loam or loam containing 10–20 percent sandstone clasts; and in the mountains the till is pebbly and sandy.

Directional data from striae, drumlins, and grooved drift and a study of till lithology by A. W. Postel indicate that the upper Wisconsin (Woodfordian) ice sheet moved southwest across the mountains and in a more southerly direction in adjacent parts of the St. Lawrence and Champlain Lowlands. However, the surface till in two small areas north of the mountains near Ellenburg contains many clasts of rocks found only in the mountains to the south, suggesting northward transport of these clasts by glacier ice or other processes prior to late Wisconsin time.

In the mountains, the history of deglaciation involves a series of episodes of moraine building, outwash deposition, and drainage diversion; in the lowlands, ice-dammed lakes lengthened northward, to be replaced eventually by the Champlain Sea. Deglaciation began with the building of massive outwash plains by southwest-flowing melt-water streams in the southwestern part of the area, probably not earlier than about 12,700 years B.P. Deglaciation proceeded in a general northeast direction and involved thinning of the marginal zone of the ice sheet. In the mountains, parts of the terminus stagnated, and melt-water streams built kames and outwash plains on and adjacent to masses of dead ice. In the lowlands, the ice sheet maintained an active front and built small moraines and ice-marginal kames. In the lowlands north of Plattsburgh, an esker about 10 miles long was built by a subglacial stream that discharged south into glacial Lake Vermont.

The principal streams, the Salmon, Trout, and Chateaugay Rivers on the edge of the St. Lawrence Lowlands in Franklin County and the Great Chazy, Saranac, and Ausable Rivers in the Champlain drainage basin were ponded at the retreating ice front and diverted across interstream divides, cutting channels, now abandoned, in drift and bedrock, removing the drift from large areas of bedrock, and emptying into the glacial lakes in the adjacent lowlands.

Along the western edge of the Champlain Lowlands, small moraines were built in or near the mouths of valleys draining northeast. Northeast-flowing streams were diverted southward along the ice edge where they eroded channels in drift and in rock. The Champlain Valley ice lobe held in glacial Lake Vermont and was probably only a few tens of miles long. The edge of the lobe retreated northward, for the oldest ice-marginal features are to the south. The Saranac River was diverted south into the Ausable River, perhaps contributing to the formation of a large delta built by the Ausable River in Lake Vermont near the end of the Coveville stage, perhaps about 12,600 years B.P. A prominent moraine dams the Saranac River valley where it enters the lowlands. Abandoned channels in the moraine indicate that when the ice built the eastern part of the moraine, the Saranac River entered Lake Vermont only 2 or 3 miles south of the present course of the river. A prominent moraine near Ellenburg Depot in the Great Chazy drainage was built after that on the Saranac, perhaps at the beginning of the Fort Ann stage of Lake Vermont about 12,400 years B.P.

As deglaciation proceeded, glacial Lake Iroquois, an ice-dammed lake in the St. Lawrence Lowlands southwest of Montreal, overflowed to the east across the divide into the Champlain Valley, cutting a deep rock-walled gorge across the divide just north of the International Boundary near Covey Hill, Quebec. Water escaping over the divide flowed southeast along the ice front into an ice-dammed lake in the valley of the Great Chazy River. This lake overflowed, perhaps catastrophically, to the southeast across the Great Chazy–Saranac River divide, where it washed clean Flat Rock, an area of essentially bare Potsdam Sandstone near Altona that measures about 2.5 by 5 miles.

Farther northeasterly retreat of the ice front caused the water coming through the gorge near Covey Hill to flow south and empty into a small ice-dammed lake in the English River valley. This lake in turn overflowed to the southeast along the ice margin into another ice-dammed lake, cleaning off the bedrock ridge between the lakes. In this fashion, several bedrock ridges between valleys were washed clean of their drift cover.

The final withdrawal of the ice sheet from the northeast prong of the Adirondack uplands caused Lake Iroquois in the St. Lawrence Lowlands to drain down to the Fort Ann

level of Lake Vermont in the Champlain Valley, and the two lakes merged. This withdrawal may have taken place about 12,200 years B.P.

The strandlines of the late-glacial water bodies in the New York part of the Champlain drainage basin are marked by beaches and deltas. Beaches are prominent features where the drift is pebbly to bouldery, that is, in areas of Potsdam Sandstone. Beaches are best developed in the bouldery deposits of the ice-marginal streams. Deltas were built where the principal streams from the mountains emptied into the late-glacial water bodies. Deltas built at low levels are composed in part of material eroded from those at higher levels.

The presence of beaches of the Fort Ann stage in areas of washed bedrock and of Fort Ann deltas whose tops are below the highest stand of that stage suggest that the level of Lake Vermont rose perhaps 50–75 feet during Fort Ann time.

The Champlain Sea came into existence when the retreating ice front reached a position a short distance north of Quebec City, thereby allowing the sea to invade the St. Lawrence and Champlain basins. In the Champlain Valley the marine submergence lasted from about 12,000 to about 10,500 years B.P. The deposits of the Champlain Sea cannot be distinguished lithologically from those of glacial Lake Vermont; the distinction is based on altitude and the remains of marine or brackish-water organisms, largely mollusks.

Streams from the mountains built deltas in the sea. The size of the deltas appears to be related to the size and average slope of the drainage basins supplying the sediment. This relationship suggests that the environment of Champlain Sea time was not periglacial but more like the present. The older Lake Vermont deltas do not show such a relationship; they were built largely by melt-water streams.

Lake Champlain came into existence about 10,000 years B.P., when gradual uplift caused the marine level to fall until the connection with the ocean was cut off. Except for the Ausable River, none of the larger streams emptying into the lake has built a delta into it comparable with those of earlier date. The crows-foot delta of the Ausable River is unique. This delta was built sometime after the formation of Lake Champlain, when the river's course below the chasm was changed by piracy. Delta building was apparently much more rapid during Champlain Sea time than it has been during the life of the modern lake.

INTRODUCTION

The northeast prong of the Adirondack Mountains rises between the broad St. Lawrence and Champlain Lowlands. During deglaciation the ice sheet blocked these valleys, ponding the streams that flowed north and east from the mountains. Some streams were diverted across divides into adjacent valleys. In the mountains, the marginal zone of the ice sheet stagnated, and kame terraces were built by ice-marginal streams. In the broad lowlands, the ice sheet maintained an active front and built small moraines largely of water-laid materials. As the ice front retreated out of the mountains, the ice-dammed lakes grew in size until glacial Lake Iroquois (Coleman, 1937) in the St. Lawrence Valley overflowed across the divide into glacial Lake Vermont (Chapman, 1937) in the Champlain Valley. The overflow washed clean large areas of bedrock. The ice front moved back and forth, opening and closing lake outlets. Water levels fell, rose, and fell again. Ultimately salt water invaded the St. Lawrence River lowlands, initiating the Champlain Sea episode. Deltas and beaches record the presence of the ancient water bodies that smoothed the contours of the glaciated landscape.

The author spent about 14 months in the field, from 1961 to 1969, and mapped in detail the lowland and adjacent foothills west of Lake Champlain between the Ausable River and the Canadian border; this includes the Dannemora and Mooers 15-minute quadrangles and the New York part of the Plattsburgh and Rouses Point 15-minute quadrangles (Denny, 1967, 1970). A reconnaissance study was also carried on in the remainder of Clinton County and adjacent parts of Franklin County (pl. 1).

In company with the late A. Williams Postel of the U.S. Geological Survey, the lithology of stones in the till was analyzed in the field. Rapid mechanical analyses of sandy sediments were also made in the field. No samples of drift were studied in the laboratory.

No attempt is made here to summarize the previous geologic work in the region. The bedrock has been studied by Postel and his associates (Postel, 1952; Postel, Dodson, and Carswell, 1956; Postel, Wiesnet and Nelson, 1956; Nelson and others, 1956) and by Fisher (1968), Buddington (1937, 1953), and Miller (1919). The area is covered by the Adirondack Sheet of the Geologic Map of New York (Fisher and others, 1962). The chief contributors to the knowledge of the Quaternary geology of the region are Woodworth (1905a, b), Fairchild (1919), Chapman (1937), and MacClintock and Stewart (1965). MacClintock and Stewart present summaries of previous work. The Vermont Geological Survey has recently issued "The Surficial Geologic Map of Vermont," scale 1:250,000, prepared by Stewart and MacClintock (1970; see also Stewart and MacClintock, 1969).

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BEDROCK GEOLOGY AND TOPOGRAPHY

Uplands and mountains of Precambrian crystalline and metamorphic rocks are bordered by foothills of Paleozoic sedimentary rock that descend northward to the St. Lawrence Lowlands (Bostock, 1970) and eastward to the Champlain Valley. The northeast prong of the upland forms the St. Lawrence-Champlain divide and extends across the International Boundary into southern Quebec Province.

The Precambrian rocks form broad valleys and dome-shaped hills and low mountains commonly not more than 1,000 feet high, except for a few isolated peaks such as Lyon Mountain and Mount Whiteface. Granite gneiss and anorthosite each constitute about 30 percent of the area of uplands and mountains. These rocks are generally medium to very coarse grained and massive, with widely spaced joints. The rest of the area is underlain by other gneisses, metagabbro, and metasedimentary rocks (pl. 2; see also Fisher and others, 1962; Broughton and others, 1966). Sheeting is a common weathering feature of the gneissic rocks, and exfoliation domes are prominent topographic forms, as, for example, Silver Lake Mountains about 10 miles northwest of Ausable Forks (pl. 2). The Precambrian rocks tend to weather into large clasts, commonly of boulder size, and these in turn break up into fine gravel and sand.

A foothills belt, which is a few to as much as 20 miles wide, borders the uplands and mountains on the northeast and north. It includes both isolated hills and dissected plateaus underlain by the Potsdam Sandstone (pl. 3), a quartz sandstone or arkose that is locally conglomeratic and that includes beds of shale. The Potsdam Sandstone forms beds a few inches to a few feet thick. Crossbedding and ripple marks are common. Dips are low, commonly less than 5°.

Till derived from Potsdam Sandstone contains many clasts of pebble and cobble size, whereas that from anorthosite and granite gneiss contains a high proportion of sand-sized particles.

The lowlands adjacent to the St. Lawrence River and Lake Champlain are formed on sedimentary rocks, largely of Ordovician age (pl. 3), commonly calcareous, and generally having a larger proportion of thin-bedded units than is found in much of the Potsdam Sandstone. In the area near Plattsburgh, dolostone, limestone, and dolomitic quartz sandstone are common (Fisher, 1968). Along English River in Quebec Province, and in some areas near Lake Champlain, the Potsdam Sandstone extends into the lowlands.

DEPOSITS AND LANDFORMS ASSOCIATED WITH THE ICE SHEET

GLACIAL DRIFT

In the mountains the glacial deposits are, in large part, pebbly and sandy till that covers valley floors and lower mountain slopes. Many valleys also contain thick masses of sand and lesser amounts of gravel. Outcrops of bedrock are abundant along some of the larger streams and on upper slopes and mountain summits. Elsewhere, bedrock outcrops are scarce. (See Postel, 1952, pl. 1; Postel, Dodson, and Carswell, 1956; Denny, 1967, 1970.) On lower mountain slopes the drift may be more than 100 feet thick. W. A. Hobba, Jr. (written commun., 1967) reported that a water well on the upland about a mile south of Dannemora penetrated 130 feet of drift without reaching bedrock. No samples of drift were studied in the laboratory.

The composition and texture of the till reflect the kind of bedrock. In areas underlain by crystalline rocks of Precambrian age (pl. 3), the till is commonly a sandy loam or loamy sand containing pebbles and boulders of crystalline rocks and of Potsdam Sandstone. In most exposures clasts constitute only a few percent by volume of the deposit. The scarcity of clasts may be the result of two factors: First, the Precambrian crystalline rocks weather to boulders and then to sand without the formation of abundant pebble-sized clasts; second, the valleys drain toward the ice edge and were dammed by it, so that the glacier overrode and incorporated sandy alluvial deposits, either proglacial outwash or older alluvium. For example, in the valley of Alder Brook, a northeast-flowing tributary of the Saranac River, the till is a massive loamy sand containing only a few pebbles and boulders (less than 1 percent by volume).

In the foothills underlain by Potsdam Sandstone, the till is stonier and has a finer grained matrix (fig. 1) than the till in the mountains. Pebbles and boulders of Potsdam Sandstone may compose perhaps a fifth of the volume of the drift; the matrix is a sandy loam or loam.

In the lowlands near Lake Champlain and in the St. Lawrence Valley, where the bedrock is chiefly dolostone, limestone, sandstone, and shale, largely of Ordovician age (pl. 3), the till is a pebbly loam to silt loam, has an alkaline reaction, and contains clasts of sedimentary rock that make up 5–10 percent of the total volume.

FIGURE 1.—Bouldery till at east end of the upland between the Saranac and Salmon Rivers. Boulders are chiefly of Potsdam Sandstone. Exposure in borrow pit about 2.5 miles north of Peasleeville.

The water-laid drift, largely sand, also reflects the nature of the adjacent till and bedrock. The deposits laid down in association with ice form kames, outwash plains, and ice-channel fillings that are described in the sections on the various episodes of deglaciation.

TILL LITHOLOGY

The late A. Williams Postel identified in the field the stones or clasts in samples of till collected by him and the author (Denny and Postel, 1964) and placed each clast in one of several lithologic groups. The lower size limit of the stones collected was about 0.5 inches. Isopleths showing the percentage of clasts of Paleozoic sedimentary rock, granite gneiss, and anorthosite in the till (pls. 2, 3) trend roughly from northwest to southeast at right angles to the general direction of ice movement shown by directional features. Clasts of Paleozoic sedimentary rock, largely, and in many samples exclusively, Potsdam Sandstone, are persistent. In most samples of till from the area underlain by granite gneiss, clasts of Potsdam Sandstone are more abundant than those of granite gneiss; clasts of Potsdam Sandstone appear to be more resistant to wear during glacial transport than those of the granite gneiss.

It should be emphasized that the abundance (the number) of clasts in the till is not necessarily a true measure of the lithology of the till matrix. The clasts may constitute only a few percent of the total volume of till at a sample locality. Thus, the till as a whole might be 90 percent local material on a volume basis, whereas 70 or 80 percent by number of the clasts might be from a distant source.

Some of the pebble counts appear anomalous. In a sample of till from the southwest corner of the map area (pl. 3), 50 percent of the clasts are sandstone. This sample comes from a broad lowland north of the high peaks where the drift cover is extensive. Perhaps beneath the drift there are small outliers of Potsdam Sandstone (Kemp, 1921, p. 65).

On a high plateau of Potsdam Sandstone north of the mountains near Ellenburg (pls. 2, 3), the drift contains many clasts of granite gneiss and other crystalline rocks. The presence of these clasts 2–4 miles north of the nearest outcrop of crystalline rocks suggests either an inlier of Precambrian rock north of Ellenburg, now buried by drift, or the northward movement of the clasts by glacier ice, mass movements, or running water.

The till exposed in a roadcut about 2 miles west of Ellenburg contains about 25 percent clasts of Precambrian crystalline rocks (pl. 2) ; the rest are clasts of Paleozoic sedimentary rocks. Twenty-one percent of the clasts of Precambrian rocks are of the Lyon Mountain Granite Gneiss, and 2 percent are Hawkeye Granite Gneiss (Postel, 1952). A 16-inch boulder of the Hawkeye Granite Gneiss (not in the count) lay on the roadbank, and in the adjacent pasture are many boulders of Precambrian crystalline rock more than 4 feet in diameter and one about 6 feet in diameter. The till in an exposure about 5 miles south-southeast of Ellenburg and north of the Precambrian-Paleozoic rock contact contains about 29 percent clasts of Precambrian crystalline rocks including 12 percent Lyon Mountain Gneiss and 12 percent Hawkeye Granite Gneiss. The clasts of Hawkeye Granite Gneiss at the locality south of Ellenburg could have come from outcrops of this granite gneiss on the crest of Ellenburg Mountain (fig. 2), about 3–4 miles to the south and about 600 feet above the elevation of the sample site. The nearest source of Lyon Mountain Granite Gneiss is only about 1.5 miles south of the sample site. At the roadcut about 2 miles west of Ellenburg, the clasts of Hawkeye Granite Gneiss are about 7 miles north of their nearest source area on Ellenburg Mountain, and those of Lyon Mountain Granite Gneiss are about 4 miles from their nearest source to the southwest.

At both localities, the drift that contains the clasts of Precambrian crystalline rocks is thin and rests on Potsdam Sandstone. The clasts do not have conspicuous weathering rinds or shells. Some are stained throughout; others are essentially fresh. It is probable that the drift at the two localities was deposited by the last ice sheet, and that this ice picked up these clasts from a nearby source and moved them southward a short distance. Prior to this last movement, the clasts may have been carried northward by glacial ice from the mountains to the south (Cushing, 1899, p. 8). This transport could have taken place during the onset of the last glaciation, or it could have taken place during an earlier Wisconsin or pre-Wisconsin glacial age. The inferred northward movement of these clasts could also have been by mass movement or by catastrophic flood.

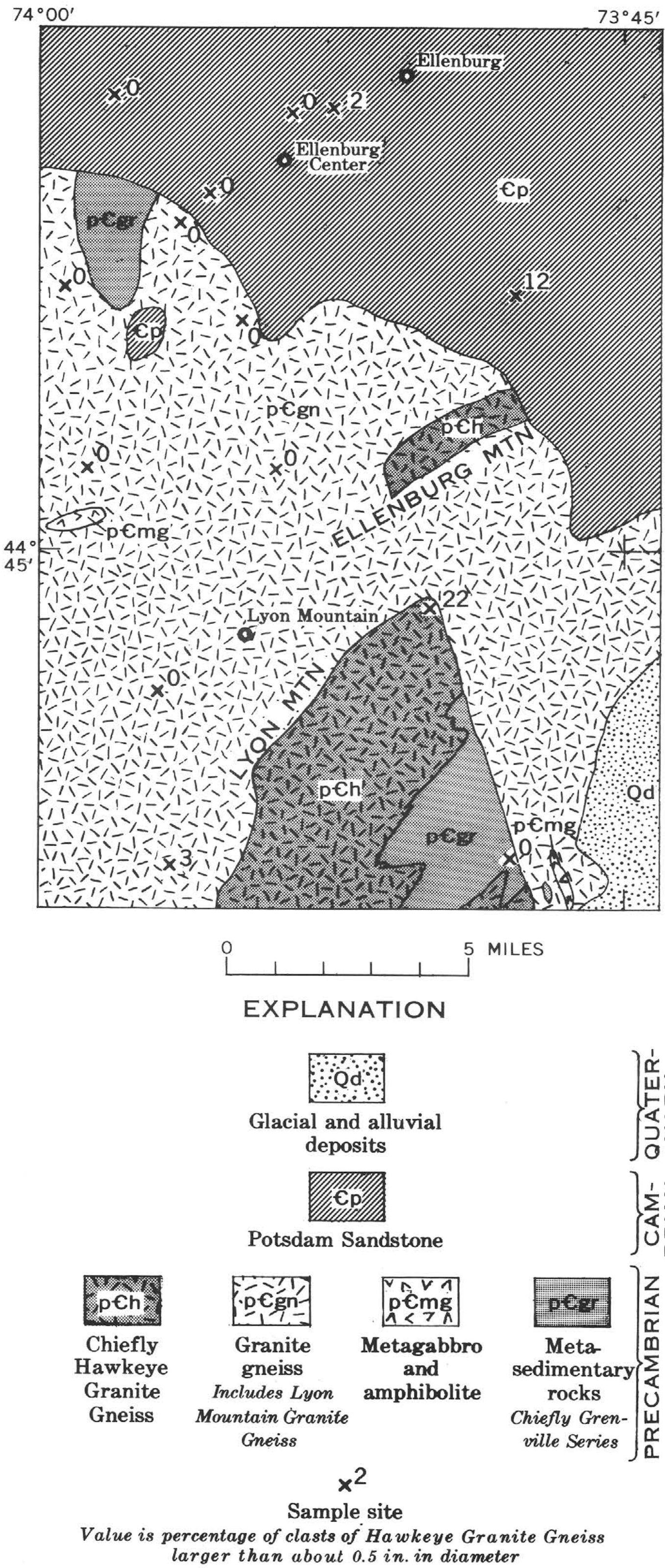
DIRECTION OF ICE MOVEMENT

Striae, drumlins, grooved drift, and till lithology indicate ice movement southwest across the northeast prong of the Adirondack Mountains (pls. 2, 3). In the lowlands south of the St. Lawrence River west of Clinton County, ice movement was south and southeast into the mountains, whereas in the lowlands west of Lake Champlain, ice movement was south up the Champlain Valley. The consistent trend of all the directional features suggests that they record ice movement during deglaciation when the ice edge was nearby. The few recessional moraines that can be traced for several miles—near Ellenburg Depot, north of Clinton Mills, and south of Miner Lake (pl. 1)—trend roughly at right angles to the direction of ice movement inferred from striae or grooved drift. The preservation of grooves in drift implies that the ice was not loaded with debris sufficient to conceal these streamlined forms when the ice had disappeared.

The lithology of the stones in the till is consistent with the direction of ice movement recorded by the directional features, except for the samples from the two small areas near Ellenburg. In general, the percentage of clasts of Paleozoic sedimentary rock (pl. 3) decreases southwestward and also as the altitude of the sample site increases (compare Lyon, Terry, and Whiteface Mountains). The trend of the isopleths suggests movement of ice southwest up the Saranac River valley. The pebble counts do not suggest southward ice movement across the Lyon Mountain–Johnson Mountain highland north of the river, nor is there evidence in the counts for a northward ice movement from the high peaks region (Craft, 1969) south of the area shown on plate 2. Isopleths for anorthosite are consistent with a southerly ice movement.

Small drumlins are abundant in the northeast corner of Clinton County near Champlain (Denny, 1970), where the bedrock is chiefly dolostone, limestone, sandstone, and shale, and where the till forming the streamlined features is a pebbly silt loam or loam. The azimuth of the long axis of 50 of the drumlins near Champlain ranges from N. 9° W. to N. 14° E. The estimated mean value is N. 6° E.

Glacial grooved or fluted drift consisting of long narrow ridges and swales is found in small areas near Miner Lake and on the crest of the ridge between Saranac and Peasleeville, where the bedrock is Potsdam Sandstone. Striae adjacent to grooved drift trend in approximately the same southwesterly direction. The individual ridge does not have a bedrock knob at its northeast (up-ice) end as do the ridges described by McDonald from the Appalachian region of southeastern Quebec Province (McDonald, 1966, 1967).



MORAINES IN THE SARANAC VALLEY

At Cadyville on the Saranac River, a belt of recessional moraine consisting of linear ridges and knolls of till and kames of sand and gravel crosses the valley from north to south (pl. 1, from ice-front position 4b east to West Beekmantown). The belt ranges in width from about 1 mile to 2.5 miles. East of Cadyville, the river descends 300 feet in a rock-walled gorge. The wide belt of moraine crosses the valley on the tread of this giant bedrock step and was perhaps localized in part by it. The moraine was built by a small southwest-moving ice tongue from a large ice mass in the Champlain Valley. The moraine is a part of Taylor's DeKalb moraine (Taylor, 1924, p. 665).

The linear till ridges (pl. 4) range in height from a few feet to about 60 feet above the adjacent swales and in length from several hundred feet to a quarter of a mile. The ridges are curving or sinuous in plan and are spaced commonly 200–500 feet apart from crest to crest. The crests are rounded and boulder covered. The swales between the ridges are 200–1,000 feet wide and are floored with sand and a few boulders. These linear features are neither as regular nor as continuous as the 20-foot contours of the map indicate.

The ridges are composed of pebbly, sandy till that includes small masses of stratified sand and gravel. The stones in the till ridges are largely Paleozoic sedimentary rock; pebbles of Precambrian rock generally make up less than 5 percent (pl. 3) of the total clasts. Bedrock outcrops are scarce in the area of the moraine.

The linear ridges in the moraine south of the river form arcs bowed out in an easterly direction, whereas the moraine as a whole is bowed in the opposite direction. The explanation for these opposing trends may involve stagnation and minor oscillations of the ice near the front. The ice tongue first advanced to its southwest limit about a mile west of Elsinore and stagnated. Then active ice east of Elsinore moved southwest diagonally up the valley wall south of the dead ice mass to form a narrow tongue of ice about a mile wide in the area between Duquette Road and the areas of bare rock south of Hardscrabble Road. The arcuate ridges bowed out to the east are inter-

FIGURE 2.—Percentage of clasts of Hawkeye Granite Gneiss in till in the Ellenburg-Lyon Mountain area, Churubusco and Lyon Mountain 15-minute quadrangles. Geology generalized from Fisher and others (1962). Percentage of clasts estimated by rapid method described by Denny and Postel (1964).

preted to be lateral moraines on the northwest side of this small tongue of ice.

North of the Saranac River the moraine includes prominent linear till ridges, abandoned stream channels eroded in drift, and bodies of sand and gravel believed to be ice-marginal kames.

Most of the sand and gravel in the moraine probably came from the ice sheet, although some of the material may have come from the erosion of channels in the drift. Deposition was largely by melt-water streams flowing southwest and south along the ice margin. In addition, water ponded north of the Saranac Valley-Smith Wood Brook divide may have spilled south across the divide and down the north slope of the Saranac Valley where it eroded the drift-walled channels, now largely abandoned. As the edge of the ice sheet retreated northeast, the place where the water spilled south also moved east. Thus, the abandoned channels on the north side of the Saranac Valley are progressively younger northeastward.

The linear ridges in the moraine suggest radially inward shrinkage of the ice tongue in the valley. Water, ponded by ice, overflowed along the southern margin of the ice tongue around the eastern end of the hill between the Saranac and Salmon River (Clinton County) valleys (pl. 1) and eroded channels both in drift and in bedrock; one of the latter is cut about 60 feet into nearly horizontal beds of the Potsdam Sandstone. The water also washed away most of the drift cover from bedrock areas as large as 0.5 mile in diameter. As the edge of the ice tongue retreated to the northeast, channels at lower and lower altitudes were utilized by overflow from the Saranac Valley. The materials eroded by the water flowing along the south side of the ice tongue were deposited just southeast of the areas of bare bedrock as bouldery glaciofluvial deposits on the north slope of the Salmon River valley.

The eastern edge of the belt of recessional moraine (pl. 4) is marked by a low ridge that extends south from Cadyville for about 0.5 mile and north of the river to a sand and gravel kame along State Route 374. Farther north, another ridge follows the same course for a short distance before turning northeast. An exposure in the ridge on the north side of State Route 3 showed till resting on sand, suggesting a minor oscillation of the ice front. The kame north of Cadyville is probably an ice-marginal deposit from south-flowing water in a channel that crossed the site of the present Saranac River and continued south, entering Lake Vermont, where it built a small delta (Qs, pl. 4), about 1.5 miles south of Woods Mills.

The broad open valley of the Saranac River between Cadyville and Saranac contains no traces of an ice-dammed lake, yet the moraine near Cadyville demonstrates that an ice tongue dammed the valley. Either the moraine-dammed lake was too short lived to leave conspicuous evidence of its presence, or the broad valley upstream from the moraine was filled with stagnant ice. It is not known whether the moraine near Cadyville marks a stationary ice-front or a readvance of the ice up the Saranac Valley from a position in the Champlain Valley.

MORAINES IN THE GREAT CHAZY VALLEY

An 8-mile-long belt of recessional moraine crosses the North Branch Great Chazy River near Ellenburg Depot (ice-front position 8, pl. 1), and shorter moraine segments occur to the northwest near the Canadian border north of Clinton Mills and to the southeast in the Great Chazy River valley south of Miner Lake. If these three segments are essentially contemporaneous, as seems probable, they show that the ice front crossed the valley of the Great Chazy River and the North Branch with only a slight bulge upvalley to the southwest.

The morainal belt attains its greatest dimensions at Ellenburg Depot where it is bisected by the North Branch Great Chazy River (pl. 5; Taylor 1924, fig. 12; MacClintock and Stewart, 1965, p. 58 and pl. 1B) South of the river, a north-trending ridge ranging from about 80 to 100 feet in height and from about 1,000 to 1,600 feet in width is composed chiefly of sand and gravel. West of the ridge, the valley walls are smooth or gently rolling, but to the east they have a pronounced local relief of knolls and ridges.

In exposures in a borrow pit near the river (point *A*, pl. 5), deformed beds of gravel and masses of till are separated by essentially vertical faults from southwest-dipping beds of coarse sand and pebble gravel that grade into horizontally bedded sand near the west side of the ridge (fig. 3). These beds probably are part of a delta built by melt-water streams from the ice sheet on the east side of a small ice-marginal lake that was west of the moraine and was dammed by it.

In a second borrow pit (at point *B*, pl. 5), about 4 feet of till overlies 10–15 feet of thin-bedded medium sand. At the base of the exposure is a second till, largely a structureless mixture of sand and silt containing a few striated stones. Between the lower till and the bedded sand is a 10-foot horizon of mixed till and sand consisting of tilted and faulted beds of sand and masses of till. In places, the till masses are

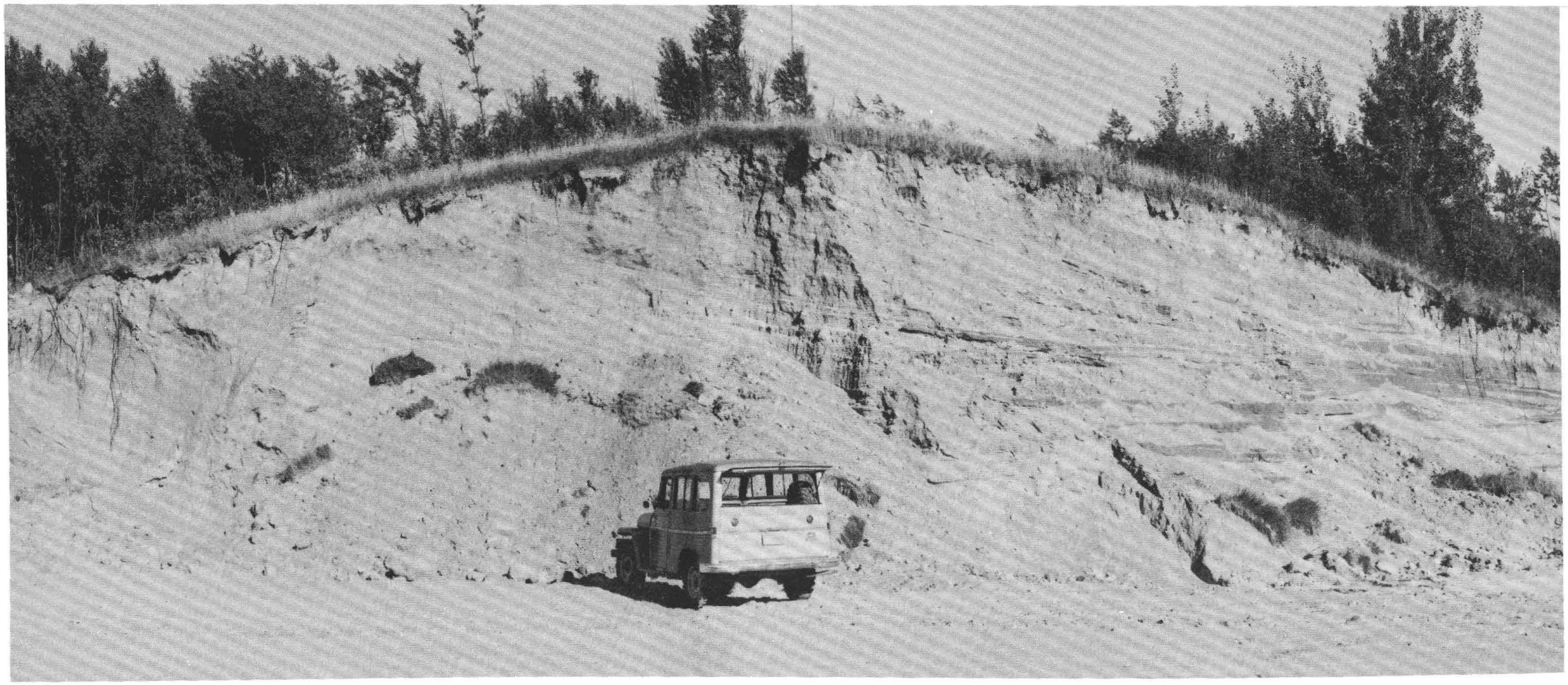


Figure 3.—Cross section through a part of the moraine near Ellenburg Depot. Deformed beds of gravel and masses of till at north (left) end of section separated by high-angle faults from west-dipping beds of coarse sand and pebble gravel that grade into horizontally bedded sand at south (right) end of section. Materials are part of a small ice-marginal delta. Exposure in borrow pit near Ellenburg Depot, point *A*, pl. 5.

vertical plates 2–3 feet thick and at least 10 feet long that lie against beds of sand cut by many closely spaced high-angle faults. It appears that the ice edge oscillated back and forth across the floor of the ice-dammed lake, depositing a relatively stone-free till, deforming its own delta deposits, and laying till on top of them.

At its north and south ends, the moraine near Ellenburg Depot rises to an altitude of about 1,100 feet or about 100 feet above the crest of the moraine along the river, suggesting that when the ice-dammed lake overflowed, it was at least 200 feet deep. If so, the sands near the top of the moraine at the river may have been deposited at a depth of more than 100 feet below the surface of the lake. There is no other evidence for an ice-dammed lake in the headwaters of the North Branch beyond that presented by the moraine itself.

South of Dannemora Crossing (pl. 5) the moraine ascends the south slope of the valley of the North Branch. In sand pits along Plank Road, deformed delta deposits of sand and gravel pass upward into lake deposits of thin-bedded fine sand containing a few boulders. West of the pits, the slope of the ridge is covered with boulders. Where the moraine crosses Old Military Turnpike (fig. 4), it is a small ridge only 6 feet high and about 80 feet wide.

North of Ellenburg Depot, the moraine is a broad low ridge of till 5–20 feet high and 500–1,000 feet wide. Near its northern end, it becomes a narrow steep-sided boulder-covered ridge as much as 50 feet high and 400 feet wide (pl. 6).

The short segment of recessional moraine north of Clinton Mills extends east-west across the uplands

FIGURE 4.—Along the Old Military Turnpike the moraine near Ellenburg Depot is a boulder-covered ridge about 6 feet high and 80 feet wide. Photograph is looking north across the Turnpike (pl. 5).

for about 1.5 miles at altitudes ranging from 1,050 to 1,100 feet (pl. 1). It is a low sinuous ridge (fig. 5), 5–20 feet high and about 100–300 feet wide. Side slopes are steep, the surface is bouldery, and the material appears to be till. West of the well-defined segment of moraine for a distance of about 3 miles are other small east-trending linear ridges (not shown in fig. 5) that may also be recessional moraines, perhaps modified by wave action. MacClintock and Stewart (1965, pl. 1B) mapped some of these as beach and nearshore deposits.

The other small segment of recessional moraine south of Miner Lake is a series of long narrow ridges on the south slope of the valley of the Great Chazy River (pl. 1). It extends from a point near the Old Military Turnpike (fig. 6), west for about 4.5 miles to Alder Bend Road (Woodworth, 1905a, p. 11).

The individual ridges range in length from about 300 to 4,000 feet, in width from about 150 to 600 feet, and in height from about 2 to 30 feet. The ridges are boulder covered; they are probably composed of sandy till. Just east and west of the Great Chazy River are sand and gravel kames.

SPILLWAYS ON THE ST. LAWRENCE-CHAMPLAIN DIVIDE

The plateau on the Potsdam Sandstone northeast of the mountains (pls. 1, 3) forms the divide between the St. Lawrence and Champlain Lowlands and extends as far as Covey Hill, Quebec Province. North and east of Covey Hill, the surface drops steeply to the floor of the St. Lawrence Lowlands, the descent being nearly 800 feet in about 2 miles. Abandoned stream channels or cols, which served as spillways for ice-dammed lakes, cross the divide (pl. 1).

The abandoned stream channels south of Churubusco in northwestern Clinton County, at altitudes ranging from about 1,200 to 1,300 feet, are generally not more than 100 feet wide and 10–20 feet deep; one is about 300 feet wide. (See Churubusco 7½-minute quadrangle.) Their east ends "hang," and their floors slope west, indicating that they carried water from east to west, perhaps into ice-marginal streams that cut some of the abandoned stream channels west of Churubusco and south of Chateaugay (pl. 1; see also MacClintock and Stewart, 1965, fig. 21). About 1.5 miles north of Clinton Mills (3 miles east of Churubusco) is a large channel about 500 feet wide and 10 feet deep at an altitude of about 1,085 feet (fig. 5). This channel hangs at both ends and could have carried water either east or west. No bare-rock areas are nearby.

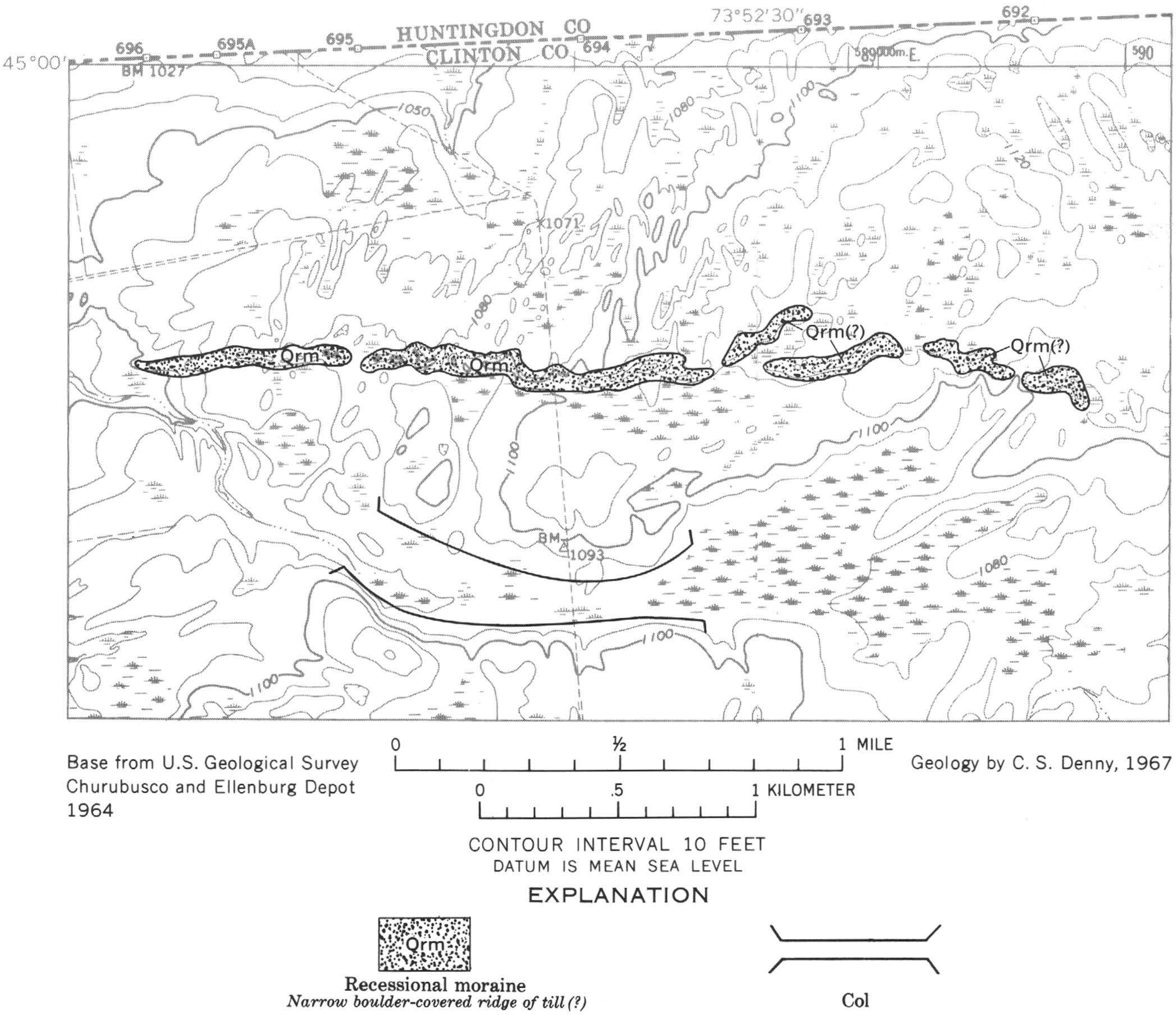


FIGURE 5.—Recessional moraine north of Clinton Mills and abandoned stream channel or col, altitude about 1,085 feet, on the St. Lawrence Valley–Champlain Valley divide.

The abandoned stream channel west of Covey Hill, Quebec, just north of the Canadian border (pl. 1; see also Chateaugay sheet, Quebec-New York, west half [31 H/4 WEST] scale 1:50,000, published by the Canadian Dept. Mines and Tech. Surveys) was first described more than 100 years ago (Emmons, 1842; Woodworth, 1905a, Fairchild, 1912; and Goldthwait, 1913. The following description is based largely on MacClintock and Terasmae, 1960). The channel is at an altitude of about 1,010 feet and is about 90 feet below the high point on the upland about 2 miles to the northeast. It is about half a mile wide and about a mile long; the floor of this channel is bedrock, which in part is mantled by a 6-foot layer of peat. At the east end of the channel is a box canyon, 75 to 100 feet deep and about a mile long, cut in the Potsdam Sandstone. An abandoned waterfall at the head of the canyon is more than 125 feet high; about 75 feet of this precipice is now beneath the surface of a small pond. Half a mile from the dry fall is a second stagnant pool, The Gulf (pl. 6). The channel near Covey Hill carried the outflow of glacial Lake Iroquois (Coleman, 1937) into the Champlain basin, supplying much of the water that washed the drift from large areas of bedrock.

AREAS OF BARE ROCK

At the northeast end of the Adirondack uplands in northern Clinton County are large areas of bare rock (Potsdam Sandstone). The largest, Flat Rock, near Altona (Woodworth, 1905a, p. 16–24), is about 5 miles long, 2.5 miles wide (pl. 1), and presents a striking appearance on aerial photographs.

The "flat rocks" are either bare rock or bedrock that supports an open forest with a luxurious undergrowth of low shrubs which grow in a thick mat of organic matter resting directly on bedrock (fig. 7). There is little inorganic mineral matter in the soil. White pine, pitch pine, red oak, and other species common to the northern hardwood forest compose the open forest. Jack pine is also abundant, although it is rarely found growing in the adjacent drift-covered areas.

Ripple marks and curving joints that are the outcropping edges of foreset beds are a characteristic minor surface feature. The bedrock surface is slightly weathered, and solution pits are common (fig. 8). At only one locality was the surface polished, presumably by glacier ice.

The large areas of abundant outcrop have little local relief beyond that of risers a few inches to 10 feet high between the essentially horizontal bedding planes (fig. 9). The flat surfaces commonly extend for hundreds of feet between risers. The sandstone may be broken by joints as much as 1-foot wide and several feet deep. In some places steep-sided bedrock masses rise 10–20 feet above the surrounding flat. About 1 mile northwest of Cobblestone Hill (pl. 7), a steep-sided ridge more than 50 feet high is surrounded by a flat bare-rock surface on three sides. The ridge has a closely packed mantle of angular sandstone blocks ranging from 1 to 15 feet in diameter; 2–6 feet is the most common range in size. The mantle of angular blocks extends down to a sharp contact with the bare rock. Bedrock crops out on the crest of the ridge near its north end.

The large areas of bare rock are on divides between the larger streams (pl. 1) and are restricted

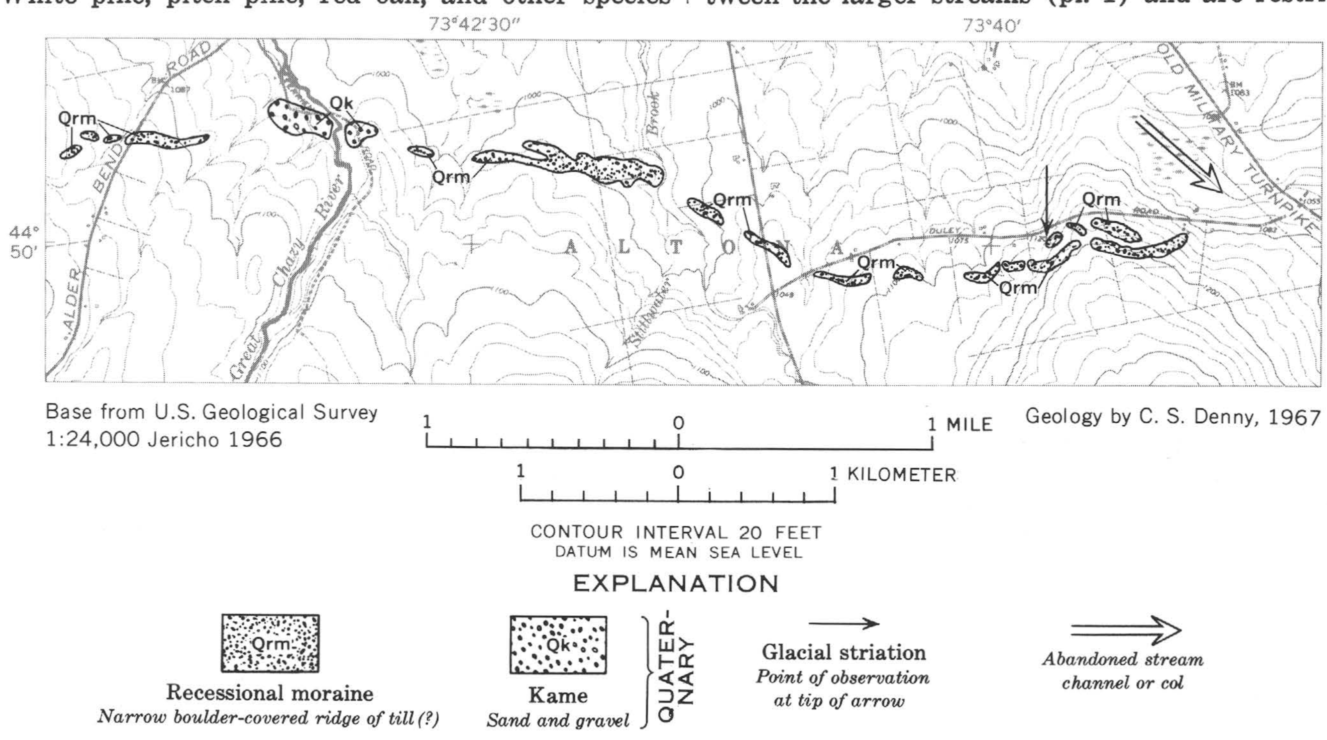


Figure 6.—Moraine south of Miner Lake.

FIGURE 7.—Flat Rock, near Altona. Potsdam Sandstone with patches of low shrubs and scattered trees. View looking west just east of Miner Lake.

largely to a belt about 3 miles wide, extending from the Canadian border near Cannon Corners southeast nearly as far as West Chazy, a distance of about 20 miles. The belt ranges in altitude from about 1,000 feet to about 600 feet.

FLAT ROCK, NEAR ALTONA

Flat Rock, southeast of Altona (pl. 1), is about 5 miles long and 2.5 miles wide. The bedrock surface slopes to the north and east from an altitude of more than 1,000 feet down to an altitude of about 600 feet. There are several narrow rock-walled gorges in the southeastern part of the bare-rock area, such as Bear Hollow (pl. 7). The bare-rock area is drained by Cold Brook, which has its source in the Dead Sea, a small rock-walled pond shown in the northwest corner of plate 7. The Dead Sea is in a former plunge pool, part of an abandoned stream channel (Woodworth, 1905a, pls. 7, 8, 9, and 10). The border of Flat Rock is generally sharp but highly irregular. The surface of bare rock passes under the surrounding drift or other surficial cover without a prominent topographic break.

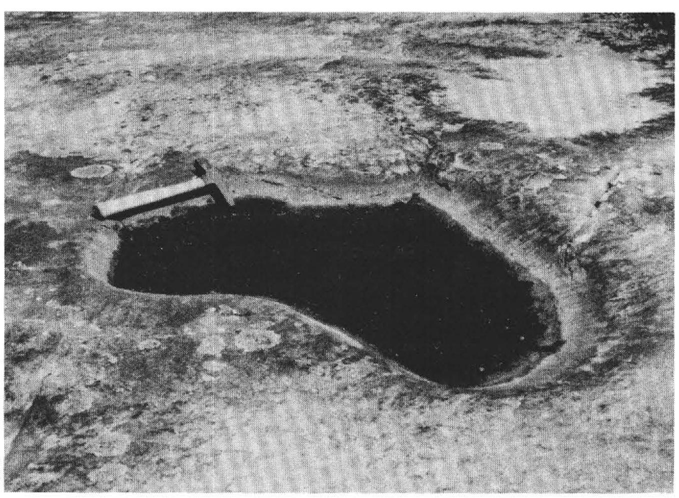


FIGURE 8.—Solution pit in Potsdam Sandstone in bare-rock area south of English River.

Flat Rock is ringed on the north and east by segments of the beaches of glacial Lake Vermont (Chapman, 1937). The highest stand of Lake Vermont in this area, the Fort Ann stage, overlapped the existing outer limit of the area of bare rock by as much as 0.5 mile and stood almost 100 feet above the lowest elevation of bare rock (pl. 7).

Bouldery material at the edge of Flat Rock was presumably in large part washed off the bare-rock area and redeposited by ice-marginal streams along the border. The total volume of such material is small. The gravel that forms the well-developed bouldery beach ridges on the east slope of Cobblestone Hill could have been reworked from material washed from the bare-rock area to the northwest. South of Cobblestone Hill along the east edge of the bare-rock area (pl. 7) are two small deposits of bouldery material. Along the stream that crosses the southern deposit, 10–20 feet of boulder gravel is exposed. Bedrock crops out nearby, and the total volume of gravel is small. A borrow pit near Flat Rock (fig. 10) excavated in boulder gravel shows many clasts ranging from 1 to 5 feet in longest diameter.

Southeast of Cobblestone Hill are bouldery deposits that may be recessional moraines composed of material washed from the bare-rock area, deposited in Lake Vermont at the edge of the ice, and subsequently molded into moraine during a slight advance of the ice edge. Woodworth made the suggestion that the abundance of gravel beaches and the scarcity of bedrock outcrops in the valley of the Little Chazy River west and southwest of West Chazy might be due to the presence of "unusually thick deposits [that] are probably to be attributed primarily to the wash from the flat rock districts" (Woodworth, 1905, p. 20). However, even assuming that all the materials described above were washed from Flat Rock, the total volume of material is so small that one must conclude that Flat Rock probably never had a thick drift cover.

SOUTH OF ENGLISH RIVER

The area of bare rock on the divide between English River and North Branch Great Chazy River ("Blackman's Rock," Woodworth, 1905a, pl. 5) is about 3 miles long and about 1 mile wide at its widest point (pl. 6; fig. 11). The bare-rock surface slopes northeast and is bordered by recessional moraines trending at right angles to the divide. The moraines record the presence of a southeast-trending ice edge that dammed the English River, causing it to overflow into the valley of the North Branch. Such overflow progressively removed the drift from the slopes of the divide as the ice front retreated to the northeast. Figure 11 shows the stages of the northeasterly retreat of the ice front across the bare-rock area south of English River (ice-front positions 10–14).

Material washed from the higher or southwestern part of this area of bare rock was deposited as gravel and sand adjacent to the recessional moraines shown near the southern edge of plate 6. Material washed from the lower or northeast end was deposited in Lake Vermont where it was reworked by lake water into beaches and spits.

A mass of gravel and sand about 0.8 mile south of Cannon Corners (pl. 6) extends from the bare-rock area east of the highest beaches of the glacial lake. The surface of the deposit is mantled with boulders as much as 6 feet in diameter and ranges in altitude from about 790 feet near the flat rocks down to about 710 feet along the highest stand of the glacial lake. The boulders could have been concentrated on the surface by the washing of underlying till, but the

FIGURE 9.—Vertical joint face between two bedding-plane surfaces on the Potsdam Sandstone south of English River (loc. *A*, pl. 6). Block of sandstone at base of joint face at left. Note shovel for scale.

FIGURE 10.—Boulder gravel, probably debris washed from an area of bare rock. Exposure in borrow pit about 1 mile south of Altona.

presence of shallow ice-block holes (not shown in pl. 6; see Denny, 1970) suggests that much of the surface is underlain by water-laid material deposited in association with wasting glacial ice.

At one place near the southeast edge of the bare-rock area south of English River (loc. A, pl. 6), the surface of the riser and adjacent tread is highly polished but not striated. Presumably, this is glacial polish that elsewhere has been removed by weathering. The polished surface extends for perhaps 50 feet along the edge of the bare rock. To either side of the polished surface, the riser and the tread are essentially smooth or slightly pitted.

There are no sandstone blocks at the base of the polished riser, but one or two blocks lie on the adjacent flat at the base of the riser to either side. None of the surfaces of these blocks is polished; all appear slightly weathered.

NORTH OF ENGLISH RIVER

The area of bare rock ("Stafford's Rock," Woodworth, 1905a, pl. 5) north of English River is about 3 miles long and 1 mile wide (pl. 6) and is much like that south of the river. At the north end, however, are several bedrock knolls that range from about 200 to 1,000 feet in diameter and that rise 5–20 feet above the surrounding flat bedrock surface. These small steep-sided residual knobs probably were carved by streams that flowed east from the gorge near Covey Hill, Quebec (The Gulf), roughly along ice-front position 13 (fig. 11). The streams eroded

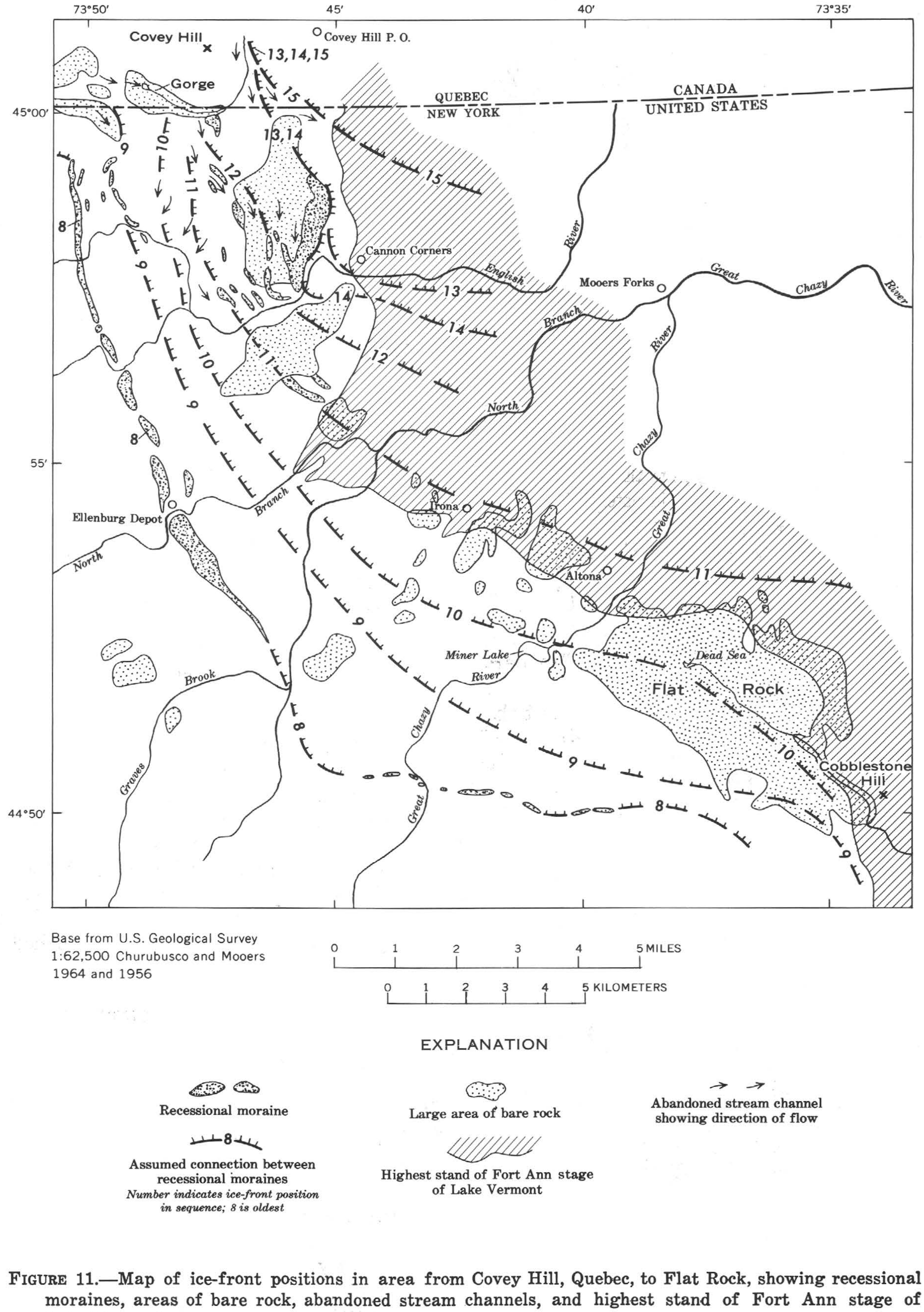


FIGURE 11.—Map of ice-front positions in area from Covey Hill, Quebec, to Flat Rock, showing recessional moraines, areas of bare rock, abandoned stream channels, and highest stand of Fort Ann stage of Lake Vermont.

a bedrock ridge (cuesta) to form the knolls and cleaned off the drift to expose the adjacent bedrock.

A prominent recessional moraine about 2 miles long and as much as 0.4 mile wide lies east of the bare-rock area (ice-front positions 13 and 14, fig. 11), and several small narrow morainal ridges lie on the western part of the bare-rock area (fig. 12). At the south end, just north of White Road, two moraines unite to form a prong whose surface is exceedingly bouldery; blocks 6–10 feet in diameter are common. The two moraines enclose a triangular-shaped swamp floored with organic material that appears to rest directly on bedrock (pl. 6).

FIGURE 12.—Boulder-covered recessional moraine, at right, and turf-covered pasture and pond on bedrock at left. View north from White Road at a point about 1.2 miles west of Cannon Corners (pl. 6).

ORIGIN

The presence of large areas of bare rock (Potsdam Sandstone) or rock thinly veneered with soil and vegetation indicates that in some areas, conditions at or near the ice edge were such that drift was not deposited, or once deposited, has since been removed. Because most of the large areas of bare rock are found in a limited region, whereas the Potsdam Sandstone is a widespread formation, it is clear that the bare surfaces do not owe their origin to some characteristic of the bedrock.

I believe that the bare-rock areas were swept clean by the outflow of glacial lakes held between the ice front and the northeast slope of the uplands. Water flowing in spillways along the ice edge removed the drift that overlay the bedrock. This is the explanation presented by Woodworth, following a suggestion by G. K. Gilbert (Woodworth, 1905a, p. 16–24), and my observations support their thesis.

Only a few small masses of bouldery material have been found adjacent to the bare-rock areas, and it is logical to assume that such areas probably never had a thick drift cover.

The cleaning of the drift cover from large areas of bedrock was accomplished by the sudden, perhaps catastrophic, draining of ice-dammed lakes. In the St. Lawrence Lowlands northwest of Covey Hill, glacial Lake Iroquois (Coleman, 1937) stood at an altitude of nearly 1,100 feet, its outlet near Rome, N.Y., southwest of the Adirondacks (fig. 14). A slight rise in lake level due either to uplift of the Rome outlet or to a slight northeasterly retreat of the ice front could have caused Lake Iroquois to overflow into the Champlain basin, supplying the ice-marginal streams with large volumes of sediment-free water.

At first, the outflow from Lake Iroquois may have been by way of cols or abandoned channels just south of the Canadian border (fig. 11). Later, the outflow came by way of the channel near Covey Hill, Quebec. By this means, the volume and perhaps the velocity of the water that flowed along the ice front were increased to such an extent that over large areas all the debris on the bedrock was removed. As the ice front retreated (ice-front positions 9 to 15, fig. 11), the ice-marginal streams—the spillways of Woodworth (1905a)—migrated north and east and cleaned off large areas of bare rock.

HISTORY OF DEGLACIATION

In the mountains, deglaciation involved stagnation of the marginal zone of the ice sheet, whereas in the St. Lawrence and Champlain Lowlands the ice sheet maintained an active front. Presumably this contrast in mode of deglaciation is related to the topographic contrast between the two areas. In the lowlands, as the marginal zone of the ice-sheet thinned, it remained active and built small moraines. In the mountains, the thinned marginal zone became separated from the actively moving ice. Melt water built kame terraces and outwash plains of debris, probably obtained from both stagnant and active ice bodies.

Drainage was toward the ice front, and ice-marginal lakes were formed. The Ausable, Salmon (Clinton County), and Saranac Rivers (pl. 1) flow northeast into the Champlain Valley, that is, in the direction of ice retreat. As deglaciation proceeded in a generally northeast direction, the headwaters of the rivers were uncovered first. The streams were ponded at the ice front. These ice-dammed lakes, fed by melt water and by runoff, overflowed across divides into adjacent valleys. Ultimately, the overflow reached glacial Lake Vermont, which itself lengthened northward as the ice edge retreated down the Champlain Valley.

When the Adirondack uplands were largely uncovered but the ice sheet still rested on the St. Lawrence-Champlain divide near the International Boundary, ice-dammed lakes were present on both sides of the divide. At first, the lake on the northwest side overflowed to the southeast. Then the flow was reversed and small ice-dammed lakes southeast of the divide flowed into the St. Lawrence Lowlands. Later, the glacial lake west of the divide, Lake Iroquois (Coleman, 1937), again overflowed to the east, cutting the gorge near Covey Hill, Quebec, cleaning off large areas of bedrock, and emptying ultimately into Lake Vermont. Finally the two lakes merged.

Only in the Champlain Valley are ice-marginal features continuous enough for one to draw, with reasonable certainty, the form of the ice front at any one time. Elsewhere, these features are widely separated, and their correlation one with another is based only on inference. As previously mentioned, the recessional moraines in the Great Chazy and Saranac valleys suggest that the ice sheet in the foothills and adjacent parts of the Champlain Valley had a straight to gently curved front, influenced in only a small way by the local topography. Narrow tongues of active ice apparently did not project far up the major east-draining valleys, at least not at the time when the edge of the ice lobe in the Champlain Valley lay along the east edge of the Adirondack uplands. If we assume that throughout deglaciation the ice edge was straight or gently curved, it is possible to reconstruct a logical picture of deglaciation of the northeast Adirondack region from the scattered patches of recessional moraine and other ice-marginal features that remain.

Plate 1 shows several inferred positions of the ice front beginning with number 1 in the southwest corner of the region and progressing in a general northeast direction toward Covey Hill, Quebec. The dating of these ice-front positions is discussed in the final section of this paper. The history of the various features shown on the map is summarized in table 1 and briefly described in the following pages.

LOON LAKE EPISODE (1)

When the edge of the southwest-moving ice sheet was at ice-front position 1 (pl. 1), melt-water streams flowed southwest, laying down thick masses of glaciofluvial deposits. This is the oldest event recognized in the region. The ice-contact slope on the northeast side of the recessional moraine is as much as 200 feet high. Drainage was to the west by way of the St. Regis River.

OWLS HEAD-REDFORD STAND (2)

In the St. Lawrence River drainage, Franklin County, the most prominent feature of the Owls Head-Redford stand is the south-sloping outwash fan at Owls Head (2a, Watertown moraine of Taylor, 1924, p. 660). The fan is apparently composed of sand and gravel as much as 200 feet thick and rests on a bedrock sill, north of which the Salmon River drops 300 feet in less than a mile. The ice front was on the north side of the fan and discharged melt water to the south. The water ultimately escaped west by way of the col between Humbug and Titusville Mountains. The Salmon River valley south of Owls Head probably still contained masses of stagnant ice.

Southeast of the St. Lawrence-Champlain divide, the ice sheet blocked the Saranac Valley near Redford (2b), and the river was diverted south into the headwaters of the Salmon River (Clinton County). The Salmon River valley also was blocked by ice, and the drainage went south across another low divide into a south-flowing tributary of the Ausable River. The position of the ice dam in the Salmon River valley is marked by a small recessional moraine consisting of steep-sided and boulder-covered knolls (fig. 13) that merge to the west with a small pitted outwash plain of sand and gravel, whose upper surface slopes to the south. The recessional moraines on the east slope of the uplands northwest of Clintonville are cut by the abandoned channels of streams that flowed south along the west side of a tongue of ice in the valley of the Little Ausable River (pl. 1), perhaps to join the Ausable River at the delta of the Coveville stage near Clintonville.

The diversion of the Saranac River near Redford, south to the Ausable River near Ausable Forks, is marked by several abandoned stream channels. The older channels, formed when the ice front was slightly southwest of position 2b, are on the Saranac River–Ausable River divide southeast of Redford. They range in altitude from about 1,200 to 1,230 feet and carried Saranac River water into the headwaters of Blake Brook (pl. 1) and thence south to the Ausable River. Northeasterly retreat of the ice edge to position 2b opened other channels at lower altitudes, 1,130–1,200 feet, that conveyed water from the Saranac Valley into headwaters of the Salmon River and thence south to the Ausable River valley. North of the village of Black Brook, a flat-floored valley 800 feet wide and 3–4 miles long ends near the village in a series of terraces that are composed of boulder and cobble gravel and that extend south for about 1.5 miles to the Ausable River. The terraces decline in altitude from about 1,060 feet to 900 feet, and the material becomes finer grained. Such extensive bouldery deposits are not common in this region; presumably they were deposited as an alluvial fan or ice-marginal kame on the north side of the Ausable River valley. The Ausable River has since incised its bed 100–200 feet. The delta west of Clintonville built by the Ausable River in the Coveville stage of glacial Lake Vermont has its top at an altitude of about 670 feet. It is possible but by no means certain that water coming down the stream channel north of Black Brook contributed directly to the building of this delta.

TROUT RIVER-MOFFITSVILLE STAND (3)

West of the St. Lawrence-Champlain divide, a recessional moraine and south-draining ice-marginal channels near the mouth of the Trout River valley mark the Trout River-Moffitsville stand (3a). To the east, the ice front is marked by recessional moraines in the valley surrounding Upper Chateaugay Lake.

In the Champlain basin, the ice-front (3b) is not well defined. Recessional moraines and other ice-marginal features are scarce, except for the small moraine, composed of till, sand and gravel, between Harkness and Keeseville.

The overflow from the Saranac Valley abandoned its southerly course to the Ausable River and turned east down the Salmon River valley as an ice-marginal stream, depositing gravel and sand and escaping around the east end of Terry Mountain. As the ice in the Salmon River valley gradually shrank, the late-glacial Saranac River cut down nearly 100 feet into the recessional moraine 4 miles southwest of Peasleeville and spread gravel on top of some of the sand deposited earlier. Near the moraine, the gravel-covered flood plain of the Salmon River slopes east at about 50 feet per mile and has conspicuous bars and swales. About 2 miles east of the moraine, the valley floor descends 100 feet in a distance of about a quarter of a mile. Farther east, the river meanders on a sandy and silty flood plain at a gradient of less than 20 feet per mile. The relations suggest that near the moraine the floor of the valley is essentially that of the flood plain of the late-glacial Saranac River

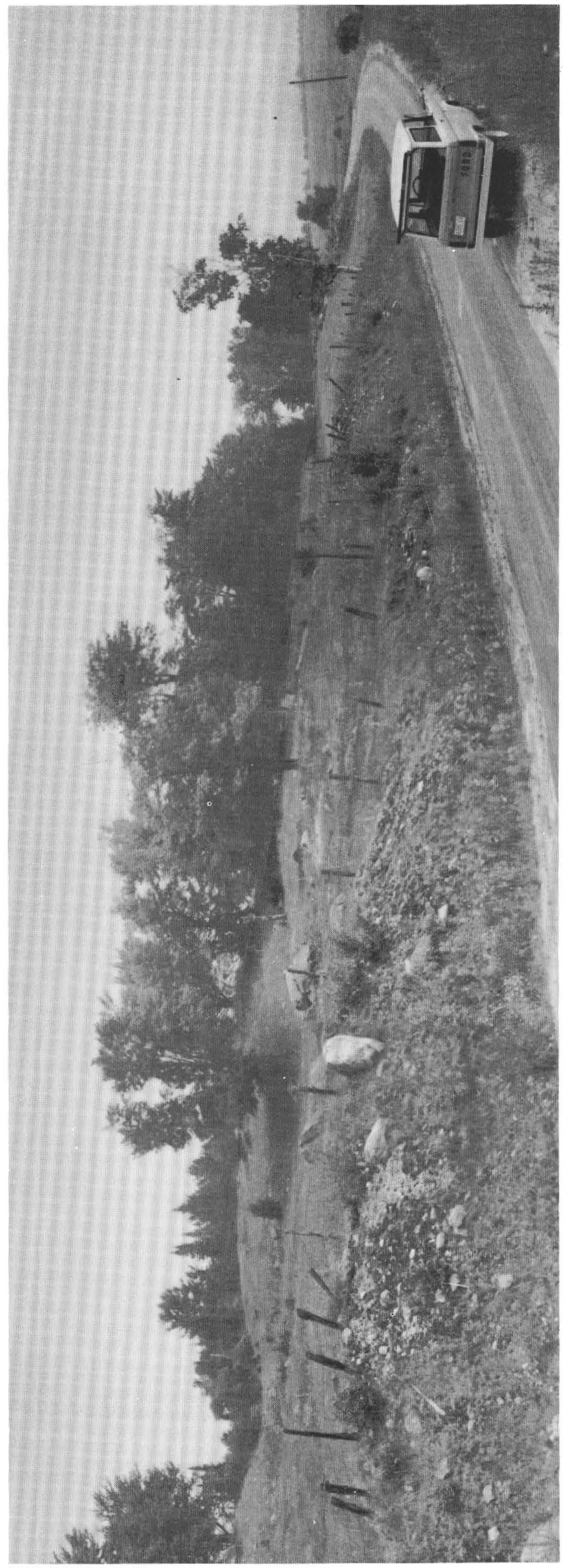


FIGURE 13.—Boulder-covered knolls form small recessional moraine in Salmon River valley (Clinton County) about 4 miles southwest of Peasleeville. View to north and northeast. Ice stood on east side of moraine and discharged melt water to the west.

TABLE 1.—*History of deglaciation of the northeast Adirondack region, New York*

Episode or stand (numbers refer to ice-front position shown on plate 1)	Location of stand or area deglaciated during episode (points near ice front listed in order from northwest to southeast)	Description and interpretation	Age of episode or stand (estimates in years B.P.)
Loon Lake episode (1)	Duane Center–Walker Mill–Loon Lake–Merrilville.	Knolls and ridges of gravel and till form recessional moraine with ice-contact slope on northeast side as much as 200 feet high. Glaciofluvial deposits that form pitted outwash plains were laid down by southwest-flowing melt-water streams from the ice front. Drainage was west by way of St. Regis River valley. Kames and ice channel fillings west of Duane Center, not shown on pl. 1, mark head of older esker system extending southwest for about 40 miles (Chadwick, 1928; Buddington and Leonard, 1962, fig. 2).	About 12,700.
Owls Head–Redford stand (2).	Owls Head–Ragged Lake–Middle Kiln–Redford–Clintonville	In the St. Lawrence basin, the ice front retreated 4 to 8 miles northeast as far as position 2a, leaving masses of stagnant ice in some of the valleys. Knolls and ridges of sand, gravel, and till form recessional moraine near Owls Head. Extensive glaciofluvial deposits south and southeast of the moraine were laid down by south-flowing melt-water streams that escaped west around and across masses of dead ice into the valley now occupied by Lake Titus. Glaciofluvial deposits south of Ingraham Lake, recessional moraine near Ragged Mountain, and abandoned ice-marginal stream channels near Middle Kiln probably mark the same ice-front position (2a). In the Champlain basin, the Saranac River valley was blocked by the ice sheet near Redford (2b), and the waters were diverted by way of several abandoned stream channels into the Ausable River Valley, perhaps contributing to the formation of the Coveville delta of glacial Lake Vermont near Clintonville. The ice front (2b) is marked by small recessional moraines near Redford, in the Salmon River valley west of Peasleeville, and on edge of the uplands northwest of Clintonville.	About 12,600.
Trout River–Moffitsville stand (3).	Whippleville–Lyon Mountain–Moffitsville–Keeseville.	In the St. Lawrence basin, the ice front retreated 3 to 6 miles north, leaving masses of stagnant ice in the Salmon River valley and probably elsewhere. The ice front (3a) is marked by knolls and ridges of drift (recessional moraines) near mouth of Trout River valley and in the valley south of Upper Chateaugay Lake. In the Champlain basin, the ice front retreated 2 to 4 miles to position 3b, leaving dead-ice masses in the valleys of True Brook, Salmon River, and probably elsewhere. Melt-water streams built kames adjacent to the stagnant ice. The ice front (3b) is marked by small recessional moraines east of Peasleeville, abandoned stream channels, and glaciofluvial deposits east of Terry Mountain, and small recessional moraines between Harkness and Keeseville. The Saranac River flowed east down the Salmon River valley, dissecting the moraine (2b) southwest of Peasleeville.	
Malone–Schuyler Falls stand (4).	Malone–Brainardsville–Dannemora–Schuyler Falls.	In the St. Lawrence basin, the ice front retreated 3 to 7 miles to the mountain front (4a), leaving masses of stagnant ice in some of the valleys. Ice-marginal channels and small masses of glaciofluvial deposits are present near Brainardsville and south of Ellenburg Center (just east of the St. Lawrence-Champlain divide). In the Champlain basin, the ice front retreated 2 to 6 miles leaving stagnant ice masses in the Saranac and the Salmon River valleys. During the retreat, glaciofluvial deposits (kames) were laid down by melt-water streams, and south of the Salmon River, other ice-marginal streams cut channels in drift emptying into glacial Lake Vermont at the Coveville stage. Ice-front position 4b marks the western limit of the prominent recessional moraine near Cadyville. To the south, this ice front formed the dam at the north end of Lake Vermont.	
Chateaugay–Cadyville episode (4, 5, 6, 7, and 8).	Area between Malone–Brainardsville–Dannemora–Schuyler Falls stand and Frontier–Clinton Mills–Ellenburg Depot–West Beekmantown stand.	In the St. Lawrence basin, the ice front retreated 8 to 10 miles, almost to the International Boundary (5). During the retreat an ice-dammed lake in the St. Lawrence basin drained east across cols into the Champlain basin, removing the drift to expose bedrock in an area extending from near Clinton Mills south to the vicinity of Ellenburg. The ice front readvanced southwest to position 6, about 3 miles south of Churubusco. Outlet streams from small ice-dammed lakes in the Champlain basin crossed into the St. Lawrence basin where, augumented by melt water, they cut many west-draining ice-marginal channels in the area between Belmont Center and Chateaugay as the ice front retreated from position 6 to position 7. The Trout, Little Trout, and Chateaugay Rivers built deltas into glacial Lake Iroquois (Coleman 1937).	
		A short retreat of about 2 miles brought the ice front to position 8 where several prominent moraines were built damming a lake in the valley of the North Branch and English Rivers which overflowed to the west through a col north of Clinton Mills into the St. Lawrence basin.	About 12,400.
		In the Saranac River valley near Cadyville, a belt of recessional moraines (4b to 8) was built by a southwest-moving ice tongue that blocked the mouth of the valley, causing runoff and melt water to flow south along the ice edge. The moraine may record a minor readvance. As the ice front retreated northeast into the Champlain basin, it built several linear morainal ridges, and south-flowing ice-marginal streams	12,400 to 12,500.

TABLE 1.—*History of deglaciation of the northeast Adirondack region, New York*—Continued

Episode or stand (numbers refer to ice-front position shown on plate 1)	Location of stand or area deglaciated during episode (points near ice front listed in order from northwest to southeast)	Description and interpretation	Age of episode or stand (estimates in years B.P.)
Chateaugay–Cadyville episode (4, 5, 6, 7, and 8)—Continued		cut successive channels on both the north and the south slopes of the Saranac River valley, washed clean, large areas of bedrock, and emptied into Lake Vermont at the Coveville stage.	
Covey Hill episode (8, 11, 14, and 15).	Area between Frontier—Clinton Mills–Ellenburg Depot–West Beekmantown stand and Franklin Centre–Covey Hill–Cannon Corners stand.	Near the St. Lawrence—Champlain divide, the ice front retreated about 2 miles to the north and east, and Lake Iroquois in the St. Lawrence basin overflowed to the east across the divide at a point 2 to 3 miles west of Covey Hill, Quebec. At about this time, the level of Lake Vermont dropped 150 to 175 feet to that of the Fort Ann stage. The overflow from Lake Iroquois emptied into a series of ice-marginal lakes in the northeast-draining valleys of the English, North Branch, and Great Chazy Rivers .The southeasterly overflow of these lakes, perhaps in part catastrophic, cleaned off bedrock divides in an area extending roughly from West Chazy northwest to the International Boundary. The outflow from Lake Iroquois cut the deep rock-walled gorge on the divide west of Covey Hill and discharged into Lake Vermont at the Fort Ann stage. During this interval, the Fort Ann lake rose 50 to 80 feet to its highest stand. There were probably several minor advances and retreats of the ice front as Lake Iroquois in the St. Lawrence basin was lowered to the level of Lake Vermont in the Champlain Basin. During the Covey Hill episode, a subglacial stream built the Ingraham esker (Woodworth, 1905a) in the lowlands north of Plattsburgh.	12,400 to 12,200.
Lowering of level of Lake Vermont.		Ice front retreated to vicinity of Montreal. St. Lawrence Lowlands still dammed by ice sheet near Quebec City. Differential uplift closed southern outlet of Lake Vermont, and it drained northeast along the southeast side of St. Lawrence Lowlands between the ice sheet and the valley wall (glacial Lake New York, Wagner, 1969).	
Champlain Sea ------		Ice front retreated to north side of St. Lawrence River, and marine waters invaded the St. Lawrence and Champlain valleys.	12,000 to 10,500.
Formation of Lake Champlain.		Differential uplift closed the connection to the ocean, and the marine waters were replaced by a fresh-water lake.	10,500 to 10,000.

rather than the product of the small modern stream. The late-glacial river dissected the moraine west of Peasleeville to form the broad gravel-covered flood plain and to build a gravel delta into a small ice-dammed lake. The eastward growth of the delta stopped when the ice front in the Saranac River valley to the north had retreated east to the point where the Saranac River no longer overflowed south into the headwaters of the Salmon River but assumed an easterly course toward Cadyville. The stream in the Salmon River valley was greatly reduced in size and its load greatly diminished.

MALONE-SCHUYLER FALLS STAND (4)

In the St. Lawrence basin this ice-front position (4a) is largely inferred.

In the Champlain basin, the Malone-Schuyler Falls stand (4b) is drawn along the western edge of the moraine near Cadyville in the Saranac Valley and the western edge of the areas of bare rock on the uplands to the south. This stand marks the ice-front position at the beginning of the formation of the moraine and of the abandoned stream channels, kames, and bare-rock areas associated with it.

In the Saranac Valley during the ice-front retreat from Moffitsville (3b) to the west side of the moraine near Cadyville (4b), glaciofluvial deposits were laid down as ice-marginal(?) kames. Sand bodies near Moffitsville rise to altitudes slightly above 1,000 feet and presumably were deposited by the Saranac River as it flowed east toward Cadyville along the north side of the ice (stagnant?) in the Saranac Valley. The floor of the lowest abandoned stream channel on the Saranac-Salmon divide is more than 100 feet above the top of the sand bodies near Moffitsville.

East of Terry Mountain, small recessional moraines and south-draining ice-marginal channels that end in glaciofluvial deposits were formed during ice retreat from position 3b to 4b.

CHATEAUGAY-CADYVILLE EPISODE (4, 5, 6, 7, AND 8)

This episode includes the uncovering of the uplands from the northeast front of the mountains (4a and b) north almost to the Canadian border, and the moraines in the valleys of the Great Chazy River and its North Branch (8) and of the Saranac River near Cadyville (4b to 8).

In the St. Lawrence basin, the ice edge retreated northward from the mountain front to a point on the St. Lawrence-Champlain divide about 3 miles south of Churubusco at an altitude of about 1,350 feet (ice-front position 6, pl. 1). A small ice-dammed lake east of the divide overflowed west, cutting small west-sloping stream channels, now abandoned, at altitudes of 1,335, 1,325, and 1,105 feet (pl. 1).

Five to 10 miles west of these small channels are many large west-draining abandoned stream channels at altitudes ranging from about 1,300 feet to about 1,050 feet. These channels are presumably ice-marginal channels cut by melt water and (or) by overflow from the small lake east of the divide as the ice edge retreated northward (positions 6 and 7).

Evidence suggests that, prior to the cutting of the abandoned stream channels between Churubusco and Malone, the ice edge may have retreated nearly to the Canadian border (ice-front position 5) and then readvanced to position 6, about 3 miles south of Churubusco. In northern Clinton County, west of the St. Lawrence-Champlain divide, as MacClintock and Stewart pointed out (1965, p. 58, pl. 1B), the bedrock is largely concealed beneath thick drift, whereas east of the divide between Clinton Mills and Ellenburg, the drift is thin and patchy. Over large areas the bedrock is covered only by a thin layer of rubble or of peat. This contrast in extent and thickness of surficial mantle suggests the washing of the east slope of the divide by running water that spilled over the divide from an ice-dammed lake to the west.

This easterly flowing water could have entered the Champlain basin by way of several cols on the divide at altitudes ranging from just above 1,300 feet about 4 miles south of Churubusco down to about 1,050 feet near the International Boundary. The easterly flowing water washed the east slope of the divide roughly from Clinton Mills south to Ellenburg and presumably entered a local ice-marginal lake in the headwaters of the North Branch Great Chazy River. This washing antedates that which cleaned off the large areas of bare rock between Covey Hill, Quebec, and West Chazy, because these latter bare-rock areas are at lower elevations than the area between Clinton Mills and Ellenburg.

The ice-dammed lake west of the divide that overflowed through the cols must antedate the cutting of the west-draining abandoned stream channels between Churubusco and Malone. No deltas or beaches of this high-standing lake (alt 1,050–1,300 ft), have been positively identified. This lake may have been of limited extent, or it may have been an arm of Lake Iroquois.

In the valley of the North Branch Great Chazy River, the Chateaugay-Cadyville episode ended with the building of the moraine near Ellenburg Depot (ice-front position 8, pl. 1 and fig. 11), where the front may have paused for a time. The ice built the moraine along the east edge of an ice-dammed lake that drained to the west, probably by way of the large col north of Clinton Mills at an altitude of 1,085 feet. If the ice that built the moraine south of Miner Lake dammed the valley of the Great Chazy River and formed a lake, this lake drained north into the valley of the North Branch, perhaps washing the drift cover from part of the area of bare rock on the divide between the two valleys. There is no evidence of the escape of lake waters from the Great Chazy Valley around the east end of the moraine south of Miner Lake (ice-front position 8). However, the area of bare rock on the Great Chazy–North Branch divide extends down to an altitude of about 1,050 feet, too low relative to the col at 1,085 feet to have been washed completely by northwest-flowing lake waters. Before the washing of the area on the divide was completed, the direction of flow changed from northwest to southeast.

At about the time the ice blocked the valley of the North Branch and built the moraine near Ellenburg Depot (ice-front position 8), a stream flowing south in or under the ice sheet in the lowlands north of Plattsburgh began to build the Ingraham esker, a ridge of gravel about 10 miles long (Woodworth, 1905a, pl. 4). The esker stream discharged into Lake Vermont and, as the ice edge retreated north, the bouldery ridge formed in the ice was partly buried by sand deposited in the glacial lake. Later, submergence of the esker in the Champlain Sea resulted in extensive modification of the ridge by erosion and deposition (Denny, 1972).

V. K. Prest (1970), in his thoughtful review of the present state of knowledge of the Quaternary geology of Canada, has presented a slightly different interpretation of the late-glacial history of the area northeast of the mountains, as shown in table 2. Glacial Lake Iroquois (Coleman, 1937) in the Lake Ontario basin drained east near Rome, N.Y. (fig. 14), southwest of the Adirondacks, into the Mohawk River valley. Coleman (1937) and most later workers believe that the northward retreat of the ice front in the Ontario basin and the St. Lawrence Valley in time opened a new outlet for Lake Iroquois, the channel near Covey Hill, altitude 1,010 feet. The Rome outlet was then abandoned. Field observations made by Prest in 1967, in company with E. P. Hen-

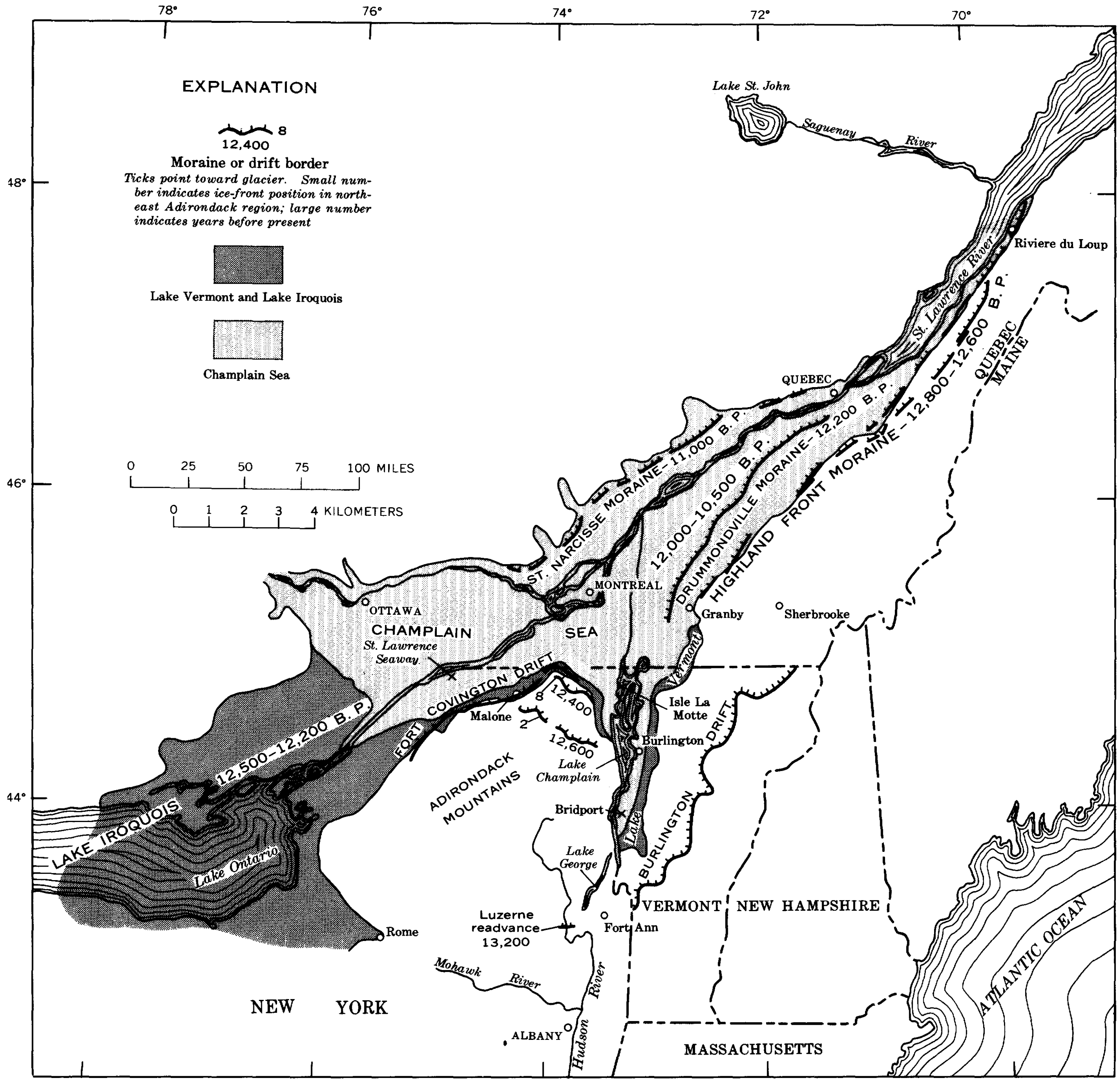


FIGURE 14.—Generalized map showing location and possible age of moraines and late-glacial water bodies in parts of the St. Lawrence and Champlain Lowlands. As used here, Lake Iroquois includes the Frontenac and Sydney phases of the post-Iroquois lakes as defined by Prest (1970, p. 727 and fig. XII–16f and g). Compiled largely from MacClintock and Stewart (1965), Prest, Grant, and Rampton (1968), and Stewart and MacClintock (1969).

derson and the author, led Prest (1970, p. 727) to question the earlier interpretation.

Though the Iroquois shoreline is reported at only 1,100 feet southwest of Covey Hill there is evidence of shoreline features to about 1,250 feet, 1¼ miles south of Churubusco. This is considered to be an Iroquois shore rather than that of local lake.

Prest also suggested that

> a small embayment of Lake Iroquois extended eastward from Churubusco [into the Champlain basin] as a reentrant between the receding northern and eastern ice-fronts around Adirondack Mountains. This ice, probably Fort Covington [MacClintock and Stewart, 1965, p. 60], built an end moraine

and ribbed moraine complex south and southwest of Covey Hill.

These are the moraines near Ellenburg Depot and north of Clinton Mills (ice-front position 8, pl. 1).

C. S. Denny (USGS, 1968) believes that an outlet opened southeast of Ellenburg, N.Y., which allowed main Lake Iroquois to breach the end moraine and discharge into glacial Lake Vermont farther south in Lake Champlain valley.

This discharge was thought to mark the beginning of the washing of Flat Rock, near Altona.

Prest (1970, p. 727) also described how the draining of the arm of Lake Iroquois southeast of Ellenburg lowered lake level about 75 feet to what he named the "Ellenburg phase of the post-Iroquois lakes." He went on to say

as glacier recession was resumed, Covey Hill outlet (sill elevation 1,010 feet) was uncovered and the post-Iroquois lake level was lowered a further 75 feet. This lake phase was named glacial Lake Frontenac and is herein considered the Frontenac phase of the post-Iroquois lakes * * *.

I now question some parts of Prest's interpretation of the history of the divide area. The abandoned stream channels trending diagonally down the west slope of the St. Lawrence-Champlain divide between Malone and Churubusco (pl. 1; see also MacClintock and Stewart, 1965, fig. 21) are best explained as the work of ice-marginal streams. Successive channels were cut as the ice edge retreated northward (positions 6 and 7). Because the channels do not appear to have been modified by ice or by wave action, I believe that they record the retreat of the ice margin when Lake Iroquois was still discharging into the Mohawk River valley by way of the Rome outlet.

The moraine near Cadyville was built by a short tongue of ice moving southwest up the Saranac Valley from a large ice mass in the Champlain Lowland. There may have been stagnant ice in the Saranac Valley west of the moraine. At the beginning of the episode of moraine building, the ice front stood near the west edge of the moraine (4b) and crossed over the east end of Johnson Mountain into the headwaters of the Great Chazy River, which presumably were dammed by the ice edge, although no record of this lake has been found. South of the Saranac River, the ice front (4b) pressed against the east end of the ridge to the south for a short distance before turning east into the Champlain basin to form the ice dam at the north end of Lake Vermont. Under the reconstruction of glacial history presented here, the building of the moraine near Cadyville took place during the time that the ice front in the St. Lawrence basin retreated from the mountain front (4a) nearly to the Canadian border (8).

The ice tongue in the Saranac Valley built small moraines as its edge retreated to the northeast. The Saranac River, blocked by the ice, was diverted south around the east end of the bedrock hill that separates the Saranac and Salmon River valleys; the waters carved channels in drift and in bedrock, washed drift from areas of bedrock, and deposited their load close to the north end of Lake Vermont. Drainage ponded north of Johnson Mountain flowed south into the Saranac Valley, cutting channels in drift and depositing gravel and sand against the ice in the valley.

As the ice edge retreated, many channels were cut on the uplands west of West Beekmantown by water ponded north of the Saranac Valley–Smith Wood Brook divide. The water escaped southward into the head of Lake Vermont, which was then only a few miles south of the Saranac River.

COVEY HILL EPISODE (9 THROUGH 15)

The interval between the withdrawal of the ice from the moraine near Ellenburg Depot (8) and the uncovering of Covey Hill and the merging of the glacial lakes in the St. Lawrence and Champlain Valleys is called the Covey Hill episode. During this episode, the channel near Covey Hill was eroded, large areas were swept clean of their drift cover, and the ice-dammed lake in the upper St. Lawrence and Ontario basins was lowered more than 200 feet to the level of the glacial lake in the Champlain Valley.

The ice edge had to maintain a position across the divide close to Covey Hill in order to hold the ice-dammed lake in the St. Lawrence Valley southwest of Montreal in a position where it could overflow into the similar lake in the Champlain Valley. If the ice front had retreated a mile or two northeast, the lakes would have become confluent, and the water in the upper St. Lawrence and Ontario basins would have discharged catastrophically into the Champlain Valley. The channel near Covey Hill, altitude 1,010 feet, that controlled lake level in the upper St. Lawrence and Ontario basins, was about 270 feet above the level of Lake Vermont in the Champlain Valley. On the other hand, if the ice front had readvanced southwest a few miles, it would have blocked all the possible outlets by which St. Lawrence Valley waters could have overflowed into the Champlain Valley. During the Covey Hill

TABLE 2.—*Ice-marginal and shore features near Covey Hill, Quebec, as interpreted by Prest (1970) and by Denny (this report)*

Feature and altitude	Denny (this report)	Prest (1970)
Gravel and sand near Churubusco, N.Y.; alt 1,250 ft.	Glaciofluvial deposits (not shown on pl. 1).	Beach deposits of Lake Iroquois.
Large area of abundant outcrops of Potsdam Sandstone on divide between Graves Brook and North Branch about 2 miles southeast of Ellenburg (pl. 1); alt 1,200–1,300 ft.	Ice-dammed lake in valley of North Branch overflowed across divide into valley of Graves Brook and washed clean the bare-rock area between the two streams.	Ellenburg phase of the post-Iroquois lakes ("Outlet * * * southeast of Ellenburg"; Prest, 1970, p. 727).
Channel near Covey Hill, Quebec; alt 1,010 ft.	Outlet of Lake Iroquois.	Frontenac phase of the post-Iroquois lakes.
Large area of abundant outcrops of Potsdam Sandstone north of English River; alt 750–880 ft.	Lake Iroquois overflowed by way of an ice-marginal stream coming around the north side of Covey Hill. The stream is assumed to have been the outlet that controlled the level of Prest's Sydney phase of the post-Iroquois lakes, alt about 885 ft (Prest, 1970, p. 727). The lake water washed clean the bare-rock area.	
Beaches marking the highest stand of Fort Ann stage of Lake Vermont; altitude at International Boundary about 740 ft.	Lake Vermont merged with lake in upper St. Lawrence and Ontario basins.	Belleville-Fort Ann phase of the post-Iroquois lakes.

episode, as here interpreted, the ice edge moved back and forth across the area of the present International Boundary several times, and the washing of the bedrock and the lowering of the level of ponded water in the St. Lawrence basin took place during several intervals, not all at one time.

Retreat of the ice front from the moraines north of Clinton Mills and near Ellenburg Depot (ice-front position 8, fig. 11) uncovered the col (1,010 ft) on the divide west of Covey Hill; the opening of this col permitted Lake Iroquois to overflow into the Champlain basin and presumably to abandon the outlet near Rome. Southwest of the col, the divide is bare rock up to an altitude of about 1,050 feet, suggesting that Lake Iroquois may have stood at that level when it first crossed the divide. Presumably the col at 1,010 feet soon became the control for the level of outflow of Lake Iroquois.

At first the ice-marginal streams turned to the south down the slope (ice-front position 9, fig. 11) just east of the moraine near Ellenburg Depot and emptied into a long narrow ice-marginal lake or series of lakes between ice-front positions 8 and 9. These ice-dammed lakes drained east across the southern edge of Flat Rock.

Continued retreat of the ice front caused the water escaping from Lake Iroquois to flow south along the ice edge (position 10) at altitudes ranging from about 900 to 950 feet; the flow cut bedrock(?) channels a mile or two to the south of the International Boundary and emptied into a long narrow ice-marginal lake or chain of lakes that overflowed to the east across Flat Rock, washing it clean and cutting narrow gorges in the bedrock. Some of the material removed from Flat Rock was deposited to the east on Cobblestone Hill (pl. 7), where it was later heaped up into conspicuous beach ridges.

On Flat Rock, the floor of the Dead Sea at the head of Cold Brook (pl. 7) is said to be 90 feet below the rock rim on the west side of the rock basin. Woodworth thought that such a deep pool could not have been eroded by a small ice-marginal stream descending over "so slight a fall as the rock cliff at its head, but it is quite conceivable that a heavy torrent might have produced the results" (Woodworth, 1905a, p. 20).

On the north and east sides of Flat Rock, beach deposits of the Fort Ann stage of Lake Vermont extend up to a point 80 feet above the lowest part of the area of bare rock. The overlap of the Fort Ann beaches on the bare rock suggests that during the washing of Flat Rock the level of glacial Lake Vermont may have stood as much as 80 feet below its highest stand. If this was so, then the level of Lake Vermont rose during the Fort Ann stage, probably because of uplift of the outlet at the south end of the Champlain basin near Fort Ann, N.Y., about 100 miles to the south (fig. 14). Woodworth (1905b, p. 162) interpreted the overlap of beach on bare rock as indicative of a rise in lake level after the drift cover had been washed from the bare-rock area by ice-marginal streams.

When the ice edge had retreated as far as position 11 (fig. 11), water from the Covey Hill channel turned south into a lake in the English River valley. This lake in turn overflowed to the southeast along the ice edge across the divide between the English River and the North Branch. Small northwest-trending recessional moraines just south of the English River–North Branch divide mark positions of the ice front that controlled the overflow. Perhaps much of the morainal material and the gravel associated with it was washed off the bare-rock area to the northwest by ice-marginal streams. Removal of drift by overflow is essentially the explanation advanced by Woodworth (1905a, p. 12).

As the ice edge retreated to the northeast (toward ice-front position 12), water from the Covey Hill channel escaped south across a col at an altitude of about 860 feet (pl. 6) and flowed into a prominent bedrock channel that runs south to drain into the ice-dammed lake in English River valley. This lake continued to drain across the divide to the south, cleaning off the bedrock.

As the ice front continued to retreat (position 12), water from the Covey Hill channel followed a more southeasterly course across the western part of the bare-rock area north of English River, along the west side of a prominent recessional moraine. Near English River, the ice front is assumed to have turned east and crossed the English River–North Branch divide where water ponded in the English River valley continued to overflow to the southeast, removing the drift.

When the ice edge had retreated to the northeast down the English River–North Branch drainage divide a short distance beyond ice-front position 12, the overflow from the lake in the English River valley deposited bouldery material at the northeast end of the bare-rock area, about 0.75 mile south of Cannon Corners (pl. 6). Shallow kettles in these bouldery deposits suggest deposition in association with glacier ice where the overflow entered Lake Vermont.

The large area of bare rock north of English River could have been washed by lake water coming either by way of the gorge west of Covey Hill or around the north side of the hill. The existing segments of moraine are not sufficient to define the ice edge precisely, but in the reconstruction of figure 11, I have assumed that at ice-front position 12, water came from the gorge and at ice-front positions 13, 14, and 15 the water was coming around the north side of Covey Hill.

The history of deglaciation presented here calls for the lowering of the level of glacial Lake Iroquois in several stages rather than in one catastrophic outflow. If the lake in the St. Lawrence Valley at the level of the Covey Hill outlet, altitude 1,010 feet, had been lowered at once to the Fort Ann level, at about 750 feet, one would expect to find on the north and east slopes of Covey Hill large areas of bare rock swept clean by the flood; instead, these slopes are boulder covered. I assume that the lake to the west was lowered gradually, perhaps because the ice edge moved up and down the north slope of the hill. Some of the water may have escaped beneath the ice.

Prest (1970, p. 727) believes that the existing strandlines in the Ontario basin call for a brief still-stand of a post-Iroquois lake on the north side of Covey Hill below the level of the lip of the gorge.

> When the ice withdrew from the northern and eastern flanks of Covey Hill there was a major drop in lake levels of some 125 feet. The short stand at this level, probably occasioned by an ice-marginal fluctuation on the northern side of Covey Hill, is termed the Sydney phase lake by E. Mirynech (1967).
>
> In the Trenton embayment in eastern Ontario the lake was lowered a further 30 to 75 feet according to Mirynech, prior to a significant halt responsible for development of the Belleville beach. Isobases drawn on the Belleville beach would place the strandline on Covey Hill at about 750 feet. This is the same as that of the Fort Ann phase of glacial Lake Vermont which expanded northward as the eastern side of Covey Hill was uncovered by the ice.

I suggest that the removal of drift from part of the bare-rock area north of English River (positions 13, 14, and 15) was accomplished during the "Sydney phase" by lake water escaping from the St. Lawrence Lowlands around the north side of Covey Hill. This water eroded low bedrock knolls near the Canadian border (pl. 6), washed bare the bedrock area now exposed, and probably entered Lake Vermont near Cannon Corners.

The position of the recessional moraines in the English River valley suggests a minor readvance of the ice front. The reconstruction of figure 11 shows the ice front first retreating to position 13 near Cannon Corners, then advancing up the English River valley about a mile (position 14), and then retreating again to a line north of Cannon Corners (position 15).

The evidence for this minor readvance is as follows: The ridge of recessional moraine on the east side of the bare-rock area north of English River (pl. 6) joins at its south end a small northwest-trending segment of recessional moraine; together the two enclose a small swamp. These ridges would have partly blocked the southerly flow of water escaping around the north side of Covey Hill. Perhaps the segments of moraine near the swamp are giant gravel bars on the bed of a south-flowing river. If so, the sequence of events may have been as follows: When the edge of the ice sheet retreated to the east side of the bare-rock area, water from the north side of Covey Hill flowed south along the ice edge, cleaned off an area of bedrock somewhat larger than the area of bare rock existing today, and deposited its load in the valley of English River. During that interval, the level of a lake in the St. Lawrence Valley was lowered about 125 feet and was stabilized at about 885 feet for a short time (the "Sydney phase lake" of Prest). In the valley of English River, following this brief episode of erosion, the ice edge readvanced about a mile west across the southern part of the area previously cleaned off, plowing up bouldery material from the valley floor to form the moraine along White Road. The water from the north side of Covey Hill escaped around the west end of the lobe (point *B*, pl. 6) into the English River valley and east across the divide to the south into Lake Vermont just south of Cannon Corners.

During the shrinkage of this minor lobe, the water escaped across the bouldery ridges, perhaps first near point C and later near point D (pl. 6) on the east side of the bare-rock area. Immediately thereafter, the water followed the channel at the north end of the bare-rock area (ice-front position 15) to enter Lake Vermont near the present International Boundary (point E, pl. 6). The ice sheet soon retreated off Covey Hill for the last time, and Lake Vermont merged with the lake in the upper St. Lawrence and Ontario basins (fig. 14).

DEPOSITS AND SHORE FEATURES OF LATE-GLACIAL WATER BODIES

The late-glacial history of the Champlain Valley involves an ice-dammed fresh-water lake followed by an incursion of the sea, the latter ending about 10,500 years B.P. Woodworth, who made detailed studies throughout the Hudson-Champlain lowland (Woodworth, 1901, 1905a, b), recognized both marine and fresh-water deposits. Later, Fairchild (1919) concluded that the beaches and deltas described by Woodworth were formed in an arm of the sea extending from New York City to Montreal. Chapman's (1937) study of the Champlain Valley largely reaffirmed Woodworth's belief that the higher shoreline features are lacustrine and the lower ones marine.

Lake Vermont, named by Woodworth (1905b, p. 190–206), was the fresh-water lake that occupied the Champlain Valley during late-glacial time. It was dammed on the north by the edge of the ice sheet and overflowed to the south across a bedrock divide into the headwaters of the Hudson River. Since Woodworth's day, the only descriptions of the lake as a whole are by Chapman (1937) and by Stewart and MacClintock (1969). Chapman recognized and named two stages of lake history. The older stage, the Coveville, he believed was graded to Woodworth's Coveville outlet, an abandoned channel and dry waterfall on the west bank of the Hudson River near Schuylerville about 30 miles north of Albany. (See Schuylerville 15-minute quadrangle.) Chapman named the younger stage the Fort Ann for its abandoned outlet, which contains giant potholes, near Fort Ann, N.Y., east of Glens Falls (Chapman, 1937, figs. 6, 7, and 8). (See also Fort Ann 15-minute quadrangle.) A still higher and older stage, the Quaker Springs, has been recognized in the southern part of the Champlain basin (Woodworth, 1905b, p. 103; Stewart and MacClintock, 1969, p. 163; Connally and Sirkin, 1971). The marine invasion that followed Lake Vermont has long been known as the Champlain Sea (Karrow, 1961; Elson, 1969).

In the lowlands of the Plattsburgh area west of Lake Champlain, the deposits of glacial Lake Vermont resemble those of the Champlain Sea; they cannot be separated on a lithologic basis. The deposits of these late-glacial water bodies are in two belts that parallel the edge of the uplands. The western or higher belt consists of elongate bodies of sand and gravel (Lake Vermont, pl. 1) that are separated by areas of ground moraine from the fossiliferous sand, gravel, silt, and clay of the eastern or lower belt (Champlain Sea, pl. 1). Only along the principal streams are the deposits of the two belts in contact.

Chapman studied the beaches and deltas of the two belts in parts of the Dannemora quadrangle and made a precise survey of the beaches of the upper group in a small area west of Peru (Chapman, 1937, fig. 9). The upper belt as far north as the Saranac River is considered by Chapman to include both the Coveville and Fort Ann stages of Lake Vermont, but north of the Saranac River only the Fort Ann stage is present. I have adopted Chapman's terms, Fort Ann and Coveville, for the two groups of deposits and shoreline features of the upper belt. The fossiliferous deposits and shoreline features of the lower or eastern belt I assign, as did Chapman, to the Champlain Sea.

The deposits of glacial Lake Vermont cannot be traced continuously from the Plattsburgh area south to the region of the outlet, nor are there fossils or carbon-14-dated materials with which to correlate the isolated deposits from place to place. It is a well-established fact that the glaciated area of northern United States and Canada was uplifted when the ice sheet disappeared, the amount of uplift increasing inward from the periphery of the glaciated area (Daly, 1934) Therefore, if the altitudes of the shore features of glacial Lake Vermont are plotted on a profile running north-south more or less in the assumed direction of maximum tilt, the upper limit of these features should define the position of the highest stand of the lake. The limit rises northward and is probably, but not certainly, a time line; all features on the line formed at about the same time. The correlation of the shore features (beach, spit, terrace, wave-cut cliff, or delta) throughout the Champlain basin has been and is based largely on their location and altitude in relation to possible outlets, as shown on such north-south topographic profiles (Woodworth, 1905b, pl. 28; Chapman, 1937, figs. 15 and 16; Denny, 1967, 1970).

The deposits of the late-glacial water bodies are dominantly sand; gravel is less abundant and commonly occurs only near the larger streams. Near Lake Champlain in the drainage of the Great Chazy River, silt and clay are abundant. In favorable localities, such as on the east slope of the uplands near West Beekmantown, the upper limit of wave action on recessional moraine is clearly marked (Stephens and Synge, 1966, figs. 9–12). In many places the upper foot or two of the till is a faintly stratified gravelly material, and the local relief of the surface of the ground moraine or of the glaciofluvial deposits is less than on similar materials above the limit of wave action.

BEACH DEPOSITS

Beach deposits closely resemble the adjacent drift from which they were derived. Where the drift contains many fragments of Potsdam Sandstone, the beach ridges are composed of a flaggy gravel (fig. 15). (See also Denny and Goodlett, 1968, figs. 2–5.) Where the drift contains abundant large clasts, the beach-forming materials are boulder or pebble gravel, commonly massive or faintly stratified (fig. 16). Only a few sandy beach ridges are recognized, in part because such features have been smoothed by frost action, tree-throw, or plowing. The beach ridges commonly occur in groups, some of which may be 3 or 4 miles long. Individual ridges range in length from a few hundred feet to nearly a mile, but most are less than half a mile long. The width of a ridge from trough to trough ranges from about 40 to 100 feet, the height from crest to trough, from about 1 to 5 feet.

The availability of gravelly materials and exposure to wave action appear to be important factors in the location and abundance of beaches. Beaches are well developed on headlands. The most conspicuous examples are those near and southwest of Sciota (Denny, 1970), where a headland underlain by the Potsdam Sandstone projected northeast into the adjacent water body. Just south of Sciota, west of State Highway 22, are beach gravels of the Champlain Sea at the base of low sandstone bluffs that Woodworth (1905a, pls. 22–24) called sea cliffs.

The most spectacular beaches, on Cobblestone Hill at the southeast end of Flat Rock (pl. 7), were described in detail by Woodworth (1905a, p. 32–35, pls. 12, 13, and 14). Pebble to boulder gravel and sand mantle the north and east sides of Cobblestone Hill; the deposit is more than a mile long and as much as 0.3 mile wide and ranges in altitude from about 580 feet at the base of the hill, where the material is chiefly sand, up to about 670 feet where the material is boulder gravel. The gravel is heaped up into beach ridges, as much as 3 feet high and 100 feet wide, that extend along the east side of the hill for at least a quarter of a mile to its south end, where they curve around to the west and north enclosing small depressions between adjacent ridges.

At the north ends of the beach ridges, rounded clasts of white sandstone form a pebble and cobble gravel (fig. 17), but near their south ends the ridges are mantled by angular blocks of sandstone (fig. 18), suggesting that the beaches did not grow southward but rather were formed in place. Thus, the shape of the beach ridges, even though resembling a curved spit, is related instead to the form of Cobblestone Hill. The crest of the hill declines southward to pass beneath the position of the highest stand of the Fort Ann stage. The curvature of the beaches near their southern ends appears to be the result of the shape of the hill crest where it passes below lake level rather than the result of the southward movement of material by longshore currents. The beach deposits are reworked alluvial materials washed off the adjacent bare-rock area just before beach formation. The sand and gravel at the east base of the hill are probably material removed by wave action from the gravel near the top of the hill.

DELTA DEPOSITS

Sand and gravel form deltas at the mouths of the larger streams where they debouched into the late-glacial water bodies (pl. 1). Similar materials that blanket gently sloping plains between deltas are called, for convenience, nearshore deposits. These deposits are, in part, the bottomset beds of deltas and, in part, material spread by longshore currents.

The deltas are composed of medium to fine sand and lesser amounts of coarse sand and pebble to cob-

FIGURE 15.—Cross section of flaggy gravel forming beach ridge of Champlain Sea. View looking north along Mooers-Chazy town line and 0.6 mile east of State Route 22.

ble gravel. Gravel is most abundant near the apex of a delta (figs. 19, 20), where the channel of the delta-building stream was restricted by the adjacent valley walls. Farther downstream the material is largely a well-sorted medium sand (fig. 21). The rivers were fed largely by local runoff from their drainage basins, except perhaps during Lake Vermont time when glacial ice may have contributed melt water and load. If Craft (1969) is correct in supposing that ice remained for a time in the high peaks region after the lowlands had been deglaciated, then perhaps the rivers continued to carry some melt water as late as Champlain Sea time.

Some of the deltas retain much of their initial form; others built at high stands were partly eroded, and the material was redeposited to form deltas at lower levels. The town of Keeseville is on top of a delta built by the Ausable River in Lake Vermont during the Fort Ann stage. Except where dissected by the river, the initial top and foreslope of the delta are well preserved. Along the Saranac River, the sands form a series of discontinuous terraces from the top of Lake Vermont to Lake Champlain. Waves and currents spread a thin blanket of sand along the shore between the deltas. In broad interstream areas near Lake Champlain these sands may be several tens of feet thick.

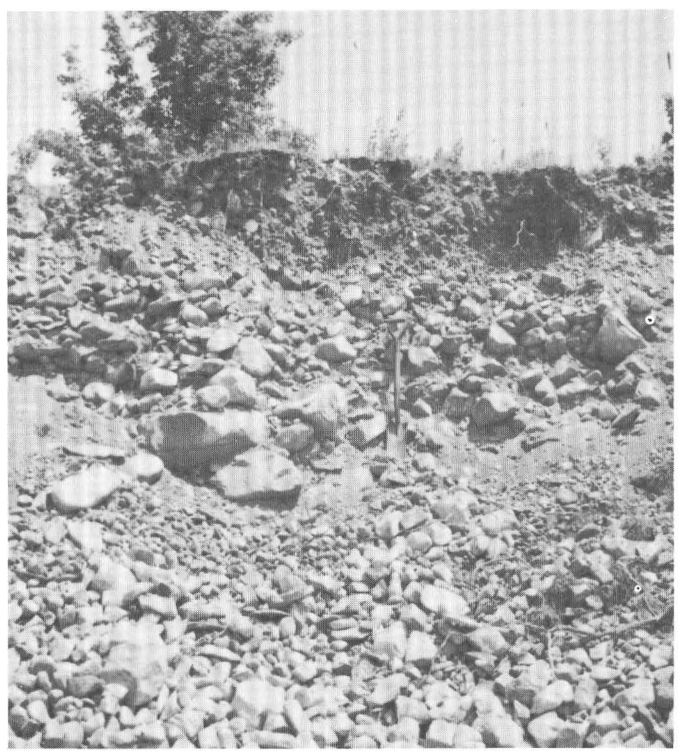


FIGURE 16.—Beach gravel of the Champlain Sea. Material contains a few fragments of marine shells. Exposed in borrow pit on ridge crest, altitude about 300 feet, about 4 miles west of Champlain.

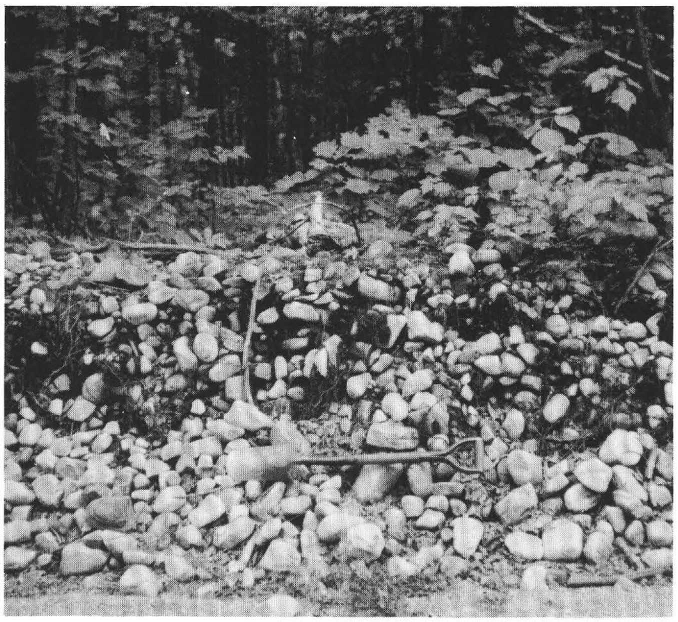


FIGURE 17.—Water-worn cobble and boulder gravel in beach of Lake Vermont. Exposure is in highest beach of Fort Ann stage along east-west road crossing south end of Cobblestone Hill (pl. 7).

TEXTURE

For the Plattsburgh area, mechanical analyses of sandy delta and nearshore deposits and sandy glaciofluvial deposits were made in the field, using small sieves as the sorting unit and a 10-cc graduate as the measuring unit (Azmon, 1961). Most of the approximately 120 samples, whose locations are shown in figure 22*B*, came from a depth of about 3 feet.

GRAIN SIZE

The sand is dominantly medium and fine (fig. 22*A*). Coarse sand is generally restricted to deltaic deposits near the large rivers and to the glaciofluvial deposits. The marine sands are largely deltaic, and size tends to decrease away from the rivers, except along the Salmon River below Schuyler Falls. Grain size shows no consistent relation to slope as measured on a topographic map. Dune sand and sand of the Fort Ann stage are a little finer and more uniformly sized than the marine sand or that of the Coveville stage (fig. 23). In the area between the Salmon and Saranac Rivers (fig. 22*A*), layers of uniformly sized sand parallel the ancient shorelines of Lake Vermont, suggesting that here the sand was spread by wave action near shore. Each of the four rivers in the study area built deltas into Lake Vermont and the Champlain Sea. Size tends to decrease away from the apex on some of these deltas (fig. 24), but, in general, the relation is not consistent.

SORTING

The sands are well sorted. The Trask sorting coefficient ranges from about 1.2 to 1.8 (fig. 22*B*); in general, the coarser the sand, the poorer the sorting.

INTERNAL STRUCTURE

Neither the delta nor the nearshore deposits were well exposed during the course of this study. Excavations in the delta near Keeseville showed both topset and foreset beds. Pits open in the nearshore deposits during the construction of Interstate Highway 87 showed a more or less horizontal stratification and some rather massive beds a few feet thick. A few lenses of crossbedded sand and gravel were observed (fig. 20).

In some of the delta deposits, contorted beds between horizontal strata suggest deformation contemporaneous with deposition, the result of the loading of sand on silty beds, of slumping, or of dewatering.

West of Peru, sands of the Fort Ann stage form small knolls and ridges where the deposits lap up against the upland east of Terry Mountain (west and south of Clark Corners; Denny, 1967). The knolls are 10–30 feet high and 100–300 feet in diameter. They are on a drift-covered surface that slopes eastward with a gradient of about 3° to 4°. Beds of fine sand are highly deformed. In cross section the folds are fan shaped or irregular and measure 3–5 feet across. The deformation appears to die out at depths of 5 or 6 feet below the surface. Perhaps the knolls of deformed sand are material that slumped down the till-covered slope when Lake Vermont drained away.

LOCATION AND AREAL RELATIONS

Deltas were mapped along four large rivers (pl. 1), the Salmon (Franklin County) and its principal tributary the Trout, the Great Chazy, the Saranac, and the Ausable. The drainage basins of these rivers range in size from about 225 to 600 square miles (table 3). Deltas were also mapped along five smaller streams, including the Little Trout and Chateaugay Rivers in Franklin County and Park Brook and the Salmon and Little Ausable Rivers in Clinton County. The drainage basins of the five smaller streams range

FIGURE 18.—Boulder beach on Cobblestone Hill at southeast end of Flat Rock near Altona (pl. 7). View is looking north from near south end of the highest beach. Photograph by G. K. Gilbert.

in size from about 25 to 150 square miles. The Saranac and Ausable Rivers head in the high peaks of the Adirondacks; the maximum relief of the Ausable drainage basin is about 5,000 feet. Of the four large river basins, only one, that of the Great Chazy River, is underlain predominantly by Paleozoic sedimentary rocks, chiefly the Potsdam Sandstone (pl. 3). The others are largely in crystalline rocks of Precambrian age. About two-thirds of the Ausable River drainage basin is underlain by anorthosite (pl. 2).

VOLUME

The volumes of the deltas have been calculated (table 3) in order to compare them with the topographic and lithologic character of the drainage basins upstream from the deltas. Many of them are dissected by streams that have cut down to bedrock, so that along the stream the thickness of the delta deposits can be measured. Well logs gathered from various sources by W. A. Hobba, Jr., of the Geological Survey (written commun., 1967), are also available. The estimated volumes are rough but appear to be the right order of magnitude. For some of the deltas, two estimated volumes are given, one for the original volume and one for the existing remnants.

The younger deltas of the Ausable River are built, in part, of material eroded from older ones. The Coveville delta along the Ausable River west of Clintonville is estimated to have lost 203 million cubic yards of sediment by erosion since it was formed (table 3). Of this, about 122 million cubic yards are estimated to have been eroded by the Ausable River, the remainder by the Little Ausable River. Because the next younger and lower Ausable River delta, the Fort Ann delta near Keeseville, has an estimated original volume of about 261 million cubic yards, nearly half its sediment could have come from ero-

TABLE 3.—*Upper Pleistocene and Holocene deltas of the northeast Adirondack region, New York*

River	Characteristics of drainage basin (above highest stand of Champlain Sea)							Deltas (estimate volumes in millions of cubic yards)		
		Main stream			Lithology[1] (percent of total area of basin)			Coveville stage[2]		
	Maximum relief (ft)	Length (miles)	Slope (ft per mile)	Drainage area (sq miles)	Anorthosite	Other Precambrian	Paleozoic and younger	Original	Existing remnants	Loss by erosion
Salmon and Trout (Franklin County)	2,855	26	110	228	--	67	33	--	--	--
Trout	2,030	16	127	47	--	55	45	42	--	--
Little Trout	1,930	17	114	23	--	44	56	36	--	--
Chateaugay	3,110	26	120	159	--	83	17	24	--	--
North Branch Great Chazy	2,164	19	114	105	--	13	87	--	--	--
Park Brook	780	8	98	27	--	--	100	--	--	--
Great Chazy	3,340	26	128	91	--	40	60	--	--	--
Great Chazy and North Branch	3,340	26	128	196	--	25	75	--	--	--
Saranac	4,472	68	66	589	34	63	3	48	28	20
Salmon (Clinton County)	1,910	15	127	37	--	48	52	141	125	16
Little Ausable	1,700	9	189	36	--	45	55	--	--	--
Ausable	4,984	48	104	501	60	34	6	352	149	[3]203

[1] Fisher and others (1962).
[2] "Lake Iroquois" stage in St. Lawrence Valley (MacClintock and Stewart, 1965, pl. 1B).
[3] 81 to Little Ausable River and 122 to Ausable River.

sion of the older delta. The Fort Ann delta of the Ausable River near Keeseville lost by erosion about 54 million cubic yards (table 3), but this figure is only about one-ninth of the estimated original volume of the younger Champlain Sea delta downstream from Ausable Chasm (446 million cubic yards). I assume, therefore, that most of the delta at the mouth of the chasm is built of debris that came from points farther upstream than the Coveville delta near Clintonville.

The volumes of the deltas and the characteristics of their drainage basins are compared to see if there are any consistent relationships between them. The volume of a delta does not show a close relation to the length of the delta-building stream (fig. 25), except for the deltas of the Champlain Sea, which show a slight tendency to increase in volume as length of streams increases. This is readily understandable because the older Lake Vermont deltas were built in part by melt-water streams, and thus the volumes of these deltas need have no relation to present drainage basins. Of the six streams that built deltas into the Champlain Sea, the delta of the combined Salmon and Trout Rivers in Franklin County is large compared with those of the other five, whereas that of the Great Chazy River and North Branch is small. The large size of the Salmon-Trout River delta could be related to the large mass of glaciofluvial deposits upstream, a readily available source of delta-building material. Perhaps the small volume of the Great Chazy–North Branch delta is related to the large area of Potsdam Sandstone in the river's drainage basin (75 percent). The lowlands of crystalline rocks in the Adirondack region are buried beneath a sandy glacial drift, whereas in areas of Potsdam Sandstone the drift cover is more stony and perhaps thinner and less extensive. As most of the bedload carried by the rivers probably was supplied by drift, it follows that the late-glacial rivers in areas of Precambrian rock may have carried a larger volume of sandy sediment than those in areas of Potsdam Sandstone.

LAKE VERMONT

The deposits and shoreline features of Lake Vermont in the lowlands from the Ausable River to the Canadian border record only the latter part of the history of this glacial lake. During most of the Coveville stage, the Plattsburgh area was beneath the ice sheet.

COVEVILLE STAGE

The history of the Coveville stage in the northeast Adirondack region opened with the deposition of the large delta in the Ausable River valley west of Clintonville during the Owls Head-Redford stand (ice-front position 2, pl. 1). The original volume of this delta is estimated at 352 million cubic yards (table 3). This figure is several times larger than that for the Saranac River delta (48 million cubic yards), although the Saranac is a somewhat longer stream than the Ausable. The Ausable River delta of Coveville age is also considerably larger than that of the younger Fort Ann stage near Keeseville (261 million cubic yards). The explanation for the large volume

TABLE 3.—*Upper Pleistocene and Holocene deltas of the northeast Adirondack region, New York*—Continued

	Deltas (estimated volumes in millions of cubic yards)							Rate of delta formation in Champlain Sea (assuming that it lasted about 1,500 years)	
	Fort Ann stage			Champlain Sea			Lake Champlain	Unit rate (acre ft per square mile of drainage basin per yr)	Rate (millions of cu yd per yr)
	Original	Existing remnants	Loss by erosion	Original	Existing remnants	Loss by erosion	Original		
Salmon and Trout (Franklin County)	529.0	441.0	88.0	382	---	---	---	0.692	0.255
Trout	----	----	---	---	---	---	---	----	----
Little Trout	6.5	----	---	---	---	---	---	----	----
Chateaugay	29.0	----	---	---	---	---	---	----	----
North Branch Great Chazy	10.4	5.7	4.7	---	---	---	---	----	----
Park Brook	0.6	----	---	---	---	---	---	----	----
Great Chazy	3.3	----	---	---	---	---	---	----	----
Great Chazy and North Branch	----	----	---	80	---	---	---	.169	.053
Saranac	95.0	70.0	25.0	335	291	44	---	.235	.223
Salmon (Clinton County)	77.0	53.0	24.0	100	---	---	---	1.120	.067
Little Ausable	----	----	---	118	---	---	---	1.351	----
Ausable	261.0	207.0	54.0	446	239	207	[4]212	.367	.297

[4] Includes both the delta at mouth of river and the submerged delta near Wickham Marsh.

of the delta near Clintonville during Coveville time is probably that the Ausable was augmented by overflow from the Saranac drainage basin.

The ice dam at the north end of Lake Vermont blocked the Champlain Valley near Keeseville, and the delta was built into an arm of the lake extending upvalley to Clintonville. The ice front gradually retreated north during late Coveville time to the vicinity of the Saranac River. The time of retreat corresponds to that involved in the Trout River-Moffitsville stand, the Malone-Schuyler Falls stand, and much of the building of the moraine near Cadyville.

Only one or two small beaches were formed during Coveville time in the area between the Ausable and Saranac Rivers; perhaps at this latitude the Champlain Valley was largely ice filled and only a narrow arm of the Coveville lake extended north between the ice and the edge of the uplands. During the retreat of the ice front from the Ausable to the Saranac, ice-marginal drainage flowed south along the ice edge where it rested against the uplands, cut channels in drift and in bedrock, removed the drift from large areas, and discharged into the lake, as described earlier under the Cadyville episode of deglaciation. The surface of the deltas at the mouths of the Salmon (Clinton County) and Saranac Rivers is irregular, suggesting collapse and slumping after adjacent glacial ice had disappeared. That part of the Coveville stage which is represented by shore features in the Plattsburgh area probably lasted less than a hundred years.

The delta east of Peasleeville at the mouth of the Salmon River (Clinton County) is large (141 million cubic yards) in relation to the length and size of the Salmon River (fig. 25). Perhaps the delta deposits came from the erosion of the large sandy kames in the valley upstream. The delta on the Saranac River east of Cadyville is small (48 million cubic yards), only about a third the size of the delta along the nearby and very much shorter Salmon River. The explanation may be that when the level of Lake Vermont dropped to that of the Fort Ann stage, the ice dam in the Champlain Valley had retreated to a line only a few miles north of the Saranac River. Thus, the river perhaps had time to build only a small delta into the Coveville lake.

FORT ANN STAGE

The level of Lake Vermont dropped nearly 100 feet, perhaps more than 150 feet, to that of the Fort Ann stage when the ice dam was a short distance north of the Saranac River, perhaps about at ice-front position 8 (pl. 1), that is, during the building of the moraine near Ellenburg Depot. Deltas began to be built at the lower lake level by the Ausable, Salmon, and Saranac Rivers, in part of material derived from older Coveville-stage deltas upstream.

Along the Ausable River, the original volume of the Fort Ann delta at Keeseville is estimated at about 261 million cubic yards. Of this, 122 million cubic yards probably came from the erosion of the Coveville delta upstream near Clintonville (table 3), leaving about 139 million cubic yards of "new" sediment to come from erosion in the drainage basin or

FIGURE 19.—Cobble and boulder gravel in Champlain Sea delta of the Saranac River. Massive topset beds exposed in borrow pit at apex of delta just east of Morrisonville. Clasts are subangular to slightly rounded; sand lenses suggest a faint horizontal stratification.

from wasting glacial ice. The volume of "new" sediment is about 2.5 times less than that deposited by the Ausable River in Coveville time. Presumably this smaller amount of sediment reflects a decrease in streamflow (no contribution from the Saranac drainage basin) and in volume of melting ice upstream. It is not clear whether this decrease in amount of sediment argues for or against Craft's (1969) suggestion that glacial ice was still present in the high peaks region during Lake Vermont time.

The original volume of the Fort Ann delta of the Saranac River, 95 million cubic yards, is nearly twice that of its Coveville-stage delta. Both are smaller than the Fort Ann delta of the Ausable River. Because the volume of melting ice in the drainage basin was probably decreasing, perhaps to zero, in Fort Ann time, the increase in size of the Saranac delta suggests deposition over a much longer period of time; that is, the Fort Ann stage lasted twice and perhaps several times as long as that time interval during which the Saranac River built its Coveville delta. The Saranac River delta is only slightly larger than that of the Salmon River (Clinton County). It is not clear why the Lake Vermont deltas of these two rivers, so different in length, are so similar in size.

Lake level may have risen during a part of Fort Ann time. The presence of Fort Ann beaches on Flat Rock, as mentioned earlier, suggests a rise in lake level after the bedrock had been cleaned off. Along the Ausable River the top of the Fort Ann delta, at Keeseville, is about 60 feet below the highest stand of the Fort Ann stage (Denny, 1967); this delta was built, therefore, when lake level was below the maximum level of the Fort Ann stage. Along the Salmon

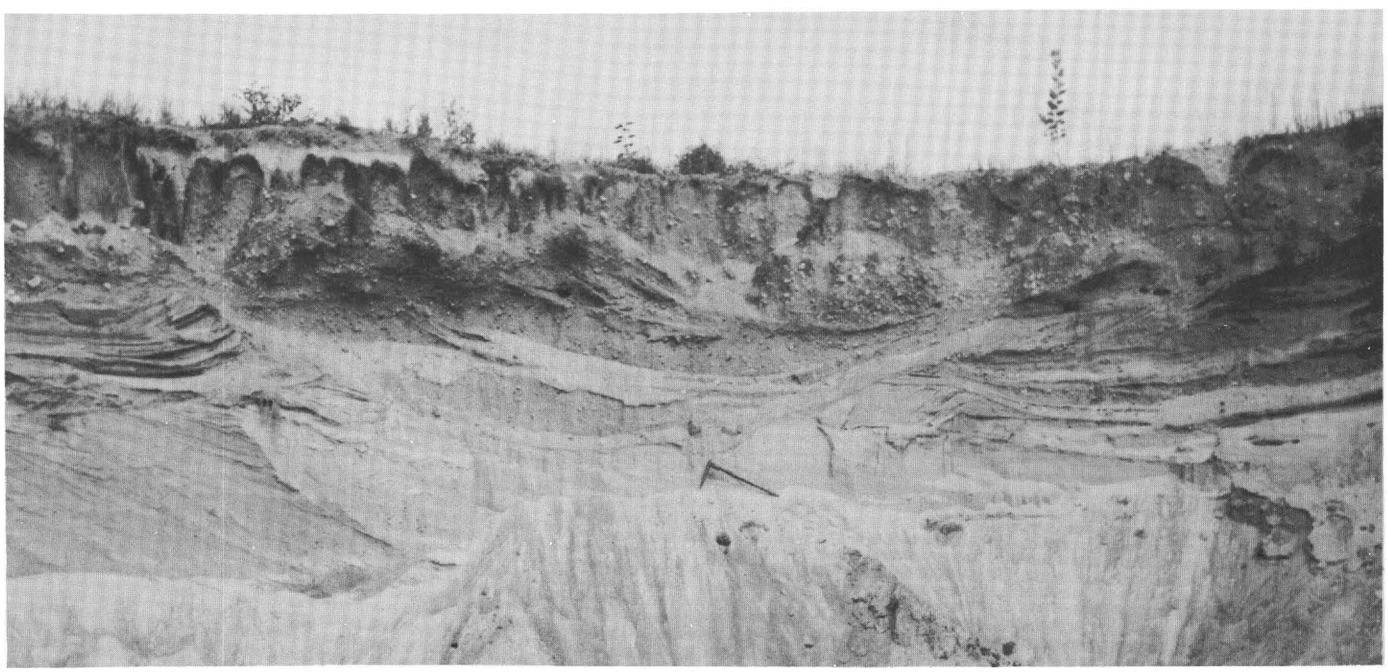


FIGURE 20.—Crossbedded pebbly sand capped by pebble to cobble gravel. Probably topset beds in Fort Ann delta of the Great Chazy River at Altona. Prominent soil tongues at top of bank.

and Saranac Rivers the relation of the top of the Fort Ann deltas to the highest lake stand is not clearly defined. The northward retreat of the ice front in the Champlain Valley (from ice-position 8) uncovered the lower course of the Great Chazy River and its North Branch, and they began to build small deltas into the glacial lake. The tops of these deltas are also about 50 feet below the highest stand of the Fort Ann stage.

Delta building by the Ausable and Great Chazy Rivers could have taken place during either rising or falling lake level or both. It is easier, however, to account for the absence of delta deposits at the maximum stand and for the occurrence of beaches on Flat Rock by assuming that delta formation was largely completed before a 50–80-foot rise in lake level near the end of Fort Ann time.

In profile, the Fort Ann beaches form a belt 50–75 feet wide, the top of which marks the maximum stand of the stage (Denny, 1967, 1970). Whether the beaches in the lower part of the belt were formed before the rise in lake level or during the subsequent fall from the highest stand is unknown. Presumably the rise was in response to a rise of the outlet of Lake Vermont near Fort Ann, N.Y.

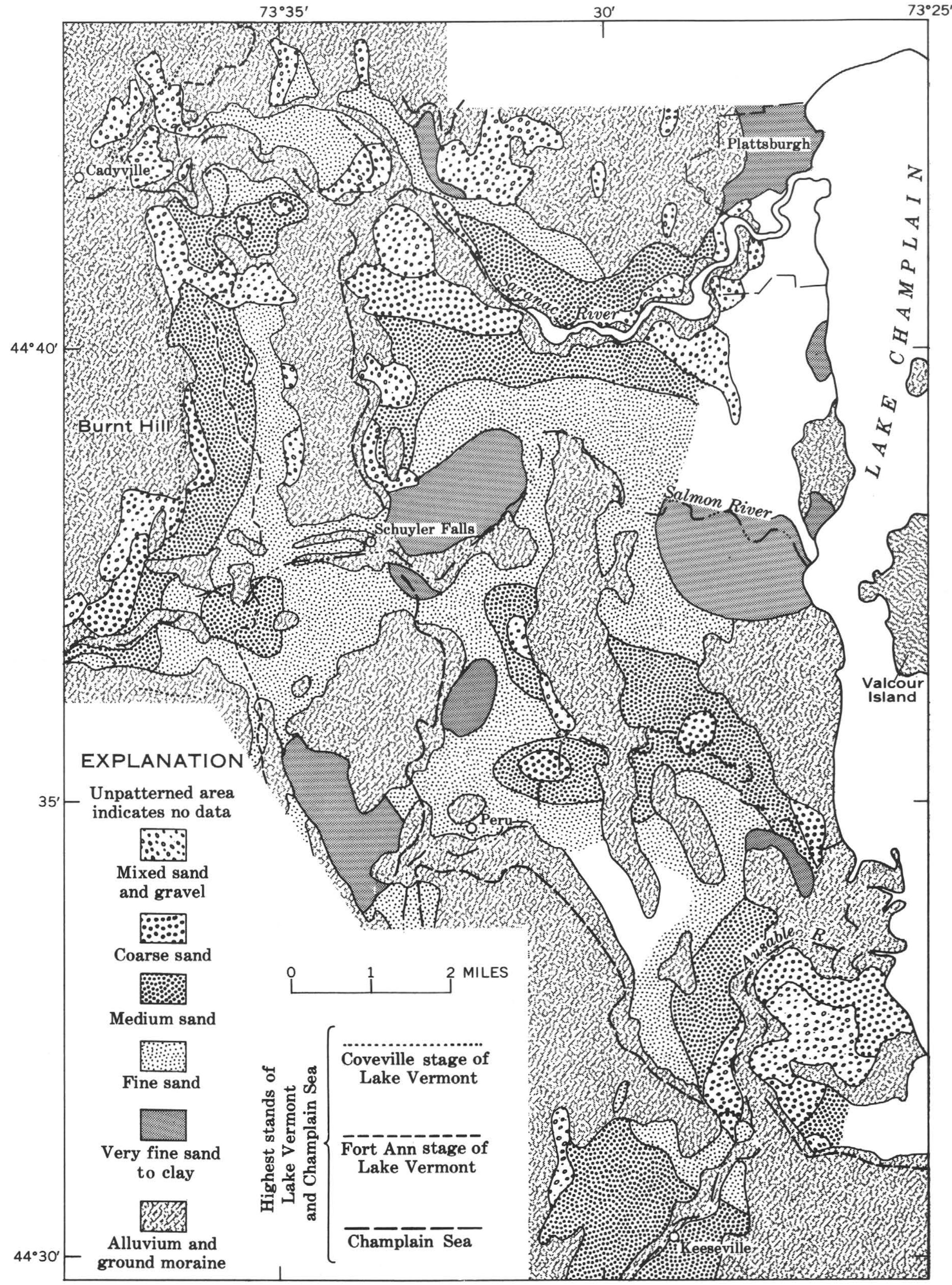
The extensive beaches at the maximum stand of the Fort Ann stage near Cannon Corners were formed of material washed off the adjacent areas of bare rock along English River (pl. 6). Perhaps the rise in the Fort Ann lake took place during the washing of the drift from the bare-rock areas near English River.

In Franklin County, during Fort Ann time, deltas formed at the mouths of the Salmon, Trout, and Chateaugay Rivers where they entered glacial Lake Iroquois (pl. 1), which drained east by way of the gorge near Covey Hill. In time, the ice front retreated to the northeast off Covey Hill, and the level of Lake Iroquois dropped to that of Lake Vermont in the Champlain Valley.

During the lowering of the Fort Ann lake prior to the incursion of the sea, there may have been a short interval when the lake in the Champlain Valley drained northeast between the retreating ice front and hills southeast of the St. Lawrence River (Wagner, 1969). Ultimately, drainage was opened to the Atlantic Ocean by way of the St. Lawrence Lowlands northeast of Montreal, and marine waters invaded the Champlain Valley. This invasion dates from about 12,000 years B.P. (McDonald, 1968; Prest, 1970).

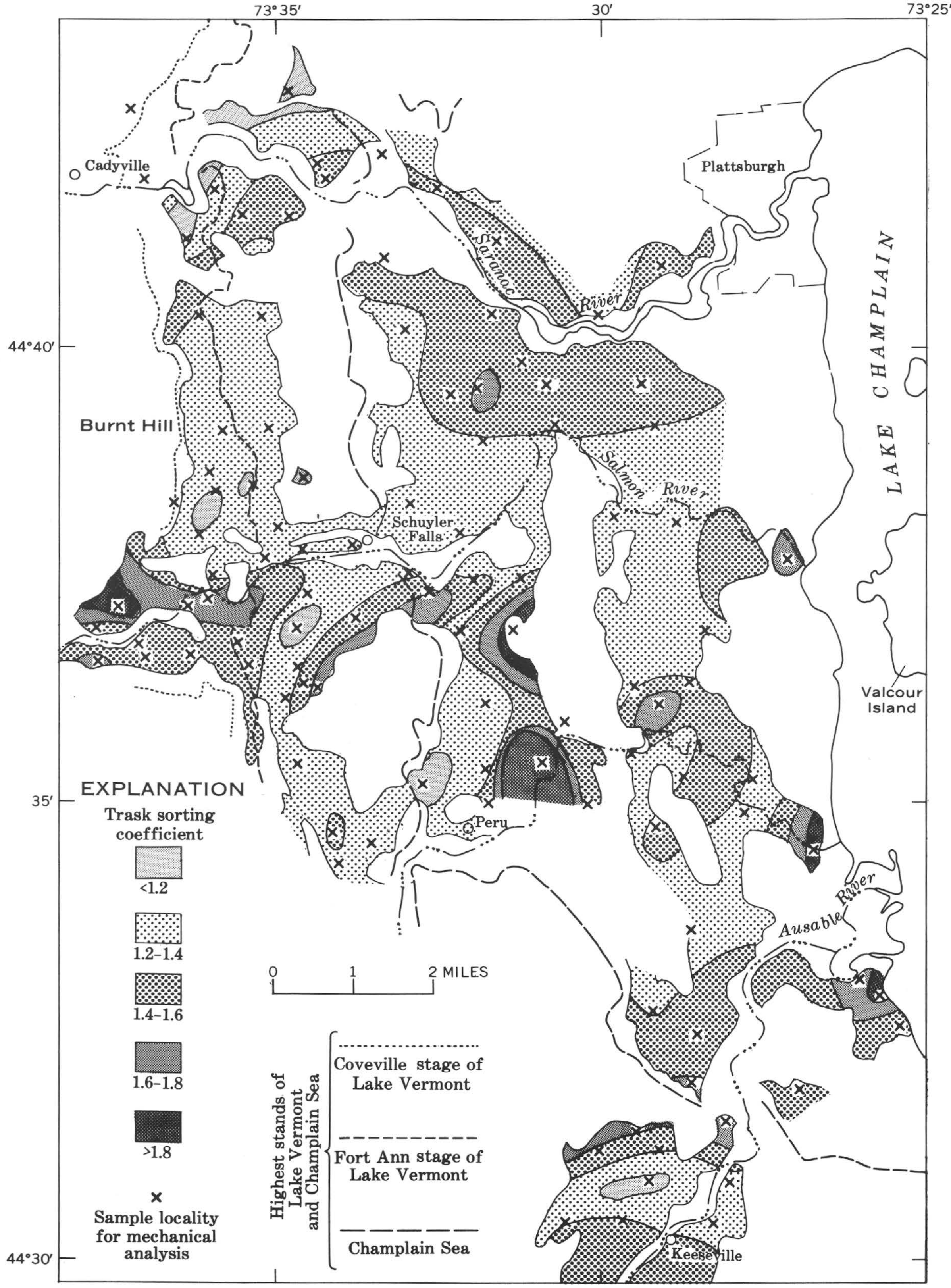
The duration of the Fort Ann stage is unknown, but it probably was only a few hundred years. Prest (1970), in his reconstruction of glacial lake phases in the Great Lakes region and the St. Lawrence Lowlands, shows the Fort Ann stage beginning about

FIGURE 21.—Foreset beds of medium well-sorted sand in delta of the Champlain Sea about 1 mile southwest of Port Kent. Exposure is about 15 feet high. Top of bank has been stripped.



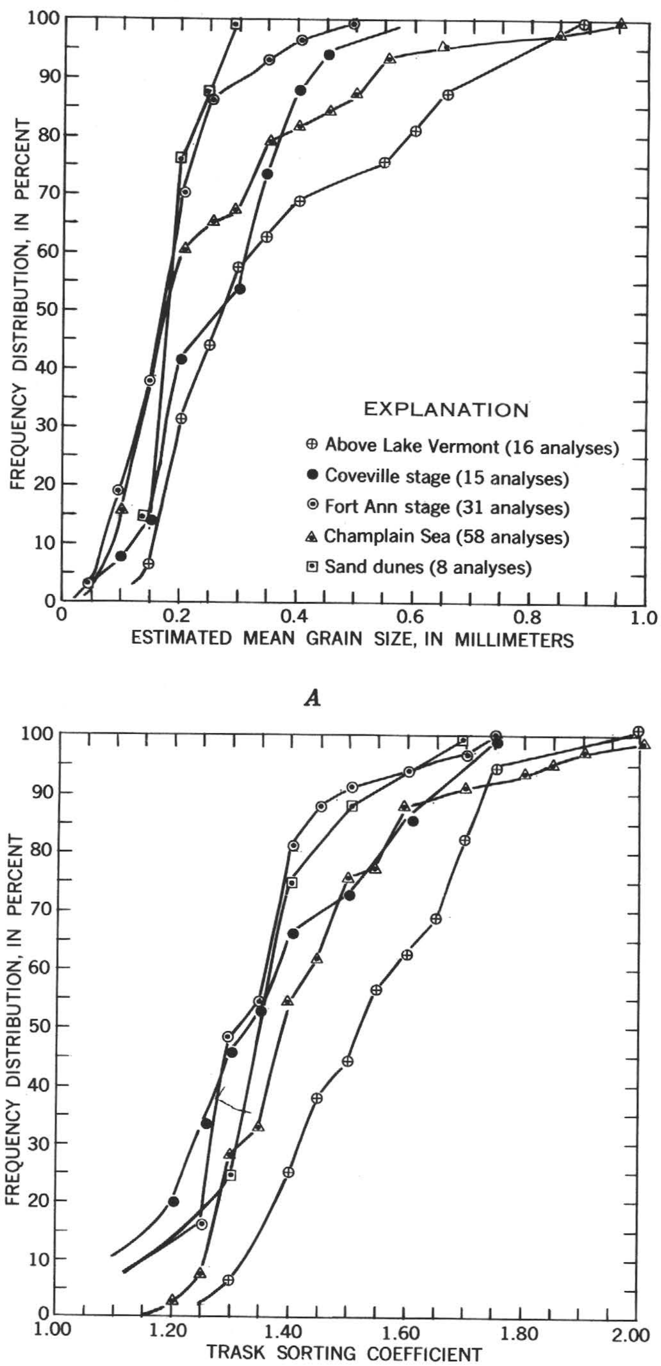
A

FIGURE 22.—Grain size and sorting of late-glacial lake, stream, and marine deposits in the Plattsburgh clay from Denny (1967, 1970). *A*, Areal distribution of



B

area. Based on rapid mechanical analysis of sand made in the field. Distribution of gravel and silt and textural classes. *B*, Areal distribution of Trask sorting coefficient.



B

FIGURE 23.—Grain size and sorting of late-glacial sands. Cumulative curves showing the range in estimated mean size (*A*) and in Trask sorting coefficient (*B*) of sands of Lake Vermont and the Champlain Sea. Data from a few analyses of dune sand and of glacial sand above the highest stand of Lake Vermont are included for comparison.

12,500 to 12,400 years B.P. (Prest, 1970, fig. XII–16f). The ice sheet left Covey Hill, and the lakes in the St. Lawrence and Champlain Valleys became confluent shortly before 12,000 years B.P. (Belleville-Fort Ann phase, Prest, 1970, fig. XII–16h). The Champlain Sea invaded the St. Lawrence Lowlands and the Champlain Valley by about 12,000 years B.P. (Prest, 1970, fig. XII–16i).

CHAMPLAIN SEA

The deposits of the Champlain Sea are indistinguishable lithogically from those of Lake Vermont. The marine deposits in places contain the remains of salt- or brackish-water organisms, largely mollusks. Localities where fossils have been found are shown on the published geologic quadrangle maps (Denny, 1967, 1970). Plate 1 shows that the marine deposits form a belt at lower altitudes than those of Lake Vermont. The principal streams built deltas into the Champlain Sea. Delta building continued along most of the large rivers until gradual rise of the land caused the sea to drain down to the level of present-day Lake Champlain. The Great Chazy River is an exception because it ceased to build a delta when sea level had dropped to a point about 100 feet above the present level of Lake Champlain.

North of the Saranac River, beaches are common near the marine limit. Along the Saranac River north of Morrisonville and along the Salmon River (Franklin County) at Malone, the marine limit is represented by steep bluffs cut in the sand of the older Fort Ann delta by wave action (MacClintock and Stewart, 1965, fig. 22b). These cliffs resemble the present-day bluff along the shore of Lake Champlain just south of the mouth of the Ausable River. In the area north of the Saranac River, beaches continued to form during the gradual lowering of the sea level until the shoreline reached a point about 200 feet above the modern lake.

AGE

The Champlain Sea lasted for about 1,500 years. It came into existence about 12,000 years B.P. This is the figure given by McDonald (1968) and by Elson (1969). Mott (1968) listed 27 Champlain Sea radiocarbon dates; 25 of them are from shell material. The oldest date, 11,800± 160–180 years B.P., was measured on samples from two localities (Geol. Survey Canada locs. GSC–505 and GSC–588). Prest, in his reconstruction, gives a range of 12,000 to 11,800 years B.P. for the sea's birthday (Prest, 1970, fig. XII–16h, i). The Champlain Sea withdrew from the Champlain Valley about 10,500 years B.P., perhaps as late as 10,000 years B.P. Elson (1969) gives a figure of 10,000 years B.P. for the close of the Champlain Sea episode in Quebec Province, and Prest

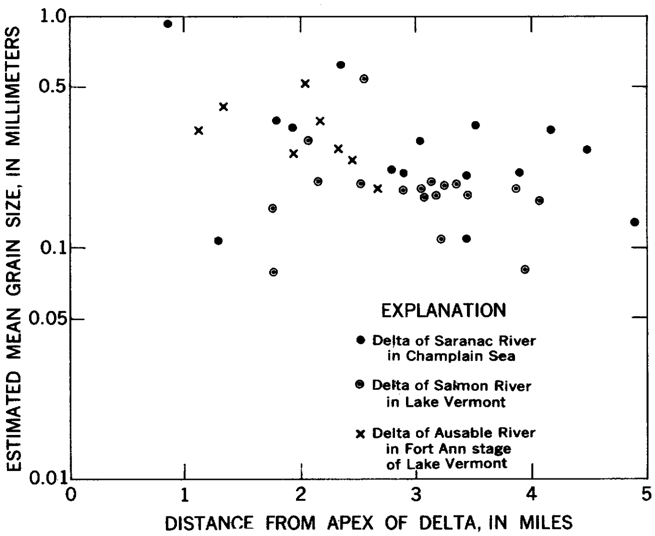


FIGURE 24.—Grain size of delta sand and distance from apex. Semilogarithmic scatter diagram showing the relation between the estimated mean size of sand at sample sites on ancient deltas of the Saranac, Salmon, and Ausable Rivers and the distance of these sites from apex of delta.

places the withdrawal of the sea from the Champlain Valley between 10,300 and 10,000 years B.P. Of the 27 Champlain Sea radiocarbon dates listed by Mott (1968), only five are less than 10,500 years B.P. (10,200 to 10,450).

A sample of shells from Champlain Sea deposits exposed in a borrow pit about 1.5 miles west of Chazy gave a radiocarbon age of 10,560 $\pm$ 350 years B.P. (W–1109, Ives and others, 1964). The sample came from a pebbly sand at the base of about 10 feet of boulder and cobble gravel on the crest of the broad ridge named the Ingraham esker by Woodworth (1905a). Presumably these shells, largely mollusks, lived on the ridge during the decline in sea level. The shell locality is at an altitude of about 210 feet, nearly 250 feet below the marine limit.

RATE OF DELTA FORMATION

Using the estimates of volume of the Champlain Sea deltas and assuming that the embayment lasted for about 1,500 years, it is possible to calculate rates of delta formation for several rivers emptying into the Champlain Sea. The rates range from 53,000 cubic yards per year for the Great Chazy River and North Branch to 297,000 cubic yards per year for the Ausable River (table 3). In terms of the size of the drainage area upstream from the delta, the rate for the Great Chazy and North Branch is 0.169 acre-feet per square mile per year and for the Ausable River 0.367 acre-feet per square mile per year.

ENVIRONMENT

When the sedimentation rates have been calculated, what do they mean in terms of the environment of Champlain Sea time? Was the environment strictly periglacial, a time of increased bedload in the streams because of strong frost action and (or) sparse vegetation on slopes, and of increased streamflow because of ice melting or high precipitation? Or was the environment nonglacial, much like the present?

Schumm and Hadley (1961) found a good correlation between the rate of sedimentation in small reservoirs in the western United States and the relative relief of the drainage basin supplying the sediment. To compare the rate of formation of the Champlain Sea deltas with the rate of sedimentation in the small reservoirs, I have plotted (fig. 26) rate of delta formation against the relative relief of the drainage basin. In this comparison, I ignored the suspended load carried by the rivers into the deeper parts of the sea. The rate of delta formation, expressed in acre-feet per square mile per year, increases with the relative relief of the drainage basin, expressed as maximum relief divided by main stream length (table 3), in the same manner but at a slightly faster rate than does the sedimentation rate in the small drainage basins of the western United States (Wyoming to Arizona). The increase in sedimentation rate with relative relief is a common feature of drainage basins and is probably related to increased average slope. The deltas of the smaller streams have the higher rates and higher average slopes, and the unit rate of delta formation decreases with increase in drainage area of delta-building stream (fig. 27). In both examples, the unit rate of delta formation for the Great Chazy and North Branch is small, suggesting again the influence of the Potsdam Sandstone on the size and, therefore, on the rate of formation of the deltas along the two streams.

The comparison between the Champlain Sea deltas and the sediment in the small reservoirs suggests that delta formation was conditioned by the same

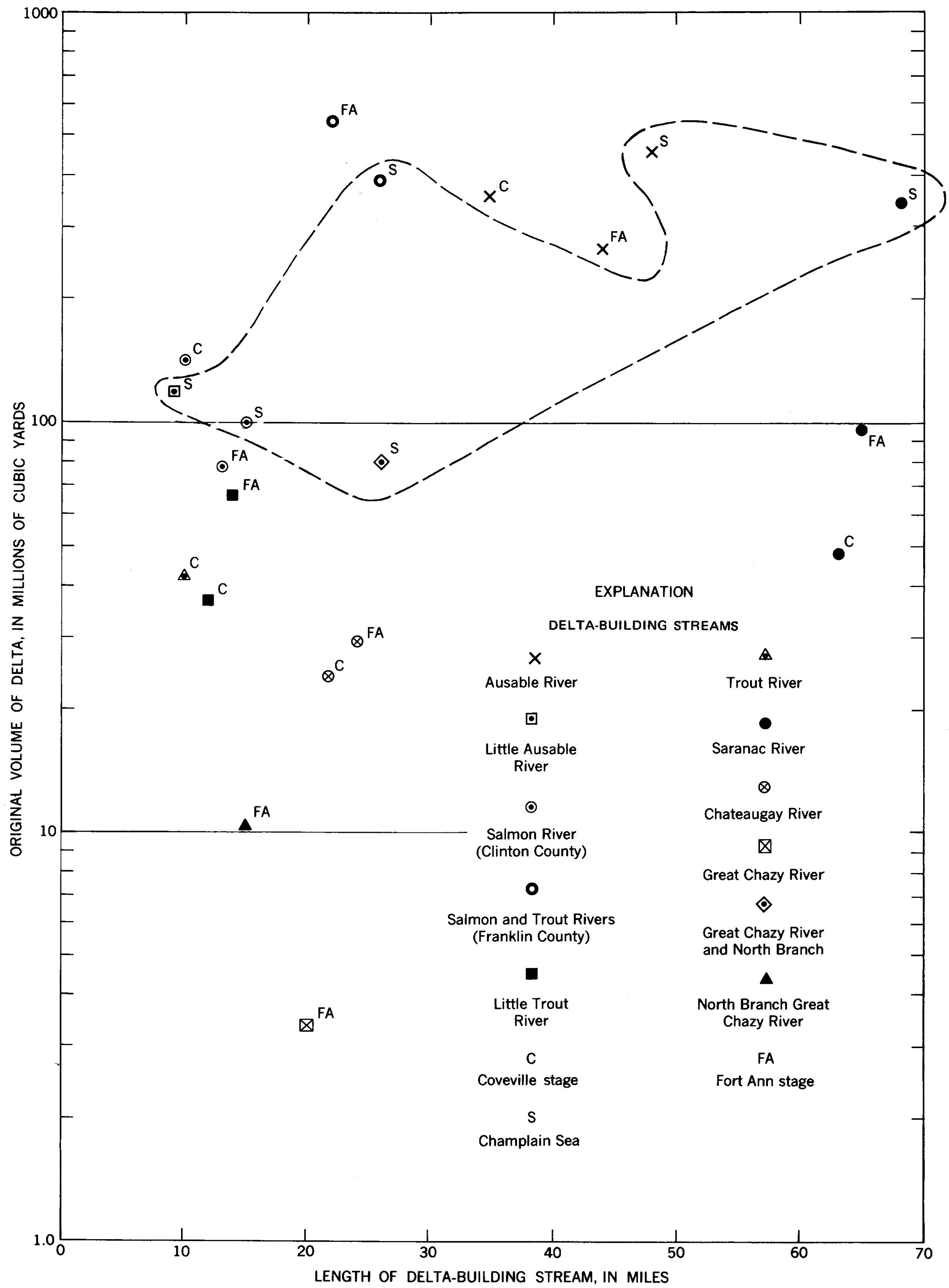


FIGURE 25.—Upper Pleistocene deltas and their source areas. Semilogarithmic scatter diagram showing the relation between the original volume of a delta and the length of the delta-building stream. Deltas of the Champlain Sea are surrounded by a dashed line.

variables that control modern sedimentation in parts of western United States rather than by climatic factors operating during deglaciation. Insofar as such a comparison is valid, it also suggests that the environment of the Champlain Sea episode was more nonglacial than periglacial. But was the environment much like that of the present? The modern delta of the Ausable River suggests that sedimentation rates were much greater in Champlain Sea time than they are at present.

AUSABLE RIVER DELTA IN LAKE CHAMPLAIN

The Ausable River has built a crow's-foot delta in Lake Champlain (fig. 28). If the lake has been in existence for nearly 10,000 years, an estimate of the volume of the modern delta will yield a rate of delta formation for post-Champlain Sea (postglacial) time. On the basis of depth curves and soundings, the volume of the modern delta is estimated at about 212 million cubic yards (table 3). The rate of delta formation would thus be about 0.021 million cubic yards per year (0.026 acre-feet per square mile per year), suggesting that the bedload of the Ausable River is now very much less than it was in Champlain Sea time (figs. 26, 27).

Delta building during Champlain Sea time may well have been much greater than at present. However, the size of the modern Ausable River delta is not dependent on the length of post-Champlain Sea time; rather, the delta appears to be a young and unique feature. Several large streams empty into Lake Champlain: the Great Chazy, Saranac, Ausable, Winooski, Lamoille, and Missiquoi Rivers, yet the Ausable is the only one that has built a crow's-foot delta graded to the modern lake. Even the Saranac River has no conspicuous delta at its mouth, in spite of the fact that the topography and geology of its drainage basin are similar to those of the Ausable River (table 3), and even though it discharges into Cumberland Bay, a shallow arm of Lake Champlain, whereas the Ausable River discharges into deep water.

The modern delta of the Ausable River in Lake Champlain was not formed until the river's course had been changed by piracy. The piracy could have taken place several thousand years after Lake Champlain came into existence. Thus, the modern delta may have been built rapidly.

The history of the delta is interpreted to be as follows. The Ausable River first built a delta into the Champlain Sea when the sea stood near its highest

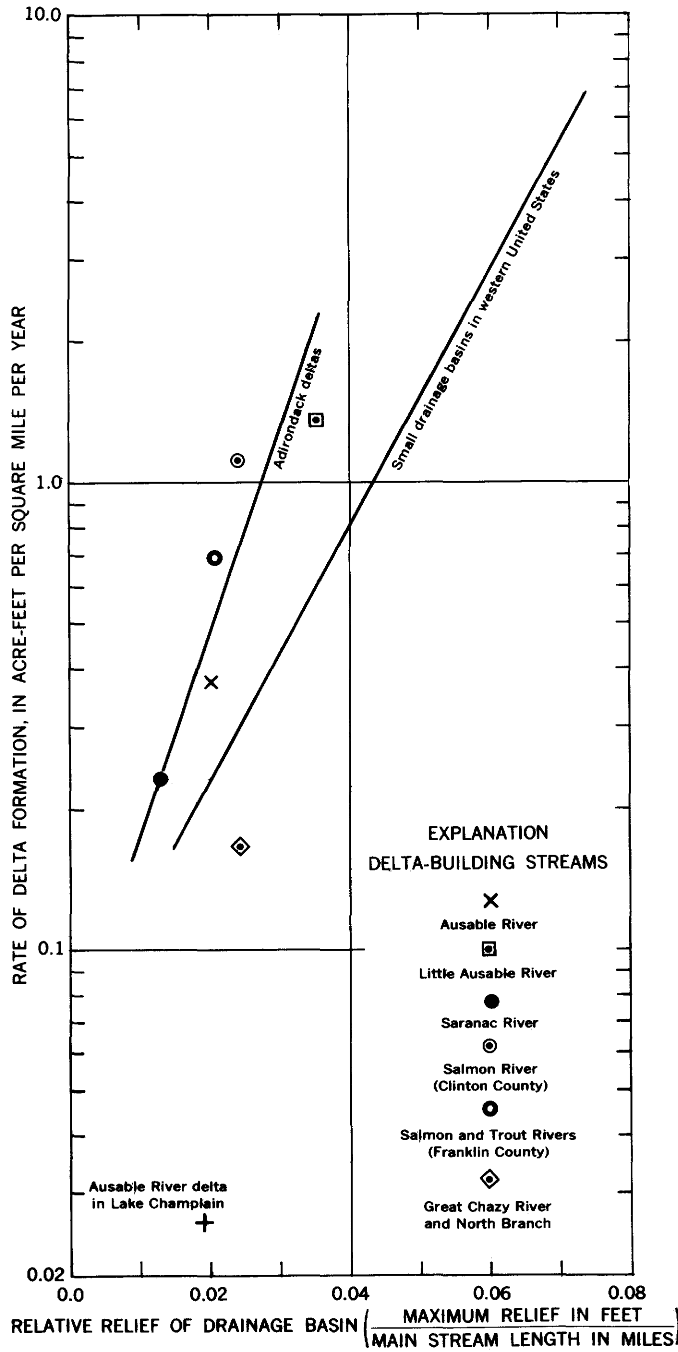


FIGURE 26.—Rate of formation of Champlain Sea deltas and relief of drainage basins. Semi-logarithmic scatter diagram showing the relation between the rate of delta building and relative relief of the drainage basin above apex of each delta. Champlain Sea assumed to have lasted about 1,500 years. Line representing a generalization from data on mean annual sediment yield for small drainage basins in western United States shown for comparison (from Schumm and Hadley, 1961, fig. 1). Point representing rate of formation of Ausable River delta in Lake Champlain assumes that the lake has been in existence for about 10,000 years.

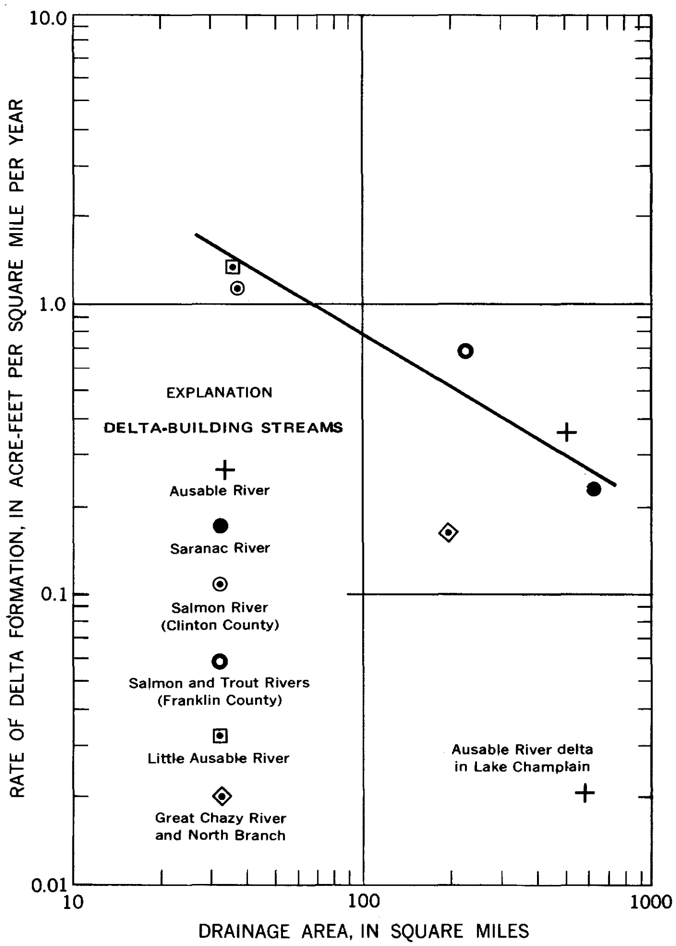


FIGURE 27.—Rate of formation of Champlain Sea deltas and size of drainage basins. Logarithmic scatter diagram showing the relation between the rate of delta formation per square mile of drainage area and the drainage area upstream from the delta. Champlain Sea is assumed to have lasted about 1,500 years. Point representing Ausable River delta in Lake Champlain assumes that the lake has been in existence for about 10,000 years.

point at an altitude of about 360 feet (fig. 28). Ausable Chasm was not yet in existence, for its rim ranges in altitude from about 250 to 350 feet. The surface of the delta above the chasm slopes toward the lake to an altitude of about 260 feet.

Sea level gradually lowered. The river began to cut the upper part of the chasm and to erode part of its older delta. A new delta was built, and the chasm was eroded to a depth of as much as 60 feet. The new delta ranged in altitude from about 260 feet at its apex to about 180 feet near the present shoreline. The deposits of the younger delta overlap those of the older one; the two can be distinguished only on the basis of their topographic position and form.

A further drop in water level, down to the present lake level or below, caused the river to cut down to an altitude of about 200 feet, where erosion was checked for a time by a bedrock sill about half a mile east of the mouth of the chasm (point W, fig. 28). Upstream from the sill, the river developed a meandering course and by cut and fill spread coarse sand and gravel on top of delta sand. Remnants of this meandering channel are still preserved upstream from the rock sill. At this time, the floor of the chasm was cut down to within about 60 feet of its present level.

The Champlain Sea drained from the Champlain Valley, and the modern lake came into existence. The Ausable River did not at once lower its channel to the new lake level because the river was held up by the rock sill (point W). Downstream from the sill, the river eroded a deep channel in the delta deposits, forming a series of rapids on bedrock, and flowed into a basin about half a mile wide, now occupied by Wickham Marsh. A delta was built east of the marsh out into the lake. The outline of this delta below the surface of the lake is shown by the 12-foot-depth curve in figure 28. Presumably, the now-submerged delta near Wickham Marsh is built primarily of material eroded by the river downstream from the rock sill.

The rate of sedimentation on the delta near Wickham Marsh decreased once the rapids below the rock sill had been formed. Waves and currents began to erode the lower end of the older Champlain Sea delta deposits and form a cliff that today is about 100 feet high and is perhaps a quarter of a mile west of its original position. The modern delta at the river's mouth laps against the northern part of this cliff (near point X, fig. 28) and therefore that part of the modern delta is assumed to be younger than the cliff. Wave erosion formed the cliff and destroyed all of the subaerial part of the delta east of Wickham Marsh. The time required to erode this delta and form the cliff along the shore cannot be estimated on the basis of available information. Certainly some of the material was removed after the Ausable River had assumed its present course into the lake, and this diversion must have taken place before peat began to accumulate in Wickham Marsh.

The diversion of the Ausable River from its abandoned channel at an altitude of about 190 feet to its present course was caused by piracy. Dry Mill Brook joins the Ausable River from the north at the head of the modern delta. The brook probably lowered its bed rapidly when Lake Champlain came into existence. A gully on the south side of the brook, now destroyed, cut into the deposits of the Champlain Sea

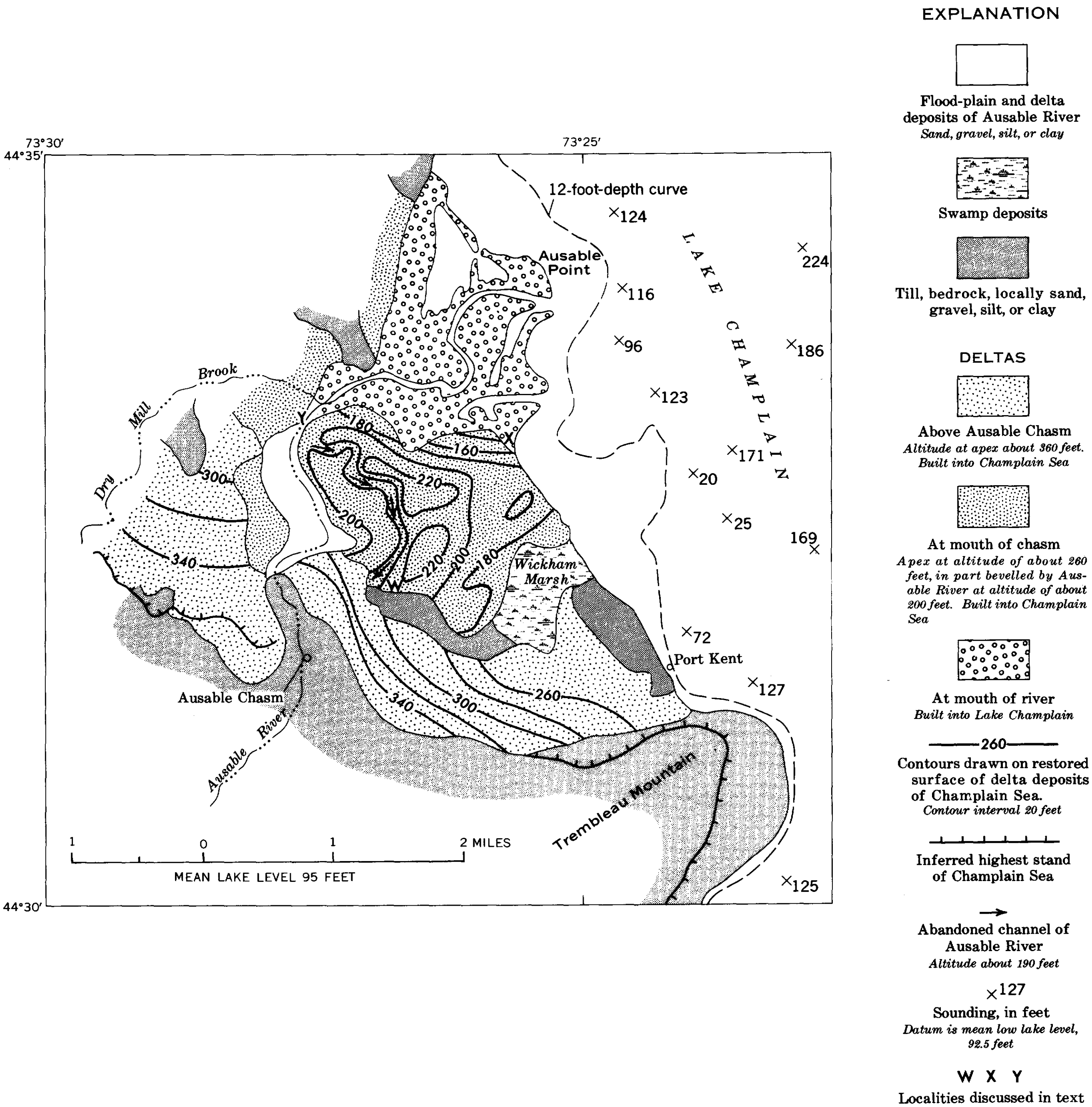


FIGURE 28.—Ausable River delta in Lake Champlain. Geology generalized from Denny (1967). Base from U.S. Geological Survey Plattsburgh 15-minute quadrangle, 1956.

delta near the mouth of the chasm. As the gully deepened, its head approached the bank of the Ausable River (near point Y, fig. 28), where the riverbed would have been about 100 feet above the mouth of the gully.

The river, perhaps in time of flood, overtopped its north bank (near point Y) and flowed north down into the gully and down Dry Mill Brook to the lake. The river was then diverted into Dry Mill Brook and cut down rapidly into the underlying sandy deposits, as there was no rock sill at the point of capture to hinder erosion. The channel above Wickham Marsh was abandoned.

The river cut down rapidly into the older deltaic deposits, and the chasm has been gradually deepened to its present level. The valley at the mouth of the chasm is about 60–80 feet deep and about a quarter of a mile wide. The material removed in the erosion of this valley was swept into the lake to form the modern delta. The volume of material removed is estimated to be about 144 million cubic yards, and the volume of the modern delta at the mouth of the river, about 133 million cubic yards.

The delta of the Ausable River in Lake Champlain is unique. It owes its development to a stream diversion that caused rapid, perhaps catastrophic, erosion of older deltaic deposits and deposition of sand in the lake, in part on top of older delta sands. Near Ausable Point, the shore of the delta is now being eroded back (westward), and sand is piled by storm waves among the trees along the shore.

This analysis of the modern delta of the Ausable River and the absence of similar deltas along the other large rivers emptying into Lake Champlain suggest that sedimentation during the Champlain Sea episode was greater than at present. Although the marine deltas appear to be the result of runoff from the drainage basin upstream, the rate of sediment (bedload) transport was considerably greater than now. The environmental differences that caused the change in rate of sedimentation are unknown. It is reasonable to suppose that the climate was still cooler than the present and that the forests had not yet completely covered the landscape. The fossil record supports this interpretation (Terasmae, 1959; Davis, 1965; LaSalle, 1966; Terasmae and LaSalle, 1968). For New England, Davis suggested that during the interval 12,000–10,500 years B.P. ". . . the vegetation may have resembled park-tundra, or, alternatively, spruce-oak woodland similar to modern vegetation near the prairie in Manitoba" (Davis, 1967, p. 11). The fossil assemblage in the Champlain Sea deposits that overlie the Ingraham esker (Woodworth, 1905a) suggest shallow cold marine water (Denny, 1972). J. E. Hazel (written commun., 1971) feels that the ostracode assemblage indicates subfrigid or frigid climatic conditions (that is, averaging about 10°C or colder in shallow water during the warmest month).

TILT OF WATER PLANES

The shorelines of the late-glacial water bodies in the Champlain Valley rise northward (Chapman, 1937; Farrand and Gajda, 1962; Wagner, 1969). In the northern part of the valley on the New York State side near the Ausable River (Denny, 1967, 1970), the Lake Vermont deposits range in altitude from about 490 feet to 670 feet, whereas near the Saranac River the range is from about 560 feet to 720 feet. The upper limit of the marine deposits ranges from about 360 feet near the Ausable River to about 510 feet near the International Boundary. The beaches and deltas of these two belts were formed in extensive water bodies, and the northward rise demonstrates that the region has been warped upward toward the north since the shore features were formed.

As mentioned earlier, the correlation of the shore features throughout the Champlain basin is largely based on their location and altitude as shown on north-south profiles. The profiles for the areas mapped in detail show a line representing the inferred position of the highest stand of the Fort Ann stage of Lake Vermont and of the highest stand of the Champlain Sea. The inclination of these inferred water planes amounts to about 4.5 feet per mile for the top of the Fort Ann stage and about 4.9 feet per mile for the top of the Champlain Sea. Theoretically, the tilt of a marine-water plane cannot be greater than that of an older lake plane and might well be less. Thus, these values of tilt have a possible error of at least 0.2 foot per mile.

The north-south profiles for the New York side of the northern Champlain basin furnish only a two-dimensional picture of the tilt. A three-dimensional model can be constructed by comparing the north-south profile showing the inferred water planes for New York with a similar north-south profile for Vermont (W. P. Wagner, written commun., 1970). A point at a given altitude on the New York profile is several miles south of the latitudinal position of a point at the same altitude on the Vermont profile. It is a simple problem in plane geometry to determine that the water planes have been tilted upward in a direction about N. 10°–15° W. along an axis trending at right angles thereto.

The 10°–15° deviation of the true direction of tilt from the north-south orientation of the published profiles (Denny, 1967, 1970) increases the amount of tilt by only about 0.1–0.2 foot per mile, probably within the error of measurement of the individual shore features.

SUMMARY AND CORRELATION

NORTHEAST ADIRONDACK REGION

The late-glacial history of the Northeast Adirondack region began with the removal of the ice sheet from the southwest corner of the region and the building of massive outwash plains by southwest-flowing streams (ice-front position 1, pl. 1). There is no way to date this episode, but it probably followed the Luzerne readvance of the Hudson-Champlain ice lobe on the southeast side of the Adirondacks (fig. 14). The Luzerne readvance, originally described by Woodworth (1905b, p. 139), has been dated by Connally and Sirkin (1971) at about 13,200 years B.P. These authors believe that this readvance antedates the Quaker Springs and Coveville stages of Lake Vermont that were coextensive with lakes in the Hudson Valley (LaFleur, 1965). The Bridport readvance (Connally, 1970) in the southern part of the Champlain basin entered into the Coveville-stage lake.

The late-glacial history of the northeast Adirondack region can be said to end with the invasion by the Champlain Sea about 12,000 years B.P. I assume that the carbon-14 dates for the Champlain Sea deposits are essentially correct. Although they are based largely on pelecypod shells, algae material (seaweed) from marine deposits near Ottawa gives a carbon-14 date consistent with that of shells found immediately above and below the algae material (Mott, 1968).

There are no carbon-14 dates from which to estimate the age or duration of any episode or stand in the northeast Adirondack region. The ages given here (table 1) are based merely on a uniform rate of ice-front retreat of about 600 feet per year, from the line of the Luzerne readvance north to a point in the St. Lawrence Lowlands about 25 miles south of Montreal, assumed to be equivalent to the Drummondville moraine (LaSalle, 1966, p. 98, fig. 3). Schafer (1968) suggested an average rate of retreat for northern New England of at least 1,000 feet per year. My reconstruction (fig. 14) makes the Loon Lake episode (ice-front position 1) contemporaneous with the Highland Front moraine (Gadd, 1964).

Many workers believe that Lake Iroquois, in the Ontario basin, came into existence about 12,100–12,000 years B.P. (Karrow and others, 1961; Goldthwait and others, 1965; Calkin, 1970) and that it drained about 11,000 years B.P. Prest (1970, fig. XII–16f) suggested slightly earlier dates—that Lake Iroquois came into existence about 12,500 years B.P. and lasted, with gradually declining levels ("post-Iroquois lakes"), until shortly after 12,000 years B.P.

My reconstruction (fig. 14) follows that of Prest, using the older dates for Lake Iroquois. This reconstruction avoids the problem of how to hold Lake Iroquois within the Ontario basin when marine waters were present in the St. Lawrence Lowlands to the northeast.

The building of the outwash plains near Duane Center and Loon Lake (ice-front position 1, pl. 1) began about 12,700 years B.P. The edge of the ice sheet gradually retreated to the northeast, ice-dammed lakes formed in north- and east-draining valleys, and streams were diverted across divides, washing debris from large areas of bedrock and carving extensive channels. The reconstruction of ice-front positions suggests that the retreat to position 2, extending from Owls Head southeast to Clintonville and including the diversion of the Saranac River into the Ausable River, represents a short interval in late Coveville time, perhaps about 12,600–12,500 years B.P. The ice front in the Champlain Valley retreated north of the Saranac River, Lake Vermont began to drain south by way of the outlet near Fort Ann, N.Y. (fig. 14), and lake level dropped to that of the Fort Ann stage.

The highest stand of Lake Iroquois in Franklin County, east of Malone (pl. 1), would, under my reconstruction, date from about 12,500 to perhaps 12,200 years B.P. Because the stand is at an altitude of about 1,000 feet, it appears to have been graded to the Covey Hill channel and, therefore, may well date from the Covey Hill episode, that is from perhaps 12,400 to just prior to 12,200 years B.P.

The north and northeast retreat of the ice front was probably interrupted by minor oscillations, during which the moraines near Ellenburg Depot and near Cadyville were built. The moraine building may have taken place about 12,500–12,400 years B.P. During possible later oscillations, drift was removed from areas northwest of West Chazy, and the level of Prest's "post-Iroquois" lake was lowered, probably in several stages, from that of the Covey Hill channel (1,010 ft) to that of Lake Vermont (740 ft); these later oscillations were part of the Covey Hill episode.

The retreat of the ice front from Covey Hill into the St. Lawrence Lowlands caused the final lowering of the level of Prest's "post-Iroquois" lake to that of Lake Vermont in the Champlain Valley. The merging of the lake waters north of Covey Hill probably took place only a short time, perhaps only 200 years, before the lowering of Lake Vermont and the inflow of the marine waters of the Champlain Sea.

Wagner (1969) suggested that there was a time, just before the marine invasion, when the Fort Ann outlet ceased operation and the ice-dammed lake in the Champlain basin drained northward. The northward-draining lake he named Lake New York.

The northeast trend of the moraines in southeast Quebec (fig. 14) and the dates for the marine invasion suggest, as noted by McDonald (1968, p. 675), that the front of the main mass of the ice sheet retreated northwest across the St. Lawrence Valley. The merger of the ice-dammed lakes in the St. Lawrence and Champlain Valleys probably occurred shortly before the advent of the Champlain Sea, that is, about 12,200 years B.P. (Prest, 1970, fig. XII–16h). Carbon-14 dates suggest that the submergence lasted about 1,500 years. Uplift of the land closed the connection to the ocean, and Lake Champlain came into existence about 10,500–10,000 years ago (Prest, 1970, fig. XII–16o).

ST. LAWRENCE LOWLANDS IN NEW YORK STATE

The late-glacial history of the St. Lawrence Lowlands in New York State has been studied in detail by MacClintock and Stewart. In excavations for the St. Lawrence Seaway (fig. 14; MacClintock and Stewart, 1965, p. 81–95, figs. 25–33), a fivefold stratigraphic sequence was demonstrated, from oldest to youngest as follows: (1) A lower Malone Till deposited by ice from the northeast as indicated by striae on bedrock beneath the till; (2) an upper Malone Till with a fabric indicative of ice from the northeast but interbedded with sand, gravel, silt, and varved clay; (3) Fort Covington Till with a fabric indicating ice movement from the northwest; in places this till overlies bedrock with striae indicating ice movement from the northwest; (4) varved lake clay; and (5) fossiliferous marine clay, silt, and sand.

These authors interpreted the sequence as follows: An advance of Malone ice to the southwest followed by an ". . . oscillatory waning of Malone ice while standing in waters of an ice-dammed lake" (MacClintock and Stewart, 1965, p. 87). Next, the Fort Covington ice advanced toward the southeast. The waning of the Fort Covington ice was followed by an ice-dammed lake that later was replaced by the Champlain Sea. Terasmae (1965, p. 35, pl. 5) suggested that there may have been a minor post-Fort Covington advance from the north.

In northern New York and also in the Champlain Valley in Vermont (Stewart and MacClintock, 1969, p. 177), these authors postulated a period of emergence between deposition of the varved clays (unit 4) and deposition of the marine clays (unit 5); that is, a period of dry land between the withdrawal of the fresh-water lake (Fort Ann stage of Lake Vermont) and the incursion of the sea. As far as I am aware, no evidence has been found elsewhere to suggest that the St. Lawrence Lowlands were completely drained of ice-dammed lakes prior to the invasion of the Champlain Sea (Terasmae, 1965, p. 34; McDonald, 1968, p. 674).

MacClintock and Stewart traced their twofold division (Fort Covington and Malone) throughout the St. Lawrence Valley in New York State by studying the fabric of the tills (MacClintock and Stewart, 1965, p. 140; see also, MacClintock, 1959). The Fort Covington Till extends up the south slope of the St. Lawrence Lowlands almost to the foothills of the Adirondacks (fig. 14). Farther south, the drift is Malone Till. Two and a half miles northeast of Malone, these authors found gray Fort Covington Till separated by 2 feet of sand and silt from the underlying red-brown Malone Till.

MacClintock and Stewart traced the southern limit of their Fort Covington Drift (fig. 14) northeast to Covey Hill and south into the Champlain basin. They suggested that the Ingraham esker (Woodworth, 1905a), a narrow ridge of glaciofluvial deposits extending from Ingraham northward to a point west of Chazy (pl. 1), is actually a belt of ice-marginal kames built along the western edge of a lobe of Fort Covington ice. I favor Woodworth's interpretation of an ice-channel filling (Denny, 1972). In Vermont, the eastern part of the Fort Covington ice lobe, according to Stewart and MacClintock (1969, 1970), deposited their Burlington Till (fig. 14) that has a northwest fabric.

My reconstruction of the ice-front positions near Covey Hill is not far different from that of MacClintock and Stewart (1965, fig. 19). Their Fort Covington Drift border crosses Covey Hill along ice-front position 14 (pl. 1). The westerly trend of the streams that run diagonally down the north slope of Covey Hill suggests that the streams were initially ice-marginal channels. MacClintock and Stewart's reconstruction calls for the Fort Covington Drift to extend as a lobe far up the Champlain Valley, but I believe that ice-front position 14 crossed to the Vermont side of the valley along a line only a few miles south of the International Boundary.

I have not found evidence for more than one drift in the northeast Adirondack region. Perhaps the younger Fort Covington ice advance was only a minor feature and not the widespread advance postulated by MacClintock and Stewart. Differences in till fabric are their chief criteria for separating the two drifts outside the Seaway area. There is an alternative explanation. Near the center of a large ice lobe, such as the one that advanced up the St. Lawrence

Lowlands to the Ontario basin, the ice moves in the direction of the major lowland, in this case to the southwest. Near the margins of the lobe, however, the ice moves outward, perhaps nearly at right angles to the direction of movement of the lobe as a whole. This phenomenon is well shown in the Ontario basin where the general movement of the Erie lobe was to the southwest, whereas the margins of the lobe at times moved at right angles to the regional trend (Prest and others, 1968). Under this hypothesis, the Malone Drift was deposited by a southwest-moving ice lobe some distance back from the ice margin, whereas the Fort Covington Drift near the Seaway was deposited by southeast-moving ice near the margin of the same lobe.

For example, the exposure near Malone (MacClintock and Stewart, 1965, p. 68–69) of gray till (Fort Covington) overlying reddish till (Malone) may record the following sequence of events: Deposition of red drift, derived from adjacent red beds near base of Potsdam Sandstone, by southwest-moving ice some distance back from the front. Later, southeast-moving ice near the glacier front deposited gray drift derived from rocks that crop out to the northwest.

CHAMPLAIN LOWLAND IN VERMONT

The deglaciation of the northeast Adirondack region encompasses a part of what Stewart and MacClintock (1969, 1970) named the Burlington glacial stade. During this interval, they postulated an ice advance from the north-northwest that covered the entire Champlain lowland and crossed the Green Mountains to the east (fig. 14). Deglaciation involved northerly retreat of the ice front by calving into the waters of glacial Lake Vermont. Recessional moraines were not formed. It is surprising that there should be such well-developed moraines as that near Cadyville on the west side of an ice lobe in the Champlain Valley and none on the east. Perhaps the extensive deposits of kame gravel along the east edge of the lowlands south of Burlington (Stewart and MacClintock, 1970) are ice-marginal features that mark the east side of a Champlain Valley ice lobe (Denny, 1966).

Wagner believes (written commun., 1970) that minor oscillations of the ice front permitted several brief incursions of marine waters into the Champlain basin.

APPALACHIAN REGION OF SOUTHEASTERN QUEBEC

Deglaciation in the Appalachians east of the St. Lawrence Lowlands involved the northwest retreat of an active ice front that built prominent moraines. Gadd (1964) described what he called the Highland Front morainic system along the northwest flank of the Appalachian highlands (fig. 14), extending from Rivière-du-Loup, about 100 miles northeast of Quebec City, to Granby, east of Montreal. The moraine is at altitudes ranging from about 375 to 700 feet. In part, it consists of a belt of kame moraine 5–6 miles wide. The local relief is 100 feet or more. Gadd suggested that the ice sheet blocked both the St. Lawrence and Champlain Valleys when it stood along the moraine, damming Lake Vermont and preventing the sea from invading the St. Lawrence Lowlands. The northwest retreat of the ice edge let marine waters flood the St. Lawrence Lowlands and the Champlain Valley. The St. Narcisse moraine (fig. 14; LaSalle, 1966, 1970; Osborne, 1950; Karrow, 1959) on the northwest side of the St. Lawrence Valley was built by a readvance of the ice sheet into the Champlain Sea, perhaps about 11,000 years B.P. (Terasmae and LaSalle, 1968).

McDonald (1968), building on the work of Gadd and of Lee (1962, 1963), carried on detailed studies in southeastern Quebec near Sherbrooke (fig. 14). He believes that the Highland Front moraine east of Montreal was formed before 12,000 years B.P., possibly about 12,600 years B.P. (McDonald, p. 675), and near Rivière-du-Loup between 12,800 and 12,000 years B.P. The moraine apparently is not significantly time transgressive, and because there is at least local evidence of readvance, McDonald (p. 675) suggested that it may correlate with mappable features in New York State and regions farther west. On this basis, the moraine is approximately equivalent to the Loon Lake episode of northeast Adirondack region (ice-front position 1).

The Drummondville moraine (Gadd, 1964) in the lowlands east of Montreal, at altitudes ranging from only about 200 to 300 feet, was built just before the invasion of the Champlain Sea, about 12,200 years B.P. (Prest, 1970, fig. XII–16g).

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