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Iydrogeologic Framework of Stratified-Drift Aquifers in the Glaciated Northeastern United States

Regional Aquifer-System Analysis

Professional Paper 1415–B

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Hydrogeologic Framework of Stratified-Drift Aquifers in the Glaciated Northeastern United States

By ALLAN D. RANDALL

REGIONAL AQUIFER-SYSTEM ANALYSIS—
NORTHEASTERN UNITED STATES

U.S. GEOLOGICAL SURVEY PROFESSIONAL PAPER 1415–B

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FOREWORD

THE REGIONAL AQUIFER-SYSTEM ANALYSIS PROGRAM

The RASA Program represents a systematic effort to study a number of the Nation's most important aquifer systems, which, in aggregate, underlie much of the country and which represent an important component of the Nation's total water supply. In general, the boundaries of these studies are identified by the hydrologic extent of each system and, accordingly, transcend the political subdivisions to which investigations have often arbitrarily been limited in the past. The broad objective for each study is to assemble geologic, hydrologic, and geochemical information, to analyze and develop an understanding of the system, and to develop predictive capabilities that will contribute to the effective management of the system. The use of computer simulation is an important element of the RASA studies to develop an understanding of the natural, undisturbed hydrologic system and the changes brought about in it by human activities and to provide a means of predicting the regional effects of future pumping or other stresses.

The final interpretive results of the RASA Program are presented in a series of U.S. Geological Survey Professional Papers that describe the geology, hydrology, and geochemistry of each regional aquifer system. Each study within the RASA Program is assigned a single Professional Paper number beginning with Professional Paper 1400.

Charles G. Groat
Director

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CONVERSION FACTORS

Multiply	By	To obtain
	Length	
foot (ft)	0.3048	meter
mile (mi)	1.609	kilometer
	Area	
square mile (mi^2)	2.590	square kilometer
	Gradient	
foot per mile (ft/mi)	0.1894	meter per kilometer
	Flow rate	
gallon per minute (gal/min)	0.06309	liter per second
	Transmissivity	
foot squared per day (ft^2/d)	0.09290	meter squared per day
	0.0108	centimeters squared per second

Sea level: In this report, “sea level” refers to the National Geodetic Vertical Datum of 1929 (NGVD of 1929)—a geodetic datum derived from a general adjustment of the first-order level nets of both the United States and Canada, formerly called Sea Level Datum of 1929.

HYDROGEOLOGIC FRAMEWORK OF STRATIFIED-DRIFT AQUIFERS IN THE GLACIATED NORTHEASTERN UNITED STATES

By ALLAN D. RANDALL

ABSTRACT

The Northeastern United States was probably glaciated several times. In most valleys, however, the last ice sheet eroded down to bedrock, and all of the stratified drift that now overlies bedrock can be ascribed to the last deglaciation. Only in northeastern Ohio, northwestern Pennsylvania, and parts of southwestern and central New York do valleys commonly contain complex stratigraphy that results from repeated glaciation: multiple till sheets interlayered with lacustrine silt and with sand or gravel aquifers. Most stratified drift was deposited in the major valleys or lowlands when they were inundated by rising sea level or, more commonly, by proglacial lakes. Coarse-grained sediment was deposited in these water bodies as deltas and subaquatic fans, commonly amid stagnant ice blocks near the margin of the active ice, and also in channels within the ice sheet. The sediment was derived partly from debris-laden basal ice and subglacial till, entrained by meltwater flowing through tunnels, and partly from fluvial erosion of recently deposited drift in the uplands that bordered the valley-bottom lakes. The stratified drift consists of three facies deposited successively: proximal coarse-grained heterogeneous ice-contact deposits, followed by distal fine-grained lake-bottom sediment, and finally by coarse-grained surficial sediment deposited in shallow lakes or stream channels. One or two facies may be absent at any given site, but all three can be identified in many places along every major valley or lowland. Coarse-grained ice-contact deposits commonly constitute the bulk of the stratified drift in narrow or shallow valleys, whereas in broad lowlands they are widely scattered and occupy only a small fraction of the valley floor. In valleys where depth to bedrock exceeds about 100 feet, the bulk of the stratified drift commonly is fine grained, and transmissivity is not generally proportional to saturated thickness. Coarse sand and gravel tend to be more abundant in the southern part of the glaciated Northeast than farther north.

Several concepts or generalizations are widely applicable in interpreting the distribution of coarse-grained aquifers within glacial drift in the glaciated Northeast. In many localities, stratified drift can be divided into a series of morphosequences, which represent successive time intervals during deglaciation. Grain size decreases distally within each morphosequence; coarse, heterogeneous ice-contact sand and gravel predominates at the proximal end, whereas coarse sand commonly overlies lake-bottom fines at the distal end. Successive morphosequences can be difficult to distinguish, however, and coarse proximal deposits that are especially prominent and easily recognized commonly overlie relatively high bedrock and have small saturated thickness. Some investigators have inferred that water-yielding coarse sand and gravel are widely distributed at the base of the stratified drift, overlying till or bedrock, as a result of continuous deposition of subaquatic fans at the ice margin during retreat, even in broad lowlands where surficial stratified drift is predominantly fine grained. Other studies indicate, however, that subaquatic fans did not form in all valley reaches and that within broad lowlands they are restricted to relatively narrow zones that follow former subglacial channels. Straight or sinuous narrow ridges composed of coarse sand and gravel extend a few thousand feet, generally northward, from the heads of at least some morphosequences. These ridges, termed eskers, were deposited within or at the mouths of subglacial tunnels and are potential sources of large water supplies. Narrow zones of coarse stratified drift, typically eskers bordered or interspersed with other ice-contact deposits, are nearly continuous for several miles along some valleys and across low saddles. These zones seem to be the product of meltwater flow through persistent subglacial tunnel systems, are spaced 5 to 15 miles apart, and typically constitute aquifers more productive than any in intervening valleys. Coarse sand and gravel derived from upland watersheds was deposited where tributary streams entered proglacial lakes in the valleys. In deep valleys of the Appalachian Plateau and perhaps elsewhere, this coarse sand and gravel was deposited largely as subaquatic and surficial fans that now provide vertical hydraulic continuity between those tributaries and deep aquifers. In many regions, however, much sand and gravel was deposited near the proglacial lake surface in the upper parts of deltas, overlying fines and largely above the water table today. Where the deltas or fluvial sand deposits were built against or atop stagnant ice, collapse or settlement as the ice melted lowered the surficial sand and gravel, thereby increasing its saturated thickness.

Stratigraphy, water-transmitting properties, and saturated thickness of stratified drift can vary greatly over distances of a few hundred feet. Such variations are seldom fully known and are difficult to represent at map scales much smaller than 1:50,000. The glaciated Northeast can, however, be classified into several broad hydrophysiographic regions that are distinctive in their typical aquifer geometry, although many boundaries are gradational. In most of these regions, valleys generally sloped away from the ice sheet, and stratified drift was deposited in local lakes that formed successively as the ice retreated. Regions of low to moderate relief

characterized by abundant, generally coarse-grained stratified drift deposited in small, closely spaced valleys occur chiefly in the southern part of the glaciated Northeast. Locally, sandy outwash was thick enough to bury the pre-existing topography, resulting in small regions in which water-level response to pumping resembles what is expected in an infinite aquifer. Abundant sandy valley trains extend for miles in major valleys south of the limit of glaciation. By contrast, regions of low to moderate relief in northern Maine contain little surficial stratified drift; the most central of these regions has almost none and is bordered by regions of sparse stratified drift that is chiefly fine grained, or largely buried beneath till, or limited to narrow eskers that follow but do not fill the valleys. The Tug Hill Plateau in north-central New York and the Pocono Plateau in northeastern Pennsylvania also are regions of slight local relief (except near the plateau margins); their small valleys drain radially and contain little sand or gravel. The mountain regions in northern New England and northern New York have narrow valley floors where scattered bedrock outcrops imply even narrower aquifers; much stratified drift is perched above stream grade on the valley sides. Relief is comparably high in the Catskill Mountains and Appalachian Plateau, but stratified drift here is not generally perched on the valley sides, and bedrock does not crop out on the valley floors. In the Catskills, valley fills seldom exceed 150 feet in saturated thickness and include thick till deposited on lee (down-ice) sides of hills. In the Appalachian Plateau, valley fills commonly exceed 150 feet in saturated thickness; those that do are mostly silt and clay, capped and commonly underlain by sand and gravel, although west of Salamanca, New York, some valleys contain multiple drift sheets. Valleys that drained toward the ice along the northern margin of the Appalachian Plateau are unusually deep and also commonly contain multiple drift sheets. The drift is predominantly diamicton and fine-grained stratified sediment, but sand and gravel occur as discontinuous lenses at several depths. Broad lowlands along the Great Lakes, Lake Champlain, coastal Maine and New Hampshire, the St. Lawrence, Connecticut, and Hudson Rivers, and a few other places were inundated by extensive proglacial lakes or marine waters. These lowlands are characterized by widespread silt, clay, and fine sand interrupted by till-covered hills. Ice-contact sand and gravel deposits underlie small parts of the inundated areas, commonly as ice-channel fillings or esker deltas. Extensive surficial sand-plain aquifers are common only where tributary watersheds are large in relation to lowland area. These lake-dominated lowlands offer few opportunities for induced infiltration from large streams, whereas in many regions induced infiltration is the principal source of water to most large-capacity wells.

INTRODUCTION

The Northeastern United States was repeatedly invaded and covered by continental ice sheets during the past 2.5 million years (Ruddiman and Wright, 1987). The ice greatly deepened some valleys, transported vast quantities of sediment, and deposited the sediment upon the scoured bedrock as a mantle of glacial drift. Meltwater was released seasonally as the ice melted, particularly during periods of ice retreat (deglaciation), and deposited much sediment as stratified drift in valleys or lowlands at or beyond the ice margin. The stratified drift includes sand and gravel deposits that now constitute the Northeast's most productive aquifers, which are capable of supplying well yields at least an order of magnitude larger than generally obtainable from the underlying bedrock. Evaluation, development, and protection of these aquifers require knowledge of what might be termed "aquifer geometry"—that is, the three-dimensional extent and distribution of saturated, water-yielding coarse sand and gravel relative to bedrock, to fine-grained non-water-yielding drift, and to recharge sources such as streams and infiltrating precipitation. These relationships constitute the hydrogeologic framework that constrains any appraisal or simulation of hydraulic behavior, yield, or use of the aquifers.

The grain size, water-transmitting properties, and saturated thickness of stratified drift can vary greatly over distances of a few thousands, hundreds, or even tens of feet. Aquifer geometry in some localities has been reasonably well defined through examination of earth materials in excavations, abundant borehole records, geophysical exploration, and hydrologic testing. Commonly, however, hydrologists must evaluate localities where data are not readily available in sufficient detail that aquifer geometry is unmistakably obvious. Particularly in such localities, interpretations depend greatly on the conceptual models that the investigator uses to organize available data. Useful concepts or generalizations have been developed from well-documented studies of particular localities and from recognition of typical patterns based on experience or literature covering many localities. The first part of this paper explains several concepts that are widely applicable in interpreting aquifer geometry in the glaciated Northeast. The latter part describes distinctive aspects and examples of aquifer geometry in each of several regions, which are delineated on plate 1. Readers primarily interested in a particular locality would probably need to look at only one or two of these regional descriptions.

The glaciated Northeast, as described in this paper and illustrated in figure 1, includes nearly all parts of the United States east of Cleveland, Ohio, that were covered by the most recent (Wisconsinan) ice sheet: northeastern Ohio, northern Pennsylvania and New Jersey, nearly all of New York and New England—an area of 122,000 square miles. Long Island, N.Y., eastern Cape Cod, and the islands of Massachusetts are excluded because even though they were glaciated, their geology and hydrology are fundamentally different: they are largely surrounded by saltwater and are generally underlain by poorly consolidated Cretaceous sediments that include aquifers much more permeable and productive than the bedrock elsewhere in the glaciated Northeast. South of the Wisconsinan glacial border (fig. 1), some valleys were glaciated during older

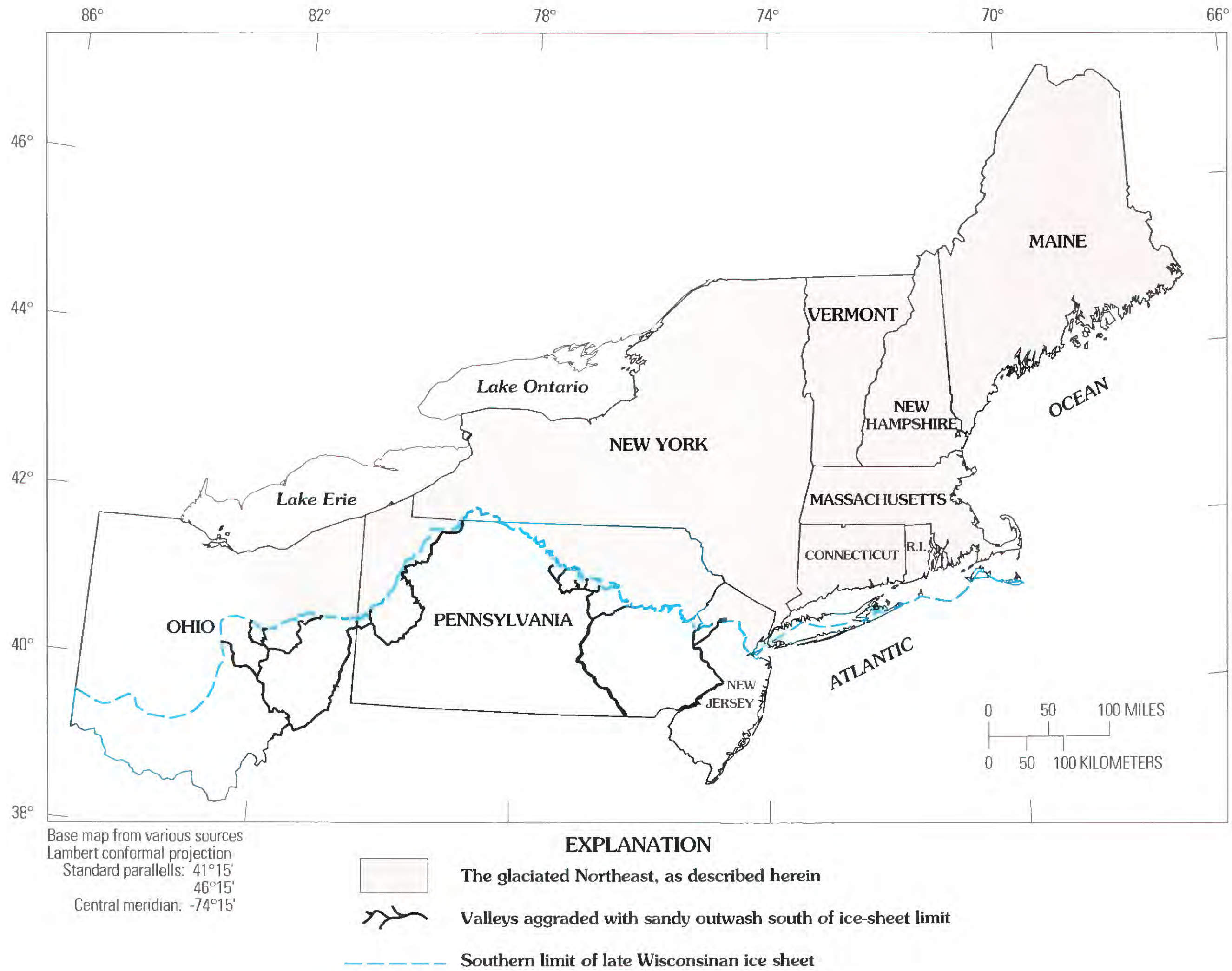

FIGURE 1.—Extent of Wisconsinan glaciation in the Northeastern United States.

glacial epochs and (or) contain sediments deposited beyond the ice sheets by meltwater; they are briefly discussed. Many place names are mentioned in this paper that could not be conveniently shown on plate 1; their locations are shown in cited references that contain information about the particular localities.

The information and interpretations in this paper were drawn together during a study of stratified-drift aquifers in the glaciated Northeast conducted as part of the U.S. Geological Survey's Regional Aquifer-System Analysis program (see Foreword). The principal findings of the study are presented in three chapters of Professional Paper 1415. Chapter A (Haeni, 1995) describes the application of surface geophysical methods to delineation of those aquifers. Chapter B (this paper) presents geologic insights and generalizations that can be used as conceptual templates in interpreting the hydrogeologic framework of the stratified-drift aquifers in localities within the region. Chapter C (Kontis, Randall, and Mazzaferro, in press) sets forth some typical values of hydraulic properties of stratified-drift aquifers, methods of computing those properties, and hydrologic principles, all useful in constructing ground-water flow models. Chapter C also presents a new method of simulating boundaries of those aquifers and recharge from the bordering uplands.

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THE SINGULAR RESULTS OF MULTIPLE GLACIATION

The chronology of glaciation and deglaciation is long and complex. Studies of ocean-bottom cores have revealed cycles in oxygen isotopes, microbiota, sea-surface temperature, and the ratio of detrital sediment to calcium carbonate that have been interpreted to indicate many episodes of global cooling over the past 2.5 million years. Each episode was presumably accompanied by ice-sheet expansion in North America. Most episodes apparently incorporated minor oscillatory temperature fluctuations and were followed by relatively rapid warming and deglaciation (Imbrie and others, 1984). The most recent major glacial episode, termed Wisconsinan, reached its maximum extent in most of eastern North America about 21,000 years ago (Mickelson and others, 1983; Stone, 1995). The previous major glacial episode, termed Illinoian, peaked 150,000 to 130,000 years ago (Ruddiman and Wright, 1987). Each ice sheet eroded soil and bedrock and eventually redeposited the debris as a widespread but highly variable mantle of glacial drift. Table 1 lists the principal episodes of ice-sheet expansion (stades) and deglaciation (interstades) in Wisconsinan time. Events more recent than 60,000 years ago have been dated chiefly from radioactive-carbon isotopes in wood and other organic matter, a method which has been found to slightly understate the actual age. Thus, the Wisconsinan maximum measured in true (sidereal) years occurred about 24,000 years ago (Stone, 1995).

The maximum extent of Wisconsinan drift in the Northeastern United States is shown in figure 1. Discontinuous remnants of older drift are found south of the Wisconsinan drift border in parts of New Jersey, Pennsylvania, and New York. Most of these remnants have been considered Illinoian in age (White and others, 1969; Marchand and others, 1978; Crowl and Sevon, 1980; Evenson and others, 1985, p. 85; Stone and others, 1989). The southern fringe of the drift in northeast Ohio and northwest Pennsylvania is likewise older than that farther north and has been considered middle Wisconsinan in age (White and others, 1969; White, 1982; Szabo and Miller, 1986). Regardless of the specific age assignments, many of which have been revised as new information became available, field evidence makes clear that the Northeastern United States has been covered by continental ice sheets more than once, and that the most recent deglaciation was interrupted by several readvances, some of regional extent that reflected changes in climate and others of more local extent that could be attributed to ice dynamics and interaction of ice sheets with proglacial lakes (Teller, 1987).

TABLE 1.—*Terminology and timing of Pleistocene ice-sheet expansion and deglaciation in the Northeastern United States (after Muller and Calkin, 1993)*

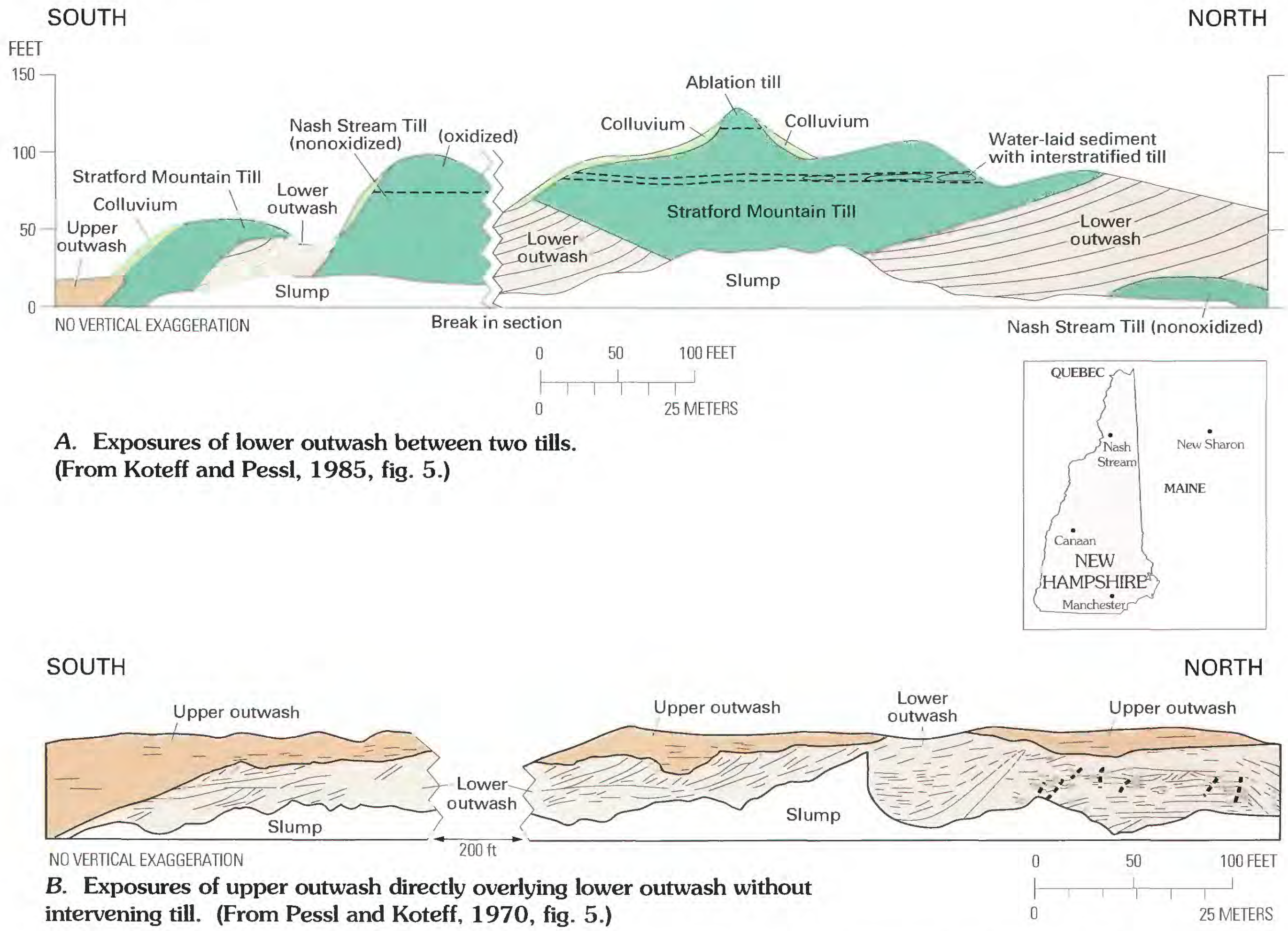
Subdivisions of Geologic Time				Thousands of years before present
PLEISTOCENE EPOCH	WISCONSINAN STAGE	LATE SUBSTAGE	Two Creeks Interstade	11.9
			Port Huron Stade	13.0
			Mackinaw Interstade	
			Port Bruce Stade	15.5
			Erie Interstade	16.5
			Nissouri Stade (Kent)	23
		MIDDLE SUBSTAGE	Plum Point Interstade (Port Washington)	35
			Cherrytree Stade	40
			Port Talbot Interstade	64
		EARLY SUBSTAGE	Guildwood Stade	
			St. Pierre Interstade	
			Nicolet Stade	117
	SANGAMON INTERGLACIAL			130
	ILLINOIAN STAGE			

Each cycle of ice advance and retreat resulted in deposition over most of the landscape of a sheet of till, which is a nonsorted and generally nonstratified mixture of grain sizes that can range from clay and silt to large boulders. Multiple layers of till are observed at many locations in the Appalachian Plateau of northeastern Ohio and northwestern Pennsylvania and in the northern fringe of the Appalachian Plateau of New York. The till layers are commonly interlayered with stratified drift, especially in the major valleys, and the resulting complex stratigraphy is what might be expected as the product of repeated glaciation. Two or more tills interlayered with stratified drift also are preserved near the massive moraines that approximately mark the southernmost extent of Wisconsinan ice in New Jersey (Vecchioli and others, 1967), Martha's Vineyard (Kaye, 1964), and Cape Cod (Sayles and Knox, 1943). Elsewhere in the Northeast, however, glacial drift seems to be largely the product of a single cycle of ice advance and retreat; that is, till is found only atop bedrock, and any overlying stratified drift can reasonably be ascribed to a single deglaciation. Apparently the late Wisconsinan ice eroded away most older drift, especially in the valleys where ice thickness, velocity, and abrasive pressure were greatest (Sugden and John, 1977, p. 154, 165, 182). Alternatively, if the Wisconsinan ice did not erode to bedrock, the sediments deposited during successive deglaciations must be so similar that they are indistinguishable. Examples of apparent single-stade stratigraphy are legion. Many papers explicitly interpret local deglacial history or aquifer geometry in single-stade terms, and a single-stade conceptual model is adequate for most geohydrologic site studies. In most regions, although scattered deposits provide evidence for multiple ice advances, those deposits are so located or so small in volume that they have little effect on the geometry of stratified-drift aquifers, as explained in the next few paragraphs.

Pre-Late Wisconsinan Drift.—In most of the glaciated Northeast, remnants of unconsolidated deposits older than late Wisconsinan have been found chiefly in upland localities remote from extensive stratified-drift aquifers. Several authors have described sites in New England where loose or compact nonoxidized till is underlain by compact oxidized till, the upper several feet of which are characterized by chemical alteration of clay minerals and by abundant joints whose surfaces are darkened by iron and manganese stains. Pessl and Schafer (1968), Newton (1978), Koteff and Pessl (1985), Weddle and others (1989), and Newman and others (1990) reviewed and augmented these observations and concluded that the oxidized till was significantly older than the late Wisconsinan surface till. The older till has been identified chiefly in drumlins and similar smooth upland hillsides where till is unusually thick (Weddle and others, 1989). Other authors have described sites where remnants of deeply weathered bedrock are preserved beneath till (Goldthwait and Kruger, 1938; Randall, 1964, p. 21; Feininger, 1965; Muller, 1965b, p. 50; Larsen, 1987b; Ridge, 1988, p. 176–178). Two tills with differing directional fabric and provenance are superposed in a small valley in southern Connecticut (Flint, 1961). Two small valleys carved into upland hillsides bordering Cayuga Lake in central New York are partly or entirely filled with drift that includes till overlain by fossiliferous stratified sediments deposited at least 35,000 and 54,000 years ago, as dated from radiocarbon (Maury, 1908; Shumaker, 1957; Bloom, 1967; Muller and Cadwell, 1986). Tills about 27,000 and 38,000 years in age are interbedded with stratified sediments and overlain by late Wisconsinan till at a site on the side of the Genesee River valley within the lowlands south of Lake Ontario (Young, 1993). In a small valley at Tahawus, N.Y., near the headwaters of the Hudson River, surficial till and three older tills are separated by sand, gravel, and (above the lowest till) wood-bearing clay that is more than 55,000 years old (Muller, 1965b; Craft, 1969; Muller and Calkin, 1993). Near New Sharon, Maine, the Sandy River has cut a gorge 100 feet deep through a ridge in which several diamicton layers are interbedded with clay, silt, fine sand, and one layer of sandy gravel (Caldwell and Weddle, 1983), all of which overlies organic debris more than 35,000 years old that, in turn, overlies till (Caldwell, 1960) or 80 feet of compact drift (Weddle and others, 1989). The entire section predates stratified drift from the last deglaciation that fills nearby reaches of the Sandy River valley. Along Nash Stream in northern New Hampshire, late Wisconsinan till and fluvial outwash overlie till and deltaic ice-contact coarse sand (fig. 2) that are attributed to an earlier glaciation (Pessl and Koteff, 1970; Koteff and Pessl, 1985). Whether the older sand extends below stream grade is unknown. None of these pre-late Wisconsinan stratified deposits has been shown to have enough saturated thickness and areal extent to be considered a significant aquifer.

Differences in the depth of leaching of carbonates from the drift, and in ice-flow directions as indicated by till fabric and striations, were at one time interpreted as evidence for deglaciation followed by readvance over large areas in the Appalachian Plateau of New York (MacClintock and Apfel, 1944), in the St. Lawrence lowland of New York (MacClintock and Stewart, 1965), and in Vermont (Stewart and MacClintock, 1969). Subsequent studies showed that the differences in depth of leaching were due largely to differences in the initial carbonate content of the drift, which, in turn, was a reflection of bedrock lithology, and also showed that ice-flow indicators did not support the distinct boundaries between drift sheets previously inferred; therefore, the surface drift in each of these areas was judged to be the product of a single glaciation (Dreimanis, 1957; Merritt and Muller, 1959; Larsen, 1972, 1987a; Wagner and others, 1972; Clark and Karrow, 1983).

Stratified Drift Deposited as Late Wisconsinan Ice Advanced.—Volumes of seasonal meltwater were probably much smaller when late Wisconsinan ice was advancing than in the warmer climate of deglaciation, and any sediment deposited by that meltwater was readily subject to erosion by the advancing ice. Near the terminal moraine in New Jersey, however, late Wisconsinan till overlies extensive fine-grained sediments and deltaic sands deposited in ponds ahead of the advancing ice (Vecchioli and others, 1967; Stanford, 1989b). Fine-grained deposits beneath late Wisconsinan till also have been recognized in a few other localities (Larsen, 1972, stop 1; Randall, 1980; Ridge, 1988). Along the Connecticut River at West Lebanon, N.H., pebbly sand exposed in two excavations was interpreted as advance outwash beneath 15 to 40 feet of late Wisconsinan till (Larsen, 1987a). Discontinuous thin lenses of stratified sediment occur in many places within the late Wisconsinan till; most probably were deposited in subglacial cavities, and some yield small amounts of water to springs or dug wells.



***A.* Exposures of lower outwash between two tills. (From Koteff and Pessl, 1985, fig. 5.)**

***B.* Exposures of upper outwash directly overlying lower outwash without intervening till. (From Pessl and Koteff, 1970, fig. 5.)**

FIGURE 2.—Generalized geologic sections based on exposures along Nash Stream north of Groveton, N.H., showing outwash that predates the most recent till.

Late Wisconsinan Readvances.—The disappearance of the late Wisconsinan ice sheet has long been recognized to have been an unsteady process, marked by pauses and temporary readvances that complicated the stratigraphic record in some localities. Mickelson and others (1983) concluded from stratigraphic evidence and radiocarbon dates that the southern margin of the ice sheet readvanced roughly every 1,000 years between 21,000 and 13,000 years ago in the region west of central Ohio, but state that regionally extensive readvances have not been documented in the glaciated Northeast before about 15,500 years ago. Teller (1987, p. 46, p. 62) reported that glacial retreat since 14,000 years ago was interrupted by several rapid readvances, many of which may have been ice surges facilitated by the presence of lakes and fine-grained sediments in front of the ice margin. Readvances in several localities in the glaciated Northeast during the last deglaciation have been proposed (table 2). Evidence for these local readvances, unlike the evidence for earlier glaciations discussed above, is largely in areas of stratified drift in the major lowlands. The principal evidence cited is scattered occurrences of till and (or) disturbed sediments (thrust faults, folds, or erosion by ice) within or atop stratified drift.

Most features attributed to the ice readvances listed in table 2 fall into one of two categories, either of which implies that the readvance did not alter aquifer geometry.

1. Minor interruptions in continuous deposition of fine-grained sediments, such as a few feet of till or disturbed bedding within lake-bottom fine sands or rhythmites. For example, a sparsely pebbly zone(s) within an extensive layer of fine sand that caps lacustrine silt and clay between Albany and Saratoga, N.Y., has been interpreted as outwash

from the Delmar readvance (Dineen and Hanson, 1983). A readvance is not required to account for the upward coarsening from silt to sand in this region; the sand can be explained by increased water velocities as the lake filled with sediment and lake level declined, and by wind blowing fine sand into the lake from adjacent sandy terraces (Dineen and Rogers, 1979; Dineen and Hanson, 1983).

2. Significant sand and gravel aquifers whose stratigraphy seems equally consistent with readvance or with a pause in retreat without readvance. For example, a 5-mile-long, deeply kettled valley train near Lake Luzerne, N.Y. (Connally and Sirkin, 1971), deltas near Rosendale, N.Y. (Dineen, 1986a) and near Middlebury, Vt. (Connally, 1982), and a deltaic outwash plain near Westfield, Mass. (Larsen and Hartshorn, 1982, p. 123) were ascribed to deposition from readvanced ice. None of the authors cited, however, presented evidence that rules out deposition of these features partly or entirely during the initial retreat (whether or not the ice subsequently readvanced to the head of each feature), nor do they mention any package of sediments underlying these features that could be ascribed to initial retreat prior to the proposed readvance. Fleisher and Cadwell (1984) ascribed thick, pitted outwash gravel overlying fine-grained lake deposits near Cooperstown, N.Y., to readvance, and Fleisher (1991a, b) later hypothesized that ice tongues could easily readvance as much as 12 miles across the deformable bed provided by the fine-grained lake deposits. A similar stratigraphic section would have resulted from a prolonged pause in retreat north of the locality described, however, whether or not the ice readvanced.

The most extensive of the proposed readvances listed in table 2 is the readvance in northeastern Maine. It represents a well-documented but quantitatively minor interruption in deposition of coarse-grained sediments and may be of hydrologic significance in that the discontinuous near-surface diamicton could slightly decrease the transmissivity of valley-fill aquifers and recharge to those aquifers. The diamicton is a distinctive feature of northeastern Maine, described further on in the section "Low-Relief Fringes of Maine Residual Ice Cap."

In summary, although the Northeastern United States was glaciated several times, multiple drift sheets in which sand and gravel aquifers alternate with confining layers are generally lacking. Apparently, the Wisconsinan ice generally eroded to bedrock in the major valleys. Remnants of pre-late Wisconsinan drift, chiefly till, are preserved in some upland localities, and evidence for readvance by late Wisconsinan ice has been reported from several valleys or lowlands. Some of the proposed readvances probably never occurred or were quite localized, however, and others apparently had little effect on the depositional environment or the topography that controlled it. Consequently, sediment deposition was commonly much the same before and after these minor readvances, or changed in ways that would be expected whether or not readvance occurred. By contrast, multiple drift sheets are widespread and hydrologically significant in three regions identified on plate 1 and discussed later in this paper: the Western Appalachian Plateau, the northern rim of the Appalachian Plateau, and the New Jersey terminal moraine.

TABLE 2.—*Proposed local readvances during late Wisconsinan deglaciation of the Northeastern United States*

Locality	References and remarks
Eastern New England	
Northeastern Maine	Kite, 1983; Borns and Borns, 1986; diamicton at or near top of sand and gravel at many sites in lower St. John and Aroostook valleys.
Pineo Ridge, Maine	Borns, 1973; Borns and Hughes, 1977.
Kennebunk, Maine	Bloom, 1963. Readvance reinterpreted as minor ice-front oscillation by Smith, 1981.
Manchester, N.H.	Stone and Koteff, 1979. Interpretation based on one exposure of till over disturbed rhythmites.
Connecticut River lowland	
Middletown, Conn.	Flint and Cushman, 1953; Flint, 1956. Readvance refuted by Stone and others (1982) and London (1985).
Chicopee, Mass.	Larsen and Hartshorn, 1982; Larsen, 1982.
Hudson-Champlain lowland	
Tappan, N.Y.	Averill and others, 1980. Readvance disputed by Stanford and Harper, 1991.
Rosendale, N.Y.	Connally and Sirkin, 1970; Dineen, 1986a. Readvance disputed by Duskin, 1986.
Delmar, N.Y. (also Yosts, N.Y.)	Dineen and Rogers, 1979; Dineen and Hanson, 1983; Dineen and Hanson, 1985.
Luzerne, N.Y.	Connally and Sirkin, 1971. Readvance disputed by DeSimone and LaFleur, 1986.
Bridport, Vt.	Connally, 1970, 1982.
Lake Ontario lowland	
Niagara County, N.Y.	Smith and Calkin, 1990. Two tills separated by silt in Lake Ontario bluffs.

THE ENVIRONMENT OF STRATIFIED-DRIFT DEPOSITION: PROGLACIAL LAKES AT THE ICE MARGIN

Glacial drift is collectively the end product of erosion and deposition by the ice itself and by meltwater. Ice eroded most effectively in the valleys and deposited till most abundantly as hills or localized accumulations on upland hillsides. Meltwater, by contrast, eroded channels in the uplands and deposited stratified drift chiefly in the major valleys. The stratified drift was deposited in proglacial lakes or marine embayments that developed in major valleys or lowlands during deglaciation and, to a lesser extent, along stream channels that spread across the valley floors after the proglacial water bodies had filled with stratified drift or had drained.

The importance of proglacial water bodies as sites of sediment deposition has been long and widely recognized. Davis (1890) and Salisbury (1892) recognized that sand or "wash" plains (outwash plains) in Massachusetts and New Jersey were predominantly of deltaic origin. Flint (1930, p. 63, map) interpreted the stratified drift of Connecticut as "successions of marginal ice-contact terraces of debris built delta-wise into temporary glacial lakes." Smith and Ashley (1985, p. 136, p. 169) mentioned that glaciolacustrine deposits are major components of the glacial drift in North America, and that the most common ice-contact lakes are small, short-lived lakes that form at glacier margins. Teller (1987, fig. 21) calculated that more than 40,000 square miles along the ice margin in eastern North America were inundated by large lakes during most of the last deglaciation, not counting the numerous small lakes or the extensive areas where marine water bordered the ice sheets.

Nearly 10,000 years elapsed between initial retreat of the late Wisconsinan ice sheet from its southernmost margin in Pennsylvania and New Jersey (fig. 1) and the disappearance of the last ice from north-central Maine. In any given locality, however, uncovering of the landscape and deposition of stratified drift took place within a few hundred to, at most, a few thousand years, during which depositional conditions changed and after which incision and regrading by postglacial streams generally prevailed. In this paper, discussions of events or conditions during deglaciation in any locality refer to the relatively short period when proglacial lakes existed there and the ice margin was retreating, or perhaps briefly readvancing, through that locality.

WHY PROGLACIAL LAKES WERE UBIQUITOUS

Proglacial lakes developed along ice margins during deglaciation for several reasons:

1. Deep erosion by ice in some valley reaches created closed basins that trapped water after the ice melted. The Finger Lakes of New York, some of which were eroded far below sea level, are prime examples (Coates, 1968; Mullins and Hinchey, 1989); ice-scoured closed basins also have been documented in other localities (Newman and others, 1969; Handman and others, 1986; Fleisher and others, 1992; Moore and others, 1994).
2. Many valleys drained or sloped toward the ice sheet, which blocked the natural outlet of runoff. Thus, water accumulated as proglacial lakes in these valleys during ice advance and retreat. Each lake spilled across the lowest saddle on the divide or, in a few instances, across or under the ice. Among the many proglacial lakes of this type were those in the Genesee and former Allegheny valleys of western New York (Frimpter, 1974; Muller and others, 1988), the northern Adirondack region (Denny, 1974), the Nashua valley of eastern Massachusetts (Koteff, 1982), and the northwestern Green Mountains (Wagner, 1972). The apparent absence of lacustrine deposits in a few north-draining valleys several miles in length has been ascribed to the presence of stagnant ice, which was inferred to have occupied all space below the lowest saddle on the watershed perimeter (Denny, 1974, p. 7; Cadwell, 1985, p. 83) or to have allowed water to drain out through crevasses or tunnels to a lower, more northerly outlet (Randall, 1978a, p. 13; Miller and Randall, 1991).
3. In the glaciated Northeast, far more valleys drained away from the ice sheet than toward it. During deglaciation, most of these valleys also contained proglacial lakes, each of which was ponded behind drift recently deposited downvalley. Whenever the surface of the ice melted down to an altitude lower than the surface of the drift downvalley, a proglacial lake developed. Overflow from the lake spilled across the earlier deposits, slowly incising them—or, in a few localities, spilled across a bedrock saddle that was slightly lower than the deposits choking the main valley. Commonly, sediment filled these lakes almost as fast as they formed, thereby lengthening the dam or spillway that controlled the lake level. Examples of self-damming sequences of stratified drift are found in the lower Connecticut River valley (Stone and others, 1982, p. 9; Koteff and others, 1987, p. 3),

the Susquehanna River valley (Randall, 1986a), and the Merrimack River valley (Koteff and others, 1984).

4. Northward tilting caused by isostatic depression of the Earth's crust beneath the ice sheet decreased gradients of south-draining valleys, thereby increasing the size of lakes ponded behind drift dams. The depression also allowed some postglacial streams to flow in directions opposite to their preglacial ancestors. For example, the Farmington River in Connecticut once flowed southward from Farmington center but now flows 13 miles northward at low gradient to cross Talcott Mountain through a saddle at Tariffville, a diversion facilitated by isostatic depression of the saddle during deglaciation (Flint, 1934; Lougee, 1938). The potent effects of isostatic depression on the relative altitude of the widely separated saddles that successively accepted overflow from the proglacial Great Lakes are described by Teller (1987) and Larsen (1987).

STREAM DEPOSITS AND THEIR RELATION TO PROGLACIAL LAKES

Stratified drift ranges in grain size from cobble-boulder gravel to laminated or massive silty clay. The fine-grained sediments obviously settled on the bottoms of proglacial lakes or marine embayments, so are termed glaciolacustrine (or glaciomarine). Most sand and gravel deposits are so located with respect to upland topography, planar depositional surfaces, delta foresets, and (or) fine-grained sediments that they also must have been deposited below the surface of proglacial water bodies. Thus, they are also glaciolacustrine (or glaciomarine). Nevertheless, many such deposits have been termed glaciofluvial in some publications because their coarse grain size, heterogeneity, and sedimentary structures such as channels, plane and trough crossbedding, and imbricated gravel indicate deposition by rapid unidirectional flows of meltwater. Presumably, these coarse sediments were deposited where meltwater streams entered the water bodies. This type of depositional setting has been described in some detail by Rust and Romanelli (1975), Smith (1982), Thompson and Smith (1983), Shaw (1985, p. 64), and Smith and Ashley (1985, p. 171). Two idealized models of depositional processes that occur where sediment-laden meltwater enters proglacial water bodies are illustrated in figure 3; the subaquatic fan generates a stratigraphic section that coarsens downward, whereas the delta model coarsens upward.

Sinuous or linear ridges of sand and gravel, termed ice-channel fillings, are found in many localities. Some were deposited in ponded water by the mechanisms illustrated in figure 3—that is, they constitute deltas built along linear lakes in ice crevasses (LaSala, 1961) or successive subaquatic fans built at the mouth of a subglacial tunnel as the ice margin retreated in ponded water (Banerjee and McDonald, 1975, p. 135). Others, known as eskers, are inferred to have been deposited in tunnels beneath the ice that, although water-filled, could not be described as ponded because the meltwater was flowing rapidly under pressure (Banerjee and McDonald, 1975; Shreve, 1984, 1985).

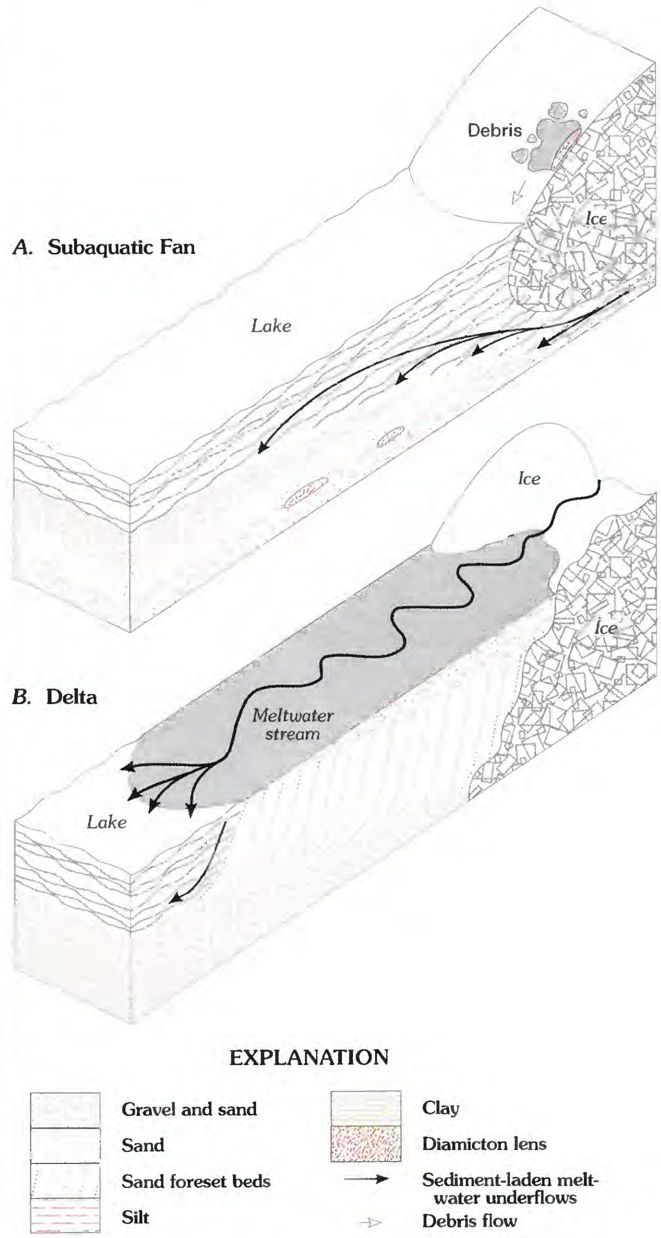


FIGURE 3.—Idealized stratigraphy resulting from two depositional mechanisms in proglacial lakes.

Sand and gravel also were deposited along meltwater streams whose channels were open to the atmosphere rather than beneath ice. Such glaciofluvial sediments are fairly widespread at land surface but generally are thin; thus, they constitute only a small fraction of total stratified-drift volume. They occur chiefly as topset beds of deltas and as veneers deposited as streams aggraded or incised the former beds of water bodies that had drained or had filled with sediment (Gustavson and others, 1975, p. 268; Stone and others, 1982). Thus, they are generally found within the boundaries of the former proglacial water bodies and overlie fine-grained sediments that were deposited in those water bodies. Beyond the limit of glaciation, however, many tens of feet of glaciofluvial outwash unrelated to lakes accumulated in some south-draining valleys or lowlands (Kaye, 1960; Larson, 1982; Stone and Peper, 1982; White, 1982).

THE NATURE OF THE ICE MARGIN

Deposition of stratified drift depended not only on the presence of proglacial lakes but also on the nature of the ice margin in those lakes. The manner of retreat of the last ice sheet has long been a subject of debate, particularly with regard to the extent of stagnant, motionless ice relative to active, advancing ice. Koteff and Pessl (1981), Connally (1982), and Dineen (1986a) reviewed and appraised the history of contrasting opinions. Two conceptual models of ice-margin retreat are widely accepted:

Stagnation-Zone Retreat.—Ice ceases to deform and flow under the influence of gravity when its thickness decreases to the point that gravity forces no longer exceed the combination of atomic bonding forces within the ice and friction at the base of the ice (Hughes, 1981, p. 236). The minimum thickness that will flow on land is 200 feet (von Engeln, 1961, p. 63), but thinner ice tongues can advance in ponded water because buoyancy reduces basal friction. Some authors envision the ice sheet as bordered by a marginal zone in which the ice is thin, motionless, and melting in place. Active ice to the north gradually thins by ablation (downwasting), so periodically another marginal segment stagnates, generally south of some topographic irregularity that provides resistance to flow. Evidence that stratified drift was deposited against or atop stagnant ice exists in many places in the form of kettleholes, hummocky terrain, irregular collapsed margins of level terraces, and downwarped beds or downthrown faults bordering these collapse features. The width of the zone of continuous stagnant ice in southern New England has been estimated to be 0.3 to 1.2 miles typically (Stone and others, 1982) and as much as 1.5 miles locally (Koteff and Pessl, 1981, p. 12); large, detached ice blocks commonly remained over a much larger area. The stagnation zone was estimated to be at least 1.5 miles wide in the Champlain valley of Vermont (Connally, 1982) and even wider in eastern Vermont (Larsen, 1987).

Active Retreat.—Other authors envision the ice sheet as remaining active and solid all the way to the margin. Layers of till, and deformation attributable to lateral pressure or drag from overriding ice, are both found locally within stratified drift and are taken as evidence of an active ice margin. Channels carved into some upland hillsides at gentle gradients (rather than heading directly downslope) are believed to mark where the sloping margin of solid ice once intersected the hillside and bordered the channel (Goldthwait and Mickelson, 1982; Stanford, 1989a; Larsen, 1992; Stanford and others, 1998). Active ice margins and rapid retreat are commonly observed where modern valley glaciers terminate in deep, extensive water bodies. Large blocks of ice break off and float away, a process known as calving. The ice may be grounded at the edge of the lake (or ocean) and terminate in a steep cliff, or it may terminate in an ice shelf or ramp that is floating (Smith and Ashley, 1985, p. 172; Teller, 1987).

In summary, depositional features of stratified drift suggest that during deglaciation the outermost active ice was commonly but not universally bordered by a fringe of decaying stagnant ice. Sediment deposition into open water beyond or beneath an active, floating ice margin was most likely in large, deep proglacial water bodies, whereas deposition in progressively expanding interlaced ponds amid and over stagnant ice was most likely in narrow, shallow valleys.

SOURCES OF SEDIMENT

Much of the sediment deposited in proglacial lakes as stratified drift was entrained by meltwater directly from the ice sheet, as basal ice was eroded by subglacial streams or melted at the ice margin. Some of the sediment was eroded from recently deposited drift in uplands bordering the lakes, as streams incised their channels in a landscape shaped by the ice. Jokulhlaups—catastrophic floods resulting from sudden draining of proglacial lakes—locally provided large volumes of coarse sediment by erosion of upland channels (Hand and Muller, 1972; Hildreth and Moore, 1993).

Studies of debris transport by modern ice sheets and valley glaciers have revealed that most of the sediment is carried in the lowest 10 to 50 feet of the ice (Boulton, 1975; Lawson, 1979; Shaw, 1985). Some melting occurs at the base of the ice sheet, caused by friction and pressure and the geothermal gradient, but the volume of seasonal melting at the ice surface is far greater. Active ice can override stagnant ice or recently deposited drift near the ice margin. Some authors have suggested that this upward shearing process can bring much basal ice up to the glacier surface where its sediment is readily accessible to melt-water (Goldthwait, 1951; Koteff, 1974; Koteff and Pessl, 1981). Under some conditions, sediment-laden subglacial meltwater freezes onto the base of the glacier, and the addition of multiple layers of sediment-laden ice by repetition of this process can raise the uppermost sediment far above the base (Boulton, 1967; Lawson, 1979). Other studies have concluded, however, that entrainment of sediment by surficial meltwater at the ice margin is relatively minor. Evenson and others (1984) estimated from field observation and mass-flux calculations that only about 10 percent of the drift at the terminus of a glacier in an Alaskan valley was delivered by the ice itself. The rest was delivered by subglacial and ice-marginal streams and was derived from glacially or nonglacially fed tributaries and from melting of the glacier some distance above its terminus. Surficial meltwater commonly disappears into moulins or crevasses over the entire length, or at least the distal part, of modern glaciers (Gustavson, 1984; Shreve, 1984, 1985), then flows through subglacial channels to the ice margin (Kamb and others, 1985). By this mechanism, surficial meltwater would obtain access to abundant basal sediment. Sediment-laden meltwater discharged from tunnels at the toe of a glacier into ponded water may jet upward as fountains because of density differences, thereby carrying suspended sediment up to the water surface (Gustavson, 1984; Cowan and Seramur, 1990). When distal flow of basal meltwater is blocked by accumulation of sediment in a subaquatic fan at the ice margin (Boothroyd, 1984), presumably similar meltwater fountains carry sediment up to the top of the fan, which may eventually build up to the water surface and become a surficial outwash fan or fan delta (Gustavson, 1975).

Meltwater in subglacial tunnels obtains sediment from the melting of sediment-laden ice along the tunnel walls by heat derived from viscous friction in the flowing meltwater. The tunnels do not enlarge appreciably because the ice flows toward them by plastic deformation in response to a hydraulic pressure gradient (Shreve, 1984, 1985). Creep of subglacial till also can deliver sediment to tunnels (Alley, 1990), and migration of subglacial tunnels can erode drift beneath the ice (Gustavson, 1984). Some eskers are bordered by bedrock surfaces that have been scoured free of till for many tens of feet on either side of the esker (Banerjee and McDonald, 1975, p. 134).

In the glaciated Northeast, till mantles the bedrock not only in uplands, where it is normally the only unconsolidated sediment, but also in lowlands, where it underlies stratified drift. Several water-resource appraisals in Connecticut (Ryder and others, 1970; Mazzaferro and others, 1979; Handman and others, 1986) report that till thickness beneath stratified drift in lowlands averages 5 to 10 feet, much less than the average of 25 feet in the surrounding uplands. Melvin and others (1992, p. 8) reported that till in southern New England is generally less than 10 feet thick where overlain by stratified drift and 10 to 30 feet thick where exposed at land surface. Randall (1972) compiled lithologic logs of 1,110 wells and test borings in the Susquehanna River basin of New York. Of these logs, 256 penetrated to bedrock in major valleys and were sufficiently explicit for a fairly confident determination of whether the unit immediately overlying bedrock was till. More than one-third of these logs reported no till atop bedrock, and about half reported less than 3 feet of till. The other half penetrated 3 to 67 feet of till; these larger thicknesses were mostly near the valley sides or in a few hills composed of till on the valley floor. Canace and others (1993) drew geologic sections across several valleys north of the terminal moraine in Morris County, N.J., where many feet of till separate stratified drift from bedrock in some places, but more commonly geologic mapping showed that till mantles bedrock on the valley sides, whereas well records indicated sand and gravel atop bedrock beneath the valley floor (fig. 4). The absence or negligible thickness of till in so many places beneath stratified drift in valleys could result from erosion by subglacial meltwater. Subglacial meltwater channels tend to follow valleys because valleys deflect hydraulic equipotential contours in the ice sheet (Shreve, 1984, 1985) and because subglacial melting occurs first where the ice is relatively close to geothermal heat. Many subglacial channels persisted throughout the later stages of deglaciation and deposited sand or gravel atop bedrock near the retreating ice margin, as described further on.

In the glaciated Northeast, meltwater issuing directly from the ice and streams from bordering uplands each contributed substantially to the deposition of stratified drift, but the ratio of sediment derived from these two sources differs from place to place. Along the Wallkill valley in New Jersey, pebble counts from stratified drift generally indicate a marked lack of mixing: 30 to 85 percent of the pebbles in gravel along the valley axis were derived from the carbonate bedrock that underlies the valley, suggesting a subglacial source, whereas nearly all pebbles in gravel along the sides of the valley were bedrock types from the adjacent uplands (Stanford and others, 1998). In Berkshire valley, New Jersey, a linear valley with only a few small tributaries, pebbles derived from bedrock that underlies the valley predominate in all gravel deposits (Stanford, 1989b). In the Susquehanna River basin of New York, erratic-rich gravel was transported from the north along most large valleys by ice flow and (or) subglacial meltwater, but an influx of erratic-poor gravel from tributaries draining adjacent uplands locally diluted the erratic-rich gravel (Moss and Ritter, 1962) and (or) preceded it (Randall, 1978a). In the Delaware River basin, most large accumulations of stratified drift are near the mouths of tributaries, which must have been major sediment sources (Kirkland, 1973). The central lowland of Connecticut and Massachusetts received much inwash from tributaries draining the adjacent uplands during deglaciation, as indicated by the relative abundance of coarse-grained deposits along the margins of the lowland (Stone and others, 1979) and by the substantial admixture of crystalline rock fragments from the uplands with red sedimentary rock fragments from the lowland in those deposits (Lougee, 1938; Randall, 1970). In localities where sediment derived from upland sources differs in lithology from that derived up-ice, the contrast has proved useful in stratigraphic correlation and in explaining spatial variations in water quality.

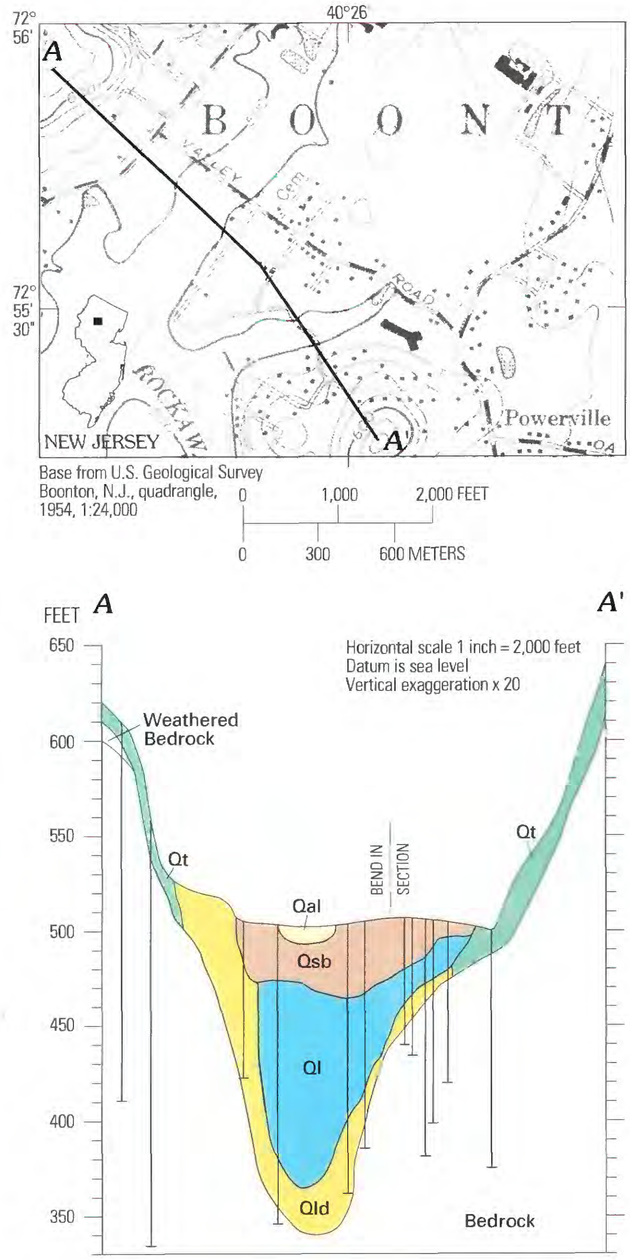
THE UNIVERSAL THREE DEPOSITIONAL FACIES

In nearly all valleys or lowlands throughout the glaciated Northeast, the stratified drift consists of three facies that are successive in any given locality, although transgressive over longer distances:

Facies 1. Early deglacial, proximal, or ice-contact facies: predominantly coarse-grained sediment, heterogeneous, varying widely in degree of sorting, deposited close to active ice and (or) amid abundant stagnant ice masses.

Facies 2. Mid-deglacial, distal facies: predominantly fine-grained sediment, deposited in moderately large bodies of water, farther than facies 1 from the active ice.

Facies 3. Late-deglacial or postglacial, surficial facies: coarse-grained sediment, commonly well sorted, deposited in shallow water and commonly atop facies 2 or 1.



EXPLANATION

Qal	**Postglacial alluvium**—Sand, silt
Qsb	**Outwash**—Sand, minor gravel
Ql	**Lake-bottom deposits**—Fine sand, silt, minor clay
Qld	**Delta, subaquatic fan**—Sand and gravel
Qt	**Till**
⊥	**Well, casing length and generally driller's log available**
A——A'	**Line of geologic section**

FIGURE 4.—Stratigraphy of stratified drift beneath the floor of Stony Brook valley near Denville, N.J. (Modified from Canace and others, 1993, pl. 3.)

Facies 1 was deposited near the ice margin by rapidly flowing meltwater wherever its velocity decreased—in tunnels beneath stagnant ice and probably beneath active ice as well, in crevasses, moulins, or erosional channels within or beside stagnant ice, and in lakes or the ocean where subglacial streams emerged from the front of the ice sheet. Much of the fine-grained sediment remained in suspension, to be deposited downvalley as facies 2. Diamicton layers are commonly interbedded within the stratified drift, having been deposited by mass movements of loose sediments freshly exposed on the ice surface or valley sides, or by the ice sheet during minor oscillations. Two alternative conceptualizations of the depositional environment for facies 1 are shown in the top panel of figure 5.

As the ice margin retreated, the locus of coarse facies-1 sedimentation retreated with it, and any stagnant ice left behind gradually melted. Consequently, in valley reaches not already choked with facies-1 sediments, proglacial lakes expanded, meltwater velocities slowed proportionately, and fine-grained sediment was deposited. Sediment was transported within proglacial lakes largely by density underflows and turbidity currents that traveled along the bottom but also, in part, by turbid overflows that fanned out at the surface (Powell, 1981; Smith and Ashley, 1985). Drifting and subsequent melting of sediment-laden icebergs dropped sediment pellets, sand, and stones as large as several feet in diameter into the lake-bottom muds. In some places, diamicton layers similar to till were formed by this process and (or) by subaquatic slumping of unstable sediments (Powell, 1981; Ridge, 1985).

As long as the rate of retreat was reasonably steady, facies 1 and 2 could potentially form a physically continuous time-transgressive onlapping stratigraphic sequence in any given valley, interrupted only where high ground limited the extent of the proglacial lake or when the lake ceased to exist because water drained out through some outlet that had previously been plugged by the ice sheet. Continuity of the facies-2 fine sediments also could be interrupted if a pause in the rate of retreat allowed coarse sediment to accumulate up to the lake surface near the ice margin.

Facies 3 consists of coarse to fine sand and gravel that were deposited in shallow ponded water or along stream channels late in deglaciation or even after the complete disappearance of the ice. It typically occurs atop facies 2 but can overlie facies 1 as well and is commonly, but not necessarily, finer grained and better sorted than facies 1. Several processes resulted in the widespread presence of surficial sand and gravel atop fine-grained sediment. For example, when a substantial influx of sediment took place at the same location for some time, whether derived from a tributary stream or from an ice-marginal or subglacial stream during a pause in retreat, coarse sediment eventually prograded outward across fine sediment in the form of a delta. When lake levels declined because erosion lowered the spillway, or because ice-margin retreat exposed a lower outlet, deltas could fill the lake more quickly and prograde farther. When lakes drained completely through new outlets, tributary streams commonly covered the former lake floor with sand and gravel derived from erosion of older stratified drift along the valley margins and from erosion in upland drainage basins. Postglacial stream incision converted many of these alluvial deposits to terraces.

Each of the three depositional facies can be identified in many places along almost any major valley or lowland throughout the glaciated Northeast. For this reason, the three-facies single-deglaciation model is termed universal. Where all three facies occur at one location, they form a characteristic coarse/fine/coarse stratigraphic section. At many locations, however, only one or two facies are present. For example, in narrow valleys where small, shallow lakes developed, meltwater velocities generally remained too large for deposition of facies 2. In broad lowlands, surficial sediment is fine grained over large areas that were too distant from major streams to be capped by facies-3 sand or gravel after the lake drained, and facies 1 is commonly restricted in areal extent, occupying the entire stratigraphic section at some locations but completely absent nearby. Despite these variations, each facies is widely distributed.

Locally, the three-facies model can be an oversimplification of the true complexity of stratified-drift stratigraphy. For example, localized complexity can result where ice-marginal deltas prograde over fine-grained sediment but are then abandoned by retreat before the lake is filled so that continued fine-grained deposition overlaps the delta (Thompson, 1978, fig. 9; Smith, 1982, fig. 9). Complexity also can result from the persistence of large stagnant ice blocks long after the retreat of the ice margin. Randall (1986) mapped several kettleholes in which fine-grained organic-rich sediment accumulated atop collapsed deltaic outwash typical of facies 3, and illustrated how these complexities can be accommodated by a quasi-three-dimensional digital aquifer model designed to represent the concept of three successive depositional facies. Fleisher (1986a) described a depositional setting in which large, detached ice blocks create depressions (termed dead-ice sinks) that are crossed by late-deglacial or postglacial rivers. As the detached ice blocks melt, a succession of small ponds

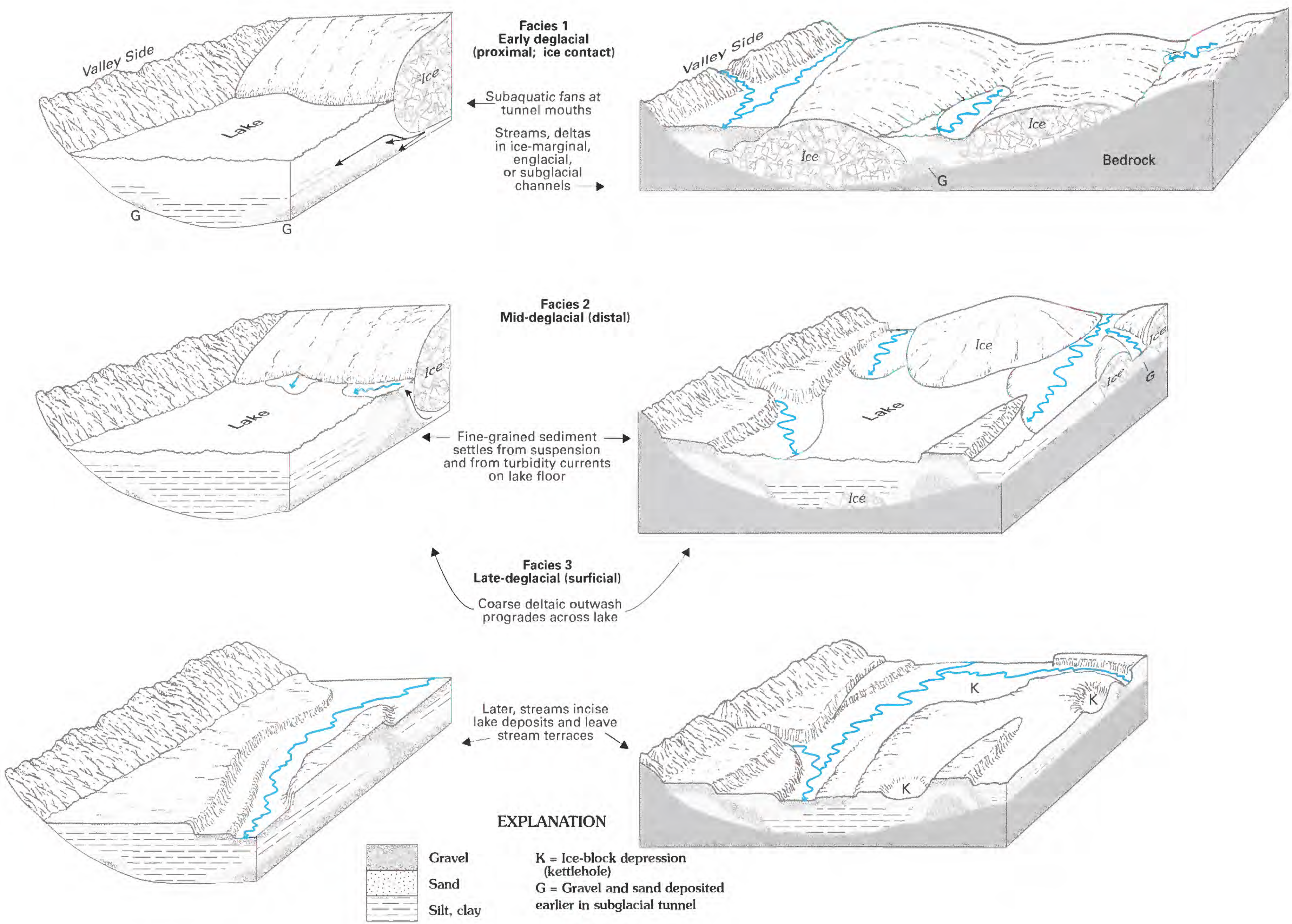


FIGURE 5.—Idealized development of three successive depositional facies. Left panel assumes active retreat; right panel assumes stagnation-zone retreat. Under both assumptions, similar stratigraphic suites can develop: all coarse grained in some places, coarse over fine in other places, and elsewhere coarse over fine over coarse.

form and trap both coarse-grained and fine-grained sediment. Collapse and subsidence disrupt the deposits in these ponds, and the resulting stratigraphic heterogeneity makes correlation of units very difficult.

Despite local complexities and absences of individual facies, the concept of three depositional facies is widely applicable throughout the glaciated Northeast and is useful in geohydrologic analysis. Many geologic reports have discussed and illustrated in maps and sections the distribution of early deglacial proximal or ice-contact stratified drift bordered or buried by deltaic outwash that grades down into fine-grained sediments (for example, Borns and Hagar, 1965a, b; Hartshorn, 1967; Stone and Randall, 1977; Stone and others, 1982; Dineen and Hanson, 1983; Koteff and others, 1984, stop 2), and a few reports have explicitly acknowledged a three-facies stratigraphy to be typical (Fleisher, 1986, p. 132; Canace and others, 1993, p. 5; Randall, 1981, p. 149). Several geohydrologic maps have depicted aquifer geometry within the stratified drift by units such as coarse, coarse/fine, and fine/coarse that, in effect, represent the distribution of facies 1, 2, and 3 as described in this paper (for example, Meade, 1978, and the 10 reports on which Meade's map is based; Cederstrom and Hodges, 1967; Frimpter, 1972; Crain, 1974; MacNish and Randall, 1982; Randall, 1986; Miller, 1987; Moore, 1990; Canace and others, 1993). Map units of this type represent reasonably well the geometry of aquifers and confining layers and can be useful in guiding exploration for water supplies. Saturated thickness must be taken into account because facies 1 and 3 can be thinly saturated where they overlie bedrock or facies 2 at altitudes above stream grade, as discussed in more detail further on. Such map units also can guide regulation of land use because contaminants spilled into facies 2 (or 3 over 2), where surficial aquifers are absent or thin, can be contained and removed more easily with less risk to groundwater supplies than can spills into facies 1 (or 3 over 1), where no barrier is present to inhibit downward flow.

TYPICAL PATTERNS OF AQUIFER DISTRIBUTION WITHIN STRATIFIED DRIFT

The delineation of aquifers in any particular locality requires site-specific information and also a conceptual framework within which to organize that information. Maps of surficial geology and (or) soils, which are available for many areas in the glaciated Northeast, are useful in that they show the distribution at land surface of coarse-grained stratified drift relative to fine-grained stratified drift, till, and bedrock. However, the distribution of stratified-drift aquifers is by no means coincident with the surficial distribution of coarse-grained stratified drift. For example, some coarse facies-1 deposits constitute productive aquifers but are buried beneath facies-2 fines and, thus, are not depicted on surficial geology or soils maps; others are surficial but useless as aquifers because they rest on shallow bedrock and are thinly saturated. In many localities, some site-specific information can be obtained from borehole logs and records, surface geophysical exploration, examination of exposures, and streamflow records. Generally, however, the interpretation of that information depends as much on what patterns of aquifer geometry the investigator expects or recognizes as it does on the information itself. The following sections describe several typical patterns or generalizations regarding aquifer distribution within stratified drift that are widely applicable across the glaciated Northeast.

AQUIFER GEOMETRY AS A FUNCTION OF WIDTH AND DEPTH OF VALLEYS

The stratified drift in most narrow, shallow valleys consists largely of ice-contact deposits (facies 1); the small lakes that formed during deglaciation were promptly filled with coarse sediment, while most fine sediment remained in suspension to be deposited elsewhere. By contrast, the stratified drift in valleys that are unusually broad or deep is largely fine grained, presumably because the amount of coarse sediment available during deglaciation was insufficient to fill the large lakes that developed. In broad valleys and lowlands, widely scattered coarse ice-contact deposits occupy only small fractions of the proglacial lake bottom or valley floor. In deep valleys, basal coarse proximal deposits (facies 1) and surficial coarse deposits (facies 3) commonly are separated by great thicknesses of very fine sand, silt, and clay (facies 2).

Transmissivity of stratified drift is roughly proportional to saturated thickness in the narrow, shallow valleys where ice-contact deposits predominate. This generalization cannot be extended to deeper valleys, however. Several studies have reported that an increase in depth to bedrock is likely to be accompanied by a comparable increase in thickness of very fine sand, silt, and clay. For example, depth to bedrock beneath stream grade is as much as 500 feet in the Susquehanna River basin of New York, but thickness of water-yielding gravel and coarse sand rarely exceeds 150 feet and is much less in some localities (MacNish and Randall, 1982). In the Taunton River basin of Massachusetts, wherever stratified drift approaches 200 feet in thickness, it consists of silt and clay interbedded with fine sand and yields little water; productive aquifers in this basin generally range from 20 to 80 feet in saturated thickness (Lapham, 1988). In the lower Quinnipiac valley of Connecticut, saturated thickness of stratified drift exceeds 200 feet in many places (Mazzaferro and others, 1979), but only 1 of 16 drillholes that reached refusal or bedrock penetrated more than 100 feet of saturated sand and gravel (fig. 6). In deep valleys, therefore, transmissivity generally does not correlate well with depth to bedrock or saturated thickness of stratified drift.

REGIONAL TRENDS IN ABUNDANCE OF COARSE-GRAINED STRATIFIED DRIFT

In a general or regional sense, coarse-grained stratified drift seems to decrease in abundance northward, from areas near the limit of glaciation to the Canadian border, probably because the rate of ice retreat increased and meltwater flux decreased as deglaciation progressed. Grain size of glacial drift also varies regionally as a function of bedrock lithology. These broad regional trends are commonly overwhelmed by variations in

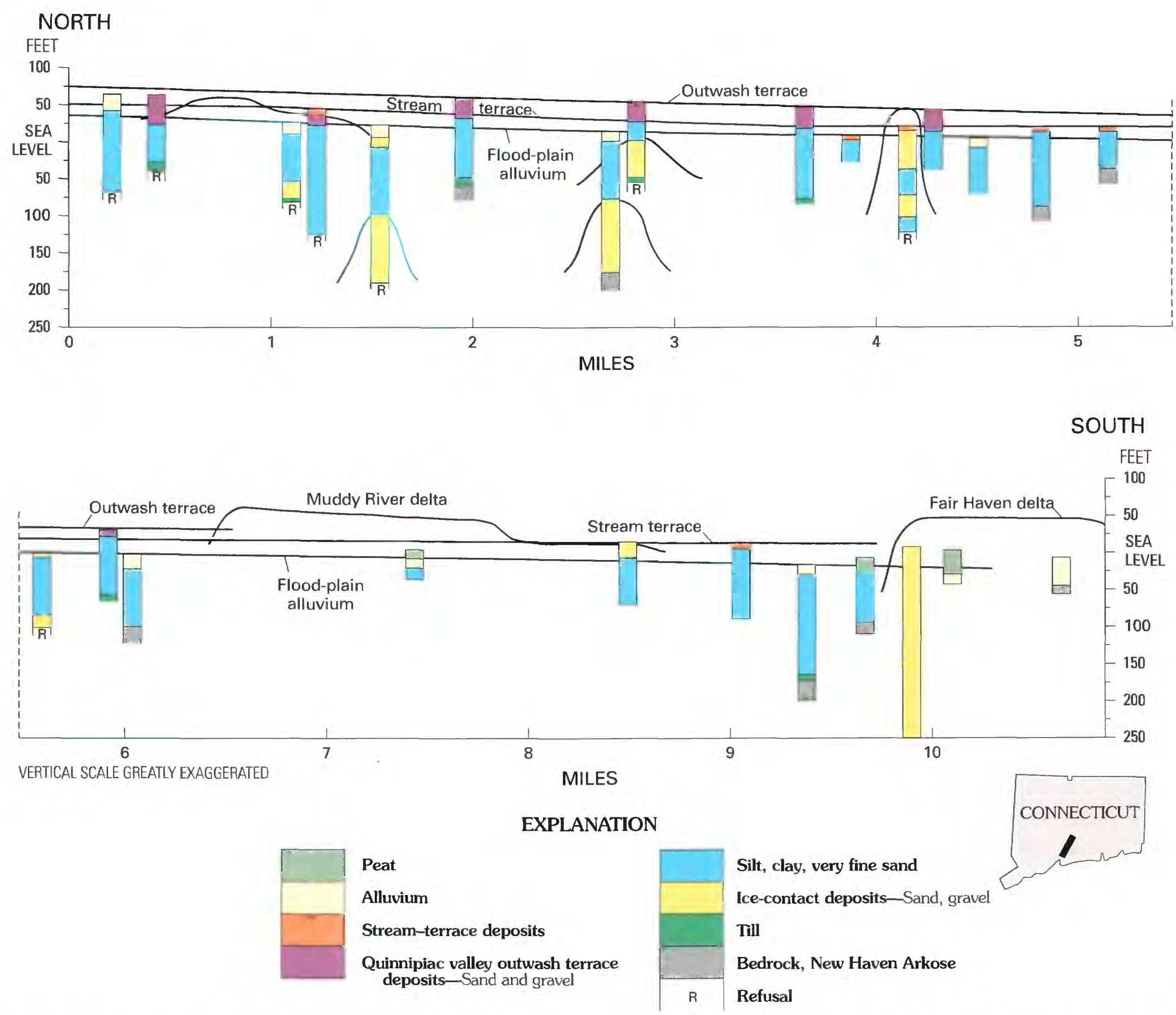


FIGURE 6.—Stratigraphy in the valley of the Quinnipiac River, Connecticut, from Wallingford center to Fair Haven. Longitudinal profiles of flood plain and terraces were constructed by projection of highest surface altitudes of each deposit to valley center line. The upper surfaces of ice-contact deltas and some buried ice-contact deposits are depicted by lines that terminate downward at the apparent limit of the deposit. Locations and narrative logs of boreholes are given by Mazzaferro (1973) or Haeni and Anderson (1980). (Modified from Stone and others, 1985, fig. 8.)

abundance of coarse stratified drift over short distances caused by more localized aspects of bedrock topography and ice dynamics, but nevertheless deserve consideration in regional studies.

Several authors have inferred that the net rate of ice-margin retreat increased with time during deglaciation in New England and eastern New York; some interpretations (Schafer, 1968, 1979; Mickelson and others, 1983) were based on a few radiocarbon dates, others (DeSimone and LaFleur, 1986, p. 226) on volumes of ice-marginal stratified drift. When the Wisconsinan ice sheet was at its maximum extent, it flowed approximately southward from central Canada across all of the northeastern United States (Andrews, 1987, p. 15; Hughes, 1987); as the ice thinned, however, its flow was more affected by relief, and after about 14,000 years ago, most ice from Canada flowed eastward along the St. Lawrence lowland or westward along the axes of Lakes Ontario and Erie (David and Lebuis, 1985; Hughes, 1987, fig. 21). These diversions could be expected to decrease the flux of ice and sediment toward the ice margin between Maine and Pennsylvania, resulting in faster retreat, less meltwater generated per unit area uncovered, and less erosion and transport of sediment by meltwater. Furthermore, the direction of both superglacial and subglacial meltwater flow is controlled by the slope of the ice surface (Shreve, 1985) and parallels the flow of the ice itself. At the Wisconsinan maximum, therefore, the ice margin in the glaciated Northeast must have received whatever seasonal meltwater was generated in much of southeastern Canada, as well as all of the Northeastern United States. Late in deglaciation, however, seasonal meltwater flux from Canada would have been diverted in the same manner as the ice flux. All of these changes could be expected to result in a decrease from south to north in volume of sand and gravel transported and deposited by meltwater.

The ratio of clay and silt to sand and gravel within the glacial drift varies considerably across the glaciated Northeast as a function of bedrock lithology. At one extreme, coarsely crystalline metamorphic bedrock yields many hard, durable cobbles, pebbles, and sandsize mineral grains. By contrast, shale and siltstone bedrock yield weak fragments that readily separate into their constituent particles of clay and silt size during transport by ice and meltwater. Many authors have reported that till consists largely of fragments of the underlying bedrock. For example, Denny and Lyford (1963, p. 6) stated that till in the Appalachian Plateau of New York and Pennsylvania has a sandy texture near coarse-grained sandstone and conglomerate of Pennsylvanian age, but has a silty-clay-loam or silt-loam texture where the bedrock is fine sandstone and siltstone of Devonian age. Clark and Karrow (1983) reported that till on dolomite and sandstone in the St. Lawrence lowland contains 57 percent sand and 35 percent silt, whereas till on the Precambrian metamorphic bedrock at the north edge of the Adirondack Mountains contains 66 percent sand and 27 percent silt. Melvin and others (1992, fig. 4) showed that till derived from granitic crystalline rocks in southern New England is consistently more sandy and less silty than till derived from phyllite, marble, or from metasedimentary rocks in eastern Massachusetts. Where multiple drift sheets occur in Ohio, Pennsylvania, and western New York, the lowermost sheet closely reflects the local bedrock, whereas one or more overlying tills are enriched in silt and clay because the ice readvanced across interstadial silt and clay in the Lake Erie and Lake Ontario basins (LaFleur, 1979b; White, 1982).

Because stratified drift includes many beds that differ widely in grain size and sorting, its average grain size in a particular region cannot be determined as readily as that of till. Some striking regional differences are apparent, however. For example, in New England and the Adirondack Mountains of New York, most sand grains are individual mineral species, and fine sand to coarse silt are abundant in deltaic stratified drift. In the Appalachian Plateau of New York, by contrast, many sand grains are tiny fragments of shale or siltstone; deltaic foresets are commonly fine to coarse sand to gravel, and fine to very fine sand is much less abundant than silt and clay. In a reach of West Canada Creek valley that borders the western Adirondacks, lake-bottom sediments derived from the Adirondacks contain thin clay layers alternating with thick fine-sandy silt layers, whereas lake-bottom sediments derived from regions of shale and carbonate bedrock to the west and southeast contain more clay and finer grained silt (Ridge, 1985, p. 62, 104–112). In central Connecticut, poor sorting and interstitial silt and clay are more abundant in ice-contact deposits derived from Mesozoic shale and arkosic sandstone than in nearby deposits derived from Paleozoic gneiss and schist (Randall, 1964, p. 54). These qualitative observations suggest that both aquifers and confining units within the stratified drift could be somewhat more permeable in regions of metamorphic bedrock than in regions of shale or siltstone, but whether the differences are enough to affect ground-water flow significantly has not been demonstrated.

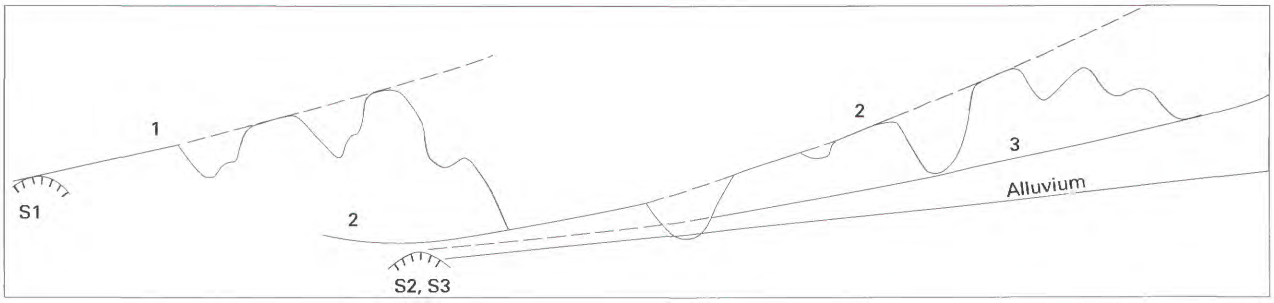
STEPWISE RETREAT AND THE MORPHOSEQUENCE CONCEPT

Stratified drift in much of southern New England and New Jersey has been classified into a series of units, known as morphosequences, that represent successive time steps or intervals in deglaciation. The development of the morphosequence concept has been summarized by Koteff (1974) and Koteff and Pessl (1981). The concept is essentially as follows: whenever the ice margin remained in the same location long enough for stratified sediment to build up to the surface of a proglacial lake, streams flowing across that sediment imparted a smooth fluvial profile graded to lake level or (if the lake filled completely) to the spillway across which it drained; all the stratified deposits that underlie that profile are of nearly the same age and constitute a single morphosequence, within which grain size decreases distally. Coarse, heterogeneous, poorly sorted gravel (facies 1) predominates near the proximal (upslope) end but grades distally into deltaic sands that, in turn, prograde across fine-grained sediment, such that the distal (downslope) end of the morphosequence commonly consists of fine gravel and sand (facies 3) overlying lake-bottom silt, clay, and (or) very fine sand (facies 2).

Progressive melting of the ice sheet resulted in frequent creation of new morphosequences, as illustrated in figure 7. When the ice margin bordering a small lake retreated past a spillway that was lower than the existing lake surface, the lake began to drain through the new outlet; subsequent stratified deposits were graded to the new lake level and generally can be distinguished by their lower elevation from earlier deposits. When the ice margin retreated without exposing new outlets, the proglacial lake expanded headward past the former proximal sediments without change in lake level, and eventually a new morphosequence formed with its head against the repositioned ice margin. In this situation, the distal parts of successive morphosequences approached the same lake level and, thus, cannot be easily distinguished by means of elevation. Whether lake levels remained constant or declined abruptly, however, several successive profiles developed along each valley system as the ice retreated, and generally overlap in shingled fashion (fig. 7). Their gradients are typically about 20 or 25 feet per mile, but approach 40 feet per mile near the heads of some morphosequences and can be as gentle as 8 feet per mile in distal outwash plains (Cadwell, 1972, table 6: Koteff, 1974, p. 125; C. Koteff, U.S. Geological Survey, oral commun., 1994). Generalization as to gradient is difficult because initial depositional gradients varied and have been variably altered by isostatic rebound. Koteff and Larsen (1989) and Larsen and Koteff (1988) presented evidence that isostatic rebound in central New England did not commence until the ice margin had retreated at least as far as northern Vermont, and that lake shorelines and morphosequence profiles south of northern Vermont have all been upwarped by 4.74 feet per mile to the N. 21° W. Farther north, rebound began before the ice was gone; therefore, early depositional surfaces such as morphosequences were upwarped more than later lake shorelines (Larsen, 1987). In western New York and northeastern Ohio, the direction of maximum rebound is N. 27° E., and the magnitude of rebound increases northeastward (Calkin and Feenstra, 1985).

In principle, delineation of morphosequences can be useful in geohydrologic appraisals because thin surficial sand or fine gravel can be expected to overlie fine-grained sediments near the distal end of a morphosequence, whereas coarse sand and gravel, in part highly permeable, can be expected to predominate near the head of a morphosequence. In practice, however, delineation of individual morphosequences can be difficult, for several reasons. Large parts of many morphosequences were deposited over or against stagnant ice and subsequently collapsed to the point that the fluvial depositional profiles are unrecognizable. Parts of some morphosequences were planed off by meltwater or postglacial streams. In some deep valleys or where the ice melted quickly, subaquatic fans never built up to lake surface, so no fluvial profile ever developed. In some places, successive ice-marginal deltas butt against one another to form a large terrace that seems to be one morphosequence but is actually several morphosequences, some of which can be so short that they are predominantly coarse grained throughout. Because these complications are so common, delineation and geohydrologic interpretation of morphosequences on the basis of topography or elevation alone are unlikely to be reliable; some information from strategically placed excavations and (or) boreholes is generally needed as well. Examples of successive abutting deltas that can be distinguished by elevation and by lithology in excavations are described by Evenson and others (1985, stops 8–9), Stanford and others (1998), Koteff and others (1984), and Larsen (1992). A detailed example of morphosequence analysis that considers several complications is presented farther on in the section "Eastern Hills and Valley Fills."

Reports that evaluate stratified-drift aquifers in localities within the glaciated Northeast generally include a brief account of the study methods; few such reports mention identifying ice margins or basing interpretations on morphosequence maps. Some maps of transmissivity or ground-water availability bear a close resemblance to maps of bedrock topography or saturated thickness; this resemblance suggests that



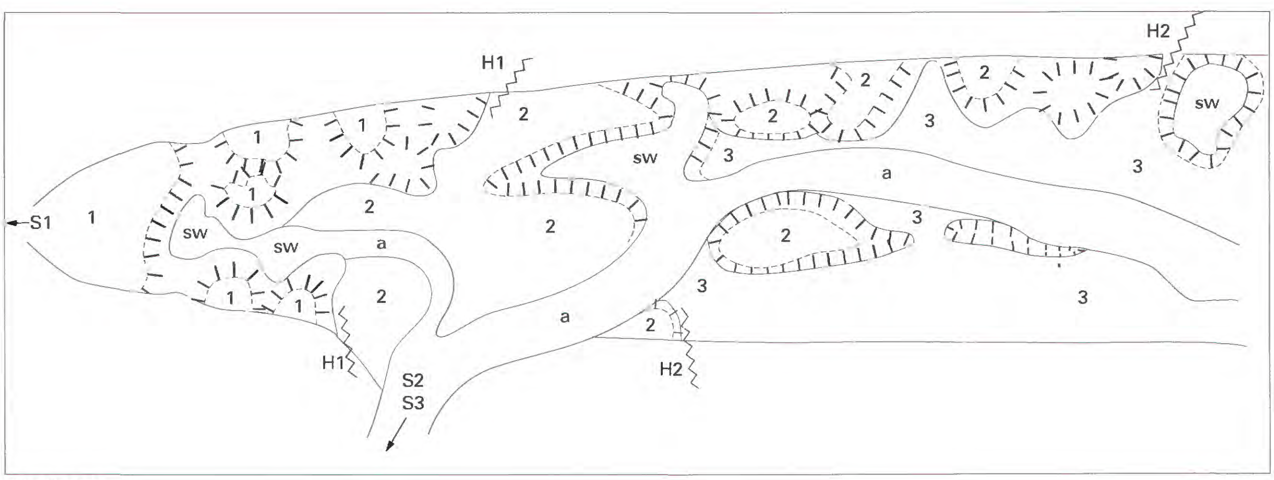
A. **Longitudinal profiles along valley**

EXPLANATION

2 **Land-surface profile and morphosequence number**—1 is oldest

Inferred depositional profile—Shown where land surface is now lower as a result of collapse

S1 **Spillway, on till or bedrock**—Numbered morphosequence 1 is graded to spillway S1



NOT TO SCALE

B. **Plan view of same valley reach**

EXPLANATION

Remnant of depositional surface and morphosequence number—1 is oldest

Steeply sloping or hummocky land surface—The result of collapse when buried ice melted

a **Alluvium**

sw **Swamp**

S2 **Spillway to which numbered morphosequence is graded**

H2 **Ice margin at head of numbered morphosequence**—Shown only along valley walls

Contact between units of different age

FIGURE 7.—Idealized arrangement of successive morphosequences that comprise the stratified drift in a typical valley.

the association of permeable coarse stratified drift with morphosequence heads was not consistent or not recognized from the data available when the maps were drawn. The morphosequence concept has been used in geohydrologic reports from several geographic areas, however, including the following examples.

1. *Eastern Massachusetts.*—Hartshorn (1960, 1967) mapped and described several outwash plains that grade distally (downgradient) into lake-bottom deposits that border the collapsed heads of older morphosequences. He did not identify the outwash plains as morphosequences, but his

description clearly indicates that they are; he specifically mentioned the heads of outwash plains as proximal coarse stratified drift. The geohydrology of the same area was evaluated by Lapham (1988), who generally inferred transmissivity to be higher near the heads of morphosequences than farther downgradient. Deltas a few miles to the northwest were assigned to a secondary rank of aquifer potential by Williams (1968) because they tend to become finer with increasing depth.

2. *South-Central New York.*—MacNish and Randall (1982, fig. 6) identified several localities in which ice tongues deposited recessional moraines during pauses in retreat. In these localities, proximal sand and gravel deposits many tens of feet thick prograde south across lake clay. Surficial aquifers more than 40 feet thick underlie most of these localities (MacNish and Randall, 1982, pl. 1).

3. *Northeastern Connecticut.*—Randall and others (1966) described the northern or proximal parts of several morphosequences (Randall and Pessl, 1968, units Qq 4, 5, and 6, Qw2, Qb2, Qc2) as coarse grained, and the southern or distal parts as fine grained (capped by a coarse surficial layer that is largely above the water table). Some other morphosequences in the same area were not similarly described, however.

In summary, the concept of ice-margin retreat as a stepwise process with morphosequences recording successive increments of retreat is potentially of fundamental importance in delineating aquifer geometry. The morphosequence concept has not been easily or widely applied in hydrogeologic studies, however, probably because sediment deposition was continuous and often included small or gradational steps that are intermediate between the most prominent morphosequences and are not obvious from examination of topographic maps or sparse exposures. The concept could prove increasingly useful as geohydrologic studies become increasingly detailed.

ASSOCIATION OF EXPOSED FACIES 1 WITH SHALLOW BEDROCK

As implied in the foregoing discussion of morphosequences, in most valleys or lowlands throughout the glaciated Northeast, remnants of several gently sloping, terracelike surfaces underlain by stratified drift can be identified. In each locality, the highest surfaces formed first, when the retreating ice margin was nearby, and are likely to be underlain by proximal ice-contact deposits (facies 1). Low surfaces may represent the floors of large lakes underlain by fine-grained deposits (facies 2), or the distal parts of morphosequences that head farther north (facies 3, commonly overlying facies 2), or terraces and flood plains incised by postglacial streams into one or more of the above categories. Thus, the stratified drift that stands highest in altitude generally consists of relatively coarse sand and gravel (facies 1), whereas lower surfaces are likely to be underlain by a substantial thickness of fine-grained sediment (facies 2). Nevertheless, many high-standing stratified deposits, easily recognized as facies 1, are not productive aquifers because the underlying bedrock is too high to allow appreciable saturated thickness. This is true regardless of the mode of ice retreat. As melting lowered the surface of a retreating ice tongue, stagnation was likely to occur first where ice was thinnest—over bedrock knolls and along the sides of the valley or lowland. The stagnant ice was not replaced as it melted, and, thus, the earliest ponds developed over shallow bedrock and were filled with coarse-grained facies-1 deposits. Where a thick ice tongue extended into a deep lake and remained active as it retreated by calving, any sublacustrine bedrock knoll high enough to provide resistance to flow was likely to cause a pause in retreat and, at the same time, provide a shallow platform from which proximal facies-1 sediments could easily build up to lake surface. This principle of preferential facies-1 deposition over bedrock highs is illustrated in figure 8 and has been observed in many localities, including the following:

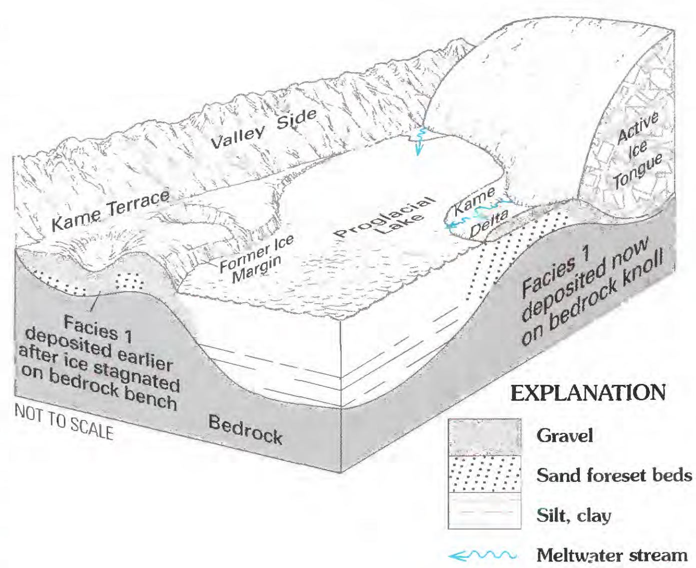


FIGURE 8.—Idealized example of preferential deposition of coarse-grained proximal stratified drift (facies 1) in valley areas where bedrock is shallow.

1. *Southern Maine.*—Where isostatic depression allowed the ocean to abut the retreating ice margin, many marine deltas were formed on the distal side of bedrock ridges, fed by meltwater streams through gaps in those ridges (Thompson and Smith, 1983, p. 8), and subaquatic outwash was commonly localized on the proximal sides of other bedrock ridges (Thompson, 1982, p. 222). The bedrock ridges seem to have temporarily anchored the active ice margin, which elsewhere retreated rapidly by calving. This concept of bedrock highs as localized anchors (pinning points) for floating ice shelves is commonly invoked in the geological literature (for example, Powell, 1981).
2. *North-Central Connecticut.*—West of Route 10 in southern Granby and northern Simsbury, Conn., level depositional surfaces at altitudes of 270 to 310 feet cap deltaic ice-contact deposits (facies 1) that are sandy and permeable but thinly saturated, as evidenced by well records and by exposures that indicate the sand rests on till or bedrock at altitudes at or above adjacent streams or lowlands (Randall, 1970; Melvin and Bingham, 1991). Near Route 10, the bedrock surface declines eastward to altitudes below sea level, and east of Route 10, broad terraces of stratified drift at altitudes of 200 feet or lower are underlain by more than 100 feet of fine-grained facies-2 lake sediment (fig. 18).
3. *Southeastern New York.*—Frimpter (1972, pl. 2) mapped many localities as underlain by unsaturated surficial sand and gravel. Most of these features are topographically prominent ice-contact deposits (facies 1) whose unsaturated condition could result only from high bedrock beneath.

Thus, although some facies-1 ice-contact deposits can be easily identified from their relatively high altitude and position at the heads of morphosequences, as well as from exposures of coarse-grained sediment, the typical occurrence of such deposits over buried knolls or valley sides of bedrock or till precludes interpreting such deposits as potentially productive aquifers without investigating bedrock topography and saturated thickness.

Some maps of surficial geology depict units characterized by constructional topography of knobs and hollows, or undulating gentle slopes, and by exposures described as sandy diamicton and (or) variably sorted, variably stratified deformed sand. These materials are commonly interpreted as the result of resedimentation, mass movement, and collapse of earlier deposits, at least some of which had been transported and deposited by meltwater. They have been mapped as ablation till, morainal till, complex or mixed deposits, kame moraine, or inwash according to the perception of the investigator and the array of sediments visible in typically poor exposures. They occur particularly in areas of sandstone or coarse crystalline bedrock and particularly on lower hillsides in regions of high relief. Predominance of diamicton and position above stream grade severely limit the aquifer potential of most such deposits.

DISTRIBUTION OF BURIED FACIES 1 BENEATH FINE-GRAINED SEDIMENTS

As previously explained, many broad valleys or lowlands are underlain by extensive fine-grained sediments (facies 2), in part capped by a thin surficial sand or gravel layer (facies 3). Landforms commonly include broad flood plains, swamps, former lake bottoms, and low terraces. Coarse ice-contact deposits (facies 1) are locally evident as high-standing landforms that border the valley floor or interrupt low-lying areas. The preceding section pointed out that many of these ice-contact deposits were built where bedrock is relatively high or shallow and, thus, are too thinly saturated to be major aquifers. An important question is whether facies 1 also is present beneath facies 2 in the low-lying areas and constitutes a buried, basal aquifer.

No simple, universal answer to this question is yet available. On the one hand, conceptual models of the retreat of an active ice margin in ponded water call for deposition of gravel and sand in subaquatic fans near the mouths of subglacial meltwater conduits (Shaw, 1985, p. 53; Smith and Ashley, 1985, p. 171). Multiple subglacial streams flow beneath some modern glaciers and deposit their load of coarse sediment near the glacier snout (Gustavson, 1975; Powell, 1981). Lateral continuity of the resulting sand and gravel across the valley presumably depends on the sediment load, rate of retreat, and spacing of conduits (Rust and Romanelli, 1975, fig. 14). Continuity along the length of the valley seems likely, because meltwater must have been generated continuously during retreat, at rates that often increased when the rate of retreat increased. On the other hand, any blocks of stagnant ice that lingered south of the active ice margin would have prevented proximal coarse stratified drift from being deposited over the entire width of the valley. Furthermore, where the ice margin retreated by stepwise stagnation, coarse sediment would likely be concentrated in crevasses between shrinking ice blocks near the head of each stagnation zone and might be completely absent in the distal part of each zone. Many maps provide evidence of ice-block depressions and ice-walled channels in the glaciated Northeast, and many borehole records document sites where facies-2 fines directly overlie till or bedrock.

Studies of several lowlands or valleys in the glaciated Northeast have inferred that basal sand and gravel deposits are widespread and have interpreted those deposits as the products of multiple subaquatic fans. For example:

1. *Coastal Maine.*—Bingham (1981), Smith (1982, 1983), and Smith and others (1982) reported that, in much of coastal Maine, a discontinuous but widespread layer of sand and (or) gravel overlies bedrock and is, in turn, mantled by clay (facies 2) that was deposited in the marine waters that fronted the retreating ice and inundated the region for a time after deglaciation. Smith (1982, 1983) considered the basal layer to be ice-contact deposits or subaquatic outwash and ascribed it to numerous small meltwater streams or sheetflow that emerged from the base of the ice and (or) to a "tidal pumping" process wherein a floating ice shelf rises and falls with the tide, creating cyclical landward and seaward flows in the thin wedge of water beneath the ice and thereby winnowing the upper part of the underlying till and any sediment dropped from the floating ice. The stratigraphy described by Smith was based largely on exposures on low hills; he did not collect subsurface data to investigate whether an extensive sandy layer also was present beneath the thicker marine clay in major valleys. Reports on the ground-water hydrology of coastal Maine and New Hampshire generally do not mention such a thin, widespread aquifer.

2. *Rockaway River Basin, N.J.*—Canace and others (1993) presented an interpretive map that represents the vertical sequence of unconsolidated sediments in much of Morris County, N.J. The map and accompanying geologic sections indicate valley floors north of the terminal moraine to be underlain in most places by all three depositional facies, including a basal confined aquifer. Stanford (1988; 1989a, b) inferred the presence of a generally active ice margin retreating in a stepwise manner during deglaciation, and Stanford and others (1990) drew conceptual diagrams of aquifer geometry in New Jersey that indicate a nearly continuous blanket of coarse subaquatic-fan deposits atop bedrock in narrow valleys (fig. 9). The well logs and yields compiled by Canace and others (1993) demonstrate that such a basal layer is indeed virtually continuous in the Lockjaw River-Stony Brook valley northeast of Denville, N.J., a reach 5 miles long by 0.5 mile wide (fig. 4); water depth was about 125 feet. Logs of wells in nearby valleys, however, indicate that the basal aquifer can be very thin or absent locally or are too sparse to demonstrate continuity. Large areas in two broad valleys in eastern Morris County apparently lack a basal aquifer. Most well logs in Morris County do not suggest a progressive downward coarsening from facies 2 to facies 1, as would be expected under conditions of continuous deposition of subaquatic fans at a retreating ice margin.

3. *Appalachian Plateau, N.Y.*—MacNish and Randall (1982, pl. 1) inferred basal aquifers to be widespread in deep valleys of south-central New York and considered them to have resulted, in part, from retreat so rapid that coarse sediment could not build up to lake surface. Lake depths probably ranged from 150 to 500 feet. Well logs from this area (Randall, 1972) included 110 wells north of latitude 42°14' that penetrated through a substantial thickness of fine sediment (facies 2) into sand, gravel, or bedrock at depths greater than 100 feet. Of these, 21 percent did not penetrate a basal aquifer above bedrock; 56 percent reported a sharp contact of fine sediment over gravel, sand and gravel, or coarse sand; 17 percent indicated some downward coarsening, generally clay over "quicksand" (silt to fine sand) over gravel; and 8 percent penetrated fine sediment over "sand," which might have proved to coarsen further had drilling continued. Drilling generally stopped as soon as an adequate water supply was obtained, so thickness of the deep aquifer(s) cannot be estimated. The abundance of deep aquifers beneath fine sediments seems consistent with the hypothesis of subaquatic-fan deposition; the sparse evidence for systematic downward coarsening does not. Perhaps coarsening downward may be more widespread than indicated by the data cited, inasmuch as most logs were reported by drillers who, after drilling through many feet of non-water-yielding fine sediment, may have been little interested in subtle differences until they reached coarse layers that produced water.

4. *Central Massachusetts.*—Walker and Caswell (1977) inferred that a basal layer of sand and gravel occurs "widely but not everywhere" beneath the fine-grained sediments of glacial Lake Hitchcock in central Massachusetts. More detailed studies, however, have emphasized the absence of such a basal layer in particular localities. For example, in an area near the Connecticut River where the lake had been 250 to 400 feet deep, the fine-grained

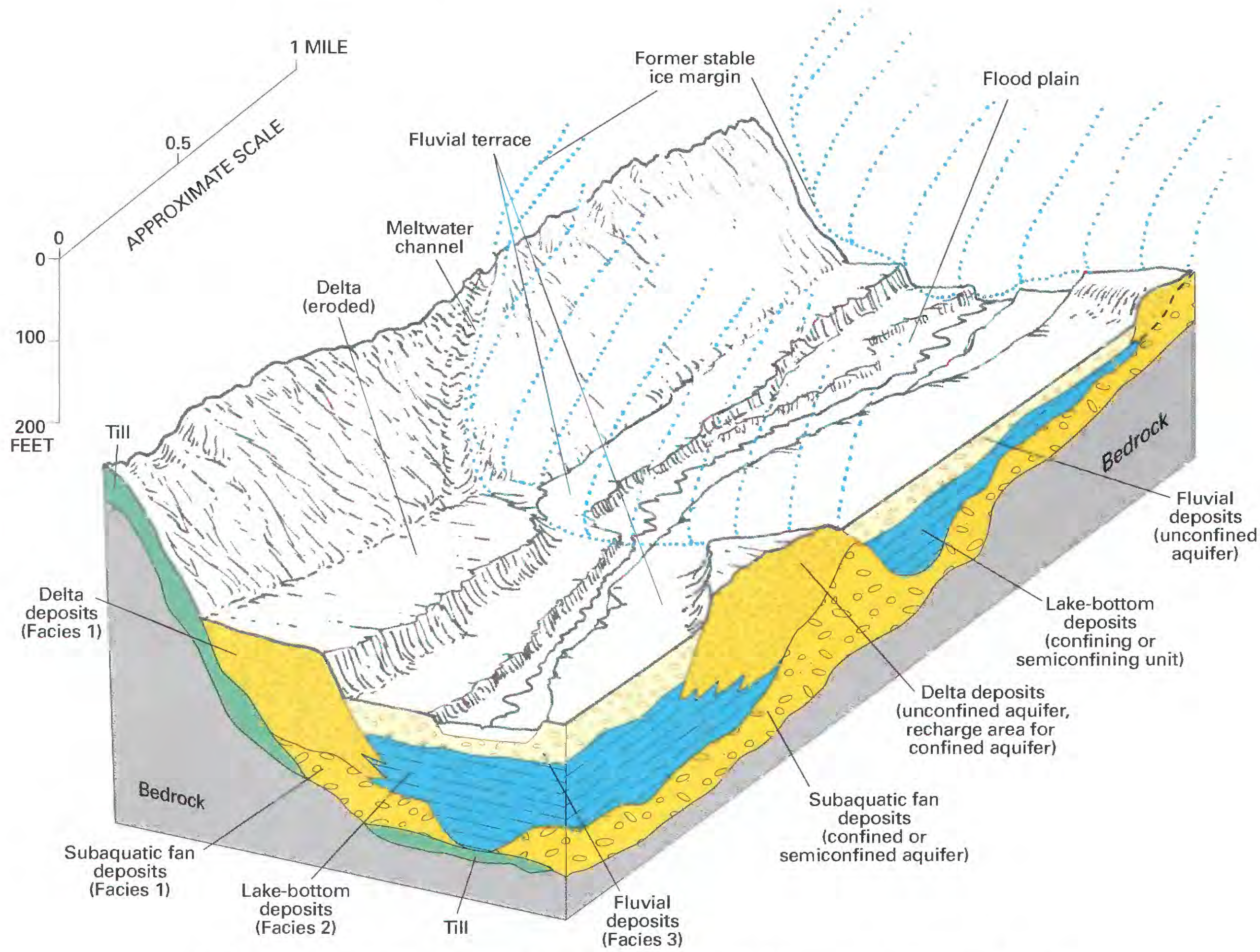


FIGURE 9.—Geometry of unconsolidated sediment typical of narrow valleys in New Jersey. (Modified from Stanford and others, 1990.)

sediments generally coarsen downward from silt and clay to coarse silt and very fine sand, but none of nine test wells penetrated a basal coarse aquifer (Hansen, 1986); a few municipal wells outside the area tested did penetrate a basal aquifer, however.

In two other valleys, by contrast, study of exposures along deeply incised streams has shown that proglacial subaqueous fans were uncommon.

1. *West Canada Creek, N.Y.*—Ridge (1985) described in detail the areal and vertical distribution of several glacial lithostratigraphic units exposed in the bluffs along West Canada Creek and its tributaries in Herkimer County, N.Y. (fig. 10). Most of the section consists of subglacial till, proglacial sediment flows that were generally derived from till or fine-grained lacustrine sediments, silty clay turbidites, and silt-clay varves that commonly contain drop pellets of unconsolidated sediment. The drift was deposited in a persistent proglacial lake(s) that varied in extent and elevation as West Canada Creek valley was repeatedly invaded by ice tongues from the west, southeast, and (more distantly) from the north. The diamicton units shown in figure 10 include till and proglacial sediment flows. Only a few sand or gravel units are present in this overwhelmingly fine-grained sequence. One of these, within the upper Newport beds (fig. 10), was identified as a subaqueous fan (Ridge, 1985, p. 108) or esker-subaqueous fan complex (Ridge and others, 1991, p. 1040 and appendix). The southeast (proximal) facies of that fan is as much as 40 feet thick and includes gently dipping boulder-cobble gravel, muddy gravel, very sparsely to moderately stony silty-clay diamicton, megarippled coarse sand to granule gravel, rippled and laminated fine-medium sand, and

muddy fine sand to silt with interbedded varves. Where the fan is best exposed (Ridge, 1985, sites 324, 111, 116), a tendency to upward fining is evident. The fan extends nearly 4 miles northwest from the inferred ice margin at the time of deposition (fig. 11), becoming thinner and grading to medium-fine sand interbedded with silty, sandy diamicton, and finally to fine sand interbedded with silty-clay rhythmites. The subaquatic fan was interpreted by Ridge (1985, p. 109) as a brief, unusual event that interrupted normal varve deposition in the proglacial lake. It immediately overlies the West Canada diamicton (fig. 10) and was deposited during retreat of the southeastern (Mohawk) ice tongue that deposited that diamicton. Evidently, the subglacial meltwater conduit that created the fan did not function continuously during the retreat of the ice tongue from the position of maximum advance shown in figure 11; instead, subglacial meltwater suddenly burst forth in this locale midway in the retreat. Otherwise, the stratigraphic record along West Canada Creek contains only sparse evidence of subglacial meltwater flow associated with the ice tongues that advanced and retreated at other times through the proglacial lakes in this locality. Most sand or gravel units recognized by Ridge (1985) are apparently small and (or) were interpreted as deltaic inwash or ice-marginal deltas. The ice lobes that deposited nearly all of the drift studied by Ridge advanced from the west and from the southeast; during the early stages of deglaciation, however, the outlet for meltwater was not ahead of these advancing ice lobes into the

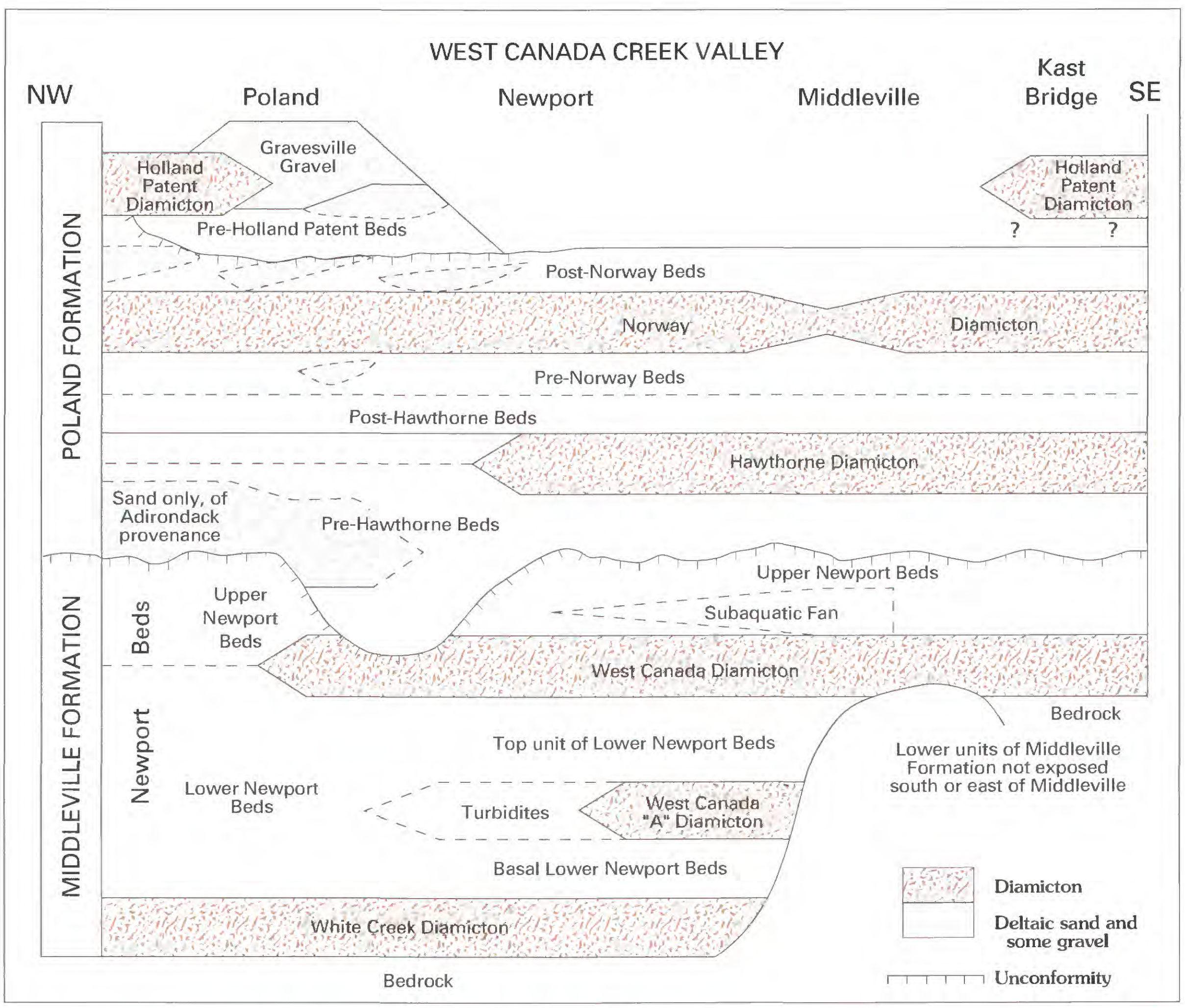


FIGURE 10.—Generalized stratigraphy in West Canada Creek valley downstream from Gravesville, N.Y. (Modified from Ridge, 1985, fig. 17.)

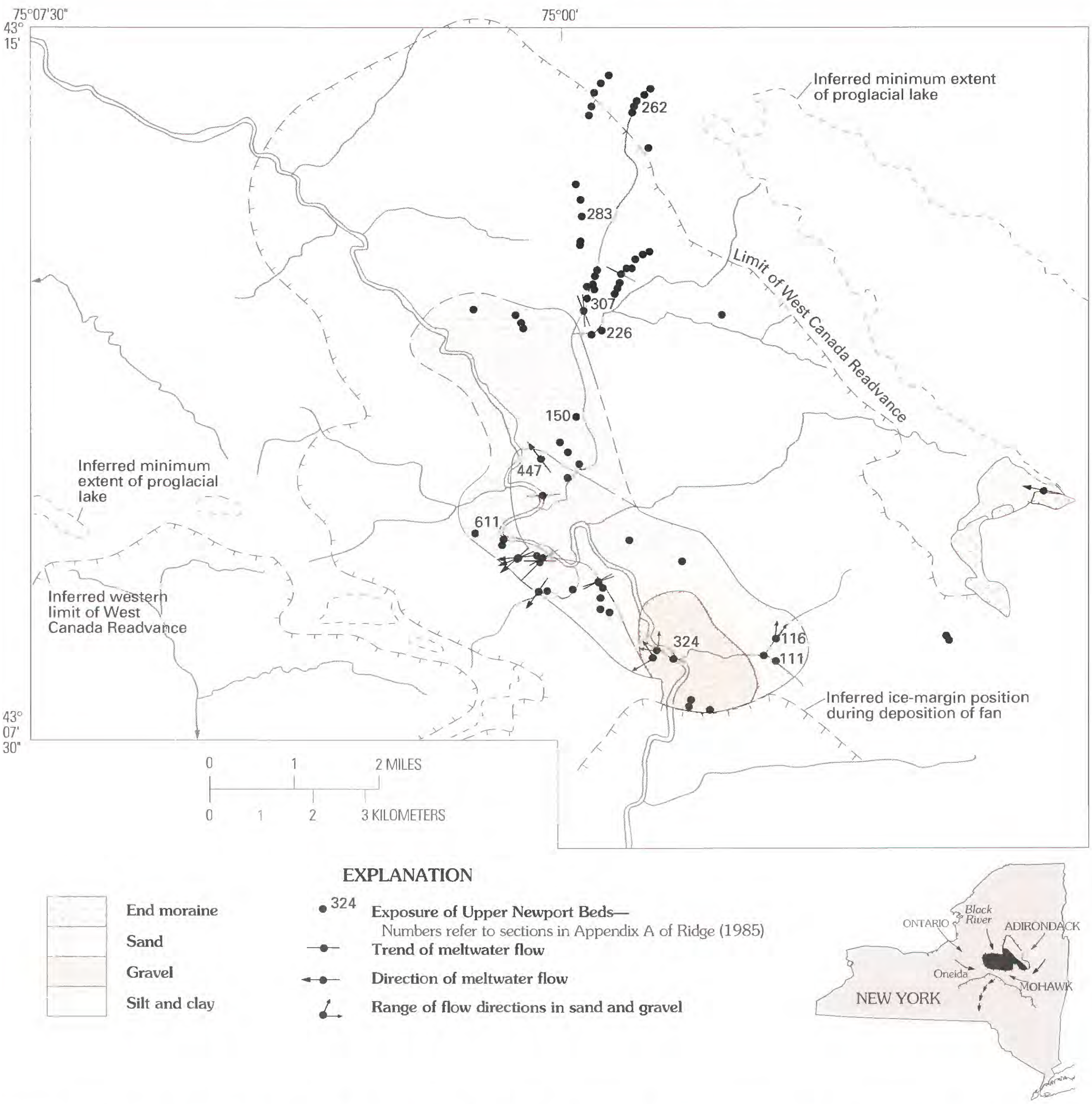


FIGURE 11.—Grain-size distribution and directions of meltwater flow during deposition of subaquatic fan within upper Newport lacustrine beds in West Canada Creek valley between Middleville and Newport, N.Y. (Modified from Ridge, 1985, fig. 20.) Inset map shows ice lobes, ice margin, proglacial lake in West Canada Creek valley, and meltwater drainage route during deposition of the diamicton that immediately underlies the subaquatic fan (fig. 10). (From Ridge, 1985, fig. 24.)

proglacial lake, but rather to the south, through subglacial tunnels to saddles in the Appalachian Plateau (inset map, fig. 11). Perhaps, therefore, regional meltwater drainage systems within these lobes did not generally continue to the ice margin in West Canada Creek valley, but instead followed the deeper Mohawk River valley and tributary valleys that led to the Plateau saddles. After deposition of the West Canada diamicton (fig. 10), the southeastern (Mohawk) and western (Ontario) ice tongues separated for the first time, allowing water to flow unobstructed by ice to the lowest nearby Plateau saddle (Ridge and others, 1991, fig. 12). A sudden decline in lake level of at least 225 feet resulted, which increased hydraulic gradients within the ice and may have temporarily retarded its retreat (Ridge and others, 1991, p. 1046), thus stimulating construction of a subaquatic fan into the proglacial lake that had become the lowest outlet for meltwater. West Canada Creek valley would always have been the lowest outlet for meltwater beneath the Black River lobe (fig. 11), which advanced southward to the northern end of the valley where exposures of interlayered gravel and sand were attributed to esker-fed subaquatic fans from that lobe by Ridge and others (1984, p. 248).

2. *Sandy River, Maine.*—Several exposures along the Sandy River near New Sharon, Maine, have been described by Caldwell (1959, 1960), Caldwell and Weddle (1983), and Weddle and others (1989). They consist primarily of massive sandy and silty diamictons, stratified diamicton with thin layers of silt, clay, and sand, rhythmic silt and clay with thin layers of sand and diamicton, and one discontinuous layer of medium sand several feet thick. This stratigraphy is rather similar to that studied by Ridge (1985) and was similarly interpreted by Weddle and others (1989) as till and subaquatic debris flows deposited amid fine-grained sediment as an ice lobe advanced into a proglacial lake. Subglacial meltwater flow was sufficient to deposit only one sand layer, perhaps because the ice at New Sharon was a southwest-flowing sublobe of an ice sheet that flowed primarily southeastward down the Kennebec valley where subglacial meltwater flow may have been concentrated.

In summary, the concept that multiple subaquatic fans deposited coarse sand and gravel early in deglaciation over large areas on the floors of proglacial water bodies is an attractive hypothesis that implies widespread, laterally continuous, basal aquifers, especially where proglacial water bodies were deep and the ice margin could retreat by calving without leaving large stagnant ice blocks. This hypothesis is supported by borehole data in some places, but borehole data and exposures also demonstrate that basal aquifers differ widely in thickness and are absent in many places, and geomorphic evidence of stagnant ice blocks also is widespread. The coarse sediment carried by subglacial conduits, whether deposited in ice tunnels or in subaquatic fans at tunnel mouths, seems commonly to be concentrated in narrow, channel-like strips, as discussed in the next sections.

MELTWATER CHANNELS AND COLLAPSE AT THE ICE MARGIN

Two geomorphic features, ice-channel fillings and collapsed ice-contact faces, are typical of many ice-marginal positions or morphosequence heads and serve to enhance the potential for productive aquifers at these locations. The nature and hydrologic significance of both features are described below and illustrated by several examples.

Subglacial or englacial meltwater channels are important avenues of sediment transport that often become sites of coarse-grained sediment deposition as well. Linear or sinuous ridges composed of sand or gravel have been observed in areas freshly uncovered by the retreat of modern glaciers and have been interpreted as recording the former courses of meltwater channels below or within the ice (Price, 1964, 1966; Goldthwait and Mickelson, 1982). Theoretical studies (Röthlisberger, 1972; Shreve, 1972, 1984, 1985) have offered explanations for the formation of subglacial conduits and deposition of sediment therein. Gravel ridges deposited in subglacial conduits or tunnels are known as "eskers," whereas the more general term "ice-channel fillings" also includes deposits in narrow ice-walled channels open to the sky.

In the glaciated Northeast, ice-channel fillings were deposited near the retreating ice margin in many localities. Some topographic maps reveal large, commonly flat-topped constructional landforms from which one or more linear or sinuous ridges extend generally northward (fig. 12). Such tadpole-shaped landforms are readily interpreted as ice-contact deltas fed by meltwater that flowed through tunnels or crevasses in the ice (feeder channels) in which sediment also accumulated and is now preserved as ice-channel fillings (Lougee, 1938; Hartshorn, 1967, p. 31, map; Thompson, 1982, p. 221; Stone and others, 1985, p. 569; Warren and Stone, 1986, fig. 6; Larsen and Koteff, 1988, p. 122; Thompson, 1989, p. 178). Topographically distinct ice-channel fillings, some as long as 1.5 miles, extend headward from many morphosequence heads in southern New England (Koteff, 1974, p. 134). Elsewhere, exposures and (or) borehole logs have revealed narrow deposits of coarse sand and gravel that seem to have a linear, ridgelike form but are mantled by sand and, in some localities, further mantled or bordered by fine-grained sediment. The coarse sand and gravel deposits are interpreted to be the product of ice-tunnel or proximal tunnel-mouth deposition (facies 1), overlapped by subsequent deltaic and lacustrine sediment deposited as the ice margin retreated further.

Examples from Canada are described by Parsons (1970) and Rust and Romanelli (1975). Examples from the Northeastern United States are presented in several guidebook articles, including Thompson (1983, stops 1–2 near Augusta, Maine), Mayewski and Birch (1984, stops 1–2 near Farmington, N.H.), Koteff and others (1984, stop 2 near Nashua, N.H.), Larsen (1987a, stop 5 near Sharon, Vt.; 1987b, stops 7–8 near Northfield, Vt.), Larsen and Koteff (1988, stops 2–3 near Brattleboro, Vt.), and Connally and others (1989, stop 2 near Middletown, N.Y.).

Some ice-channel fillings are merely deltas built in narrow lakes within ice crevasses. LaSala (1961) described ice-channel fillings whose linear segments with angular intersections suggest crevasse systems in ice sheets, and whose stratigraphy resembled nearby deltas. Stone and others (1985) observed 11 feet of coarse-sandy foreset beds exposed beneath 12 feet of coarse gravel topsets in a sinuous ice-channel filling at the head of a kame delta in Southington, Conn. The foresets indicate that, late in the depositional history, the proglacial lake extended into a crevasse at this location. The foresets may, however, be underlain by

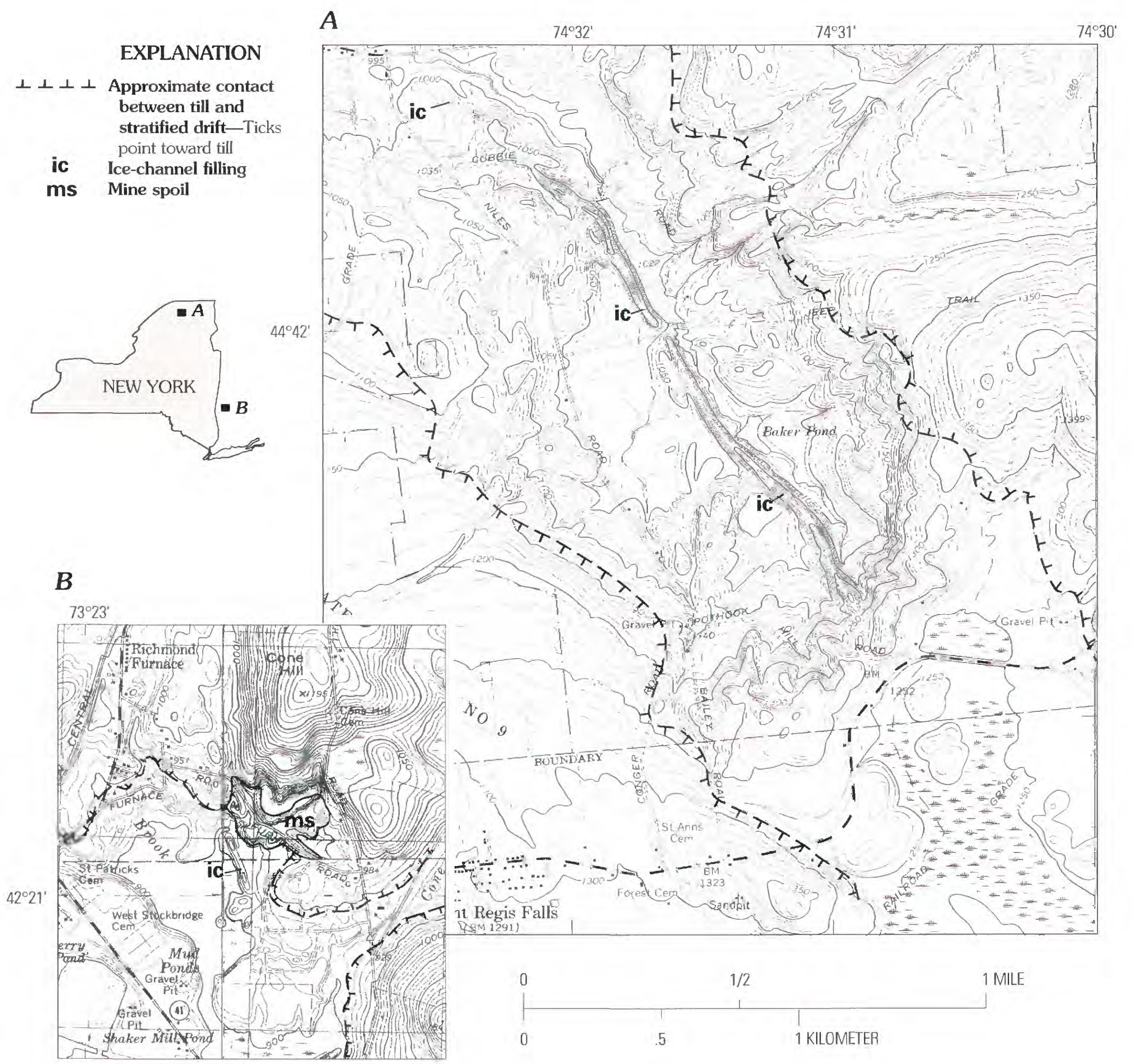


FIGURE 12.—Examples of ice-channel fillings inferred to mark former subglacial channels along which meltwater transported sediment to adjacent ice-marginal deltas. *A*. From U.S. Geological Survey St. Regis Falls, N.Y., 7.5-minute quadrangle, 1964, 1:24,000. *B*. From U.S. Geological Survey State Line and Stockbridge, Mass., 7.5-minute quadrangles, 1975, 1:25,000.

coarse sand and gravel deposited earlier along a subglacial tunnel that conveyed sediment to the kame delta, just as early subaquatic-fan deposits presumably underlie part of the delta. Thus, some ice-channel fillings may owe their sinuous form to buried eskers that predate the sediments exposed at land surface.

Even more common than ice-channel fillings at the heads of deltas or morphosequences are steep, highly irregular proximal slopes characterized by gravel and scattered boulders and by knoll and kettlehole topography that reflects collapse of the proximal stratified drift as buried ice melted. The gravel exposed at land surface may or may not extend down to the water table beneath the proximal slopes, but saturated gravel can be expected immediately north of the toe of that slope, buried beneath younger stratified drift. These buried proximal deposits might be simply the toe of a collapsed ice-contact face, or they might be feeder-channel deposits that lack topographic expression because they are thin or deeply buried. Several examples of collapse and (or) feeder channels at former ice-marginal positions are described below.

1. *Arlington, Vt. (fig. 13).*—The topographic setting in figure 13 is typical of many morphosequences. The mesalike feature at *A* is probably a kame delta; its south face was not exposed, but fine sand predominated in exposures at *B*. The nose at *C*, on the north face, was capped by at least 15 feet of coarse gravel, including openwork cobble gravel; very coarse sand was exposed lower on the 70-foot-high slope. The surface at 800-foot altitude north and east of the kame delta is underlain by lake-bottom deposits, predominantly silt but locally capped by coarse, pebbly sand. No information is available on depth to bedrock in this locality, nor on whether the coarse sediments that constitute the north face of the kame delta extend northward beneath the alluvium and lake-bottom deposits, as might be expected.

2. *Danielson, Conn. (fig. 14).*—The Quinebaug River at Danielson is bordered by the distal or downvalley part of a morphosequence designated Qq5 in figure 14. Exposures in this unit on both sides of the river reveal a few feet of pebbly, coarse sand over silt and very fine sand, typical of the downvalley end of an ideal morphosequence. The fine sediment was deposited in a lake that expanded northward in the Quinebaug valley as the ice retreated. The downvalley end of morphosequence Qq6, 2 miles to the north (fig. 14), is similar in stratigraphy and altitude to the downvalley end of Qq5; the same lake was base level for each, and the two depositional profiles merge near Danielson. Excavations in East Brooklyn, just west of Danielson, revealed very coarse gravel in a linear ridge bordered and overlapped by the fine-grained sediment of Qq5. This ridge was interpreted by Randall and Pessl (1968) as an ice-channel filling, part of the next older morphosequence, Qq4 (fig. 14). Although morphosequence Qq4 heads farther north along both sides of the Quinebaug lowland, near the valley axis this ice-channel filling represents the most headward part of the sequence. A municipal well was drilled on the axis of the gravel ridge (fig. 14); its log indicates 80 feet of gravel and coarse sand from which 350 gallons per minute were pumped (Thomas and others, 1966). No other buried feeder-channel deposits headward of the mapped extent of morphosequence Qq4 were penetrated by boreholes. Such deposits might be expected, however, especially north of the proximal end of the ice-channel filling and kame terrace east of Long Brook (fig. 14), as inferred by Randall and others (1966).

3. *Cromwell, Conn. (fig. 15).*—Extensive surficial sands and gravels were deposited near the Cromwell-Rocky Hill town line in central Connecticut as coalesced deltas (Langer, 1977; Stone and others, 1982; Koteff and others, 1987) in water ponded behind older morphosequences downvalley (London, 1985). Excavations and logs of boreholes (Ryder and Weiss, 1971; Grady and Handman, 1983) revealed that the deltaic sands are underlain by coarser sand and gravel that seems to represent a feeder-channel system. A large pit has exposed and largely removed the stratified drift between Dividend Pond and the railroad (fig. 15). The northeast end of the pit, near the end of the dashed extension of Dividend Road (fig. 15), revealed coarse, heterogeneous proximal ice-contact deposits, including vertical and chaotic bedding, resedimented diamicton layers, steeply dipping cobble-boulder gravel, and horizontal silty fine sand; Stone and others (1982) report up to 15 feet of collapse. Bedrock is exposed at the Dividend Pond dam (Deane, 1967) and might not be far below the pit floor. The coarse ice-contact deposits were overlain by sand that coarsened upward, contained depositional structures typical of deltas, and was capped by 10 feet of gravel topset beds (Stone and others, 1982; Koteff and others, 1987). About 2,000 feet northeast of the pit, beneath a younger terrace surface that formed

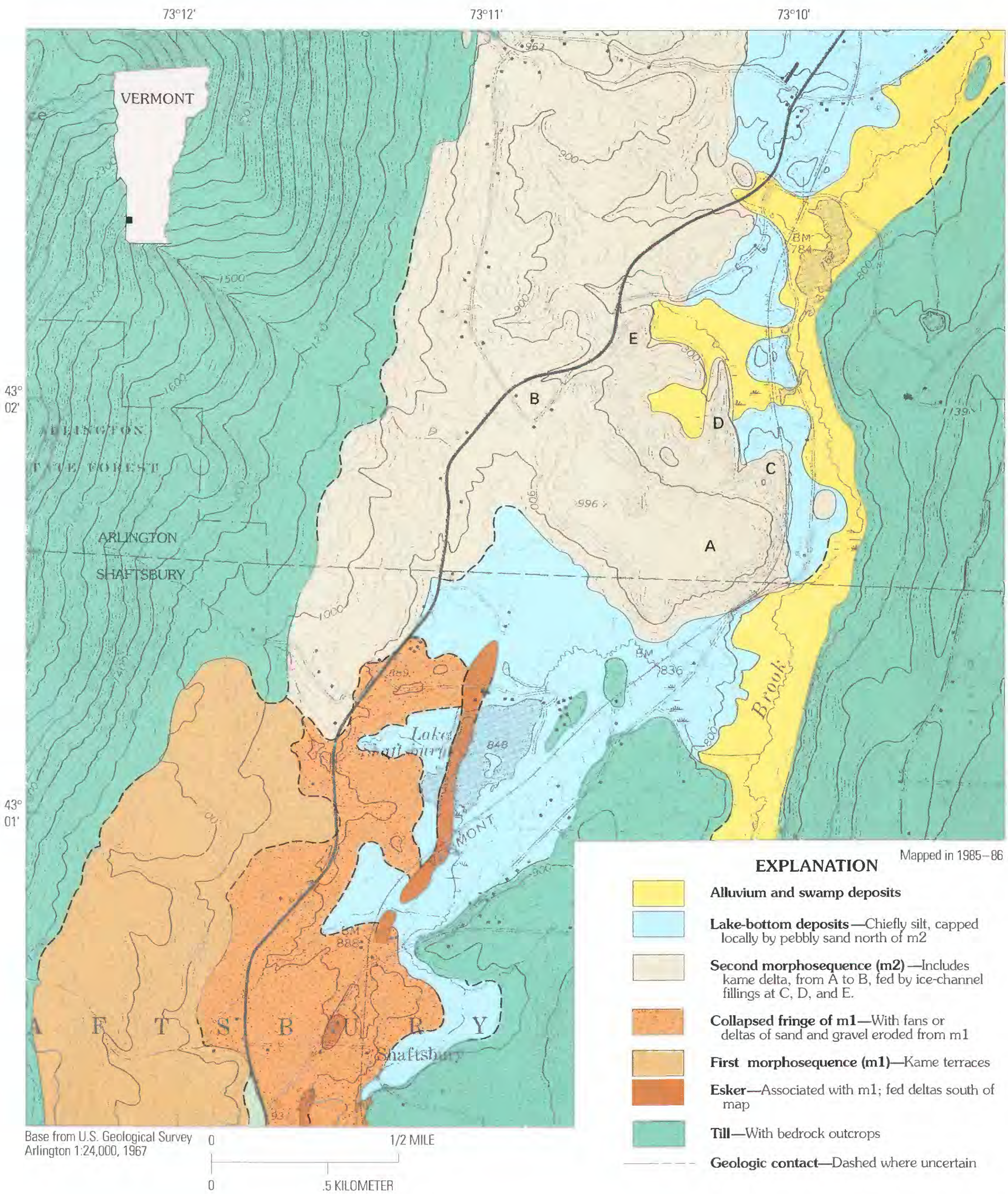


FIGURE 13.—A collapsed morphosequence head in the town of Arlington, Vt.

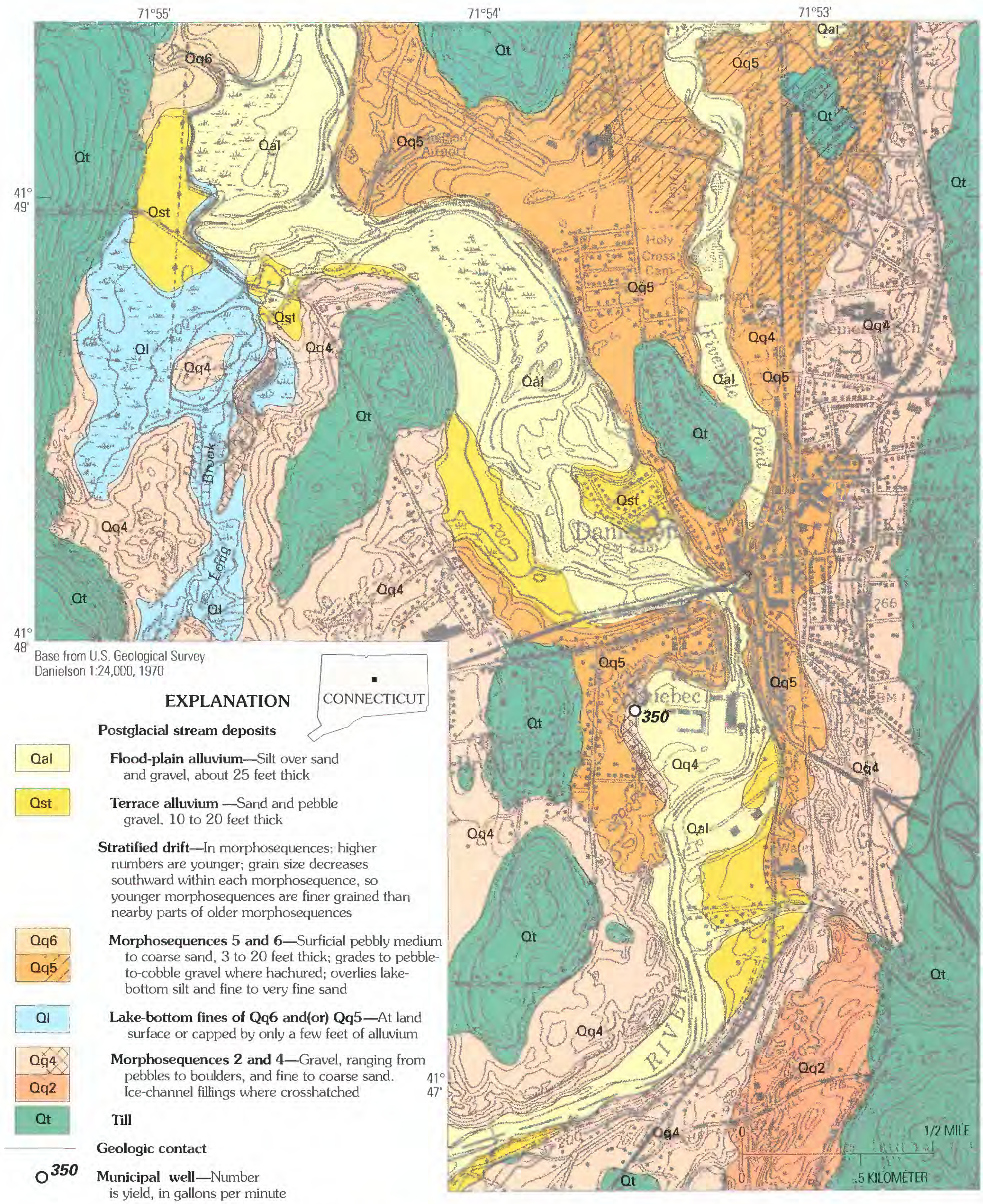


FIGURE 14.—Feeder channels at head of morphosequence near Danielson, Conn. (Modified from Randall and Pessl, 1968.)

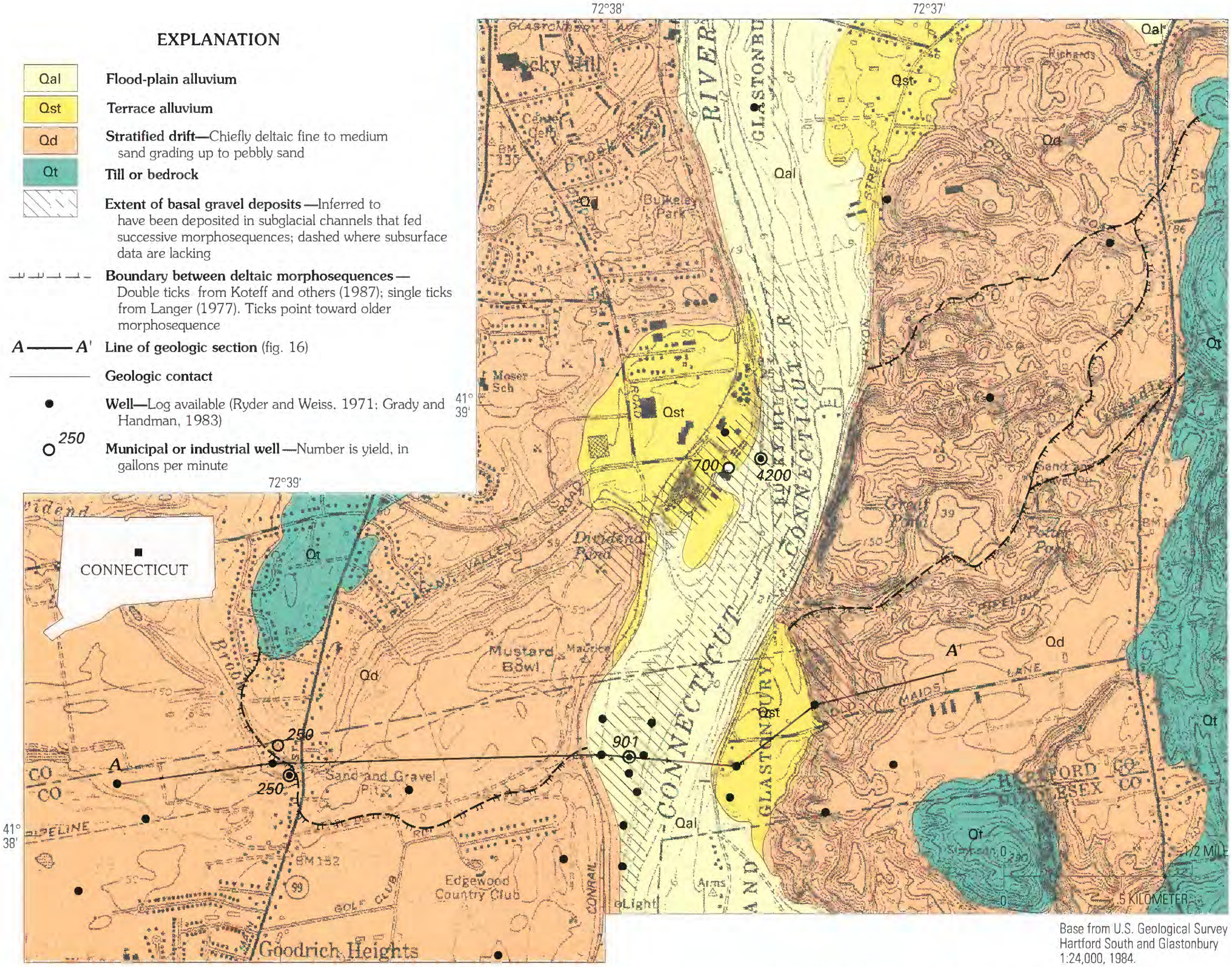


FIGURE 15.—Feeder channels at head of morphosequence in Cromwell, Conn., and adjacent towns. Surficial geology from Deane (1967) and Langer (1977).

when the Connecticut River was diverted to its present course, two wells obtain yields of 700 and 4,200 gallons per minute from sand and gravel that may well have been deposited as part of the same ice-channel system that was responsible for the ice-contact deposits exposed in the pit.

About 0.5 mile south and slightly east of the pit, several municipal test wells were drilled on the Connecticut River flood plain. Most penetrated the following materials:

Very fine sand and silt, 25 feet thick; overlies

Sand, varying from very fine to medium-fine to medium-coarse, 20 to 60 feet thick; overlies

Sand and gravel, water-yielding, 30 to 60 feet thick.

The upper 25 feet, at least, is alluvium deposited by the Connecticut River, which has aggraded as sea level rose in Holocene time (Upson and Spencer, 1964). The basal sand and gravel unit seems to form a buried ridge, as does the overlying fine to coarse sand (fig. 16), inasmuch as a well just west of the river penetrates only very fine sand to silt, and a well just east of the river also penetrates fine sand and silt above fine-coarse sand. The basal sands and gravel penetrated by these test wells are probably slightly older than that exposed in the pit near Dividend Pond; they could have been deposited along a channel that fed the slightly higher, earlier delta at the Edgewood Country Club, south of the ice-marginal position shown in figure 15. East of the Connecticut River near Old Maids Lane (fig. 15), two ice-channel fillings occur side-by-side along the same inferred ice margin; they are probably the product of other feeder channels active at about the same time. The sand and gravel from land surface to 35 feet below sea level just east of the river may have been deposited during incision of the river through the deltaic stratified drift, when base level downstream was deeper than at present (London, 1985).

The data cited above constitute reasonably convincing evidence for the presence in this locality of linear gravel aquifers deposited atop bedrock at the ice margin within or at the mouths of subglacial tunnels. An inferred distribution of these deposits is sketched in figure 15. That figure also illustrates the difficulty of predicting the distribution of buried aquifers in the absence of borehole records. Morphosequence heads are not topographically distinct and were differently interpreted by different investigators, probably because they represent only small changes in base level within a continuous depositional process, and each may include several successive deltas deposited as the ice margin retreated (Koteff and others, 1987). Except for the two ice-channel fillings east of the Connecticut River, the topographic map provides no clue as to where buried channels may be.

4. *Mount Warner, Mass. (fig. 17).*—The Mount Warner delta, one of several prominent kame deltas deposited at the retreating ice margin in proglacial Lake Hitchcock, has been described by Webb (1957), McIlvride (1982), Hansen (1986), and Koteff and others (1987, stop 9). Its location in midvalley, remote from any tributary or ice-marginal drainage, requires that it was fed by one or more subglacial or englacial channels. On the southwest face, delta foresets of coarse sand with lesser amounts of pebble gravel and fine sand (Koteff and others, 1987) become finer downdip and interfinger with lake-bottom varves (Webb, 1957). The north face is an ice-contact slope; collapsed beds and possible flowtill have been reported. At the bottom of the slope are two municipal wells (fig. 17) that yield 550 and 1,080 gallons per minute from sand and gravel beneath 140 to 190 feet of fine-grained stratified drift (Hansen, 1986). Three nearby wells also tap this basal aquifer. The aquifer is probably a proximal subaquatic fan at the mouth of a feeder channel, but might instead be collapsed coarse ice-contact sediment at the head of this deltaic morphosequence.

5. *Other Examples.*—Collapse of ice-contact margins has created many other aquifers. For example, in Proctor, Vt., a municipal well penetrates 72 feet of sand and gravel beneath 52 feet of fine sand and clay (Willey and Butterfield, 1983) just north of the steep proximal slope of a kame terrace mapped by Stewart and MacClintock (1970). The proximal slope coincides with an abrupt southward decrease in width and depth of the bedrock valley. In Simsbury, Conn., collapse along the east side of a deltaic morphosequence is indicated by multiple ice-block depressions and by wells farther east that tap sand beneath younger fine-grained sediments (fig. 18). The collapsed margin coincides with the side of a bedrock valley (Randall, 1970).

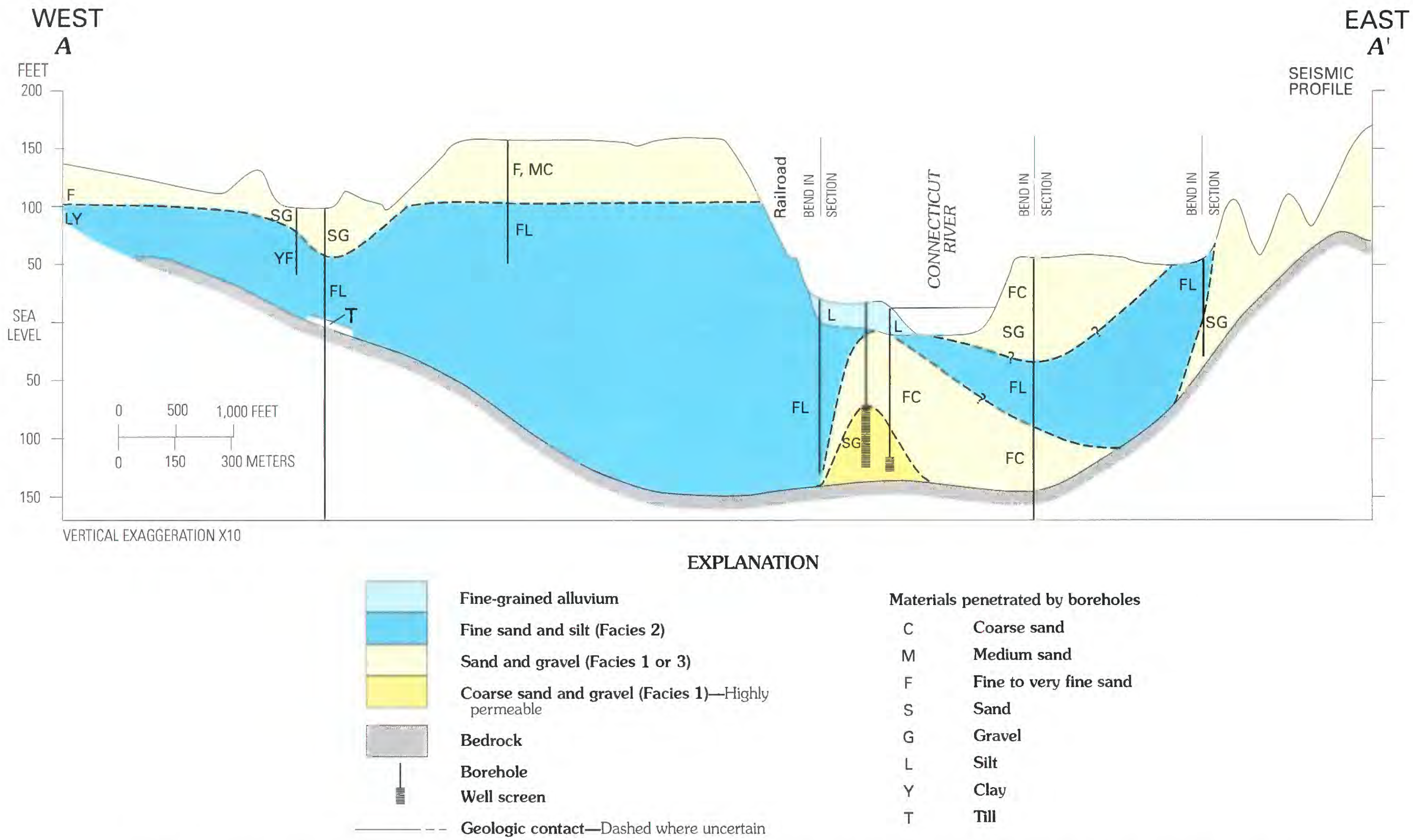
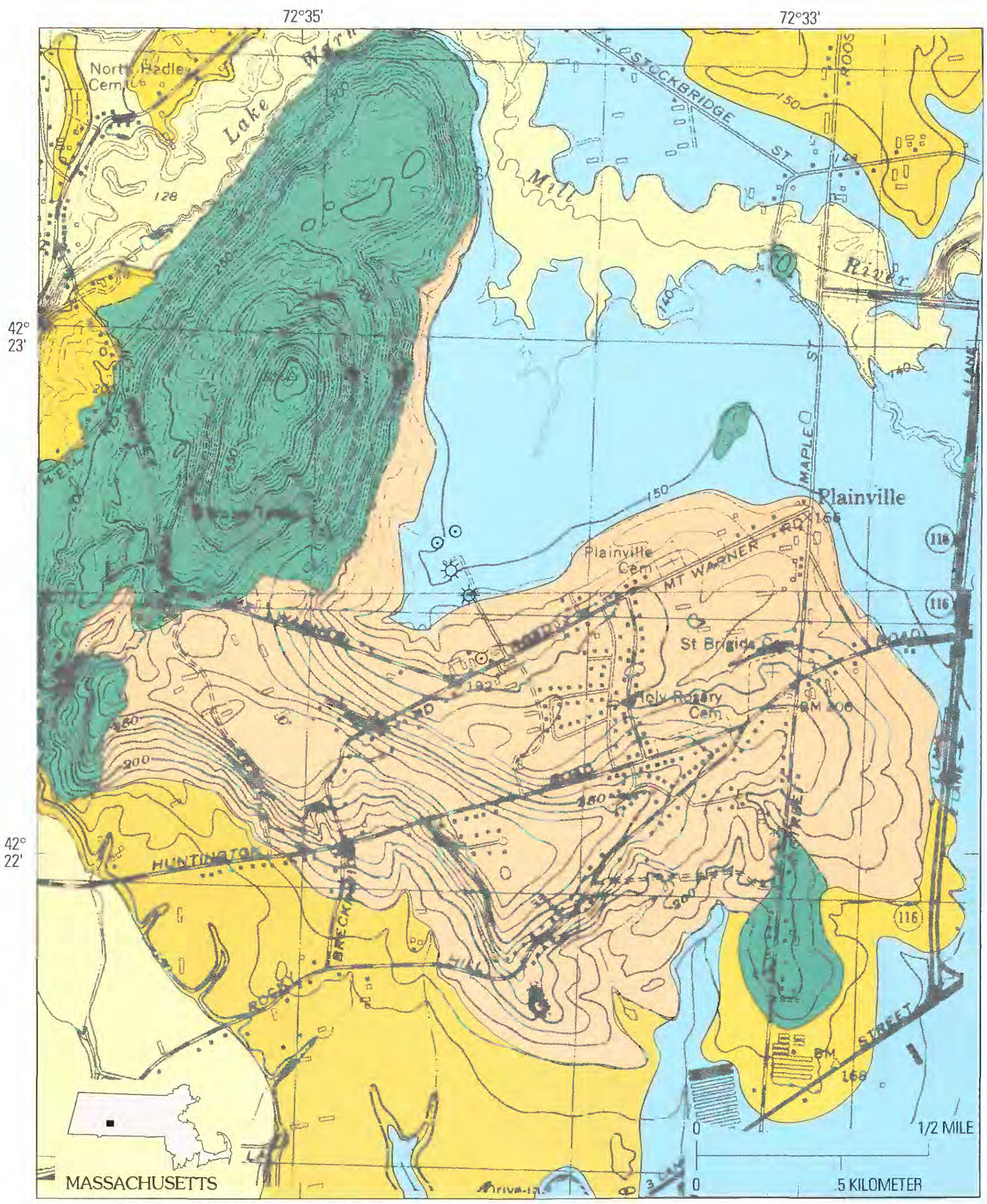


FIGURE 16.—Geologic section across the Connecticut River valley in Cromwell and Glastonbury, Conn. Location of section shown in figure 15.



Base from U.S. Geological Survey Mt. Toby (1971) and Mt. Holyoke (1979) 7.5-minute quadrangles, 1:25,000

EXPLANATION

Till and bedrock

Mount Warner kame delta—Graded to glacial Lake Hitchcock; exposures reveal 6 feet of gravel over more than 50 feet of coarse sand

Municipal production well—Taps coarse sand and gravel beneath fine-grained sediment

Well

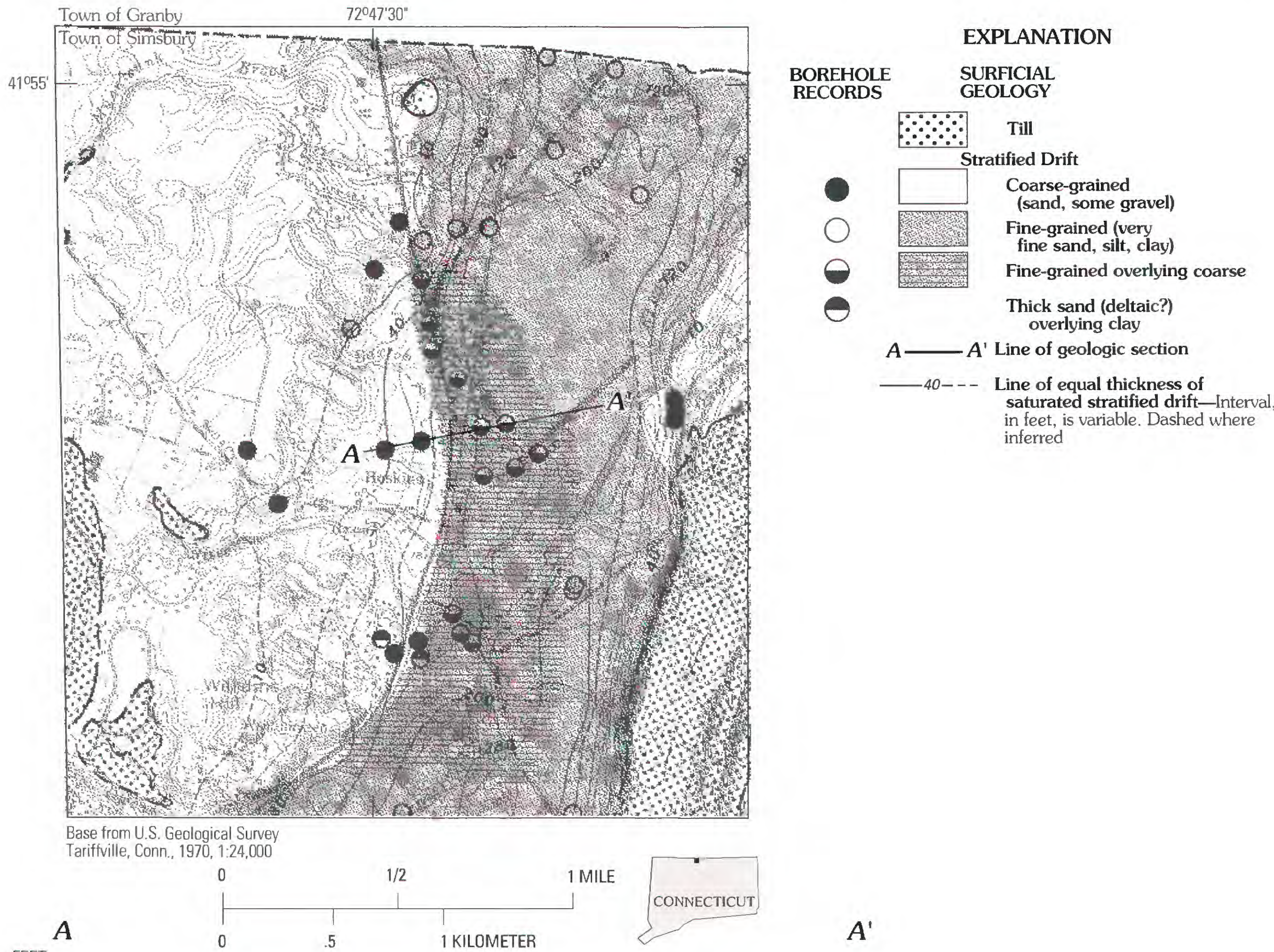
Lake-bottom varved clay, commonly grading downward to silt or silty very fine sand, and overlain by:

Little or no surficial sand—Clay is at or near land surface

Sand, fine to coarse—Generally 10 to 30 feet thick, in part fluvial and dune deposits on former lake floor

Alluvium—Silt and sand on terraces and flood plains incised below former lake floor

FIGURE 17.—The Mount Warner delta in the town of Hadley, Mass. (Adapted from Jahns, 1951; Balk, 1957; Stone and others, 1979; and Hansen, 1986.)



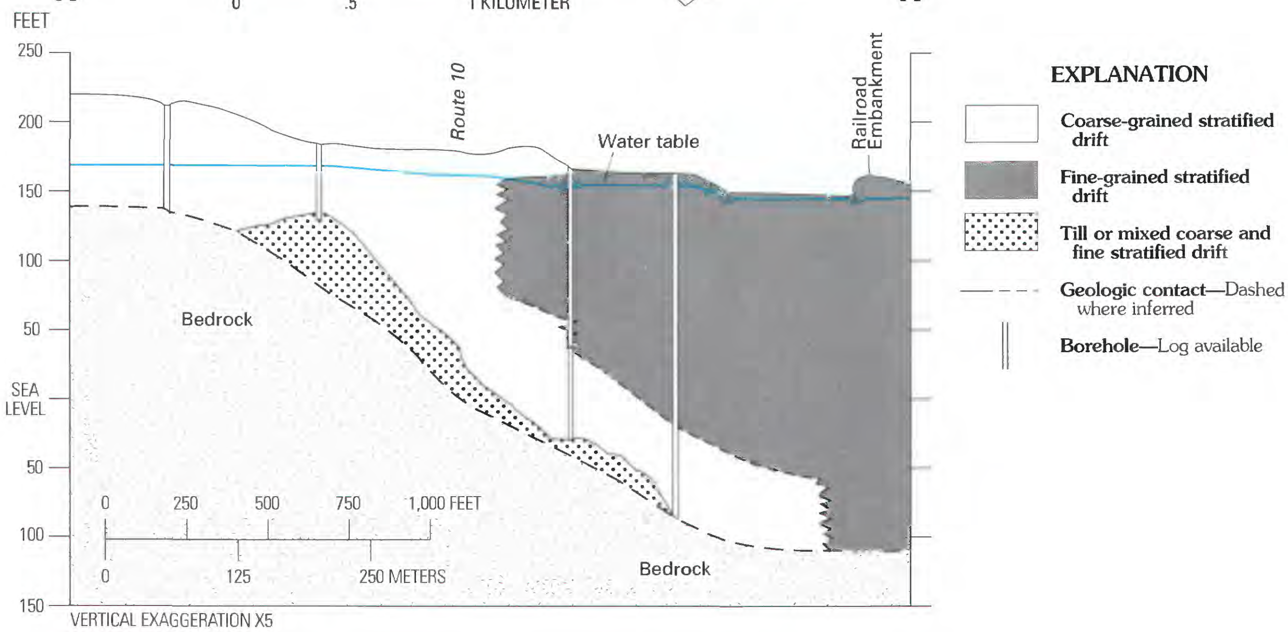


FIGURE 18.—Buried aquifer along collapsed margin of morphosequence in Simsbury, Conn. (From Melvin and Bingham, 1991.)

Other examples of proximal ice-channel fillings also can be cited. An esker, clearly depicted along Dolf Brook on the Hopkinton, N.H., topographic quadrangle, once fed a delta in a small proglacial lake and has been tapped by a municipal well (Hildreth and Moore, 1993, p. 19). Eskers in the towns of Northfield, East Montpelier, Norwich, and Hartland, Vt., are bordered and partly buried by younger and finer grained sediments and yielded large water supplies to test wells (Hodges and others, 1976a, 1976b). Ice-contact deposits form a buried ridge that extends for several miles beneath younger sediment along the Chenango and Susquehanna River valleys near Binghamton, N.Y., and are tapped by several municipal wells (Randall, 1986a).

In summary, ice-channel fillings that extend north from distinct morphosequence heads are likely to consist of coarse sand and gravel. One or more feeder channels should be anticipated on the proximal side of each morphosequence head or ice-margin position, in addition to collapsed coarse sediment at the base of proximal ice-contact slopes. Morphosequence heads can be difficult to identify with certainty, however, especially where hummocky stratified drift extends more or less continuously for several miles along a valley; and if deposits along feeder channels are covered by younger sediments, their location may not be marked by topographic ridges. Pumping tests and oriented resistivity soundings can be used to identify the orientation of buried aquifers that are not apparent from surface topography.

REGIONAL MELTWATER DRAINAGE SYSTEMS

The previous section pointed out that some of the coarsest sand and gravel transported by meltwater was deposited along narrow subglacial channels where they approached or reached the ice margin. This section presents evidence that large meltwater channel systems were spaced several miles apart, generally followed the lowest valleys and saddles on the bedrock surface, generally continued to function along the same route as the retreating ice margin encroached upon their lower reaches, and resulted in concentration of coarse stratified drift in semicontinuous narrow bands at least several miles in length, each perpendicular to a succession of temporary ice margins. Seasonal meltwater drainage systems must commonly have headed a few hundred miles north of the ice margin. Near the margin, meltwater flowed for at least several miles through tunnels at the base of the ice where it entrained sediment and, farther downstream, deposited some of that sediment. Coarse sand and gravel are likely to be more abundant, better sorted, and more continuous along persistent trunk channels that drained large watersheds on the ice sheet than in intervening areas where less meltwater was available. Several examples of nearly continuous linear sand and gravel deposits (facies 1) that are ascribed to persistent regional meltwater channel systems are described on the following pages.

ESKER SYSTEMS IN MAINE

The most obvious evidence of extensive subglacial channels in the glaciated Northeast are the esker systems of Maine, whose distribution is shown on a statewide map (Thompson and Borns, 1985) and on many geologic and geohydrologic maps of larger scales. The esker systems have been described in detail by Stone (1899) and more briefly in many subsequent publications. Eskers typically consist of medium to fine gravel, capped by coarser gravel (Leavitt and Perkins, 1935). In form, they are typically single steep-sided, sharp-crested ridges on the order of 60 feet high and 500 feet wide (Shreve, 1985). They commonly extend with only short interruptions for several tens of miles. In southern Maine, where local relief is less than 1,000 feet (Denny, 1982), they are subparallel, spaced 5 to 15 miles apart, and commonly branched like river systems. They tend to follow valleys for several miles, occupying only part of the valley-bottom width, then cross low saddles, thereby maintaining a gross orientation that reflects the slope of the ice surface at the time of deposition (Shreve, 1985). The large, complex Pine River esker system in eastern New Hampshire (Goldthwait, 1968) is a continuation of this pattern. In west-central Maine, where local relief generally exceeds 1,000 feet, eskers follow valleys consistently, seldom crossing saddles, and commonly occur as a series of discontinuous short segments. Although eskers are typically sharp-crested ridges, they locally grade into broad hummocky forms termed "reticulated kame fields" (Leavitt and Perkins, 1935) or "multiple-crested ridges" (Shreve, 1985). Eskers also grade into broad ridges with flat tops and, where the meltwater channel bordered a valley wall, into kame terraces. Shreve (1985) explained several variations in form as well as local discontinuities near saddles in terms of glacier physics—specifically, heat generated within the meltwater by internal friction interacting with changes in the pressure and melting temperature of the ice as subglacial tunnels ascend or descend in elevation or as the ice sheet crosses obstacles such as hills. These changes in esker form are reflected on the map of Maine by Thompson and Borns (1985) in that some of the linear stratified deposits that align with eskers are mapped as other ice-contact landforms.

Most esker systems in Maine terminate in deltas, and some are mantled by deltas at several separate points along their lower reaches (Leavitt and Perkins, 1935). The multiple deltas must have formed sequentially, which led Thompson (1982) to conclude that the eskers were built in successive segments. Alternatively, Shreve (1985) argued that continuity of the Katahdin esker system over 90 miles and a downstream increase in size indicate that the entire system was deposited simultaneously. He explained some features that had been thought to indicate sequential deposition—intervals of nondeposition, changes in pebble lithology that correspond to underlying bedrock units—in terms of glacier physics. Perhaps each subglacial meltwater conduit system initially functioned throughout its length, but its lower reaches were progressively abandoned as the ice margin retreated, and deltas were built at successive ice-margin positions by the continuing sediment flux along the conduit.

Eskers in Maine are commonly significant aquifers, inasmuch as they consist predominantly of coarse sand and gravel and tend to follow valleys. A few esker reaches are largely unsaturated, however, because they are perched above the present stream grade on the valley sides or where the esker crosses a saddle. Apparent gaps between topographically distinctive esker segments are likely to result from (a) deposition of enough younger sediment along the valley to completely bury low-lying esker reaches, or (b) postglacial stream erosion that planed off the crest of the esker flush with an adjacent terrace or flood plain, or (c) gradation of a typical sharp-crested esker into some less distinctive landform. Each of these explanations implies that the sand and gravel deposited along the former meltwater channel is likely to be continuous between, as well as within, visible esker segments and probably constitutes a continuous linear aquifer. Gaps between esker segments also can represent reaches of the subglacial conduit system in which sediment was not deposited, but Shreve (1985) pointed out that such reaches occur chiefly where the eskers cross saddles. Where an esker is overlapped by younger sediment, it can be tapped by wells alongside the exposed esker as well as on or in line with it.

An esker in the town of Oxford, west-central Maine, is illustrated in figure 19. Several esker segments, mapped as ice-contact deposits (Prescott, 1968; Morrissey, 1983), extend in slightly sinuous alignment along the Little Androscoggin River valley and a tributary. The segments lie along or slightly west of the thalweg of the bedrock valley (Morrissey, 1983, pl. 2) and are less than 500 feet wide at land surface. Most of the valley is filled with silt, clay, and very fine sand (facies 2) overlain by fine-sandy outwash (facies 3), as illustrated in the geologic section in figure 20*A*. Just north of the Little Androscoggin River, a pit (identified in fig. 19) revealed sand overlying marine clay that, in turn, overlapped the esker. A municipal well, drilled just north of an esker segment south of the river, obtained 325 gallons per minute from 52 feet of sand and gravel underlying 30 feet of fine sand and clay (Prescott, 1967). Test holes and a geophysical survey described by Haeni (1995) led to the interpretation shown in figure 20*B*—a buried esker segment thinly mantled and bordered by fine-grained sediment. Prescott (1968) inferred that this esker constitutes a nearly continuous aquifer more than 10 miles long and 1,000 feet or more wide, including the esker flanks and gaps between exposed segments where the esker is buried beneath younger fine-grained sediments.

ESKERS IN THE CONNECTICUT RIVER VALLEY

Esker segments have been identified along two reaches of the Connecticut River valley and also along the Passumpsic River valley, which continues the straight northeast-southwest trend followed by the Connecticut River valley farther south. Larsen and Koteff (1988) reported at least 10 successive ice-contact, apparently esker-fed deltas between the Massachusetts State line and Putney, Vt., a distance of 15 miles; Moore and others (1994) identified a buried esker beneath fines just north of the State line. Several esker segments, one 7 miles long, occur between Hartland, Vt., and Lyme, N.H., a distance of 18 miles (Stewart and MacClintock, 1969, p. 91; Larsen, 1987a, p. 36). Hodges and others (1976b) demonstrated that these segments are productive aquifers. An esker is nearly continuous for a distance of 24 miles along the Passumpsic River. In each of these three valley reaches, esker segments are, in part, overlapped or buried by clay, silt, and fine sand deposited later in glacial Lake Hitchcock. Continuity of coarse sand and gravel deposits has not been demonstrated for even the short gaps between visible esker segments within the three reaches mentioned, let alone the many miles between reaches. Nevertheless, the known distribution of esker segments suggests a persistent subglacial meltwater conduit along this straight valley, and the substantial depth to bedrock (Hodges, 1968a–d; Moore and others, 1994) allows for the possibility that a basal aquifer might be nearly continuous, deposited at a conduit mouth as the ice margin retreated, but commonly buried beneath the abundant younger deltaic and lake-bottom sediment mapped by Stewart and MacClintock (1970).

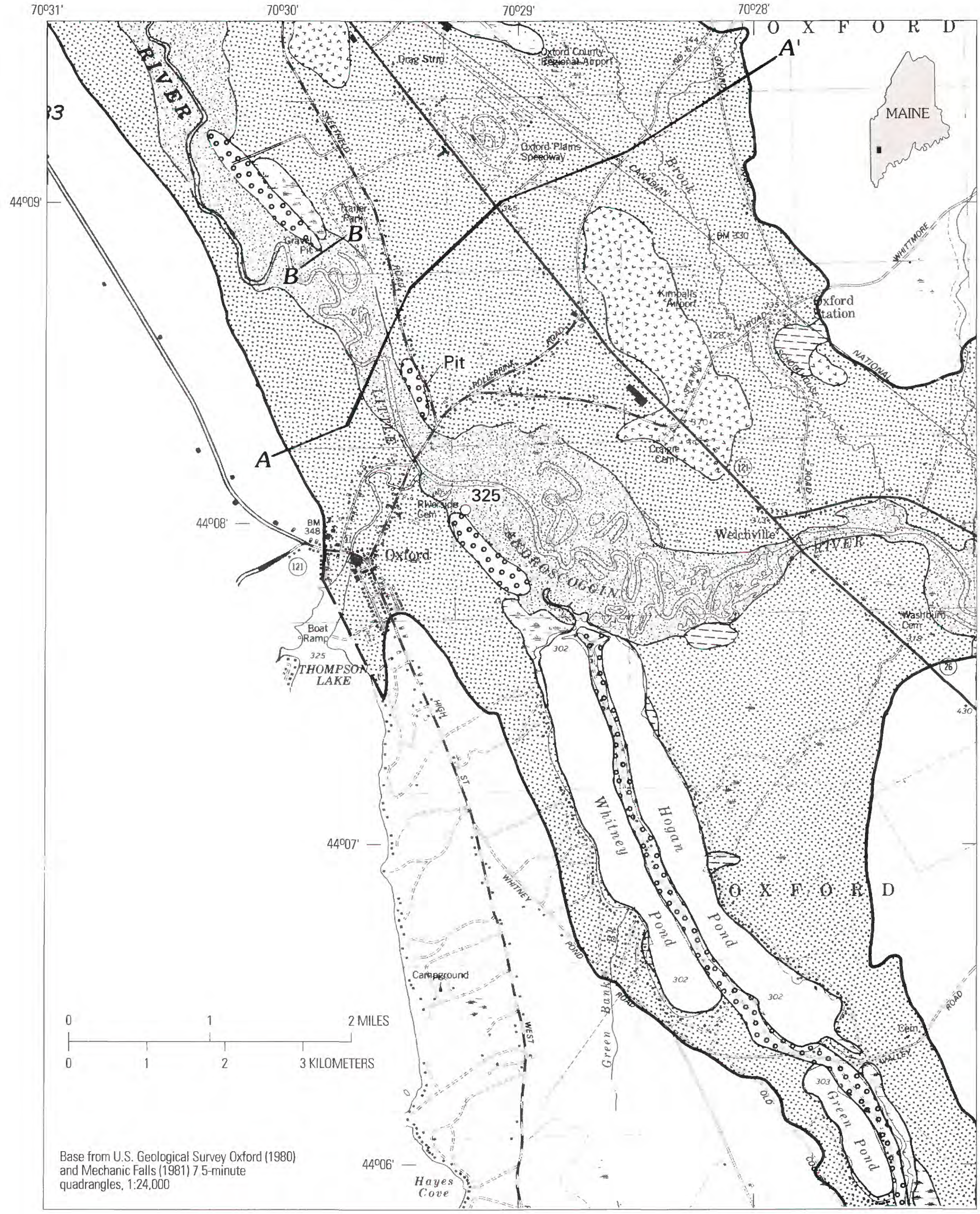


FIGURE 19.—Surficial geology in part of the town of Oxford, Maine. (Modified from Morrissey, 1983, pl. 3.)

EXPLANATION

Alluvium—Sand, silt, clay, and some gravel beneath stream channels and in flood plains and low terraces

Swamp deposits—Peat and organic muck and some interbedded silt, clay, and sand

Outwash (facies 3)—Stratified deposits of sand and gravel in outwash plains and valley trains; overlies or is interfingered with marine deposits

Marine deposits (facies 2)—Dark blue to gray silt, clay, and fine sand, tan where weathered. Contains layers of sand and gravel. Underlies outwash and may crop out in stream valleys

Ice-contact deposits (facies 1)—Well- to poorly stratified deposits of sand, gravel, and cobbles with some silt and boulders. Landforms include eskers

Till and bedrock—Till is an unsorted, unstratified mixture of clay, silt, sand, gravel, and boulders. Below the first few feet, particularly where thick, till is clay rich and very dense. Till covers the bedrock in the upland areas with a mantle of varying thickness

Contact between stratified drift and till or bedrock

A——*A'* **Line of geologic section (fig. 20)**

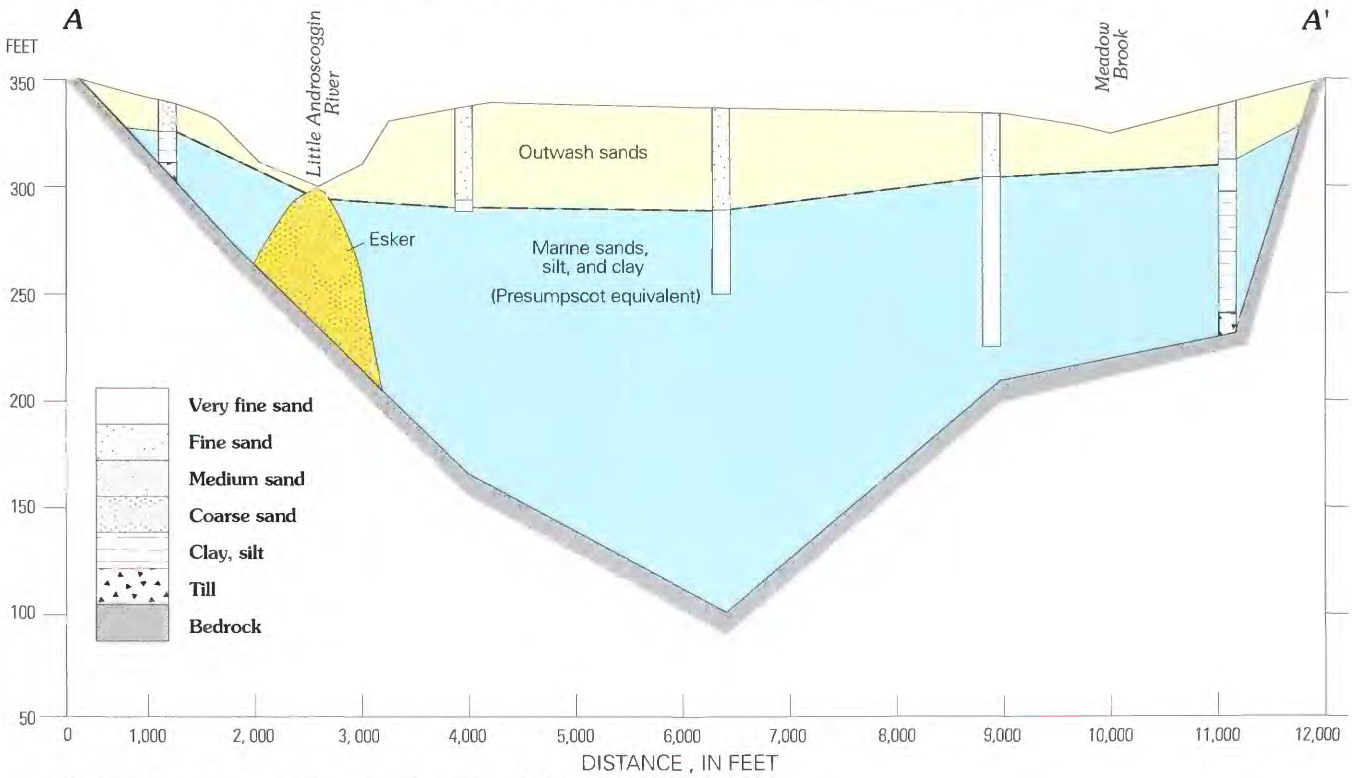
325 **Municipal well; number is yield in gallons per minute**

FIGURE 19.—Surficial geology in part of the town of Oxford, Maine. (Modified from Morrissey, 1983, pl. 3.)—Continued

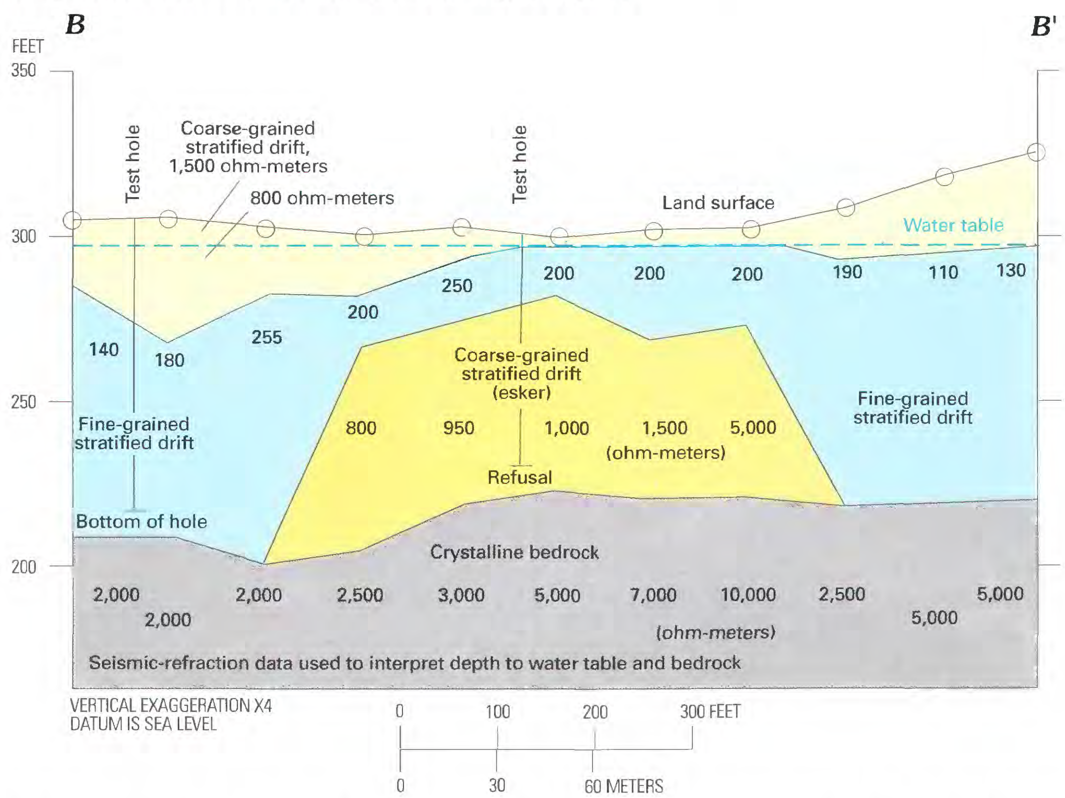
THROUGH VALLEYS IN RUGGED TERRAIN OF CENTRAL VERMONT

Meltwater generated during deglaciation in the mountainous headwaters of the Winooski River in Vermont drained south across the modern divide into the White River watershed because the lower reaches of the west-draining Winooski watershed were still occupied by ice. Five saddles, ranging in altitude from 1,660 to 915 feet, link the heads of major tributaries across that divide (fig. 21), but the distribution and character of stratified drift suggest that only the lowest two saddles were crossed by meltwater drainage systems that carried large volumes of sediment. The White River and its tributaries, south of the divide, were inundated by glacial Lake Hitchcock below an altitude of roughly 770 feet. Lakes also existed during deglaciation in valleys north of the divide (Larsen, 1987a, b). The five saddles are as follows:

1. *Washington Heights (1,660 feet altitude)*.—No evidence of stratified drift associated with meltwater drainage through this saddle is shown by Stewart and MacClintock (1970) or mentioned by Larsen (1987b).
2. *Brookfield Gulf (1,420 feet altitude)*.—No coarse sand or gravel is graded to the north end of Brookfield Gulf, and little occurs at the south end (Stewart and MacClintock, 1970; Larsen, 1987b). Farther south, the valley of Ayers Brook is filled with silt, clayey silt, and some very fine sand from somewhat north of East Braintree to Randolph. The fines are capped by a few feet of fluvial sand and gravel deposited after Lake Hitchcock drained, and small deltas of coarser sand and gravel occur where tributaries entered the lake (F.D. Larsen, Norwich University, written commun., 1985; A.D. Randall, unpub. data).
3. *Granville Notch (1,410 feet altitude)*.—A lake was ponded north of this saddle during deglaciation, as evidenced by small inwash deltas along tributaries at the altitude of the saddle (Larsen, 1987b; Ackerly and Larsen, 1987, stop 3). Coarse sediment tentatively interpreted as esker or subaquatic fan deposits occur along the valley axis near Warren (Ackerly and Larsen, 1987), but little or none occurs within 5 miles north or south of Granville Notch (Stewart and MacClintock, 1970).
4. *Roxbury (1,010 feet altitude)*.—A delta graded to the Roxbury saddle heads in an esker that extends north along the axis of the Dog River valley for at least 10 miles. The esker is composed of poorly sorted gravel that grades upward into pebbly medium to coarse sand and then into fine sand to silt. All of these sediments contain structures indicating southward flow (Larsen, 1972) and were deposited while ice still choked the valley (Larsen, 1987b). This subglacial channel was not only responsible for most of the stratified drift north of the Roxbury saddle but also probably functioned earlier as a source of meltwater and sediment to Third Branch valley south of Roxbury. No coarse ice-contact deposits have been recognized in Third Branch valley, perhaps because it was above the level of Lake Hitchcock, but the entire length of the valley has a smoothly graded floor underlain by coarse gravel that was interpreted by Stewart and MacClintock (1970) as outwash, and an unusually large delta was built into Lake Hitchcock at Randolph (Larsen, 1987a, stop 7; also written commun., 1985). All these features suggest that Dog River and Third Branch valleys, linked at Roxbury, constituted a through valley, a major avenue for sediment transport during deglaciation.



A. **Geologic section across the entire valley. (From Morrissey, 1983, pl. 3.)**



B. **Geoelectric section across buried esker, interpreted from direct-current resistivity, seismic refraction, and very-low-frequency apparent resistivity data. (From Haeni, 1995, fig. 57.) The interpreted resistivity in ohm-meters of each unit is shown; the numbers suggest that the fine-grained stratified drift is more resistive (coarser grained) near the esker than at a distance, as indicated by test-hole logs (Haeni, 1995) and as would be expected if the esker were mantled by upward-fining subaquatic outwash deposited at the tunnel mouth (fig. 3). The numbers also suggest that grain size within the esker increases toward the right (northeast) side.**

FIGURE 20.—Geologic and geoelectric sections across a buried esker in the Little Androscoggin River valley, Oxford, Maine. Location of sections shown in figure 19.

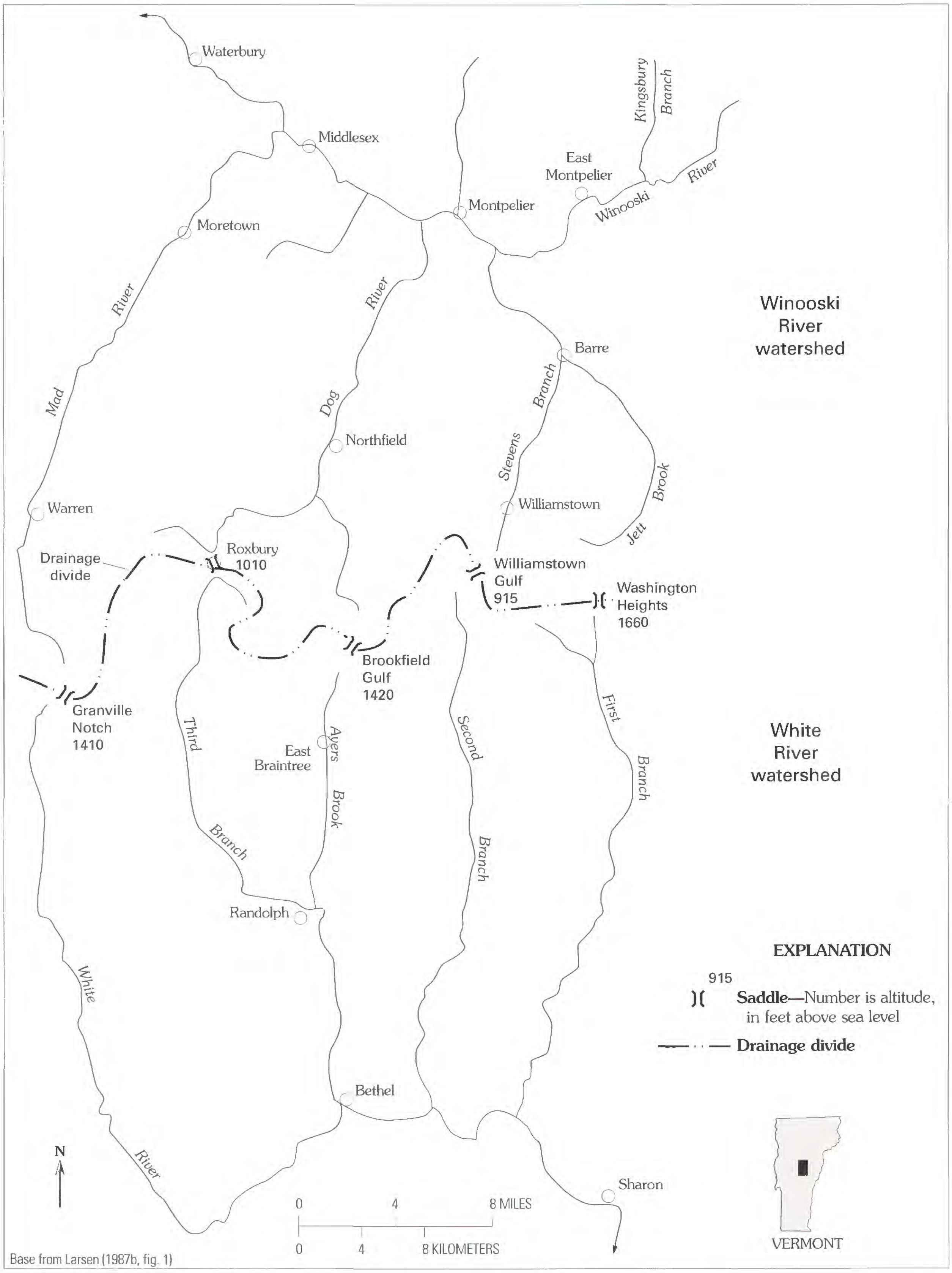


FIGURE 21.—Drainage patterns and saddles in central Vermont.

5. *Williamstown Gulf (915 feet altitude).*—The abundant ice-contact deposits in Second Branch valley were deposited in Lake Hitchcock, which extended north as far as the foot of Williamstown Gulf. These severely faulted deposits of coarse sand and gravel imply large meltwater flows while ice still choked the valley. The valley floor is characterized by low, streamlined landforms composed of gravel and interpreted by Larsen (1987b) to be longitudinal channel bars deposited by large meltwater flow that continued after Lake Hitchcock drained; the gravel may have been derived from ice-contact deposits planed off by meltwater. The valley of Stevens Branch, which heads on the north side of Williamstown Gulf, contains many esker segments and other ice-contact deposits of south-flowing meltwater (Larsen, 1972). Like similar deposits south of the gulf, they suggest abundant meltwater and sediment flow along englacial or subglacial channels early in deglaciation. Data and interpretive maps by Hodges and others (1976a) indicate a nearly continuous band of coarse, permeable stratified drift from Williamstown Gulf north along Stevens Branch to Barre, then north to the Winooski valley near East Montpelier, and farther north along the Kingsbury Branch valley to Lake Woodbury and beyond. All these valleys are aligned nearly north-south and seem to have constituted a regional meltwater channel system that resulted in coarse-grained stratified drift being more abundant than in other valleys.

In summary, surficial geology in the Winooski and White River watersheds of central Vermont indicates that meltwater channels, each largely subglacial and many miles in length, carried sediment through saddles at Roxbury and Williamstown Gulf during deglaciation. A channel along Second Branch extended northward through Williamstown Gulf, then along Stevens Branch, part of the Winooski valley, and Kingsbury Branch. About 10 miles to the west, a channel along Third Branch extended northward through Roxbury saddle, along much of the Dog River valley, and perhaps farther north. Either or both of these channels might have fed an esker at Sharon (Larsen, 1987a, stop 5). Coarse sand and gravel seem to be more abundant along these valleys than in parallel valleys heading at higher saddles, and, if so, greater aquifer potential could be expected.

THROUGH VALLEYS IN RUGGED TERRAIN OF NORTHWESTERN MASSACHUSETTS

The Hoosic River drains northwesterly from Massachusetts into the Hudson River lowland of New York. During deglaciation, ice retreated from southeast to northwest (Warren and Stone, 1986), so water was ponded in the upper reaches of the Hoosic watershed for a time while active ice still blocked the lower reaches. Most stratified drift was deposited in glacial Lake Bascom, which originally drained south through the Berkshire spillway (fig. 22) and later through a succession of lower outlets to the southwest (Bierman and Dethier, 1986; DeSimone and Dethier, 1988). Major valleys in this region are developed on interbedded marble and quartzite and form a crude H-pattern (fig. 22). The distribution and lithology of the stratified drift within these valleys, described below, suggest that a regional subglacial channel system may have carried meltwater and sediment south and east through the combination of valleys and saddles that offered the lowest route toward the margin of the ice sheet at the time.

1. *Northwest Limb: Vermont valley.*—Ice-contact stratified drift, including kame fields, deltas, and eskers, is abundant throughout this broad valley from beyond the northern limit of figure 22 through Bennington to Pownal Center (DeSimone and Dethier, 1988; A.D. Randall, unpub. data). The hummocky topography, numerous eskers, and the presence of sandy diamicton atop constructional topography suggest ice stagnation and subglacial deposition by meltwater (DeSimone and Dethier, 1992).

2. *Central Limb: Hoosic River valley west of North Adams.*—Ice-contact stratified drift is exposed near North Adams (Bierman and Dethier, 1986). Lacustrine silt and clay of Lake Bascom mantle valley walls below 750 feet (A.D. Randall, unpub. data) and underlie the valley floor, where they conceal a continuous basal gravel aquifer (Hansen and others, 1973; Bierman and others, 1988). The aquifer (fig. 23) was presumably deposited in or at the mouth of one or more subglacial channels.

3. *Southeast Limb: Upper Hoosic valley.*—Ice-contact landforms are abundant (Bierman and Dethier, 1986) and were once even more abundant (Hitchcock, 1841). The ice-contact deposits form a continuous aquifer on the valley side and (or) beneath clay on the floor (fig. 23), except locally near Adams where the river occupies a gorge (Hansen and others, 1973). Some deposits were graded to the Lake Bascom water plane, others to

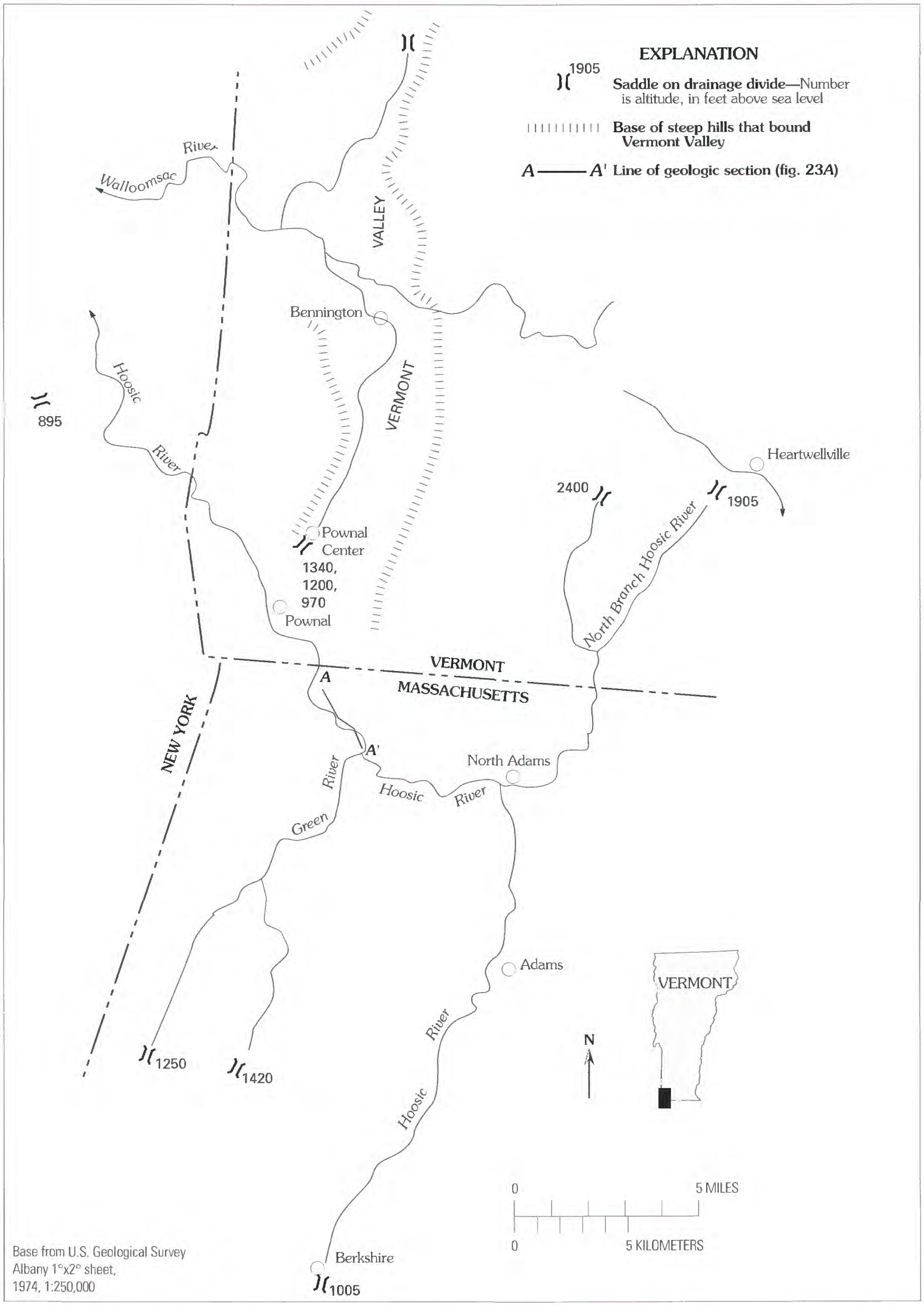
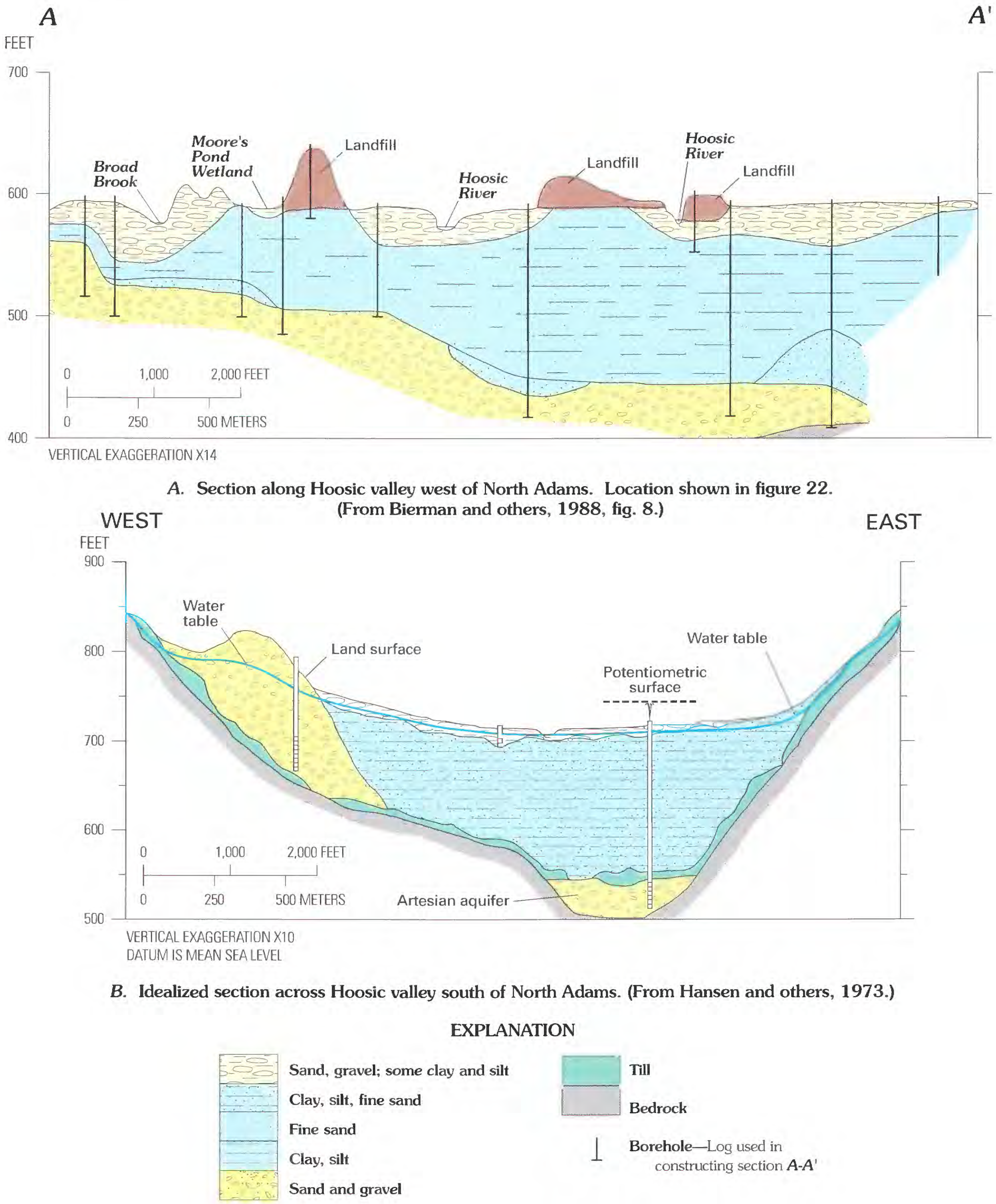


FIGURE 22.—Drainage patterns and saddles in southwestern Vermont and northwestern Massachusetts.



A. **Section along Hoosic valley west of North Adams. Location shown in figure 22. (From Bierman and others, 1988, fig. 8.)**

B. **Idealized section across Hoosic valley south of North Adams. (From Hansen and others, 1973.)**

FIGURE 23.—Geologic sections in Hoosic River valley, Massachusetts.

higher lake levels. Meltwater and sediment poured through the Berkshire spillway to build a large delta near Dalton (Warren and Stone, 1986).

4. *Northeast Limb: North Branch valley.*—Surficial sand and gravel deposits are sparse and largely associated with tributaries. They include alluvial fans and flood plains, small deltas and ice-channel deposits on the valley sides, and a broad plain near Roaring Brook, interpreted by Bierman and Dethier (1986, p. 37) as a delta into Lake Bascom. A band of constructional topography bordering the valley floor is mantled by sandy ablation till that, at least locally, overlies and is interlayered with sand and silt (A.D. Randall, unpub. data). The stratified sediment may have been deposited subglacially, but it seems too small in volume, lacking in gravel, and distant from the Heartwellville saddle to suggest meltwater flow through this valley.

5. *Southwest Limb: Green River valley.*—Sand and gravel are limited to flood plains and to a few widely scattered deltas, fans, and kames, mostly bordering modern streams (A.D. Randall, unpub. data) at altitudes that suggest base level was the Berkshire spillway of Lake Bascom. Silt and clay mantle land surface locally, but its apparently sparse distribution suggests early Lake Bascom may have existed here only in a few channels amid a widespread cover of stagnant ice. No evidence of early meltwater flow through the watershed toward saddles to the south has been reported.

In summary, the abundance and apparent continuity of coarse-grained ice-contact stratified drift along the Vermont valley and along the Hoosic River valley from Pownal Center east and south to Berkshire suggest that these valleys, which link the lowest saddles in the region, carried a large flow of meltwater early in deglaciation, at least in part along subglacial channels. Adjacent parallel valleys that head at higher saddles contain much smaller volumes of ice-contact deposits whose distribution generally suggests deposition by local tributaries amid crevassed stagnant ice that drained readily to the Hoosic valley. The inferred meltwater conduit system had an important effect on aquifer distribution: a continuous basal aquifer (facies 1) is present beneath silt and clay (facies 2) in the reach of the Hoosic valley from North Adams through Williamstown to Pownal Center, even though few ice-contact deposits are exposed at land surface in that reach.

OTHER TEMPORARY MELTWATER CHANNELS IN UPLANDS

At a few places in northern New England, meltwater channel systems extend for several miles through saddles and across uplands at orientations discordant with postglacial drainage. For example, the Cherry Valley esker, near Fabyans, N.H., extends at least 9 miles through an area of very high relief (Goldthwait and Mickelson, 1982, p. 175). It connects two saddles (notches) and constitutes the lowest possible route for subglacial meltwater southward across the White Mountains. A channel system at least 18 miles in length extends from the upper reaches of the north-draining Mississquoi River in Vermont, south through Eden Notch, past Eden Mills and Garfield, to a large delta built across the present Lamoille River valley east of Morrisville (fig. 24). Sand and gravel deposits mapped as kames and kame terraces by Stewart and MacClintock (1970) north of Eden Notch are mostly below the altitude of the notch and hence are presumed to have been deposited in tunnels and (or) as subaquatic fans at the mouth of a subglacial conduit along the axis of the Mississquoi valley. Probably, they were later reworked and slightly augmented by the small north-flowing tributaries. The reach between Eden Mills and Garfield includes a short esker and other constructional landforms and was interpreted by Hodges (1967a) as a continuous gravel aquifer; the map by Stewart and MacClintock (1970), however, implies the channels are mostly erosional. These deposits probably did not all form simultaneously, but collectively they reflect a persistent channel system that carried considerable meltwater. North of Lowell, Vt., the Mississquoi valley is largely mantled by fine sand, silt, and clay deposited later in large lakes that drained across saddles to the east and west at or below 840 feet in altitude. Stewart and MacClintock (1969, p. 171), however, mention localities near the Mississquoi River at Troy, Vt., where these lake fines mantle ice-contact kame gravels that may well represent headwater segments of the earlier channel system.

THROUGH VALLEYS IN THE APPALACHIAN PLATEAU

Several lines of evidence indicate that extensive subglacial drainage systems followed the major valleys of the Susquehanna River basin during deglaciation. The absence of stratified drift in many north-flowing tributaries indicates that the outermost several miles of the ice sheet were porous and capable of internal subglacial drainage that was controlled by bedrock topography. The lithology and volume of the stratified drift in the major valleys provides further evidence that meltwater drainage systems apparently extended for long distances through or beneath sediment-laden basal ice.

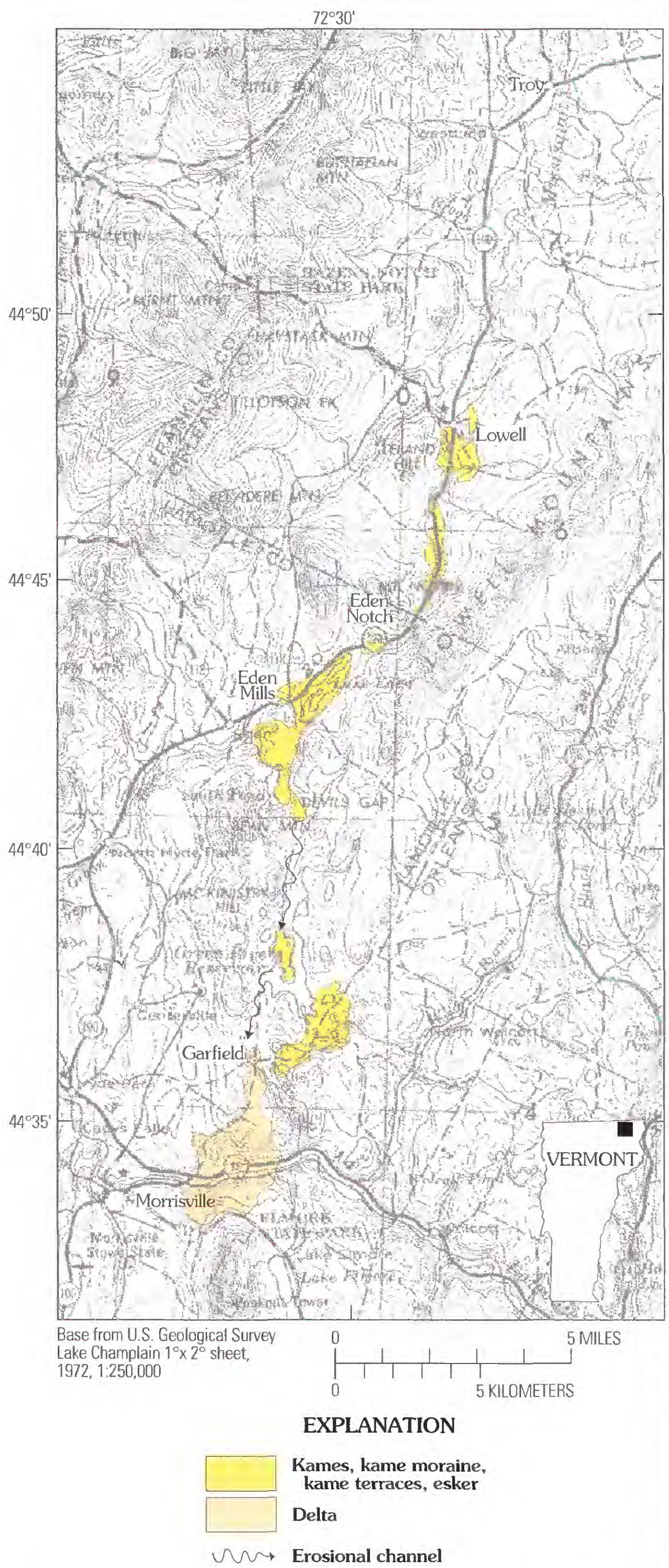
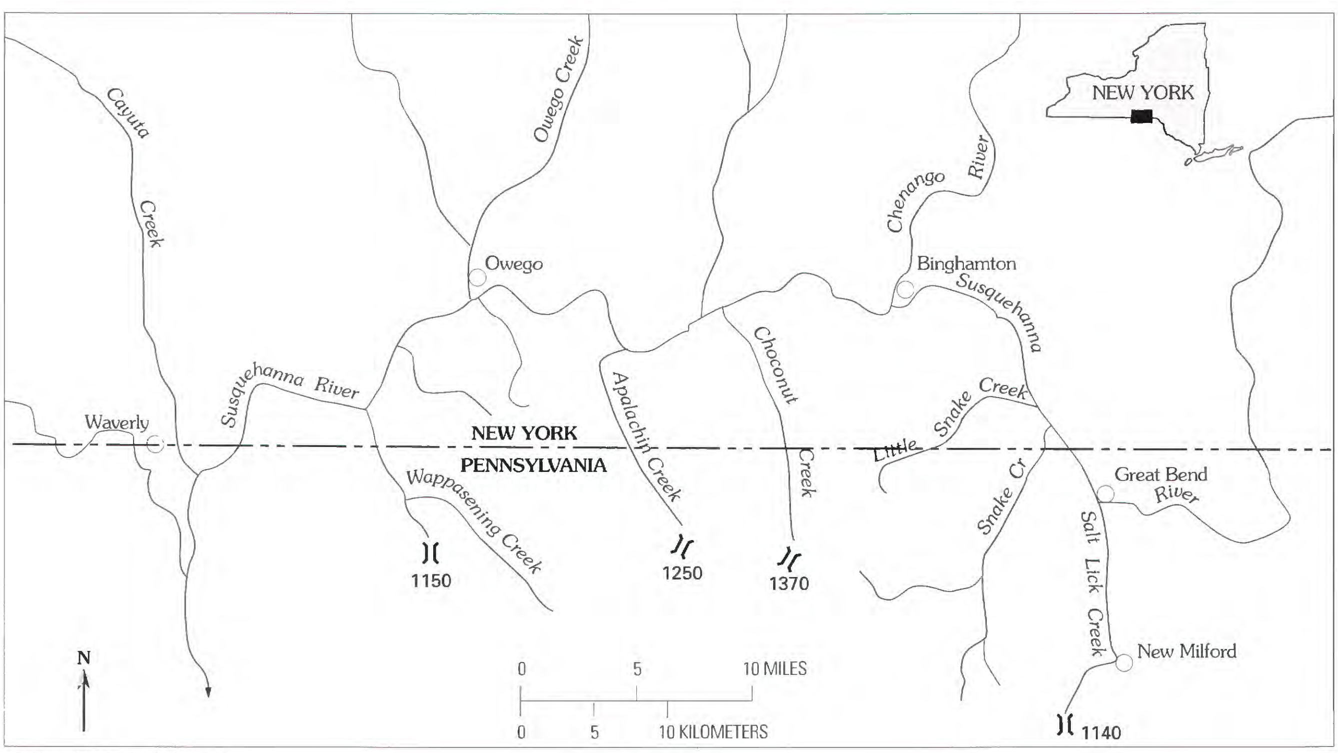


FIGURE 24.—Stratified deposits and erosional channels produced by meltwater flow through Eden Notch in northern Vermont. (Modified from Stewart and MacClintock, 1970.)

In south-central New York, the Susquehanna River flows generally westward for 45 miles from Great Bend, Pa., through Binghamton and Owego to Waverly, N.Y., before turning south into Pennsylvania (fig. 25). In this reach, it is joined by several north-flowing tributaries that range in length from 5 to 12 miles. Saddles at the heads of these tributary valleys range upward from 1,140 feet in altitude, whereas the depositional surfaces of stratified drift in the Susquehanna River valley are generally between 800 and 900 feet in altitude. The four lowest saddles (fig. 25) are narrow, straight-sided gorges interpreted to be spillways carved by large volumes of meltwater during advance and (or) retreat of one or more ice sheets (Denny and Lyford, 1963; Harrison, 1966; Coates and King, 1973). Only one of these tributaries, Salt Lick Creek, contains appreciable amounts of stratified drift graded to a headwater saddle. Several deltaic ice-contact deposits line the walls of Salt Lick Creek valley (Harrison, 1966; Coates and King, 1973), all reaching altitudes approximately level with the saddle near New Milford, Pa., which would have been the lowest potential outlet for meltwater from the Susquehanna River basin anywhere east of Waverly, N.Y. These ice-contact deposits were apparently derived largely from inwash rather than subglacial flow along the valley; Coates and King (1973, also supplemental field trip itinerary) indicate that three deltas on the east side of Salt Lick Creek valley were deposited by meltwater from the next valley to the east, which poured across saddles on the intervening north-south ridge and eroded gullies as it descended 250 feet or more to the 1,150-foot level of the lake in Salt Lick Creek valley. No deposits graded to the New Milford spillway have been identified on the walls of the Susquehanna River valley near or northwest of the mouth of Salt Lick Creek (Harrison, 1966), which suggests that when the locus of ice-contact deposition reached the north end of Salt Lick Creek valley, meltwater began to drain beneath or through the ice down the Susquehanna River valley to Binghamton and beyond. The other north-flowing tributaries contain postglacial alluvial fans and flood plains, but virtually no surficial stratified drift. One small ice-contact deposit was recognized by Denny and Lyford (1963); other constructional landforms in several valleys are composed of or mantled by till (Grubb, 1986; Reber, 1973); all these features are much lower in altitude than the headwater saddles and distant from them. The spillway in Wappasening Creek valley (fig. 25) contains a small plug of drift so located as to deny the possibility of large meltwater flow during the last deglaciation. No soils derived from surficial lake-bottom sediments have been recognized in any of these valleys (Grubb, 1986; Reber, 1973). Thus, surficial geology and topography indicate that:

1. The ice was so porous or crevassed that proglacial lakes seldom developed in these north-draining valleys. Instead, early in deglaciation, meltwater readily drained 5 to 12 miles northward through the decaying ice, then westward down the Susquehanna River valley.
2. Sediment-laden subglacial meltwater from the north did not generally flow southward (upslope) through these tributary valleys during deglaciation. Instead, it apparently followed drainage routes along deep valleys such as the Susquehanna and the Chenango.

Contrasts in drift lithology within the Appalachian Plateau (Moss and Ritter, 1962; Denny and Lyford, 1963; Randall, 1978a, 1981) also are consistent with the concept of extensive subglacial flow systems. Drift in the uplands and in the Susquehanna River valley upstream from Binghamton is composed almost entirely of fragments of local shale and siltstone bedrock. So are the earliest stratified deposits all along the Susquehanna River valley from Binghamton to Waverly. These early deposits comprise kame terraces along the sides of the valley and buried ridges along the valley axis that suggest subglacial conduits (Randall, 1986a). Later deposits, however, contain much larger percentages of erratic bedrock types derived from areas far to the north and transported to the Susquehanna valley by way of the Chenango River, Owego Creek, and Cayuta Creek valleys (fig. 25), which extend as continuous, deep through valleys to and across the divide near the northern perimeter of the Appalachian Plateau. This change in stratified-drift lithology suggests that early in deglaciation of the Susquehanna valley west of Binghamton, most sediment was derived from melting ice and fluvial erosion in the uplands roundabout, probably augmented by meltwater flux from the southeast along the Susquehanna valley, whereas later in deglaciation, sediment transport along the south-flowing tributary valleys predominated. Transport of erratic pebbles has been ascribed to



Base from U.S. Geological Survey State Base Maps, 1:500,000

EXPLANATION

1370)(**Spillway incised into saddle on divide**—Number is altitude of spillway, in feet above sea level

Figure 25.—Drainage patterns and saddles in south-central New York and north-central Pennsylvania.

rapidly moving ice streams within the ice sheet but confined to the through valleys (Muller, 1965a) and to subglacial streams along the through valleys (Randall, 1978a) that not only transported erratic pebbles southward but also increased the erratic content of the stratified drift by attrition of the less durable stones (Bugh and others, 1970).

Fleisher (1991a, b) calculated from well records that the valleys of the Susquehanna and Unadilla watersheds upstream from Sidney, N.Y., contain 14 cubic miles of fine-grained lacustrine sediment. Removal of an average of 50 feet of till from the intervening uplands during or shortly after deglaciation would be required to account for all this fine sediment as inwash. Fleisher argued that this much erosion by stream incision, mass wasting, sheetwash, or winnowing would have left conspicuous evidence of dissection and widespread bouldery lag gravel, neither of which exist, and concluded that the bulk of the fines must have entered proglacial lakes from conduits beneath active ice tongues in the valleys, having been eroded by meltwater from basal ice or subglacial sediment. If so, two corollaries follow: (a) the subglacial drainage systems must have headed a long distance north of the ice margin to generate sufficient sediment; and (b) the sand and gravel eroded from the basal ice or subglacial sediment must have been deposited concurrently atop bedrock (facies 1) near the mouths of the subglacial conduits, especially in through valleys, which would be the most likely routes of long subglacial conduits.

Farther north, seismic-reflection surveys across Cayuga, Seneca, and Owasco Lakes (Mullins and others, 1991) disclosed a sinuous ridge in each valley that overlies bedrock near the valley axis and can be traced for a few miles. The ridges (fig. 26) are 1,000 to 1,600 feet wide at the base, 65 to 115 feet high, and are composed of seismically chaotic, unconsolidated sediment like that which blankets bedrock to lesser

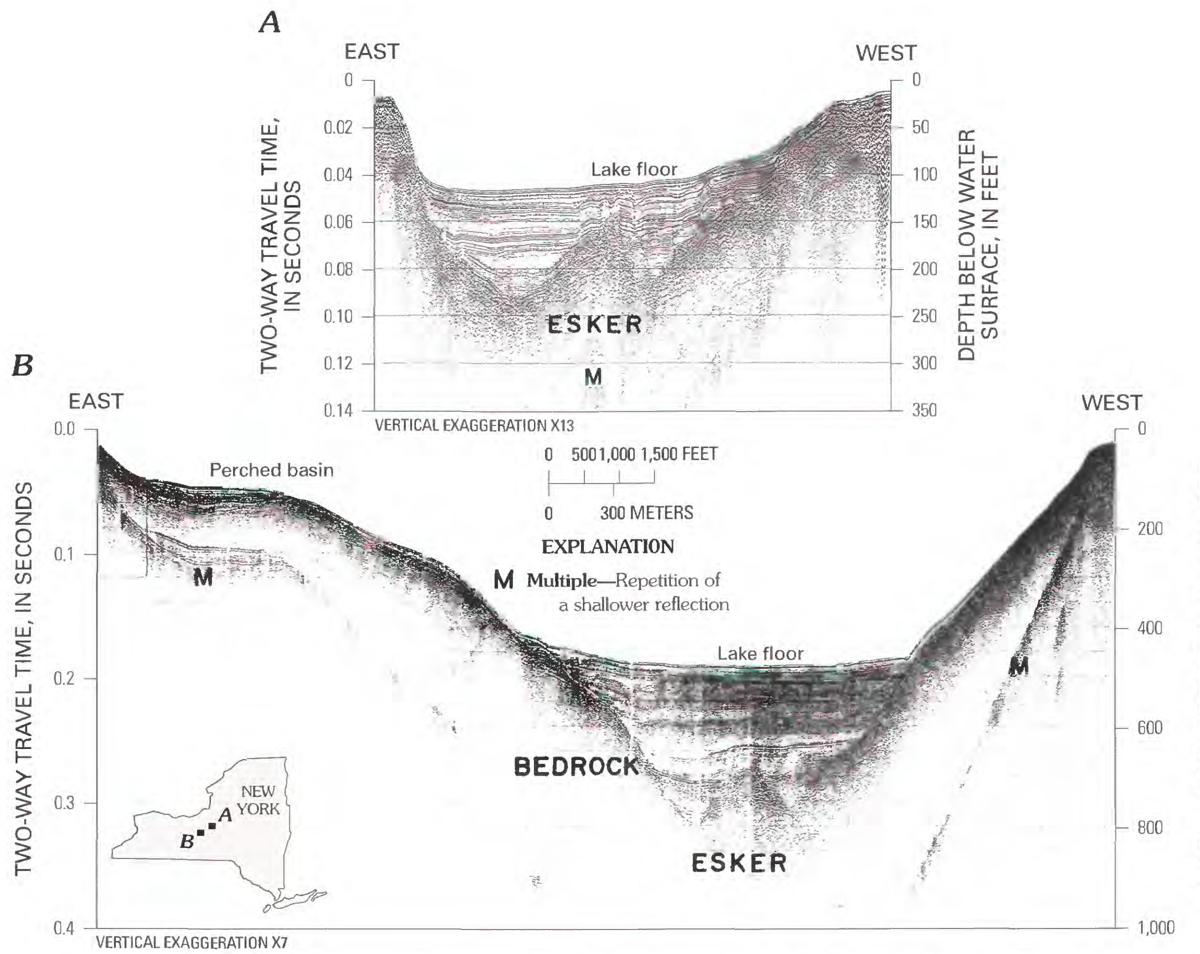


FIGURE 26.—Seismic-reflection profiles across (*A*) northern Owasco Lake and (*B*) central Seneca Lake, New York. (From Mullins and others, 1991.) Near the axis of each valley is a buried ridge interpreted as an esker.

thickness on either side of the ridges; a piston core at one site recovered clast-supported gravel. The ridges are recognized only beneath the northern half of each lake, where bedrock is appreciably shallower than farther south, and are interpreted as eskers deposited in extensive subglacial conduits.

MELTWATER CHANNELS IN LOWLANDS

Linear bands of coarse-grained ice-contact deposits have been demonstrated or inferred to be continuous over distances of several miles at several localities in broad lowlands, where they are overlapped or buried to varying degrees by fine-grained sediment.

Ontario, Canada.—Two esker systems in lowlands outside the Northeastern United States are cited briefly because they provide exceptionally clear examples of the spacing and continuity of meltwater channels.

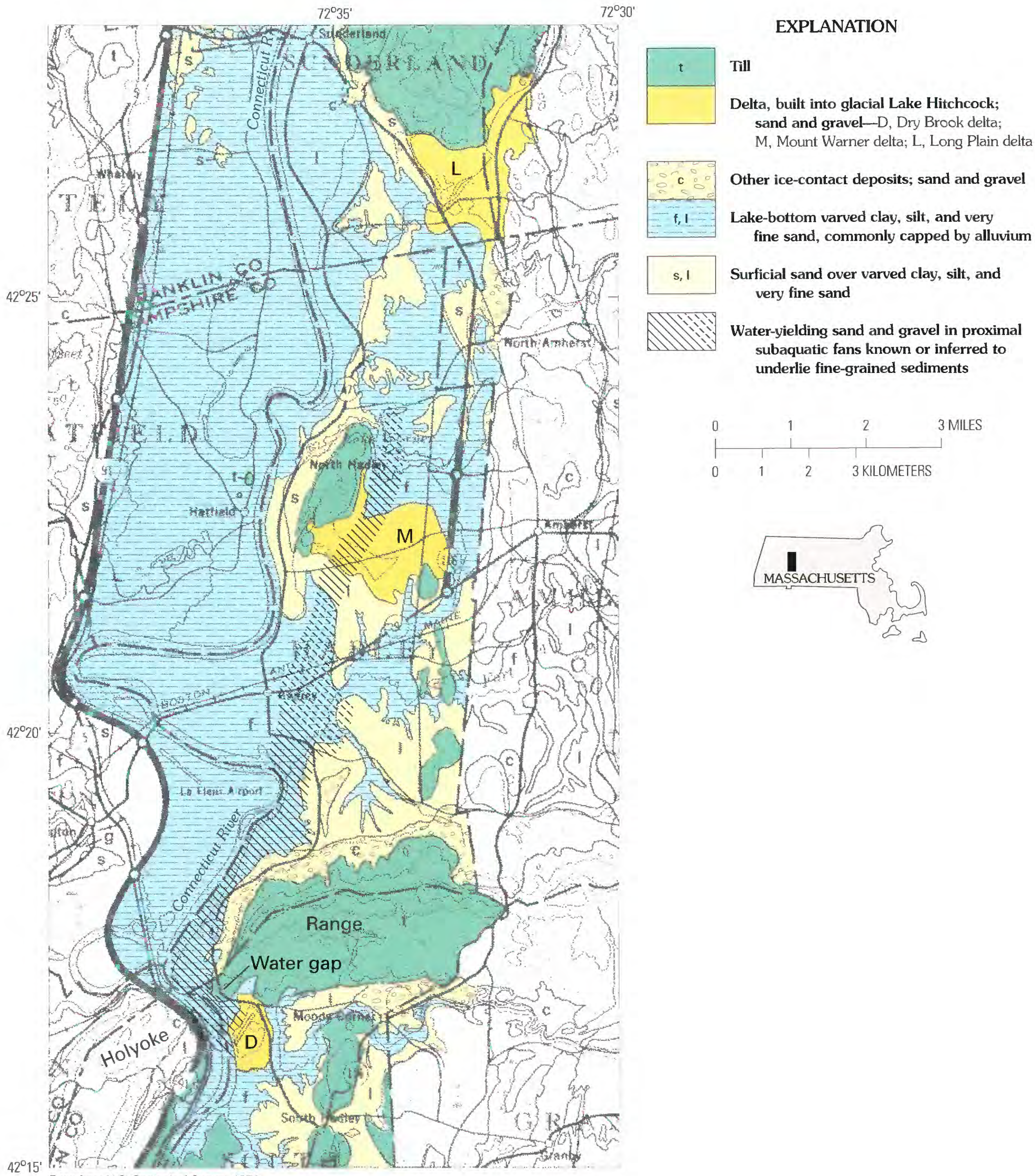
Parsons (1970) described a lowland area near Cochrane, Ontario, that was part of a large proglacial lake during deglaciation. Two parallel esker complexes are 7 miles apart and more than 12 miles long; each includes a linear gravel ridge as much as 1,000 feet wide that is locally mantled, bordered, or replaced by relatively flat sand plains (presumably deltas or subaquatic fans) as much as 8,000 feet wide. The remainder of the area is a swampy plain underlain by varved clay that grades down into silt and very fine sand that overlaps the esker complexes.

Rust and Romanelli (1975) described a broad lowland south of Ottawa, Ontario, that also is crossed by eskers mantled by sandy subaquatic fan deposits. These features occur as ridges 5 to 8 miles apart, 8 to 10 miles long, and less than 50 feet high. They are generally oriented parallel to each other, and parallel to ice movement as indicated by striations. They were deposited below at least 150 feet of water and are bordered and overlapped by marine silty clay.

Eastern New York.—LaFleur (1965, fig. 3; 1979a, figs. 2–6) recognized within the Hudson River lowland near Albany, N.Y., a set of ice-contact deposits, chiefly kames or kame deltas, aligned roughly north-south and overlapped in part by fine sand, silt, and clay deposited later in proglacial Lake Albany. LaFleur interpreted the ice-contact deposits as the product of a major subglacial or englacial drainage system that persisted in approximately the same alignment while the ice front retreated 15 miles over irregular topography. Dineen and Hanson (1983) traced this alignment of ice-contact deposits farther north, for a total length of at least 21 miles. They interpreted it as a zone 1 to 3 miles wide in which basal ice-contact deposits are generally present, mostly thin, but locally 25 to 100 feet thick. Much of the zone follows a buried bedrock valley, but the thickest ice-contact deposits commonly occur on the valley sides rather than along the thalweg; thick ice-contact deposits presumably reflect pauses in retreat where sediment carried along the subglacial drainage system had time to accumulate. The interpretation by Dineen and Hanson (1983) is illustrated later in this paper (fig. 62).

Wallkill Valley in New York and New Jersey.—The valley of the north-draining Wallkill River was inundated by large proglacial lakes during deglaciation. In northern New Jersey, ice-contact deposits form a continuous line about 1,000 feet wide for 8 miles along the west side of the Wallkill valley south from the New York border, interrupted by a large delta at Sussex, N.J., and terminating in another large delta near Hamburg, N.J. (Stanford and others, 1998). Over most of this distance, the ice-contact deposits are probably largely unsaturated, perched on bedrock that is probably above stream grade and certainly well above the thalweg of the bedrock valley to the east. The ice-contact deposits were laid down by south-flowing currents and are interpreted by Stanford and others (1998) as sublacustrine fans deposited at the mouth of a persistent meltwater channel within the retreating ice. In New York, the Wallkill valley is broader than in New Jersey and is generally characterized by fine-grained lacustrine sediment at land surface. Ice-contact deposits are locally exposed near the west side of the valley, however, and were interpreted by Frimpter (1972) as constituting a continuous buried aquifer as much as a mile wide that extends for 11 miles along the west side of the valley. Only a few well records (Frimpter, 1970) were available to document the inferred continuity of the basal aquifer in New York, but it seems evident that early in deglaciation, meltwater flow and ice-contact deposition were concentrated along the west side of the Wallkill valley in both New York and New Jersey, over a north-south distance of at least 19 miles.

Central Massachusetts.—The geologic setting of the Mount Warner delta in central Massachusetts was described earlier (fig. 17) as an example of a sand and gravel aquifer buried beneath younger fine-grained sediment at the head of a morphosequence. The sand and gravel probably were deposited along a meltwater channel that fed the delta. A similar feature, known as the Dry Brook delta (Saines, 1973), is about 7 miles to the south, southeast of the water gap where the Connecticut River cuts through the Holyoke Range (fig. 27). Immediately upstream from this delta, chiefly west of the Connecticut River, several wells penetrate coarse ice-contact or deltaic deposits (Saines, 1973; Cederstrom and Hodges, 1967).



Base from U.S. Geological Survey, 1974

FIGURE 27.—Surficial geology and geohydrology along part of the Connecticut River valley in central Massachusetts. (Adapted from Stone and others, 1979; Hansen, 1986; and Saines, 1973.)

Immediately northeast of the water gap, wells penetrate water-yielding gravel beneath 125 feet of fine-grained lake-bottom deposits (Saines, 1973; Hansen, 1986). Several seismic-reflection and well records farther north along the Connecticut River led Hansen (1986) to infer that a continuous band of ice-contact deposits lies buried beneath fine sediment from the water gap northeastward at least as far as the Mount Warner delta (fig. 27). Both of these deltas are so located that they could not have received sediment directly from upland tributary watersheds; the sediment must have come through a meltwater channel(s) within or under the ice sheet. Collectively, these observations suggest a persistent subglacial channel at least 7 miles in length, which formed here because the water gap is the lowest place at which subglacial meltwater could drain south through the Holyoke Range. Meltwater flowing along the channel built the deltas during pauses in retreat that occurred while the ice was grounded on the Holyoke Range and later on Mount Warner.

Four miles north of the Mount Warner delta is the Long Plain delta, another large accumulation of coarse-grained sediment. The Long Plain delta was deposited by meltwater from a tributary valley aligned with the meltwater channel inferred in figure 27. Several test holes on the intervening lake plain penetrated thick lake-bottom sediments but did not detect a northward extension of the basal gravel aquifer.

SUMMARY

The preceding sections have described or cited examples of linear bands of coarse-grained ice-contact deposits (facies 1) that are known or inferred to extend without interruption for several miles. They are inferred to be the products of meltwater flow along persistent subglacial channels whose spacing was controlled by interaction of glacier physics and topography. They commonly include eskers, but also include other constructional landforms. They commonly have sufficient saturated thickness to constitute productive aquifers. In areas of low relief, they tend to be spaced 5 to 15 miles apart and to be oriented down the ice-surface slope, perpendicular to the ice margin, but are confined to some extent in valleys and cross from one valley to another at low saddles. In areas of rugged terrain, their orientation conforms more closely to major valleys, but they still link some valleys across the lowest available saddles and are absent or less prominent in parallel valleys that head at higher saddles. In the Appalachian Plateau, they probably follow all the deep, glacially eroded troughs, although they are not as well exposed so their continuity is difficult to demonstrate. Throughout the glaciated Northeast, where surficial or subsurface data indicate the existence of linear channel deposits in particular localities, interpolation or extrapolation of the linear trend is a reasonable method of exploration for water supplies.

TRIBUTARIES AS SOURCES OF COARSE SEDIMENT

The stratified drift in valleys of the glaciated Northeast was derived in part from subglacial flow systems that followed these valleys, as described in preceding sections, but also in part from tributary streams that drained watersheds in the adjacent uplands. The tributaries entrained sediment from melting ice in those uplands as well as from the freshly exposed drift into which they were incising their channels.

Wherever a proglacial lake existed temporarily at a level much higher than the modern valley floor, deltas or deltaic ice-contact landforms are common along the former lake shore near upland tributaries. Such deltas are obvious evidence that the tributaries were sources of coarse sediment to the major valleys, although they generally are not important aquifers today. For example, in the 5-square-mile upland watershed of Cold Brook near South Tamworth, N.H. (fig. 28), an ice-channel filling about 100 feet wide and a few hundred feet long lies alongside each of several small tributaries; these landforms were deposited in channels or tunnels that had been carved in the ice by the tributaries, then ponded up to an altitude of about 1,060 feet. The Mad River watershed of central Vermont contained a proglacial lake that spilled through Granville Notch (fig. 21), and small deltas have been recognized along the inferred shoreline near each tributary (Ackerly and Larsen, 1987, stop 3; F.D. Larsen, Norwich University, oral commun., 1987). Along the west side of the Green River watershed south of Williamstown, Mass., isolated deltas or fans more than 100 feet above the valley floor were recognized near most small tributaries (Holmes, 1965) at altitudes that probably correspond to the earliest level of proglacial Lake Bascom (Bierman and Dethier, 1986). A vast complex of coalesced sandy deltas was formed along the east side of the Black River valley in northern New York by streams descending from the Adirondack Mountains (Waller and Ayer, 1975, p. 15; Waller, 1976). Deltas also were built by all major streams flowing north from the Adirondacks into a succession of lakes in the St. Lawrence lowland (Clark, 1984). All the deltaic deposits cited above are perched high on the sides of valleys or lowlands and, thus, are thinly saturated and of minor importance as aquifers, although many are sources of perennial springs or seepage that contribute significantly to low flow of streams and are locally used for water supply. Several studies, however, have reported saturated stratified drift within valley fills to be especially coarse where a prominent tributary enters the valley (Randall, 1964, p. 53; Kammerer and Hobba, 1967; Schiner and Gallaher, 1979, p. 9), and other studies have presented pebble counts or mineralogic analyses that indicate some of the coarse sediment in valley fills to be derived from bedrock in the adjacent uplands (Langer, 1977; Randall, 1978a; Ridge, 1985; Stanford, 1989a, b; Stanford and others, 1998).

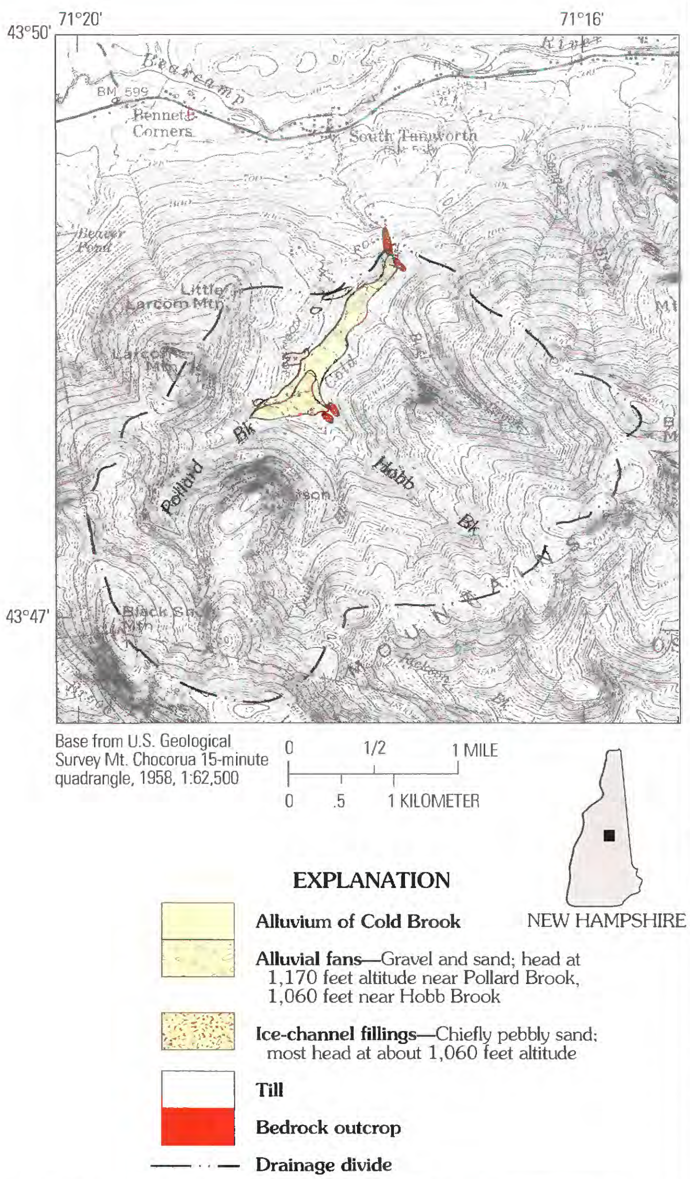


FIGURE 28.—Surficial geology of the Cold Brook watershed, Tamworth, N.H.

In the Appalachian Plateau, sand and gravel deposited where tributaries enter major valleys are significant aquifers and transmit recharge to deep aquifers. Erosion by ice scoured the bedrock surface in most major valleys to depths 100 to 500 feet below the thalwegs of tributary valleys. Where the tributaries descend from these "hanging" valleys in the uplands to the modern floor of the main valley, they have deposited large alluvial fans composed of variably silty, sandy gravel. Apparently the tributaries also contributed coarse sediment to the proglacial lakes that occupied the major valleys during deglaciation. Fleisher and others (1992) inferred from geophysical surveys that most of the sediment beneath the southern half of Otsego Lake was derived from ice-free upland tributaries during deglaciation. Crain (1966) demonstrated the presence and hydrologic function of the deglacial and postglacial fans deposited by tributaries in Cassadaga valley near Jamestown, N.Y.; his interpretation can be summarized as follows: A gravel aquifer ranging from 10 to 40 feet in thickness lies below the floor of Cassadaga Creek valley over a distance of 10 miles. It might have been deposited as a subaquatic fan(s) by meltwater issuing from a retreating ice tongue or by north-flowing streams as fan and channel alluvium during an interstadial interval. It is overlain by 80 to 120 feet of silty clay deposited in a proglacial lake that persisted for several hundred years or was reestablished by ice readvance. It is tapped for municipal supply near Jamestown and is known as the Jamestown aquifer. The coarse sediment that was deposited near the mouths of tributaries merges with the Jamestown aquifer and interfingers with the fine sediment that subsequently accumulated over most of the lake floor (fig. 29). The coarse deposits of the tributary streams were termed deltas by Crain (1966), but the stratigraphy and form (fig. 30) are suggestive of subaquatic fans rather than flat-topped Gilbert-type deltas. Teller (1987, p. 44) mentioned that the sediment load deposited in proglacial lakes by rivers commonly occurs as sublacustrine splays with concave-up surfaces and bedding that dips toward the center of the lake. The tributaries in Cassadaga Creek valley lose large amounts of water as they cross their alluvial fans, and on one fan (fig. 30) an annual temperature fluctuation from 3° to 13°C recorded in the well nearest to the head of the fan showed that seepage losses from the tributary moved quickly to depths of at least 55 feet. The potentiometric surface in the same fan paralleled land surface in the spring, but sloped toward the valley wall in September (fig. 30). From this evidence,

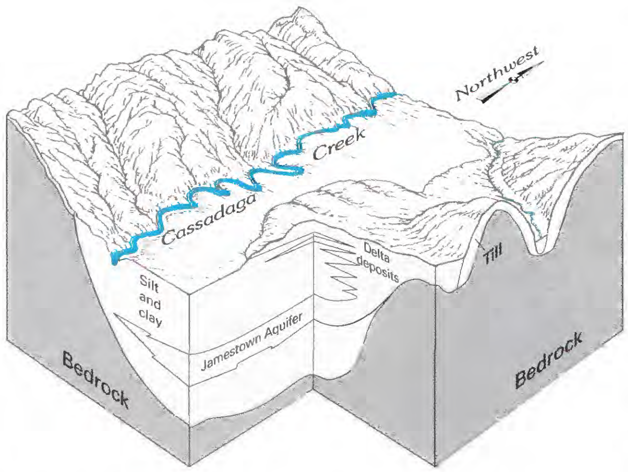


FIGURE 29.—Idealized relation of tributary delta (or fan) to the Jamestown aquifer in Cassadaga Creek valley, New York. (From Crain, 1966, fig. 22.)

Crain (1966) deduced that coarse sediment near the head of the tributary fan is continuous from land surface to a depth of at least 125 feet and constitutes a window or conduit through which recharge from precipitation and tributary-stream seepage on the alluvial fan can migrate downward and recharge the Jamestown aquifer beneath the valley floor (figs. 29, 30).

The concept of tributary fans as localized aquifers along the valley walls and as conduits that convey recharge to buried aquifers has been adopted by other investigators in the Appalachian Plateau. For example, MacNish and Randall (1982, figs. 8–9, pl. 1) found the concept consistent with scattered borehole records and applied it in interpreting aquifer geometry in the Susquehanna River basin. Randall (1978b) drew geologic sections based on borehole records beneath seven alluvial fans deposited by tributaries in major valleys of south-central New York. All seven fans overlie coarse sand and gravel near the valley wall; but nearer the center of the valley, a wedge of fine sand or silt separates four of the fans from the deeper aquifer (fig. 31). Williams (1991) drew sections beneath several tributary fans in north-central Pennsylvania that indicated most of the valley fill to be sand and gravel. Both Randall (1978b) and Williams (1991) inferred that the sand and gravel beneath postglacial alluvial-fan gravels were deposited by meltwater flowing along the major valley, perhaps augmented by sediment from tributaries, rather than solely a product of tributary inflow as inferred by Crain (1966) for Cassadaga Creek valley.

INCREASED SATURATED THICKNESS AROUND ICE-BLOCK DEPRESSIONS

Ice-block depressions, a common feature of stratified drift in the glaciated Northeast, can significantly influence aquifer geometry. Blocks of stagnant ice were buried or overlapped by rapidly accumulating stratified drift in many places. As those ice blocks melted, the loss of support caused the overlying or adjacent stratified drift to sag or settle, by gentle downwarping and by abrupt displacement along normal faults, resulting in irregularly shaped depressions known as kettleholes or ice-block depressions. Examples abound throughout the glaciated Northeast. In figure 15, for example, Great Pond and Potter Pond occupy linked ice-block depressions that extend below the modern water table, whereas the "Mustard Bowl" 1 mile southwest of Great Pond is a single, dry ice-block depression. Where an ice block melted before local proglacial lakes drained, the kettlehole is likely to contain fine-grained lake-bottom sediment.

The origin and potential hydrologic significance of ice-block depressions are illustrated in figure 32. Deltaic or fluvial outwash overlies lake-bottom fines in the distal parts of many morphosequences, but modern streams commonly are incised below the base of facies 3, leaving the surficial aquifer thinly saturated. Where the outwash is downwarped or collapsed around the margin of ice-block depressions, however, its saturated thickness is thereby increased, and wells at such locations could effectively capture recharge from surrounding areas of smaller saturated thickness. Organic matter accumulates in many ice-block depressions, however, and trash as well in some; decay of these materials can result in low pH, reducing conditions, and elevated dissolved iron and manganese in ground water pumped nearby (Randall and others, 1966; Gay and Frimpter, 1984).

Several examples of downwarping of surficial outwash beneath ice-block depressions and of groundwater flow along the downwarped aquifers have been documented along the Chenango and Susquehanna River valleys in Broome County, N.Y. The stratigraphy in these valleys is summarized in table 3. Complex downwarping just west of the Chenango River is illustrated by a map (fig. 33), three geologic sections (fig. 34), and a block diagram (fig. 35). Ice-contact deposits and outwash in the sections were distinguished by pebble lithology, as explained in table 3. Water-level measurements and temperature profiles in several wells (Randall, 1977) indicated that at least 78 percent of the 1,500 gallons per minute pumped by

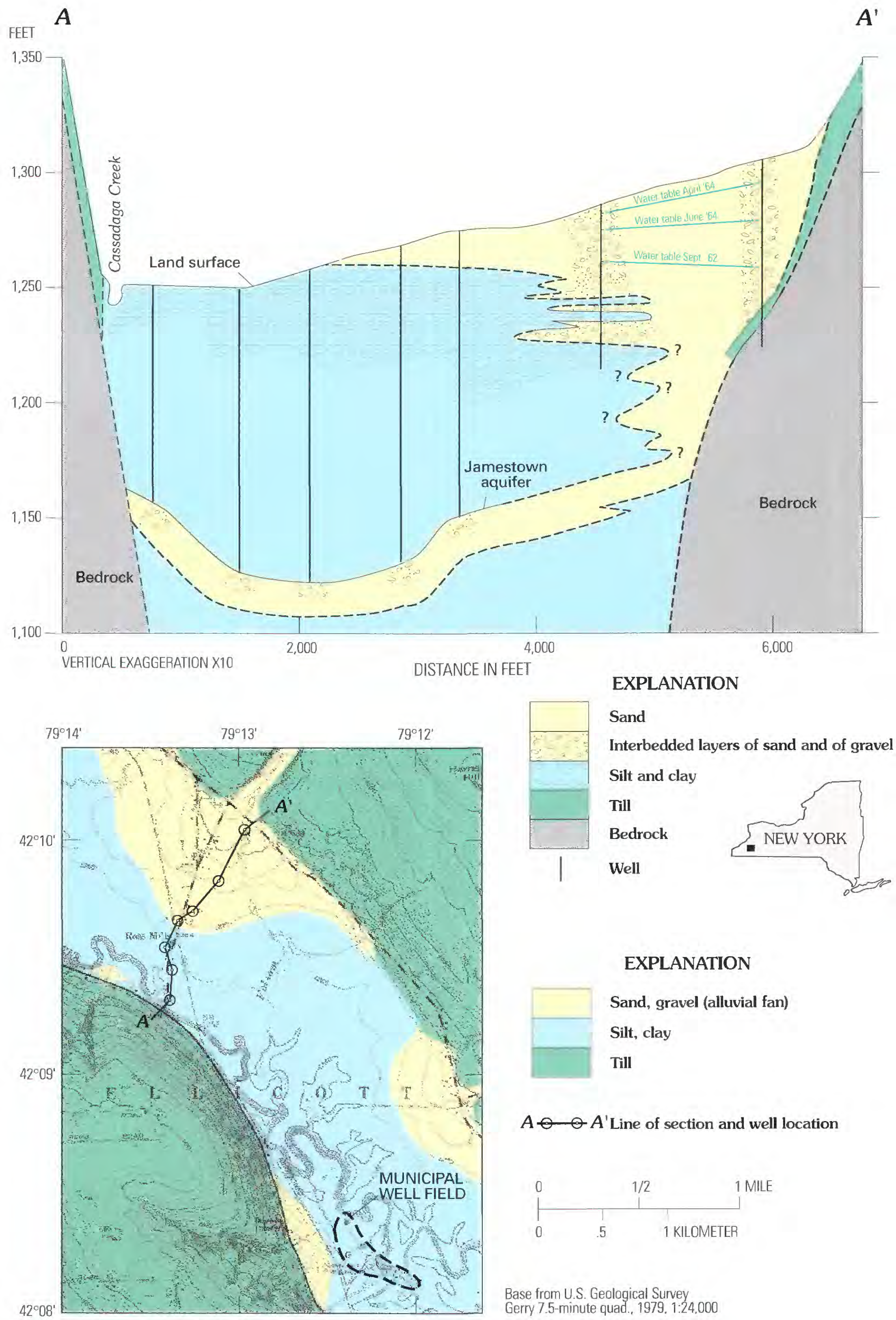


FIGURE 30.—Geologic section across alluvial fan and delta (or subaquatic fan) in Cassadaga Creek valley north of Jamestown, N.Y. (Modified from Crain, 1966, fig. 18.)

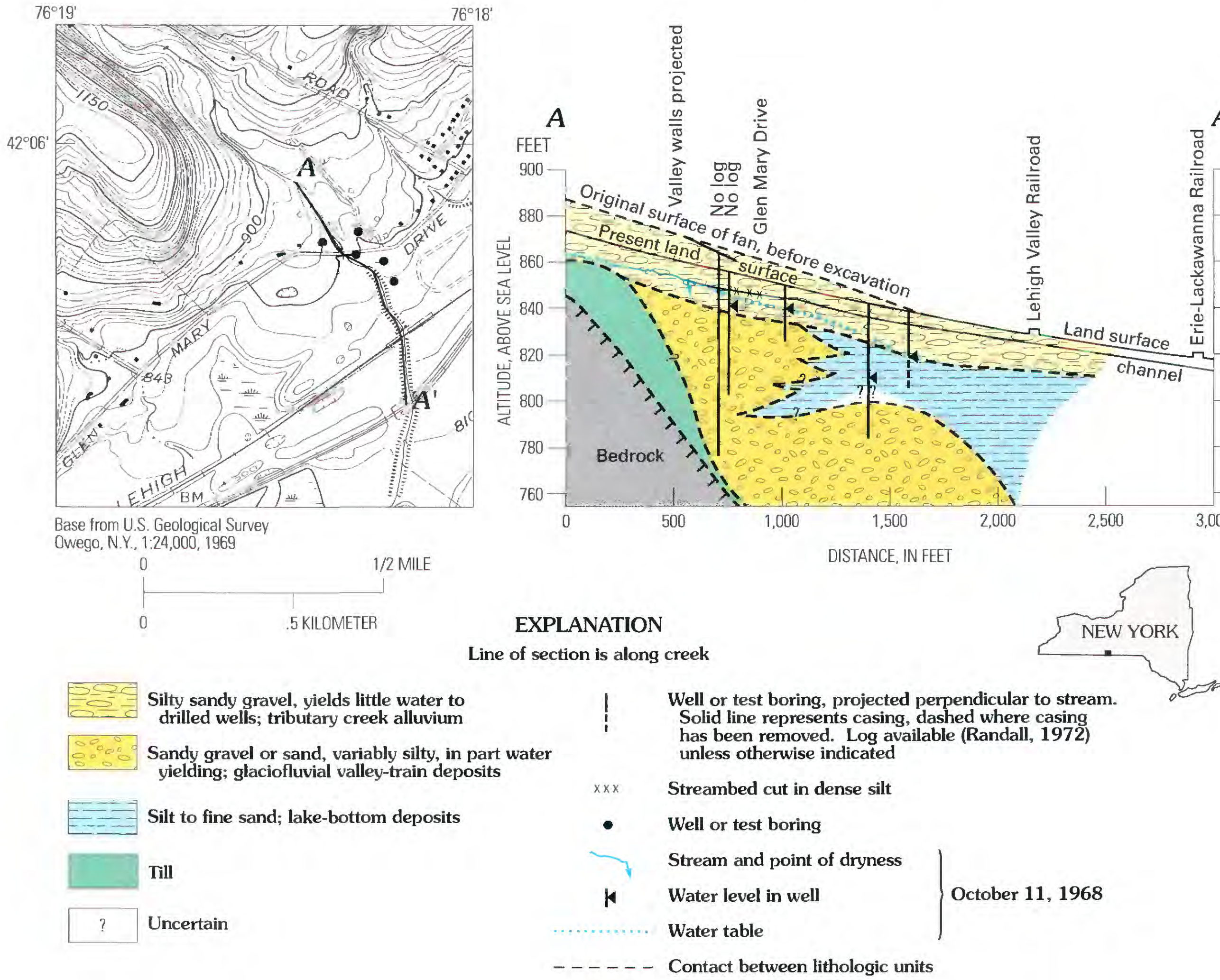


FIGURE 31.—Geologic section across alluvial fan in the Susquehanna River valley near Owego, N.Y. (From Randall, 1978b, fig. 1.)

production well 7 was derived from the Chenango River by induced infiltration, which moved to the well along the downwarped outwash layer on the northeast side of an ice-block depression (fig. 34, section B–B'). The path of the induced infiltration is shown by the solid arrows in figure 35. A few miles to the west, in the towns of Union and Vestal, at least four ice-block depressions contain fine-grained lake-bottom sediment (fig. 36). Borehole logs demonstrate that one of these ice-block depressions, near Snapp Jr. High School, is underlain by downwarped surficial outwash (fig. 37); the boreholes are not generally deep enough to indicate what underlies the outwash, but calibration of a digital aquifer model suggested some hydraulic continuity between the downwarped outwash and basal ice-contact deposits tapped by several municipal wells nearby (Randall, 1986, p. 28). Borehole logs suggest that downwarped outwash beneath another depression, on the south side of the Susquehanna River, also abuts ice-contact deposits (fig. 37). Continuity and ground-water flow through downwarped outwash beneath the depression near the mouth of Nanticoke Creek and the sewage disposal (fig. 36) are demonstrated by numerous borehole logs and by movement of organic solvents from a landfill west of Nanticoke Creek to a public-supply well (Endicott Ranney well, fig. 36) east of the creek (Adams and Grant, 1984). These examples indicate that ground water often can flow readily through downwarped outwash beneath ice-block depressions and suggest that well yield would benefit from the greater saturated thickness and potential drawdown at these locations.

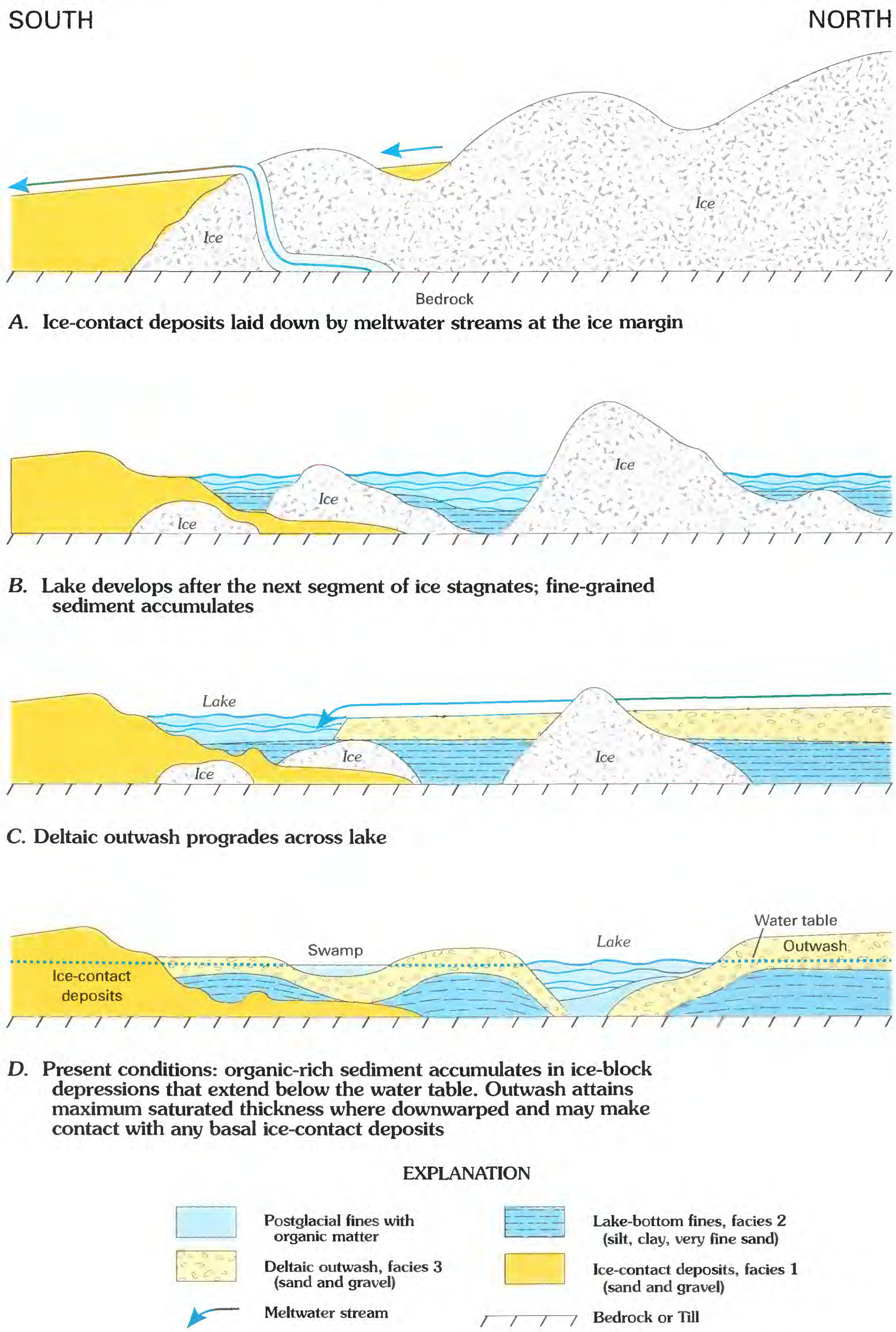


FIGURE 32.—Development and hydrologic significance of typical ice-block depressions in outwash deposits.

TABLE 3.—*Geologic units in the glacial drift along the Susquehanna and Chenango River valleys in southwestern Broome County, N.Y.*

Geologic unit	Distribution and lithology
Fill	In natural and excavated depressions, chiefly trash and ashes; in levees and embankments, natural materials.
River alluvium	Beneath flood plains of large streams; silt to fine sand that may contain organic-rich layers, overlying a few feet of sandy pebble-cobble gravel.
Older river alluvium	Caps low terraces, interfingers with postglacial lake beds in some depressions; sand and gravel.
Postglacial lake beds	In ice-block depressions; silt and very fine sand with some clay and scattered plant fragments; commonly grades to peat or highly organic silt at the top; may contain thin lenses of sandy gravel near streams.
Stratified drift	
Outwash	Caps broad terraces; sandy pebble gravel and coarse-fine sand; commonly deltaic; bright appearance because of numerous exotic pebbles. Facies 3.
Lake beds	Silt; silt with clay layers; silty very fine sand: commonly underlies outwash. Facies 2.
Ice-contact deposits	Sandy pebble-cobble gravel and pebbly sand, clean to very silty; locally includes thick silt layers; drab appearance because of few exotic pebbles. Facies 1.
Till	Immediately overlies bedrock; commonly thin, but forms rounded hills in some localities.
Bedrock	Shale, siltstone.

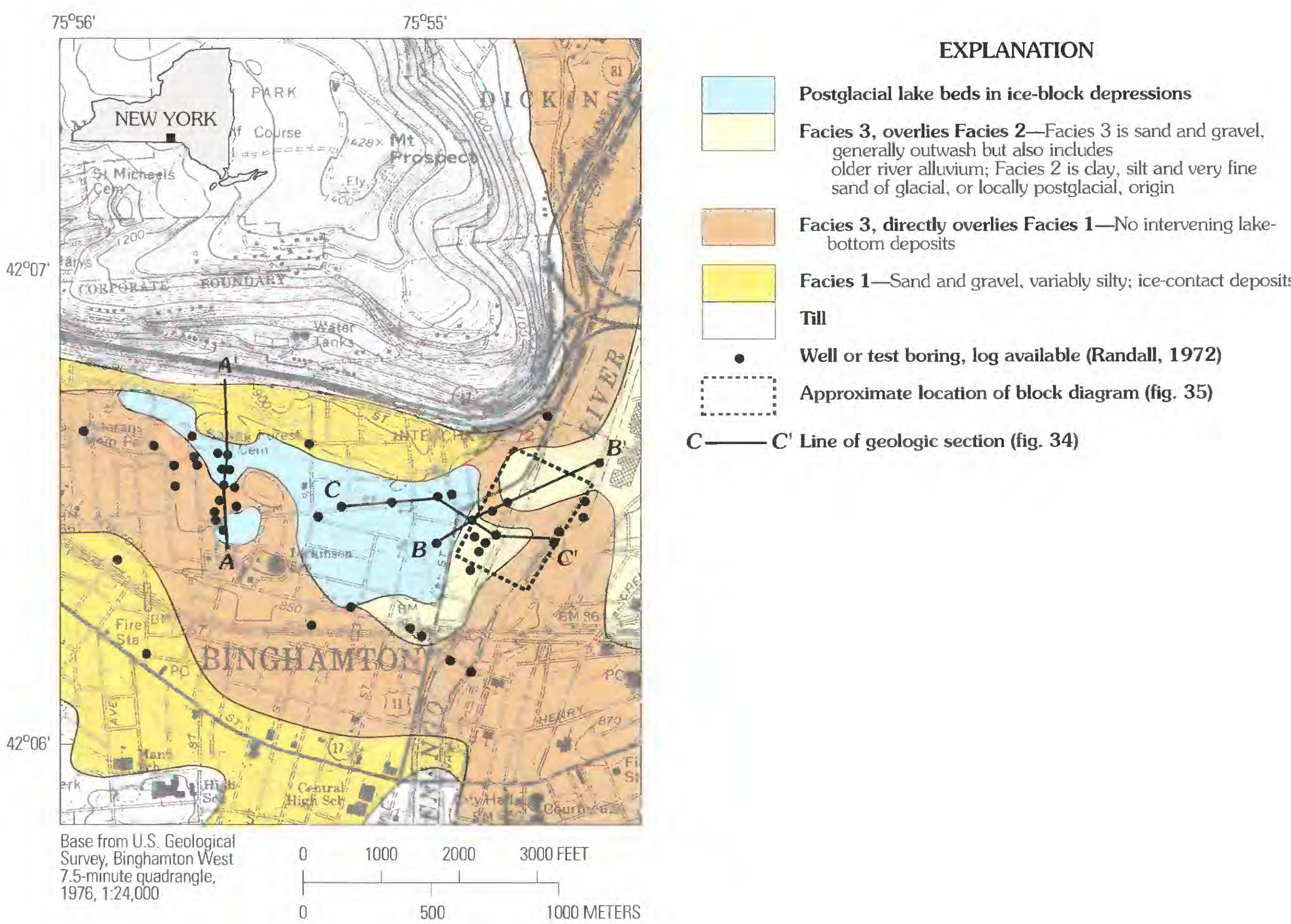


FIGURE 33.—Geohydrology of the eastern part of the Clinton Street-Ballpark aquifer, Binghamton, N.Y. (Modified from Randall, 1986a, pl. 1.)

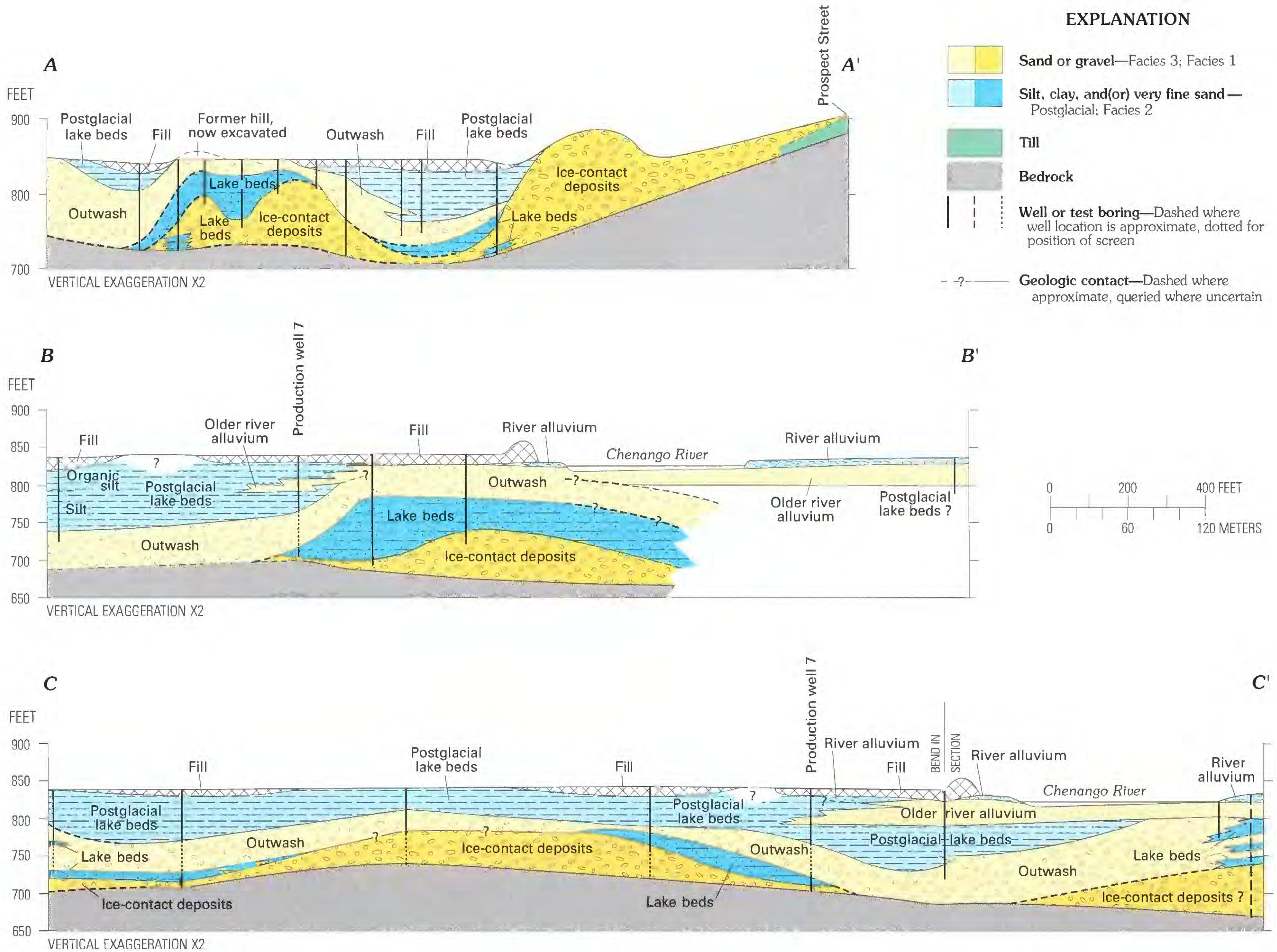


FIGURE 34.—Geologic sections in the eastern part of the Clinton Street-Ballpark aquifer, Binghamton, N.Y. (From Randall, 1977, fig. 3.) Location of sections shown in figure 33.

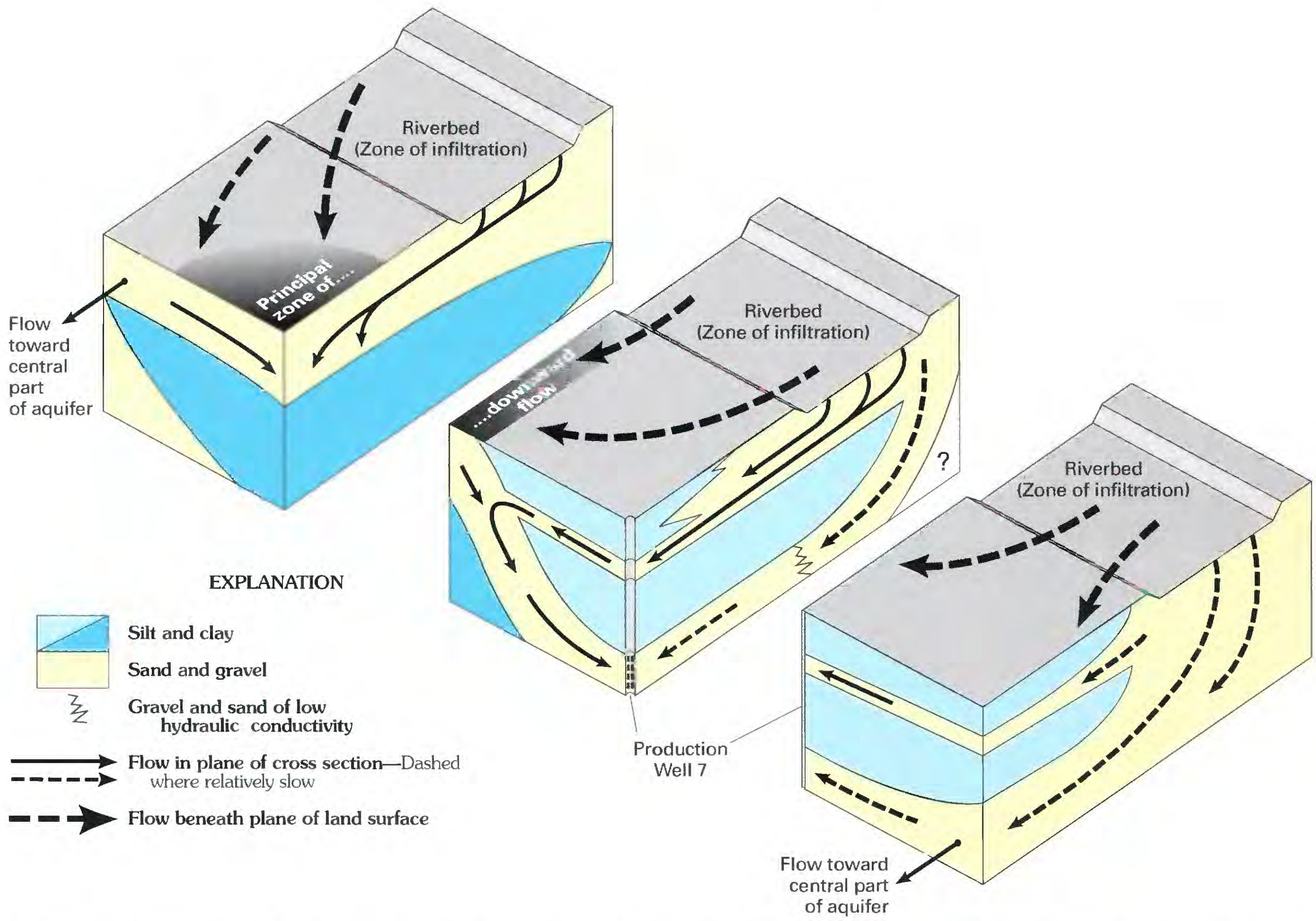


FIGURE 35.—Movement of induced infiltration from Chenango River to a production well. The three blocks represent adjacent segments of valley fill, sliced and separated to show the aquifer geometry. (Modified from Randall, 1977, fig. F1.) Location of blocks shown in figure 33. Not to scale.

CLASSIFICATION AND MAP REPRESENTATION OF AQUIFERS

The foregoing sections have described several typical patterns or concepts that generally characterize aquifer geometry in the glaciated Northeast, and have suggested how they can be recognized amid the sometimes bewildering heterogeneity of the stratified drift. The discussion has emphasized the three-dimensional spatial variation in saturated thickness of sand and gravel and in distribution of sand and gravel relative to fine-grained sediments. These aspects are important in locating sites for large-yield wells, in delineating capture areas around such wells, and in identifying ground-water flow paths. When the water-transmitting properties of earth materials and the location and magnitude of recharge sources also are considered, aquifer yield can be estimated.

Despite the many influences on aquifer form and distribution and the heterogeneity that is so common within even single exposures of stratified drift, aquifer geometry can be and has been mapped. For example, the three-dimensional spatial distribution of the three universal depositional facies can be represented effectively on maps by "stacked units" such as coarse/fine, fine/coarse, coarse/fine/coarse, coarse only, and fine only (Randall and others, 1966; Meade, 1978; Stone and others, 1992; Canace and others, 1993). Alternatively, saturated thickness or transmissivity of surficial aquifers (facies 1 and 3) could be represented by a set of map patterns or contours, while contrasting patterns or overlays could be used to represent any facies-1 aquifers confined or buried beneath fines. For example, MacNish and Randall (1982) divided surficial sand and gravel in the Susquehanna River basin of New York into three ranges of saturated thickness and divided buried aquifers into three depth intervals. Moore

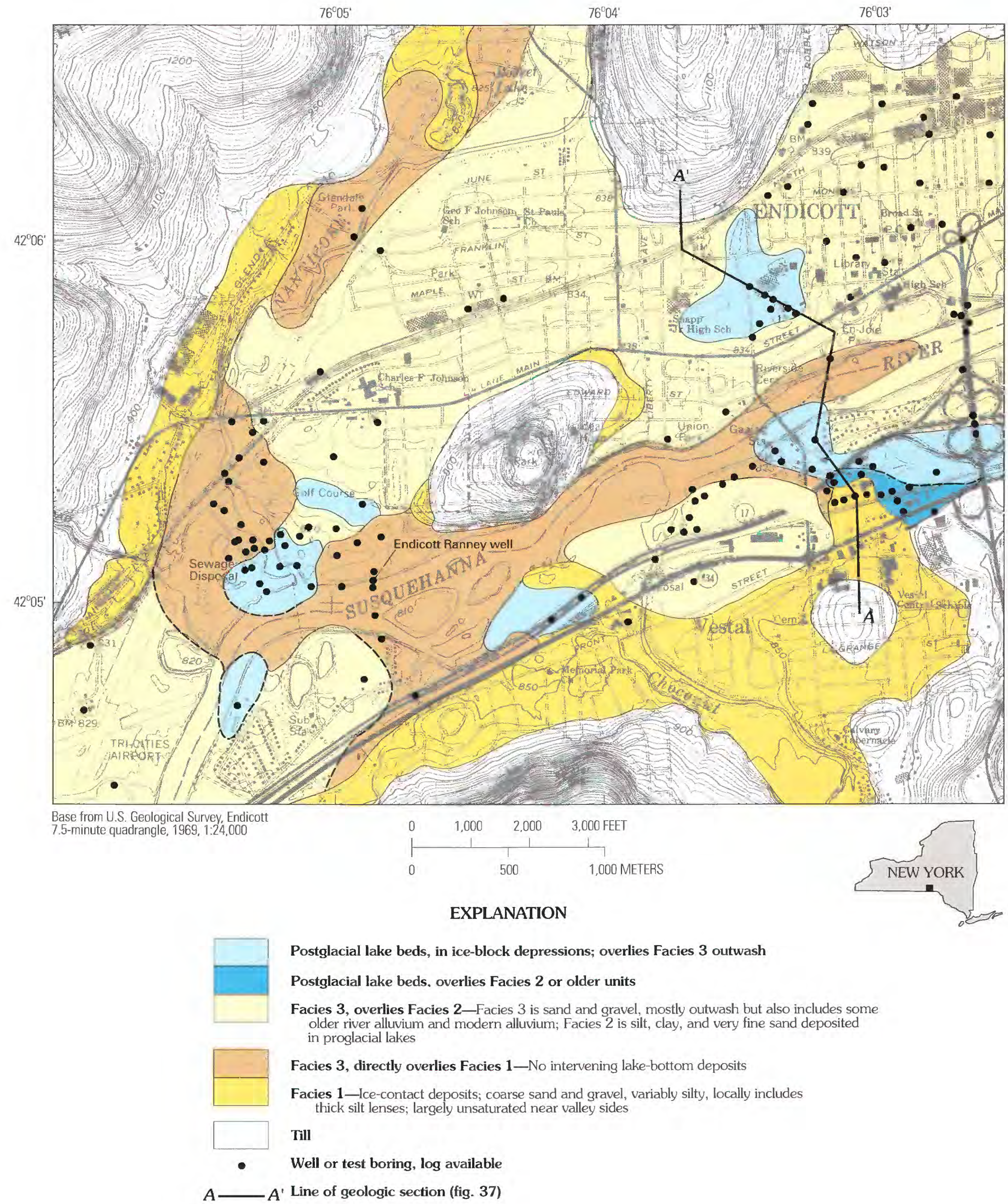


FIGURE 36.—Geohydrology of the Susquehanna River valley in the towns of Union and Vestal, N.Y. (Modified from Randall, 1986a, pl. 1.) Logs of wells are given in Randall (1972), Martin and others (1983), Adams and Grant (1984), and Ecology and Environment, Inc. (1986).

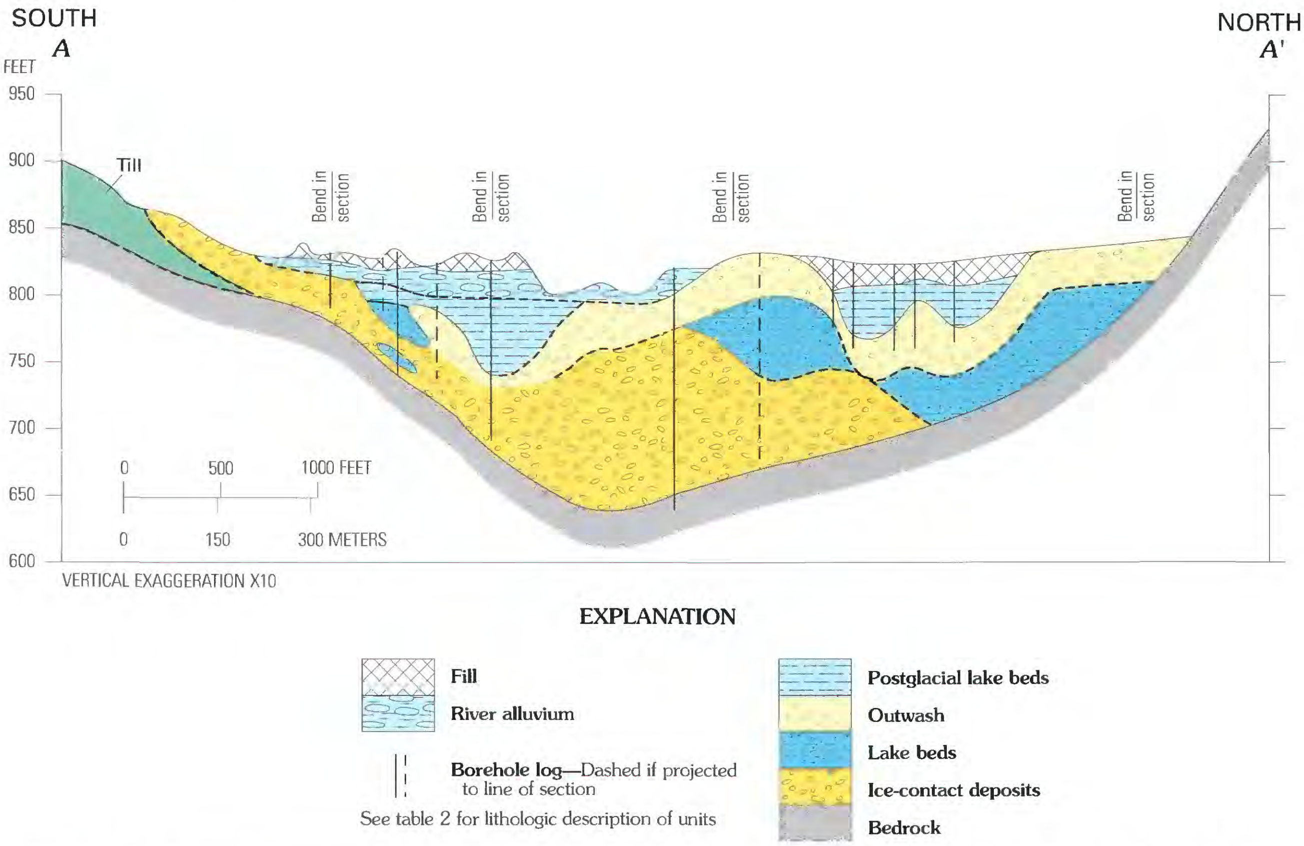


FIGURE 37.—Outwash downwarped beneath ice-block depressions in the Susquehanna River valley in Union and Vestal, N.Y. Location of section shown in figure 36.

(1990) and Stekl and Flanagan (1992) depicted aquifer geometry in southern New Hampshire by stacked units expressed as patterns. They also presented contours of saturated thickness and patterns for several ranges of transmissivity; the saturated thickness contours pertain to the entire unconsolidated section rather than only to aquifers, however, and the patterns pertain only to the principal aquifer (surficial or buried) at any location. Extremely complex maps or multiple overlays would be required to fully describe the valleys in southwestern New York, Pennsylvania, and Ohio that contain multiple drift sheets.

Aquifer geometry, saturated thickness, and transmissivity of the stratified drift can change radically over distances of a few hundred feet and therefore cannot be delineated accurately at scales much smaller than 1:50,000. A map(s) delineating any of these properties over the entire glaciated Northeast would be unwieldy at a large scale or overgeneralized at a small scale. Such a map(s) also would be deceptively unreliable because surface and subsurface data and comprehensive interpretation are lacking for so many localities. The remainder of this paper addresses regional differences in aquifer geometry in general terms, however, by delineating several regions and characterizing distinctive aspects of the hydrogeologic framework of the stratified-drift aquifers within each region by brief descriptions or sketches.

Aquifer yield is a function not only of the dimensions and water-transmitting properties of saturated sand and gravel, but also of aquifer location in relation to sources of recharge. The most productive aquifers are surficial riparian aquifers whose yields can be sustained by induced recharge from large streams in addition to recharge from local precipitation. Some aquifers, however, are useful precisely because they do not adjoin large streams and, thus, can be pumped down seasonally or during droughts without depleting flow in any large stream when streamflow is already critically low. This concept is explained by Randall and others (1988). A classification of aquifers on the basis of their relation to hydraulic boundaries and recharge sources is illustrated in figure 38. Those relations are important in evaluating aquifer yield and designing aquifer models, as discussed briefly below and in more detail in Chapter C of this Professional Paper (Kontis, Randall, and Mazzaferro, in press).

Most stratified-drift aquifers are in valleys, bordered laterally by till and bedrock at the base of upland hillsides (fig. 38*A*1–*A*4). Some valley fills are predominantly coarse (fig. 38*A*1), whereas others contain all three depositional facies, with facies 1 linked to facies 3 (fig. 38*A*2), or separated from facies 3 (fig. 38*A*3), or with multiple aquifers (fig. 38*A*4). Because the adjacent bedrock transmits ground water into the stratified drift only at slow rates, the valley walls are commonly simulated as barrier boundaries in analytical or digital models. Nevertheless, the uplands are a major source of

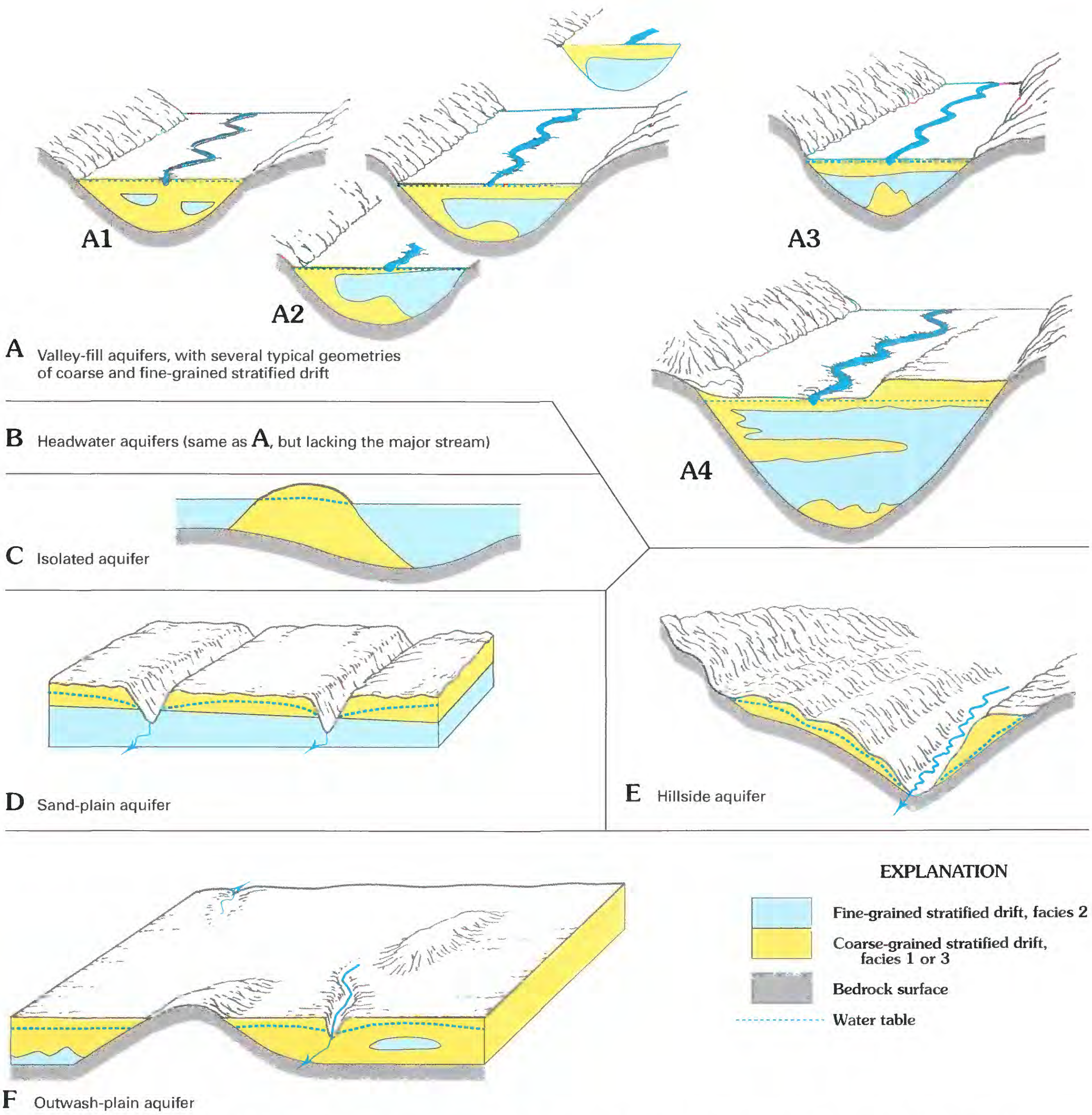


FIGURE 38.—Geometry of idealized aquifer types that differ in their relations to hydraulic boundaries and recharge sources.

recharge to the valley-fill aquifers (Morrissey and others, 1988), chiefly by means of the runoff in channels of tributary streams but also by unchanneled surface or subsurface runoff. Most surficial valley-fill aquifers are crossed by large trunk streams that could provide abundant induced recharge if ground-water development were to take place; those that are crossed only by small, ephemeral tributaries that originate locally in the adjacent uplands are classified separately as headwater aquifers (fig. 38*B*).

Other types of aquifers are less numerous than valley-fill aquifers but nevertheless are widely distributed across the glaciated Northeast. For example, broad lowlands commonly contain isolated kames, kame deltas, or ice-channel fillings (facies 1) bordered and overlapped by fine-grained sediment (fig. 38*C*). Some coarse ice-contact deposits (facies 1) on hillsides are perched above stream grade (fig. 38*E*). Some outwash or alluvial deposits (facies 3) on valley floors are also perched above stream grade, where postglacial streams have incised their channels into the underlying fine-grained sediment (fig. 38*D*). All these perched deposits are distinctive because saturated thickness is commonly small and induced recharge is impossible. In a few localities, surficial outwash is so thick and extensive that it inundates the bedrock topography (fig. 38*F*) and, thus, approaches the condition of an infinite aquifer without boundaries, in which upland hillsides are of minor importance as recharge sources. A few aquifers are completely buried beneath fine-grained sediment and, being separated from surficial aquifers, receive recharge only by slow seepage through bedrock or the fine-grained sediment (fig. 38*A*3, *A*4).

A common approach in modeling flow in valley-fill aquifers has been to treat the entire valley fill as the aquifer. This approach was used in some valleys only because the aquifer geometry was not well enough understood to do anything else. This approach is entirely appropriate, however, for valley fills that consist almost entirely of facies 1 (and/or 3) with only small lenses of fine-grained sediment (fig. 38*A*1). It also may be appropriate for valley reaches that contain extensive lake-bottom deposits (fig. 38*A*2–*A*3) if those deposits are chiefly fine sand to coarse silt that is sufficiently permeable to transmit water fairly readily to wells that tap coarser sand and gravel. A more sophisticated approach that can accommodate any combination of aquifer geometry represented in figure 38*A*1–*A*3 would be a two-layer quasi-three-dimensional model (Randall, 1986; Bergeron, 1987). By contrast, the aquifer geometries shown in figure 38*B*–*E* could all be simulated by models with a single active layer and various boundary conditions. Kontis, Randall, and Mazzaferro (in press) discuss in more detail the design of models to simulate typical aquifer geometries and the estimation of hydraulic properties of aquifers for use in those models. Except near the edges, valley-fill and outwash-plain aquifers typically contain 50 to 100 feet of saturated sand and gravel that typically ranges from 50 to 100 feet per day in hydraulic conductivity, resulting in transmissivity between 2,500 and 50,000 feet squared per day. Saturated thickness of sand-plain and hillside aquifers tends to be inversely proportional to hydraulic conductivity, so transmissivity is likely to average around 2,000 feet per day in sand-plain aquifers, probably less in hillside aquifers.

HYDROPHYSIOGRAPHIC REGIONS IN THE GLACIATED NORTHEASTERN UNITED STATES

The glaciated Northeast has been divided into regions (pl. 1) that differ in the typical distribution or geometry of coarse-grained stratified drift with respect to fine-grained stratified drift, till, bedrock, and streams. The regions are termed “hydrophysiographic regions” because their delineation is based on aspects of hydrology that can be important in the evaluation of aquifer yield or water-resource development, and yet those aspects are, in part, functions of such physiographic properties as relief, slope orientation, drainage density, and size of tributary watersheds, and boundaries of some regions on plate 1 correspond closely to physiographic boundaries delineated by Fenneman (1938), Cressy (1966), and Denny (1982). The salient characteristics of these regions are summarized in the next few paragraphs. The remainder of the paper is devoted to capsule descriptions of individual regions.

In about 70 percent of the glaciated Northeast, major valleys sloped generally away from the ice sheet, and waterborne sediment was deposited as stratified drift in a succession of local lakes within those valleys. The lakes drained across recently deposited stratified drift downvalley or across saddles on the divides. This large area contains a wide variety of terranes, however.

1. Southern New England, eastern New York, and northern New Jersey are generally characterized by low to moderate relief with closely spaced, small valleys. Most proglacial lakes in these valleys were relatively small and shallow, deglaciation took place slowly, and large seasonal meltwater volumes were generated from the vast

expanse of ice to the north. Consequently, coarse-grained stratified drift (facies 1 and 3) is abundant, especially at the numerous morphosequence heads or ice-marginal positions. In a few areas of low relief, sandy deltaic or fluvial outwash is abundant enough to bury the preexisting topography, creating extensive outwash-plain aquifers.

2. Other regions, however, have equally low to moderate relief but much less stratified drift. In northern and northeastern Maine, volumes of coarse-grained stratified drift are small because a nearly stagnant residual ice cap that no longer transported sediment was the only source of meltwater at the time this region was deglaciated. The central part of this region, which contains almost no stratified drift, is bordered on the northwest by an area where valley floors are underlain by silt and clay, on the northeast by an area where valley floors are underlain by sand and gravel largely buried beneath till, and on the southeast by an area where narrow eskers commonly follow the valleys but do not fill them. Other areas characterized by low internal relief and by small valleys that generally lack significant stratified-drift aquifers include the Tug Hill Plateau of New York and Pocono Plateau of Pennsylvania (although relief is high along plateau margins).

3. Several regions have consistently high relief. Those in northern New England and northern New York are characterized by small areal extent of coarse stratified drift, much of which is perched above stream grade on the valley sides where it is easily drained. Bedrock outcrops are fairly common in the flood plains, so valley-fill aquifers must be even narrower than the narrow valley floors. By contrast, the Catskill Mountains and Appalachian Plateau of New York, Pennsylvania, and Ohio have little stratified drift perched on the valley sides, and bedrock outcrops do not occur on the valley floors. In the Catskill Mountains, valley fills rarely exceed 150 feet in saturated thickness; they include till (deposited chiefly on the lee side of hills) and silty to clean sand and gravel but relatively little fine-grained stratified drift. In the Appalachian Plateau from the Catskills west to Salamanca, N.Y., valley fills commonly exceed 150 feet in saturated thickness, so silt and clay are correspondingly thicker and more extensive, although commonly capped and underlain by sand and gravel. The sand and gravel are especially thick at ice-marginal positions and in wedges of valley-train outwash. West of Salamanca, stratigraphy in some valleys is similar to that farther east, but relief is smaller, the network of intersecting valleys more complex, and many valleys contain multiple till layers or drift sheets.

In about 10 percent of the glaciated Northeast, deep valleys sloped northward toward the ice sheet. Valleys in the northern 10 to 30 miles of the Appalachian Plateau were deeply eroded and later filled with thick drift, predominantly layers of diamicton and fine-grained stratified drift. Sand and gravel occur as discontinuous lenses at multiple depths, especially near tributaries. In the northern reaches of some valleys in this region, and also in northwestern Vermont where ice similarly advanced against a steep slope, most valleys contain chiefly fine-grained sediments, but stratigraphy seems to conform to the 3-facies single-deglaciation model.

About 20 percent of the glaciated Northeast consists of broad lowlands that were inundated by large proglacial lakes or marine waters during deglaciation. Clay, silt, and fine sand widely mantle the lower parts of the landscape, interrupted by till-covered hills. Ice-contact sand and gravel underlie only a small fraction of the inundated areas, commonly as ice-channel fillings or esker deltas, but also as thin basal sand sheets in some places. Surficial facies-3 sand aquifers are absent or insignificantly thin in many lowland areas but are extensive in the relatively narrow Connecticut and Hudson River lowlands and elsewhere near mountain fronts. Very few opportunities for induced infiltration exist anywhere in these lowlands.

South of the glacial border, in Ohio, Pennsylvania, and New Jersey, valleys that drained away from the ice contain sandy valley-train deposits that, despite post-glacial incision, still constitute aquifers in many valley reaches.

The capsule descriptions of individual regions that follow were written from the bottom up; that is, any generalizations or insights about the hydrogeologic framework of each region that could be extracted from maps or other literature were summarized and illustrated by typical examples, rather than attempting to select representative values from each region to populate some preconceived array of aquifer characteristics or dimensions. Persons who are primarily interested in one or more specific localities need not read all the material that follows but should examine the description of the appropriate region(s), as shown on plate 1, for concepts and references that apply to those particular localities.

GLACIATED REGIONS THAT GENERALLY SLOPED AWAY FROM THE ICE

In about 70 percent of the glaciated Northeast, the major valleys sloped away from the retreating late Wisconsinan ice sheet, chiefly as a result of pre-Pleistocene geomorphic development, but locally because streams were diverted southward when their former valleys were blocked by earlier ice sheets. Generally, these major valleys contained a succession of small proglacial lakes during deglaciation and received large volumes of sediment-laden meltwater. They now contain nearly continuous surficial deposits of coarse stratified drift that also extend across saddles between valleys in some areas of low relief and are locally underlain or interrupted by fine-grained stratified drift. Some tributary valleys and some segments of major valleys sloped toward the ice and contained proglacial lakes that spilled through saddles on the watershed perimeter, but commonly these lakes were similar in size and depth to the lakes that spilled over older stratified drift in the more numerous valleys that sloped away from the ice. The stratified-drift-filled valleys are separated by much more extensive uplands in which the bedrock is mantled predominantly by till. The uplands are incised by small tributary valleys whose floors are above the grade of the stratified drift in the major valleys. The tributary valleys are floored by several feet of alluvium but contain no stratified drift, other than a few isolated deposits on hillsides. A few uplands are large enough to be shown as areas of scant stratified drift on regional maps.

REGIONS WITH MODERATE TO LOW RELIEF AND ABUNDANT COARSE STRATIFIED DRIFT

Much of southern New England, eastern New York, and northern New Jersey is a region of moderate to low relief in which valleys that contain stratified drift are closely spaced but separated by till-mantled ridges or uplands. Sand and gravel underlie 15 to 35 percent of most large watersheds. This nearly contiguous region (pl. 1) is referred to herein as the Eastern Hills and Valley Fills region. A few parts of southeastern New England are designated as outwash plains regions on plate 1 because relief is especially low and the entire landscape is inundated by abundant sand and gravel. Elsewhere in the glaciated Northeast, several small areas have been assigned to the outwash plains and Eastern Hills and Valley fills regions because the sand and gravel they contain are similarly distributed.

EASTERN HILLS AND VALLEY FILLS

Local relief in the Eastern Hills and Valley Fills region is less than 1,000 feet, commonly much less, as indicated for much of the region by Denny (1982, fig. 10). Deglaciation took place chiefly by stagnation-zone retreat, during which a succession of morphosequences developed along most valleys. Proglacial lakes were generally small and shallow enough that sandy or gravelly facies-1 deposits predominate in many places. Silt and very fine sand, generally capped by coarser sand and (or) fine gravel, were deposited in the distal parts of morphosequences.

Stone and others (1982, p. 6) categorized the stratified drift of Connecticut into four depositional environments: (1) Major glacial lakes in lowlands, (2) ice-marginal ponding in north-draining valleys in uplands, (3) glaciolacustrine-glaciofluvial systems in south-draining valleys, and (4) glaciofluvial systems. Their descriptions and example maps indicate considerable stratigraphic similarity among the four categories, however. All categories generally consist of surficial fluvial sand and gravel (delta topsets or fluvial terraces) overlying delta foresets and (or) bottomsets, and all include fine-grained sediments at the distal ends of morphosequences. Categories 1 and 2 both include small areas in which surficial stratified drift is fine grained because deltas never prograded completely across lake-bottom deposits to fill the lake. Categories 1, 2, and 3 all include areas in which successive deltas abut one another and are difficult to distinguish. Only the largest proglacial lakes, where fine-grained lake-bottom deposits are continuous for miles and large tributary deltas and (or) fluvial terraces result in extensive thin surficial sand-plain aquifers, were deemed sufficiently distinctive in aquifer geometry to warrant separate classification in this paper.

The northwestern part of New York's Adirondack uplands is assigned to the Eastern Hills and Valley Fills region on plate 1, even though all major streams drain northward, because the volume, landforms, and relief of the stratified drift are similar to that in southern New England. During deglaciation, meltwater drained away from the ice sheet southwestward across low saddles and eventually to the Black River lowland. Stratified drift is extensive; it covers more than half of some quadrangles, and in some places it forms an anastamosing network of valley fills that connect across former saddles and surround till-mantled hills (Cadwell and Pair, 1991). Most hills rise only about 200 feet above the adjacent valley fills. Few drillhole records are available in this sparsely populated area, but surficial stratified drift is predominantly coarse grained, and morphosequences have been identified in many localities from topography and scattered exposures. In the Brother Ponds quadrangle, for example, cobble gravel at the head of an esker-fed morphosequence grades southeastward to sand beneath a smooth depositional surface that ends near a former saddle incised by the postglacial Grass River (J.T. Gurrieri, field maps, New York State Geological Survey files). Gurrieri (1983) described proximal subaquatic-fan or delta deposits at two ice-marginal positions near Saranac Lake and a gross morphosequence (probably multiple) extending southwestward from Loon Lake.

Three small areas within or bordering the Ontario lowland of western New York were slightly above the level of proglacial lakes that inundated most of this region and are characterized by distinct valleys containing well-defined morphosequences. One area, along the Wayne-Ontario County line, is described in the subsection "Interdrumlin Outwash" under "Erie and Ontario Lowlands." Another area, south of Tug Hill, contains several valley-fill aquifers, including much of the Tug Hill aquifer described by Miller and others (1988).

Example: Quinnipiac Valley in Southington, Conn.—Aquifer geometry in Southington, Conn., and adjacent towns (fig. 39) is reasonably well understood as a result of detailed surficial geologic mapping and compilation of borehole records. It is described here in some detail to illustrate the closely spaced variability that is typical of stratified drift in the Eastern Hills and Valley Fills region and how that variability can be deciphered. Early in deglaciation, meltwater drained southward through three saddles in a low divide in Cheshire (near the south edge of fig. 39). Later, meltwater drained eastward through the Quinnipiac Gorge (1.5 miles north of the southeast corner of fig. 39), although it was ponded 100 feet above the present river level for a while by drift barriers in the gorge (LaSala, 1961) or immediately to the west (Stone and others, 1982). Deposition of stratified drift in water ponded behind these saddles and drift barriers was undoubtedly a continuous process, but topography and lithology indicate that the resulting deposits can be divided into several areally distinct morphosequences (fig. 39) that reflect stagnation of successive marginal segments of the ice.

The distribution of water-yielding gravel and medium to coarse sand within the stratified drift was affected by at least four factors:

1. Deltaic Deposition.—Meltwater flow was consistently southward, and deposition was commonly deltaic. Hence, coarse sand and gravel are most abundant near the north margin of each morphosequence, whereas deposits farther south tend to be fine grained at depth but coarsen upward toward the delta surface, as illustrated in a geologic section (fig. 40). Cobble gravel layers can be found even at the south margin of deltas, however (Stone and others, 1982, p. 19).
2. Collapse.—Each morphosequence was built against or atop stagnant ice, which eventually melted, leaving irregularly shaped depressions. At least the margins of the depressions are commonly underlain by sediments similar to, and collapsed from, the adjacent landforms. Hanshaw (1962) observed anticlinal structure in several ice-channel fillings and ascribed it to collapse as the channel walls melted. Undisturbed delta surfaces are many feet above the modern water table, but surficial coarse sand and gravel that collapsed along the delta margins are now saturated.
3. Depth to Bedrock.—The retreating ice margin stagnated in successive segments, often at narrow or shallow valley reaches that offered maximum resistance to flow. Hence, the northern (proximal) ends of morphosequences commonly lie on shallow bedrock.
4. Feeder Channels.—Ice-channel fillings occur at the north margin of many morphosequences; some extend as much as a mile farther north as continuous or segmented sinuous ridges (fig. 39). Some probably formed in tunnels at the base of the ice sheet through which pressurized meltwater carried sediment southward to feed subaquatic fans and deltas at the ice margin—an interpretation suggested by their position at the north end of deltas and by coarse sediments observed in exposures and well records (Lougee, 1938; Simpson, 1959; LaSala, 1961; Mazzaferro, 1973, well S235; Stone and others, 1982, p. 19). Others, however, are merely deltas built in narrow lakes within ice crevasses. One well on an ice-channel filling in Cheshire penetrated a typical downward-fining deltaic section of gravel over sand over silt (Mazzaferro, 1973, well Cs186). Short north-pointing noses at the north margin of a few morphosequences may mark lesser feeder channels.

Eastern Southington.—Morphosequences 2, 3, and 4 (fig. 39) are separated by lowlands underlain by clay, silt, fine to very fine sand, and a little coarser sand, all of which were deposited on the lake bottom during and after deposition of these sequences. By contrast, sequences 4, 5, and 6 are part of a continuous pitted outwash plain (Hanshaw, 1962) of nearly uniform slope (fig. 40), interpreted as a single morphosequence by

LaSala (1961) but separated into three morphosequences by Stone and others (1982) at places where topography, excavations, and (or) well records indicate coarse, severely collapsed deposits that could mark an ice margin. Large stratified deposits in the Eastern Hills and Valley Fills region commonly consist of multiple deltas that are difficult to distinguish because they are built against each other and graded to a common lake level. Sequence 8 is much younger than the others; a thin fluvial sand and gravel layer overlies thick and extensive fine-grained lake-bottom sediments that separate sequence 7 from younger deltaic deposits to the north.

Most boreholes on the deltaic outwash plains in eastern Southington did not penetrate thick sections of saturated coarse-grained materials (fig. 39). Typically, surficial coarse sand and gravel either overlie bedrock at shallow depth (where deltas were built over bedrock knolls) or grade downward into finer sediment. Thick sections of saturated coarse sand and gravel were penetrated, however, by five of the six holes drilled on ice-channel fillings or in gaps between closely spaced segments thereof. The presence and general alignment of ice-channel deposits in successive morphosequences suggest that a continuous subglacial channel system traversed the entire valley in eastern Southington from north to south and was a major source of water and sediment to the retreating ice front. The ice-channel fillings identified in figure 39 were by no means the only sources of coarse sediment, however. Virtually all boreholes near the west side of sequences 5 and 6 penetrated predominantly coarse sands and gravels with substantial saturated thickness; this suggests that southward flows of meltwater and sediment through subglacial channels may have been greater in the valley west of Churchill Hill than in the valley to the east, which is wider but less deep. If so, the younger valley train and underlying lake-bottom deposits west of Churchill Hill (sequence 8) may conceal gravel deposited along a subglacial channel(s) that fed sequences 5 and 6. A few boreholes have penetrated coarse sediment at depth west of Churchill Hill, but continuity has not yet been demonstrated.

Many wells or test holes drilled in low areas north of, beside, or between deltaic outwash plains or ice-channel fillings penetrated saturated coarse-grained sediments that could be collapsed extensions of adjacent deposits, at least in part. Collapse at the north end of each sequence is depicted in figure 40, and borehole records (figs. 39, 41) demonstrate that coarse sand and gravel buried beneath fines extend at least 1,000 feet northward from the exposed heads of sequences 3 and 7. A few wells penetrate through deltaic sediments into gravel or coarse sand that are probably either early deposits in subglacial channels or the collapsed heads of ice-contact deltas intermediate between the morphosequences recognized in figure 39. One such deposit is partly exposed immediately south of morphosequence 4 (figs. 39, 40).

Western Southington.—Only three morphosequences were recognized by LaSala (1961) in the western part of the lowland in Southington. The second of these, sequence 2W in figure 39, is approximately time-equivalent to sequence 1. Ice-channel fillings are not obvious on the valley floor at the head of sequence 2W in Cheshire, but clearly sequence 2W was fed, in part, by meltwater that flowed between the ice and the upland to the west and deposited kame terraces high on the west side of the lowland (figs. 39, 41). The third morphosequence (3W) heads 6 to 8 miles farther north near the Bristol-Southington town line, where ice next stagnated in an area of shallow bedrock. Here, most deposits are coarse grained, thinly saturated, and much collapsed. A delta was built southward several miles across fine-grained lake-bottom sediments, ending in a natural delta front in southern Southington. Some shallow wells tap saturated deltaic medium to fine sand just below the water table, but otherwise, saturated sand and gravel are scarce beneath and south of this large delta, except in two places:

1. Along the west side of the valley, several boreholes penetrated thick coarse sand and gravel (figs. 39, 41). These deposits are likely to be collapsed margins of the higher kame terraces (sequence 2W) and later deposits of meltwater that continued to flow southward along the west side of the lowland for a while after base level dropped to that of sequence 3W, but before enough ice melted near the Bristol-Southington town line to trap coarse sediment there. An ice-channel filling near Lake Compounce in northwest Southington trends southward into gravel knolls that are higher than the rest of sequence 3W, suggesting deposition along a meltwater channel early in the history of the morphosequence.
2. South of the delta front, near the Cheshire-Southington town line, many (but not all) wells penetrate 10 to 40 feet of gravel or sand between the lake-bottom deposits and bedrock. These coarse sediments do not thicken southward toward the head of sequence 2W but may thicken westward toward the kame terrace (fig. 41). They are commonly interbedded with fine sand to silt and could represent local subaquatic fan deposition beneath a floating margin of the ice tongue.

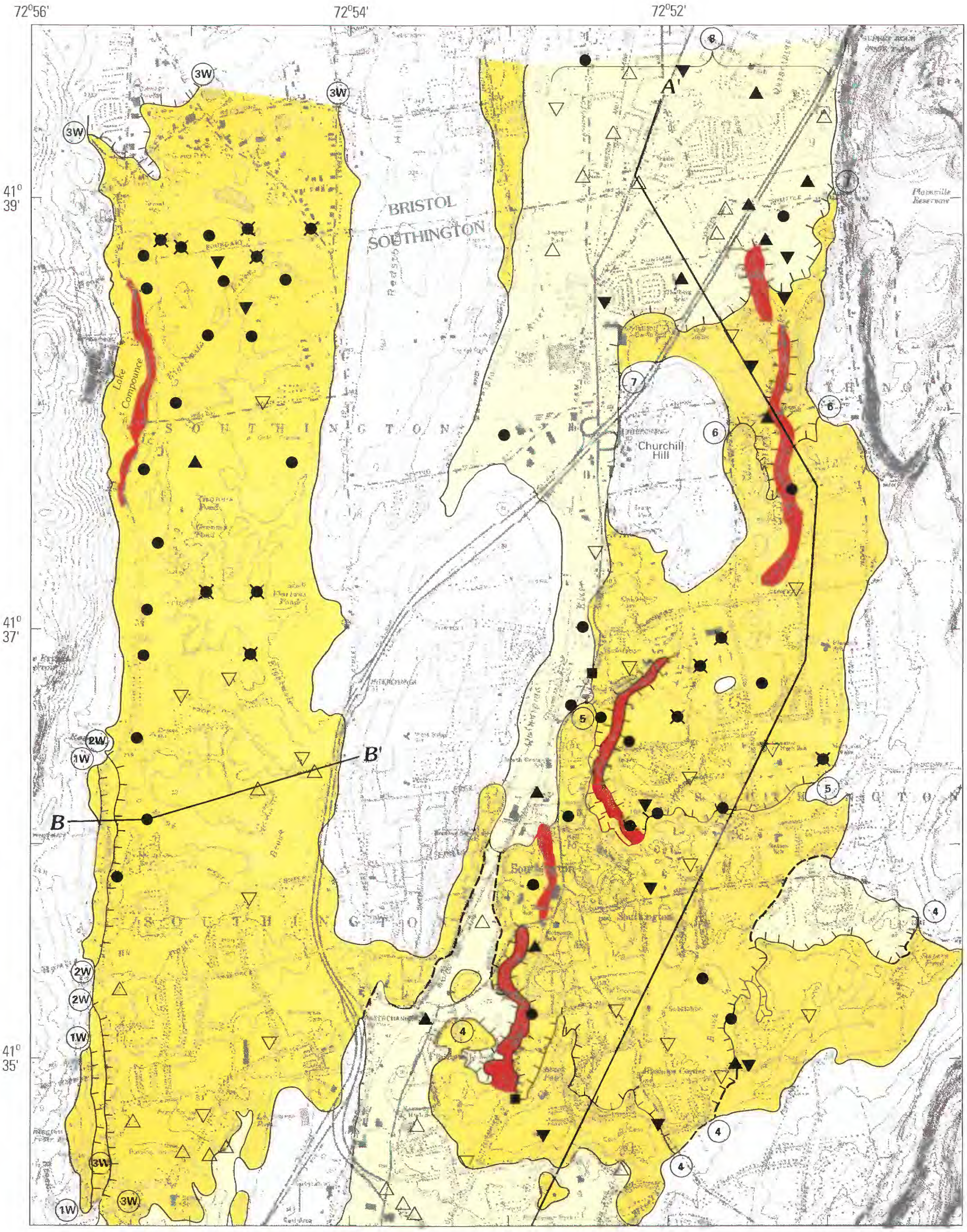
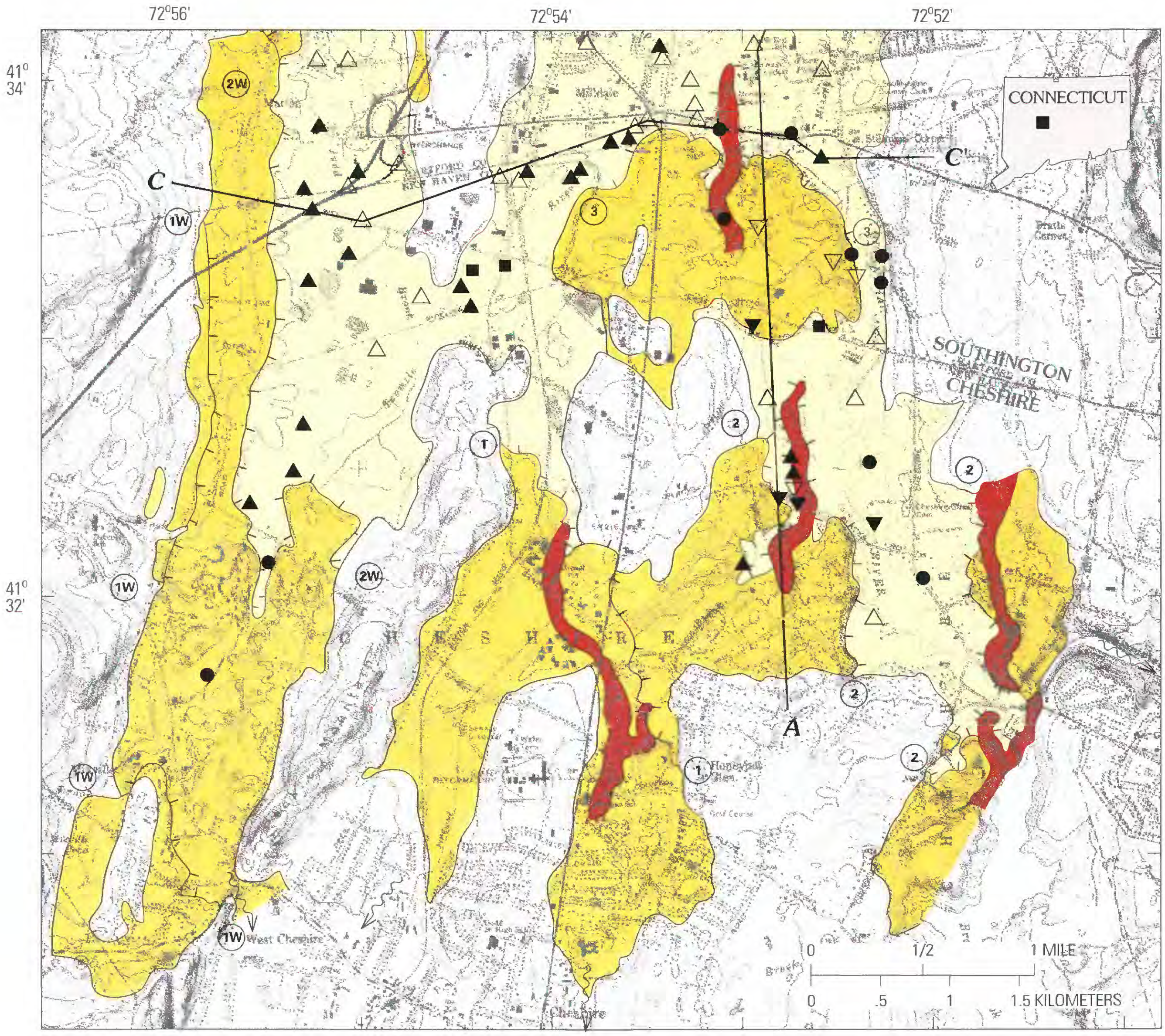


FIGURE 39.—Surficial geology in Southington and adjacent towns, Connecticut. Geology adapted from Stone and others (1982), LaSala (1961), Hanshaw (1962), and Simpson (1959). Borehole data chiefly from Haeni and Anderson (1980) and Mazzaferro (1973).



Base from U.S. Geological Survey, Bristol, New Britain, Meriden, and Southington, Conn., 7.5-minute quadrangles, 1984, 1:24,000

EXPLANATION

Till, bedrock

Ice-contact stratified drift, predominantly deltaic

Lake-bottom deposits, commonly capped by thin fluvial outwash or alluvium

(1W) (1W) Approximate proximal limit of morphosequence, including collapsed deposits not known to be mantled by younger sediment—Ticks point toward ice. Numbers at ends of limit lines indicate relative age (1 is oldest); numbers followed by W indicate location in western Southington

(8) Morphosequence that heads north of map area

Trend of ice-channel filling, or closely spaced segments thereof

C —— C' Line of geologic section (see figs. 40 and 41)

Saddle or channel through which meltwater drained

LITHOLOGY OF STRATIFIED DRIFT AT BOREHOLES

▽ Coarse to medium sand to gravel, grading down into medium-fine sand, very fine sand, or silt (all deltaic)

△ Fine sand to clay (lake bottom) commonly capped by thin sand or gravel (fluvial)

▲▼ Same as open triangles, but fines are underlain by medium-coarse sand or gravel

■ Well ends in sand or gravel, but no record of lithology at shallow depth

● Stratified drift is predominantly medium-coarse sand or gravel; X indicates saturated thickness is less than 25 feet over till or bedrock

FIGURE 39.—Continued.

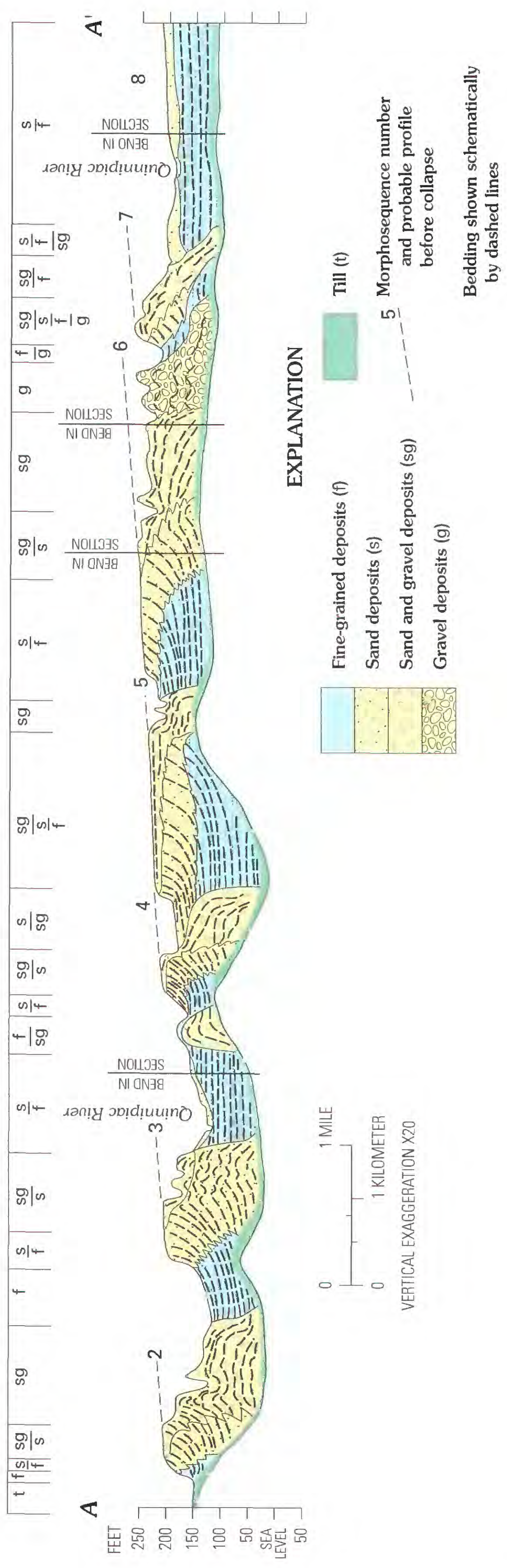


FIGURE 40.—Geologic section along the Quinnipiac lowland in Southington, Conn., and adjacent towns. Trace of section shown in figure 39. The section crosses seven morphosequences, numbered in order of deposition as in figure 39. Stacked symbols in bar across top of section indicate vertical distribution of lithology in segments of the section. (Modified from Stone and others, 1992, fig. 2.)

Evaluation.—Valleys in the Eastern Hills and Valley Fills region may contain multiple deltas, each partly collapsed and overlapped by subsequent deltas. If so, aquifer geometry is likely to be complex, and detailed surface and subsurface exploration may be required to decipher it. The first step in characterizing aquifer geometry is to map the extent of stratified drift. Commonly, the easiest second step is to identify and delineate any extensive deposits of fine-grained sediment, either surficial or capped by thin facies-3 deltaic or fluvial sand. The remaining stratified drift ordinarily constitutes an aquifer of variable thickness and transmissivity. Within such an aquifer, relatively coarse deposits that include the most suitable sites for large-capacity wells can be located through application of several concepts: northward coarsening within individual morphosequences, continuous coarse-grained deposits in feeder channels that are now visible as discontinuous ice-channel fillings, and collapse of coarse deltaic sand and gravel below the water table along ice-contact margins. In Southington, the large delta in the southern part of sequence 3W, the fluvial valley train of sequence 8, and the lowlands south of sequences 3, 4, and 3W are readily identified as the only extensive areas of fine-grained deposits, in part capped by a thin surficial aquifer. The localized presence of buried aquifers beneath these fine sediments, but remote from any collapsed margins, defies prediction so far.

OUTWASH PLAINS

In some regions of low relief, coarse-grained stratified drift is so abundant that it not only fills valleys and crosses saddles but also completely buries low hills or ridges, surrounds higher hills, and blankets 50 to 100 percent of the land surface. A few such regions, each several tens of square miles in areal extent, are delineated on plate 1. Most of these regions include outwash plains or sandurs: broad expanses of sand and gravel deposited by multiple coalescing braided meltwater streams in front of an ice margin. Locally, the outwash progrades southward across fine-grained sediments of lacustrine or marine origin. Most of these regions are termed interlobate because the meltwater and stratified drift were concentrated in temporary lowlands between walls or lobes of ice. The sand and gravel was deposited partly against or atop ice and is commonly marked by large ice-block depressions; some deposits are so severely collapsed that they can be described as kame fields or kame moraines.

Where best developed, an outwash plain region approaches that rarest of hydrologic conditions in the glaciated Northeast, an aquifer of infinite areal extent.

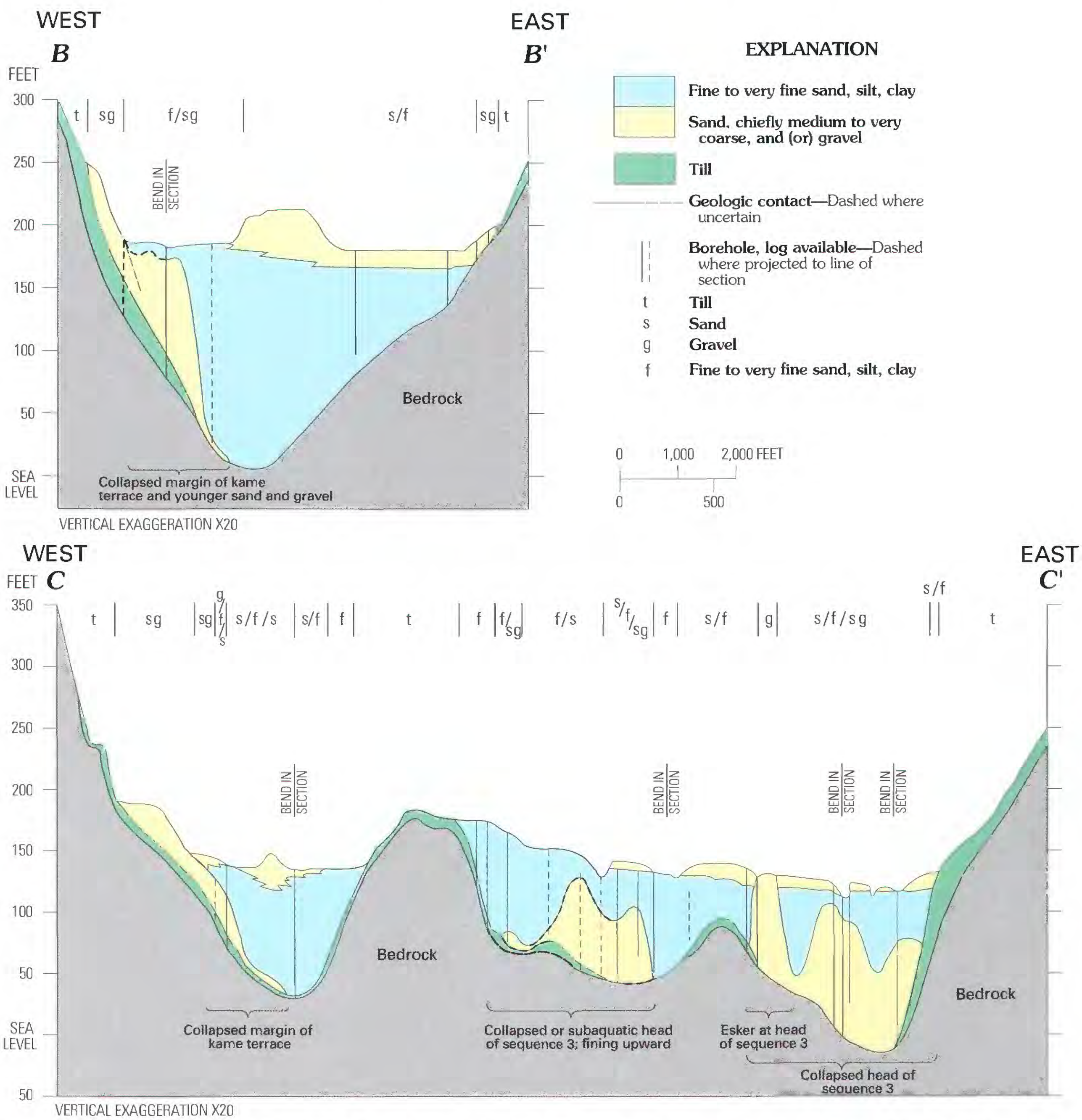


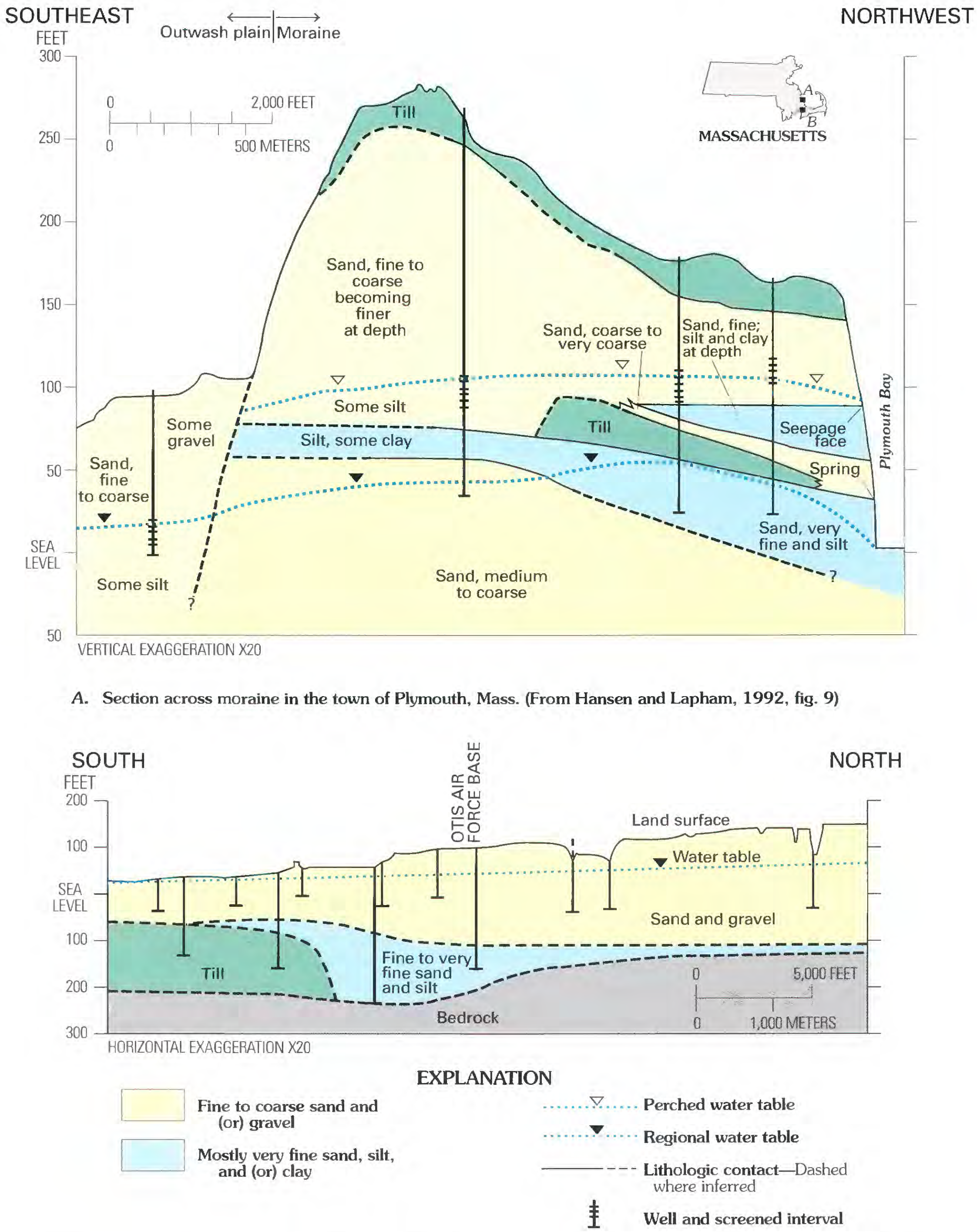
FIGURE 41.—Geologic sections across the Quinnipiac lowland in Southington, Conn., and adjacent towns. Traces of the sections shown in figure 39. Stacked symbols in bar across the top of each section indicate vertical distribution of lithology in segments of the section.

Permeable, water-yielding sand and gravel extends well below stream grade, and streamflow consists entirely of discharge from the aquifer, except for brief episodes of storm runoff from wetlands and from a few till-covered hills that protrude through the outwash. Locally, however, buried ridges of till or bedrock beneath the broad expanse of sand and gravel greatly reduce saturated thickness and transmissivity, such that the continuous outwash-plain aquifer grades into linear valley-fill aquifers typical of much of the glaciated Northeast. Locally, where the outwash overlies fine-grained lake or marine sediments, streams are incised into the fines, and the outwash-plain aquifer grades into a perched sand-plain aquifer (fig. 38*D*, *F*).

Southeastern New England.—A strip 10 to 15 miles wide along the coast of Massachusetts, from Kingston to Falmouth, is continuously mantled by sand and gravel that was deposited largely as outwash south of four ice margins identified by Larson (1982) in an interlobate area that was fed by meltwater from ice lobes in Narragansett Bay and Cape Cod Bay. The ice margins are marked by moraines that have highly irregular topography and are generally capped by sandy till and large boulders but are composed at least partly of stratified sand and gravel similar to that in the extensive outwash plains south of each moraine (fig. 42). The moraines also contain layers of dense clayey till, and well yields and water-table gradients indicate that transmissivity is lower within the moraines than in the outwash plains (LeBlanc and others, 1986). Mather and others (1942, p. 1153) described the outwash in the coastal strip as generally becoming siltier with depth and distance from the source. Hansen and Lapham (1992) also indicated that well logs and geophysical data generally show increased siltiness with depth, but interpret the outwash as consisting, in detail, of stacked packages that represent different depositional environments and were laid down during nonsynchronous pulses of ice retreat and readvance. Their ground-water flow model simulated hydraulic conductivity of the outwash as decreasing with depth but areally uniform. A model of the Mashpee outwash plain and vicinity by Guswa and LeBlanc (1985) simulated hydraulic conductivity as variable both laterally and vertically; moraines were described as areas of extreme lithologic variation over short distances, but transmissivity was simulated as between 8,000 and 30,000 feet squared per day at most locations, about the same as the adjacent outwash. Smith (1985, p. 124) reported that areally uniform, vertically repeated sedimentary packages have been recognized in some outwash plains worldwide, but the more common condition is "a bewildering array of vertical and lateral assemblages of varying sedimentary textures, structures, and bedding relationships."

Immediately west of the continuous coastal outwash in Massachusetts is the Narragansett basin, a broad lowland with less than 200 feet of relief underlain by sedimentary bedrock of Pennsylvanian age. Three ice-marginal zones are readily traced from northeast to southwest across the southern and north-central parts of that basin, through the hamlets of Middleboro, Bridgewater, and East Bridgewater, respectively (Hartshorn, 1976; Stone and Peper, 1982). Each is marked by a concentration of kames, ice-channel fillings, and collapsed stratified drift, interrupted by till-covered hills and by ice-block depressions, at and near the heads of several outwash plains and kame deltas (Hartshorn, 1960, 1967, 1976; Koteff, 1964). Large low-lying areas between the successive ice-marginal zones are underlain chiefly by fine-grained lake sediments but also, locally, by buried sand and gravel. The water-transmitting capacity of these stratified deposits is a function of depositional history and bedrock topography. Hartshorn (1960, p. D38; 1967) pointed out that outwash plains, kame deltas, and kame terraces associated with ice margins in the Narragansett basin consist largely of coarse gravel near the head or proximal end but grade southward to pebbly coarse sand overlying fine sand to silt. Williams (1968) described several localities in which a layer of saturated deltaic sand and gravel 20 to 120 feet thick overlies a finer grained layer, each layer becoming coarser and thicker northward. Williams and others (1973) and Lapham (1988) inferred that transmissivities greater than 1,300 feet squared per day are more common along the ice-marginal positions than elsewhere, but delineated zones of even higher transmissivity that seem to be associated with the large saturated thickness in bedrock valleys more than with the coarse lithology near ice margins or meltwater channels. Much of the coarse surficial stratified drift apparently overlies buried till hills and is thinly saturated.

Ohio.—Parts of Stark, Summit, and Portage Counties of northeastern Ohio are shown on surficial geology maps (Winslow and White, 1966; White, 1982, 1984) as mantled continuously by kame and kame-terrace deposits, except for younger outwash and alluvium along some streams and a few small areas of till. This region is largely part of the Kent and Bucks Hill moraines, described as a broad belt of sharp knolls and kettleholes arranged in random patterns. The morainal deposits consist largely of gravel or sandy gravel with subordinate till, deposited on or against stagnant ice in an interlobate area between the Killbuck and Grand River lobes (White, 1982, fig. 5). The bulk of the deposits are ascribed to the Titusville ice sheet, but the Kent ice sheet readvanced across much of the moraine, leaving a layer of till at or near land surface in many localities and depositing some additional stratified drift. The widespread ice-collapse topography, multiple drift sheets (White, 1984, fig. 9), localized till cover (Rau, 1969), and complex stratigraphy preclude geologic classification of these deposits as an outwash plain; but in this paper, they can reasonably be classified as an outwash-plain hydrophysiographic region because saturated sand and gravel constitutes a continuous aquifer, surficial or nearly surficial, over such a large area. Most ground-water appraisals of this part of Ohio (Smith, 1953; Winslow and White, 1966; Ohio Drilling Company, 1971; Schmidt, 1979; Walker, 1979a, b) present map units defined in terms of well or well-field



A. **Section across moraine in the town of Plymouth, Mass. (From Hansen and Lapham, 1992, fig. 9)**

B. **Section south of moraine in the towns of Sandwich and Falmouth, Mass. (From LeBlanc and others, 1986, sheet 1)**

FIGURE 42.—Geologic sections in outwash plains in southeastern Massachusetts.

yield. These appraisals do not discuss the surficial kame moraine deposits or their yield separately from deeper layers that may be present locally, but emphasize the distribution of "buried" valleys in which drift is 100 to 500 feet thick and generally yields several tens to hundreds of gallons per minute to wells. The surficial sand and gravel reaches its maximum saturated thickness within the buried valleys. It is probably thinly saturated where it crosses the intervening bedrock ridges and the upper flanks of the valleys, as documented in some localities (Smith, 1953, p. 36, p. 50) and suggested by sections (Rau, 1969, fig. 10) and by the apparent predominance of wells that tap bedrock (Smith, 1953; Walker, 1979a, b; Schmidt, 1979). Even if the stratified drift in this outwash-plain region was not continuously saturated, its abundance at shallow depths would result in greater ground-water recharge, storage, and yield than are available in till-mantled areas that border the region.

Immediately northwest of the outwash-plain region in Ohio, delineated on plate 1, surficial sand and gravel is extensive but largely restricted to valleys separated by till-mantled ridges (Winslow and White, 1966; White, 1984). White (1984, p. 15–16, p. 22) mentions that gravels correlative with those in southern Summit County underlie large areas in north-central Summit County, overlain by thick till and lacustrine silt; it seems unlikely that the buried gravels constitute a continuous aquifer. Both of these areas are included in the Western Appalachian Plateau region on plate 1.

New Jersey.—A broad outwash plain covers more than 25 square miles in and south of Plainfield, N.J., immediately south of the terminal moraine (Stanford and others, 1990). Drift thickness is generally 40 to 60 feet (Nemickas, 1974), and a few wells are screened in the outwash (Nemickas, 1976).

LIGHTLY DISSECTED UPLANDS WITH NEGLIGIBLE STRATIFIED DRIFT

Four large uplands are identified as individual hydrophysiographic regions on plate 1. Like many smaller uplands that border stratified-drift aquifers within other regions, each was affected only by locally derived meltwater that deposited little stratified drift. Ice dynamics varied from a residual ice cap to high plateaus bordered by lowlands occupied by active ice lobes, but the result was similar: a region nearly devoid of stratified-drift aquifers. All four regions are assigned the same color pattern on plate 1 to reflect their common lack of stratified drift.

TUG HILL PLATEAU

Tug Hill is a high but relatively undissected plateau in north-central New York. It rises steeply 1,000 feet above the Black River lowland to the north and east but slopes more gently southward and westward toward the Ontario lowland. Within the plateau, local relief is small. Streams drain radially outward from the center of the plateau and, thus, are rather small except near the plateau margin. Many streams are incised into bedrock for long distances as they approach the plateau margin (Wright, 1972; Miller, 1980a; Cadwell and Pair, 1991).

Little stratified drift is indicated on maps of soils or surficial geology of the plateau. Nearly all soils are derived from till; most are stony, acid soils with fragipans (Pearson and others, 1960). Large areas, particularly in Oswego and Jefferson Counties, are mapped by Cadwell and Pair (1991) as ablation moraine, deposited by downwasting; topography in these areas is a maze of small knolls and hollows superimposed on low parallel ridges. Street (1966) describes many scattered exposures that reveal a complex assortment of materials: loose unsorted debris with angular boulders, loose till grading down into very compact till, poorly sorted gravel and sand with faint stratification, and well-sorted glaciofluvial sediments. These descriptions suggest a variable combination of meltout till, stratified drift, and resedimented derivatives thereof; a few small areas identified as ice-contact kames by Jordan (1978) or Cadwell and Pair (1991) are topographically similar to their surroundings. Gravel on the floor of some incised channels in the southwestern part of the plateau has been mapped as outwash (Miller, 1980; Jordan and Miller, 1980; Cadwell and Pair, 1991) but could be considered Pleistocene or modern alluvium and is likely to be thin. A narrow, discontinuous band of thinly saturated sand and gravel extends for more than 25 miles along a bench on the east side of the plateau at an altitude of about 1,300 feet (Pearson and others, 1960; Jordan, 1978); these sediments formed in the same lake at the same time as massive deltas on the east side of the Black River lowland but are far smaller in volume because the watershed area was smaller and meltwater runoff was minimal.

In summary, the Tug Hill Plateau is an upland area of small local relief and radially outward drainage that was apparently deglaciated by downwasting without much throughflow of meltwater from an ice sheet to the north and without formation of proglacial lakes in which stratified sediments could accumulate. Meltwater throughflow was probably concentrated in trunk channels along lowlands east and west of the plateau. The result is a region devoid of significant stratified-drift aquifers.

POCONO PLATEAU

Another lightly dissected plateau is the Pocono Plateau, which approximately coincides with Pike and Monroe Counties of northeastern Pennsylvania (pl. 1). Local relief is generally 100 to 300 feet, except along the Delaware and Lehigh Rivers, which bound the region, and along the Pocono Mountain scarp in central Monroe County. As on Tug Hill, small streams drain radially outward to the bordering valleys or lowlands, and stratified drift is neither abundant nor concentrated in valley fills.

Ice-contact stratified drift is widely distributed across Pike County (Davis, 1989, pl. 2) but is found mostly as irregular hummocky masses on hillsides, north of low saddles, or on one side of shallow valleys, in positions that suggest it is largely above stream grade and, therefore, unsaturated. Outside the valleys of the Delaware River and its major tributary, the Lackawaxen River, surficial stratified drift and alluvium underlie only 5 percent of Pike County, and only 9 of 252 wells inventoried by Davis (1989) are finished above bedrock; two of these penetrate surficial ice-contact deposits or outwash, and seven presumably penetrate morainal or buried stratified drift in areas mapped as till. Large areas in northern Monroe County and vicinity are mantled by end moraine, described as ranging from sandy or very sandy till to ice-contact stratified drift with hummocky topography, draped continuously over hills and valleys (Sevon, 1975a). End moraine in southwestern Monroe County is described as till with minor sand and gravel (Crowl and Sevon, 1980). Hollowell and Koester (1975) mention morainal hills of sand and gravel in adjacent southernmost Lackawanna County. Northwestern Monroe County is a high plateau in which surficial stratified drift is virtually absent (Sevon, 1975a; Carswell and Lloyd, 1979); bouldery till mantles the hills, and a surficial carpet of closely spaced boulders blankets valleys or swales. East-central Monroe County is a lower plateau where drift thickness is locally appreciable, but only 5 of 112 wells inventoried by Carswell and Lloyd (1979) tap sand and gravel.

As on Tug Hill, small streams on the Pocono Plateau are incised into drift or bedrock where they approach the plateau margins and descend to major valleys or lowlands. These tributary valleys include V-shaped gorges but more commonly are floored by flood plains of gravelly alluvium a few hundred feet wide, some of which are interpreted as underlain by Wisconsinan outwash (Sevon, 1975a, b; Davis, 1989). Well records compiled by Carswell and Lloyd (1979) indicate that the lower slopes and floors of tributary valleys in east-central Monroe County are underlain by 50 to 150 feet of drift in several localities, but the well records provide little information as to lithology.

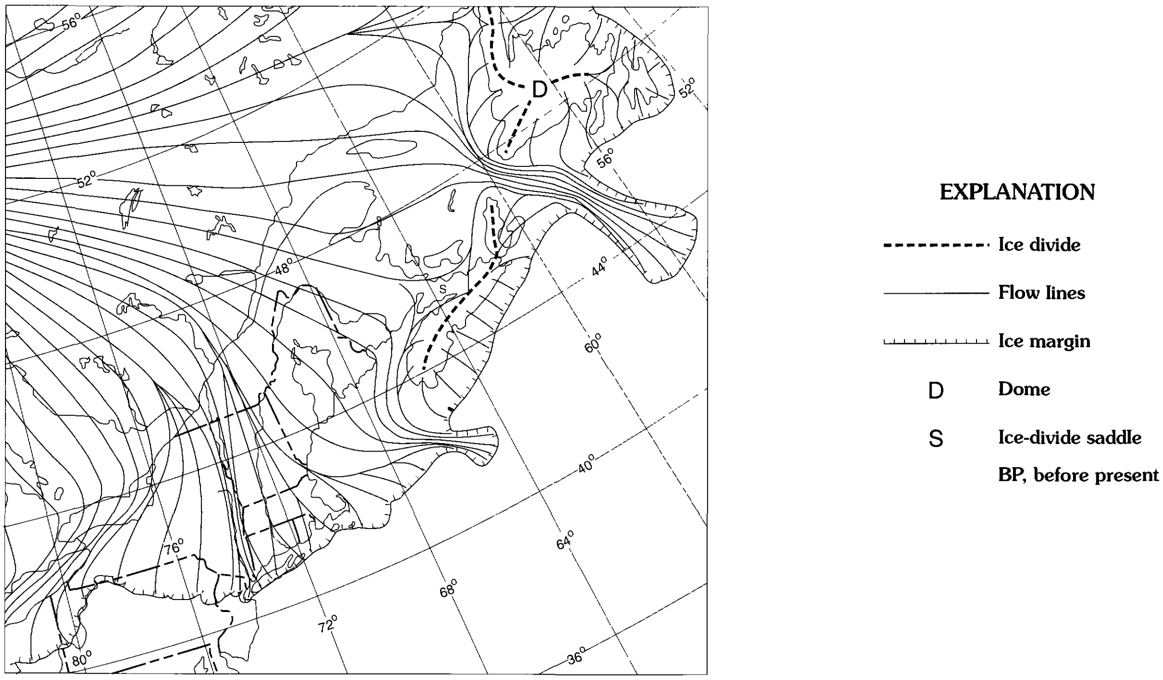
The Pocono Plateau region, as outlined on plate 1, is bounded on the northeast, southeast, and southwest by deeply incised valleys. Perhaps, as inferred for Tug Hill, meltwater from regions to the north was largely captured by these incised valleys during deglaciation. Also, the consistently high altitude and lack of large valleys in Pike and Monroe Counties hindered the development of proglacial lakes in which stratified drift could have been deposited. The area to the northwest, in Wayne County, has somewhat greater relief and is drained by south-flowing tributaries whose valleys apparently contain appreciable stratified drift; thus, it is tentatively assigned to the Eastern Hills and Valley Fills region.

NORTHEASTERNMOST ADIRONDACKS

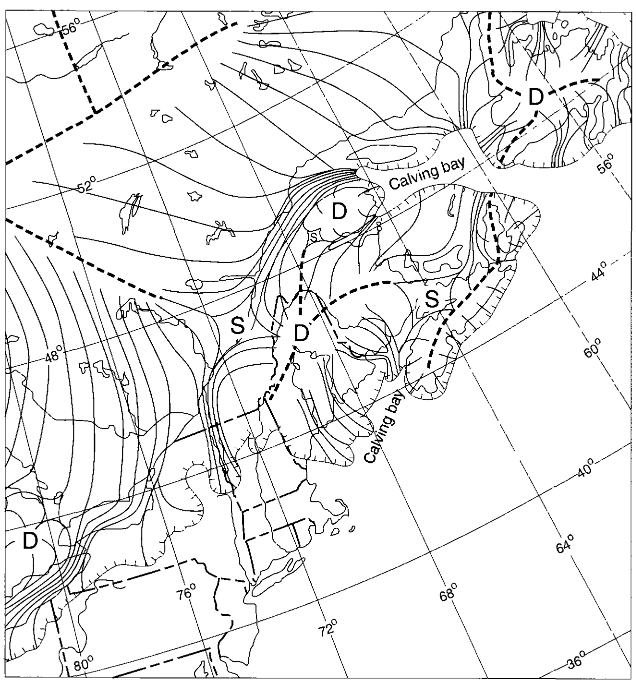
Another upland with small local relief and minimal stratified drift is in northern Clinton County, N.Y., at the northeasternmost corner of the Adirondack uplands (pl. 1). The region is underlain by the Potsdam sandstone of Cambrian age and includes several areas as large as 12 square miles in which the flat-lying bedrock is totally exposed, having been swept clean of glacial drift by the catastrophic drainage of Lake Iroquois (Denny, 1974). Tributaries of the Great Chazy River drain the region and are generally incised several feet into bedrock. Stratified drift occurs as part of recessional moraines and as small, scattered gravel knolls (Denny, 1970, 1974; Cadwell and Pair, 1991) that are outside the shallow valleys and above stream grade and are presumably unsaturated. As inferred for Tug Hill, subglacial meltwater flow during deglaciation was probably captured by the adjacent St. Lawrence and Champlain lowlands, so did not cross this intervening upland spur.

CENTER OF RESIDUAL ICE-CAP AREA IN MAINE

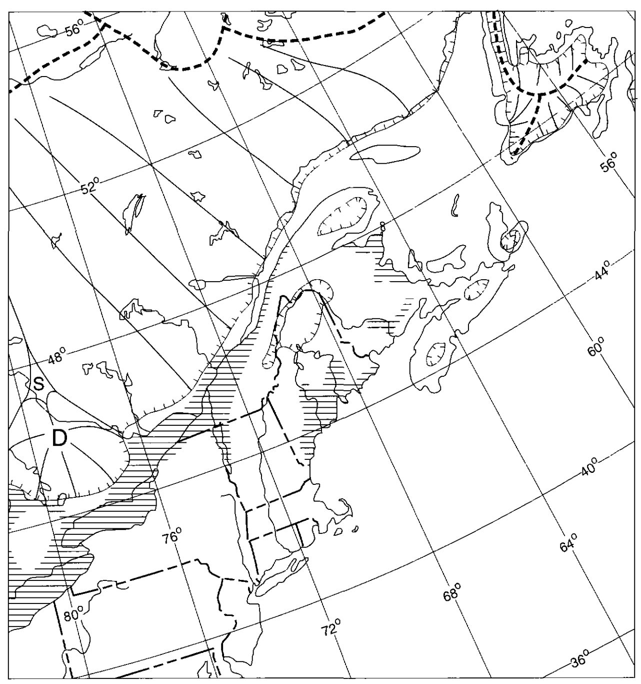
The last stages of deglaciation in New England and maritime Canada involved splitting the vast Laurentide ice sheet along the St. Lawrence valley, reorganization of ice south of the St. Lawrence to form an independent ice dome centered in northern Maine and Quebec's Gaspe peninsula, followed by radial shrinkage of that ice dome and, finally, melting of the last stagnant remnants in northern Maine and Gaspe. This process is illustrated in figure 43 and has been described in several papers, including LaSalle and others (1977), David and Lebuis (1985), Hughes and others (1985), Chauvin and others (1985), and Lowell and Kite (1986b, c). In response to the early phases of Wisconsinan deglaciation, sea level rose enough to allow ice in the lower St. Lawrence valley to dissipate directly into the ocean by calving, a process that is many times faster than retreat on land by melting. Therefore, the ice margin retreated rapidly up the St. Lawrence valley, and ice began to drain out of the bordering uplands into the valley and rapidly down the valley toward the ocean. Meanwhile, the rising sea level abutted the retreating ice margin in southern Maine and presumably caused rapid calving there too. Lowell and Kite (1986b, p. 81) speculated that rapid drawdown of the ice sheet in response to marginal



A. **18,000 years BP.—Ice flows southeastward over all of New England**



B. **14,000 years BP.—Downdraw into ice stream along St. Lawrence lowland has created an independent ice dome centered in northern Maine**



C. **12,000 years BP.—Small residual ice cap is all that remains of the ice sheet in New England. Horizontal lines show lakes and marine transgressions between 14,000 and 12,000 years BP**

FIGURE 43.—The flow regime of the eastern Laurentide ice sheet 18,000 to 12,000 years ago, according to a glaciological model. (From Hughes and others, 1985.)

calving might have quickly flattened its profile enough that the ice remaining over most of Maine could barely flow and dissipated largely by melting in place.

Once a saddle developed on the ice-sheet surface over the St. Lawrence valley (fig. 43*B*), not only ice but also meltwater from Canada no longer flowed southward into Maine because meltwater flow within as well as atop the ice was controlled by the ice-surface profile. Furthermore, areas in Maine north of the crest of the ice dome (fig. 43*B*) no longer contributed sediment or meltwater to the southern margin of the ice sheet. As the ice margin continued to retreat, smaller and smaller volumes of meltwater would have been available to entrain, sort, and transport sediment. One might expect, therefore, to find the volume of stratified drift becoming progressively smaller from southern coastal Maine northwestward to the center of the residual ice cap. Field observations suggest that such is indeed the case, as described in the following paragraphs.

An elongated oval region of roughly 4,500 square miles in northern Maine that contains few stratified-drift deposits and even fewer significant stratified-drift aquifers is delineated on plate 1. This region, where the last remnants of the ice sheet melted in place, is characterized by low hills and by scattered higher peaks that rise as much as 1,000 feet above the valleys. Drainage within this region is outward in all directions. Valley floors are generally flat, swampy, and underlain by organic-rich alluvium deposited in many generations of beaver ponds. Several valley reaches are inundated by large lakes, some of which were created or augmented by dams. The drift is thin in most places (Lowell, 1983; Thompson and Borns, 1985). Shallow exposures generally consist of loosely compacted till of sandy loam to loam texture that includes lenses and stringers of sand or silt; compact silty to sandy till is exposed instead in some places (Kite, 1983, p. 57). Several areas are mapped as stagnation moraine or undifferentiated drift (Thompson and Borns, 1985; Halter, 1985), each of which may include some stratified sand and gravel, but only a few scattered ice-contact or outwash deposits are shown on published maps. Lowell (1983, p. 16) asserted that very few valleys in this region contain "appreciable or systematic gravel deposits." Lowell and Kite (1986b, p. 76, 79, 80) reported that sand and gravel deposits are scarce in this region, much less abundant than in the fringe areas to the south and northwest, and are generally less than 16 feet thick. Kite (1983, p. 66) stated that isolated kames are found throughout the St. John watershed in northern Maine, but generally have little topographic expression and are difficult to discern; such deposits would be of little significance as aquifers. Fine-grained lake-bottom stratified drift is equally scarce; Kite (1983, p. 177–181) found no silt or clay beneath peat in several upland swamps that might have been ponded during deglaciation, and found at most 1 to 3 feet of silt along the shores of several modern lakes. Low, sandy deltas have been deposited where streams now enter these lakes, however (Kite, 1983, p. 181; Halter, 1985), and may constitute aquifers.

The boundaries of the region of negligible stratified drift (pl. 1) were placed where they would be approximately equidistant from the axis of the residual ice cap, as defined by striations (Lowell and Kite, 1986a), and would exclude most of the coarse stratified deposits mapped by Lowell (1980), Thompson and Borns (1985), and Lowell and Kite (1986b). Some coarse stratified drift might be present but as yet unrecognized, however, in this uninhabited area where roads and excavations are few and boreholes nonexistent. Coarse stratified drift has been found beneath till in excavations and boreholes in valleys east of this area, as described further on.

LOW-RELIEF FRINGES OF MAINE RESIDUAL ICE CAP WITH LITTLE COARSE STRATIFIED DRIFT

On the fringes of Maine's residual ice cap, stratified drift is apparently more abundant than near the center, but less abundant than in most areas of comparable relief farther south. Aquifer geometry in three fringe areas of low relief is distinctive, as described below.

NORTHWESTERN LAKE-DOMINATED FRINGE

Northwesternmost Maine drains southeastward toward the St. John River. As the residual ice cap in Maine retreated southeastward from the Canadian border, a series of proglacial lakes formed that initially drained northwestward across saddles in the Notre Dame Mountains, but later drained northeastward across various lower saddles to the St. Francis River (Kite, 1983; Lowell, 1985). Lake-bottom fines underlie the few broad, flat valley floors. Maps by Lowell (1980, 1985) and Thompson and Borns (1985) indicate several isolated deltas, eskers, kames, and kame terraces, most of them perched above stream grade on valley sides and probably thinly saturated. Individual deposits identified by Kite (1983, p. 160–1) do not all correspond to deposits shown on these maps, an inconsistency that reflects the difficulty of mapping in this undeveloped, wooded terrane. Gravel mapped as outwash covers more than 8 percent of the land surface in part of this subregion (Lowell, 1980), but most of these deposits mantle hillsides rather than fill valleys, are less than 10 feet thick (Lowell, 1980, p. 11), and are probably discontinuous. Areas of hummocky stagnation moraine could contain lenses of sand and gravel. In summary, although coarse stratified drift is more abundant in this northwestern fringe area than near the center of the former ice cap, aquifers are probably thin and widely separated.

NORTHEASTERN VALLEY-FILL FRINGE

The distribution of stratified drift in northeasternmost Maine is similar to that in the Eastern Hills and Valley Fills region; that is, stratified drift is found generally along the larger valleys but seldom on the intervening hills. The areal extent and volume of stratified-drift aquifers seem, however, to be smaller than in southern Maine and in regions to the south or west. For example, ice-contact deposits and outwash mapped along Munsungan Stream (Thompson and Borns, 1985) are perched on the valley sides (Tolman and Lanctot, 1981; Lowell, 1983, p. 25). Only discontinuous aquifers are recognized along the upper Aroostook River (Tolman and Lanctot, 1980), and several reaches of the lower Aroostook River and its tributaries are interpreted as bordered by till and bedrock (Prescott, 1972; Weddle and Neil, 1989). Only organic-rich alluvium is mapped along the flat, swampy floors of most valleys.

Aquifer delineation in this region is somewhat hindered by the presence of a diamicton layer within or, more commonly, atop the stratified drift, as observed in many pits and river-bluff exposures and reported in several borehole logs, although at least an equal number of comparable exposures reveal no such layer. Thus, the presence of till on the floor of a valley reach does not necessarily rule out the possibility of sand and gravel at greater depth. Exposures along the Aroostook River valley are described and evaluated by Borns and Borns (1986). Exposures along the St. John River valley are described and evaluated by Genes and Newman (1980), Kite (1983), and Lowell and Kite (1986c). Logs of several boreholes are presented by Prescott (1971a, 1973a), and logs of several exposures and boreholes are summarized by Tolman and Lanctot (1980), Weddle and Neil (1989), and Weddle and others (1989). Maps by Prescott (1972, 1973a) clearly indicate a few localities where productive sand and gravel aquifers are inferred to underlie surficial till, and others can be identified by comparison of surficial geology with aquifer maps by Weddle and others (1989). In this fringe area, near-surface diamicton is apparently a common but discontinuous and minor component of narrow valley fills composed primarily of coarse-grained stratified drift.

The widespread distribution of diamicton at shallow depth within or atop stratified drift in northeastern Maine is suggestive of a regionwide ice advance. Some diamicton exposures are described as compact lodgement-type till and have fabrics consistent with the directions of ice movement inferred from striations (Lowell and Kite, 1986c; Borns and Borns, 1986). Multiple drift sheets have been documented to the west in the lowlands of southern Quebec (Blais and Shilts, 1989; Shilts and Blais, 1989). If the numerous diamicton occurrences in northeastern Maine are the result of a single ice readvance, however, that readvance caused remarkably little erosion or deposition of stratified drift. Furthermore, much of the diamicton exposed at land surface in this region has been described as sandy or gravelly and noncompact (Kite and Borns, 1980, p. 214; Kite, 1983; Weddle and others, 1989) and might constitute ablation debris resedimented by debris flows from the ice or valley sides. Lowell and Kite (1986c, p. 29) interpreted layers of sand, sandy gravel, diamicton, and sandy diamicton at a site near Fort Kent as deposits of meltwater or debris flow during deglaciation.

EASTERN ESKER-DOMINATED FRINGE

East-central Maine is a region of low relief where stratified drift was not deposited in proglacial lakes or marine waters as in most of the glaciated Northeast; hence, fine-grained stratified drift is virtually absent. Coarse-grained stratified deposits do not fill the bedrock valleys and were not built against valley walls. Deposition occurred in tunnels or channels within the ice sheet and formed narrow esker ridges or, less commonly, more complex constructional landforms along the esker alignments. The gentle ridges and valleys in this region trend approximately south-southeast, discordant to the bedrock structure but parallel to the eskers and to striations that record the direction of ice flow (Thompson and Borns, 1985). Many esker segments follow shallow, linear valleys of modern streams, generally slightly to one side of the valley axis; some segments cross low hills or saddles where they are probably thinly saturated, while others bisect large swamps.

An example of aquifer geometry, typical except that two eskers are unusually close to each other, is illustrated in figure 44. The eastern esker parallels South Branch Meduxnekeag River but only locally abuts the river. Some borehole records near where Meduxnekeag River crosses the esker and is joined by South Branch, and farther north near Houlton, indicate that esker sand and gravel extends 40 to 60 feet below the altitude of land surface adjacent to the esker (Prescott, 1971b; Tolman and Lanctot, 1981). This geometry seems to require that a meltwater channel was incised into the subglacial till before esker deposition, and (or) that the esker is bordered by younger sediment—such as till

EXPLANATION

Ice-contact sand and gravel—Including eskers (Prescott, 1973b)

Aquifer—Well yields exceed 10 gallons per minute (Tolman and Lanctot, 1981)

Aquifer—Well yields exceed 10 gallons per minute (Prescott, 1973b)

Municipal well—Yield exceeds 700 gallons per minute

FIGURE 44.—Typical esker-dominated terrane in east-central Maine. (Modified from Prescott, 1973b, and Tolman and Lanctot, 1981.)

melted out of stagnant ice, or thin outwash of local origin, or postglacial alluvium. A few other esker segments in which sand and gravel apparently extends below the altitude of land surface alongside the esker are evident from borehole records in other parts of the eastern esker-dominated fringe, as shown on a series of maps similar to the map by Tolman and Lanctot (1981). Worldwide, geological literature is replete with reports of tunnel valleys, which are channels incised in drift or bedrock beneath ice and commonly associated with eskers; generally, however, the eskers link tunnel valleys (Sugden and John, 1977, p. 305) or lie within or beside tunnel valleys (Wright and Ruhe, 1965, p. 35; Barnett, 1990) rather than entirely filling such valleys. If the sand and gravel in some eskers of east-central Maine extends deeper than the surface of the adjacent till plains or river flood plains, appreciable saturated thickness is likely. The distribution of saturated thickness is subject to interpretation, however. For example, the entire width of each esker landform in figure 44 was inferred to constitute an aquifer by Tolman and Lanctot (1981), whereas the west side of each esker was generally excluded from the aquifer by Prescott (1973b), presumably because he inferred the west side of the esker to be banked against the side of the bedrock valley and, therefore, barely below the water table. Both interpretations consider the aquifer to extend beneath some areas of flood plain, swamp, or till on the east side of the esker.

In this region, almost no stratified drift has been mapped other than eskers and related ice-contact deposits aligned with eskers. Most small stream valleys have flat, swampy floors, presumably underlain by organic-rich alluvium deposited partly in beaver ponds. Large swamps and lakes occupy natural depressions in many places. Borehole records do not indicate the presence of sand and gravel beneath any of these valleys or depressions, except alongside the esker alignments. In figure 44, the east-west reach of the Meduxnekeag River winds across a low plain that was mapped as till or swamp by Prescott (1973b) and Thompson and Borns (1985) and mapped as lacking sand and gravel aquifers by Tolman and Lanctot (1981).

South of the region of minimal stratified drift at the center of Maine's former ice cap is a generally mountainous area that extends from Mt. Katahdin west to the Canadian border. Here, too, coarse stratified drift consists chiefly of eskers and underlies only a small percentage of watershed area, but because relief is much higher (Denny, 1982), esker alignments are confined within valleys, and the combination of eskers with other forms of stratified drift and alluvium results in aquifers that can be described as valley fills. All these characteristics are consistent with the Eastern Mountains region as defined in this report; therefore, the southern fringe of the Maine ice-cap area is classified as part of the Eastern Mountains region on plate 1.

REGIONS WITH HIGH LOCAL RELIEF

Four of the hydrophysiographic regions delineated on plate 1 are deeply dissected mountains or plateaus in which the major valleys sloped away from the ice sheet. Stratified-drift-filled valleys are more widely spaced than in the regions described previously and are generally bordered by steep valley walls. Upland runoff is the principal source of natural recharge to surficial aquifers (Morrissey and others, 1988).

EASTERN MOUNTAINS

Relief in this region is variable but consistently greater than 1,000 feet, as measured by Denny (1982). The extent of this region in New England (pl. 1) is based partly on the relief map by Denny (1982, fig. 10) and partly on geologic and topographic maps. Excluded from this region are the Vermont valley from Rutland to Bennington and several valleys in western Massachusetts, all cut in carbonate bedrock, because they are wider and contain more stratified drift than other valleys in the region.

The distribution of stratified drift in the valleys of the Eastern Mountains region can be characterized as follows:

1. The percentage of the land surface mantled by surficial sand and gravel is commonly small—only about 9 percent of the land surface in high-relief watersheds in New England is mantled by surficial sand and gravel, much less than the 20 to 26 percent in watersheds of lower relief to the south (table 4). The difference probably results from the major valleys being narrower and more widely spaced in the Eastern Mountains region than in adjacent regions of lower relief. Large streams in this region typically have valley floors less than 2,500 feet wide.

2. A substantial part of the surficial sand and gravel in this region consists of alluvium—flood plains and alluvial fans (table 4). Although these alluvial deposits are highly permeable in most places, they do not necessarily overlie substantial thicknesses of coarse stratified drift.

3. Stratified drift commonly extends to depths of 50 to 150 feet below the valley floors (for example, see Cotton and Olimpio, 1996; Willey and Butterfield, 1983). Bedrock crops out in many places in stream channels and as rounded knobs on flood plains,

TABLE 4.—*Areal extent of surficial sand and gravel in watersheds representative of several hydrophysiographic regions*
[Each data set was originally compiled for analysis of the effect of watershed characteristics on low flow of streams. Area of surficial sand and gravel includes area of any swamps or lakes inferred to overlie sand and gravel. All watersheds from which this table was compiled are between 10 and 200 square miles. Percentages were computed by summing areas for all watersheds, then dividing one sum by the other as indicated in column heads. Dash indicates area of alluvium not compiled; >, greater than; <, less than]

Locality and report reference	Hydrophysiographic region(s) as shown on plate 1	Relief	Number of water-sheds	Mean drainage area (square miles)	Area of surficial sand and gravel divided by drainage area (percent)	Area of alluvium divided by area of surficial sand and gravel (percent)
Most of Vermont, New Hampshire,western Massachusetts (Wandle and Randall, 1994)	Eastern Mountains; Mountains bordering Champlain lowland	High	15	60.5	8.6	35
Rhode Island, eastern Massachusetts, southern New Hampshire, and Maine (Wandle and Randall, 1994)	Eastern Hills and Valley Fills	Low	13	56.2	26	10
Connecticut (Cervione and others, 1982)	do.	Low[1]	14	53.8	20.5	—
Delaware River basin, N.Y. (Coates, 1971)[2]	Catskill Mountains	High	12	62.5	6.3	—
Susquehanna River basin, N.Y. and Pa. (Coates, 1971)[2]	Eastern Appalachian Plateau	High	10	96.2	14.2	—
Susquehanna River basin, N.Y. (Ku and others, 1975)	do.	High	25	79.3	14.3	—
Hudson River basin, south of Troy, N.Y. (Barnes, 1986)	Eastern Hills and Valley Fills	Low[1]	32	36.4	>13.7 <15.3	—

[1]A few watersheds in adjacent areas of high relief were deleted from data set.
[2]Coates measured "valley fill area," which is equivalent to area of sand and gravel in valleys.

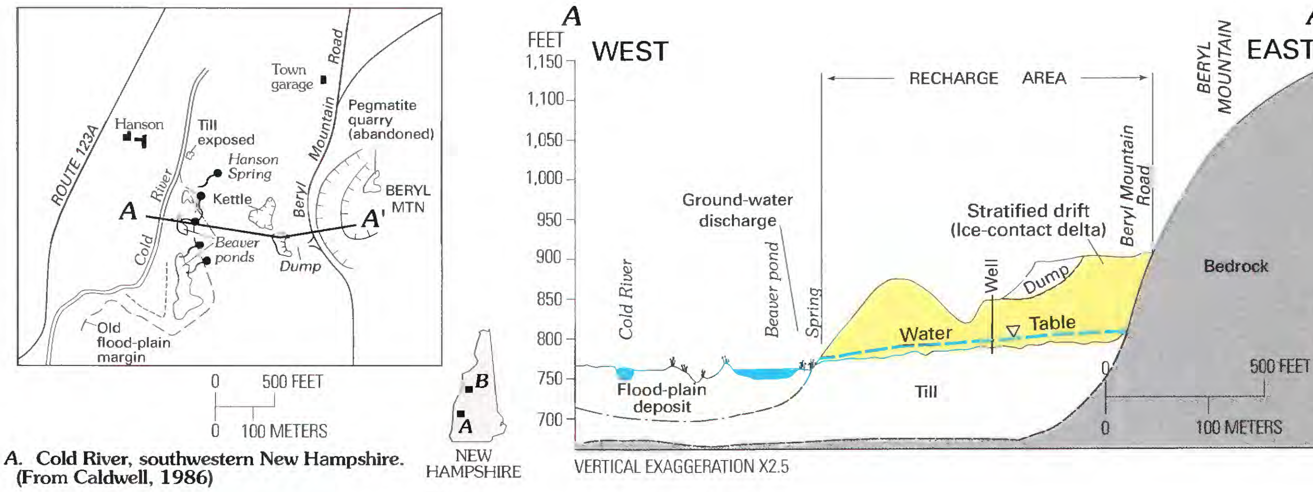
however, and short reaches of many streams are incised in bedrock gorges. These outcrops, along with drillhole records, indicate that deep stratified-drift aquifers generally underlie only parts of the narrow valley floors.

4. A substantial part of the stratified drift in this region is draped over upland ridges or banked high on valley walls entirely above stream grade and, hence, is very thinly saturated. Close examination of some such deposits has revealed exposures of till or bedrock and (or) springs at their downhill margin where postglacial streams have incised entirely through them (fig. 45). More commonly, the base of the stratified deposit is not obviously exposed, but projection of adjacent till-mantled slopes suggests that the base is above stream grade.

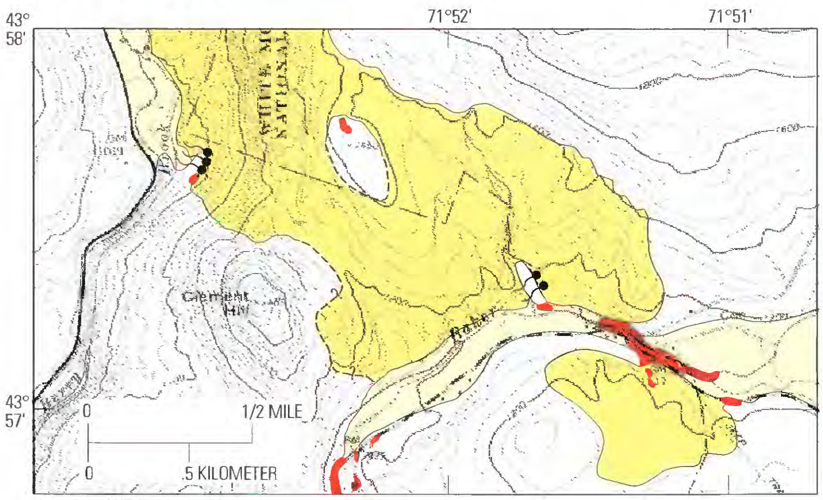
5. Large-scale ground-water development would be sustained chiefly by recharge from upland runoff (Morrissey and others, 1988) and induced recharge from the master stream. Precipitation on the aquifer and storage within it would be of less importance because stratified-drift aquifers are so narrow.

CATSKILL MOUNTAINS

The Catskill Mountains region is one of high relief, steep valley sides, and relatively narrow, flat valley floors underlain by stratified drift. Relief in gaged watersheds generally exceeds 1,300 feet and averages 1,930 feet (Coates, 1971, table 3). Valley floors along the largest streams (Delaware River and its East and West Branches) are typically 800 to 2,000 feet wide and rarely more than 2,500 feet wide. On the average, stratified drift underlies only about 6 percent of the watershed area (table 4). All or most of the valley floor generally consists of flood plain, tributary fans, and low terraces that could be outwash or early postglacial river alluvium; the small extent of ice-contact stratified-drift landforms and the inundation of several valley reaches by reservoirs may hinder use of the morphosequence concept as a guide in interpreting aquifer geometry and grain size. Kirkland (1973) divided stratified drift in the valley of West Branch into 17 morphosequences, most of which he described as grading downvalley from kame terraces to kame moraine to outwash; however, he presented little evidence of downvalley decrease in grain size within individual morphosequences, or correspondence of terrace elevations to his theoretical profiles.



A. **Cold River, southwestern New Hampshire. (From Caldwell, 1986)**

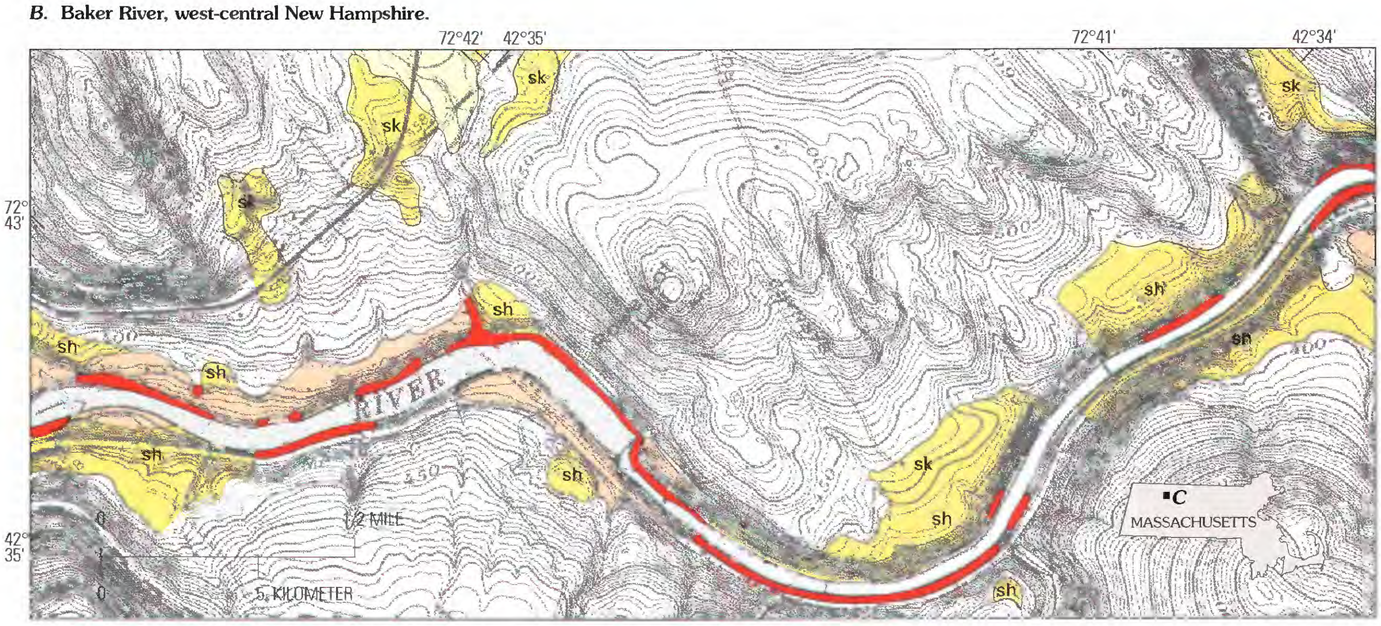


Base from U.S. Geological Survey
Warren and Mt. Kineo 7.5-minute quadrangles

B. **Baker River, west-central New Hampshire.**

EXPLANATION FOR MAPS

- **Alluvium**—Chiefly gravel
- **Till, with small bedrock outcrops**
- **Stratified drift**—Sand and gravel (sk, graded to early small lakes; sh, chiefly sand, graded to large lake in main valley)
- **Postglacial stream-terrace gravel**—Generally overlying sand (unit sh)
- **Bedrock outcrop**
- **Spring**
- **Geologic contact**
- *A*——*A'* **Line of geologic section**



Base from U.S. Geological Survey Shelburne Falls 7.5-minute quadrangle

C. **Deerfield River, northwestern Massachusetts. (From Segerstrom, 1959)**

FIGURE 45.—Examples of stratified drift perched on valley sides in three areas of high relief. Springs, till exposures, and bedrock outcrops at the base of the valley sides and (in *A*) a test-well log all indicate that stratified drift upslope rests on till and bedrock well above stream grade.

The Catskill Mountains region differs from the Appalachian Plateau to the west in that valley fills are not as wide or deep and include less fine-grained sediment. It differs from the Eastern Mountains region in that ice-contact deposits perched above stream grade on the valley sides are much less common or so narrow as to lack topographic expression, and bedrock seldom crops out in flood plains or stream channels except where the stream abuts the valley wall.

Saturated thickness of valley fills seldom exceeds 150 feet. Soren (1961, 1963) tabulated records of many wells in this region, including 105 drilled wells along the valleys of the largest streams—Delaware River and its East and West Branches and the lower reaches of Beaver Kill. Depth to bedrock below river grade, which presumably nearly equals saturated thickness of the drift, was reported to exceed 100 feet and 150 feet at only 13 and 4 well sites, respectively. The areal extent of valley fill that exceeds 100 or 150 feet in saturated thickness is undoubtedly somewhat greater than indicated by this set of well records, however, inasmuch as a majority of the wells are along roads near the base of the steep valley walls, and 28 of them were completed above bedrock.

Lithology of valley fills at several locations has been thoroughly described by geologic sections or borehole logs for dams, bridges, and tunnels constructed or proposed by the New York City Board of Water Supply, as presented by Fluhr (1953), Fluhr and Terenzio (1973, 1984), Soren (1963), and (or) in memorandum reports by T.W. Fluhr on file at the New York State Geological Survey. Maximum thickness of drift below river grade at seven sites along East Branch and West Branch ranged from 35 to 120 feet and averaged 90 feet. The sediments penetrated were mostly sand and gravel, generally coarse and variably silty, but also included much till. Differing terminology and detail in the original borehole logs complicate interpretation, but clearly fine-grained lake-bottom sediments are only a minor component of the stratified drift. An open trench 50 feet deep and about 1,300 feet long was dug across the East Branch valley at Downsville during construction of a dam, and pumpage of 4,000 gallons per minute was required to keep the water table below the trench (Fluhr, 1953; Fluhr and Terenzio, 1984), indicating that the valley fill is quite permeable at that site. By contrast, logs of boreholes at a bridge over Trout Creek (Fluhr and Terenzio, 1973, p. 300) and a dam site on Neversink River (Fluhr and Terenzio, 1984, p. 111) reveal all three depositional facies: alluvial sand and gravel (facies 3) over lacustrine fine sand and silt (facies 2) over heterogeneous ice-contact deposits (facies 1).

At one proposed dam site (fig. 46) and seven other locations in the Catskill Mountains region (Soren, 1961, 1963), water wells or test borings are reported to have penetrated water-yielding coarse sand and gravel beneath 80 to 200 feet of mostly till. All these sites are on, or at the base of, moderately gentle hillsides that slope southward or southwestward into major valleys, except for one hillside that slopes southwestward along the valley from a bedrock spur on the valley side. During the last glaciation, the ice must have selectively deposited large volumes of till on the south-facing or lee sides of some hills or valley walls; as the till accumulated, it encroached upon the valley floors and buried some sand and gravel that had been deposited ahead of the advancing ice or during the previous deglaciation. The valleys of Beaver Kill, Callicoon Creek, and the headwaters of East Branch and Neversink River, all in the eastern part of the Catskill Mountains region, include many narrow reaches that seem to have resulted from thick, localized accumulations of till that extend into or across the valley. Rich (1935, p. 29, pl. 1) described and mapped thick drift, composed predominantly of till, as universal in tributary valleys in the Catskills and common along the larger valleys; he noted that drift can be so thick in the lee of spurs on valley walls that the mouths of tributaries have been blocked and shifted downvalley. Drift as thick as 360 feet was penetrated by boreholes at three proposed dam sites in narrow reaches of Beaver Kill and Willowemoc Creek valleys; the drift extended 160 to 190 feet below stream grade and was interpreted by Fluhr (unpublished memorandum reports from 1948–49, on file at the New York State Geological Survey) and by Fluhr and Terenzio (1973) as predominantly till, with only a few scattered lenses of sand, gravel, or lake beds. Boreholes in intervening wider reaches of these valleys reached bedrock 210 to 240 feet below the valley floors; much of the drift penetrated was apparently stratified. Coates (1966b) showed that "till shadows" (accumulations) on the southerly or lee sides of hills are common in uplands in the Appalachian Plateau to the west, and later illustrated (Coates and King, 1973; Coates, 1981) how till accumulation on the south side of hills has shifted the axes of some small upland valleys (fig. 47). Similar relations probably prevail in the uplands of the Catskill Mountains region. The thick till shadows that locally protrude into valleys in the Catskill Mountains region and confine buried aquifers can be thought of as products of upland depositional processes, related not to deglaciation of the valleys but to glaciation of the adjacent uplands. Aquifers buried beneath the till shadows are presumably of small extent and may or may not intersect the principal sand and gravel aquifers deposited during retreat of the last ice. Major valleys within the Appalachian Plateau to the

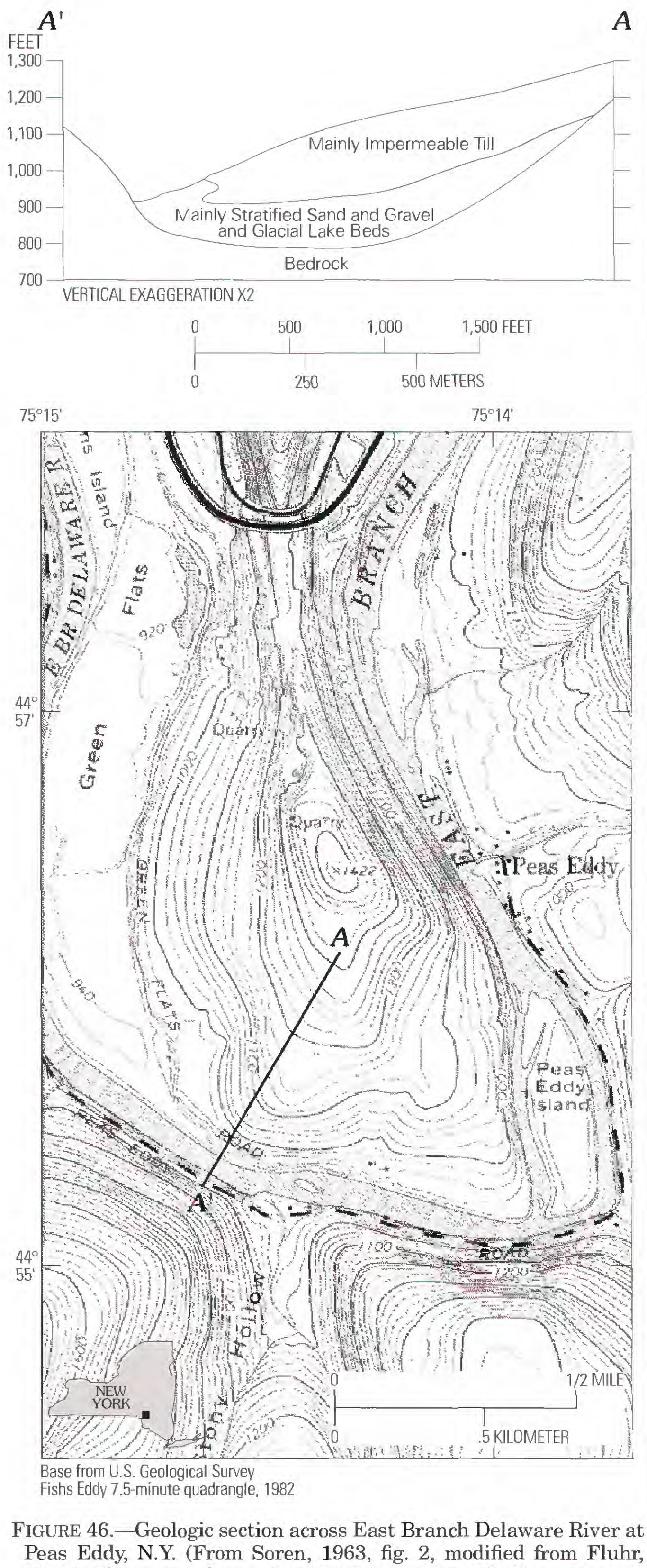


FIGURE 46.—Geologic section across East Branch Delaware River at Peas Eddy, N.Y. (From Soren, 1963, fig. 2, modified from Fluhr, 1953.) The upper layer, described by Fluhr (1953) as "mainly impermeable till," was interpreted by Kirkland (1973) as a mixture of till and colluvium (resedimented till).

west are wider than those in the Catskill Mountains region and are not locally narrowed by till shadows; probably they were more generally followed by ice streams during glaciation.

Two distinctive aspects of the hydrology of the Catskill Mountains region are not related to the geometry of stratified-drift aquifers, but result in large rates of recharge to those aquifers. The first is large precipitation, averaging 42 to 60 inches per year (Knox and Nordenson, 1955; Hely and Nordenson, 1961; Dethier, 1966; Randall, 1996), which causes abundant upland runoff. The second is bedrock lithology; the competent, well-indurated but unmetamorphosed nonmarine siltstone and sandstone that predominate in this region (Soren, 1961, p. 15) allow more open fractures and, hence, greater lateral ground-water flux into the valley fill than is likely from the finer grained marine shale and siltstone that predominate in the Appalachian Plateau to the west and in much of the Hudson lowland to the east.

EASTERN APPALACHIAN PLATEAU

The Eastern Appalachian Plateau is an extensive upland with summit elevations of 1,500 to 2,200 feet, whose major valleys have been widened and deepened by glacial erosion; relief in gaged watersheds generally exceeds 900 feet and averages 1,075 feet (Coates, 1971, table 3). On the average, 14 percent of the region is mantled by surficial sand and gravel, nearly all of it in the major valleys (table 4). The intervening uplands generally lack surficial sand and gravel; they are drained by tributary valleys up to 8 miles long whose floors are only a few hundred feet wide and underlain by several feet of alluvium over till. Most major valleys are between 2,000 and 8,000 feet wide, although a few are as narrow as 1,000 feet. No bedrock outcrops are found on the valley floors; stratified drift generally extends 70 to 200 feet below stream grade west of Binghamton, and 250 to 500 feet below stream grade northeast of Binghamton. Many major valleys near the northern perimeter of the Appalachian Plateau cross modern drainage divides, where former saddles were enlarged and deepened by glacial erosion into valleys that conveyed large volumes of meltwater and sediment southward. These valleys, which are termed "through valleys" (von Engeln, 1961; Coates, 1966a; Randall and others, 1988), contain thick stratified drift beneath the modern divide. Other major valleys head in an array of steep upland tributaries and are commonly narrower than the through valleys, particularly in northern Pennsylvania, where surficial gravel in valleys tributary to the Susquehanna River seldom exceeds 2,000 feet in width.

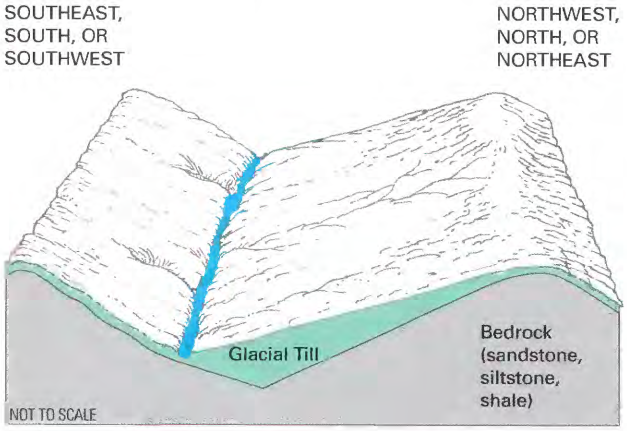


FIGURE 47.—Idealized upland valley displaced by till accumulation in the shadow of an adjacent hill. (From Coates and King, 1973, fig. 7.)

Till, commonly several tens of feet thick, mantles bedrock along the sides of the major valleys, but is a rare and minor component of the drift in those valleys, except immediately atop bedrock. Examination of undisturbed samples from many drillholes along the major valleys (Randall, 1972; Martin, 1981; U.S. Geological Survey, unpublished records) disclosed many layers of poorly permeable silt-bound sand or gravel, but very few layers with the texture and lack of sorting typical of till. Locally, however, till or resedimented diamicton mantle stratified drift. The lower side slopes of many valleys are gentler than the upper slopes and display smoothly rounded lobes or noses—all characteristics that suggest thick drift. The drift is generally till, as indicated by exposures and soils maps. In some localities, however, soils maps depict lower side slopes as hybrid soils (such as Bath-Howard) in which the parent material of one soil is till and that of the other is gravel, and excavations reveal stratified sand and gravel mantled by diamicton. Perhaps the surficial diamicton and the rounded topography result from ice readvance across kames or kame terraces. Diamicton layers that are present at or near land surface on the valley floor in several localities along the Valley Heads moraine (Randall, 1986b; Randall and others, 1988) and at the Wisconsinan drift border (Zarriello and Reynolds, 1987) have been attributed to a readvance. Alternatively, perhaps the surficial diamicton results from solifluction. Denny and Lyford (1963, p. 15, p. 40) described several sites where colluvium derived from bedrock or till had migrated down upland hillsides onto stratified drift; Denny (1956a, p. 83) reported the phenomenon to be widespread in this region. King and Coates (1973) pointed out that hillside profiles in some upland localities change from concave to convex with decreasing elevation and attributed this configuration to downslope movement of large volumes of till as solifluction lobes.

The universal three depositional facies are widely distributed along the major valleys of the Eastern Appalachian Plateau region, as stated in general terms by Fleisher (1986b, p. 132) and illustrated in many reports, including several examples presented on the next few pages. Valley fills in this region are generally broader and thicker than those in other regions that drained away from the ice, so they might be expected to offer large well and aquifer yields. The thickness of coarse-grained stratified drift does not generally increase in proportion to drift thickness, however (MacNish and Randall, 1982, p. 7). Instead, greater depth to bedrock is generally associated with greater thickness of fine-grained (facies-2) sediments. Coarse-grained stratified drift seems to be fairly widespread below as well as above the fine sediment, as discussed in an earlier section and evidenced by examples farther on, but because few boreholes reach bedrock where drift thickness is great, the distribution and thickness of deep stratified-drift aquifers are not well established in much of the region. The large drawdown available to deep aquifers favors large well yields, but data compiled by Hollyday (1969) indicate that specific capacities are greatest in surficial aquifers, are half as great in aquifers 50 to 200 feet below the water table, and are less than one-third as great in aquifers deeper than 200 feet below the water table.

The extent to which buried ice-contact deposits are concentrated along former meltwater channels within the valleys is likewise not well established. Abundant borehole data along the Chenango River, Susquehanna River, and Nanticoke Creek valleys in southwestern Broome County, N.Y., indicate that ice-contact deposits several tens of feet thick form narrow bands that approximately follow the thalweg of the bedrock valley and are bordered by areas in which ice-contact deposits are thin or absent beneath thick fines. This distribution is reflected in maps by Holecek and others (1982), MacNish and Randall (1982), and Randall (1986a) and suggests preferential deposition within or at the mouth of a persistent channel in the decaying ice. Eskers a mile or more in length are known from tributary valleys near Brisben (Cadwell, 1972) and Lisle, N.Y., and from east-west reaches of major valleys near Oakland, Pa. (Harrison, 1966), and Chemung, N.Y.; bedrock is probably less than 100 feet below stream grade at all four localities. As explained earlier in the section on meltwater-drainage systems, Fleisher (1991a, b) argued that the bulk of the fine-grained sediment in the major valleys of this region must have been contributed by subglacial meltwater conduits along those valleys, having been eroded from basal ice or subglacial sediment. If so, significant amounts of sand and gravel also would have been eroded and must have been deposited within, or at the mouths of, the subglacial conduits, especially in through valleys.

Aquifer geometry in the Eastern Appalachian Plateau can, to some extent, be inferred from landforms, which reflect depositional conditions. Several investigators have identified prominent ice-margin positions within the major valleys of this region where massive, hummocky deposits of coarse-grained, heterogeneous facies-1 stratified drift fill the valley from side to side. These deposits constitute single or multiple-abutting morphosequences, range in length from about 1,000 feet to about 2 miles, and have been termed valley-choker moraines (MacClintock and Apfel, 1944; Frimpter, 1974), valley plugs (Cadwell, 1972), ablation moraines (Fleisher, 1977; Morrow, 1989), recessional moraines (MacNish and Randall, 1982), and kame moraines (Fleisher, 1986b). Coarse-grained ice-contact deposits extend to substantial depth, at least at the heads of these features, and coarse-grained outwash (facies 3) generally progrades downvalley across fine-grained lake sediments (MacNish and Randall, 1982; Fleisher, 1986b). Although deposition was predominantly deltaic, sediment aggraded somewhat above lake level; thus, these moraines served as dams (actually, large riffles) that collectively and incrementally impounded proglacial lakes in front of the retreating ice. Some large deltas, perhaps in part supported by buried large ice blocks, also might have functioned as temporary incremental dams in other valley reaches. For example, a feature identified by Fleisher (1977, p. 39) as a delta terrace or kame delta, not a moraine, occupied the entire width of Otego Creek valley 2 miles south of the hamlet of Mt. Vision, prior to incision by Otego Creek.

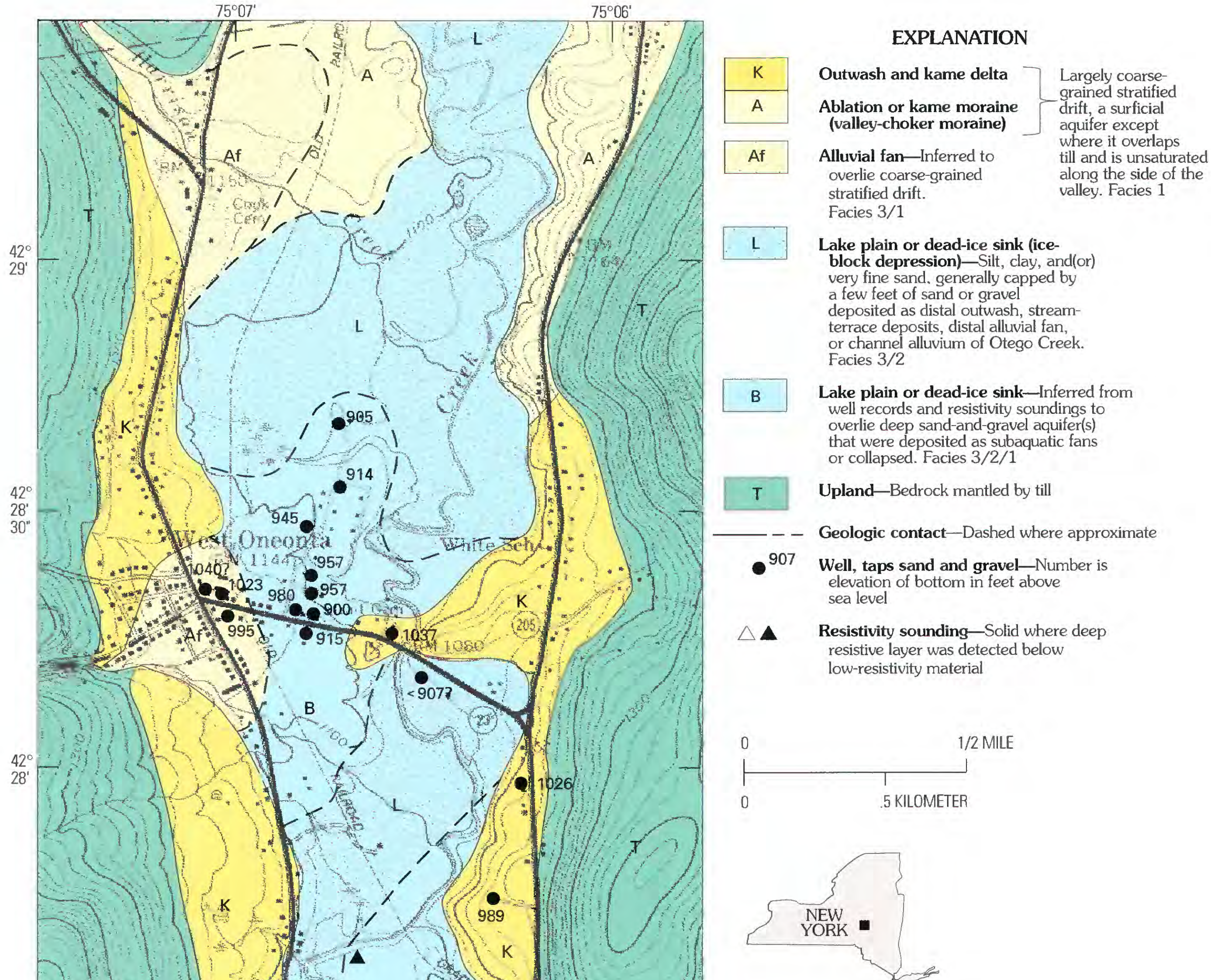
Many valley reaches between ice-marginal positions are discontinuously bordered by terraces that stand several tens of feet above the valley floor and could be kame terraces (built between stagnant ice and the valley wall), delta terraces (built at the end of a retreating ice tongue by a stream that flowed alongside the ice tongue; MacNish and Randall, 1982, fig. 7; Fleisher, 1986b, fig. 4), or tributary deltas (inwash at the mouths of tributaries). These features, which are predominantly coarse grained and are commonly found where bedrock is relatively shallow, have been interpreted as successive deposits at the terminus of an ice tongue that was retreating too rapidly for proximal coarse deposits to build across the valley to form a moraine. Broad, poorly drained flood plains and low river terraces that occupy much of the valley floor between some prominent ice-margin positions have been interpreted by Fleisher (1977, 1986a, b) as lake plains (underlain chiefly by thick silt, capped by thin sand and gravel) and dead-ice sinks (large ice-block depressions, underlain by a heterogeneous array of collapsed outwash, lake-bottom deposits, and postglacial alluvium).

Many valley reaches in the Eastern Appalachian Plateau contrast sharply with those described above in that all or most of the valley floor consists of low planar surfaces, commonly less than 20 feet above stream grade. These reaches are especially common near the Valley Heads moraine but also are found elsewhere; long reaches of the Canisteo, Cohocton, Tioga, and Tioughnioga Rivers, and Tuscarora, Catatonk, and Cayuta Creeks can be thus described. A surficial sand and gravel aquifer is present throughout these reaches and probably originated as outwash, locally reworked by postglacial streams and locally augmented by tributary fans. Commonly, the surficial outwash overlies lacustrine fines, but truncated deltas, ice-contact deposits, and ice-block depressions can complicate the stratigraphy, and the uniform surfaces offer few clues as to subsurface conditions. One possible clue is river slope, which tends to be steeper where the channel overlies gravel or till than where it overlies sandy outwash or lake sediment (Randall, 1986, p. 16).

Geologic maps and sections from five localities were selected to illustrate aquifer geometry of the Eastern Appalachian Plateau.

1. *Typical South-Draining Valley.*—The reach of Otego Creek valley depicted in figure 48 contains two ice-marginal positions identified by Fleisher (1977) and Morrow (1989)—a partial "delta moraine" at West Oneonta and an "ablation moraine" 1 mile to the north, near the Laurens-Oneonta town line. The moraines, like the terraces elsewhere along the valley walls, are composed largely of coarse to fine sand and gravel and were deposited in water ponded at about 1,140 feet altitude behind stratified drift farther down the Susquehanna valley. Borehole records and resistivity soundings (Morrow, 1989) in two localities indicate that the valley floor is underlain by fine-grained lake deposits, which, in turn, are underlain by (basal?) sand and gravel, presumably deposited by subaquatic fans. The deep aquifer was mapped only where data were available; it might be more extensive.

2. *Interlaced Valleys.*—Near Smyrna, N.Y. (fig. 49), surficial ice-contact deposits (kames and kame deltas) are concentrated where bedrock is relatively shallow, along the side of the Chenango



Base from U.S. Geological Survey
Oneonta, N.Y. quadrangle, 1982, 1:24,000

FIGURE 48.—Surficial geology and depositional facies in Otego Creek valley where it joins the Susquehanna River valley near Oneonta, N.Y. (Developed from information in Morrow, 1989.)

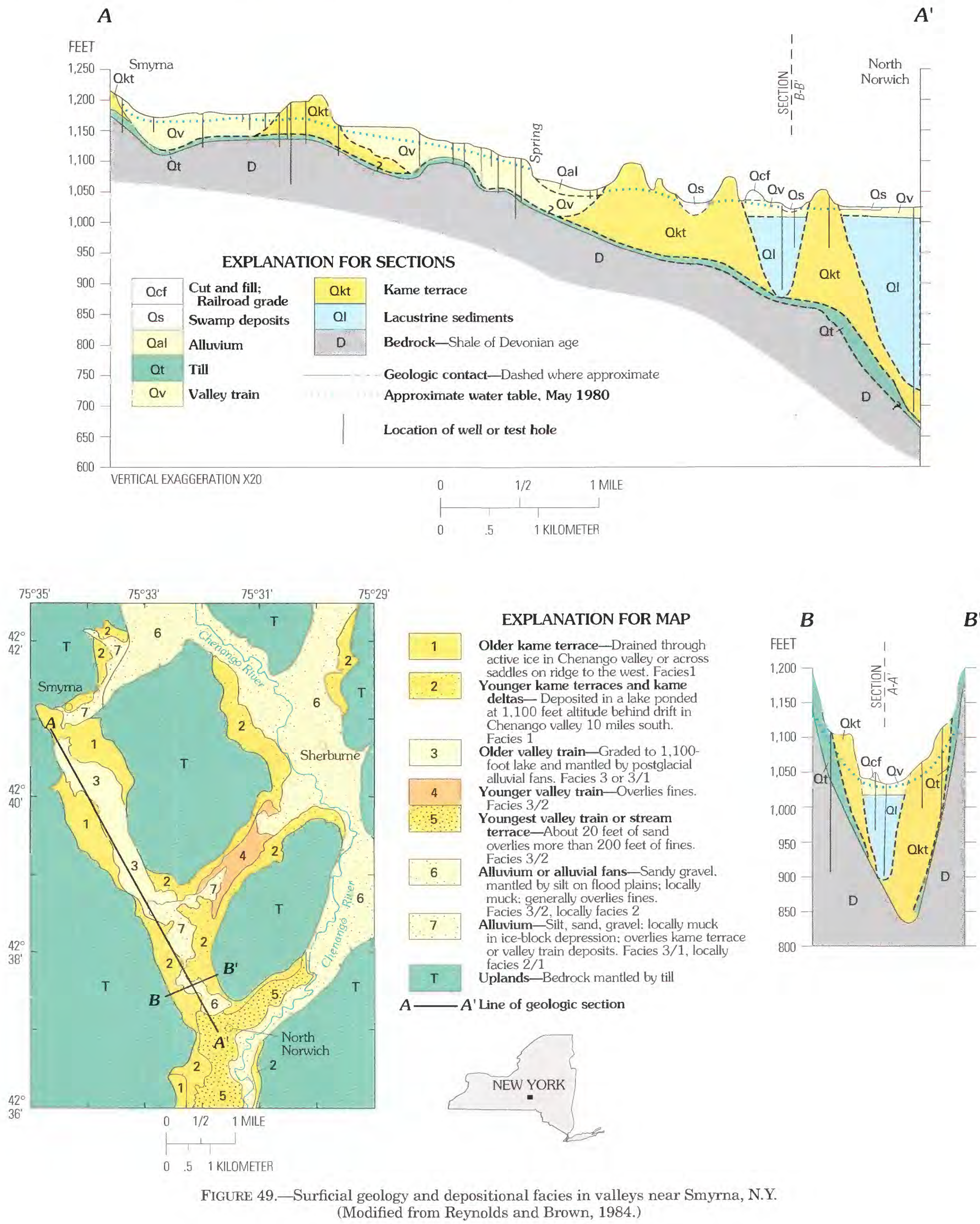


FIGURE 49.—Surficial geology and depositional facies in valleys near Smyrna, N.Y. (Modified from Reynolds and Brown, 1984.)

River valley and in subparallel tributary valleys, perhaps because meltwater flow initially became concentrated in these areas where the ice ceased flowing and melted earlier than in the deep Chenango valley. Eventually, widespread fine-grained lake-bottom deposits overlapped the ice-contact deposits in the tributary valleys (fig. 49) and accumulated to thicknesses greater than 200 feet near North Norwich and Sherburne (Randall, 1972; Reynolds and Brown, 1984). If meltwater flow was indeed concentrated in the tributary valleys early in deglaciation, basal ice-contact deposits are likely to be sparse or absent beneath the lake beds in the Chenango valley.

3. *North-Draining Valleys.*—The Tioga River and its tributaries in glaciated north-central Pennsylvania flow north or northeast (fig. 50), but their valley fills are generally similar in stratigraphy and thickness to those found in south-draining valleys of the Eastern Appalachian Plateau. Geologic sections (fig. 50) show that coarse-grained outwash or alluvium (facies 3) caps lacustrine silt, clay, and very fine sand (facies 2), which, in turn, overlaps or buries discontinuous coarse-grained ice-contact deposits (facies 1). The surficial aquifers in these valleys parallel the valley floor, commonly extend only a few feet below stream grade, and consist predominantly of modern alluvium. The fine-grained lacustrine deposits typically constitute the bulk of the valley fill, locally exceed 100 feet in thickness, and seem to be continuous for several miles (J.H. Williams, Pennsylvania Geological Survey, unpub. data, 1993).

4. *Extensive Outwash near Valley Heads Moraine.*—Cortland, N.Y., is at the junction of several valleys whose broad, level floors are underlain by outwash, in part capped by alluvium (fig. 51). The outwash thins gradually southward and eastward from the Valley Heads moraine in each valley. Geologic sections along two valleys show that lake beds underlie the outwash consistently, and nearly all of the few boreholes that extend through the lake beds have penetrated a deep gravel aquifer.

5. *Extensive Outwash near the Wisconsinan Glacial Border.*—The Allegheny River and its northside tributaries in Cattaraugus County, N.Y., contain extensive, thick, permeable surficial outwash that generally overlies fines. The Allegheny River, like other major rivers in western New York, flowed northward prior to glaciation (Calkin and others, 1974; Calkin and Muller, 1980). When advancing ice sheets blocked the former Allegheny valley, the ponded water escaped southward across a saddle near Kinzua, Pa., incising the gorge that is still occupied by the Allegheny River. Fine-grained sediments that were deposited in these proglacial lakes are found at depth along the valleys of the Allegheny River and its tributaries (Frimpter, 1974). The Wisconsinan ice sheet closely approached or barely reached the Allegheny valley all along the length of that river in New York (Frimpter, 1974; Muller, 1977), and as it did so, it released a flood of sand and gravel that was deposited as outwash atop the fine sediment in the Allegheny valley and the lower reaches of north-side tributaries (Frimpter, 1974), as illustrated in figure 52. The outwash is as much as 100 feet thick at Olean (Bergeron, 1987) and Vandalia (Frimpter, 1974). In several tributary valleys, the outwash is exceptionally thick and permeable and transmits so much underflow that the lower reaches of the tributaries are frequently dry (Frimpter, 1974). Several gravel deposits high on the sides of the Allegheny valley west of Olean have been interpreted as Illinoian (MacClintock and Apfel, 1944; Muller, 1977); apparently that ice sheet also terminated near the Allegheny River. The stratigraphic relation of these gravel deposits to the widespread lake-bottom sediments and outwash is not well documented. The general distribution of high kame terraces, lake-bottom sediments, and surficial outwash seems very similar to the rest of the Eastern Appalachian Plateau, as does the depth to bedrock of 150 to 300 feet cited by Frimpter (1974, p. 4). North of the Wisconsinan glacial border, the usual succession of facies 1 to 3 was deposited during deglaciation of south-draining tributaries to the Allegheny River. Ice-contact deposits are apparently thickest at ice-margin positions identified by Frimpter (1974, 1986), and deltas occur where smaller streams enter these valleys. Frimpter (1974, p. 42) noted that the valley of Oil Creek, a branch of Olean Creek, has few tributaries large enough to build deltas and, therefore, is filled largely with silt and clay.

WESTERN APPALACHIAN PLATEAU

Upland summits in the glaciated Appalachian Plateau west of Salamanca, N.Y., are a few hundred feet lower than those farther east, local relief is correspondingly less, and the effects of glaciation somewhat more profound. The drift in upland areas has been shown to consist of a series of sheets deposited by

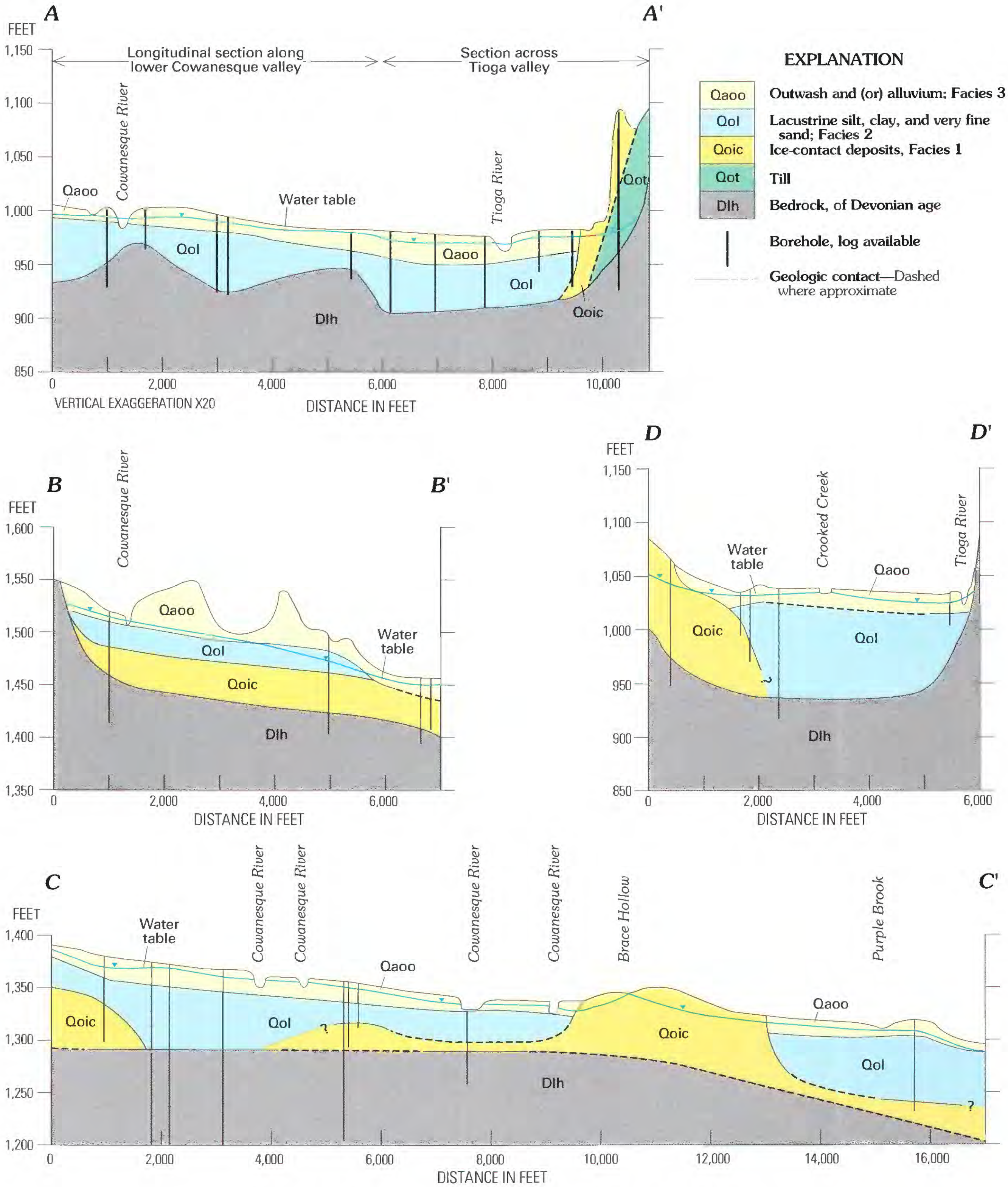


FIGURE 50.—Depositional facies in the Tioga River basin, Pennsylvania. (From J.H. Williams, Pennsylvania Geologic Survey, unpub. data, 1993.)

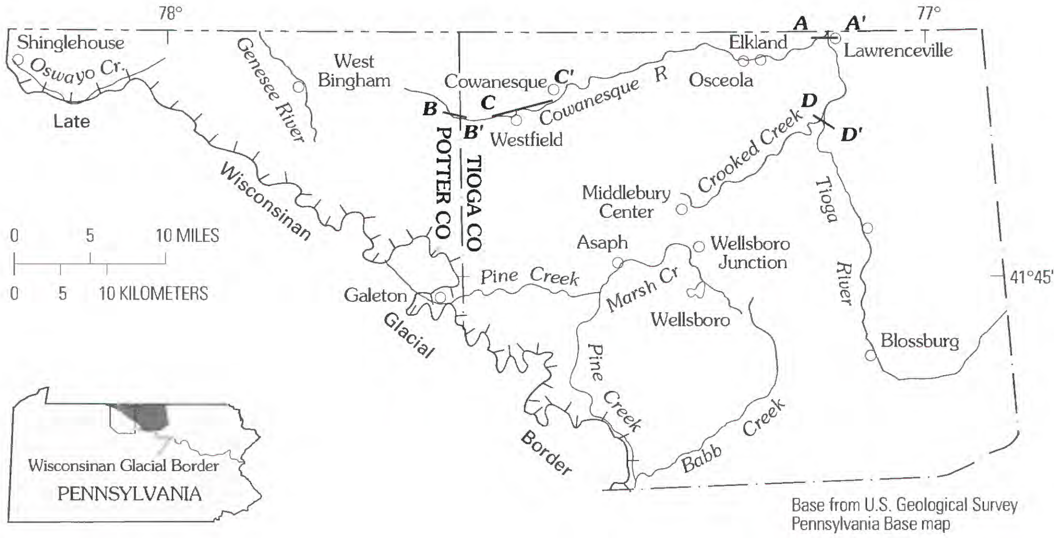


FIGURE 50.—Depositional facies in the Tioga River basin, Pennsylvania. (From J.H. Williams, Pennsylvania Geological Survey, unpub. data, 1993)—Continued.

successive ice readvances, as described below. A complex history of drainage diversion and glacial erosion has resulted in a network of intersecting bedrock valleys that also contain multiple stratigraphic units, as described in succeeding paragraphs.

Multiple Till Sheets.—The drift in most upland localities in the Western Appalachian Plateau includes several till sheets that can be distinguished by texture, weathering, or mineralogy. Each sheet varies widely in thickness from place to place; thicknesses of 2 to 20 feet are typical. A hard, compact, relatively sandy and stony till constitutes the bulk of the drift; it is termed Titusville Till in Pennsylvania and easternmost Ohio, Mogadore Till and Millbrook Till farther west (White, 1982). At least two earlier till sheets are exposed locally beneath the Titusville Till and its equivalents, and extend slightly beyond it in Pennsylvania (Shepps and others, 1959; White and others, 1969). The Titusville Till actually consists of at least three separate sheets, each of which covers a lesser area than its predecessor. The Titusville, in turn, is largely mantled by Kent Till and three younger till sheets, each of which also covers less area than its predecessor.

Layers of sand, gravel, or silt separate some till sheets in many upland localities. These layers are ordinarily only a foot or so thick but locally thicken to form buried knolls or, less commonly, channel fills, as shown in many exposures and measured sections (Winslow and White, 1966; White and others, 1969; White, 1982). Some examples are presented in figure 53. Such thin but laterally extensive aquifers commonly supply enough water for domestic wells drilled or dug in upland areas. A water-yielding sand and gravel layer is so common between the two uppermost Titusville Till sheets in Mercer and Lawrence Counties, Pa., that strip mines are ordinarily excavated in two lifts, removing the material above this layer before excavating below it, thereby avoiding the slumping this layer would cause in a vertical face (White and others, 1969, p. 29). In many places, however, multiple till sheets are stacked without stratified interbeds.

Multiple Bedrock Valleys.—Throughout the Appalachian Plateau, glaciation broadened and deepened river valleys into U-shaped troughs, some of which intersect or cross to form networks that imply profound glacial modification of the preglacial stream pattern (Coates, 1966a). In the Eastern Plateau and Catskill Mountains, virtually all of these glacially sculpted troughs served as avenues of southward meltwater flow during deglaciation. In the Western Plateau, many valleys functioned similarly but some were blocked by drift. Consequently, several large streams were diverted into narrow valley segments across former saddles, and the blocked troughs are now drained by small "underfit" streams. In this region, troughs are commonly referred to as "buried" valleys because they contain thick valley fill, and their delineation has been a major component of geohydrologic studies (Cummins, 1959; Rau, 1969; Ohio Drilling Company, 1971; Schiner and Kimmel, 1976; Richards and others, 1987). Only a very few valley segments are actually filled completely with drift (Poth, 1963, provides an example); nearly all are still recognizable as topographic valleys that contain modern streams.

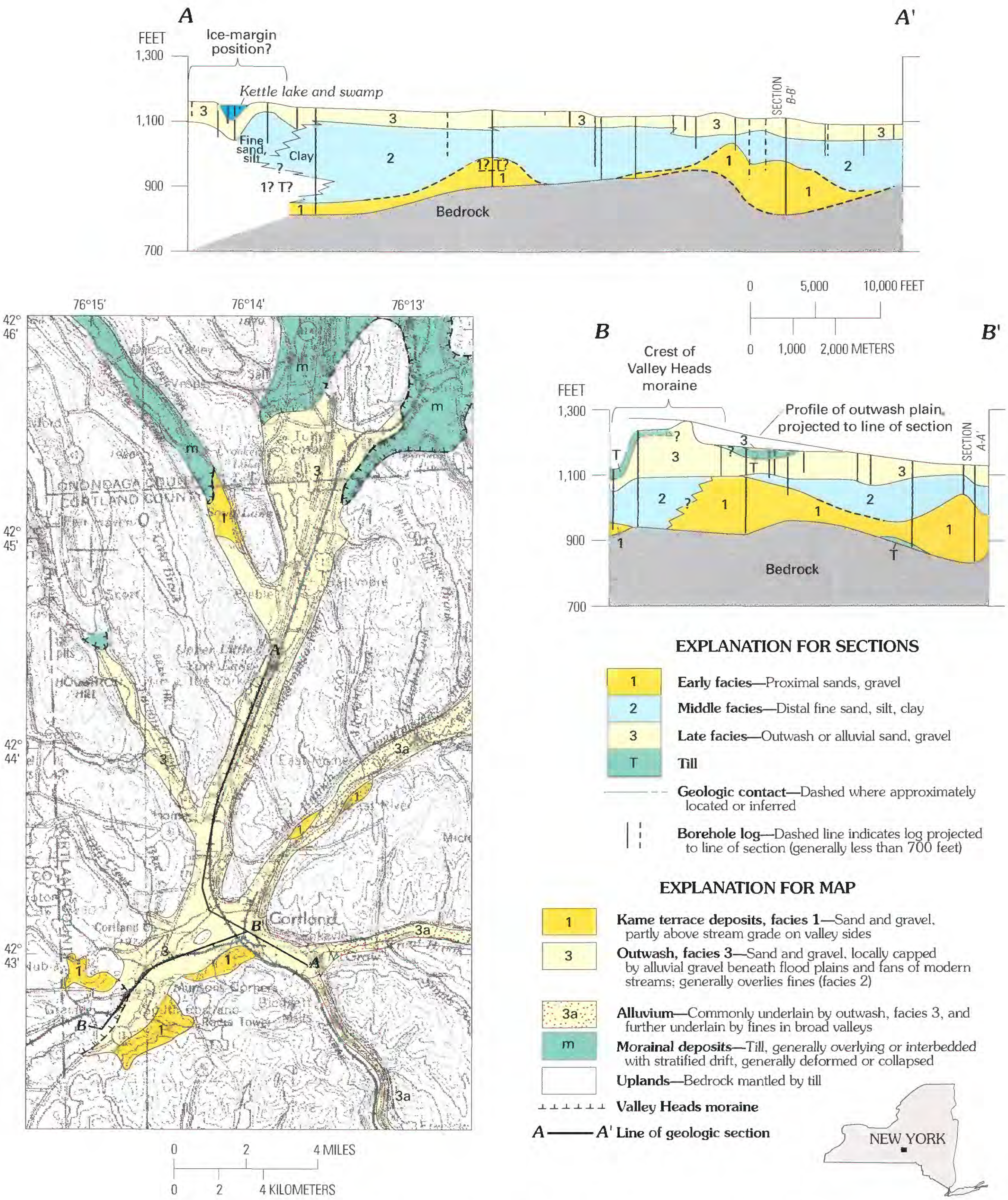


FIGURE 51.—Surficial geology and depositional facies in valleys near Cortland, N.Y. (Sections modified from Miller and Randall, 1991, fig. 2; map adapted from Muller and Cadwell, 1986, and other sources.)

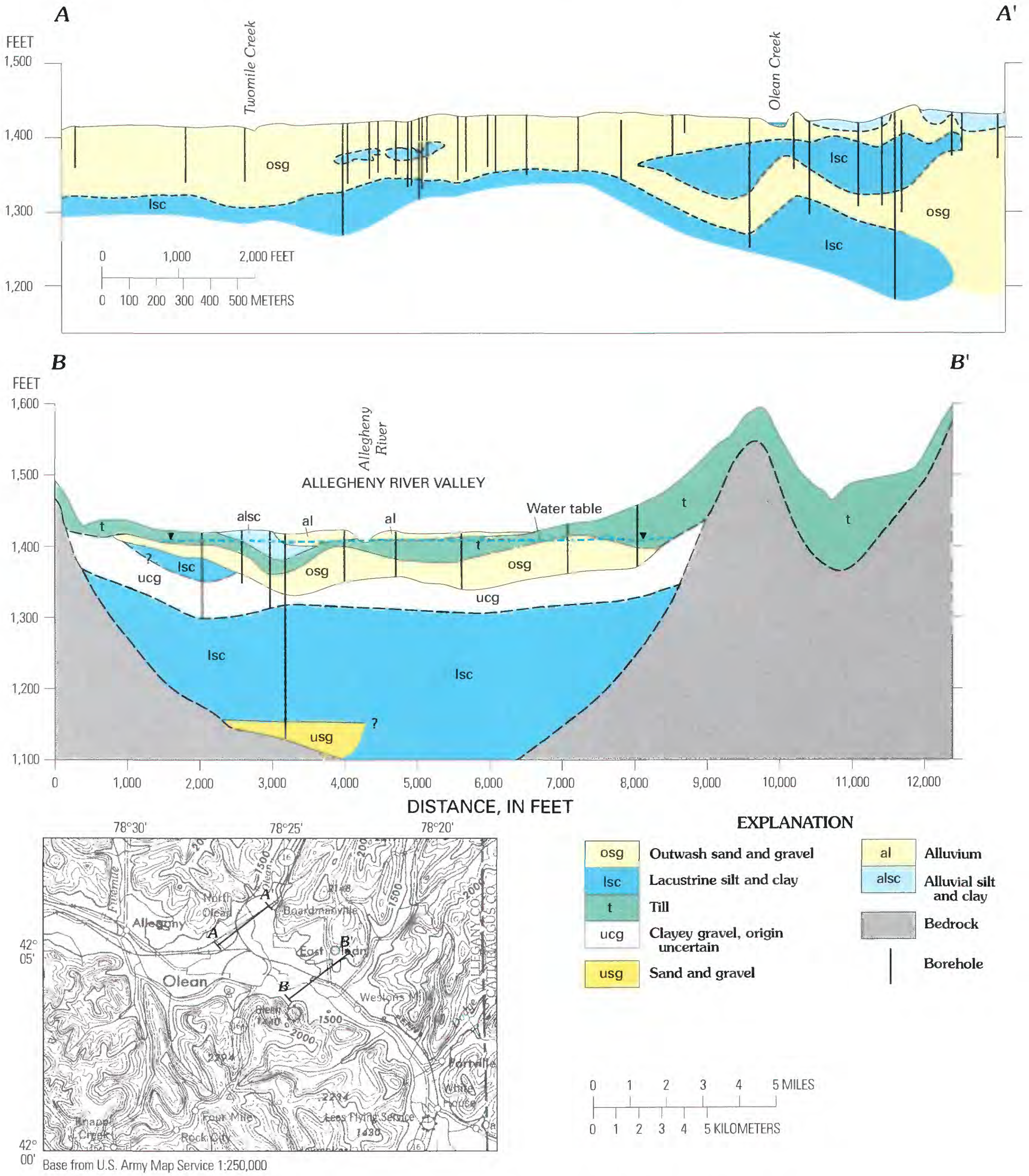


FIGURE 52.—Geologic sections in the Allegheny River valley near Olean, N.Y. (From Bergeron, 1987, fig. 2; and Zarriello and Reynolds, 1987, pl. 3.)

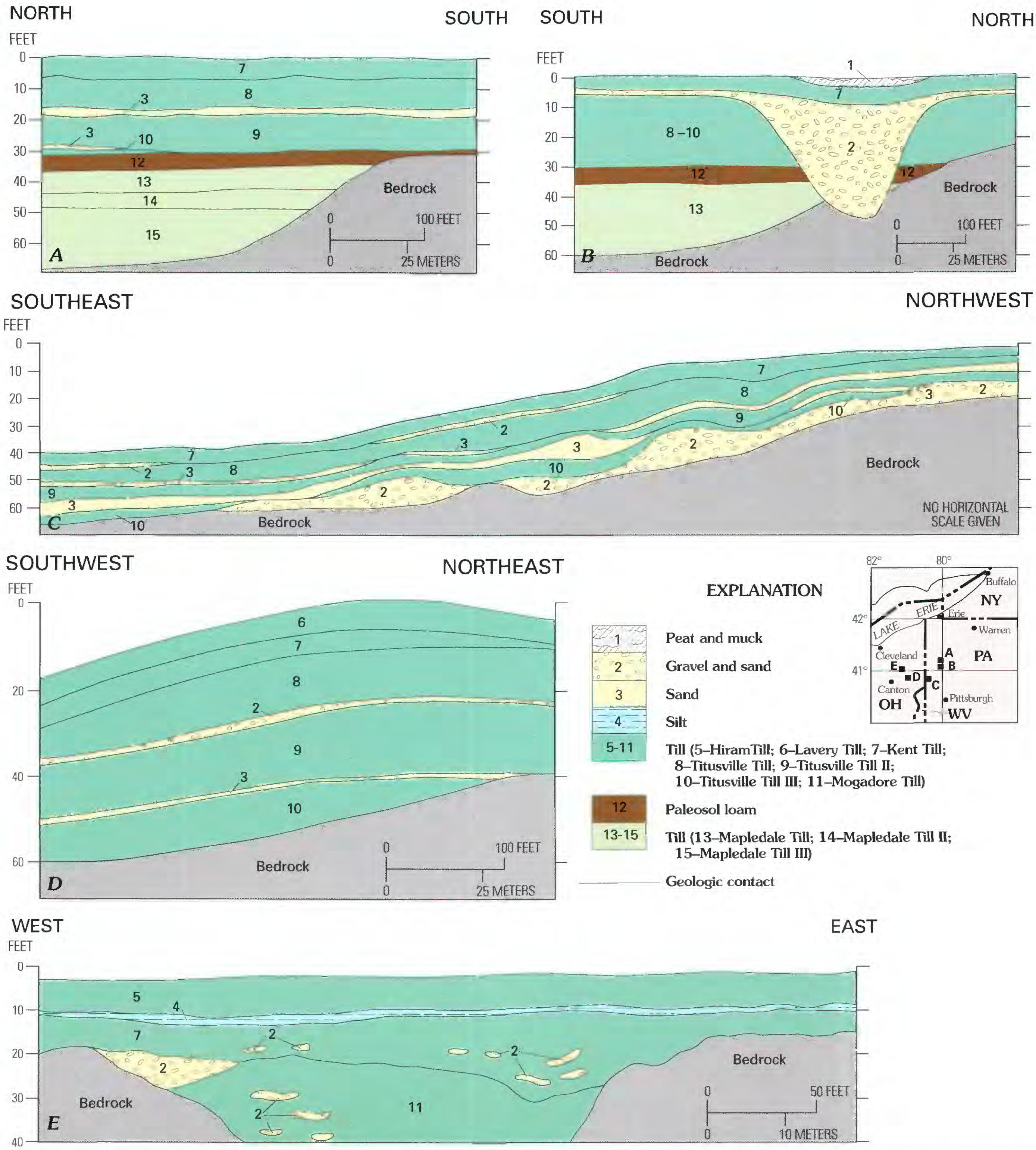


FIGURE 53.—Multiple till sheets interbedded with gravel and (or) sand, as exposed in excavations in upland localities in the Western Appalachian Plateau. Datum is land surface. (*A–C* modified from White and others, 1969, figs. 19, 26, 28; *D* modified from White, 1982, fig. 21; *E* modified from Winslow and White, 1966, fig. 10.)

Complex Valley-Fill Stratigraphy.—The multiple ice advances across the Western Appalachian Plateau left a relatively complex stratigraphic record in at least some valleys as well as in the uplands. Thick pre-Wisconsinan sand and gravel have been identified in a few valleys (fig. 54; Winslow and White, 1966; White, 1968). Many or most kames and kame terraces in northeastern Ohio are the product of early Wisconsinan (Titusville-Mogadore-Millbrook) glaciation but are mantled by till and some stratified drift

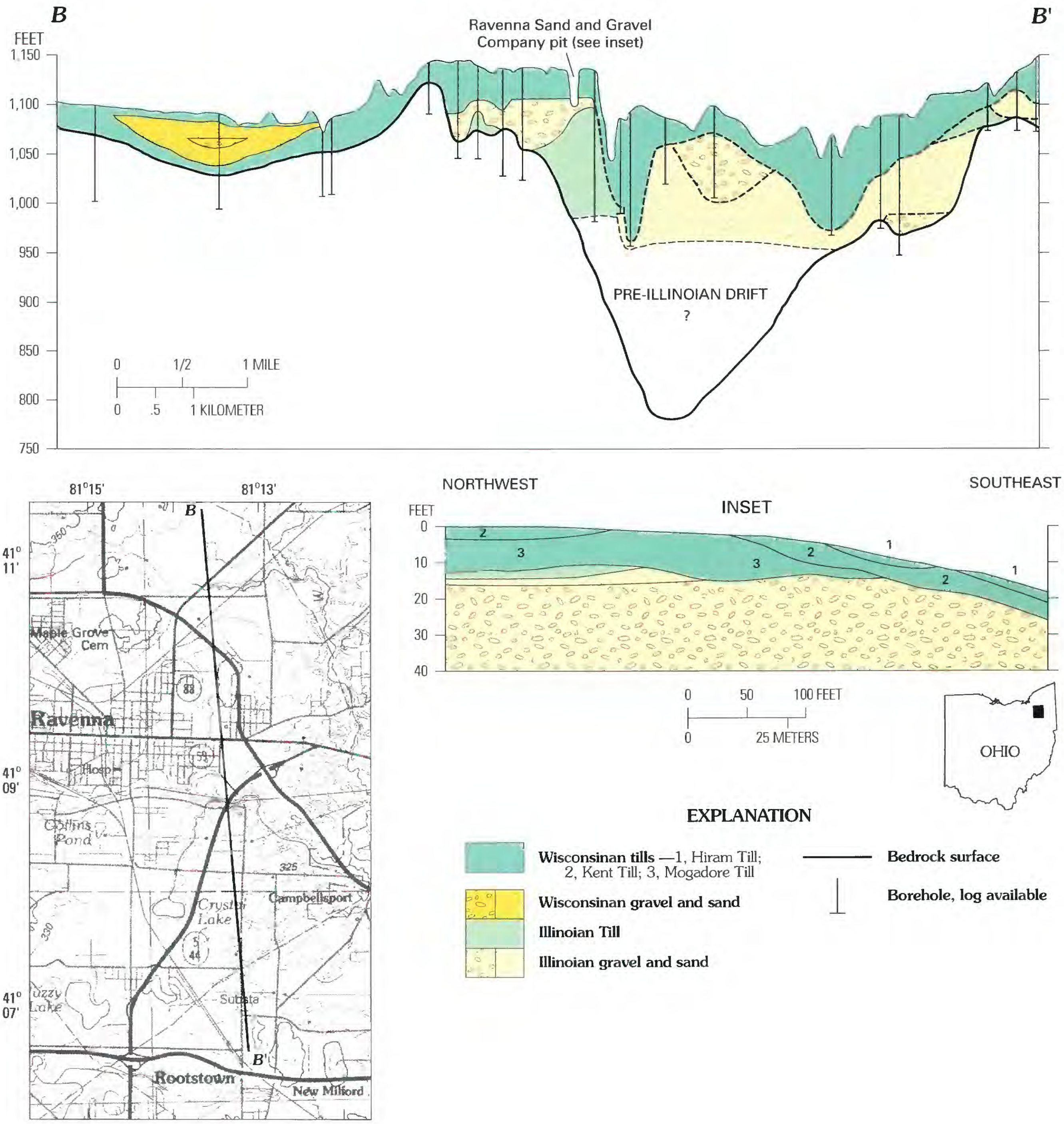


Figure 54.—Geologic section across a buried valley beneath West Branch Mahoning River near Ravenna, Ohio. (Modified from Rau, 1969, fig. 1.) Inset section shows details of till stratigraphy exposed in Ravenna Sand and Gravel Company pit. (Modified from Winslow and White, 1966, fig. 8.)

deposited during late Wisconsinan (Kent or younger) glaciation (White, 1982, p. 18, p. 39–41). Similarly, ice-contact stratified drift is extensive in valleys within the area of Kent drift in western Pennsylvania (Shepps and others, 1959) but is partly older than, and mantled by, Kent Till (White and others, 1969). Because the Kent Till is discontinuous, commonly a few feet thick at most, and (or) somewhat sandy, it is probably not a serious impediment to recharge in most places.

Some bedrock valleys in the Western Appalachian Plateau, especially those aligned east-west or northeast-southwest at a high angle to the direction of ice flow, have a thick and continuous till cover. For example, sections along a northeast-trending valley in Mercer County, Pa. (fig. 55*A*), and across the east-west valley of West Branch Mahoning River near Ravenna, Ohio (fig. 54), include thick, continuous till layers that are probably effective confining layers. Walker (1979a) recognized the buried aquifer in West Branch valley near Ravenna (fig. 54) but reported the same valley to be filled largely with fine-grained non-water-yielding sediment east of Ravenna. Other complexities, such as the presence of two till layers interbedded with stratified drift (fig. 55*C*), also have been reported in various localities but are poorly understood.

Despite the unmistakable evidence for glacial readvance throughout the Western Appalachian Plateau, the distribution of ice-contact deposits and outwash described in several reports (Shepps and others, 1959; Frimpter, 1974; White, 1982) is similar to that in regions to the east. Although few geohydrologic reports have described valley-fill stratigraphy in detail, several have reported logs of boreholes that suggest the three depositional facies that form during a single deglaciation: thin coarse-grained sediment over thick fines over coarse sediment (fig. 55*B*, *D*; Poth, 1963, p. 45; Richards and others, 1987, wells Er–808, 1423, 1648). This stratigraphy is most common in valleys aligned parallel to ice flow. Broad, flat valley floors between or bordered by ice-contact landforms mark sites of late-deglacial lakes. Wedge-shaped surficial outwash aquifers that thin downvalley and overlie lacustrine fines are documented by borehole records in several localities (Frimpter, 1974; Schiner and Gallaher, 1979, p. 9–10 and table 6).

North of the Western Appalachian Plateau is a region that generally slopes toward Lake Erie and is referred to in this paper as the Northern Rim of the Appalachian Plateau. Some valleys that now drain southward probably drained northward before glaciation and (or) during interstadial intervals. The boundary between the two regions (pl. 1) generally approximates the modern drainage divide, except that a few low-gradient valleys that currently drain southward but have little surficial outwash and probably contain till layers have been included in the north rim region. Valley-fill stratigraphy changes only gradationally across the boundary. Depth to bedrock (valley-fill thickness) and percentage of fines seem to increase northward. White (1982) reported that the thickness of surficial till atop kames in Summit County, Ohio, increases northward. Till interlayered with fine-grained lake deposits, which is typical of the north rim region, is reported in the large valleys on both sides of the present divide in western Crawford County, Pa. (Schiner and Gallaher, 1979, p. 8).

BROAD LOWLANDS INUNDATED BY LARGE WATER BODIES

The glaciated Northeast includes several broad lowlands that were inundated during deglaciation by unusually large lakes or marine embayments. Scattered hills within these lowlands protruded as islands above the proglacial water surfaces. Large, stable lakes formed in a few south-draining lowlands when drift plugged the preglacial outlets for runoff and thereby diverted meltwater across nearby bedrock saddles; these lakes expanded northward as the ice retreated, but remained approximately constant in depth because isostatic depression had temporarily eliminated or canceled the pre-glacial southward land-surface slope. These lakes were less than 200 feet deep, except where they spread across or into bedrock valleys. Large lakes also formed in lowlands along the northern border of Ohio, New York, and Vermont, where isostatic depression augmented the natural northward slope, resulting in lakes much deeper than 200 feet over large areas. Isostatic depression alone was sufficient to allow marine water to inundate coastal Maine and New Hampshire as sea level rose during deglaciation.

All these lowlands are characterized by widely distributed clay, silt, and very fine sand (facies 2). In some places, fine-grained sediments mantle the entire landscape, but more commonly they are restricted to the lower parts of the landscape and are thin, discontinuous, or absent on hills, even hills that were entirely inundated. This distribution could reflect deposition from underflows or interflows of sediment-laden water (Smith and Ashley, 1985), nondeposition in turbulent water above wave base, and (or) erosion as lake or marine levels eventually declined.

Coarse-grained ice-contact deposits (facies 1) underlie only a small fraction of the total lowland area. Commonly, they occur as linear, somewhat ridgelike features of variable width, the products of meltwater flow in subglacial channels. The channels were spaced

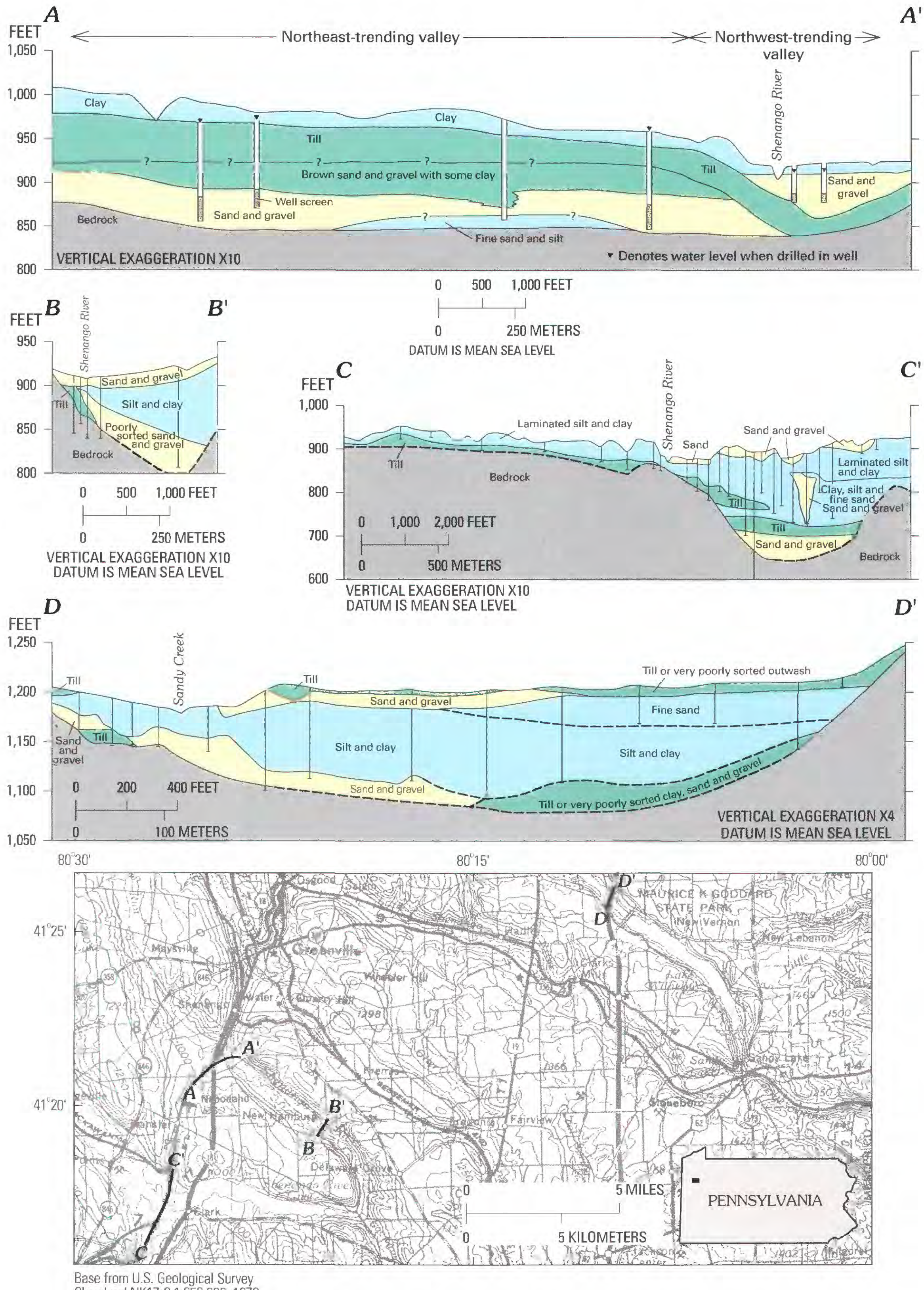


FIGURE 55.—Geologic sections in valleys in the Western Appalachian Plateau, Mercer County, Pa. (Modified from Schiner and Kimmel, 1976, fig. 8.)

a few miles apart, generally followed valleys or depressions, and were oriented subparallel to the direction of ice flow. Less common are linear accumulations of coarse stratified drift oriented perpendicular to the direction of ice flow along ice-margin positions, and semicontinuous sheets of sand a few feet thick beneath the facies-2 fines. Facies-1 deposits seem to be especially sparse in lowlands near the northern border of the glaciated Northeast—that is, on the lake plain adjacent to lakes Erie and Ontario, and in the St. Lawrence and Champlain lowlands—where large-scale calving in lakes that ranged in depth from 200 to at least 400 feet could have led to rapid deglaciation. The sparse distribution of subsurface records in these lowlands might also partly explain the lack of evidence for buried coarse sand and gravel. Lowland localities where ice-contact deposits are sparse are not separately identified on plate 1, but are noted in the locality descriptions that follow.

Coarse-grained facies-3 surficial aquifers also are much less abundant than fine-grained sediment in the broad lowlands. On plate 1, the broad lowlands are divided into three map units that reflect the distribution of surficial aquifers, as follows:

1. *Mountain-Front Perched Deltas.*—Areas where extensive inwash deltas were built by large upland streams into proglacial lakes that stood several hundred feet above the floor of the lowland; the deltas were subsequently incised by those streams and are now largely above stream grade.

2. *Lowlands Characterized by Surficial Sand Plains.*—Lowlands in which facies-2 fines are commonly (but not universally) mantled by facies-3 surficial sand-plain aquifers that consist of extensive inwash deltas, outwash plains, and (or) reworked sand deposited by postglacial streams or near-shore currents. These lowlands were occupied by long-lived water bodies and are bordered by extensive uplands.

3. *Lowlands Characterized by Surficial Fines.*—Lowlands in which the predominant surficial sediment, at least in the lower parts of the landscape, is lake-bottom or marine silt and clay, or postglacial swamp deposits or alluvium that grade down into silt or clay. Surficial facies-3 sand is absent, is of minor extent in former beaches or small deltas, or is thin and silty. These lowlands are bordered by narrow tributary uplands or are unusually broad, so only their margins received much coarse inwash from the tributary uplands.

Surficial sand-plain aquifers, although widespread in many parts of this region, are not highly productive. Commonly the sand is medium to very fine grained and becomes finer with depth. Most streams are incised through the sand into the underlying fines, and commonly near these streams less than 10 feet of the sand is saturated. Nevertheless, the sand-plain aquifers are widely tapped for domestic water supplies, and discharge from these aquifers is a persistent and important contribution to streamflow during periods of low flow.

Perhaps the most distinctive aspect of the ground-water hydrology of the broad lowlands is the slight potential for natural or induced recharge from streams. Within these broad lowlands, most streams flow across fine-grained facies-2 sediments, either on the floors of former lakes and marine embayments or incised through sand plains into the underlying fines. Ice-contact facies-1 deposits are limited in area; many form or mantle positive topographic features that constitute local topographic divides, and others are buried beneath the fines. Therefore, they are rarely in hydraulic contact with major streams that could provide induced recharge. Many tributary streams from adjacent uplands cross kame terraces, deltas, or alluvial fans along the edges of the lowlands, just as in large valleys elsewhere in the glaciated Northeast. Such coarse-grained deposits along the sides of valleys or lowlands constitute windows or conduits through which infiltrating precipitation and seepage from the tributaries can migrate downward and thereby recharge any adjacent coarse facies-1 sediments that are buried beneath fine sediment (see figs. 29–38). This mechanism could supply recharge to buried aquifers only near the margins of the lowlands, however. The many scattered hills within the lowlands that stood above the proglacial water surface were too small to have contributed appreciable inwash during deglaciation, so few of the small modern streams that descend from these hills cross windows of coarse sediment that could be sites of seepage loss and recharge to buried aquifers.

Brief accounts of the deglacial history and distribution of depositional facies in individual broad lowlands follow. Some lowlands are entirely assigned to one of the three map units depicted on plate 1, whereas others contain two or three map units.

CONNECTICUT RIVER LOWLAND AND GLACIAL LAKE HITCHCOCK

Lake Hitchcock is geologically noteworthy as the longest and the longest lived proglacial lake in New England. It extended from Rocky Hill, the geographical center of Connecticut, to Burke, Vt., only 16 miles from the Canadian border, and existed for nearly 4,000 years (Ridge and Larsen, 1990). A lowland characterized by surficial sand plains and bordering the Connecticut River (pl. 1) coincides with the former extent of Lake Hitchcock. The lake formed when deltas south of Rocky Hill (fig. 15) coalesced to block the Connecticut River valley. The deltas were not incised by subsequent runoff because a low saddle at New Britain, Conn., captured all overflow from the lake. The history of Lake Hitchcock is described by Stone and others (1982) and Koteff and others (1987). The presence of a stable lake level over a north-south distance of more than 200 miles during deglaciation has provided a basis for calculating postglacial isostatic rebound, most recently by Koteff and Larsen (1989); see also Koteff and others (1987). Lake Hitchcock was bordered by large areas of tributary upland; therefore, inwash from tributary watersheds during and after deglaciation was a major source of sand and gravel, and postglacial tributary deltas eventually prograded across the lake in some localities. Water depth was generally less than 250 feet but exceeded 300 feet locally in southern New England (Cushman, 1964; Walker and Caswell, 1977; Hansen, 1986; Koteff and others, 1987, fig. 4) and reached 650 feet in the deepest parts of the much narrower valley in southern New Hampshire (compare base of drift computed from Moore and others, 1994, with lake level from Koteff and Larsen, 1989).

The deposits of Lake Hitchcock in Connecticut and Massachusetts have yielded evidence for an active, occasionally readvancing ice tongue during deglaciation, a condition perhaps facilitated by the width of the lake and by the low relief of the Mesozoic sedimentary bedrock that underlies central Connecticut and Massachusetts. Oscillations of the ice margin do not seem to have complicated aquifer geometry, however. Most exposures cited as evidence for readvance displayed an unconformity within deltaic or lake-bottom sediments. The unconformity is commonly marked by stones and by till lenses that contain lacustrine fragments, and beds below the unconformity are deformed and thrust-faulted, but grain size and depositional structure above and below the unconformity are generally similar (Larsen, 1982; Ashley and others, 1982, stops 3–4; Larsen and Hartshorn, 1982). Larsen and Hartshorn (1982, p. 123) attribute a large delta-outwash plain in a tributary lake just west of Lake Hitchcock to deposition beyond the ice front following readvance, but they do not make any claim that a sediment package of different lithology resulted from the initial retreat through the same valley reach. Apparently, conditions controlling sediment delivery to the lake were largely independent of oscillations of the ice margin.

FACIES DISTRIBUTION IN CONNECTICUT AND MASSACHUSETTS

Lake Hitchcock averaged about 9 miles in width in central Connecticut and Massachusetts (Stone and others, 1979). The distribution of proximal ice-contact deposits, fine-grained lake-bottom deposits, and surficial sand deposits in this area is discussed below. The latter two facies are far more abundant in the broad lowland inundated by Lake Hitchcock than in the narrow valleys of the adjacent Eastern Hills and Valley Fills region.

Ice-Contact Deposits.—Ice-contact stratified drift is abundant near the margins of the lowland underlain by Mesozoic bedrock in central Connecticut and Massachusetts, most of which is drained by the Connecticut River. Regional maps by Meade (1978) and Stone and others (1979) show this pattern consistently, as do numerous local studies (for example, Campbell, 1975). Coarse-grained stratified drift was concentrated near the sides of the lowland because ice stagnated earlier there than in the middle of the lowland, as a result of higher bedrock altitudes and higher relief, and because much inwash followed tributary valleys toward the lowland. Some of the coarse stratified drift along the sides of the lowland was actually deposited in earlier, higher lakes and in tributary valleys beyond Lake Hitchcock. In particular, coarse sediment from the west was trapped in a valley parallel to the Connecticut River lowland but west of Talcott Mountain; therefore, only small areas of ice-contact deposits are found on the east side of Talcott Mountain within Lake Hitchcock.

Ice-contact deltas or kames have been identified in a few midvalley localities and seem to be associated with buried aquifers. For example:

1. Along the Connecticut River in the towns of Hadley and South Hadley, Mass., subsurface data suggest that buried sand and gravel deposits as much as 120 feet thick link two prominent kame deltas 7 miles apart (fig. 27).

2. Immediately south of the Farmington River in Windsor, Conn. (southwest corner of fig. 56), two ice-marginal deltas were built successively (Koteff and others, 1987) on shallow bedrock (Cushman, 1964). Coarse-grained stratified drift is exposed, or penetrated by wells beneath fines (fig. 56), at several points 0 to 3 miles north of the younger ice margin (Ryder and others, 1981).

3. East of the Connecticut River near the Connecticut-Massachusetts border, gravel deltas that stand above the surrounding terrain were deposited in a northeast-trending arc (fig. 56) along the margin of an active ice tongue (Larsen and Hartshorn,

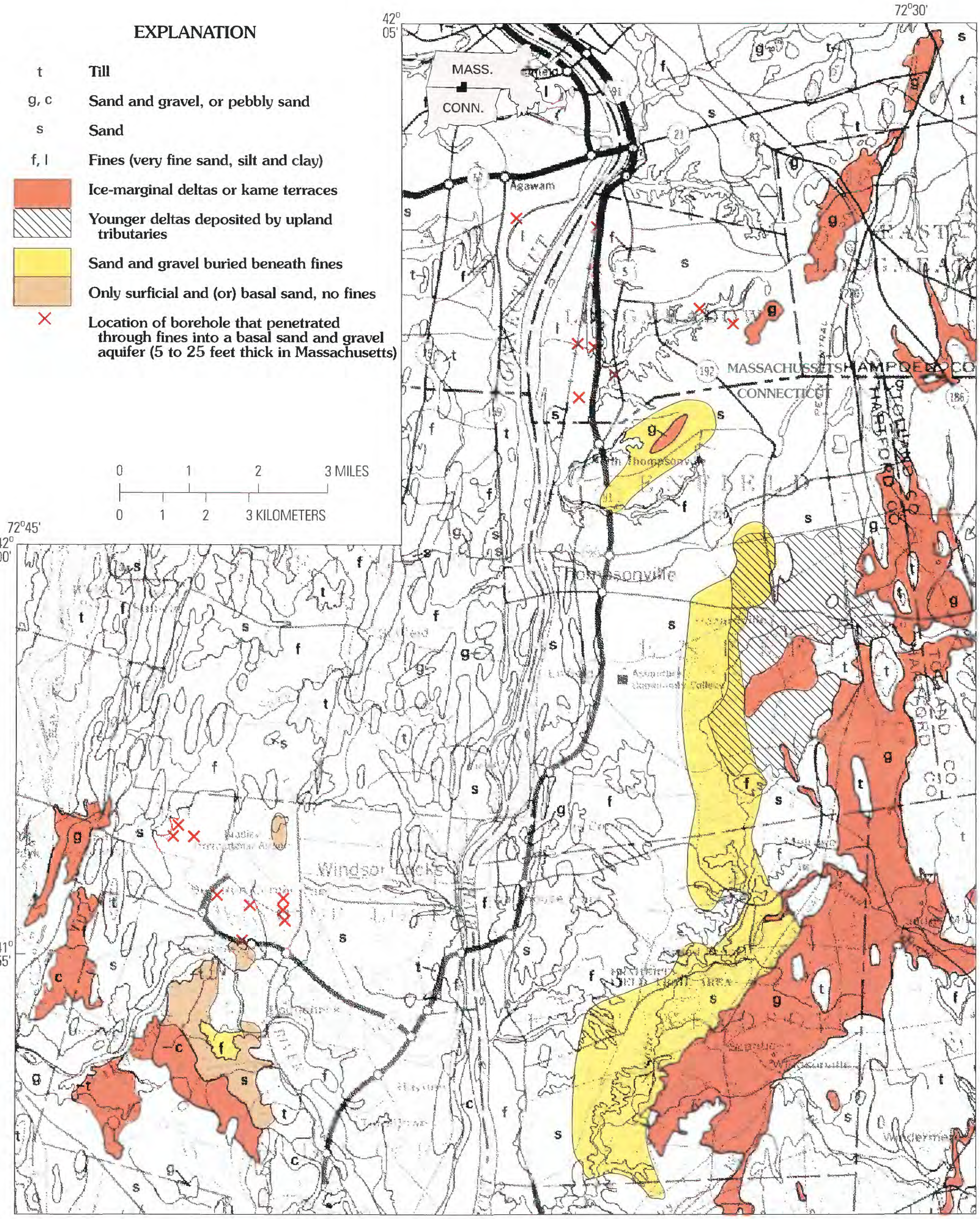


FIGURE 56.—Surficial geology and inferred distribution of buried facies-1 sand and gravel in part of the Connecticut River lowland. (Modified from Stone and others, 1979, according to interpretations by Ryder and others, 1981; Koteff and others, 1987; and Stone and others, 1992.)

1982, fig. 6, p. 124). An apron of buried sand and gravel was inferred by Ryder and others (1981, pl. B) to surround the westernmost knoll, in northernmost Connecticut (fig. 56), and might extend north into Massachusetts where several nearby boreholes penetrated sand and gravel beneath fines (Walker and Caswell, 1977).

4. In some places east of the Connecticut River in northern Connecticut, clay deposited in Lake Hitchcock is underlain by water-yielding gravel whose presence has long been known (Cushman, 1964) but poorly understood. Ryder and others (1981) and Stone and others (1992) depicted a nearly continuous north-south band of buried sand and gravel extending as much as 1 mile westward from the west edge of multiple deltas that border the lowland in Enfield and East Windsor (southeast one-third of fig. 56). According to Koteff and others (1987, p. 17), the earliest, easternmost deltas were ice marginal, but most of the deltaic complex consists of younger deltas (fig. 56) built westward into Lake Hitchcock by tributaries before and after ice disappeared from their watersheds. If so, any buried sand and gravel along the western margin of the younger deltas cannot be the collapsed head of the younger deltas; instead, it might have been deposited along several channels that fed the earlier deltas or might constitute a continuous band (as inferred by Ryder, 1981) deposited at the mouth of a single, persistent north-south channel as the ice margin retreated.

Although basal facies-1 aquifers underlie the clays of Lake Hitchcock in some places, as shown by the foregoing examples, more commonly they are absent or thin and discontinuous, as demonstrated by test borings (Cushman, 1964, pl. 3; Deane, 1967; Campbell, 1975, p. 74; Hansen, 1986) and illustrated by geologic sections (figs. 57, 58). Cederstrom and Hodges (1967), Meade (1978), and Ryder and others (1981) recognized only small areas within Lake Hitchcock where coarse-grained stratified drift is predominant or constitutes a

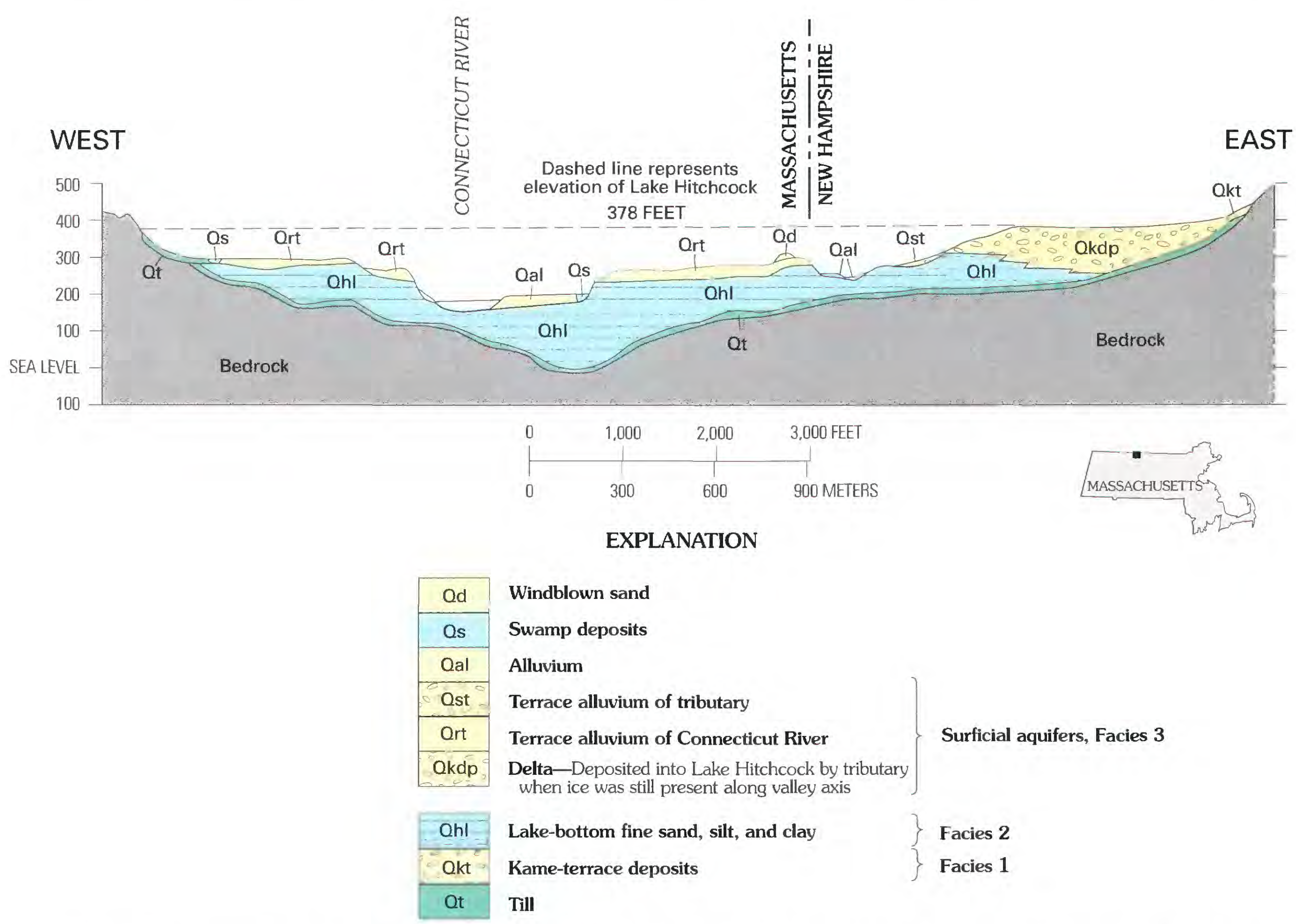


FIGURE 57.—Generalized section across the Connecticut River valley. Line of section nearly coincides with northern border of Massachusetts. (Modified from Campbell and Hartshorn, 1980.)

basal aquifer beneath fine-grained sediment. Walker and Caswell (1977), on the other hand, stated that "a basal layer of sand and gravel occurs widely but not everywhere" beneath Lake Hitchcock sediments in Massachusetts; their maps show 114 sites where boreholes penetrated a buried aquifer but indicate thickness to be generally less than 16 feet and do not show sites where no buried aquifer was penetrated.

Lake-Bottom Deposits.—The extensive bottom deposits of Lake Hitchcock have been studied in detail by Antevs (1922), Ashley (1972), and Ridge and Larsen (1990). They range in grain size from silty fine sand to rhythmic silt and clay. In most places, grain size decreases upward because the upper sediment layers were deposited farther from the northward-retreating ice margin where most sediment entered the lake

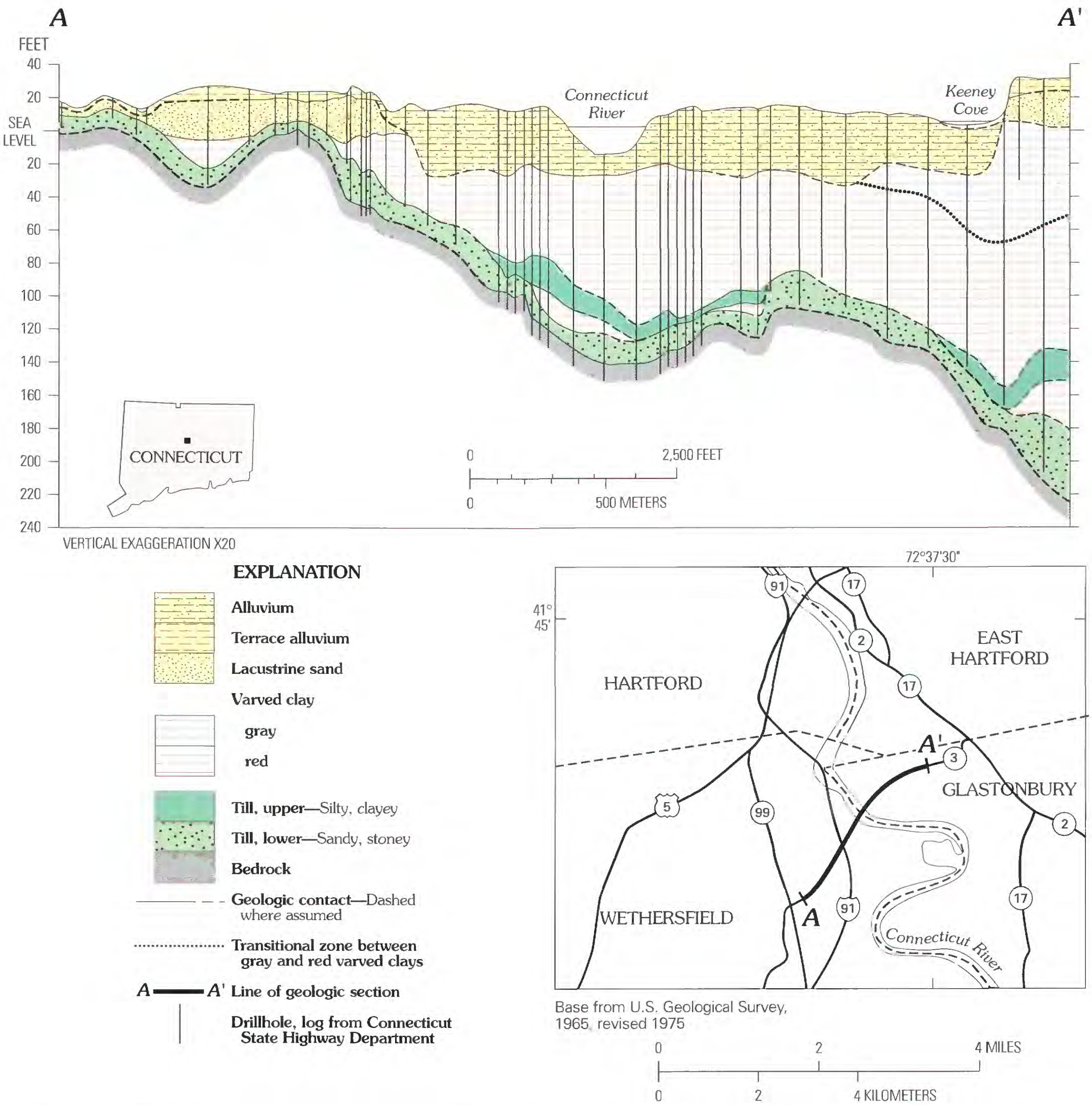


FIGURE 58.—Geologic section across the Connecticut River valley along State Highway 3 in Wethersfield and Glastonbury, Conn. (From Deane, 1967.)

(Hansen, 1986), but in the extensive tributary deltas, grain size increases upward. In Connecticut and Massachusetts, the deep deposits are red because they were derived from subglacial meltwater that contained abundant fragments of Mesozoic red sandstone and shale that underlie the Connecticut River lowland; the younger layers are gray because sediment from tributaries that drain highlands of crystalline bedrock to the east and west came to predominate as the ice margin retreated (fig. 58).

Surficial Aquifers.—Throughout the existence of Lake Hitchcock, tributaries built deltas into the lake. Sediment volumes were greatest at first, when decaying ice in the tributary watersheds and areas to the north that temporarily drained into those watersheds generated abundant meltwater, and streams were rapidly regrading their till-choked upland valleys. At the distal margins of these deltas, fine-sand foresets flatten and grade into the coarse "summer" layers of lake-bottom varves, as shown in many exposures (Webb, 1957, mile 54.6; Ashley and others, 1982, stops 3 and 5; Koteff and others, 1987, stop 9; McElroy, 1987). When Lake Hitchcock drained, redistribution of sediment was again rapid as the deltas, as well as the upland valleys, were incised; streams built fans and flood plains across the former lake floor and built deltas into small, shallow lakes that remained in depressions on the lake floor. Continued incision has lowered present channels to well below the initial post-lake flood plain.

One result of all this sediment redistribution and incision is widespread surficial sand-plain aquifers. They consist chiefly of very fine to medium sand and generally are perched above stream grade; that is, most streams are incised below the level at which sand rests upon or grades into the underlying clay-silt rhythmites, as clearly shown on maps of surficial sediments in central Connecticut (Stone and others, 1979) and in South Hadley, Mass. (Saines, 1973). Although the surficial sand is only thinly saturated near the incised streams, elsewhere it is capable of supplying 5 to 50 gallons per minute to individual springs, dug wells, or driven well points. Cushman (1964) and Ryder and others (1981) indicated that yields of 100 to 400 gallons per minute are possible at a few locations. Ryder and others (1981, pl. B) mapped most of the former lake area as a surficial sand aquifer overlying clay to very fine sand. They recognized two categories of saturated thickness of the surficial sand: 0 to 15 feet, and 15 to 80 feet. An excavation within the thinner category (Ashley and others, 1982, stop 2) cut through 5 to 10 feet of tabular cross-bedded layers of fine to very fine sand into varved clay; springs issued from the lower part of the sand. Other studies have reported surficial sand thicknesses of 0 to 50 feet (Antevs, 1922) and 30 feet (Hansen, 1986) atop clay.

FACIES DISTRIBUTION IN VERMONT AND NEW HAMPSHIRE

In Vermont and New Hampshire, Lake Hitchcock was much narrower than in southern New England (Ridge and Larsen, 1990, fig. 11), relief bordering the lake was greater (Denny, 1982), the rate of ice retreat was probably more rapid (Schafer, 1979), and the ice margin was apparently characterized by extensive stagnation rather than by active ice (Stewart and MacClintock, 1969; Larsen, 1982; Larsen and Koteff, 1988). Large volumes of water flowed south through subglacial tunnels near the valley axis, depositing esker segments of varying lengths that commonly constitute productive aquifers, although they underlie only part of the valley width and are locally discontinuous or perched on bedrock above the water table. Eskers are well documented in the following localities:

1. At least 10 successive ice-contact, apparently esker-fed deltas lie near the Connecticut River between the Massachusetts State line and Putney, Vt. (Larsen and Koteff, 1988, p. 110). Several exposures revealed a ridgelike gravel core mantled by deltaic and (or) collapsed sand and, near the top, lake-bottom very fine sand or silt.
2. Many esker segments, some continuous for several miles, are present along the Connecticut River valley over at least 18 miles from Lyme, N.H., to Hartland or Windsor, Vt. (Lougee, 1939; Stewart and MacClintock, 1969, p. 91; Larsen, 1987a). Hodges and others (1976b) demonstrated that these features constitute highly productive aquifers.
3. An esker follows the lower reaches of the White River valley, a tributary to the Connecticut River valley that also was occupied by Lake Hitchcock (Larsen, 1987a). Where best exposed between Sharon and South Royalton, Vt. (Larsen, 1987a, stop 5), the esker is largely above the water table, as indicated by bedrock exposures in the river channel, but Hodges (1968c) depicted a continuous productive aquifer along the White River valley for several miles downstream from this point.
4. The Passumpsic River valley, another tributary that was inundated by Lake Hitchcock, contains an esker that extends for 24 miles from East Haven, Vt., nearly to the Connecticut River. Stewart and MacClintock (1969, p. 93) note that near St. Johnsbury, Vt., the esker is massive but nearly buried by lake sediment.

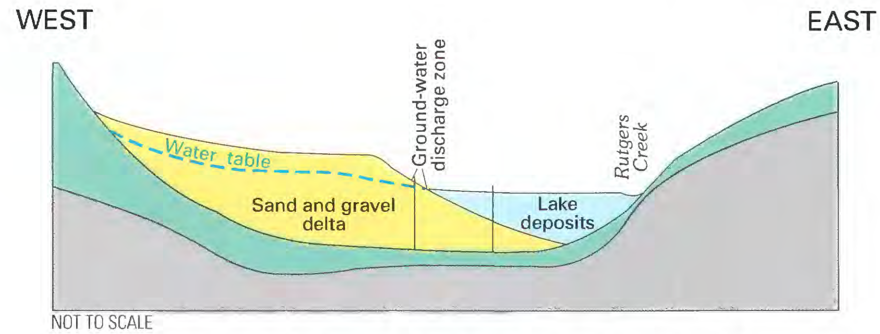
All along the Connecticut River valley and tributaries formerly occupied by Lake Hitchcock, eskers and related ice-contact deposits are continuously overlapped or buried by lake-bottom varved silt and clay (Antevs, 1922; Lougee, 1939, p. 137; Larsen, 1987a, stop 5 and mile 66.6; Larsen and Koteff, 1988, stop 5A) and by fine sand to silt that could represent both tributary deltas and fining-upward sequences deposited by retreating subglacial drainage. The proximal coarse ice-contact deposits underlie only a fraction of the valley floor; elsewhere fine-grained lake deposits rest directly on till or bedrock (Antevs, 1922; Stewart and MacClintock, 1969, p. 102; Hodges and others, 1976b). Stewart and MacClintock (1970) mapped extensive lake-bottom silt and clay. Exposures along the White River valley (F.D. Larsen, Norwich University, written commun., 1985) indicate that the entire valley was filled with fine to very fine sand rhythmically alternating with thin layers of silt or clay. Sand was deposited widely as deltas into Lake Hitchcock and as river terraces after the lake was drained (Larsen and Koteff, 1988, p. 114), especially south of White River Junction (Stewart and MacClintock, 1970), but many of these surficial sand deposits are narrow, deeply incised, and overlie very fine sand or silt that transmits water downward; therefore, they drain readily and are not dependable aquifers.

WALLKILL LOWLAND

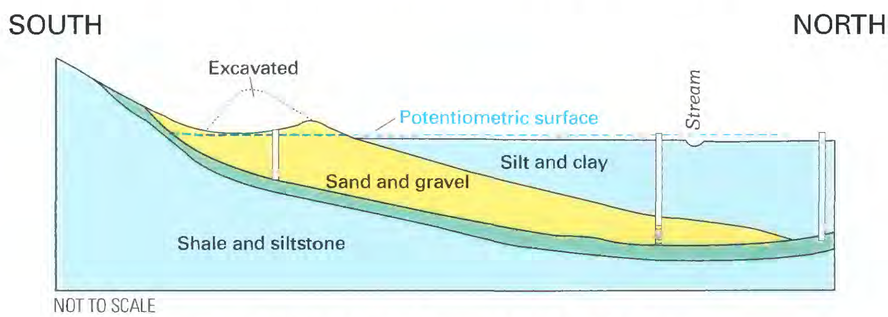
A broad lowland underlain chiefly by shale, siltstone, and carbonate bedrock lies west of the Hudson River in Ulster and Orange Counties, N.Y., and Sussex County, N.J. It is drained primarily by the northeastward-flowing Wallkill River and was extensively inundated by lakes ponded in front of the retreating Wisconsinan ice sheet, as represented by the "L" symbol on plate 1. Early in deglaciation, the ponded water drained southward to the Delaware River across saddles 520 to 500 feet in altitude near Augusta, N.J. (Connally and others, 1989; Stanford and others, 1998). Later, lake level dropped as meltwater escaped eastward to the Hudson River across a succession of progressively lower saddles. The stratified deposits in this lowland have been variously interpreted as deposited amid stagnant ice or in front of active ice (Dineen, 1986a). Diamicton layers capping or interbedded with coarse-grained stratified drift provide evidence of locally active ice at Augusta, N.J. (R.W. Witte, New Jersey Geological Survey, oral commun., 1989), Pellets Island, N.Y. (Connally and others, 1989, stop 3) and nearby areas (Bugliosi and others, 1998), and Rosendale, N.Y. (Waines, 1979). Modern streams generally flow along wide drift-filled valleys that parallel the north-northeast strike of the bedrock, but locally cut across ridges in narrow valley reaches.

Ice-Contact Deposits (Facies 1).—Early-deglacial ice-contact deposits, whose hummocky topography suggests collapse, are scattered across the Wallkill lowland. Some lie within bedrock valleys, bordered and overlapped by younger lake-bottom sediments (figs. 59, 60), but because lake levels were higher than many bedrock ridges, some ice-contact deposits are draped over or banked against ridges above modern stream grade and are largely above the water table (Frimpter, 1972, pl. 2). Frimpter (1972) and Stanford and others (1998) inferred that the ice-contact deposits exposed discontinuously along several valleys are linked by continuous basal gravel aquifers buried beneath fines. For example, as explained earlier in the section titled "Meltwater Channels in Lowlands," exposed and buried ice-contact sand and gravel are apparently continuous for at least 19 miles along the west side of the Wallkill valley from New Hampton, N.Y. (fig. 60*B*), south to Hamburg, N.J.; this suggests a persistent subglacial meltwater channel. Buried ice-contact deposits seem less abundant in the part of the Wallkill lowland north of New Hampton (Frimpter, 1972, pl. 2). The Wallkill lowland is bordered on the northeast by the Hudson River trench, on the southeast by the Hudson highlands, which drain largely east or south, and on the west by a narrow ridge, Shawangunk Mountain; therefore, inwash was slight and no concentration of ice-contact deposits is found along the lowland margins.

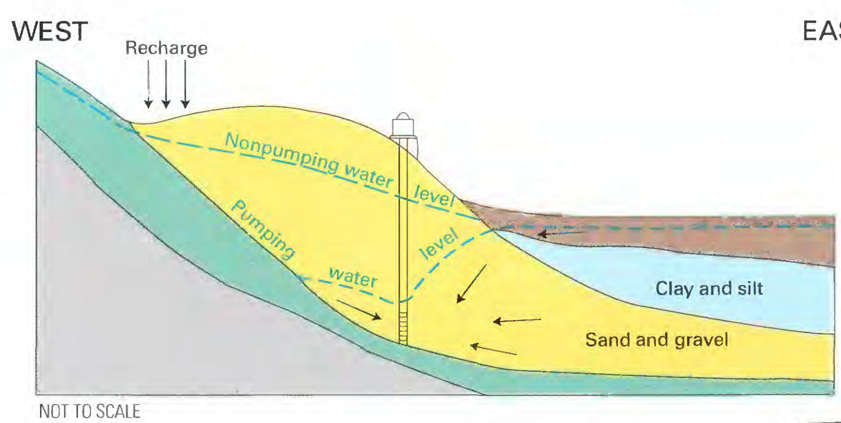
Surficial Aquifers (Facies 3).—Silt and clay are widespread in the valleys of the Wallkill lowland, at land surface or mantled by Holocene peat and muck (Frimpter, 1972; Cadwell, 1989). In only a few localities are they overlain by significant thicknesses of coarse-grained sediment. Three deltas or delta segments are recognized in the lowest reach of Rondout Creek valley, near and north of Rosendale (Frimpter, 1972; Waines, 1979; Cadwell, 1989). The largest of these, at Tillson, blocked the ancestral Wallkill valley and diverted the Wallkill River through a postglacial gorge; it constitutes a significant sand-plain aquifer (Frimpter, 1972). Exposures and well records indicate that the surficial sand that caps all three deltas is everywhere underlain by silt and clay; little evidence has been reported of a morphosequence head or ice margin where coarse sediment extends from land surface to bedrock. For this reason, and because surficial diamicton has been observed locally, these deltas have been interpreted as the product of glacial readvance into an established (or reestablished) proglacial lake (Waines, 1979; Dineen, 1986a). Surficial deltaic or lake-bottom sand also caps lacustrine silt and clay in the middle Rondout Creek valley between Wawarsing and High Falls; it is the product of abundant meltwater from ice near Ashokan Reservoir pouring south through a variety of channels into a lake in Rondout Creek valley (Dineen, 1986a). The surficial sand is thin and fine grained, however, and was not mapped or mentioned by Frimpter (1972); it is, at best, a minor sand-plain aquifer. The only other extensive surficial aquifers in the Wallkill lowland are the several feet of alluvial gravel that underlie the flood plains of the large streams.



A. **Schematic geologic section across Rutgers Creek Valley near Westtown, N.Y.**



B. **Schematic geologic section at Baird's Crossroad near Chester, N.Y.**



C. **Idealized view of hydrogeologic conditions in the Wallkill lowland**

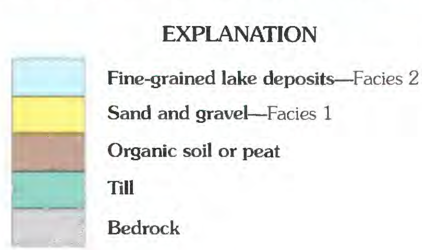
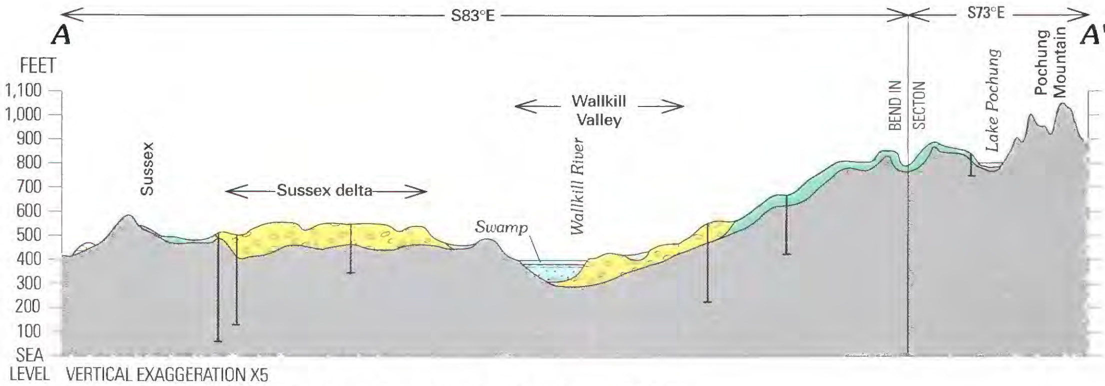
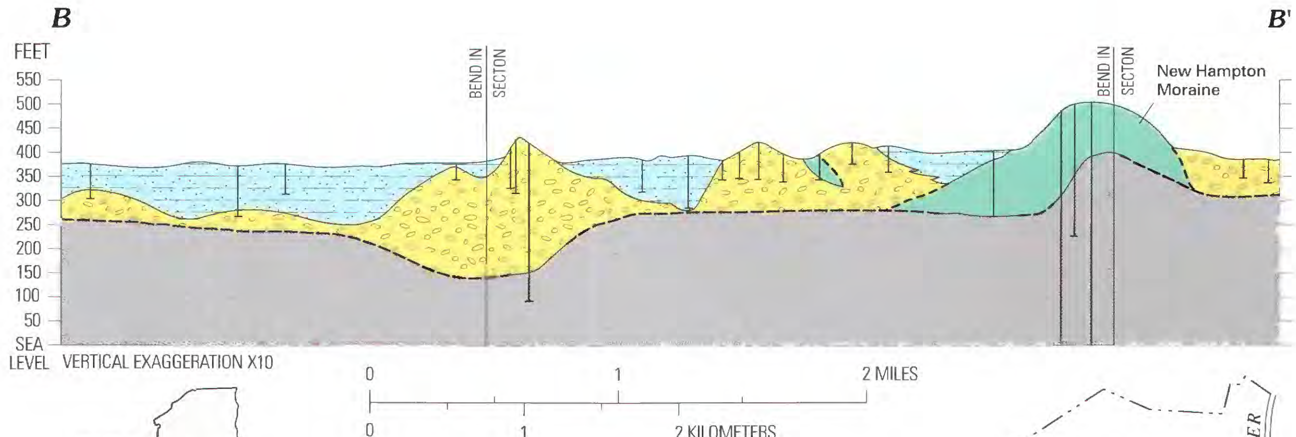


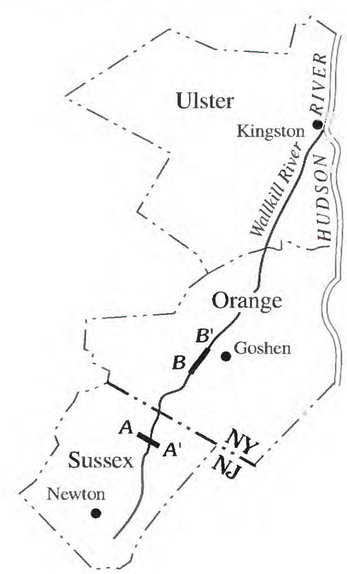
FIGURE 59.—Typical relations of facies-1 aquifers and facies-2 fines in the Wallkill lowland, New York. (From Frimpter, 1972, figs. 18, 22, 23.)



A. **Near Sussex, N.J. (From Stanford and others, 1998)**



B. **Along the west side of the Wallkill valley near New Hampton and Pellets Island, N.Y. (Modified from Bugliosi and others, 1998)**



EXPLANATION

- **Till**
- **Sand, gravel as deltas or subaquatic fans (facies 1)**
- **Clay, silt, and fine sand**
- **Bedrock**
- **Borehole**
- **Geologic contact**—Dashed where uncertain
- *A* —— *A'* **Line of geologic section**

FIGURE 60.—Geologic sections near the Wallkill River in New Jersey and New York.

Distribution of Coarse Facies Relative to Streams.—The Wallkill lowland offers few opportunities to obtain large ground-water withdrawals sustained by induced infiltration. The major streams (Wallkill River, lower Rondout Creek, lower Esopus Creek) and some tributaries flow chiefly across fine-grained lake deposits; in only a few places do they intersect ice-contact deposits (Frimpter, 1972; Cadwell, 1989). The alluvial deposits beneath flood plains and tributary fans include permeable gravel but are ordinarily so thin that large yields are obtainable only from wells of uncommonly large effective diameter or infiltration galleries. Some tributary valleys (including Wawayanda and Rutgers Creeks and upper Shawangunk Kill, also Tin Brook east of Walden) contain abundant kames or outwash deposits, but contact between these tributaries and the streams is likely only in headwater reaches where streamflow is too small to provide much induced infiltration.

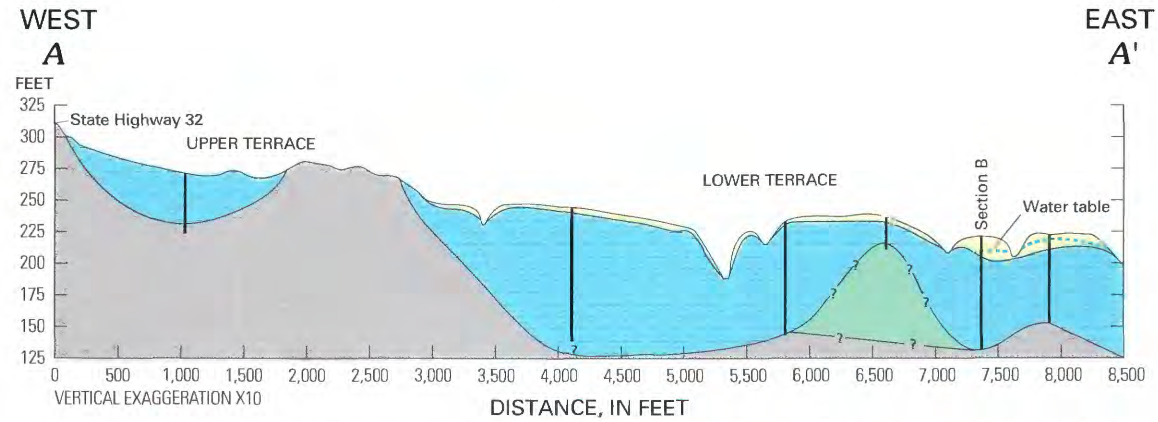
The ice-contact deposits that are exposed locally along the sides of several reaches of major valleys are inferred to constitute productive aquifers and sources of recharge to adjacent buried sand and gravel (figs. 59, 60). The description of partly buried aquifer systems along some margins of the upper Wallkill and Black Meadows valleys by Frimpter (1972, p. 39–43) suggests that large seasonal water supplies could be obtained during periods of low flow without significantly depleting streamflow (fig. 59).

HUDSON RIVER LOWLAND AND GLACIAL LAKE ALBANY

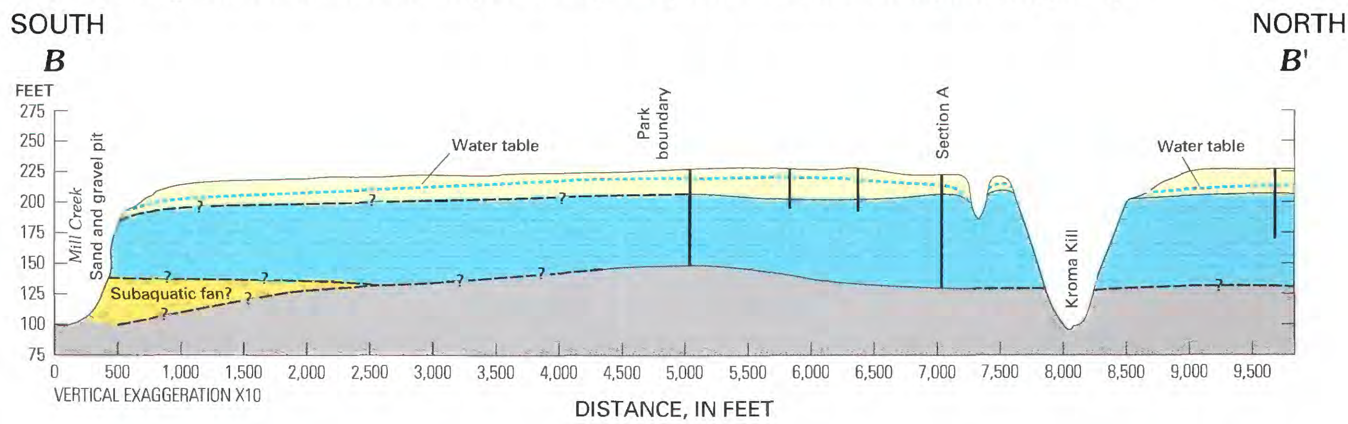
Bedrock beneath the Hudson River is at least 200 feet below sea level everywhere south of Ulster County, N.Y., and more than 700 feet below sea level in some places (Berkey and Rice, 1921; Perlmutter, 1959; Worzel and Drake, 1959; Newman and others, 1969; Soren, 1988). Nevertheless, the Hudson valley was apparently filled at New York City by the Wisconsinan terminal moraine. The lake ponded behind this drift dam has been called Lake Hudson south of Orange and Putnam Counties and Lake Albany farther north (Connally and Sirkin, 1986; Stanford and Harper, 1991). This lake expanded northward during deglaciation and eventually extended nearly to the Canadian border (Stewart and MacClintock, 1969; Denny, 1974; DeSimone and LaFleur, 1986), although its existence within the Champlain lowland north of Washington County, N.Y. (pl. 1) was brief and limited to narrow ice-marginal lakes along the edges of the Adirondack and Green Mountains. The level of Lake Albany seems to have remained stable for several thousand years, which is most easily explained if the lake spilled across bedrock rather than across the drift dam; spillways at Sparkill, N.Y. (Dineen, 1986b), and at Hell Gate in New York City (Stanford and Harper, 1991) have been proposed. After the ice margin had retreated north of Albany, N.Y., however, lake level apparently began to decline gradually and also declined in several abrupt steps that have been attributed to erosion of the drift dam by catastrophic discharges when the proglacial Great Lakes were diverted into the Hudson River valley (Hanson, 1977; LaFleur, 1979a; DeSimone and LaFleur, 1985, 1986).

South of Kingston, N.Y. (in Ulster County, pl. 1), Lake Albany was narrow, barely wider than the modern Hudson River estuary. Coarse-sandy ice-contact deltas can be seen in a few places along the sides of the estuary (Connally and Sirkin, 1986). Several of these are perched on bedrock well above river level (Secor and others, 1955). Subglacial meltwater was probably concentrated along the bedrock trench; basal gravel deposits are reported in geologic sections at every bridge and aqueduct crossing north of New York City (Newman and others, 1969), although their water-yielding properties have not been tested. Once the moraine that dammed Lake Albany was breached, the Hudson River incised its valley to the prevailing subnormal sea level. Subsequently, as sea level rose, more than 100 feet of organic-rich silt was deposited all along the estuary south of Kingston (Newman and others, 1969).

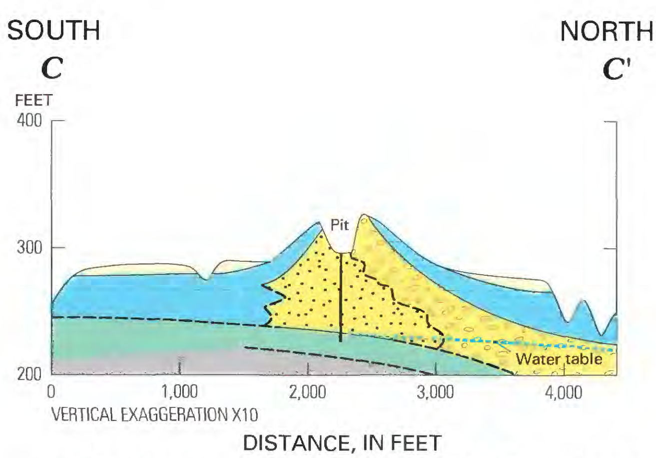
North of Kingston, low relief and isostatic depression allowed Lake Albany to inundate an area far wider than the modern estuary. All subsequent discussion of the Hudson lowland refers only to the reach north of Kingston where deposits in Lake Albany are areally extensive. In this reach, the ice retreated mainly through progressive stagnation (Dineen, 1986a), and detached ice blocks were preserved for long periods (Dineen and Rogers, 1979; LaFleur, 1979; Dineen, 1986a, p. 107; DeSimone and LaFleur, 1986). Maximum lake depth was as great as 300 feet in bedrock valleys but much less elsewhere. Several local ice readvances or surges have been inferred (Connally and Sirkin, 1971; Dineen and Hanson, 1983; DeSimone, 1985; Dineen, 1986a), but stratigraphic evidence is sparse, and some readvances have been disputed (DeSimone and LaFleur, 1986). The stratigraphy north of Kingston has been summarized by DeSimone and LaFleur (1986, p. 219) as follows: "A basal facies of interbedded gravel and sand with turbidites and minor silty diamicton (flowtill) grades and fines upward to clay and silt rhythmites of the middle facies. These rhythmites in turn grade and coarsen up to the upper facies of silt or, more commonly, sand deposited in a shoaling lake." The upper and lower facies contain aquifers and are described in greater detail below. Examples of stratigraphic relationships are presented in figure 61.

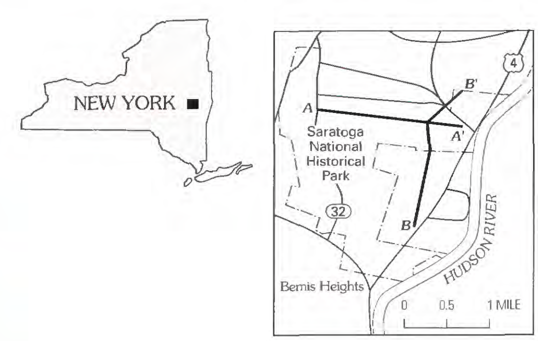


A. **Saratoga National Historical Park, Stillwater, N.Y. (From Heath and others, 1963, fig. III-6)**



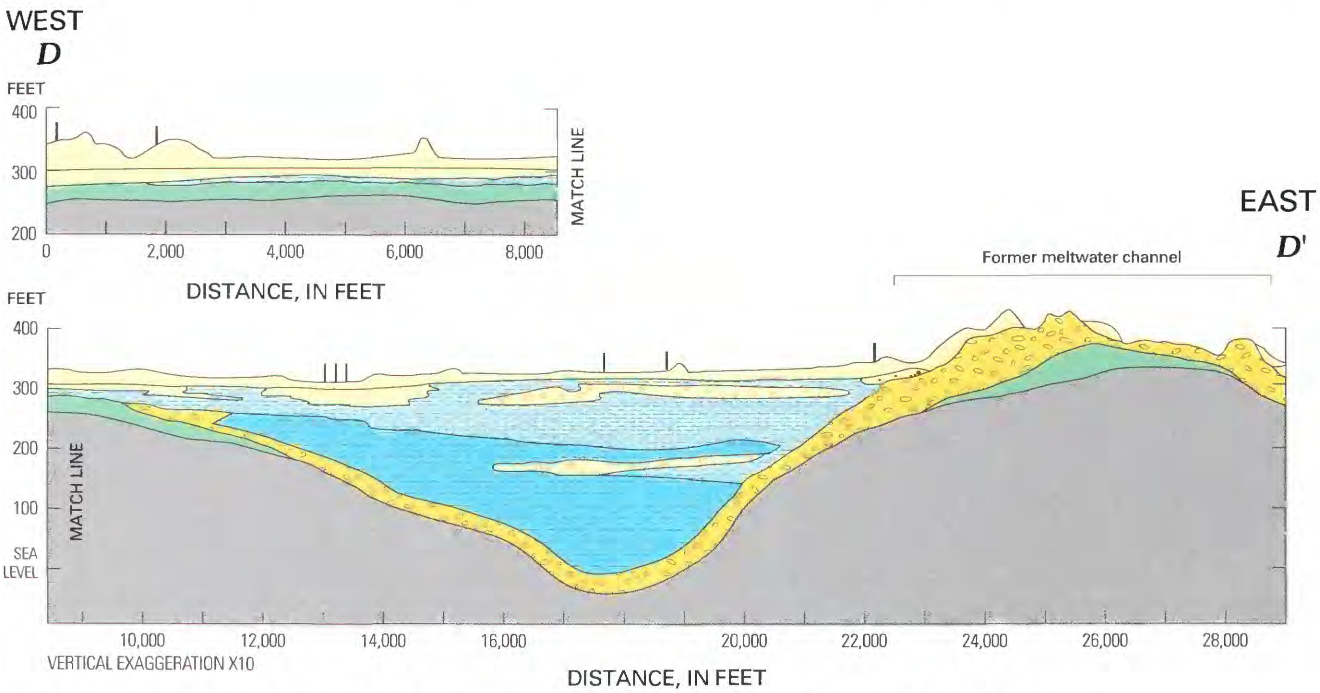
B. **Saratoga National Historical Park, Stillwater, N.Y. (From Heath and others, 1963, fig. III-6)**



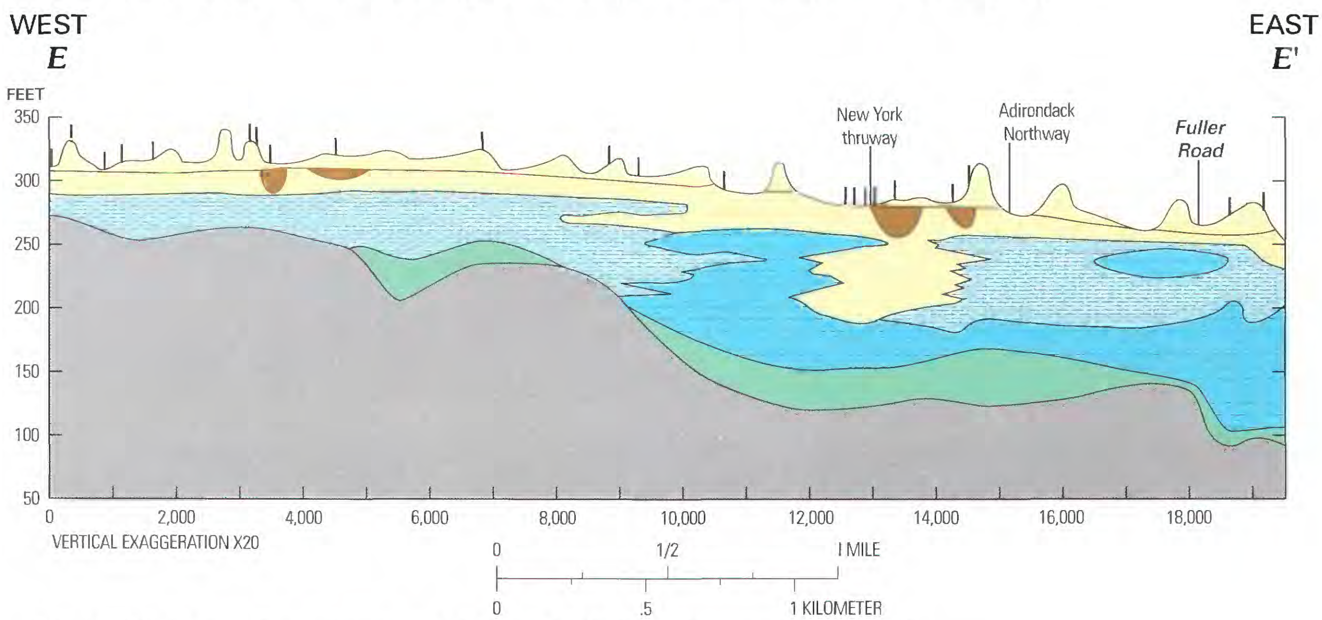


C. **Across west end of kame delta in Halfmoon, N.Y.**

FIGURE 61.—Geologic sections representative of glacial-lake deposits in the Hudson River lowland near and north of Albany, N.Y. Locations of sections *C–E* shown in figure 62.



D. **Colonie, N.Y., just north of Albany. (Modified from Dineen and Hanson, 1983, pl. 4)**



E. **Pine Bush area in western part of Albany, N.Y. (From Dineen, 1982, fig. 4)**

EXPLANATION

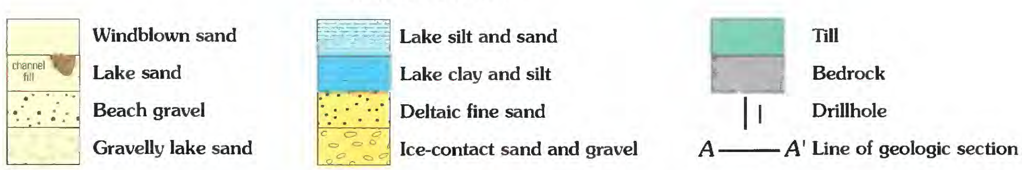


FIGURE 61.—Geologic sections representative of glacial-lake deposits in the Hudson River lowland near and north of Albany, N.Y. Locations of sections *C–E* shown in figure 62—Continued.

BASAL OR PROXIMAL ICE-CONTACT DEPOSITS (FACIES 1)

Ice-contact deposits vary widely in thickness from place to place. The thicker deposits include kame deltas, kame terraces, and ice-channel fillings that generally stand slightly higher than surrounding deposits (fig. 61*C*, *D*) and subaquatic fans that are buried by fines (fig. 61*B*; Dineen, 1986a, p. 98, p. 102; DeSimone and LaFleur, 1986). More commonly, lake fines are underlain by only a thin blanket of coarse stratified drift draped over preexisting topography (Dineen and Rogers, 1979, p. 93), and in many localities, this facies is absent (fig. 61*A–E*) or too thin and silty to yield water. The thickest ice contact deposits tend to be preferentially located along ice margins, meltwater channels, or (less likely) bedrock valleys, as discussed below.

Bedrock Valleys.—The bedrock surface in the Hudson lowland in Albany and Saratoga Counties is incised by several valleys whose configurations have been revealed by borehole and seismic records (Simpson, 1949; Stearns and Wheler, 1968; Dineen and Hanson, 1983). Such valleys are often surmised to be promising sites for exploration for large water supplies (Simpson, 1949, p. 719; Giese and Hobba, 1970, p. 44), and any thick and permeable ice-contact deposits that are found within bedrock valleys do have potential for large yields because available drawdown is greater than elsewhere. Nevertheless, test drilling and compilation of subsurface records by Stearns and Wheler (1968) and Dineen and Hanson (1983) have shown that the distribution of ice-contact deposits within the Hudson lowland does not necessarily coincide with bedrock valleys; many valley reaches contain fines but no coarse basal aquifer.

Ice Margins.—Several successive positions of the margin of the retreating ice sheet during deglaciation of the Hudson lowland have been outlined by LaFleur (1979a), DeSimone and LaFleur (1985, 1986), and Dineen (1986a). These ice-margin outlines were drawn so as to link the proximal ends of recognized ice-contact deposits or morphosequences and were inferred to represent positions at which the retreat of the ice front slowed or halted temporarily. The proximal ends of the individual deposits are known or assumed to be coarse grained and permeable. None of the papers cited, however, claimed or presented evidence that subaquatic fans, deltas, or other coarse-grained sediments were deposited contiguously in significant thickness all along the inferred ice margin.

Meltwater Channels.—LaFleur (1979a) depicted the locations of meltwater drainage lines as well as ice margins and ice-contact deposits in part of the Hudson lowland near Albany. Dineen and Hanson (1983) recompiled and extended LaFleur's interpretations; a simplified version of their map is presented as figure 62. Although a few ice-contact deposits seem randomly distributed, most fall in one of three spatially distinct groups:

1. Many ice-contact deposits are near the southwest margin of the lowland in Albany County (fig. 62). They were deposited chiefly by water draining southeastward from the Mohawk River valley (LaFleur, 1979a, fig. 2; Dineen and Hanson, 1983, p. 18).
2. A set of three kame deltas, aligned northwesterly from the confluence of the Mohawk and Hudson Rivers (fig. 62), spans a distance of 11 miles. The alignment suggests deposition along a single subglacial meltwater channel near successive positions of the ice margin.
3. A much longer set of aligned ice-contact deposits extends from the Rensselaer-Columbia County line, 5 miles east of the Hudson River, northwestward some 33 miles (fig. 62). North of the Mohawk River, these deposits follow a prominent bedrock valley; south of the Mohawk, they cross a low ridge and another bedrock valley, and farther south form a kame terrace along the east side of Lake Albany. Dineen and Hanson (1983, pl. 2) contoured the total thickness of ice-contact deposits in southern Saratoga and northern Albany Counties; the contours depict a zone 6,000 to 12,000 feet wide in which the aligned ice-contact deposits are generally 25 to 150 feet thick but locally absent; west of this zone, basal coarse deposits are absent or less than 25 feet thick. Reynolds (1985, pl. 4) contoured saturated thickness of the same deposits in the same area of Saratoga County; the two maps differ in detail partly because different sets of subsurface records were used. Both maps suggest, however, that the bulk of the buried ice-contact deposits is in the form of knolls spaced several thousand feet apart, connected at least by a thin sheet of silty sand or gravel and perhaps also by nearly continuous buried ridges no more than a few thousand feet wide.

DeSimone and LaFleur (1986) mention two subaquatic fans (exposed but mantled by lake clay) and an esker-fed delta in Argyle valley east of the Hudson River several miles north of the area shown in figure 62. These features are only 1 or 2 miles apart and suggest the presence of persistent subglacial meltwater flow along that valley.

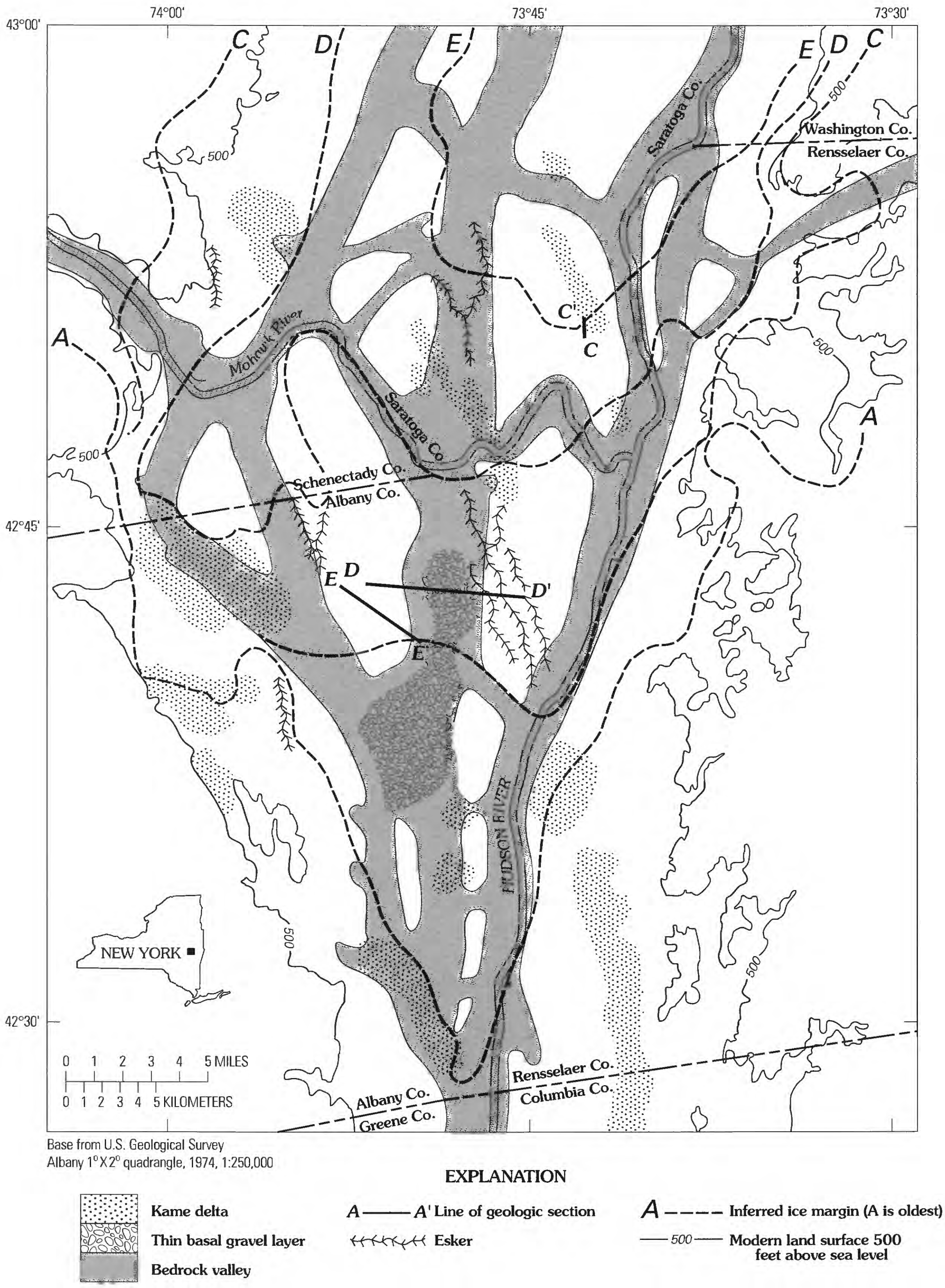


FIGURE 62.—Distribution of ice-contact sand and gravel (facies 1) in relation to bedrock valleys and ice margins in the Hudson River lowland near Albany, N.Y. (Modified from Dineen and Hanson, 1983, fig. 6.)

SURFICIAL SAND-PLAIN AQUIFERS (FACIES 3)

The silt and clay laid down in glacial Lake Albany are capped in many places by surficial sand that generally constitutes a shallow aquifer. Most of the sand was originally deposited in extensive inwash deltas fed by major tributaries. Lake currents, generated by meltwater flux, transported sand away from some deltas across the lake floor (DeSimone and LaFleur, 1986, p. 220). As the stage of Lake Albany declined in response to spillway erosion and perhaps isostatic rebound, streams and waves incised the earlier deltas and redistributed the sand widely across the shoaling lake. The surface of deltas exposed by declining lake level was reworked into dunes by prevailing winds, and some of the windblown sand was redeposited in the remaining lake as areas of especially well-sorted sand (Dineen and Rogers, 1979; Dineen and Hanson, 1983). This surficial lake-deposited sand layer of varied origin extends to depths as great as 100 feet in Schenectady and northwest of Glens Falls (Simpson, 1952, p. 32; Stearns and Wheler, 1968; Giese and Hobba, 1970) but is commonly about 60 feet thick near Albany, where it contains scattered clay lenses (Dineen, 1982), and averages about 25 feet thick in southern Saratoga County (Giese and Hobba, 1970). Where the tributaries entering Lake Albany were small or northward-draining, surficial sand deposits are localized or absent; this condition is typical of areas south of Catskill and northeast of Glens Falls (pl. 1).

Modern streams are generally incised through the surficial sand into the underlying clay or silt, except for headwater reaches of small streams (Heath and others, 1963; Snavely, 1983). The sand is typically very fine to medium grained and becomes finer with depth. Depth to water is commonly 10 to 15 feet (Heath and others, 1963; Snavely, 1983), and saturated thickness can be substantial. Large-yield wells are uncommon, but a great many domestic wells tap these sand-plain aquifers.

SCHENECTADY AQUIFER

Highly permeable sand and gravel deposits along the Mohawk River immediately west of Schenectady, N.Y., constitute one of the most productive aquifers in the glaciated Northeast. Transmissivity locally exceeds 1 million feet squared per day near the Schenectady well field (Winslow and others, 1965, p. 56). The Mohawk valley is slightly less than 1 mile wide in western Schenectady County and generally contains 50 to 100 feet of gravel, sandy gravel, or coarse sand beneath the valley floor; some of the gravel is in sand-free openwork layers. Where the Mohawk valley joins the Hudson lowland at Schenectady, the flood plain widens and is incised more than 100 feet below a broad sand plain typical of Lake Albany. Coarse sand and gravel beneath the flood plain grades downward and laterally into sand (Winslow and others, 1965, pl. 3) and grades southeastward rather abruptly into very fine sand and silt (fig. 63) that continues beneath the surrounding sand plain. The location and stratigraphy led Winslow and others (1965, p. 24, p. 36) to interpret the coarse sand and gravel as the headward part of a delta built into Lake Albany.

LaFleur (1979a) and Wall (1995) describe a more complex depositional history. Some coarse-grained inwash was deposited amid the last decaying ice in the Mohawk valley west of Schenectady. Runoff from the deglaciated Mohawk watershed later filled the Mohawk valley west of Schenectady and the Hudson lowland east of Schenectady with fine-grained sediment and deltaic sand graded to Lake Albany. Thereafter, channel incision in response to declining lake levels was intensified when runoff from the entire Great Lakes watershed was diverted down the Mohawk valley for several hundred years. Annual maximum discharges probably exceeded 1.5 million cubic feet per second (Wall, 1995); more than 40 feet of cobble gravel was deposited in western Schenectady County (LaFleur, 1979a), some of which is preserved beneath terraces that border the flood plain (Cadwell and Dineen, 1987); a huge expansion bar or delta of gravel and coarse sand occupies much of the valley at Scotia (Wall, 1995).

The distribution of the coarse sand and gravel that constitutes the Schenectady aquifer suggests that most of the aquifer is not a product of the late-deglacial torrents, but rather was deposited during ice-contact and deltaic phases that are typical of the margins of lake-filled lowlands near large tributaries. That is, the aquifer extends to depths of 50 to 100 feet below the modern Mohawk River, much deeper than the late-deglacial torrents could erode at a site upstream (Wall, 1995, p. 76). The flood plain and fluvial terraces that postdate Lake Albany overlie aquifer materials near the edge of the Hudson lowland, but overlie lacustrine fines in Schenectady (fig. 63) and apparently downstream (Winslow and others, 1965, pl. 1), which also suggests that when the late-deglacial torrents incised the Lake Albany sand plain, they did not erode and rework older sediments to depths appreciably below the modern river channel.

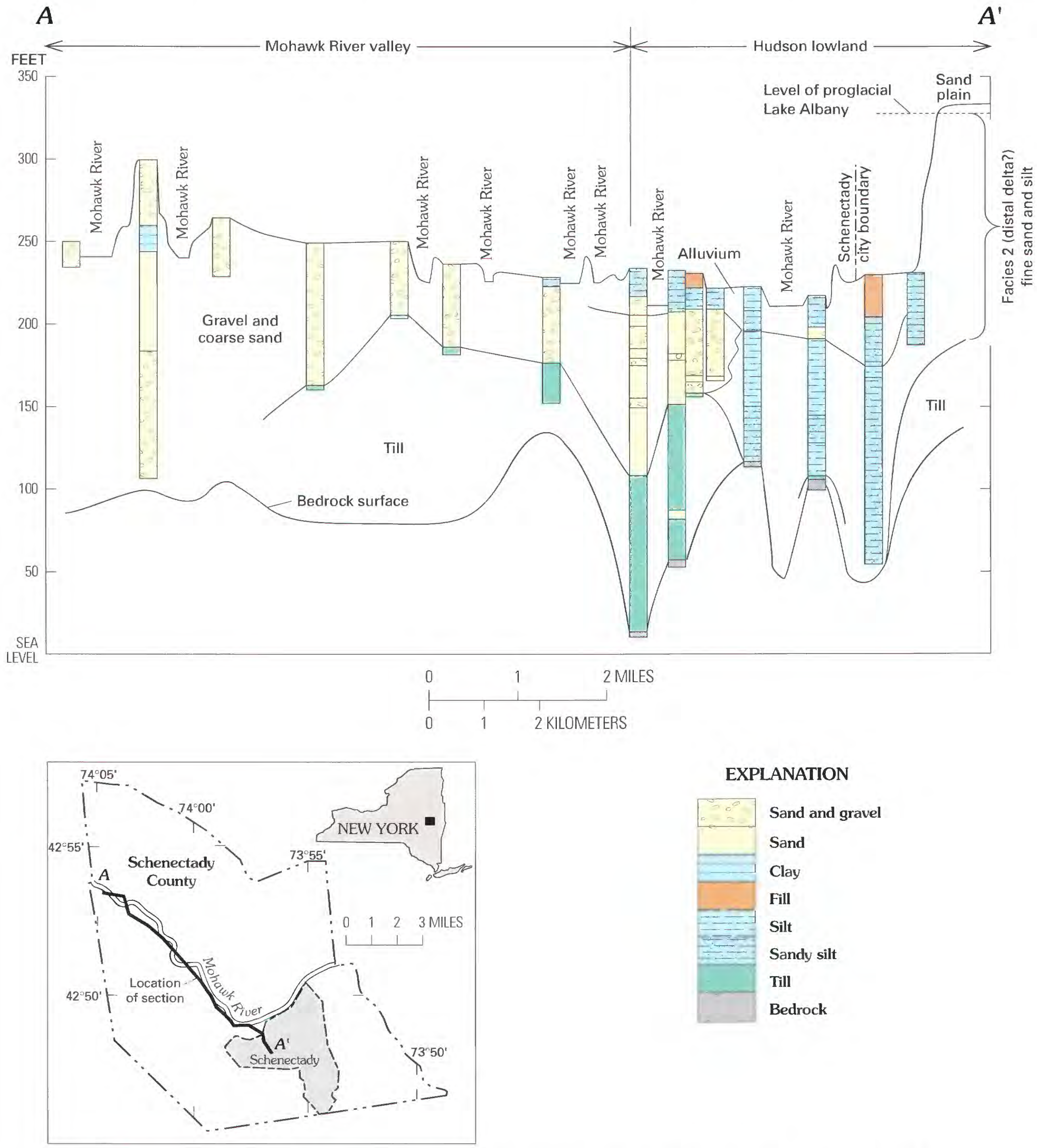


FIGURE 63.—Geologic section along the Mohawk River in Schenectady County west of Schenectady, N.Y. (Modified from Winslow and others, 1965, fig. 8, and Brown and others, 1981, sheet 3.)

ERIE AND ONTARIO LOWLANDS

Directly south of Lakes Erie and Ontario is an area of generally low relief that is underlain chiefly by shale and carbonate bedrock. Nearly all of this area was inundated at one time or another during deglaciation by broad proglacial lakes ponded between the Appalachian Plateau and the retreating ice sheet, as indicated by map symbol "L" on plate 1.

LAKE HISTORY AND EXTENT

A complex succession of proglacial lakes bordered the ice sheet in the Lake Erie and Lake Ontario basins; the level of ponding changed often as lake outlets to the east and west were blocked or unblocked by the fluctuating ice margin, or as isostatic rebound changed the relative altitude of alternative outlets. Many studies have contributed to the syntheses of lake history presented by Calkin (1970), Fullerton (1980), Muller and Prest (1985), Calkin and Feenstra (1985), and Muller and others (1988). A succession of westward-draining lakes (Lakes Whittlesey and Warren, among others) inundated the area west of Batavia, N.Y., below altitudes of at least 850 feet in New York and 765 feet in Ohio; the successive shorelines are now generally marked by sandy beach ridges (see fig. 67). Several short-lived local lakes drained east and west across spurs between the major valleys along the margin of the Appalachian Plateau from Batavia east to Syracuse, N.Y., carving or reoccupying spectacular channels in many places (Wilson, 1981; Muller and others, 1988). The final stage was Lake Iroquois, which drained east to the Mohawk River valley and inundated much of the lowland south of Lake Ontario; its former shoreline now rises in altitude from 360 feet near Niagara Falls to 435 feet near Rochester, 490 feet near Oneida Lake, 550 feet near Oswego, and more than 700 feet near Watertown (Calkin, Muller, and Drexhage, 1982; Miller, 1982; Kappel and Young, 1989).

Till and lake-bottom fine sand, silt, and clay are the predominant surficial sediments in the Erie and Ontario lowlands. The extensive proglacial lakes that inundated these lowlands left correspondingly extensive mappable deposits of lake-bottom fines. Other areas are mapped as till, however, including some areas that lie below former lake levels (LaSala, 1968; Muller, 1977; Muller and Cadwell, 1986; Kammerer and Hobba, 1986; Miller, 1988a, b); this fact, and the considerable disparity among some of these maps, suggest that the surficial lake-bottom fines are thin in many places. The surficial till is not necessarily older than all of the lake-bottom fines, however. Two episodes of substantial ice retreat and readvance have been inferred, one about 13,300 radiocarbon years ago that predated Lake Whittlesey (Muller, 1977; Fullerton, 1980; Calkin and Feenstra, 1985), and another about 12,000 years ago that drained and later reinstated Lake Warren (Fullerton, 1980; Muller and Prest, 1985, p. 218–9; Calkin and Feenstra, 1985; Muller and others, 1988; Muller, 1988b); subsequent minor readvances just south of Lake Ontario also have been inferred (Muller and others, 1988, p. 127; Young, 1988, p. 75). Such events could be expected to result in till layers interbedded with lake-bottom fines. Drift exposed in bluffs along Lakes Erie and Ontario as far east as Rochester generally consists of 10 to 30 feet of silt that is locally interbedded with clay or fine sand, overlying a comparable thickness of till (Calkin, Muller, and Drexhage, 1982; Calkin and Muller, 1992). Along Lake Ontario, however, the till layer is made up of two lodgment tills, the upper of which grades up into subaquatic resedimented diamicton that contains many lenses and wisps of sand, silt, and clay, and a few feet of silt or clay are preserved locally below each till. Along Lake Erie, till is exposed over varved clay for 20 miles (Calkin, 1982, p. 130). Near Lake Ontario and the Niagara River, many feet of stratified water-laid diamicton and (or) compact basal till interrupt the silt, clay, and minor fine sand that constitute the bulk of the drift (Smith and Calkin, 1990).

DISTRIBUTION OF SAND AND GRAVEL

The Erie and Ontario lowlands consist of two terranes that differ with respect to distribution of sand and gravel. One is a lake plain of very low relief that rises gently from Lake Erie to the foot of the Appalachian Plateau and from Lake Ontario to the broad crest of the Niagara cuesta, which parallels the lake shore. The other lowland terrane lies between the Niagara cuesta and the Appalachian Plateau and is underlain by shale and dolostone of the Salina Group and Akron Formation. On the lake plain, which is 1 to 6 miles wide along Lake Erie and about 10 miles wide along Lake Ontario, coarse-grained stratified drift is largely confined to beach ridges deposited by waves and longshore currents along the former shores of proglacial lakes; proximal facies-1 sand and gravel deposits are few, small, and widely scattered. In the lowland south of the Niagara cuesta, coarse-grained stratified drift occurs chiefly as meltwater channel deposits that are aligned subparallel to the direction of ice flow. A few head within the lake plain. Some coarse stratified drift is associated with ice margins and aligned nearly perpendicular to the direction of ice flow, and a few upland tributaries or spillway channels built large deltas where they entered proglacial lakes. The several landform types that contain sand and gravel—including meltwater channel deposits, kame moraines, interdrumlin outwash, perched deltas, beach ridges, and wave-built terraces—differ in aquifer potential and are considered in turn in the next several sections.

MELTWATER CHANNEL DEPOSITS

Many surficial and buried sand and gravel deposits in the Erie and Ontario lowlands have distributions, dimensions, or landforms that suggest deposition in or from subglacial meltwater channels. Although they tend to follow bedrock valleys, in this area of low relief they commonly form prominent topographic features. Examples in several localities are described in this section.

Oswego County, N.Y.—Multiple elongated kames and ice-channel fillings aligned in narrow bands constitute the principal stratified-drift aquifers in central and southern Oswego County (Miller, 1982). Their linear distribution (fig. 64) suggests that meltwater flow was largely concentrated in channels, which were spaced 4 to 8 miles apart near the southern border of the county. Several of these aligned deposits follow bedrock valleys at least in part (Miller, 1982; T.S. Miller, U.S. Geological Survey, oral commun. 1986), although bedrock valleys are shallow and not easily recognized in this region of low relief. Some small kame deposits shown on surficial geology maps (for example, Miller, 1980b; Muller and Miller, 1980) are not obviously linked to those in figure 64, but might nevertheless have been part of a northward-branching network. Several of the aligned channel deposits head within the lake plain and climb the gentle slope southward to or beyond the divide, but few are found in the northern part of the lake plain (fig. 64). This pattern might reflect greater buoyancy and more rapid retreat of the ice margin in deeper water. Lake Iroquois, which spread northward across Oswego County as the ice retreated, must have been less than 150 feet deep in most places, but its depth increased northward on the lake plain and exceeded 250 feet near the present lakeshore.

Sand Ridge Aquifer.—The Sand Ridge aquifer is the largest of the meltwater channel deposits in Oswego County, N.Y. (fig. 64). It extends 13 miles from the southern part of the Lake Ontario plain across a low divide (the Niagara cuesta is imperceptible in Oswego County) and down a gentle slope to the Oneida River (Miller and Sherwood, 1993). Its south-southeasterly alignment is nearly parallel to the direction of ice flow, as recorded by numerous drumlins. The southern part of the aquifer is generally a broad-crested ridge, composed predominantly of coarse sand with some gravel but mantled or overlapped by fine sand and by silt and clay (fig. 65), all of which constitutes a fining-upward sequence deposited in Lake Iroquois. The highly variable thickness of the sand and gravel (fig. 65) and the many kettleholes that partly interrupt the ridgelike landform (Miller and Sherwood, 1993, pl. 1) reflect deposition against and over stagnant ice. The northern part of the aquifer is a network of multiple narrow constructional landforms (kame terraces or kame plains banked against or around drumlins, also ice-channel fillings and aligned kames). These landforms consist of gravel and sand and are separated by large ice-block depressions underlain by peat and lake fines. The aquifer approximately follows a shallow bedrock valley and interdrumlin lowlands but is thick enough over most of its length to constitute a prominent topographic ridge that is exceeded in height by only a few drumlin crests. The ridge continues another 2 miles south into Onondaga County (fig. 64) where it may connect to buried aquifers identified by Kantrowitz (1970, pl. 1) and Miller (1988b) that extend south at least to Onondaga Lake and perhaps farther down the Onondaga Creek trough within the Appalachian Plateau. The aquifer is recharged solely by precipitation on its surface under natural (nonpumping) conditions; induced recharge would be possible only where the aquifer is crossed by the Oneida River (fig. 64). This aquifer seems to constitute a textbook example of deposition from a subglacial channel system that continued to contribute sediment as the ice margin retreated. Some of the coarse sediment was deposited in deep water in subaquatic fans that fined upward as the ice margin retreated, and the remainder in shallow water in tunnels or crevasses within the ice.

Irondequoit Bay, Eastern Monroe County, N.Y.—Stratified drift is more abundant from Irondequoit Bay southward to western Ontario County than elsewhere in the Ontario lowland. The surficial stratified drift is largely fine sand, silt, and clay and locally includes ablation till (Crain, 1974, p. 65; Waller and others, 1982, sheet 1; Muller and Cadwell, 1986), but coarse sand and gravel commonly underlie these fine sediments, are present locally at land surface, and have been inferred to constitute a continuous productive aquifer oriented generally southward from Irondequoit Bay (Crain, 1974; Waller and others, 1982; Miller, 1988a; Kappel and Young, 1989). The large volume of stratified drift here, in contrast to that in adjacent lowland areas, and the fining-upward stratigraphy suggest subaquatic-fan deposition throughout the deglaciation of the lowland from a subglacial channel system that followed the lowest available route across the lake plain, namely the preglacial valley of the Genesee River (Leggette and others, 1935; Young, 1980; Waller and others, 1982). Large amounts of sediment were delivered to this locality over the timespan represented by several successive proglacial lakes, as

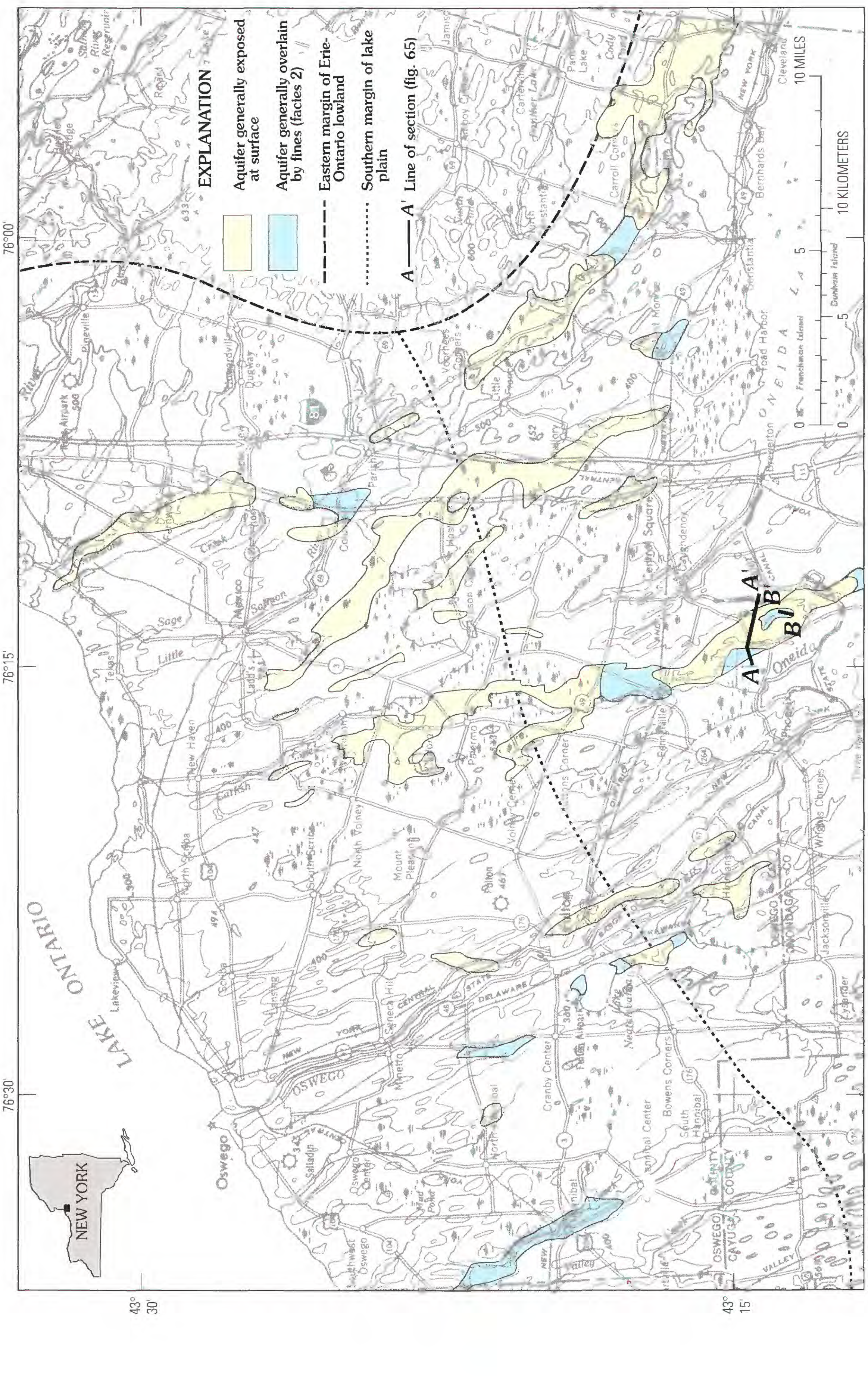
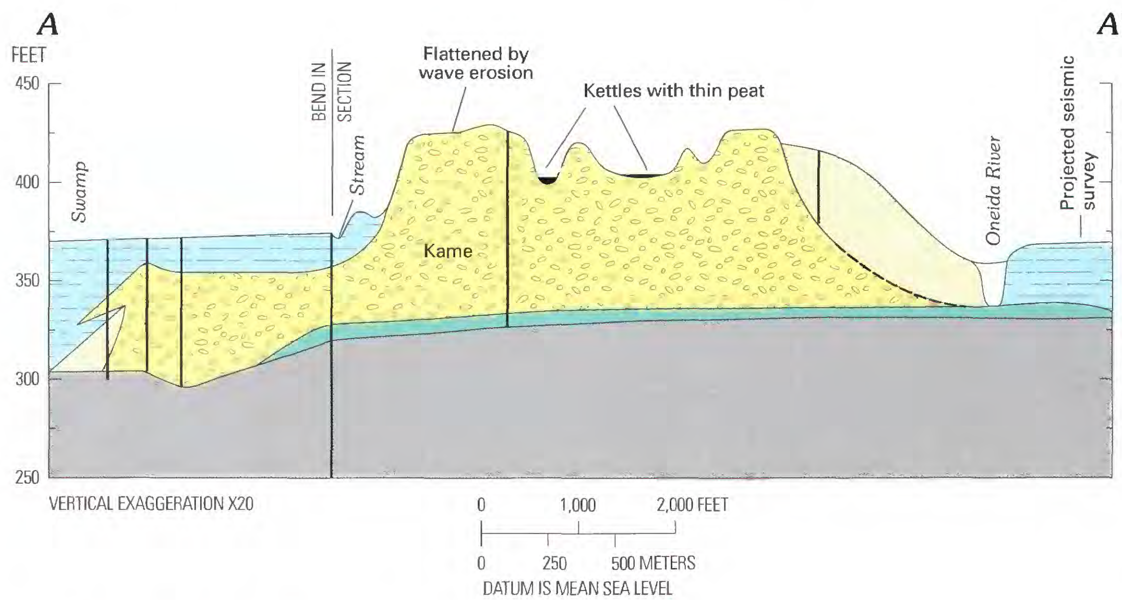
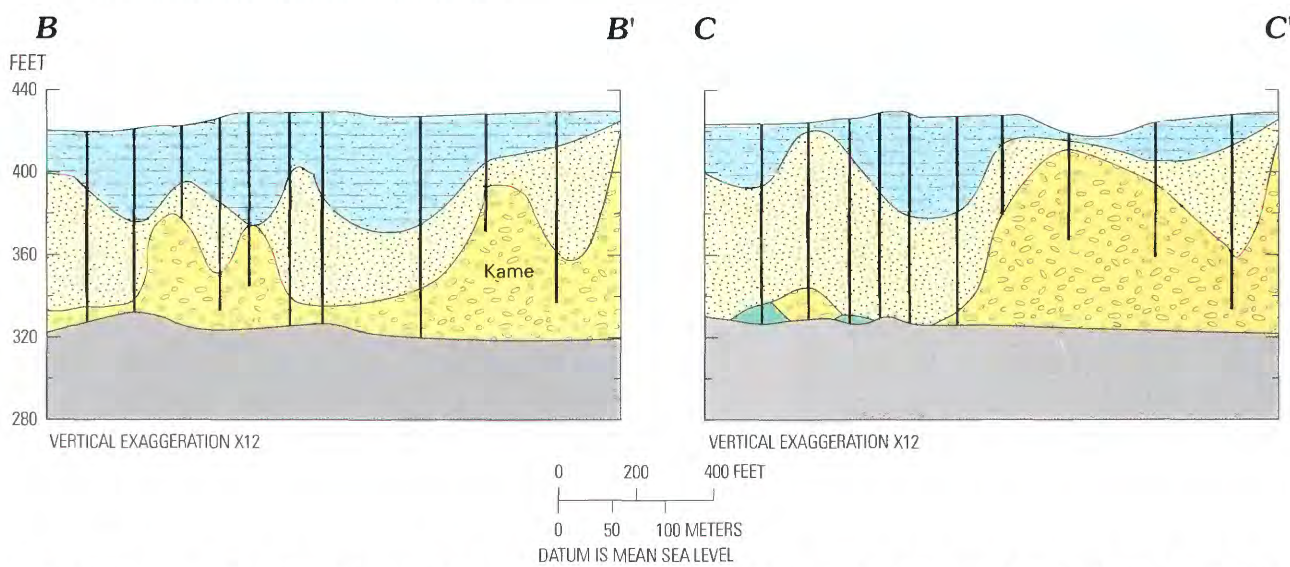


FIGURE 64.—Locations of potential facies-1 sand and gravel aquifers in the Ontario lowland in Oswego County, N.Y. (Modified from Miller, 1982, 1987, 1988a; Miller and Sherwood, 1993; and reconnaissance maps of surficial geology such as Muller and Miller, 1980. Some small, isolated kames are not shown.)



A. **Section across the entire aquifer**



B, C. **Two sections along the aquifer 1 mile south of section *A*. Section *C-C'* is parallel to *B-B'* and only 200 feet distant. These sections show that the thickness of the buried sand and gravel can be highly variable over short distances**

EXPLANATION

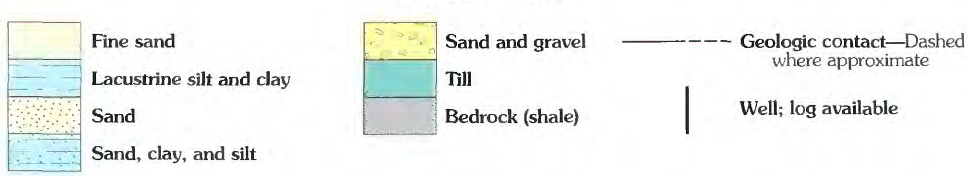


FIGURE 65.—Geologic sections near the south end of the Sand Ridge aquifer in southern Oswego County, N.Y. (From Miller and Sherwood, 1993.) Locations of sections A–A' and B–B' are indicated in figure 64.

demonstrated by several large constructional landforms with depositional surfaces that range from 1,040 to 575 feet in altitude. For example, an 870-foot outwash deposit southwest of Victor hamlet was ascribed to Lake Warren (Kappel and Young, 1989, p. 4), and lower surfaces nearby were deposited in later lakes that drained eastward through successively lower spillways near Victor. Abundant sediment delivery to the preglacial Genesee valley persisted through ice-margin oscillations, as demonstrated by the presence within the valley of a basal aquifer mantled by till south of a buried till moraine, as well as a younger basal aquifer unmantled by till north of the moraine, and by an esker segment that apparently overlies lacustrine fines (Young, 1988; Kappel and Young, 1989).

Other Localities.—Cayuga and Seneca Lakes (pl. 1) occupy two of the longest and deepest valleys on the north rim of the Appalachian Plateau. Crain (1974) and Miller (1988a) inferred that sand and gravel aquifers extend from the north ends of these lakes northward to the Niagara cuesta, generally buried beneath fine-grained sediments of Lake Iroquois but locally exposed at land surface. The continuity, northward branching, and generally basal position inferred for these aquifers are consistent with a subglacial meltwater channel system that tended to follow bedrock valleys.

Near Albion, N.Y., is the Burma Woods esker, a narrow gravel ridge that extends for at least 4 miles in a southwesterly direction between the Albion and Barre moraines. Its southern end is within a complex of constructional landforms, probably kame deltas, where it is bordered by fine sand and silt (Calkin, Muller, and Drexhage, 1982, p. 321). Muller (1977) mapped several elongated ice-contact deposits, some identified as eskers, all descending the gentle south slope of the Niagara cuesta and all standing well above the surrounding till or lake fines. Most of these features were not identified as aquifers by Miller (1988b) because their saturated thickness was thought to be small.

Small, thin deposits of facies-1 sand or gravel that seem not to be associated with persistent major meltwater channels are exposed in several localities in the Erie and Ontario lowlands. Coarse sediments that range from sandy gravel to gravelly silt and flow till are exposed at intervals of 2 to 3 miles along the Lake Ontario bluffs west of Rochester. They lack topographic expression, are commonly separated from the underlying till by fine-grained sediments and (or) an erosional unconformity, and are exposed about a mile north of a minor moraine; hence, they are inferred to be localized deposits of ephemeral meltwater channels along an oscillating ice margin (Calkin, Muller, and Drexhage, 1982, p. 291, p. 295). Coarse sediments that grade upward to medium-fine sand and might be the product of discharge from a subglacial channel were exposed in 1982 at Brighton Cliff (fig. 66); nearby deformation suggests the presence of active ice (Calkin, Muller, and Drexhage, 1982, stop 7). No such deposits have been identified as significant aquifers or shown to extend as buried linear features over thousands of feet or more. In one 4-mile reach along the Lake Erie bluffs, however, surficial lake silts are underlain persistently by facies-1 sandy gravel, which is as much as 10 feet thick (Calkin and Muller, 1992). Surficial silt and clay also grade down to thin basal sand in places near Tonawanda Creek northeast of Buffalo (Smith, 1990), and buried aquifers have been recognized in a few localities (Miller, 1988b), but the dimensions of these scattered aquifers are poorly known.

KAME MORAINES

Several end moraines record ice-marginal positions within the Erie and Ontario lowlands in New York, mainly west of Rochester (Calkin, 1970, 1982; Muller, 1977; Muller and Prest, 1985). Most consist chiefly of till, but two have been described as kame moraines—chiefly sand and gravel, in part capped by till or interlayered with till. The Pinnacle Hills kame moraine at Rochester was described in detail by Fairchild (1923) and depicted on a map by Muller and Cadwell (1986); it seems to be largely unsaturated (Miller, 1988a). Fairchild (1923, p. 167) wrote "most or all of the glacial drainage . . . issued from tunnels beneath the ice sheet," although no gravel deposits in tunnels have been identified on the north side of the Pinnacle Hills moraine. The Buffalo kame moraine, which extends from Buffalo to Batavia, was described by Calkin (1970, p. 86) as largely ice-contact stratified drift; several constructional landforms associated with it were mapped as ice-contact deposits or outwash by LaSala (1968) and Muller (1977). One of the largest of these occupies and borders a shallow bedrock valley south of a notch in the Onondaga escarpment at Clarence (Miller and Staubitz, 1985) where considerable meltwater probably flowed southward through the lowest avenue locally available.

INTERDRUMLIN OUTWASH

A drumlin field about 60 square miles in area along the Wayne-Ontario County line west of Newark, N.Y., lies within the Ontario lowland but above the level of Lake Iroquois and also above its immediate predecessors, which drained through channels incised across

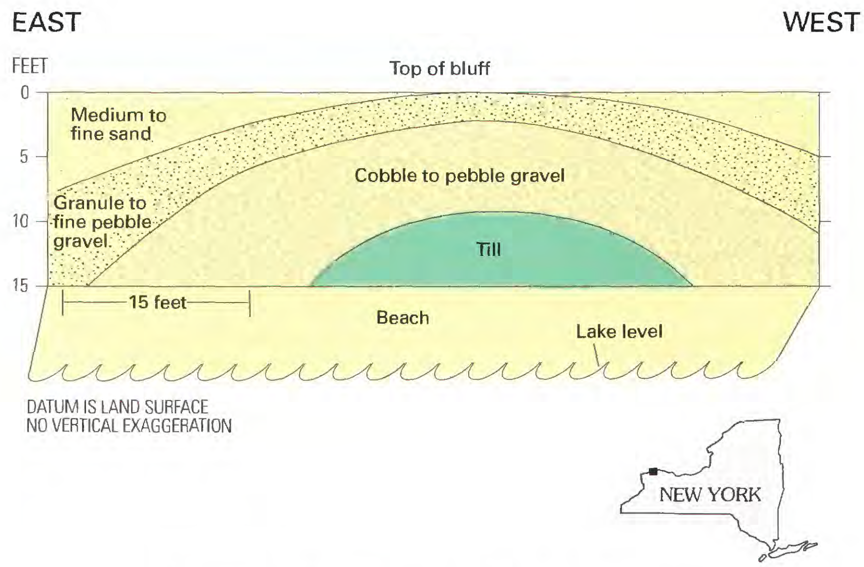


FIGURE 66.—Exposure in Lake Ontario bluff between Point Breeze and Brighton Cliff, N.Y.

the drumlin field. Apparently the area was still covered by ice during the life of Lake Warren, which was much higher than the later lakes (Muller and Prest, 1985, p. 220). During deglaciation, meltwater streams flowed southward and ultimately eastward between and around the closely spaced drumlins, depositing sand and gravel around each drumlin or cluster of drumlins. The resulting meshlike network of coarse stratified drift covers the entire width of the interdrumlin valleys, mantles any fine-grained sediment that might have been deposited, and includes several successive, topographically distinct morphosequences. These characteristics are unlike the rest of the Ontario lowland but are typical of parts of the Eastern Hills and Valley Fills region, to which this drumlin-field area is assigned on plate 1. The coarse stratified drift has been interpreted as commonly having less than 10 feet of saturated thickness (Crain, 1974, pl. 2; Miller, 1988a). Nevertheless, it represents a substantial volume of coarse sediment. Comparable volumes of coarse sediment could presumably be expected in adjacent areas of the Ontario lowland, in channels beneath the widespread surficial fines, as inferred by Crain (1974) and Miller (1988a).

EROSIONAL CHANNELS AND PERCHED DELTAS

On several occasions, while the ice sheet was retreating from the Appalachian Plateau, proglacial lakes drained catastrophically from one valley or lowland to the next across bedrock spurs that slope northward from the plateau. These jokulhlaups created local scablands of bedrock scoured free of drift, incised several striking east-west channels, and deposited prominent deltas where the channels debouched into lakes. The earlier, higher channels are purely erosional features carved in bedrock or till, such as those west of the Genesee River valley near Caledonia (Muller, 1977; Wilson, 1981), near Pearl Creek (Muller, 1977; Muller and others, 1988), and east of Syracuse (Hand and Muller, 1972). Large deltas at the downstream ends of these channels seem to be generally perched on high bedrock above modern streams and have been interpreted as thinly or questionably saturated (Miller, 1988a, b). Later, lower channels extend eastward from Rush and Fairport, N.Y. These channels were interpreted by Crain (1974, p. 55, pl. 2) as predominantly erosional features that only locally contain stratified-drift aquifers more productive than the underlying bedrock. More recent maps (Muller and Cadwell, 1986; Miller, 1988a) indicate that these channels generally contain productive outwash aquifers.

A few moderately large streams that flow north or west out of the Appalachian Plateau built large deltas into Lakes Whittlesey and (or) Warren in southwestern New York (Symecko, 1967; LaSala, 1968, pl. 2; Calkin, 1970, p. 83). These deltas also are largely perched above stream grade but could be significant as aquifers locally.

BEACH RIDGES AND WAVE-BUILT TERRACES

The bulk of the surficial sand and gravel within the lake plain occurs as beach ridges, which are linear deposits built by waves and longshore currents along the margins of large proglacial lakes. Most are many miles long. They are commonly best developed near large tributaries whose bedload was an important source of sand and gravel (Miller, 1982, p. 12; Totten, 1985) and whose earlier deltas were easily reworked by waves (Calkin, 1970, p. 83, p. 90; Calkin, 1982, p. 132). A series of beach ridges parallel the shore of Lake Erie; some of those identified in Ohio (fig. 67) have not been found in New York (Calkin, 1970), presumably because areas north of Ohio were covered by ice at the time. A single beach ridge that marks the shore of glacial Lake Iroquois parallels the modern shore of Lake Ontario, except in Oswego and Cayuga Counties, N.Y., where Lake Iroquois extended southward far beyond the lake plain.

Beach ridges are only minor aquifers. They range in grain size from fine sand to pebble gravel. Most are 500 to 1,500 feet wide; some occur as multiple adjacent ridges or spits that total as much as a mile in width. Most are 10 to 15 feet thick, although a few are as much as 30 feet thick (Calkin, 1970; Chute, 1979; Totten, 1985). Their saturated thickness is small, generally

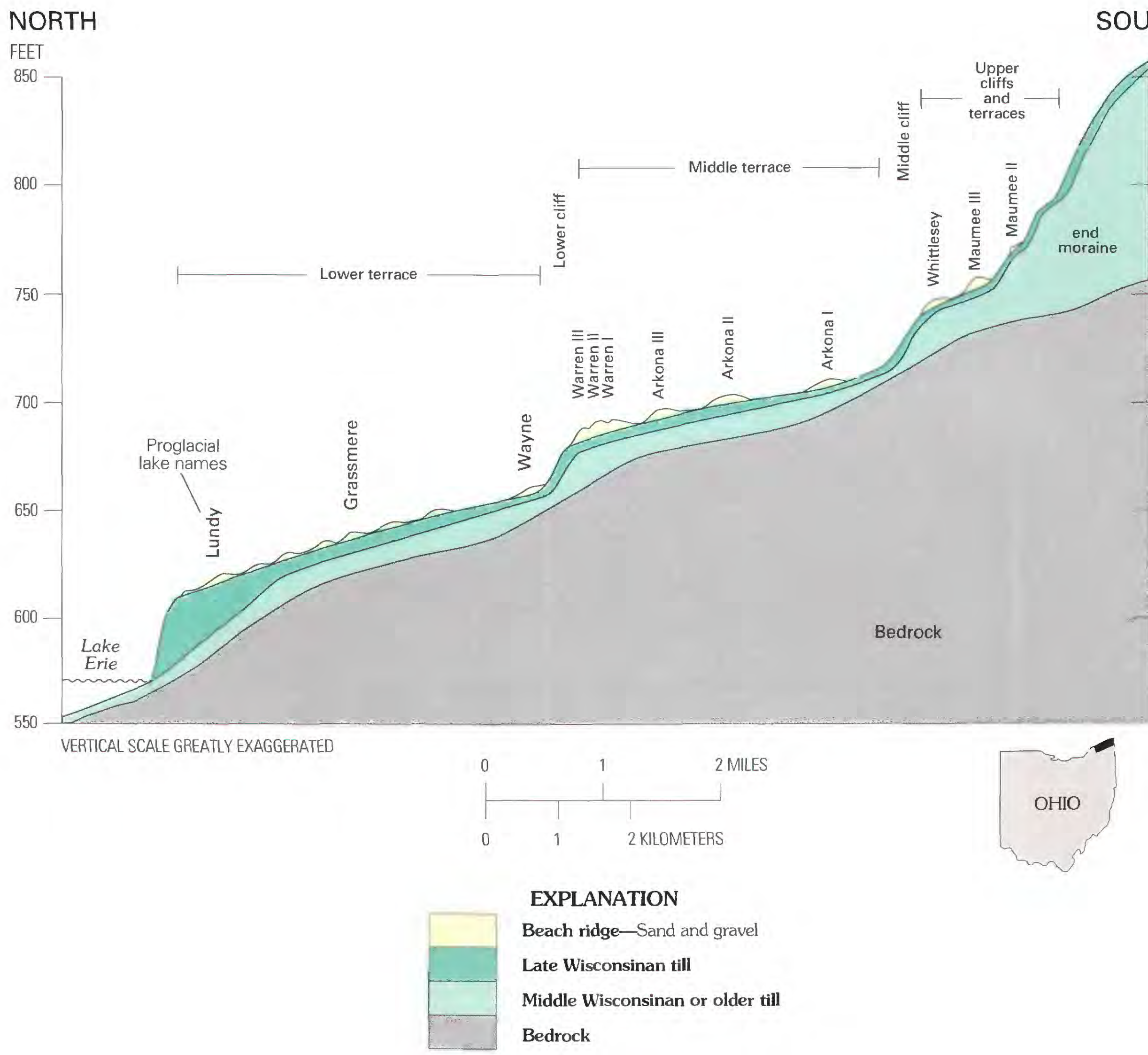


FIGURE 67.—Composite geologic section of strandlines south of Lake Erie in Lake and Ashtabula Counties, Ohio. Late Wisconsinan till mantles cliffs and terraces carved after the previous deglaciation and is capped by several beach ridges that formed along the shores of successive proglacial lakes during the late Wisconsinan deglaciation. (From Totten, 1985, fig. 2.)

only a few feet (fig. 68). Many beach ridges provide small water supplies from shallow wells and springs, however, and a few large ridges near the east end of Lake Ontario are capable of yielding a few hundred gallons per minute (Griswold, 1951, p. 16; Johnston, 1964, p. 45; Miller and others, 1988, p. 30; Zarriello, 1993).

Wave-cut platforms bevel many drumlins that were encircled or inundated by Lake Iroquois in Oswego and Cayuga Counties, N.Y. (Chute, 1979). Wave-built terraces underlain by angular gravel commonly rim the platforms and locally cap drumlins (fig. 68*D*). Saturated thickness of these terrace gravels is probably negligible.

BLACK RIVER, ST. LAWRENCE, AND CHAMPLAIN LOWLANDS

The Adirondack upland of New York is bordered on the north and east by broad lowlands along the St. Lawrence River and Lake Champlain, respectively, and on the west by a narrow lowland or valley along the Black River. As the ice sheet retreated, these lowlands were occupied by large lakes and later, in part, by marine embayments, whose collective extent is indicated by map symbols L, S, and D on plate 1. The lake history is reviewed below to introduce the subsequent descriptions of the distribution of sand and gravel in these lowlands.

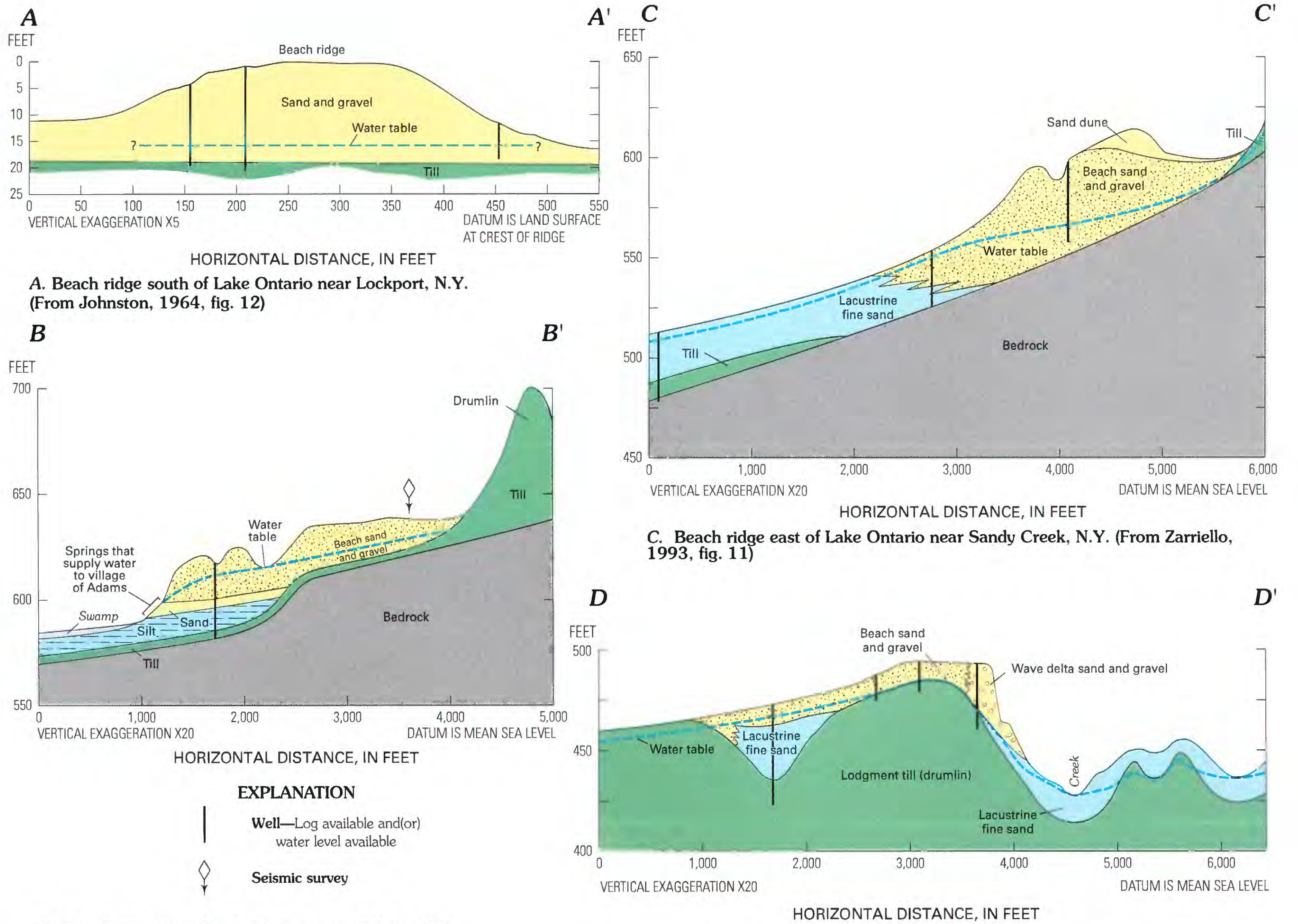
LAKE HISTORY AND EXTENT

Several successive lakes in the Black River valley drained southward to the Mohawk River through different outlets (Ridge and others, 1984). Meanwhile, the Ontario lowland farther west was occupied by Lake Iroquois, which also drained down the Mohawk valley and expanded northward as the ice margin retreated. When the retreating ice margin reached Watertown, N.Y., lake level in the Black River valley dropped to that of Lake Iroquois, which continued to expand northward along the southern margin of the St. Lawrence lowland (Clark, 1984; Clark and Karrow, 1984; Pair and Rodrigues, 1993). Shoreline elevations that imply lake depths of 300 to 400 feet, and abundant dropstones and drop pellets in the lowest 3 feet of Lake Iroquois rhythmites, together suggest a floating ice margin that retreated rapidly by calving (Pair and Rodrigues, 1993).

The Champlain lowland is a northward continuation of the Hudson River lowland in New York. Both lowlands are carved in carbonate bedrock and shale of Cambrian and Ordovician age, but the Champlain lowland is somewhat wider. As the ice melted away from the Adirondack and Green Mountains, which flank the Champlain lowland on the west and east, respectively, local proglacial lakes in mountain valleys drained, and a regional lake expanded northward from the Hudson lowland between the mountains and the shrinking ice tongue. Lake level in the Hudson lowland eventually declined below the level of a broad saddle near Fort Ann, N.Y., which is the lowest point on the modern divide. Thereafter, channels that were cut into the saddle by the southward-flowing meltwater (DeSimone and LaFleur, 1986, p. 226) controlled the level of a lake, known as Lake Fort Ann, in the Champlain lowland. Lake Fort Ann or its predecessors inundated most of the Champlain lowland to depths of 200 to 400 feet, and perhaps 600 feet near the Canadian border, as suggested by comparison of modern topography with the proglacial lake profiles inferred by Stewart and MacClintock (1969), Denny (1970), Wagner (1972), and DeSimone and LaFleur (1986). Several types of evidence indicate that the ice margin in the Champlain lowland generally remained active and retreated by calving. For example, extensive deposits of lacustrine silty clay containing dropstones of pebble to boulder size were mapped by Stewart and MacClintock (1969, 1970), who also reported many exposures where clay rests directly on bedrock without intervening till. Exposures of till interbedded with lacustrine fines were reported by Wagner (1972), Connally and Calkin (1972), and Parrott and Stone (1972) from the northern part of the lowland in Vermont. Recessional moraines along the northwestern margin of the lowland in New York were identified by Denny (1974), who ascribed them to active ice. Till overlies sand and gravel in small areas near West Beekmantown, N.Y. (Giese and Hobba, 1970), and along the Ingraham esker (Denny, 1972, fig. 4).

The northern tip of the Adirondack Mountains is known as Covey Hill and lies on the international border. As the ice sheet retreated northward past this point, all water from the St. Lawrence and Ontario lowlands began to drain eastward into the Champlain lowland through a succession of spillways or channels on Covey Hill (MacClintock and Stewart, 1965; Denny, 1974; Clark and Karrow, 1984; Muller and Prest, 1985). Soon, water levels in the St. Lawrence lowland dropped to that of Lake Fort Ann, which extended westward as far as modern Lake Ontario (Clark and Karrow, 1984; Muller and Prest, 1985).

Further ice retreat within Quebec eventually allowed water in northern New York and Vermont to drain northeastward, down the St. Lawrence valley. The Fort Ann spillway was abandoned and the level of ponding dropped to the then-current sea level, which was higher than the isostatically depressed land surface in the Champlain and St. Lawrence lowlands. The resulting incursion of marine water formed what is



A. **Beach ridge south of Lake Ontario near Lockport, N.Y. (From Johnston, 1964, fig. 12)**

B. **Beach ridge east of Lake Ontario near Adams, N.Y. (From Miller and others, 1988, pl. 4B)**

C. **Beach ridge east of Lake Ontario near Sandy Creek, N.Y. (From Zarriello, 1993, fig. 11)**

D. **Wave-built terrace on drumlin near Volney, N.Y. (From Miller, 1982, fig. 7)**

FIGURE 68.—Geologic sections across Lake Iroquois beach deposits.

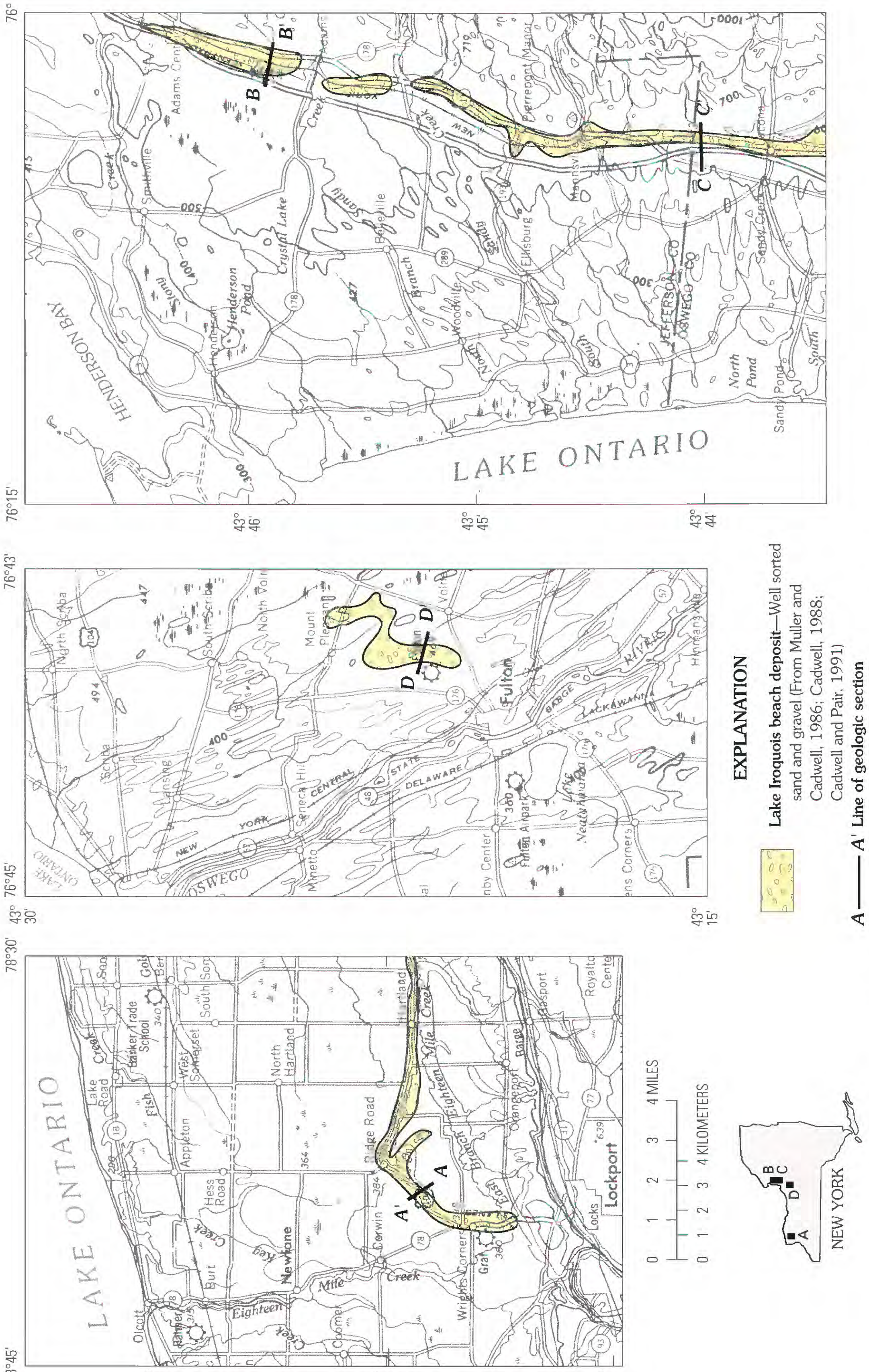


FIGURE 68.—Geologic sections across Lake Iroquois beach deposits—Continued.

known as the Champlain Sea, which initially inundated all areas below a present altitude of about 500 feet near the Canadian border (Stewart and MacClintock, 1969; Wagner, 1972; Denny, 1974). Gradually, the land rose, and after about 1,500 years (Denny, 1974; Muller and Calkin, 1993), the modern drainage pattern was established.

DISTRIBUTION OF SAND AND GRAVEL

Coarse sand and gravel deposits are few and widely scattered in the Black River, St. Lawrence, and Champlain lowlands, except along the proglacial lake shores that bordered the Adirondack upland, and many of these deposits are thin or thinly saturated. Several subregions are identified and described below.

MOUNTAIN-FRONT PERCHED DELTAS AND KAME TERRACES

The proglacial lakes that occupied the lowlands bordering the Adirondack Mountains during deglaciation were far too large to be filled by stratified drift, but several large and (or) multiple coalesced deltas and (or) kame terraces were built into those lakes by streams from the Adirondacks. Although these mountain-front deposits are not especially productive as aquifers, their large volume dominates the areas where they occur and results in a distinctive terrane that is delineated on plate 1 as map unit "D."

The mountain-front deposits have several common characteristics:

1. They were built into proglacial water bodies whose surfaces stood hundreds of feet above the present floor of the broad lowlands. Later, after the lake or marine water had drained, they were incised by the streams that built them and now are largely perched above modern stream grade.

2. Most of them are deltas composed of sand, largely fine sand, that in general progrades outward over silt or clay (fig. 69; also Waller and Ayer, 1975, p. 9, p. 99).

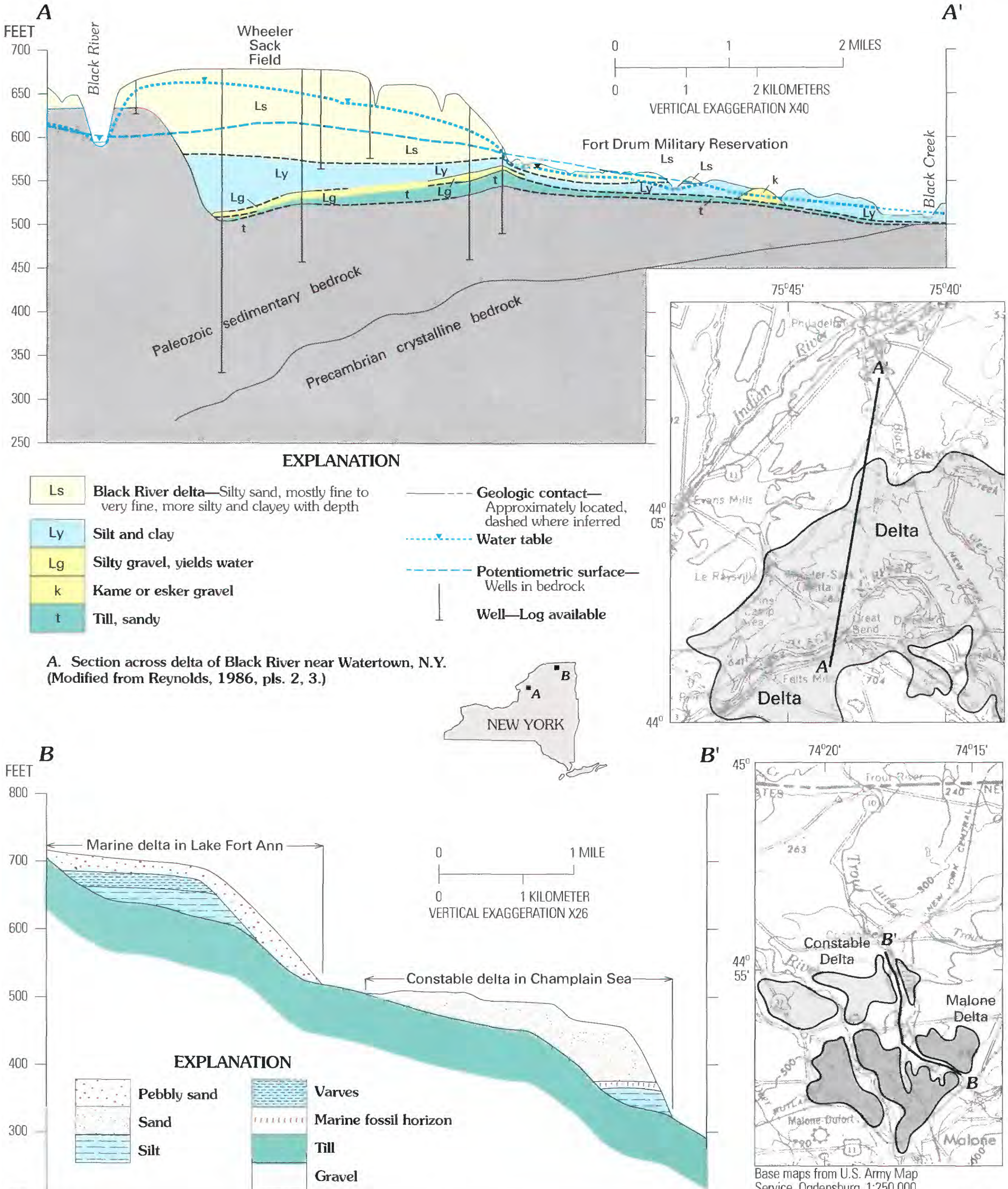
3. Deposition began when the lake was only a narrow pond between the ice sheet and the upland. The earliest deposits, which underlie the highest deltas close to the upland, are commonly coarse-grained, collapsed ice-contact deposits (facies 1). MacClintock and Stewart (1965) depict several streams as incised through deltaic sand into kamic sand and gravel. Waller and Ayer (1975, p. 9) also report that deltas cover or fill in around kames. Later, lower deltas farther from the upland are less likely to overlie ice-contact deposits, although one of the later deltas (fig. 69*A*) is known to overlie thin basal sand and gravel.

4. The concentration of mountain-front deposits near major tributaries suggests an origin predominantly as inwash. Nevertheless, some sediment was derived from the ice. For example, tributary deltas built into the Black River valley were fed by meltwater from ice retreating across the Adirondacks. Some of the kamic sands and gravels that underlie inwash deltas of Salmon River, St. Regis River, and other rivers that drain north from the Adirondacks were probably deposited in headwater segments of earlier subglacial meltwater channel systems that drained southward into and through the Adirondacks. An esker-fed delta at St. Regis Falls (fig. 12) and a large southward-sloping delta at 1,560 feet at Owls Head along upper Salmon River are evidence that large meltwater streams drained into the Adirondacks shortly before glacial Lake Iroquois expanded into the St. Lawrence lowland and the direction of flow in northern Adirondack valleys reversed.

5. Saturated thickness of permeable coarse sand or gravel is likely to be too small to allow large well yields. Geohydrology of these mountain-front deposits has not been studied in any detail, but any permeable sand and gravel in deposits that are banked against bedrock hillsides well above the grade of nearby streams or lowlands can be expected to drain readily. Some deltas are so large and are composed of such fine-grained sand that a large saturated mound is present (fig. 69*A*), but the hydraulic conductivity of the saturated materials is likely to be too low to support large-capacity wells.

BLACK RIVER VALLEY

The Black River flows generally along the western margin of the extensive mountain-front deltas shown on plate 1 and, for much of its length, occupies a narrow valley. Between Glenfield and Carthage, N.Y., however, the valley floor is 3 to 4 miles wide and is underlain by as much as 198 feet of fine-grained lake-bottom sediment (Waller and Ayer, 1975, p. 100). This reach is indicated by the "S" symbol on plate 1. The fines are generally capped by a surficial facies-3 aquifer composed of sand derived from incision of the deltas and thinly saturated (Waller, 1976; Waller, 1986, sheet 3). A buried sand and gravel aquifer is inferred to underlie the fines along part of the valley, although data are sparse (Waller, 1986) and the dimensions and origin of the inferred basal aquifer are poorly known; presumably it is the product of subglacial meltwater that followed the axis of the Black River valley southward before the expansion of Lake Iroquois reversed the direction of flow.



A. **Section across delta of Black River near Watertown, N.Y. (Modified from Reynolds, 1986, pls. 2, 3.)**

B. **Generalized section across two successive deltas of Trout and Salmon Rivers near Malone, N.Y. (Modified from MacClintock and Stewart, 1965, fig. 22b, pl. 1)**

FIGURE 69.—Geologic sections across deltas along the mountain front south of the St. Lawrence lowland, New York.

ST. LAWRENCE LOWLAND SOUTH OF CANTON

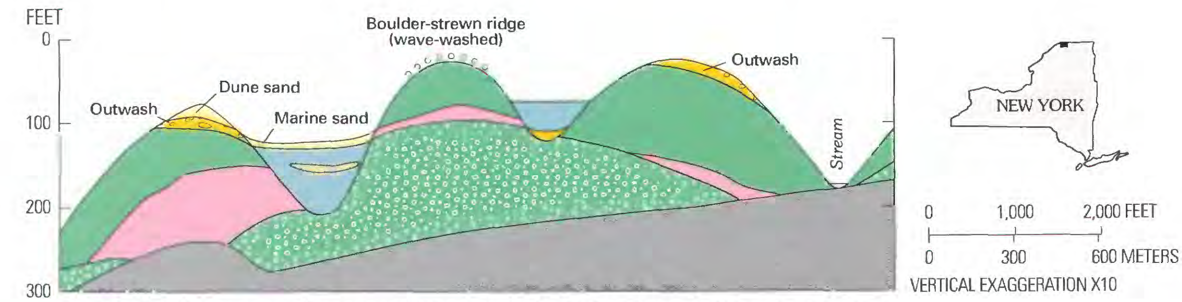
MacClintock and Stewart (1965, p. 26–37) divided the St. Lawrence lowland into different physiographic regions along an east-west line at latitude 44°35', which passes through Canton (pl. 1) and Morristown, N.Y. South of this latitude, till and coarse stratified drift are scarce and bedrock is generally exposed at land surface or mantled by a few feet of silty clay. Where the bedrock is crystalline and Precambrian in age, rounded bedrock knobs rise 30 to 50 feet above flat clay-filled swales. Where bedrock is flat-lying sandstone or carbonates of Cambrian or Ordovician age, broad tablelands are incised by steep-sided valleys. Scattered kames, subaquatic fans, and elongated masses of ice-contact stratified drift were formed during the last deglaciation (MacClintock and Stewart, 1965; Cadwell and Pair, 1991). Most are elongated and (or) aligned northeast-southwest, subparallel to (a) the grain of the bedrock topography, (b) the direction of ice flow (Clark and Karrow, 1983, fig. 12), (c) a prominent esker near Natural Bridge, N.Y. (MacClintock and Stewart, 1965, pl. 1A), and (d) the margin of the ice lobe along the side of the St. Lawrence valley (Clark, 1984, fig. 7; Pair and Rodrigues, 1993). They seem to be associated with bedrock highs, which may have acted as pinning points that briefly stabilized the retreat of the grounding line (Pair and Rodrigues, 1993), and their orientation suggests preferential southwestward meltwater flow controlled by bedrock topography and by a hydraulic gradient toward the end of the ice lobe. Perhaps buried sand and gravel aquifers underlie the clay in a few swales or valley reaches, but no evidence of such has been reported (Dunn Associates, 1968).

ST. LAWRENCE LOWLAND NORTH OF CANTON

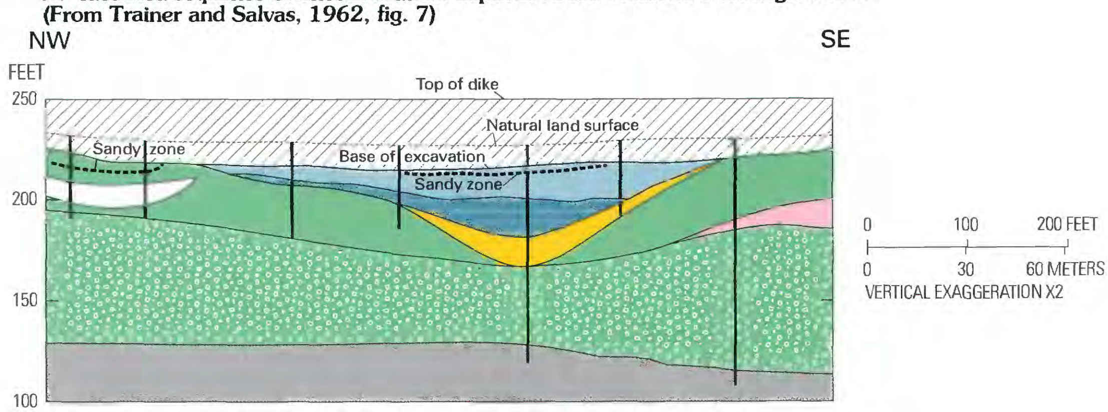
North of Canton and Morristown, N.Y., the St. Lawrence lowland is characterized by low ridges composed largely of till, elongated northeast-southwest, and separated by swales or valley flats underlain by silty fine sand and clay. Bedrock exposures are much less common here than south of Canton. Study of excavations for the St. Lawrence seaway project and records of wells (MacClintock, 1958; Trainer and Salvas, 1962; MacClintock and Stewart, 1965) established the stratigraphy summarized in table 5 and illustrated in figure 70.

TABLE 5.—*Quaternary stratigraphic units in St. Lawrence lowland north of Canton, N.Y.*

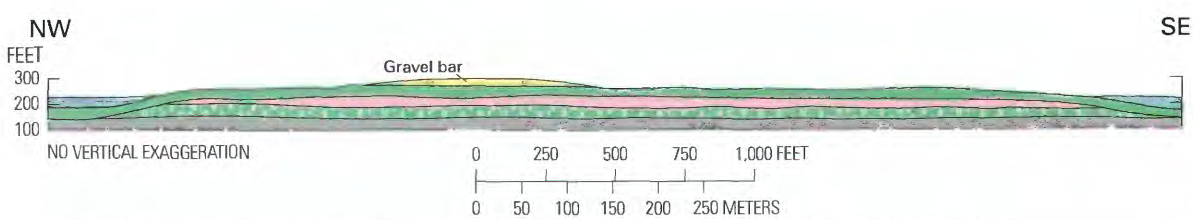
Geologic unit	Relative age (1 = oldest)	Lithology and origin	Distribution
Beach deposits	6–7	Coarse, poorly sorted gravel; overlies till; microtopography of ridges a few feet high, parallel to contours of hill; generally unsaturated.	Tops or (less commonly) sides of hills.
Mountain front inwash deltas	5–6	Chiefly sand, deposited by streams from Adirondack watersheds into proglacial lakes and Champlain Sea, also forms aprons spread along shorelines by longshore currents; generally perched above stream grade.	Along south margin of lowland at altitudes above 500 feet.
Marine sand	7	Silty, fine sand deposited discontinuously atop marine clay during recession of the Champlain Sea; commonly 1 to 10 feet thick; tapped by dug wells.	Common in swales or lowlands between hills.
Marine clay	6	Massive silty clay, nearly impermeable, locally contains sand lenses, deposited in Champlain Sea.	Universally present in swales and lowlands between hills.
Lacustrine clay	5	Varved clay and silt, nearly impermeable, locally contains lenses of resedimented (slumped) diamicton and of sand.	Do.
Subaquatic outwash	4	Chiefly sand; tapped by some test wells and domestic wells.	Locally present beneath clay in swales and lowlands.
Upper till	3	Fort Covington till as defined by Clark and Karrow (1983); compact and impermeable except for scattered lenses of sand and gravel no more than a few feet thick.	All three units underlie most of the ridges that constitute the highest topographic features in the lowland. One or more units continue beneath some underlying swales.
Middle unit	2	Diamicton layers interbedded with stratified sediments ranging from clay to sandy gravel.	
Lower till	1	Till, more compact than upper till, nearly impermeable except for scattered thin lenses of sand and gravel.	



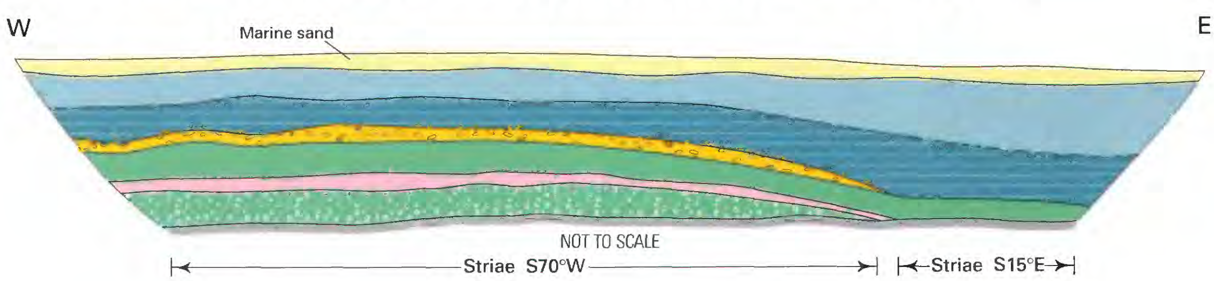
A. **Idealized sequence of unconsolidated deposits in the Massena-Waddington area (From Trainer and Salvas, 1962, fig. 7)**



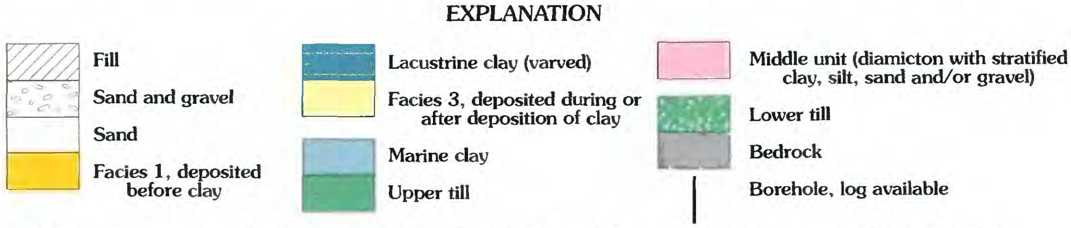
B. **Section beneath part of Richards Landing dike (From Trainer and Salvas, 1962, fig. 5)**



C. **Measured section through Eisenhower lock (From MacClintock and Stewart, 1965, fig. 14)**



EXPLANATION



D. **North face of excavation for Grass River lock (Modified from MacClintock and Stewart, 1965, fig. 25)**

FIGURE 70.—Typical stratigraphy in the St. Lawrence lowland north of Canton, N.Y.

Coarse-grained stratified drift is generally thin and capable of yielding only small amounts of water. Trainer and Salvas (1962, p. 32) report sand beneath the lacustrine clays in several swales near Massena, N.Y., atop till (fig. 70*A*, *B*). Perhaps subaquatic fans at the mouths of numerous small subglacial channels merged to form a thin but extensive sand sheet in this region. Facies-3 surficial sand was widely deposited atop clay in swales or valleys as the Champlain Sea retreated, but it is fine grained, silty, generally only a few feet thick and, hence, a very minor aquifer.

Sand and gravel lenses in the middle unit, between the principal till sheets, presumably were deposited during a previous deglaciation. They underlie some swales but are found chiefly within hills (fig. 70*A*). Some are easily drained, the overlying till limits recharge, and interlayered diamicton and fines limit hydraulic conductivity. Thus, these older stratified deposits are not important aquifers.

CHAMPLAIN LOWLAND

Deposition in proglacial lakes along the margins of the Champlain lowland began as ice-contact deposits that were overlapped or succeeded by deltas (Stewart and MacClintock, 1969, p. 127). Where such deposits are confined to valleys within the Adirondack and Green Mountains that border the Champlain lowland, they are considered part of other hydrophysiographic regions, even though some lower depositional surfaces may have been graded to regional lakes. Extensive deltas banked along the west side of the Champlain lowland near Plattsburgh and Keesville, N.Y., are mapped on plate 1 as part of the mountain-front perched delta terrane.

The abundance of inwash contrasts with the lack of evidence of facies-1 proximal deposits from the ice itself. The Ingraham esker (Denny, 1972; Franzi and Cronin, 1988), which extends for 10 miles in the towns of Chazy and Beekmantown in eastern Clinton County, N.Y., is the only known product of meltwater in a subglacial channel. A core of coarse, heterogeneous ice-contact stratified drift is overlain by well-bedded sand and gravel in a typical fining-upward subaquatic-fan sequence and is capped or bordered by fossiliferous gravel reworked from the esker materials during the declining phase of the Champlain Sea. Sand or gravel occurs elsewhere beneath clay, mostly at depths of 110 to 260 feet, as shown by several scattered wells from Plattsburgh south to Lake George, but these wells are too few to demonstrate continuity of the sand and gravel. Buried aquifers were delineated by Giese and Hobba (1970) in only two localities in New York, one of these being immediately south of the Ingraham esker. A buried aquifer along the edge of the lowland in Swanton, Vt., is documented by borehole logs (Hodges, 1967b). Connally and Calkin (1972) describe one exposure where sand underlies varved clay and till; Stewart and MacClintock (1969) report several exposures where lake deposits lie directly on bedrock. Thus, the extent of buried facies-1 deposits in the Champlain lowland is not yet accurately known.

The Winooski, Lamoille, and Mississquoi Rivers enter the Champlain lowland in Vermont near and north of Burlington. Extensive sand plains were deposited by these rivers in the Champlain Sea, regraded and extended as sea level declined, then incised by postglacial streams. In general, 10 feet or more of pebbly fine to medium sand overlies very fine sand, silt, and clay and constitutes a surficial aquifer perched above stream grade. A similar condition obtains in a small area south of Plattsburgh, N.Y. (Giese and Hobba, 1970; Denny, 1974). Both areas are depicted by map unit "S" on plate 1. Elsewhere in the Champlain lowland, facies-3 surficial aquifers are absent, or very thin and localized.

COASTAL MAINE AND NEW HAMPSHIRE

The late Wisconsinan ice sheet advanced about 200 miles beyond the present Maine coastline (Tucholke and Hollister, 1973) when sea level was abnormally low because so much water was incorporated in the ice sheet. Sea level rose rapidly during deglaciation, however, and abutted the ice sheet as it retreated inland from the present coastline of Maine and New Hampshire. At that time, land surface was depressed by the weight of the ice, and several hundred years passed before enough rebound took place to cause the sea to recede to approximately its present position (Borns and Stuiver, 1975). Although large hills stood above the ocean, marine water inundated much of the landscape within 10 to 60 miles from the present coastline, as indicated by map units "L" and "S" on plate 1. The stratified drift includes the same three depositional facies found elsewhere in the glaciated Northeast—early, coarse-grained ice-contact deposits that locally constitute the entire stratigraphic section; fine-grained deep-water deposits; and late, coarse-grained surficial deposits. The geometry, distribution, and lithology of these three facies are described in turn below.

EARLY ICE-CONTACT DEPOSITS

The geometry and distribution of ice-contact deposits (facies 1) in the area of marine inundation differ from that in most southward-draining regions in two respects: (1) Spacing is not strictly a function of spacing of preexisting bedrock valleys but rather reflects the spacing of meltwater channels within the ice sheet and the position of the ice margin during oscillations or pauses in retreat; and (2) the zone of stagnant ice that typically bordered the retreating ice margin and was the principal locus of ice-contact deposition elsewhere seems to have been generally absent here, especially near the present coast; instead, the forward edge of the ice probably stood as a cliff or floating shelf of active ice that calved into the ocean (Smith, 1981, 1982, 1984; Thompson, 1982; Moore, 1982). Ice-contact deposits formed in three or perhaps four types of settings, described in turn below.

Eskers and Related Deposits.—Much of the coarse sand and gravel in coastal Maine and New Hampshire is found in or associated with eskers, many of which extend, with only short gaps, for several miles in generally north-south orientations (Stone, 1899; Leavitt and Perkins, 1935). Eskers are described in a previous section of this paper entitled "Esker Systems in Maine." In this coastal region, generally, an esker will have a core of relatively coarse gravel, overlain or surrounded by finer gravel and sand and commonly overlapped by marine clay in a typical fining-upward subaquatic-fan sequence (Thompson and Smith, 1983, p. 5–6, 34).

Ice-Marginal Deposits.—Linear ridges parallel to the former ice front have been recognized in many places in coastal Maine (Prescott, 1974a, b; Borns, 1974; Thompson, 1980, 1982; Smith, 1982; Smith and others, 1982). These ridges consist of gravel and coarse sand, sand and silt, till, or any combination thereof. The stratified components are thickest near eskers but are widely distributed elsewhere. Some of these ridges are thought to have formed by ice push during oscillatory retreat of an active ice front, as evidenced by numerous thrust faults and recumbent folds, by intercalated till slices, and commonly by a till carapace atop the proximal slope (Smith, 1982; Smith and others, 1982; Thompson and Smith, 1983, p. 27, 32). Other low ice-marginal ridges seem to be incipient deltas that were deposited under or in front of a floating ice terminus but did not build up to sea level (Thompson, 1982; Smith, 1982).

Glaciomarine Deltas.—Deltas were built into the sea along the retreating margin of the ice sheet in at least 50 localities in Maine (Thompson, 1982, p. 221) and at least 26 in New Hampshire (Moore, 1982; Larson and Goldsmith, 1989; Koteff and others, 1989, 1993). They are characterized by:

1. A broad, level surface, graded to the maximum level of marine submergence (Thompson, 1982, p. 222).
2. Thin surficial gravel (originally fluvial topsets, commonly reworked by waves) overlying foresets of sand and fine gravel (Koteff and others, 1993).
3. Interfingering distally and downward with marine clay (fig. 71*G–I*).
4. Kettles and collapsed deposits on the proximal margin (fig. 71*H*). Less frequently, the proximal margin is a bedrock ridge, cut by one or more saddles through which meltwater carried sediment.

Glaciomarine deltas formed only where major meltwater drainage systems emptied into the sea. Many occur at or near the inland limit of marine submergence, commonly at the southern end of esker systems (fig. 71*H*). Some are laid out in series along an esker, having formed successively as the ice margin retreated (Leavitt and Perkins, 1935, p. 52; Thompson, 1982, p. 221). Others represent successive northward-retreating positions of the mouths of persistent meltwater streams not associated with eskers (Larson and Goldsmith, 1989).

Possible Widespread Sheets of Sand or Gravel.—Some scattered kames or kame fields or hummocky stratified drift are not on esker or ice-marginal trends nor associated with known glaciomarine deltas (Prescott 1974b, 1977; Thompson, 1982). Thompson (1982) interprets some of these deposits as originating because bedrock or till hills "temporarily anchored the ice margin and localized deposition of sand and gravel that washed out from beneath the ice," implying that a thinner layer of sand and gravel might have been widely deposited elsewhere. If meltwater emerged from the base of the ice as small streams or as sheet flow, in most of this region it would have entered marine water in which its sediment load could readily be deposited as stratified drift, perhaps in the form of small subaquatic fans. Furthermore, a floating ice shelf rises and falls with the tide and, thus, allows cyclical landward and seaward flow of seawater in the thin space beneath the ice, a process that might winnow the upper part of the underlying till and any sediment dropped from the melting ice shelf. These mechanisms had the potential to create a discontinuous but widespread layer of sand or gravel between the till and the marine clay that was deposited later after the ice front had retreated farther. Bingham (1981), Smith (1982), and Smith and others (1982) report that such a thin sand or gravel deposit mantles large areas on low hills in the coastal region and is thickest near eskers; Smith ascribed it to the mechanisms mentioned above. He did not collect subsurface data to show whether the same deposit extended beneath adjacent valleys, however. Subsurface data were obtained by Tepper (1980) from 25 domestic wells drilled through marine clay in valleys in the town of Arundel, Maine, 13 of which penetrated sand or gravel beneath the clay; although one well reportedly penetrated 100 feet of gravel, generally the basal sand or gravel was thin and had no obvious distribution pattern. Some wells in other localities studied by Tepper (1980) also penetrated gravel between clay and bedrock. Most reports on the ground-water hydrology of coastal Maine and New Hampshire do not mention any widespread thin basal aquifer beneath marine sediments, however.

FINE-GRAINED MARINE DEPOSITS

Fine-grained marine sediment, the Presumpscot Formation of Bloom (1960), is widespread in coastal Maine and New Hampshire. Maps by Prescott (1963, 1974a, 1977) and Bradley (1964) suggest that fine-grained marine sediment may underlie as much as 65 percent of some towns near the coast. In areas of higher relief, the marine deposits are limited to the deeper valleys, which once constituted estuaries between till-covered uplands.

Fine-grained marine sediments reach thicknesses of 100 to 200 feet in many lowlands (Prescott, 1967; Borns, 1974; Tepper, 1980; Smith, 1981; Thompson, 1982, p. 24) but are only a few feet thick on many hillsides within the inundated region. Generally, the maximum altitude of marine silty clay is 50 feet or more below the maximum sea level during deglaciation in any particular locality (Thompson, 1978).

The typical marine sediment is a blue-gray, massive, silty clay (Thompson, 1982). Thin clayey silt layers separated by sand and silt laminae were deposited in the inland estuaries (Borns and Hagar, 1965a, p. 1236) and near the top of the marine section (Bradley, 1964).

The marine deposits overlap and locally bury ice-contact deposits such as kames and eskers. Locally, near the landward limit of their deposition, marine deposits are found only in depressions such as kettleholes and the bottoms of modern lakes. A few exposures indicate that the marine silty clay was deposited before the last remaining ice melted (Goldthwait and others, 1951, p. 43; Borns and Hagar, 1965a, p. 1237). The marine silty clay interfingers with the distal parts of glaciomarine deltas and deltaic ice-frontal deposits (Smith and others, 1982, fig. 4; Moore, 1982, fig. 3; Moore, 1990, fig. 10).

LATE COARSE-GRAINED SURFICIAL SEDIMENT

Although the Presumpscot Formation occupies lowlands throughout coastal Maine and New Hampshire, commonly it is mantled by younger sand and gravel of various origins. Outwash, chiefly well-sorted fine to coarse sand, is found near the marine limit in many localities, deposited as valley trains by aggrading meltwater streams and as deltas prograding across marine fines. Outwash can be several tens of feet thick but commonly becomes finer with depth and thins southward; many test borings (Tolman and others, 1983) in the large area of deltaic outwash between Kennebunk and Sanford, Maine, reveal 35 to 90 feet of sand, the lower part fine to very fine, overlying marine clay (fig. 71*F*). As postglacial isostatic rebound raised land surface relative to the sea, streams incised the outwash and redeposited the sand farther out into the receding sea; most of this redeposited marine sand is less than 15 feet thick, thinner than the outwash as well as lower in altitude. The stream terraces incised into the outwash are capped by alluvial deposits; those along the Kennebec River are chiefly coarse gravel and were interpreted by Borns and Hagar (1965b) as a valley train from a late-glacial ice cap, but later were reinterpreted as postglacial (H.W. Borns, University of Maine, written commun., 1995). Beach deposits, chiefly silty fine sand but locally gravel, occur as fringes around hills and were derived by wave erosion of the hillsides as the sea receded.

AQUIFER GEOMETRY IN THIS REGION

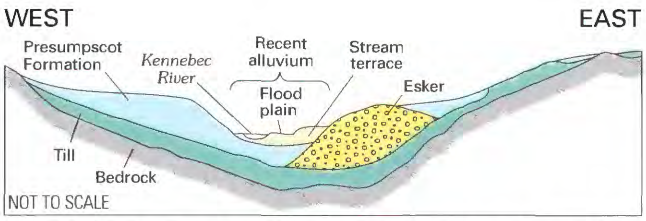
Several examples of typical aquifer geometry in the area of marine incursion in Maine, New Hampshire, and extreme northeast Massachusetts are illustrated in figure 71. Facies-1 aquifers include eskers, related constructional ice-contact deposits, widespread thin sheets, and glaciomarine deltas; facies-3 aquifers include outwash and alluvium. Two generalized sections across the Kennebec River valley (fig. 71*A*, *B*) show an esker overlapped by marine silt and clay. In the upstream reaches of this valley (fig. 71*B*), the marine fines grade up into sandy outwash (Embden Formation of Borns and Hagar, 1965a, b) that is largely above stream grade and, thus, could supply only small well yields. Both sections show alluvium, inset into the marine fines and extending below stream grade, that is mostly gravel and could perhaps be tapped by shallow wells. Near Pittsfield, Maine (fig. 71*C*), an esker that follows the Sebasticook River valley for a few miles is bordered by swampy lowlands that are capped by 15 to 20 feet of outwash or alluvium, beneath which are marine fines. A section along the Maine Turnpike north of Portland (fig. 71*D*) crosses an ice-contact deposit whose cross profile is eskerlike, but no esker is identified here on maps by Prescott (1976) or Thompson and Borns (1985). This section also crosses buried ice-contact deposits near the Presumpscot River that border a kame terrace (Prescott, 1976) and probably represent a collapsed or subglacially deposited extension of that kame terrace. The outwash deposits in this section are thin, discontinuous, and substantially below the maximum level of marine inundation in this region; probably they represent outwash redeposited during retreat of the ocean. A broad valley drained by the Picassic River at Epping, N.H. (fig. 71*E*), lies near the inland limit of marine inundation; test borings indicate that a thin sheet of proximal sand and gravel underlies at least the north half of the valley, beneath fine sediment that largely fills the valley and is capped by a thin surficial outwash aquifer (Cotton, 1988). Outwash in the town of Kennebunk, Maine (fig. 71*F*), becomes finer with depth and grades to marine clay; presence of a localized basal aquifer is inferred on the basis of seismic data and a domestic well a mile away. A section typical of glaciomarine deltas in southeastern New Hampshire (fig. 71*G*) shows how deltaic fine sand overlies and interfingers with concurrently deposited marine clay. Many deltas ceased receiving coarse sediment long before marine waters receded from this region, resulting in deposition of marine clay atop the distal foreset sands (fig. 71*H*). Pineo Ridge, an unusually large glaciomarine delta (fig. 71*I*) owes much of its apparent bulk to being built over a bedrock ridge; the relation of this delta to marine fines is speculative. At the proximal end of each of these glaciomarine deltas (fig. 71*G–I*) is saturated sand and gravel, a potentially productive aquifer.

The principal aquifers in coastal Maine and New Hampshire are the facies-1 ice-contact deposits, but many such deposits are thinly saturated because they were deposited around bedrock knolls or on the sides of valleys above stream grade. Test wells at the following types of sites are likely to penetrate relatively thick, saturated sections of permeable ice-contact deposits:

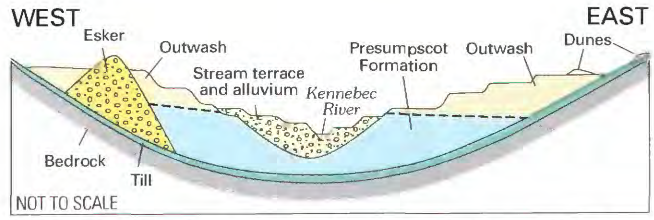
1. Anywhere within exposures of an esker or other ice-contact deposit that is near the axis of a bedrock valley (as in fig. 71*A* or fig. 71*B*) or in some other setting where the bedrock surface is several tens of feet below stream grade.
2. Near the valleyward margin—the margin closest to the valley axis—of an esker or other ice-contact deposit that is exposed along the side of a bedrock valley, because saturated thickness of the exposed deposit is likely to be greatest on the valleyward side, and furthermore, the collapsed flanks may extend valleyward a few hundred feet beneath the overlapping marine silty clay.
3. In gaps between exposed segments of an esker or a linear ice-marginal deposit, because deposition of sand and gravel along the esker or morainal trend may well have been continuous; some apparent gaps may be places where the sand and gravel deposits are buried beneath younger sediments within a bedrock valley or planed off by subsequent fluvial erosion.
4. Near the head or apex of glaciomarine deltas where the sediment is most likely to be coarse-grained from top to bottom.

Although most large-capacity wells in coastal Maine and New Hampshire are located where facies-1 ice-contact deposits are exposed at land surface, a few tap ice-contact deposits overlapped by marine silty clay. For example, three municipal wells within 300 feet of an esker ridge south of South Lincoln, Maine, obtain 2,000 gallons per minute from gravel below the clay that mantles most of the valley, and several nearby wells drilled on, beside, or between segments of the esker penetrate and (or) tap sand and gravel (Tolman and Lanctot, 1981b). The municipal well at Oxford, Maine (see fig. 19), is another example. Several test boreholes and a municipal well 1 mile southwest of Machias, Maine, penetrate through marine clay into gravel or silty sandy gravel (Weddle and others, 1988); the surrounding knolls are mapped as end moraine (Prescott, 1974b; Thompson and Borns, 1985). A municipal well at Sabattus, Maine, penetrated gravel below marine clay on the flood plain of Maxwell Brook, alongside a ridge of ice-contact deposits (Prescott and Attig, 1977). North of Sabattus Pond, gravel was inferred by Prescott (1968) and Tepper and Lanctot (1985) to underlie marine clay along a swath aligned with an esker to the north and ice-contact deposits to the south; the gravel was penetrated by a test hole (Prescott, 1967).

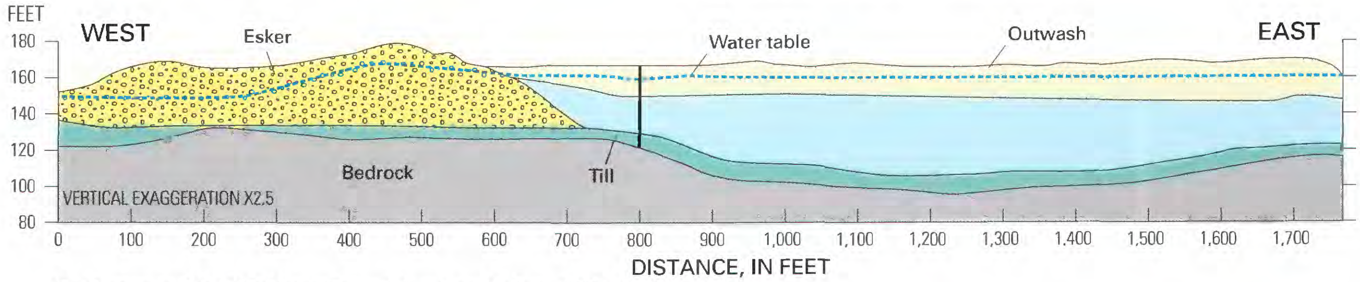
Surficial sand-plain aquifers are extensive and continuous atop marine fines in a few parts of coastal Maine and New Hampshire (pl. 1). Elsewhere in this region, sand-plain aquifers are locally significant, chiefly along large streams near the inland limit of marine inundation; but in most places, little or no sand caps the marine clay. All sand plains in the region, whether localized or extensive, have been incised by large streams, and springs are common along the contact between sand and the underlying clay. Sand-plain aquifers are tapped by many shallow dug and driven wells, but few large-capacity wells are known to be finished in these aquifers, probably because the saturated section is thin or composed chiefly of fine sand (Tolman and others, 1983).



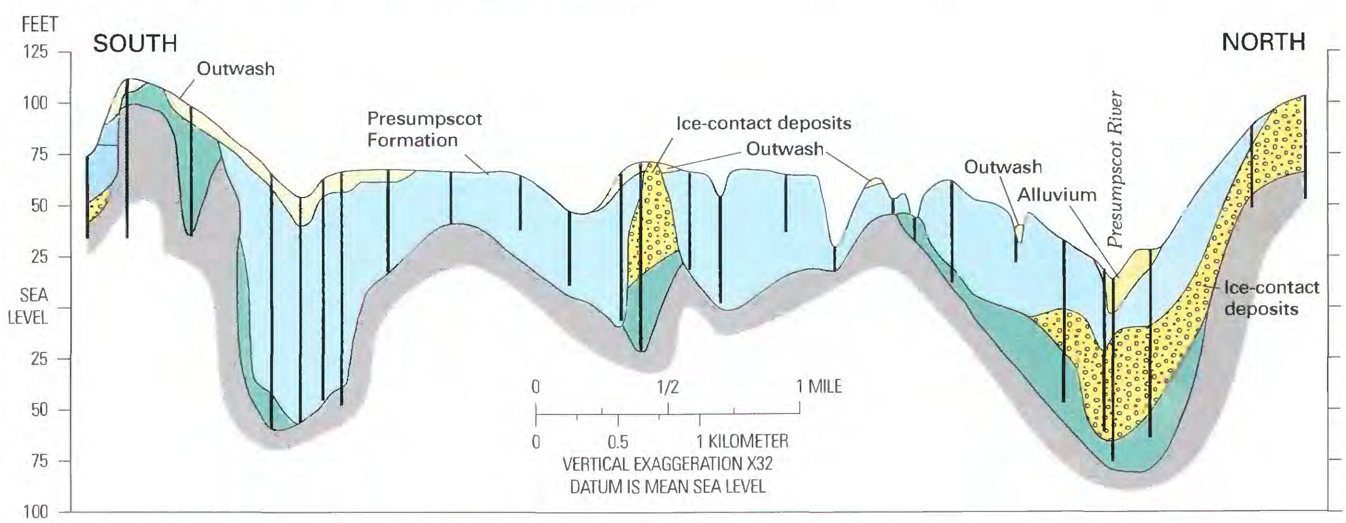
A. **Generalized section across Kennebec River valley near Augusta, Maine (Modified from Thompson, 1978, fig. 20) Bedrock is about 60 feet below river level within this valley (Tolman and Lanctot, 1975)**



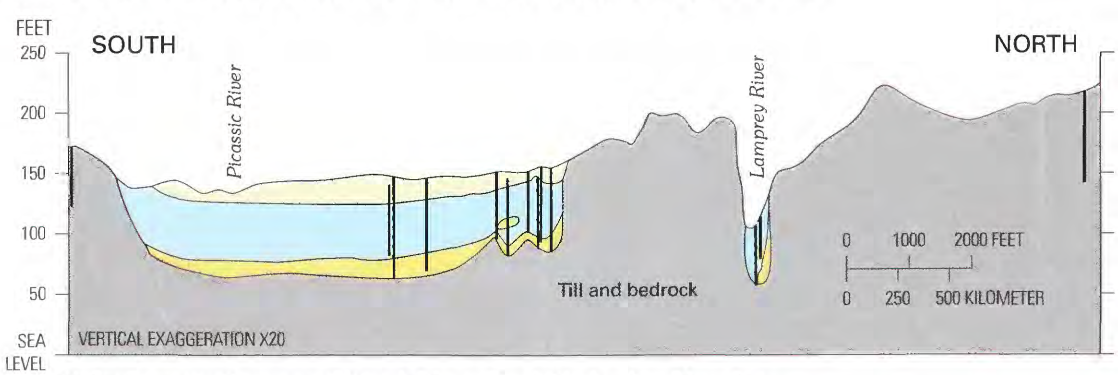
B. **Generalized section across Kennebec River valley between Norridgewock and Solon, Maine (Modified from Borns and Hagar, 1965b, fig. 1) Bedrock is 60-100 feet below river level along this valley reach (Adamik and others, 1987; Tolman and Lanctot, 1981c)**



C. **Section in Pittsfield, Maine (From Adamik and others, 1987, fig. 4)**

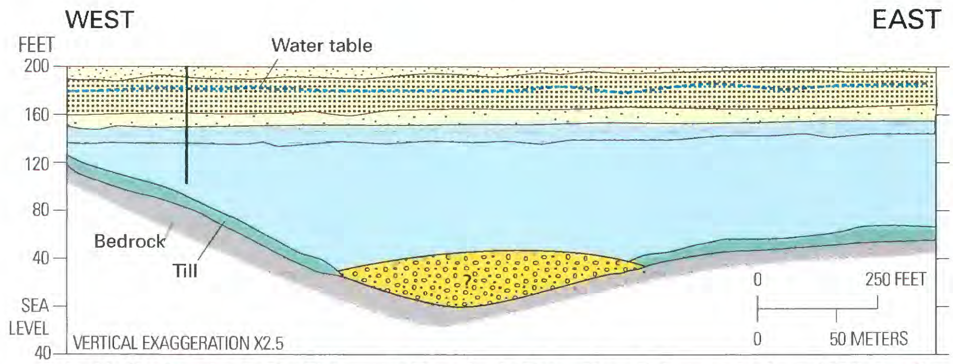


D. **Section along Maine Turnpike in Portland, Maine (Extended from Prescott, 1976, section A)**

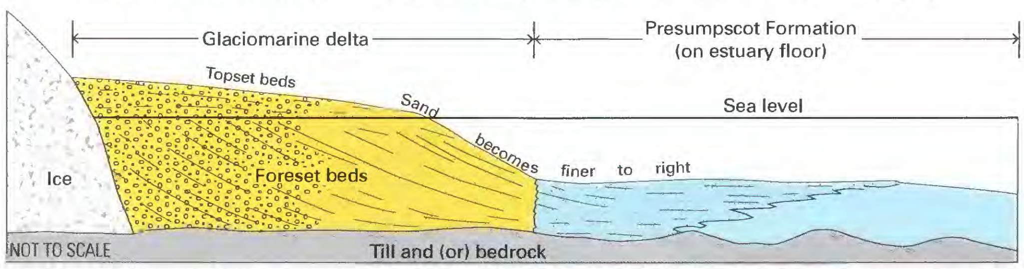


E. **Section across Lamprey and Picassic River valleys in Epping, New Hampshire (Modified from Cotton, 1988, fig. 6)**

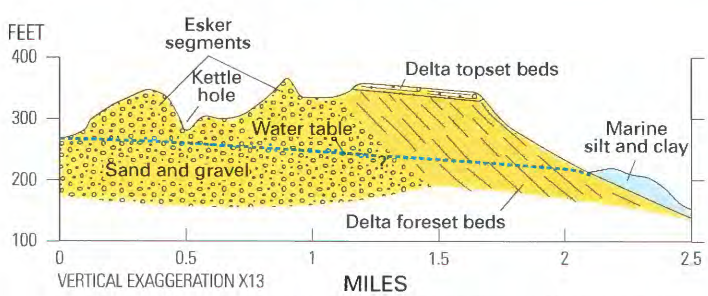
FIGURE 71.—Geologic sections representative of the area inundated by marine water during deglaciation in coastal Maine and New Hampshire.



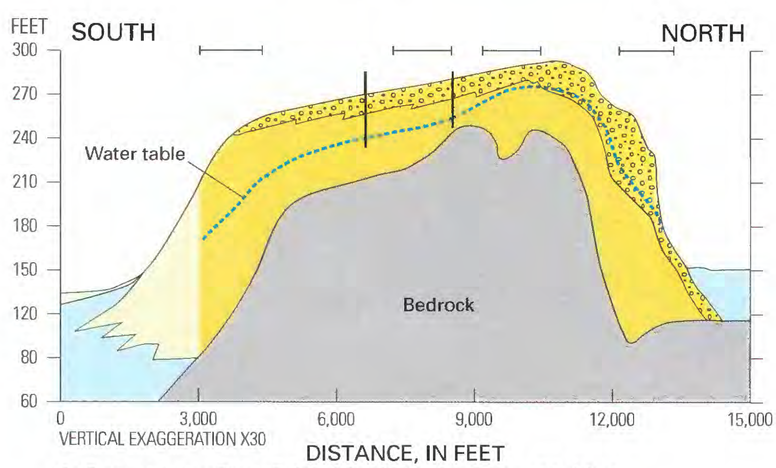
F. **Idealized section in The Plains area, Kennebunk, Maine (From Tolman and others, 1983, fig. 4)**



G. **Schematic drawing of the relation between glaciomarine deltas and the Presumpscot Formation in southeastern New Hampshire (From Koteff, 1991)**



H. **Section through esker-fed glacial-marine delta near North Augusta, Maine (From Thompson, 1978, fig. 9)**



I. **Section across Pineo Ridge delta, Columbia, Maine (Modified from Weddle and others, 1988, fig. 3)**

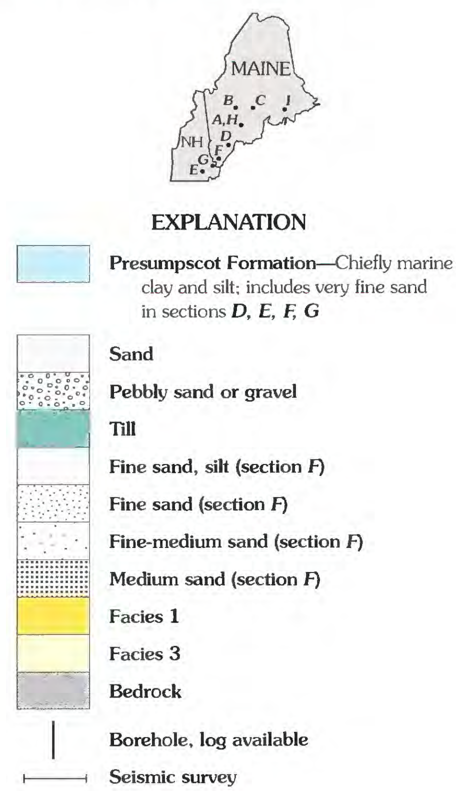


FIGURE 71.—Geologic sections representative of the area inundated by marine water during deglaciation in coastal Maine and New Hampshire—Continued.

SMALLER LOWLANDS

Four of the lowlands classified as "broad lowlands inundated by large water bodies" (map units "S," "L," and "D" on pl. 1) are smaller than the rest. They are described below. Although they could be more accurately termed "narrow lowlands" or even "broad valleys," they resemble the broader lowlands in that facies-2 fines greatly predominate and continue uninterrupted for miles; surficial facies-3 aquifers are generally absent or thinly saturated, and facies-1 ice-contact deposits are sparsely distributed. Some valleys of comparable size elsewhere also contained proglacial lakes several miles long but are not included in the same category because the fines were thought to be interspersed with coarse facies in many places; however, such distinctions are gradational and commonly poorly documented.

ST. JOHN RIVER VALLEY IN MAINE AND NEW BRUNSWICK

Glacial Lake Madawaska extended more than 80 miles up the St. John River watershed from Grand Falls, New Brunswick, where the river had been diverted over a bedrock sill by a moraine that crossed and filled the preglacial valley (Rampton, 1986). The northern headwaters were isostatically depressed at that time, perhaps to the extent that runoff could overflow northward into the St. Lawrence River watershed (Kite and Stuckenrath, 1986). These factors were apparently responsible for the unusual length of Lake Madawaska and its persistence for about 1,200 years (Kite and Stuckenrath, 1986).

The St. John valley is filled largely with silt and clay rhythmites along the international border from Madawaska, Maine, and Edmundston, New Brunswick, downstream to Grand Falls (pl. 1). Ice-contact deposits are discontinuously present along the south side of the valley, and boreholes near Van Buren, Maine, penetrated coarse, basal, stratified drift beneath the fines; but in general, the extent and continuity of any basal aquifer in this valley reach are poorly documented. Lake Madawaska extended farther up the St. John valley to Fort Kent, Maine, but because this upper reach is relatively narrow and seems to have been filled with ice-contact deposits, it is excluded from the lake-dominated lowland that is represented by map unit "S" on plate 1.

The fine-grained sediment in the St. John valley from Madawaska to Grand Falls was apparently never capped by outwash or large tributary deltas (Kite, 1983; Kite and Borns, 1980). In Holocene time, however, after the lake was filled with sediment or drained, major episodes of fluvial aggradation and incision took place (Kite, 1983; Kite and Stuckenrath, 1986). The coarse channel-bar and channel-lag gravels that are generally found at the base of the alluvial deposits may constitute a thin surficial aquifer beneath the modern flood plain, in hydraulic contact with the river, but similar gravels beneath stream terraces are probably dry or only thinly saturated.

LAKE HACKENSACK, NEW JERSEY

The lowland in northeastern New Jersey that is drained by the Hackensack and lower Passaic Rivers was deepened by glacial erosion and blocked by the Harbor Hill (terminal) moraine to the south, resulting in proglacial ponding during deglaciation. The broad lowland represented by map unit "L" on plate 1 is underlain chiefly by deposits of Lake Hackensack, whose stable level was controlled by bedrock spillways, but also by deposits of earlier, higher, less stable lakes (Stanford and Harper, 1991); varve counts indicate that the lakes persisted for at least 3,350 years (Reeds, 1926; Averill and others, 1980). This lowland is distinguished by:

1. Sparse volume of proximal ice-contact sand and gravel, insofar as reported in publications.
2. Extensive laminated clay and silt that varies widely in thickness over an irregular topography of low ridges and valleys.
3. No postlacustrine facies-3 aquifer that is of any consequence.
4. Surficial organic-rich, fine-grained sediment several feet thick.

A few reports present subsurface information from boreholes in this urban area. For example, Agron (1980) shows geologic sections along the Pulaski Skyway and at Meadowlands Arena that indicate only clay and till above bedrock. Nichols (1968, p. 20) reports that the buried valley in the Newark area is filled with clay. Carswell (1976) presents a section from Rutherford to Union City that indicates unconsolidated sediment to be mostly clay over till, except for basal gravel at the east end, which is immediately north of a small delta subsequently mapped by Stanford and others (1990).

The apparent lack of ice-contact sand and gravel in Lake Hackensack can perhaps be explained by the configuration of the watershed. The eastern margin of the lowland is a narrow basalt ridge, beyond which is the deep Hudson River trench that also truncates the headwaters of the watershed and, therefore, could have captured all subglacial meltwater from the north and east. The only substantial tributary on the west is the

Passaic River, which could not contribute much coarse sediment because the lower part of its northward-draining watershed was mantled by silt and clay from an earlier proglacial lake. The only evidence of inwash is several small deltas along the margins of the lowland, mentioned by Carswell (1976) and mapped by Stanford and others (1990). Thus, any ice-contact sediment that underlies the fine-grained lake-bottom deposits was probably derived entirely from local sources.

At its northern end, Lake Hackensack was limited to narrow arms along several valleys. These narrow valleys contain ice-contact deposits, deltas, and lake-bottom fines mantled by fluvial sand and gravel (Stanford and others, 1990; Stanford and Harper, 1991). They resemble other southward-draining valleys in New Jersey and, therefore, are not included in the broad lowland region delineated on plate 1.

Lake Hackensack drained when ice retreat uncovered a gap at Sparkill, N.Y. (Stanford and Harper, 1991). Post-lake streams flowed north across the former lake floor at first but were gradually diverted southward by isostatic rebound. They deposited an alluvial layer 10 to 20 feet thick (Averill and others, 1980) that is of little importance as an aquifer because it is mostly fine to very fine sand and silt (Carswell, 1976, p. 12). Eventually, rising sea level inundated the former lake floor, resulting in a vast salt marsh in which organic-rich sediments accumulated prior to urban development.

LAKE HOOKSETT, NEW HAMPSHIRE

The smallest lacustrine area identified on plate 1 is a broad reach of the Merrimack River valley near Concord, N.H. A narrow reach south of Concord was largely filled by profuse deposition of coarse ice-contact sediment, culminating in deltas that choked the valley to the point that meltwater was diverted across a nearby bedrock spillway, thereby creating glacial Lake Hooksett (Koteff and others, 1984). The lake expanded northward from Bow Junction at least to Franklin, a distance of 19 miles; this broad reach is characterized by widespread facies-2 fines, which grade up to deltaic sand plains that are now deeply incised. Rhythmic sediments, including varved fine sands or silts near deltas but silty clay at most localities, are exposed near river level all along this reach (Antevs, 1922). Borehole records in U.S. Geological Survey files also document the predominance of facies-2 fines. Coarse facies-1 sediments are known to underlie or border the fines in a few places along the valley sides, however, so a more extensive buried aquifer cannot be ruled out.

LAKES SOUTH OF THE TERMINAL MORAINE IN NEW JERSEY

When the margin of the ice sheet stood at the terminal moraine in New Jersey, it blocked several northward-draining lowlands or valleys, thus creating proglacial lakes that spilled southward through various saddles. As meltwater poured into these lakes, fine-grained sediments were deposited all across the lake floor, while sand and gravel prograded south atop the fines as deltaic outwash. The outwash generally pinches out about 1 mile south of the moraine (Stanford and others, 1990) or, in one valley, becomes quite thin at that distance (fig. 72); fine-grained sediment predominates at land surface over most of the lake area.

If the terminal moraine represented the southern limit of glaciation, then the extensive fine-grained lake deposits would necessarily rest on weathered bedrock or, locally along valley axes, on thin pre-glacial alluvium. Such is not the case, however. Several geologic studies have shown that the ice that deposited the terminal moraine briefly advanced well beyond that moraine, as did earlier ice sheets. For example, Ridge (1983) recognized a fringe of till and erratic boulders that extends as much as 3 miles beyond (south of) the terminal moraine in the Delaware River valley. Cotter and others (1985, p. 9) cite other studies that recognized drift south of the moraine. Geohydrologic studies of a few localities beyond the terminal moraine have reported fine-grained lake deposits to overlie sand and gravel that might be ascribed to subaquatic fans as the ice melted back from its maximum advance. For example, Great Swamp, which now occupies the basin of the largest of these proglacial lakes, is underlain by 5 to as much as 128 feet of silt, clay, and muck, beneath which is a basal sand and gravel layer probably less than 25 feet thick in most places (Vecchioli and others, 1962; Minard, 1967). Long Valley, southwest of Kenvil, contains as much as 100 feet of clay underlain by basal gravel that Harper (1978, 1979) ascribed to glacial advance beyond the terminal moraine. The northern part of the Lamington River valley, near Kenvil, also contains a basal aquifer (fig. 72), and unpublished well records 1 to 2 miles farther south reveal a complex stratigraphy of probable ice-contact origin. In summary, the proglacial lakes immediately south of the terminal moraine in New Jersey were not beyond the outer limit of ice advance, and, therefore, aquifer geometry in the area inundated by those lakes is not fundamentally different from that in areas inundated by large proglacial lakes farther north.

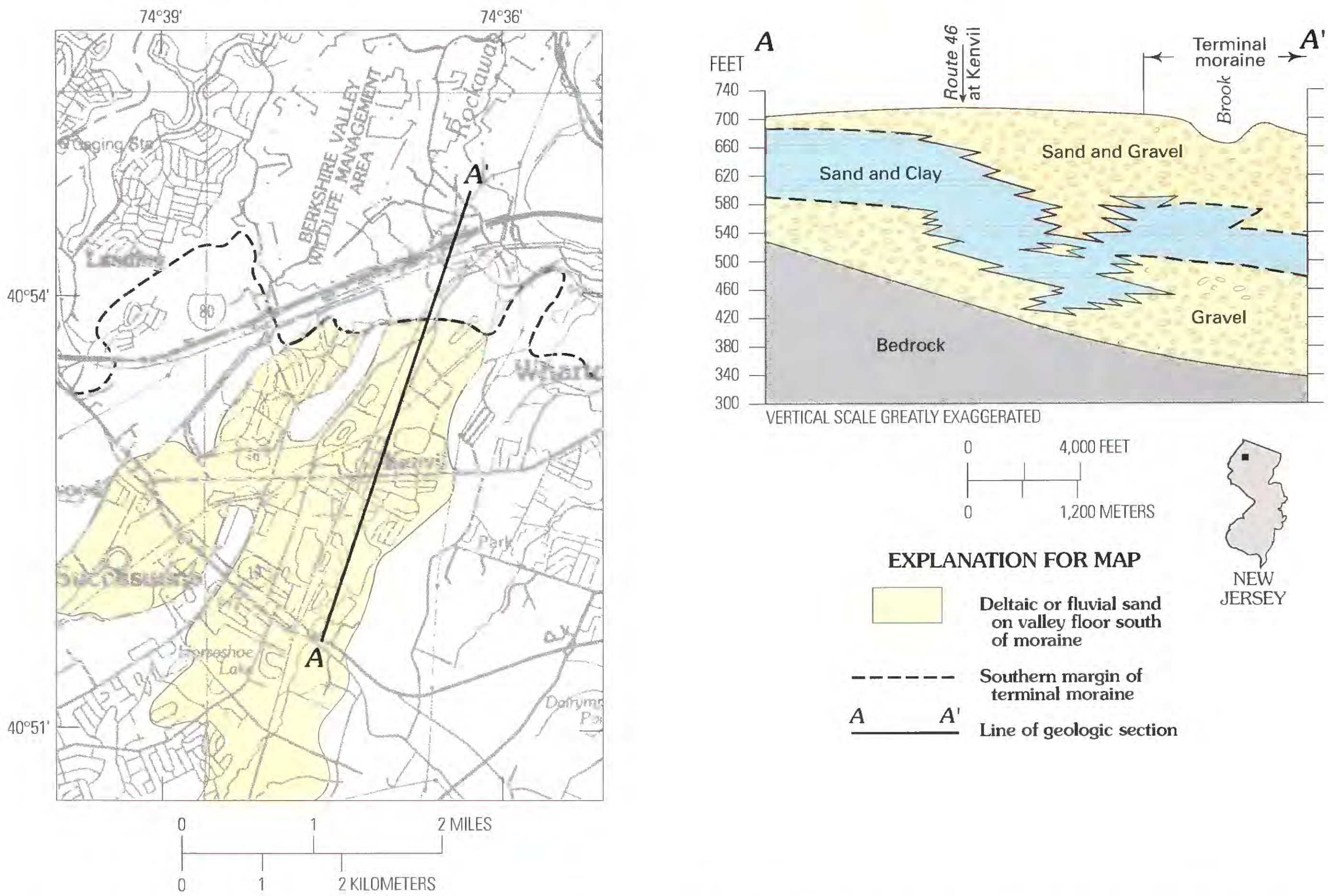


FIGURE 72.—Geologic section from the terminal moraine south through Kenvil, N.J., into the headwaters of the Lamington River valley. (Section from Hill and Pinder, 1981, fig. 25; map modified from Stanford and others, 1990.)

REGIONS OF HIGH RELIEF THAT SLOPED TOWARD THE ICE

As Pleistocene ice sheets advanced into what is now the United States from Canada, they first crossed lowlands bordering the St. Lawrence River and Great Lakes, then faced much steeper terrain that sloped north from the regional divide that separates the St. Lawrence watershed from other North Atlantic Ocean watersheds. Two sections of this north-sloping terrain are noteworthy for their high relief, steep slopes, and deep north-draining valleys; each is treated as a hydrophysiographic region in this paper. The northern rim of the Appalachian Plateau (map unit “N,” pl. 1) extends from northern Ohio and Pennsylvania eastward across central New York; for most of this distance it abruptly borders the Erie and Ontario lowlands and constitutes the first major barrier faced by the ice sheet. Mountains in northeastern New York and northwestern Vermont (map unit “H,” pl. 1) border the Champlain lowland just as abruptly, and the major valleys amid these mountains drain in a northerly direction. In both regions, the major valleys contained long, deep proglacial lakes in which abundant fine-grained sediments were deposited. In both regions, the fines are seldom capped by surficial outwash, although thin, unsaturated terrace alluvium deposited by northward-draining postglacial streams is fairly common. The two regions differ, however, in complexity of their valley fills. In general, drift in valleys draining to the Champlain lowland conforms to the universal single-deglaciation model of three depositional facies. Proximal coarse sand and gravel locally underlie the fine-grained sediments, especially in north-south tributary valleys that were the principal avenues of meltwater flow during deglaciation. Till layers are found amid the stratified drift in a few areas, reflecting minor readvances of an active ice margin, and bedrock crops out locally along stream channels.

Valleys draining northward from the Appalachian Plateau, by contrast, generally contain multiple drift sheets that consist primarily of interlayered diamicton and lacustrine fines, but also include discontinuous lenses of coarse sand and gravel at multiple depths; the drift is several hundred feet thick and bedrock never crops out on the valley floors. These stratigraphic differences between two regions of comparable relief and orientation might be a function of the hard metamorphic bedrock in the mountains bordering the Champlain lowland. Also, the flow of ice and meltwater into those mountains might have been weak during deglaciation because so much ice was being diverted down the St. Lawrence valley then (fig. 43) and (or) because within the United States, flow was primarily southward along the Champlain lowland rather than up tributary valleys. By contrast, valleys in the Appalachian Plateau were the principal avenues of southward flow of both ice and meltwater. These two regions are described in more detail in the following sections.

Another similar section of steep north-sloping terrain that bounds the St. Lawrence watershed is not treated in this paper because it lies in Canada. From New Hampshire to northernmost Maine, the regional divide approximately follows the international border. As a matter of interest, however, stratigraphy exposed in river bluffs along the St. Francis River in Quebec is similar to that along northward-draining Appalachian Plateau valleys. McDonald (1969) described 125 measured sections that collectively consisted chiefly of silt and clay, capped and locally interbedded with diamicton layers and lenses, but also including 30 percent sand with southward-oriented current structures; this stratigraphy is interpreted as reflecting multiple ice advance and retreat in proglacial lakes. The bluffs along the St. Francis River lie about 800 feet below saddles on the watershed divide, comparable to relief along the Appalachian Plateau.

MOUNTAINS BORDERING THE CHAMPLAIN LOWLAND

Active ice in the Champlain lowland blocked the lower reaches of the northwest-draining Winooski, Lamoille, and Mississquoi Rivers of Vermont and the northeast-draining Ausable, Salmon, and Saranac Rivers of New York as deglaciation progressed downstream. As a result, water was ponded far above the valley floors in these watersheds and spilled across a succession of saddles into valleys tributary to the Connecticut and Hudson Rivers. The flow of meltwater was generally southward at a high angle to the orientation of the principal valleys, but parallel to the ice margin and the grain of the bedrock topography.

The stratified drift in most watersheds in the northeastern Adirondacks and northwestern Green Mountains conforms to the universal model of three depositional facies deposited during a single deglaciation. Lake-bottom clay, silt, and very fine sand (facies 2) are widely distributed along the principal valleys and along many tributaries, mantling the lower slopes and generally underlying deltas and the valley floor (Stewart, 1961; Stewart and MacClintock, 1970; Larsen, 1972, 1987b; Hodges and others, 1976a; Craft, 1979; Diemer and Franzi, 1988; Gurrieri and Musiker, 1990, fig. 4). Facies-3 deposits include several prominent deltas built into ice-free proglacial lakes by southward-flowing streams that temporarily carried large volumes of meltwater (Stewart and MacClintock, 1970; Cadwell and Pair, 1991), and many small deltas built by lesser tributaries (Larsen, 1987b, p. 220; Ackerly and Larsen, 1987, p. 380; Diemer and Franzi, 1988, fig. 6). Because lake levels were high above the valley floors, many deltas are perched on the valley sides entirely above stream grade. The deltas generally grade down to fine sand that could be partly saturated, but the coarse sand and gravel are above the modern water table. Streams have incised the deltas and lake-bottom deposits, leaving thin facies-3 terrace gravels at various elevations (fig. 73).

The distribution of facies-1 ice-contact deposits is poorly documented in the northeastern Adirondack Mountains. Gurrieri and Musiker (1990, p. 194) mention subaquatic outwash complexes in what they term the north-central Adirondacks, but do not identify any within the Ausable River watershed. Cadwell and Pair (1991) mapped isolated kame deposits in some northeastward-draining valleys, but none that border and might, in part, underlie younger deltas or lake-bottom fines. Few borehole logs are available; Giese and Hobba (1970) list three wells that tap sand and gravel beneath the toe of a large delta near Clintondale, N.Y.

In the northwestern Green Mountains of Vermont, by contrast, facies-1 ice-contact deposits are widely documented in many north-south tributary valleys that were principal avenues of meltwater flow. Kame terraces choke some reaches of these valleys, especially near saddles, and basal coarse deposits underlie fine-grained facies-2 sediments in other reaches. Basal coarse deposits also are found where subglacial streams flowed southeastward up the principal valleys, away from the Champlain ice lobe, as inferred by Larsen (1987b, p. 220) for a reach of the Winooski valley and as suggested by scattered well records (Hodges, 1967c; Hodges and others, 1976a). The basal coarse deposits are commonly thin and discontinuous, but are as much as 100 feet thick in some north-south tributary

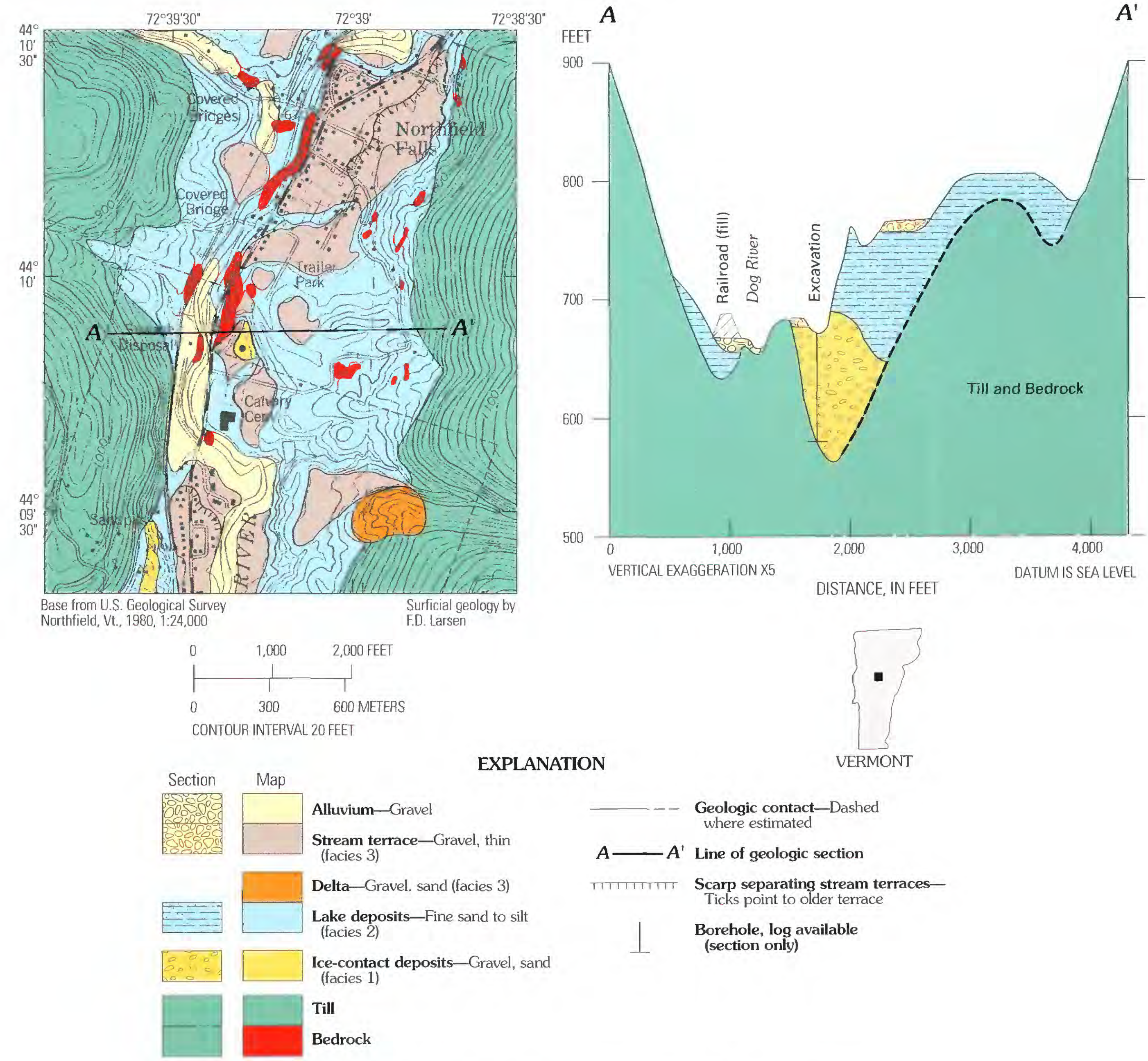


FIGURE 73.—Geologic section across Dog River valley, Northfield, Vt.

valleys. Among the north-south valley reaches in northwestern Vermont that contain proximal facies-1 aquifers are the following:

1. *Dog River Valley.*—The Dog River flows northward to join the Winooski River near Montpelier, Vt. (pl. 1). Exposures and subsurface data near the valley axis (Larsen, 1972, 1987b; Hodges and others, 1976a) indicate that basal sand and gravel is apparently continuous for at least 9 miles, generally buried by fines (fig. 73), and at least partly saturated. The lower part of the basal sand and gravel consists of poorly sorted pebble-cobble gravel interbedded with pebbly coarse to fine sand and silt, and occurs in narrow elongate esker segments; the beds dip south and are collapsed on the margins. The upper part consists of medium to very coarse sand and pebble gravel arranged in

trough-crossbed sets of variable orientation, locally truncated by channels filled with structureless sand that contains sand clasts; it was deposited as subaquatic outwash by density currents near the mouth of a subglacial channel (Larsen, 1987b, p. 224).

2. *Kingsbury Branch and Connecting Valleys.*—Maps by Hodges and others (1976a) suggest that the valley of southward-flowing Kingsbury Branch contains a larger volume of more continuous ice-contact deposits than the parallel valleys of North Branch and upper Winooski River to the west and east. A well-documented esker extends from the valley of Kingsbury Branch southward into the Winooski valley in the town of East Montpelier. The esker has a saturated thickness of 60 feet, a transmissivity of 40,000 feet squared per day, and is bordered and partly buried by fine sand, silt, and clay. Hodges and others (1976a) suggested that the esker may continue on a southward trajectory toward Barre and toward the valley of north-flowing Stevens Branch, which heads at Williamstown Gulf (fig. 21). Near the head of Kingsbury Branch, at Lake Woodbury, kame terraces (Stewart and MacClintock, 1970; Hodges and others, 1976a) commonly crest at 1,000 to 1,040 feet, which is at or slightly above the projected level of the proglacial lake (Lake Winooski) that was controlled by Williamstown Gulf, allowing for postglacial rebound (Larsen, 1987a, p. 37). The ice-contact deposits in this set of valleys are inferred to reflect a persistent subglacial channel system early in deglaciation of this region, as described earlier in the section "Regional Meltwater Drainage Systems."

3. *Mississquoi Valley, Upstream from North Troy.*—This reach of the Mississquoi valley drains north into Canada. Early in deglaciation, melt-water drained south through Eden Notch, a saddle 1,360 feet in altitude at the head of the valley (fig. 24, east of Mt. Norris). Sand and gravel deposits immediately north of the notch were probably deposited by meltwater from a subglacial channel that followed the Mississquoi valley, as discussed earlier in the section "Other Temporary Meltwater Channels in Uplands." Farther north, fine-grained lake deposits are widespread below an altitude of 840 feet, and small deltas or fans occur at this altitude near tributaries. Stewart and MacClintock (1969, p. 171) report that the lake deposits mantle kame gravels, which seem to be products of the earlier subglacial channel system.

4. *Mississquoi Watershed Downstream From East Richford.*—Surficial stratified drift downstream from East Richford, Vt., where the Mississquoi River returns to the United States, consists largely of fine-sandy deltas and lake-bottom fines deposited in valleys ponded at successive levels of Lake Fort Ann, a large water body that occupied the Champlain lowland late in deglaciation. The only extensive coarse ice-contact deposits reflect meltwater flow across local divides early in deglaciation or perhaps following a local readvance (Parrott and Stone, 1972). Ice-contact deposits between West Berkshire and Enosburg Falls, Vt., are tapped by a municipal well (Hodges, 1967b) and head in a segmented esker about 3 miles long. South of Enosburg Falls, kame terraces and eskers extend across a divide near Bakersfield, Vt. (Stewart and MacClintock, 1970; Parrott and Stone, 1972). These two masses of ice-contact deposits are aligned with each other and could reflect successive depositional events along a single meltwater-drainage system, although they are not known to be continuous, and intervening surficial gravel deposits are interpreted as younger (Parrott and Stone, 1972).

Valleys in the north-draining mountain region that borders the Champlain lowland are typically narrow and steepsided. Bedrock outcrops seem to be less common along the stream channels and valley floor than in the adjacent Eastern Mountain region (pl. 1), but in several places, streams have cut short gorges through bedrock spurs on which they were superposed when proglacial lakes drained. Hence, any basal ice-contact sand and gravel aquifers are likely to be bounded by bedrock walls only several hundred feet apart.

Ice retreated from this region by stagnation in segments that were at least a few miles in length; indeed, much of the region may have stagnated at the same time (Schafer, 1967). Evidence of active ice is reported from several localities, however. Numerous exposures of till and deformed strata within lake deposits have been attributed to one or more readvances of the Champlain lobe in the Memphremagog and Mississquoi basins near the Canadian border, where altitude and relief are the lowest in this region and decline northward (Wagner, 1972; Parrott and Stone, 1972). Denny (1974, p. 7–8) described exposures of till overlying deformed deltaic sands along the Saranac and Great Chazy Rivers in New York, also at relatively low altitudes near the Champlain lowland. Till overlying inwash terraces in the headwaters of West Branch Ausable River near Lake Placid was attributed to a minor readvance by Cubbison (1989) and Gurrieri and Musiker (1990, p. 191–2). Several exposures of till overlying deformed lake-bottom fines farther downstream along the Ausable River valley were illustrated and interpreted by Craft (1979) as deposits of local mountain glaciers late in deglaciation. Remnants of the continental ice sheet also may have functioned briefly as local glaciers in a few mountain valleys in Vermont (Wagner, 1970; Connally, 1971). Thus, till layers are abundant amid the stratified drift in parts of this region, and some exposures (for example, Craft, 1979, figs. 11, 29) suggest a complex stratigraphy with more than one drift sheet. More generally, however, the tills seem to be incidental interruptions in a typical 3-facies stratigraphy.

NORTHERN RIM OF THE APPALACHIAN PLATEAU

Nearly all of the Appalachian Plateau in Ohio, Pennsylvania, and New York drains southward to the Ohio, Susquehanna, or Delaware Rivers, but the northern 10 to 30 miles of the plateau drains northward to Lake Erie, Lake Ontario, or the Mohawk River. In this northern rim, valley fills are unusually thick and commonly include multiple drift sheets rather than the single drift sheet with three depositional facies that is typical of valley fills nearly everywhere else in the glaciated Northeast. Because the valley fills are so thick and complex, this region is treated at greater length than most other regions. A review of aquifer geometry in the region as a whole is followed by brief descriptions of several valleys that have been studied in some detail.

All major valleys that drain the north rim of the plateau were deeply scoured by glacial erosion. The Finger Lakes of New York are spectacular examples; the bedrock surface is at least 800 feet below sea level beneath parts of Seneca and Cayuga Lakes (Muller, 1965b; Mullins and Hinchey, 1989). In many through valleys, bedrock a few miles north of the modern drainage divide is 1,000 to 1,500 feet lower than the bedrock saddle, which is at most a few miles south of the modern divide. Such large differences in altitude must have resulted in unusually deep proglacial lakes during both advance and retreat of the ice. These lakes were longer as well as deeper than most lakes in southward-draining valleys and, thus, probably trapped a higher percentage of the fine-grained sediment carried by meltwater. During both advance and retreat, ice tongues must have been partially buoyant but stalled for long periods within the steep north-sloping reach at the head of each valley, which was a barrier to ice tongues just as escarpments farther north on the ridges between these valleys were barriers to the ice sheet as a whole. All these factors favored deposition of sediment, particularly fine-grained sediment. At present, all northward-draining valleys along the northern rim of the Appalachian Plateau contain several hundred feet of drift; some reaches contain as much as 1,000 feet. In general, the bulk of the valley fill consists of interlayered diamictons and fine-grained lake deposits, but in some valley reaches, the latter seem to greatly predominate. Multiple diamicton layers are evident not only within the Valley Heads and Lake Escarpment moraines, but also in reaches to the north and south.

Sand and gravel is much less abundant than diamicton and lake-bottom fines, at least in the upper part of the section, and is apparently absent in some localities. Sand and gravel occurs chiefly as discontinuous lenses of variable thickness that could have been deposited as subaquatic fans by subglacial meltwater and as alluvium in fans or channels of north-flowing streams during interstadial intervals. Tributary deltas are small, perhaps because upland ridges between valleys are too narrow to allow large tributary watersheds and because the steep valley walls below lake level may have favored slumping and subaquatic fans. Surficial sand and gravel is generally limited to the alluvial fans and channel deposits of postglacial streams; saturated thickness seldom exceeds 20 feet and is locally negligible due to Holocene incision. In a few valleys, these thin alluvial deposits have been considered potentially important aquifers in that shallow wells or infiltration galleries might obtain significant yields sustained by seepage induced from streams (Frimpter, 1972; Randall, 1979). At the southern ends of the Finger Lakes, about 50 feet of Holocene aggradation has occurred as a result of isostatic rebound that tilted the lakes southward; this alluvium is mostly organic-rich mud, but gravel aquifers are present where large tributaries reach the valley floor (Kantrowitz, 1970; Crain, 1974). In the northern half of Cattaraugus Creek watershed in Wyoming and Erie Counties, N.Y., and rarely elsewhere, southward-draining tributary valleys contain surficial valley trains of sand and gravel.

Some information suggests that a basal aquifer is commonly present in the deep northward-draining valleys of the Appalachian Plateau. Productive basal aquifers have been penetrated by boreholes in Dale Valley (Randall, 1979) and in Cayuga Inlet, Tully, northern Schoharie, and Cobleskill Creek valleys (fig. 74), as described farther on. Seismic-reflection profiles across each of the Finger Lakes (Mullins and Hinchey, 1989; Mullins and others, 1991) reveal a basal unconsolidated facies that is seismically chaotic (typical of stony, heterogeneous sediments), has a hummocky upper surface, and approaches 300 feet in thickness near the south end of some lakes but thins and becomes younger northward. It is overlain by a layer that is reflection-free (interpreted as massive muds), and then by a set of parallel high-amplitude reflectors (typical of rhythmites) that are interrupted in some valleys by southward-thinning reflection-free wedges (interpreted as massive sandy silt or diamicton deposited rapidly during individual inflow events). At the top of the section are northward-thinning postglacial sediments. All this information seems consistent with the hypothesis that valley fills in this region are largely the product of the last major deglaciation, as is generally true elsewhere. According to this hypothesis, when the Wisconsinan ice reached its maximum thickness and extended into Pennsylvania, it further scoured these valleys, which were already overdeepened by earlier glaciations. A long pause in retreat when the ice sheet fronted on the north margin of the plateau allowed time for deposition of an unusually thick basal facies that included subglacial channel gravels, subaquatic fans, and diamictons of various origins. Any later readvances may have been surges controlled by glacier dynamics in proglacial lakes rather than by climate. Surges would have involved thin ice tongues that advanced and retreated quickly (Teller, 1987), and, therefore, left little coarse stratified sediment. Valley-fill stratigraphy that contains the typical three depositional facies associated with a single deglaciation seems to become increasingly prevalent with distance north of the modern drainage divide. A fourth facies composed of fine-grained, organic-rich Holocene alluvium was deposited locally atop facies 2 or 3 when isostatic rebound reduced northward stream gradients and caused the Finger Lakes to prograde southward (Lawson, 1977; Mullins and others, 1991).

Other information, however, indicates that a large part of the drift in many valleys is older than late Wisconsinan and, thus, contradicts the foregoing hypothesis that the valley fills are essentially a single late Wisconsinan drift sheet. At Millport, N.Y., within the Valley Heads moraine along a deep trough that extends south from Seneca Lake, wood recovered from fine sand beneath till at a depth of about 175 feet was more than 40,270 radiocarbon years in age (P.E. Calkin, State University of New York, written commun., 1989). Just north of the Valley Heads moraine in a trough that extends southeast from Cayuga Lake, wood from the upper part of a thick varved sequence overlain and underlain by till was dated as 41,900 years in age (Bloom, 1972, citing Schmidt, 1947). Preliminary analysis of pollen grains

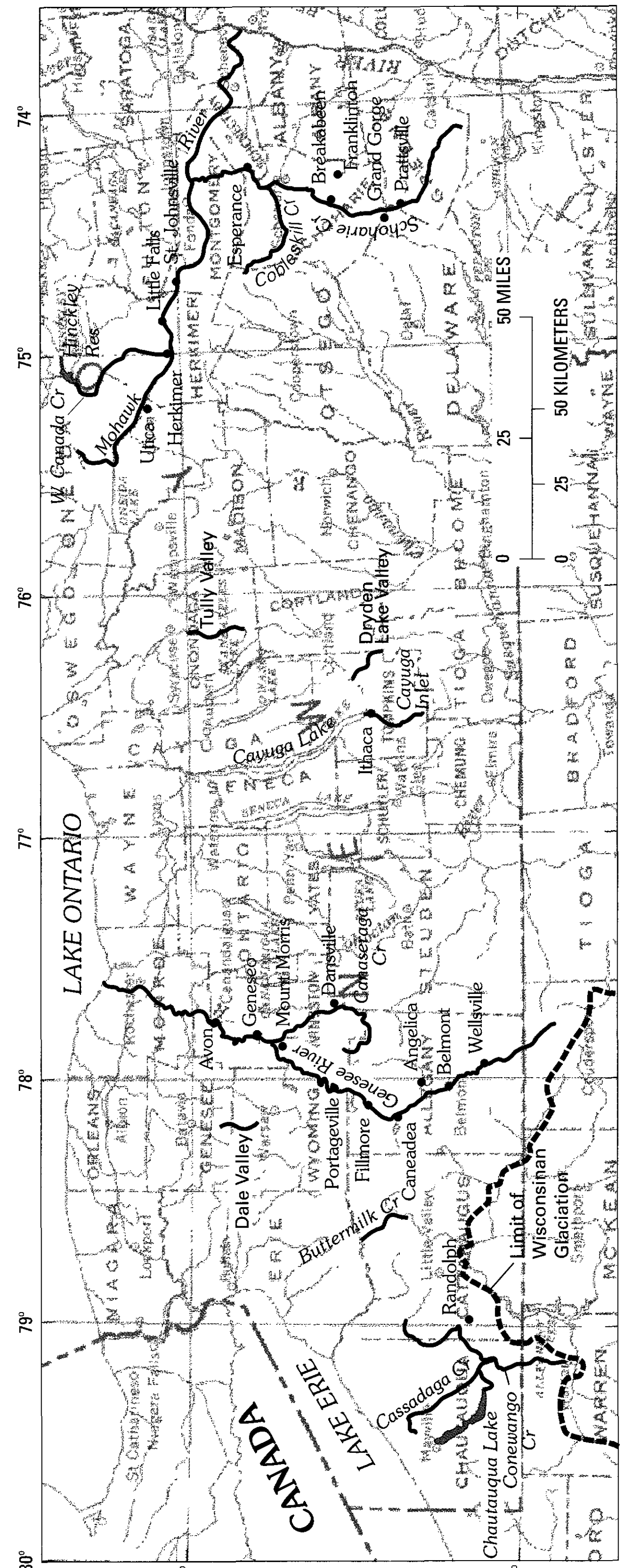


FIGURE 74.—Locations of described or cited valleys in the northern rim of the Appalachian Plateau of New York. Limit of Wisconsinan glaciation modified from Crowl and Sevon (1980), Muller (1977), and Shepps and others (1959).

in borehole samples from Conewango valley suggest mid-Wisconsinan interstadial deposits at a depth of about 130 feet and Sangamon interglacial deposits at about 500 feet (Pefley and Calkin, 1990). Peat at Otto, N.Y., and wood at Collins, N.Y., have been dated as 63,900 and 52,000 radiocarbon years in age, respectively (Muller, 1964; Calkin and others, 1982; Muller and Calkin, 1993); the Otto site lies in a narrow northeast-southwest valley segment that might not have been effectively scoured by ice, but the Collins site lies on the deep Conewango (ancestral Allegheny) trough that is aligned generally parallel to flow of ice out of the Erie basin. Drift 100 to 300 feet thick below the grade of the Genesee River south of Portageville was inferred to predate the late Wisconsinan ice advance to the glacial border (Braun, 1988). The bulk of the till and stratified drift in northeastern Ohio was deposited by the Titusville ice advance in mid-Wisconsinan time and was subsequently mantled by late Wisconsinan drift (White and others, 1969; White, 1982). All of this evidence, taken together, indicates that if ice advanced to the Wisconsinan drift border in late Wisconsinan time, as concluded by Crowl and Sevon (1980) and Muller and others (1988), then that ice did not generally remove preexisting drift from northward-draining valleys along the north rim of the Appalachian Plateau. If the drift in those valleys was deposited incrementally during multiple cycles of ice advance and retreat, coarse proximal stratified drift would not be expected to occur exclusively or predominantly at the base of the section. Borehole records from most of the valleys that are described in more detail below provide no clear evidence that coarse-grained stratified drift is any more abundant at the base of the valley fill than at shallower depths. The upper part of the section in these valleys consists of multiple drift sheets, predominantly diamicton interbedded with lacustrine fines, with subordinate lenses of coarse sand and gravel of variable thickness. A reasonable hypothesis, therefore, is that the deep part of the section was deposited under much the same conditions as the shallow part, except for deeper water that might have favored a higher proportion of fine-grained sediment.

Brief descriptions of stratigraphy in several individual valleys along the northern rim of the Appalachian Plateau follow, arranged from east to west. All but the last of these valleys are labeled in figure 74, as are several places mentioned in the text.

SCHOHARIE CREEK WATERSHED

The watershed of northward-draining Schoharie Creek includes three contrasting terranes, the first two of which resemble the north rim of the plateau farther west:

1. *Upper Schoharie Watershed, South of Breakabeen.*—Valley floors and, in most places, lower valley-side slopes are underlain by till interbedded with lake-bottom fines and scattered lenses of water-yielding sand and gravel. Rich (1935), who studied the southern part of the watershed, mapped the relatively gentle slopes that characterize the lower valley sides as "thick drift" composed mostly of "hardpan-type till," and presented sections and borehole logs that reveal 225 to 280 feet of interlayered till, clay, and fine sand with few possible aquifers. Cadwell and Dineen (1987) mapped some lower valley slopes as kame deposits, despite the general lack of flat-topped morphosequence remnants, and mapped some broad hummocky areas at higher elevation as kame moraine. Soils maps indicate that most areas mapped as kame moraine in Schoharie County are underlain by till, at least at shallow depth (Flora and others, 1969). Wells on the valley floors and lower slopes of Schoharie Creek and most tributaries are reported to penetrate chiefly hardpan (till) with lesser amounts of clay, quicksand (silt or very fine sand), sand, or gravel to depths of 100 to 250 feet (U.S. Geological Survey, unpublished records). North of Prattsville, the lower sides of valleys are generally mantled by fine sand, silt, and clay deposited in proglacial lakes that overflowed at Grand Gorge (altitude about 1,600 feet) and later at Franklinton (1,180 feet); lake fines are not found southeast of Prattsville, perhaps because stagnant ice choked these valley reaches (Cadwell, 1985).

2. *Lower Schoharie Valley (Breakabeen North to Esperance) and Cobleskill Valley.*—Thick, predominantly fine-grained drift also characterizes this region, but lake-bottom sediments predominate over till, and a basal aquifer is commonly present. Lake-bottom sediments, 100 to 300 feet thick and apparently continuous, underlie the valley floors (U.S. Geological Survey, unpublished records) and many segments of the valley sides. These sediments are commonly capped and perhaps interbedded with till reworked from lake-bottom sediments during ice readvance (Darien soils of Flora and others, 1969) and are underlain in many places by deep gravel. The deep gravel might be alluvium that predates the last glaciation, but its upper surface seems to have a steplike downvalley profile that suggests discrete subaquatic fans deposited at successive positions as ice retreated downvalley. In a few localities, gravel layers also occur higher in the valley fill.

3. *Eastern Margin of Schoharie Watershed.*—Several east-side valleys or headwater segments thereof contain much more coarse stratified drift than is evident elsewhere in the Schoharie watershed. Kame moraines, kame terraces, and deltas are banked high on valley sides and across a few low divides (Cadwell, 1985; Cadwell and Dineen, 1987) and were deposited by meltwater that flowed amid stagnant ice eastward out of the Schoharie watershed through upper Schoharie valley, or westward into the watershed through Keyser Kill, Platter Kill, Manor Kill, and perhaps Batavia Kill valleys. Accordingly, these marginal areas are classified as part of the Eastern Mountains Region on plate 1, although they generally lack the bedrock outcrops on the valley floors that characterize this region in New England and south of Greene County in New York.

MOHAWK RIVER AND WEST CANADA CREEK

The Mohawk River drains a rather narrow lowland, carved chiefly in Ordovician shale, that separates the Appalachian Plateau from the Adirondack Mountains. A major tributary, West Canada Creek, drains an extension of that lowland that lies between the Adirondacks and the south end of Tug Hill. During deglaciation, ice lobes repeatedly readvanced into these lowlands from the west, north, and southeast (Ridge and others, 1984). The ice lobes ponded deep proglacial lakes that inundated the lowland and drained southward across the Appalachian Plateau, as did lakes in valleys all along the northern rim of the plateau. The uppermost 100 to 250 feet of drift is exceptionally well exposed in bluffs along West Canada Creek and its tributaries from Hinckley Reservoir to the Mohawk River at Herkimer. These exposures have been thoroughly studied by Ridge (1985). Detailed geologic sections (Ridge and others, 1984, figs. 3, 5) indicate that about 20 percent of the drift is silt-clay rhythmites and 55 percent is till or resedimented proglacial mudflows. In places, thick sections of till are interrupted by thin layers of fine sand or silt that presumably resulted from localized meltwater flow under occasionally buoyant ice. The remaining 25 percent of the drift is described as gravel or as sand in flat beds, dunes, or ripples. Presumably part of the sand is fine to very fine grained, and only 10 to 15 percent of the total volume of drift is coarse and permeable enough to constitute an aquifer. Some sections contain only a few thin gravel units, interlayered with deformed silty sand or diamicton in a manner that suggests final emplacement by mass movements. A few thicker deposits are interpreted as subaquatic fans, the largest of which is described previously in the section "Distribution of Buried Facies 1 Beneath Fine-Grained Sediments," or as deltas (Ridge and others, 1984; Ridge, 1985). Sand and gravel lenses exposed in the bluffs along West Canada Creek drain readily and are essentially unsaturated. A few well records indicate saturated water-yielding gravel below the grade of West Canada Creek, but recharge is likely to be small if these aquifers are surrounded by till and (or) lake clay, as the exposed sand and gravel lenses are.

Surficial deltaic and kamic sand and gravel is widespread north and east of the incised gorge of West Canada Creek (Cadwell and Dineen, 1987) in a marginal part of the Mohawk lowland that was ponded but generally not invaded by the successive ice readvances (Ridge and others, 1984; Ridge, 1985). The bulk of the sand and gravel is inwash from the Adirondacks. Sparse data (Fullerton, 1971, p. 60; Ridge and others, 1984, p. 248, p. 256–257; U.S. Geological Survey, unpublished records) suggest that lake-bottom fines commonly underlie the surficial deltaic sand plains. This marginal locality is classified on plate 1 as part of the mountain-front delta terrane that also borders the Adirondacks farther north.

The Mohawk River valley was ponded and invaded by successive ice lobes, as was West Canada Creek. During at least two periods of ice retreat, however, meltwater from the entire watershed of the proglacial Great Lakes drained eastward toward Rome, N.Y., and down the Mohawk valley (Fullerton, 1980; Ridge and others, 1984). Such extremely large overflows of sediment-free lake water could be expected to incise earlier drift and leave a distinctively coarse channel or lag gravel along the Mohawk valley. Well-rounded cobble-pebble gravel more than 30 feet thick, exposed beneath till and stratified drift in the walls of the Mohawk valley 1.5–5 miles east of Little Falls, has been interpreted as a subaerial fluvial deposit of overflow during the Erie Interstade, 15,000 to 16,000 years ago (Lykens, 1983, p. 27; Ridge and others, 1984; Ridge, 1991, p. D11). Surficial terraces composed of similar gravel as much as 60 feet thick occur at a few wide places in the Mohawk valley farther east and have been ascribed to transport and redeposition by Great Lakes overflow during the latest deglaciation, about 12,500 years ago (LaFleur, 1979a, p. 339; Wall, 1995). None of these gravels have been shown to extend far below modern river grade. The floor of the Mohawk valley west of bedrock outcrops near Little Falls is a wide fluvial terrace that projects to or above the aforementioned gravels, but well records between Utica and

Little Falls (U.S. Geological Survey, unpublished records) do not suggest an unusually productive surficial aquifer. Thus, the episodes of Great Lakes discharge seem to have affected stratigraphy in the Mohawk valley only at elevations above the modern river.

From Little Falls west to Rome, the stratigraphy exposed along the sides of the Mohawk valley includes much more sand and gravel than found farther east or in the bluffs along West Canada Creek. The sand and gravel was deposited in deep proglacial lakes before, and especially after, three readvances of the ice, all subsequent to the Erie Interstade (Ridge, 1991, fig. E1), and was interpreted as esker-subaquatic fan deposits by Ridge (1991, p. E1, E23), who noted that such deposits seem limited to the deepest part of the Mohawk valley—implying that late in deglaciation, subglacial meltwater flow was concentrated along the axis of the deepest valley in the region. When proglacial ponding declined to a low level within the Mohawk valley (only 100–200 feet above the modern river), inwash deltas formed at the mouths of tributaries all along the valley (Ridge, 1991, p. E20), later to be incised by the Great Lakes overflow and modern streams.

Stratigraphy and aquifer geometry in the lowlands along West Canada Creek and along the Mohawk River from Utica to St. Johnsville are idealized in figure 75. Most wells for which records are available (Brigham, 1898; U.S. Geological Survey, unpublished records) between Utica and St. Johnsville are completed in water-yielding sand and gravel beneath 70 to 250 feet of fine-grained sediment that is commonly described as quicksand (very fine sand or silt) or as clay west of Herkimer, and as hardpan (till?) north and east of Herkimer. This stratigraphy is suggestive of the first two facies deposited in so many valleys during retreat of the last ice. Lykens (1983), Foresti (1984), and Ridge (1991, fig. E1) ascribed the drift exposed above river grade along the sides of the Mohawk valley to ice readvances following the Erie Interstade, however, and inferred that those readvances did not generally erode to bedrock. If so, most of the drift below river grade must predate the Erie Interstade and, as illustrated in figure 75, seems comparable to that exposed in tributary valleys—largely till layers and lake beds that reflect multiple ice advances, with localized thick lenses of coarse sand and gravel that originated as eskers or subaquatic fans during episodes of retreat, although their separate identity and limited extent have not yet been defined. Some subglacial meltwater may have followed this deep valley despite its east-west orientation; subglacial flow southward across this valley and up tributary valleys to saddles on the Appalachian Plateau also is plausible (Ridge and others, 1991). Sand and gravel constitutes a large fraction of the drift in the Mohawk valley only near tributaries where it was deposited as deltas late in deglaciation.

TULLY VALLEY

Onondaga Creek flows northward through Tully Valley, a broad trough in the towns of Lafayette and Tully, N.Y. Beneath the flat floor of Tully Valley are a few feet of surficial gravel or sand overlying red clay that overlies and, near tributary streams, interfingers laterally with silty fine sand. The fine sand, in turn, overlies a sand and gravel aquifer, as demonstrated by several borehole records, seismic investigations, and by the history of mudboils near Otisco Road (Getchell, 1982; Mullins and others, 1991; R.M. Waller, U.S. Geological Survey, oral commun., 1978–88; Kappel and others, 1996). The mudboils occur along a short reach of Onondaga Creek and a minor tributary, where hydraulic head in the sand and gravel aquifer is above land surface. Water carries the fine sand upward through the capping clay along fractures that result from collapse after removal of fine sand through earlier mudboils (fig. 76). The high head at depth in the freshwater flow system that feeds most of the mudboils implies hydraulic continuity from sources of recharge on alluvial fans along the valley sides, or possibly on the Valley Heads moraine, 3 miles to the south, whereas the brackish water that is discharged from some mudboils and from the few wells that tap the basal gravel aquifer to the south indicates that saline water enters

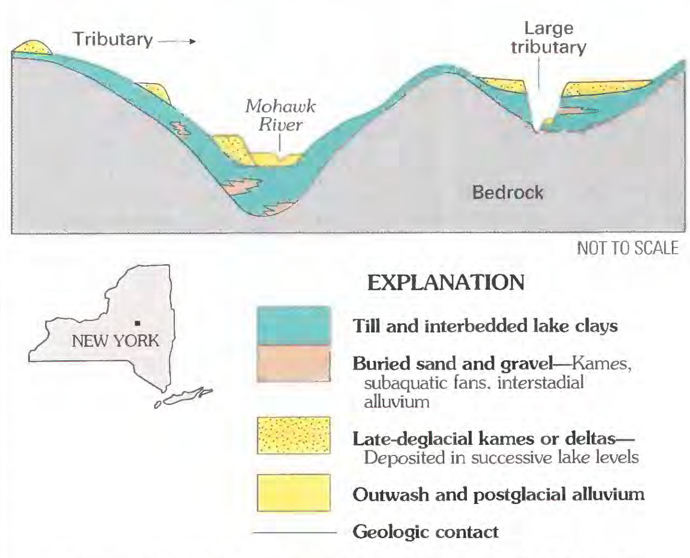


FIGURE 75.—Idealized geologic section across the Mohawk River lowland west of St. Johnsville, N.Y.

FIGURE 76.—Inactive mudboil in Tully Valley, Onondaga County, N.Y. The broad apron of fine sand in the foreground was carried up the vent marked by a pool about 3 feet in diameter at the center of the photograph. Fine sand deposited around this and nearby vents choked and ponded the flood plain of a tributary to Onondaga Creek. Removal of the sand at depth resulted in collapse of land surface along normal faults, two of which are visible as dark horizontal bands at the top and middle of the bank at the right rear. Photograph by F.A. Getchell in March 1981.

the basal aquifer from a regional flow system through the bedrock. A test hole near the mudboils revealed that the basal aquifer consists of 170 feet of sand that ranges from very fine to medium coarse and contains scattered clay laminations, interbedded with 105 feet of sandy, coarse gravel in three layers and, near the top, 12 feet of clayey till (W.M. Kappel, U.S. Geological Survey, written commun., 1994). This stratigraphy suggests a facies-1 subaquatic fan deposited at the end of a retreating ice tongue that readvanced at least once, but conceivably could be the product of multiple ice advances.

DRYDEN LAKE VALLEY

Miller and Randall (1991) and Miller (1993) described the stratigraphy of a northward-draining valley in the towns of Harford and Dryden, N.Y., immediately north of the crest of the Valley Heads moraine. The drift is predominantly till and lake deposits interlayered with less abundant sand and gravel (fig. 77). From land surface to a depth of 100 feet, till constitutes about 50 percent of the drift, lake-bottom fines 20 percent, and sand and gravel 30 percent. The sand and gravel occurs as discontinuous layers that seldom exceed 15 feet in thickness. Between depths of 100 and 300 feet, lake deposits are slightly more abundant than till, and only small amounts of sand and gravel are present. Some sand and gravel layers might be valley trains that were deposited by southward-flowing meltwater streams, partly atop ice whose subsequent melting resulted in collapse. Some might be subaquatic fans deposited in proglacial lakes that were ponded north of the divide, although the upward-fining stratigraphy that normally characterizes retreating subaquatic fans is not prominent. The last glacial event was a readvance that extended more than 1 mile south of the divide, deposited a till layer (D1 in fig. 77) atop valley-train gravel, then dissipated, leaving no lacustrine fines and only thin sand and gravel that is continuous atop the till near the valley axis (fig. 77) but not elsewhere. The absence of surficial fines implies that the ice of the last readvance was thin and stagnated quickly over a long enough distance that meltwater could drain northward through crevasses in the ice to some lower saddle, rather than forming a proglacial lake. Such circumstances would result in a discontinuous layer of ablation till, interlayered or interspersed with fluvial sand and gravel that was derived from melting of the stagnant ice and also from upland runoff, which eventually became predominant. Lithology and stratigraphy of the drift in boreholes and exposures led Miller (1993) to infer that some earlier, deeper gravel layers originated in a similar manner. Other individual stratigraphic units that are interpreted as continuous between boreholes in figure 77 may, in fact, be locally discontinuous.

CAYUGA INLET VALLEY

The broad valley occupied by Cayuga Lake continues southwestward beneath Cayuga Inlet, a principal tributary. Lawson (1977) identified four stratigraphic units beneath the valley floor over a distance of about 5 miles from the lake. The four units are illustrated in figure 78 and described below from oldest to youngest.

1. *Basal Valley Fill.*—Laterally discontinuous layers ranging from diamicton to loose sand to gravel, water yielding in part; may consist of proximal ice-contact sediment; rises or thickens southward enough to suggest correlation with constructional landforms on the valley floor 5 miles south of Cayuga Lake and beyond. Such continuity could explain hydraulic head more than 30 feet above land surface at Ithaca (Crain, 1974).

2. *Proglacial Lake Deposits.*—Clay, 150 to 200 feet thick, with occasional ice-rafted pebbles, deposited in a deep southward-draining lake.

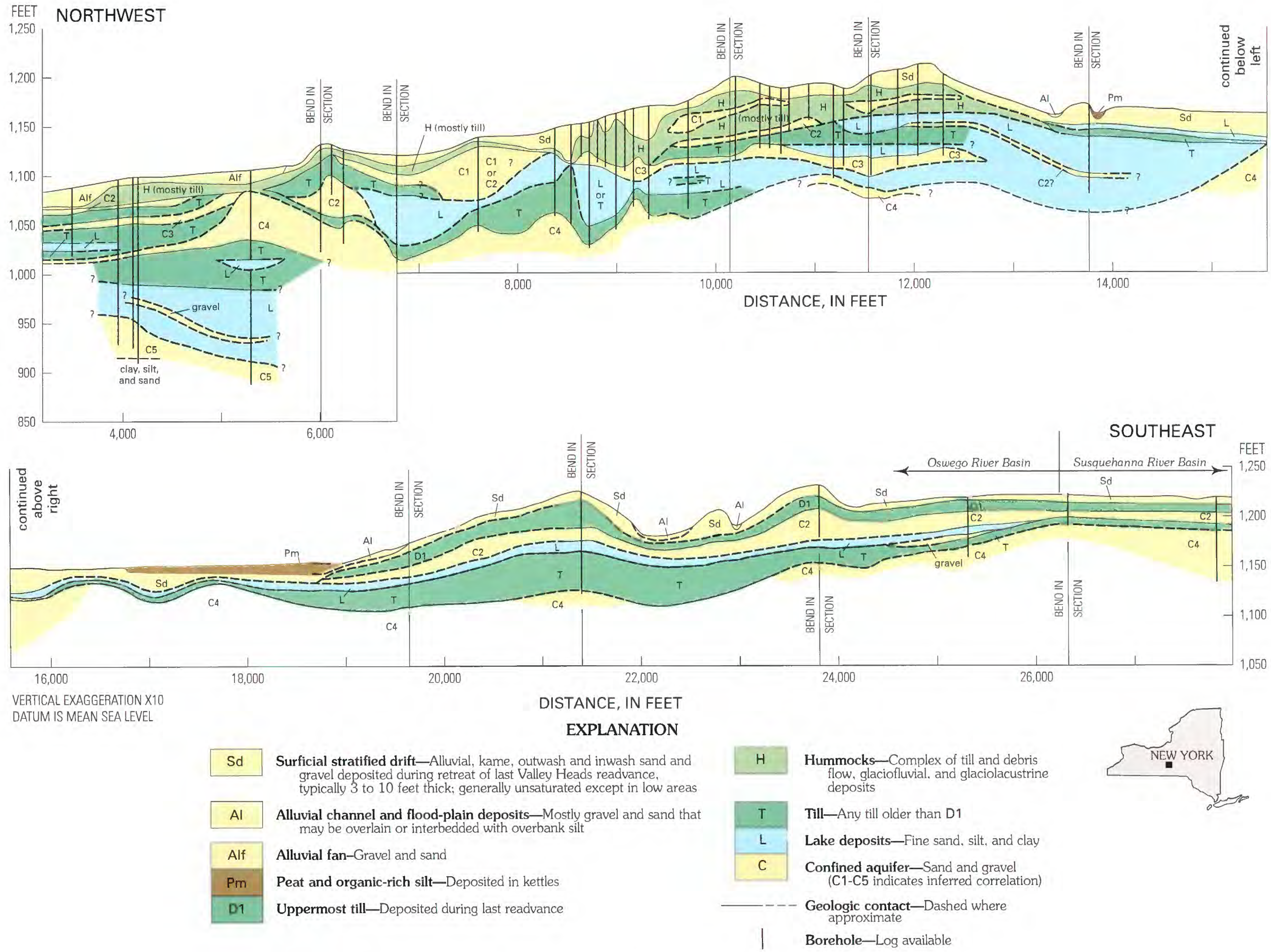


FIGURE 77.—Geologic section along the axis of a through valley in Dryden and Harford, N.Y. (From Miller, 1993). Location of valley shown in figure 74.

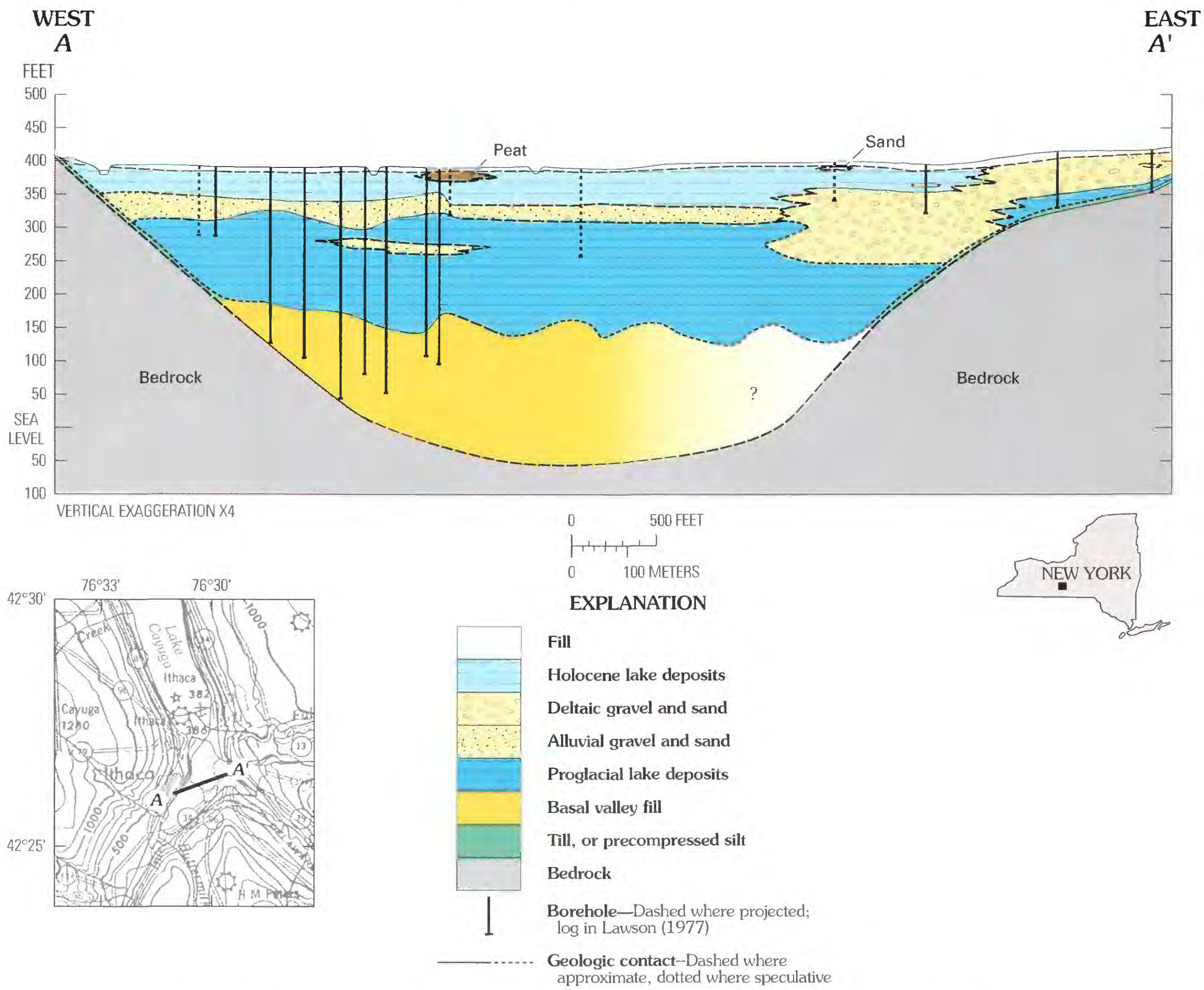


FIGURE 78.—Geologic section across Cayuga Inlet valley in Ithaca, N.Y. (From Lawson, 1977, fig. 23.)

3. *Deltaic and Alluvial Gravel and Sand.*—Coarsening upward, deposited by tributaries after drainage shifted to the north and lake level in Cayuga Inlet valley dropped to that of Lake Iroquois; as much as 100 feet thick near major tributaries, thinner elsewhere.

4. *Holocene Lake Deposits.*—Highly organic silt and clay, about 50 feet thick, deposited after isostatic rebound to the north raised lake level near Ithaca and thereby inundated the alluvial gravel and sand that was graded to Lake Iroquois.

Drillhole logs and geologic sections by Crain (1974, fig. 24D) and H&A of New York (1983) are consistent with the stratigraphy described above. Crain (1974, p. 71) inferred that yields as large as 14 million gallons per day were potentially available from the basal valley fill and the deltaic/alluvial unit.

GENESEE RIVER VALLEY

The Genesee River heads in northernmost Pennsylvania and flows north to Lake Ontario (fig. 74). From Portageville north to Mount Morris, N.Y., the river has carved a postglacial gorge, but south of Portageville and north of Mount Morris, it occupies broad drift-filled valleys. Deglaciation involved a complex chronology of retreat and readvance in which successive proglacial lakes were drained and later reestablished, as described by Muller (1988a) and in more detail by Muller and others (1988).

For at least 30 miles south of Portageville, the Genesee River is incised 100 to 200 feet into interlayered till and lacustrine deposits (fig. 79) whose extent has been traced by Braun (1988). Where the lacustrine units are best exposed, they range from rhythmic silt and clay to fine and very fine sand (Braun, 1988, stops 5 and 6). Sand and gravel is a major component of the stratigraphic section only at the Valley Heads moraine (fig. 79) where it is interbedded with diamicton and locally forms large surficial kame deltas or kames (Braun, 1988, stop 5 and mile 66.1), but some coarse sand and gravel has been reported elsewhere. For example, the river bluff near Caneadea (fig. 74) revealed surficial till and 30 feet of varved clay over till and proglacial gravel (Muller and others, 1988, p. 120), and 12 feet of pebbly fine to coarse sand were exposed near Fillmore in a streambank more than 65 feet high that otherwise consisted of till, silt, and clay (Muller, 1965b). D.D. Braun (Bloomsburg University, oral commun., 1988) recalled that the lake deposits in figure 79 include in a few places sandy units that grade northward to gravel, then abruptly pinch out, suggesting deposition in a short-lived subaquatic fan. Braun (1988) concluded that the lowest diamicton layer exposed along the Genesee valley south of Portageville (DL1 in figure 79) probably represents the recession from the limit of Wisconsinan glaciation (fig. 74), although he had not traced this layer south of Wellsville and could not rule out the possibility that it was the product of a readvance during deglaciation (D.D. Braun, oral commun., 1988). Fine-grained lacustrine deposits interlayered with till continue to depths of 125 to 200 feet below the modern valley floor, as shown by test borings for highway bridges at three sites between Caneadea and Wellsville (U.S. Geological Survey, unpublished records). Bedrock lies 100 to

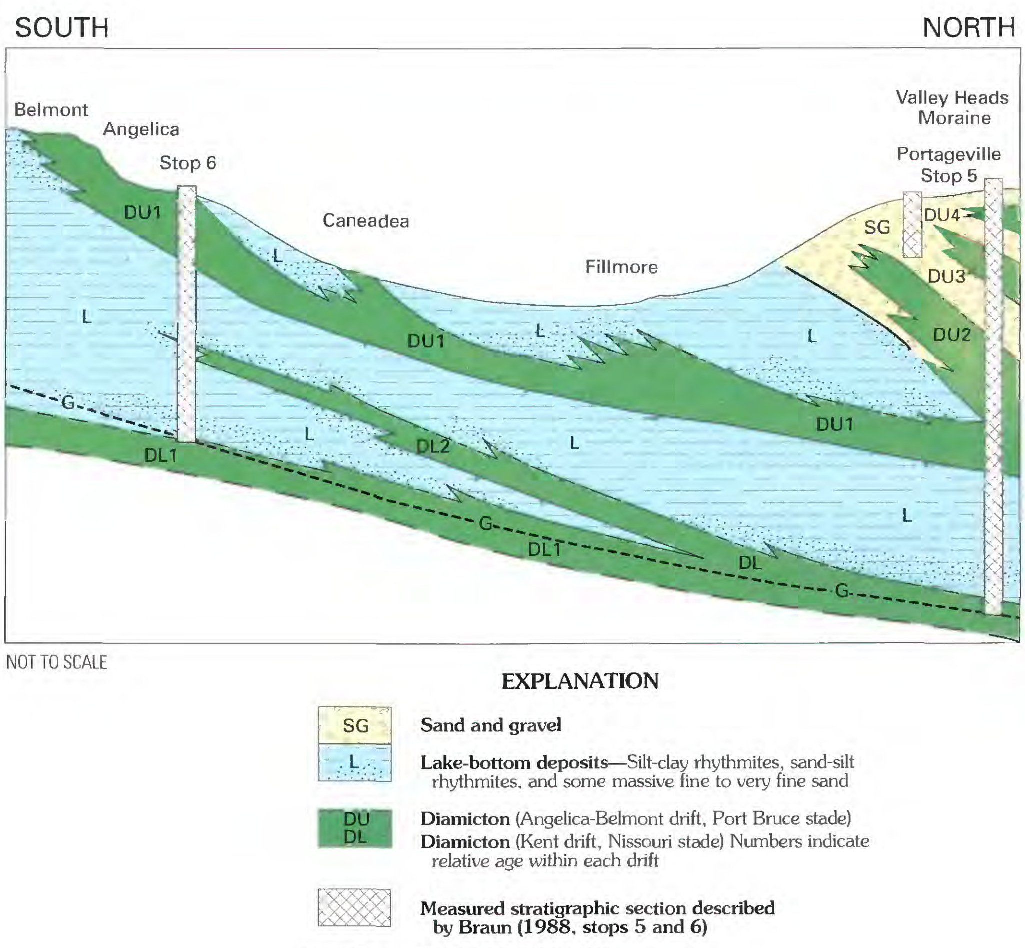


FIGURE 79.—Schematic geologic section along the Genesee River valley from Portageville to Belmont, N.Y. (see fig. 74), a distance of about 41 miles. (Modified from Braun, 1988, fig. 5.)

300 feet beneath the valley floor between Wellsville and Portageville (Kammerer and Hobba, 1967, p. 142). If the lowest exposed till was deposited when ice was at the Wisconsinan glacial border, the drift below the valley floor must represent the period of ice advance or earlier glaciations.

A broad valley extending northward from Dansville is occupied by Canaseraga Creek and, north of Mount Morris, by the Genesee River (fig. 74). Stratigraphy in this valley, as described by Muller and others (1988, p. 124) and illustrated in figure 80, is similar to that south of Portageville, but because depth to bedrock is greater in the Canaseraga-Genesee valley, lacustrine clay to fine sand is more abundant than south of Portageville. Deep boreholes, mostly near where a salt mine beneath the Genesee valley collapsed in 1994 but a few elsewhere (Yager and others, in press; W. Brennan, State University of New York, written commun., 1988), generally penetrate 10 feet or more of water-yielding gravel atop bedrock. The water-level response to the salt-mine collapse showed that this lower confined aquifer is continuous and permeable for more than 15 miles along the valley (Tepper and others, 1995; Yager and others, in press). Stratigraphy in the lower part of the valley fill, above the basal gravel, is complex and poorly documented, but the heterogeneous sediments are largely fine grained and can be interpreted as the product of deposition in a proglacial lake beneath an intermittently floating and (or) calving ice tongue during oscillatory readvance and retreat. The upper part of the valley fill is largely distal silt and clay, intercalated with till from two readvances. Figure 80 is oversimplified in that the upper confined aquifer is probably discontinuous and is hydraulically connected with the lower confined aquifer about 8 miles south of the collapse site.

Kammerer and Hobba (1986) inferred that an artesian aquifer underlies the fine-grained sediment all along both broad reaches of the Genesee valley and along most tributaries. In an earlier report (Kammerer and Hobba, 1967, p. 22), they described this aquifer south of Portageville as a basal layer 5 to 30 feet thick that varies from sand and gravel to fine sand. By contrast, Miller (1988b) inferred a buried aquifer only where several wells are known to tap water-yielding sand and gravel beneath fine-grained sediment. Nearly all the scattered deep wells south of Portageville for which records are available penetrate either clayey gravel or water-bearing gravel, but some of these coarse deposits do not directly overlie bedrock. Fining-upward stratigraphy suggestive of subaquatic fans was recorded by logs of 12 wells (Kammerer and Hobba, 1967) that penetrated thick clay, then silt and very fine sand, then sand and gravel or till. Also, Braun (1988, stop 6) described an exposure in which clay-silt rhythmites overlie sand-silt rhythmites that, in turn, overlie planar-bedded to ripple-marked fine sand.

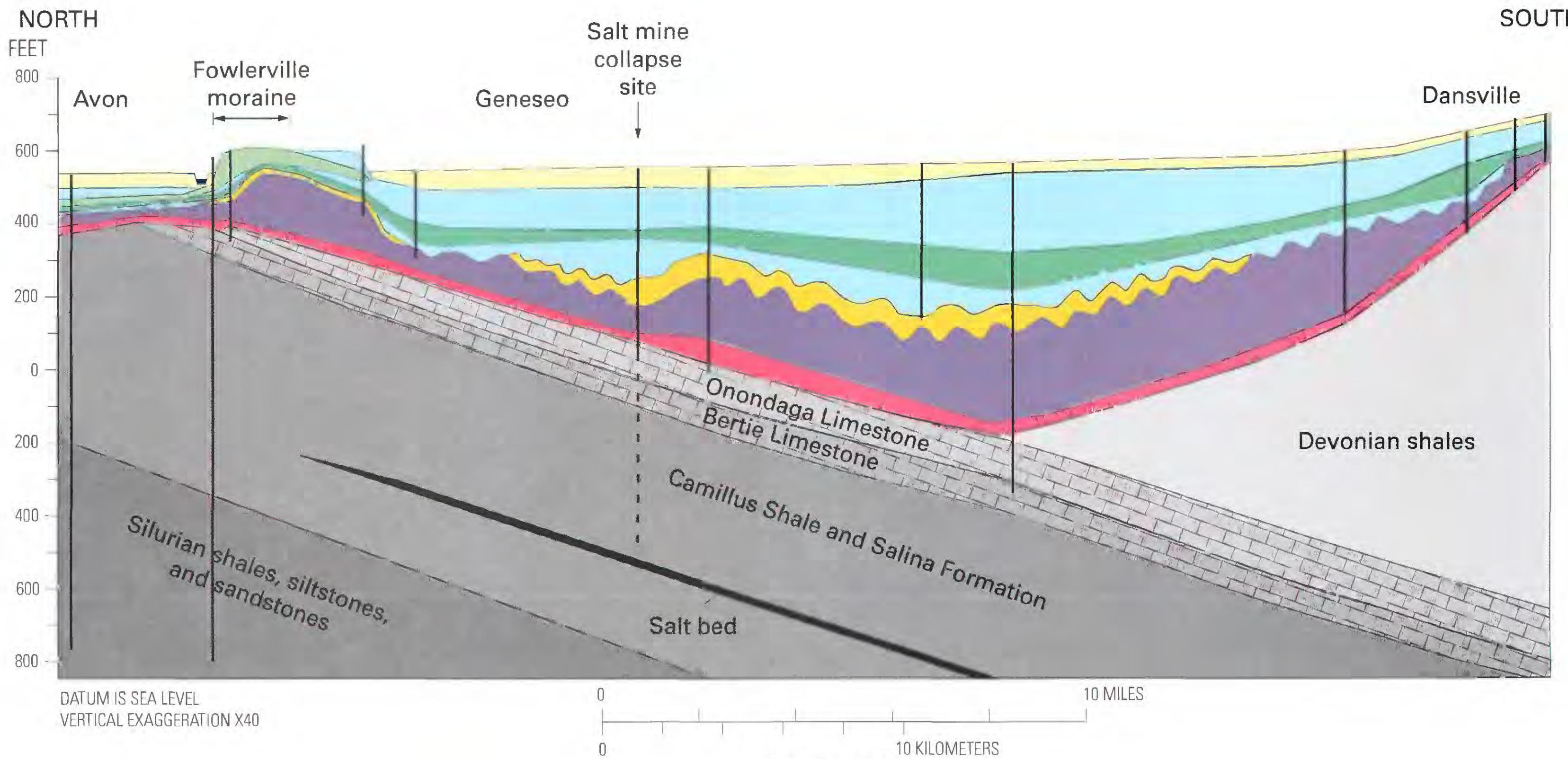
Inwash deltas or subaquatic fans formed where major tributaries enter the Genesee valley (Kammerer and Hobba, 1967, p. 23). Fluvial channel gravels and small deltas might have been deposited by northward-flowing streams on former lake floors when lake levels dropped during ice retreat, later to be buried beneath more fine sediment when readvance temporarily reestablished former lakes. However, the successive lake elevations enumerated by Muller and others (1988) are such that any sand and gravel deposited on exposed valley floors would crop out in the valley walls above the modern Genesee River rather than being potential aquifers beneath the valley floor.

Two parts of the Genesee watershed are excluded from the northern rim of the Appalachian Plateau on plate 1: (1) the area north of Geneseo (fig. 74) that lies within the Lake Ontario lowland, and (2) the area south of Wellsville and on the east side of the watershed somewhat farther north, all of which is treated as a part of the Eastern Appalachian Plateau because valleys in this area were not deeply inundated by proglacial lakes, generally received meltwater overflow from adjacent watersheds, and contain considerable sand and gravel deposited amid stagnant ice as kame terraces or deltas.

BUTTERMILK CREEK VALLEY

The geology and hydrology of the Western New York Nuclear Service Center, which lies within the valley of Buttermilk Creek in the town of Ashford, N.Y., have been studied in considerable detail. Reports by LaFleur (1980), Boothroyd and others (1982), Albanese and others (1984), Fakundiny (1985), Prudic (1986), Bergeron and others (1987), and Matuszek (1988) summarize several major investigations and cite earlier papers. The site is about 5 miles north of the divide between southward and northward drainage from the Appalachian Plateau. Oscillatory ice-sheet retreat (Muller, 1977, fig. 1a) is reflected in a succession of ice-marginal positions (LaFleur, 1979b, fig. 2).

The drift in Buttermilk Creek valley consists chiefly of till and glaciolacustrine silt or clay, as illustrated in figure 81. The uppermost layer (unit 4 in fig. 81) is a discontinuous blanket of alluvial sand and gravel that was spread across parts of the gently sloping valley floor by Buttermilk Creek and its tributaries soon after ice retreat. The uppermost till (unit 5) thins to the south and overlies a few feet of proglacial rhythmites; successively lower units include interstadial deltaic or fluvial gravel and sand, lake-bottom deposits that become increasingly fine with depth, another till, older



EXPLANATION

- **Postglacial alluvium**—Gravel near Dansville and in alluvial fans near valley sides; elsewhere silt, silty clay, fine sand, peat, underlain in part by sand or gravel
- **Lacustrine deposits**—Mostly silt and clay, grading to sand along the valley sides near tributaries
- **Fowlerville moraine till**—Clayey matrix with embedded pebbles
- **Till**—Silt and clay with embedded pebbles; locally underlain by a few feet of sand or gravel
- **Upper confined aquifer**—Silty sand and gravel, probably discontinuous, perhaps the product of subaquatic fans
- **A complex, heterogeneous unit, predominantly fine-grained**—May include till, proglacial sediment flows, lacustrine fines containing coarse englacial sediment rainout from floating ice, and subaquatic fans
- **Lower confined aquifer**—Sand and gravel, deposited by meltwater in subglacial channels and (or) proximal subaquatic fans
- **Well, log available**

FIGURE 80.—Geologic section along the Genesee River and Canaseraga Creek valleys from Avon to Dansville, N.Y. (see fig. 74). (Modified from Yager and others, in press.)

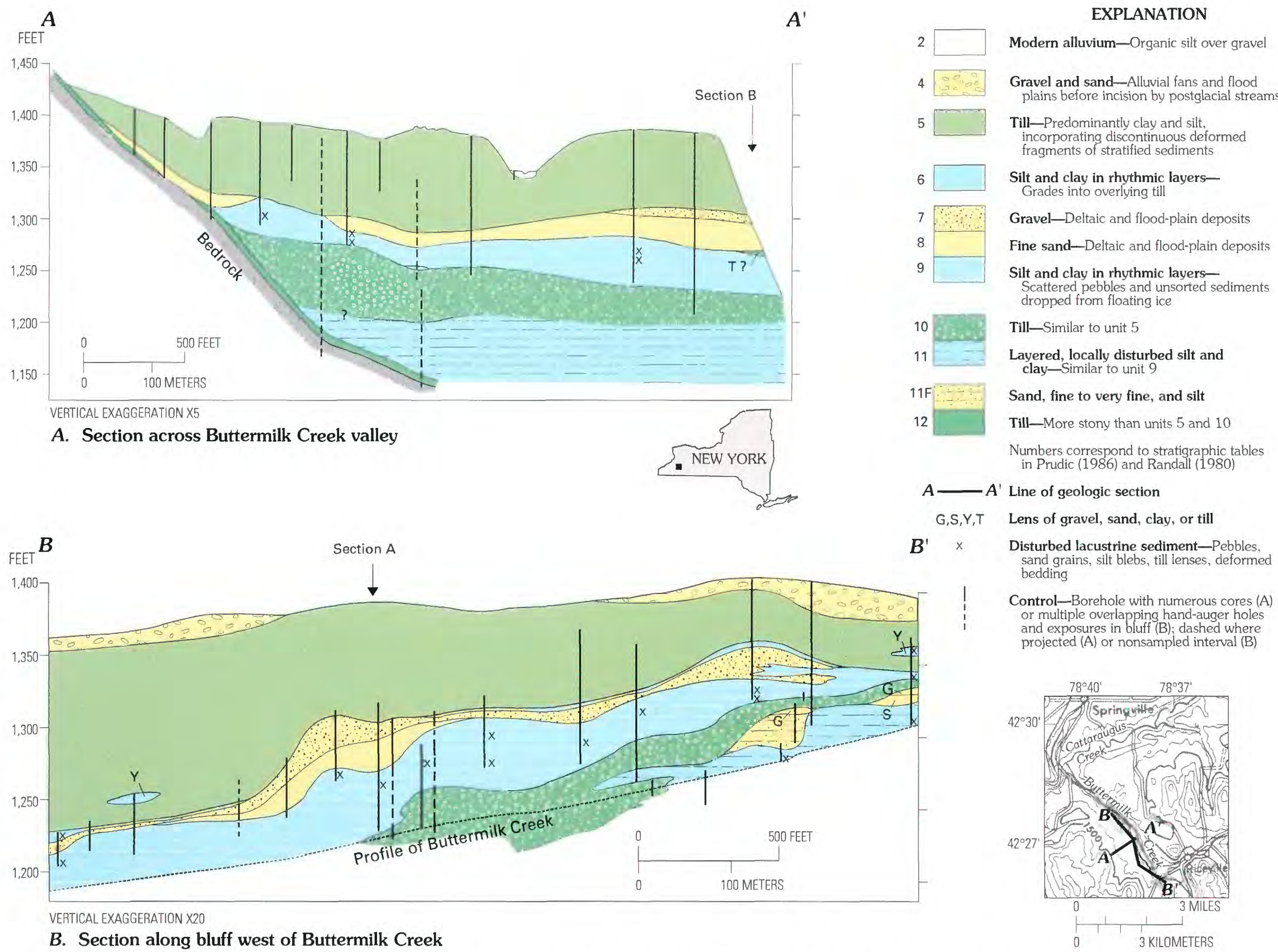


FIGURE 81.—Geologic sections in Buttermilk Creek valley, Ashford, N.Y. (Modified from Randall, 1980.) Locations and records of test borings and measured exposures are given in Bergeron (1985), Bergeron and others (1987), or Randall (1980).

lake deposits, and finally, a basal stony till atop bedrock. The uppermost till typically consists of about 50 percent clay, 25 percent silt, and 25 percent sand and stones. It interfingers randomly with a secondary facies that is similar in grain size but contains many tiny pellets of coarse silt and torn, wispy fragments of silt layers. Both facies contain randomly oriented, deformed, discontinuous pods or lenses of sand, gravel, and rhythmic silt and clay that appear to be transported from their place of origin. The rhythmites (unit 6) beneath the uppermost till were deposited in a proglacial lake ponded by the readvancing ice. The underlying sand and gravel (units 7 and 8) would constitute an aquifer if they were saturated, but postglacial incision by Buttermilk Creek (fig. 81) allows them to drain, and replenishment through the overlying till is very small (Prudic, 1986). The stratigraphy, and a few sedimentary structures that indicate northwestward flow (Randall, 1980), suggest that units 7 and 8 could have been deposited by interstadial Buttermilk Creek along its channel and as deltas into persistent late-deglacial lakes. A deep well not shown in figure 81 penetrated an additional 250 feet of clayey silt below the level of Buttermilk Creek, then several feet of gravel atop bedrock that might constitute a basal aquifer along the thalweg of the bedrock valley; the gravel yielded only 20 gallons per minute, however, and has not been recognized elsewhere.

CONEWANGO CREEK AND TRIBUTARY VALLEYS

The broad valley north of Randolph, N.Y., now occupied by upper Conewango Creek, was once occupied by the Allegheny River, which originally flowed northward to Lake Erie but was blocked and diverted to its present course by some early ice advance (Muller, 1963). Illinoian and Wisconsinan ice sheets later advanced to or nearly to the present Allegheny valley (Muller, 1977). As the ice sheets advanced and retreated, water was ponded in the long, deep valley of Conewango Creek and the parallel valleys of two principal tributaries, Chadakoin River (Chautauqua Lake) and Cassadaga Creek. These lakes served as sediment traps in which great thicknesses of mostly fine-grained sediment accumulated. Drift thicknesses of 500 to more than 1,000 feet have been documented by scattered well records in Cassadaga and Conewango valleys (Muller, 1963; Crain, 1966; Frimpter, 1974; Pefley and Calkin, 1990). Although most of the drift seems to be lake-bottom clay and silt, some evidence of intercalated till layers has been reported. Crain (1966) presented graphic logs of many wells in the Conewango Creek watershed, most of which record till only atop bedrock, but four wells in Cassadaga Creek valley penetrated 20 feet or more of till overlain by clay and underlain by more clay or by gravel. Pefley and Calkin (1990) interpreted gamma-ray logs and washed samples from boreholes in Conewango valley as recording silt-clay layers alternating with fine-grained diamicton. Wood at a depth of 583 feet (Muller, 1963, p. 49) and abundant deciduous pollen at a depth of 500 feet (Pefley and Calkin, 1990) imply interglacial conditions and, thus, multiple drift sheets. Two miles north of the divide at the head of Conewango Creek valley, along the ancestral Allegheny River valley, four till layers were exposed in a 125-foot-high exposure (Muller, 1960, p. 7), interbedded with silt and with sand and gravel.

Ice-contact stratified drift, commonly interbedded with till, borders the three major valleys in some places, generally as terraces 100 to 150 feet above the valley floor. Crain (1966, p. 86, p. 97, p. 102) noted that many of these ice-contact deposits are perched above stream grade, are thinly saturated, and discharge by springs at the base of the deposit. More productive aquifers occur beneath the valley floor as fan-shaped wedges of sand and gravel that immediately underlie the channels of major tributaries where they enter broad valleys. The wedges become thinner and lower in elevation toward the center or far side of the valley (see fig. 30). Many are overlapped by surficial clay that, under natural conditions, confines water under artesian heads many feet above the valley floor. Crain (1966) referred to these wedge-shaped gravel deposits as deltas. In Cassadaga valley and part of Conewango valley, these deltas merge into the Jamestown aquifer, a tabular body of sand and gravel 10 to 40 feet thick that slopes northward along the valley axis for at least 11 miles (fig. 82), overlain by 80 to 140 feet of silt and clay and underlain by more of the same. Crain (1966) and Frimpter (1974) inferred that these deltas (or fans) and the Jamestown aquifer were largely deposited at a time when ice retreated far enough to allow proglacial lakes to drain and streams to flow northward along the valleys; much of the sand and gravel were presumably derived from incision of the ice-contact deposits that lined the valley wall and choked some tributary valleys. Later, ice readvanced and reestablished the lakes; silt and clay accumulated across the valley floor and buried the fluvial gravels, except where tributaries entered the lakes and continued to deposit coarse bedload. The stacked layers of coarse sediment thus deposited near tributaries now allow water to infiltrate from tributary stream channels to the buried aquifer system (see fig. 30). Alternatively, the Jamestown aquifer might have been deposited as subaquatic fans from a retreating ice tongue, but the scant evidence for till below the aquifer or upward-fining sediment within it and the increased thickness near tributaries argue against this hypothesis. The only suggestion of a basal aquifer is in Cassadaga valley, where logs of 10 deep wells (Crain, 1966, fig. 43) indicate that the silt and clay beneath and beyond the Jamestown aquifer and associated deltas overlie heterogeneous interlayered silty sand, gravel, and diamicton.

Even though Conewango, Cassadaga, and Chautauqua Lake valleys drain south at present, they are classified as part of the northern rim of the Appalachian Plateau on plate 1 because of the great drift thickness, the likelihood of multiple drift sheets, the presence of moraines to the south and north, and the lack of surficial outwash. Lacustrine silt and clay seem to be more abundant relative to till here than in other valleys on the northern rim. Surficial outwash, such as commonly found in southward-draining valleys in the Appalachian Plateau, is found atop the fines only near the north ends of Cassadaga and Conewango valleys, where it is the product of late ice readvances (Muller, 1963, p. 45; 1977; Crain, 1966; Frimpter, 1986).

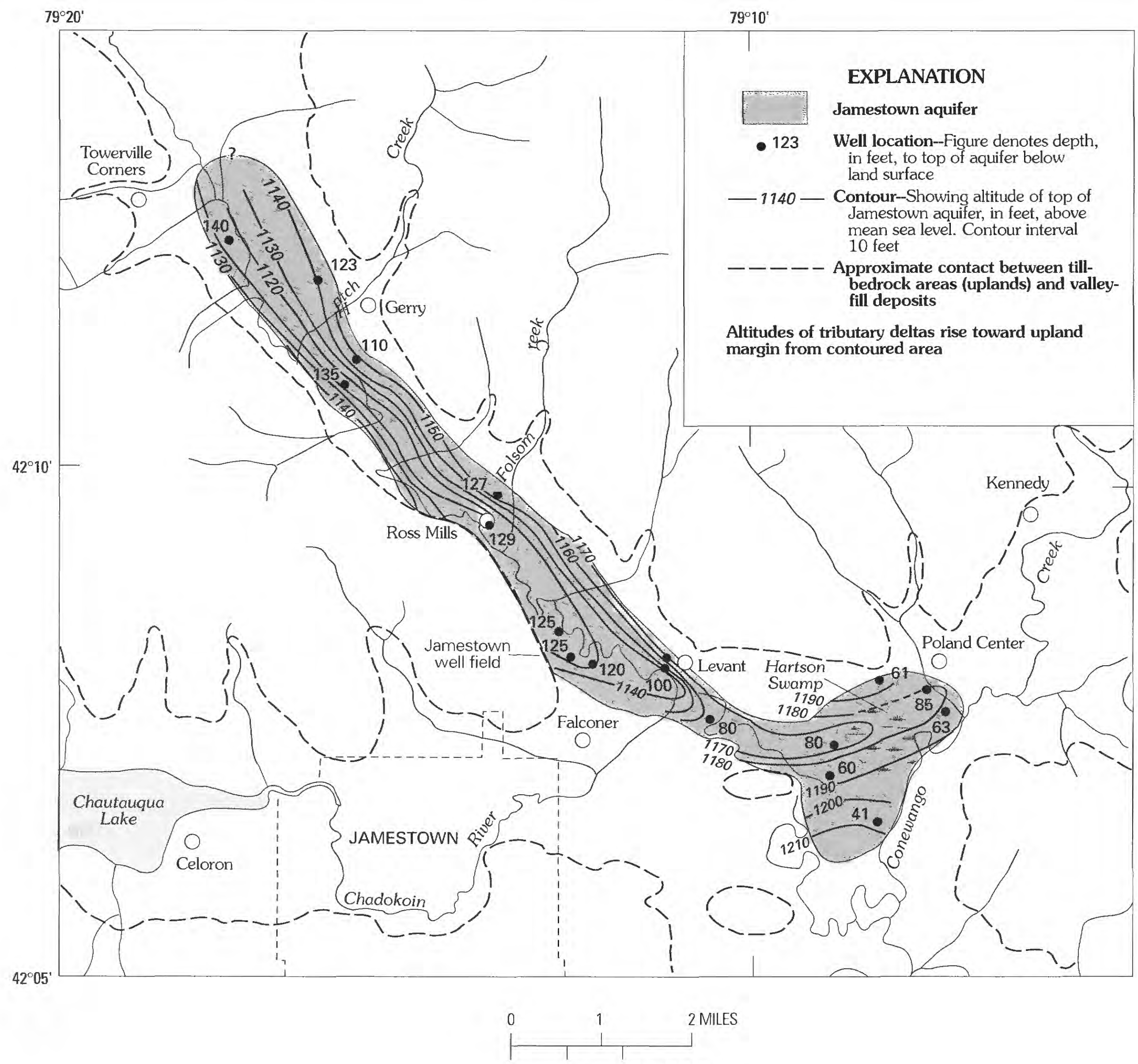


FIGURE 82.—Areal extent of the Jamestown aquifer and altitude of its upper surface. (From Crain, 1966, fig. 16.)

CUYAHOGA RIVER VALLEY

The Cuyahoga River flows north from Akron to Cleveland, Ohio (pl. 1), in a postglacial gorge incised some 300 feet below the surface of the drift that fills a broad bedrock valley. The walls of the gorge are intricately gullied and composed chiefly of fine-grained lake-bottom sediments interbedded with till sheets, whose extent is described in several theses cited and summarized by Szabo and Miller (1986). Also exposed along the valley walls are resedimented diamictons, loess, and several masses of sand interpreted by White (1984, p. 21) and Szabo and Miller (1986) as deltas built by upland tributaries, although the low altitude and upward fining in one such deposit (Szabo and Miller, 1986, stop 4) suggest a subaquatic fan fed by subglacial meltwater. The valley fill has been interpreted as generally not water yielding except for scattered lenses of sand and gravel (Smith, 1953; Crowell, 1979; Schmidt, 1979). A few wells yield from 15 to as much as 700 gallons per minute from sand or gravel 100 to 500 feet below the valley floor (Smith, 1953, p. 49; Crowell, 1979; Schmidt, 1979), but whether these wells tap a continuous basal aquifer or scattered lenses such as exposed in the valley walls is not clear.

REGIONS AT OR BEYOND THE WISCONSINAN DRIFT BORDER

Two of the hydrophysiographic regions delineated on plate 1 are at or beyond the Wisconsinan drift border and are distinctive because of that location.

NEW JERSEY TERMINAL MORAINE

The "terminal" moraine that lies just north of the farthest extent of the late Wisconsinan ice sheet in New Jersey is associated with an aquifer geometry that differs from that in nearby regions in at least two characteristics:

1. A network of bedrock valleys contains drift that predated the arrival of the late Wisconsinan ice but was not removed by that ice. Widespread sand and gravel of Illinoian age are overlain by deltaic outwash and (or) lake-bottom fines that were deposited ahead of the advancing Wisconsinan ice sheet and were capped and confined by thick till beneath and somewhat north of the terminal moraine.
2. The drift is thick enough locally to fill and conceal parts of the preglacial bedrock valleys.

Both aspects of aquifer geometry also are typical of many valleys in the Western Appalachian Plateau but not farther east. Therefore, a small, separate Terminal Moraine region is delineated on plate 1.

The buried coarse-grained Illinoian and late Wisconsinan stratified drift together constitute a highly productive aquifer 30 to 50 feet thick in eastern Morris, southwestern Essex, and northern Union Counties. These aquifers have been extensively developed and described (Vecchioli and Nichols, 1966; Vecchioli, Nichols, and Nemickas, 1967; Meisler, 1976; Stone and Hoffman, 1993). Farther west in Morris County, Stanford (1989a, b) depicted morainal and nonmorainal till draped over late Wisconsinan deltaic sand and lake-bottom fines and pre-late-Wisconsinan sand and gravel in several preglacial valleys that were locally filled sufficiently to divert the modern Rockaway River or its tributaries.

VALLEY TRAINS SOUTH OF THE DRIFT BORDER

Nearly everywhere along the Wisconsinan drift border, streams large and small carried meltwater and sediment beyond the maximum extent of the ice sheet. In eastern New Jersey, Long Island, and Martha's Vineyard, where relief is low, outwash plains composed of sand and fine gravel mantle most of the landscape between the drift border and the ocean a few miles away. West of central New Jersey, where relief is higher, outwash beyond the drift border is confined to valley trains many miles in length along the Delaware, Susquehanna, Allegheny, Ohio, and Muskingum Rivers and their southward-draining tributaries. Valley trains aggraded to thicknesses of 100 to 300 feet immediately south of the drift border in all these valleys, but thinned southward and were deeply incised by later meltwater that formed lower valley trains and by postglacial streams. Many terrace remnants along the valley sides are largely unsaturated, but 40 to 50 feet of saturated sand and gravel remains below river level in many places along valley trains in Pennsylvania; for example, along the Allegheny and Ohio Rivers above and below Pittsburgh (fig. 83), along the Susquehanna River near Williamsport, Berwick, and Sunbury (Lohman, 1937, p. 197; Lloyd and Carswell, 1981; Williams and Eckhardt, 1987), along tributaries of the Lehigh River near the Wisconsinan drift border in Carbon County (Epstein and others, 1974), and along the Delaware River as far south as central Bucks County (U.S. Geological Survey, unpublished well records). More than 100 feet of saturated outwash is present everywhere along several valley trains tributary to the Muskingum River in Ohio (Lamborn, 1954, 1956). These deposits in Pennsylvania and Ohio are productive aquifers tapped by many large-capacity wells (Leggette, 1936; DeLong and White, 1963; Norris, 1969; Lloyd and Carswell, 1981) and also are important as reservoirs that sustain the yield of wells tapping the underlying bedrock (Williams and Eckhardt, 1987, p. 28, fig. 17).

In theory, a valley train consists of gravel grading southward to uniform sand that gradually becomes finer with increased distance from the drift border. The sand is not interbedded with much silt and clay because ponds seldom develop in valleys occupied by aggrading, braided streams, although valley reaches that nearly parallel the drift border could have been locally ponded by ice tongues or by abundant proximal outwash descending tributary valleys (for example, Sandy Creek south of Canton, Ohio, as mapped by White, 1982). Reports of gravel or silt beneath sandy outwash in other valley reaches probably reflect ice

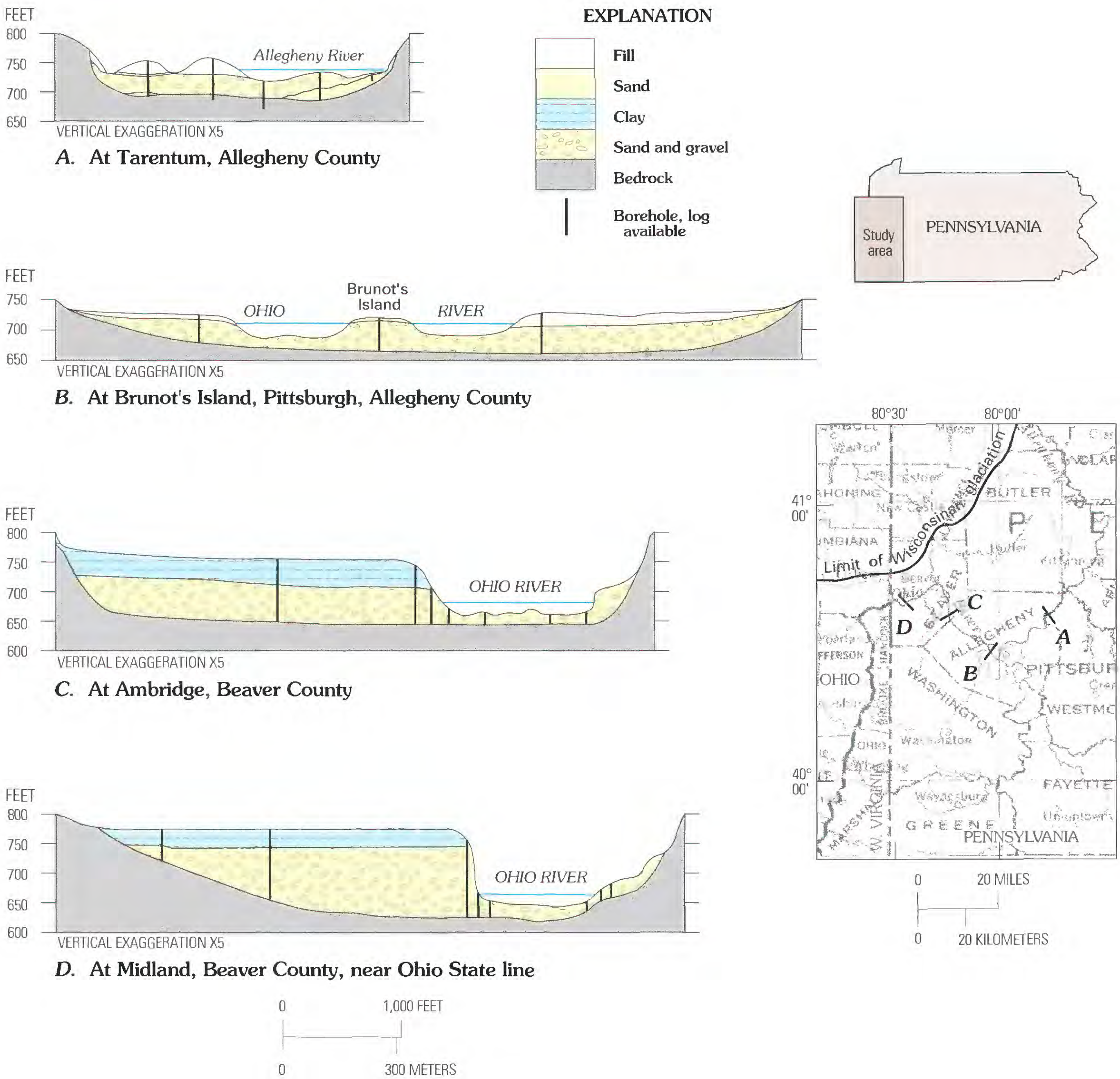


FIGURE 83.—Geologic sections across the Allegheny and Ohio River valleys in Pennsylvania. (From Adamson and others, 1949, figs. 3–4, and Van Tuyl and Klein, 1951, fig. 4.)

advances beyond the Wisconsinan drift border. Late Wisconsinan ice advanced briefly beyond the recognized drift border in some localities (for example, Ridge, 1983, cited in Evenson and others, 1985, p. 17). Pre-Wisconsinan ice sheets also spread somewhat beyond the Wisconsinan drift border in several localities, as indicated by weathered till and by gravel terraces higher than any Wisconsinan valley train (Peltier, 1949; Lockwood and Meisler, 1960; Goldthwait and others, 1961; White and others, 1969; Epstein and others, 1974; Marchand and others, 1978; Evenson and others, 1985). Peltier (1949) mentioned several localities along the Susquehanna River valley in which weathered pre-Wisconsinan gravel was exposed beneath a late Wisconsinan terrace surface. Thus, some of the stratified drift beneath the Wisconsinan valley trains could be outwash or heterogeneous ice-contact deposits of Illinoian or older age.

Streams that head south of the Wisconsinan drift border could not have carried Wisconsinan outwash; nevertheless, where they joined the large valley trains, their valleys were aggraded to the same depths with sediment of local origin. Silt and clay deposited in open lakes ponded by ice (Preston, 1950) or by aggrading outwash (Jacobson, 1983) have been reported in tributary valleys in western Pennsylvania. Most tributary valleys, however, were apparently filled with alluvial sediment at rates that approximately kept pace with aggradation of the valley train. The process was described by Shaw (1911, cited in Leggette, 1936, p. 30) as follows: when aggradation in a principal valley began to block the mouths of tributaries, each tributary dropped the coarse part of its bedload near its mouth, where the gradient was reduced. As aggradation continued, the location of coarse-sediment deposition migrated up each tributary to form a basal gravel layer, while the fine-grained alluvium accumulated downstream. The alluvial load of streams near the drift border was abnormally large because preglacial conditions created abundant rubble on upland hillsides and favored downslope movement by solifluction (Denny, 1956b). This Pleistocene alluvium forms a significant and productive aquifer in some tributary valleys (Lohman, 1939, p. 96, p. 153–187; Adamson and others, 1949) but is siltier and generally finer grained than the valley trains (Shaw, 1911; Adamson and others, 1949; Lamborn, 1954, p. 26), and saturated thickness decreases quickly up the tributary valleys (Shaw, 1911; Leggette, 1936; Poth, 1973).

A COMPREHENSIVE APPROACH TO APPRAISAL OF THE HYDROGEOLOGIC FRAMEWORK OF STRATIFIED-DRIFT AQUIFERS

An understanding of the three-dimensional configuration and boundaries of aquifers is prerequisite to development and protection of those aquifers. This section summarizes how the hydrogeologic framework in which stratified-drift aquifers occur can be defined in any locality of interest, with emphasis on the ideas presented in this and other papers prepared as part of the Northeast Glacial Aquifers Regional Aquifer-System Analysis project. Once the hydrogeologic framework has been reasonably well defined, aquifer recharge and yield can be evaluated by measurements, computational procedures, and ground-water flow models, as described by Kontis, Randall, and Mazzaferro (in press) and by Lyford and Cohen (1988).

AREAL EXTENT OF STRATIFIED DRIFT

The first step in most geohydrologic investigations in the glaciated Northeastern United States is to distinguish those areas underlain by stratified drift, which may constitute or contain highly productive aquifers, from areas where the only significant aquifer is bedrock, mantled by till of variable thickness and generally capable of only small well yields. This distinction, qualified in various ways, is depicted on a few maps at scales of 1:1,000,000 or smaller that represent all or most of the glaciated Northeast. For example, Kontis, Randall, and Mazzaferro (in press) delineated areas where stratified drift is sufficiently coarse and saturated to constitute aquifers and identified many aquifers capable of sustained withdrawals exceeding 5 million gallons per day or capable of large seasonal withdrawals from storage with minimal concurrent depletion of streamflow. Soller (1993) depicted generalized drift thickness and distinguished surficial till from coarse-grained stratified drift and fine-grained stratified drift. Olcott (1995) also distinguished coarse stratified drift from other surficial materials. All these small-scale regional maps illustrate the discontinuous dendritic distribution that characterizes coarse stratified drift in the glaciated Northeast but are unsuitable for aquifer definition in individual localities because they omit many small deposits, generalize much local variability, and can be deceptive in that their interpretations are founded on abundant data in some intensively studied places but on virtually no data in other places. Surficial geology of many quadrangles, however, has been mapped at scales of 1:62,500 or larger. County soils maps at similar scales are readily translated to surficial geology and are available for much of the glaciated Northeast, and detailed hydrogeologic maps are available for some localities. On the average, stratified drift underlies only about 15 percent of the glaciated Northeast; thus, its delineation greatly narrows the area that merits closer examination to locate productive aquifers capable of yielding large water supplies.

AQUIFER GEOMETRY WITHIN THE STRATIFIED DRIFT

Almost all aquifer appraisals must go beyond merely delineating the areal extent of stratified drift. In some places, stratified drift is too thin, too thinly saturated, or too fine grained to yield much water. In other places, productive aquifers within the stratified drift are interrupted or confined by till, fine-grained stratified drift, or shallow bedrock. Knowledge of aquifer geometry—that is, the three-dimensional distribution and extent of saturated, water-yielding coarse sand and gravel relative to non-water-yielding fine-grained drift, to bedrock, and to streams that are potential recharge sources—is essential to understanding the ground-water flow system, siting large-capacity wells, evaluation of aquifer yield, and prevention or remediation of ground-water contamination. Delineation of aquifer geometry requires some information from the locality of interest and also a conceptual model(s) to organize that information.

INFORMATION THAT IS USEFUL IN DELINEATING AQUIFERS

The following types of information are commonly used to interpret aquifer geometry to the extent they are available within the locality of interest.

Surficial Geology.—Surficial geology maps and soils maps commonly depict not only the areal extent of stratified drift but also the spatial variation in grain size and (or) depositional environment.

Well Records and Borehole Logs.—The U.S. Geological Survey, State and local government agencies, college professors and students, and consultants have over the years collected and published or archived many records and logs of drillholes, chiefly water wells but also test borings intended for engineering foundation design of highway structures or large buildings. Some States and counties currently require water-well drillers to submit records of wells soon after construction, but for many localities, no agencies have collected anything approaching a complete set of records of all wells. Many geohydrologic studies incorporate a review of published or archived records, collection of additional records, and supplemental test drilling at key locations.

Geophysical Exploration.—Several methods of surface geophysical exploration have proven useful in interpreting the geometry of stratified-drift aquifers.

Electrical resistivity or conductivity can, in general, be empirically correlated with grain size and, thus, with hydraulic conductivity of glacial drift in geographic areas where ground-water quality is uniform. Coarse-grained material is generally less conductive (more resistive) to electric current than fine-grained material, and three methods—direct-current resistivity, inductive terrain conductivity, and very low frequency (VLF) terrain resistivity—can detect conductive material below resistive material (Haeni, 1995). Thus, these methods can ordinarily distinguish areas in which thin surficial aquifers overlie fines from areas where all or most of the unconsolidated deposits are coarse grained, although such interpretations can be complicated by local topographic irregularities that result in large differences from place to place in thickness of highly resistive unsaturated surficial sand and gravel. Inductive terrain conductivity can detect lateral changes in thickness and grain size of individual conductive layers in conductive terranes, and VLF can detect lateral changes in grain size in resistive terranes.

Seismic refraction can reliably determine depth to bedrock and commonly also depth to the water table. The combined use of seismic, resistivity, and electromagnetic methods in conjunction with borehole logs and measurements of specific conductivity of ground water often can provide a reasonably comprehensive interpretation of aquifer geometry (Haeni, 1995).

Continuous seismic-reflection surveys taken from boats can distinguish coarse-grained sediments from fine-grained sediments along and beneath rivers and lakes, which is where such information is most needed because thick coarse-grained surficial aquifers that border and underlie rivers and lakes are especially promising as aquifers from which large ground-water withdrawals can be sustained by induced infiltration. Many valleys in the glaciated Northeast are characterized by reaches of predominantly coarse sediment alternating with reaches in which a thin surficial aquifer overlies thick lacustrine fines. Continuous seismic reflection also can detect significant thicknesses of coarser drift beneath the fines, although borehole data are usually required to determine whether the basal deposits are ice-contact sand and gravel deposits or morainal drift of low hydraulic conductivity (Haeni, 1988; Reynolds, 1988; Mullins and Hinchey, 1989).

Water-Level Measurements and Water-Temperature Profiles.—Measurements in wells of water level and change in temperature with depth have been used to establish continuity of aquifers, define preferential flow paths, and identify barriers to flow. For example, Ferris and others (1962) and Walton (1962) explained how aquifer boundaries can be located by applying image-well theory to the analysis of drawdown or recovery of water levels in wells. Crain (1966) cited seasonal changes in water level in a few wells (fig. 30) as evidence that valley-side alluvial fans are the principal source of recharge to a midvalley buried gravel aquifer. Randall (1977) used temperature profiles, water levels, and lithology of drill cuttings to demonstrate that a surficial aquifer had been downwarped, as part of an ice-block depression that is no longer recognizable topographically, and constituted the principal flow path whereby induced infiltration reached a well field (fig. 35). Randall (1986a, p. 14–15) used temperature profiles and water levels to demonstrate continuity of a gravel aquifer between two municipal wells despite differing stratigraphy, and to demonstrate discontinuity of a widespread silty clay confining layer beneath a river reach where induced recharge took place readily. Winslow (1962) and Winslow and others (1965) used temperature observations in many wells to determine the location and depth of the principal flow path of induced infiltration to a well field.

CONCEPTUAL MODELS THAT ARE USEFUL IN ORGANIZING INFORMATION

Seldom is information so abundant and detailed that the hydrogeologic framework of stratified-drift aquifers is unequivocally obvious. Therefore, those who organize and interpret that information commonly rely on some idealized mental images of what to expect. The following generalizations, or "conceptual models," are explained in this paper and are widely applicable in identifying productive aquifers within the stratified drift.

Expect Three Depositional Facies.—Throughout the glaciated Northeast, most tracts of stratified drift larger than 1 square mile contain some combination of three successive depositional facies:

Facies 1. Proximal or ice-contact facies, predominantly coarse grained, heterogeneous and widely variable in sorting, deposited close to active ice and (or) amid abundant stagnant ice.

Facies 2. Distal facies, predominantly silt, clay, and very fine sand, deposited in somewhat larger bodies of water than facies 1.

Facies 3. Surficial facies, coarse grained, commonly well sorted, deposited in shallow water; commonly overlies facies 2; includes prograding deltas, fluvial outwash, and postglacial alluvium, terraces, and fans.

At any given moment during deglaciation, facies 2 and 3 were being deposited farther from the ice margin than facies 1; that is, farther downvalley and in valleys that no longer received meltwater. At any given location, one or two facies may be absent, but where all three are present, facies 1 was deposited first, followed by facies 2, then facies 3. Therefore, in most valleys, the entire stratigraphic section at some locations is coarse grained (facies 1, or 3/1), whereas at other locations nearby it is coarse over fine (3/2), coarse over fine over coarse (3/2/1), or, in broad lowlands, fine over coarse (2/1).

Expect Stratigraphy from a Single Deglaciation.—The Northeastern United States was covered at least twice by continental ice sheets, and the retreat of the last ice sheet was occasionally interrupted by brief readvances. In a few regions, notably the northern rim of the Appalachian Plateau, valley fills include multiple layers of till and stratified drift (facies 1 or 2) that record this history of multiple glaciation. In most regions, however, the last ice sheet eroded to bedrock, and any readvances caused only minor perturbations in the depositional environments wherein facies 1, 2, and 3 were being deposited during deglaciation.

Identify Morphosequences.—During deglaciation, stratified drift often filled all available ponded depressions in a reach of valley, and meltwater streams established a smooth fluvial profile graded to some lake or bedrock saddle. All deposits associated with such a profile constitute a morphosequence. Thin surficial sand or pebble gravel commonly overlie fine-grained sediments near the distal end of a morphosequence, whereas coarse sand or pebble gravel predominate near the proximal end or head. Some maps of surficial geology divide stratified drift into morphosequences. Where no such maps are available, the location of coarse-grained morphosequence heads can often be inferred from altitudes of terrace remnants shown on topographic maps and from grain size in excavations or drillhole logs.

Locate Ice Margins.—Coarse-grained stratified drift (facies 1) is likely to be concentrated wherever the ice margin paused during retreat and to grade distally to fine-grained sediments. Many geologic studies have been directed to delineating the margin of the former ice sheet at one or more times during deglaciation. These inferred ice margins commonly link several morphosequence heads, which are interpreted as marking pauses during retreat. Projections of former ice margins across uplands to connect morphosequence heads in separate valleys have little hydrologic significance; but where ice margins crossed broad lowlands, the coarse stratified drift may form a semicontinuous swath, in part concealed below younger, fine-grained sediments.

Follow Subglacial Channels.—Much of the stratified drift that was deposited at the margin of ice tongues in valleys had been entrained by meltwater at the base of the ice sheet and transported along subglacial tunnels to the ice margin; some sediment entrained by meltwater in adjacent uplands also found its way to these channels. Coarse, permeable ice-contact deposits termed eskers were commonly formed in these straight to sinuous conduits and remain today as exposed or buried ridges of sand and gravel extending proximally from morphosequence heads. In localities where ice retreated rapidly by calving, subaquatic fans formed at tunnel mouths. Subglacial channels were persistent features within the retreating ice and have resulted in continuous, narrow basal aquifers as much as a few miles long in some places.

Test Near Tributaries.—Tributary streams, both large and small, were sources of sediment to major valleys during deglaciation, as they are today. The coarse fraction was deposited chiefly near the valley margin, as ice-contact deposits amid stagnant ice and later as subaquatic fans and deltas into persistent large lakes. Not only is stratified drift likely to be relatively coarse near tributaries, but the modern tributaries are potential sources of recharge that could sustain the yields of wells completed in the stratified drift. In many valleys, the coarse sediments near tributaries constitute avenues by which recharge can reach deep ice-contact deposits that are elsewhere buried beneath fines.

Look for Collapsed Margins.—A substantial fraction of the total volume of coarse-grained stratified drift in the glaciated Northeast lies above modern stream grade and, thus, is thinly saturated. Many facies-3 deltaic and outwash deposits were graded to lake levels well above the modern valley floors, and subsequent stream incision or abandonment of temporary spillways has left these deposits high and dry—except where they were built over buried ice blocks whose eventual melting allowed surficial coarse layers to collapse below the modern water table. Many facies-1 ice-contact deposits rest upon shallow bedrock along the valley sides or around bedrock knolls in midvalley and also are thinly saturated—except along their headward or valleyward margins where bedrock is deeper and some coarse stratified drift has collapsed to lower elevations.

ANTICIPATING AQUIFER DISTRIBUTION TYPICAL OF PARTICULAR HYDROPHYSIOGRAPHIC REGIONS

The abundance and distribution of stratified-drift aquifers were affected by the changing dynamics of the continental ice sheet during deglaciation and by many aspects of the subglacial landscape, such as relief, drainage density, and drainage orientation toward or away from the ice sheet. Different regions were affected differently; therefore, generalizations as to aquifer geometry in a particular region can be more specific than generalizations intended to apply to the entire glaciated Northeast. Several regions are delineated and described in this paper with the expectation that a review of aquifer geometry in each region would prove useful in guiding appraisal of individual localities. These hydrophysiographic regions have been grouped into four broad categories, as follows.

Regions that Sloped Generally Away from the Ice.—This category includes some areas that sloped toward the ice but had such low relief that only small lakes were ponded behind the low saddles on the watershed perimeter. In regions of moderately low relief in southern New England, New Jersey, and southern and northwestern New York, the principal valleys contain abundant, nearly continuous stratified drift in which coarse-grained sediment is a major component. By contrast, regions of moderately low relief in northern and central Maine have minimal amounts of stratified drift, which occurs partly as valley fills and partly as eskers that follow but do not fill valleys. Mountain regions in northern New England and northern New York are characterized by narrow valley floors that are largely capped by coarse alluvium, commonly interrupted by bedrock outcrops, and in part underlain by stratified-drift aquifers; stratified drift also is found on the valley sides, perched above stream grade. Relief is equally high in the Appalachian Plateau, but no stratified drift is perched on the valley sides, no outcrops occur on the valley floors, valley fills generally exceed 2,000 feet in width and 150 feet in thickness, and facies-2 fines predominate wherever depth to bedrock is relatively great. In the adjacent Catskill Mountains, typical valley fills contain few fines and are less than 2,000 feet wide, less than 150 feet thick, and are locally narrowed by thick till on the lee (down-ice) side of hills. The Tug Hill Plateau of New York and Pocono Plateau of Pennsylvania have low internal relief except along the plateau margins; valleys are small and generally lack significant stratified-drift aquifers.

Broad Lowlands Inundated by Large Water Bodies.—Extensive deposits of silt, clay, and fine sand in these lowlands are interrupted by many till hills, by a few ice-marginal deltas, and (at intervals of several miles) by linear sand and gravel deposits, the products of subglacial meltwater channels. Some lowland areas are capped by extensive surficial sand-plain aquifers deposited by upland runoff from adjacent regions, but generally, facies-2 fines are the uppermost sediments. Induced infiltration is seldom possible in these lowlands. The margins of adjacent uplands are commonly characterized by high-level deltas now deeply incised and perched above stream grade.

Regions of High Relief that Sloped Toward the Ice.—Deep valleys are filled with predominantly fine-grained sediments seldom capped by a surficial aquifer. Along the north rim of the Appalachian Plateau, valleys contain diamicton layers and scattered lenses of sand and gravel interbedded with fines. In northwestern Vermont and northeastern New York, most valleys contain a single extensive layer of fines, locally underlain by facies-1 sand and gravel, with bedrock outcrops common on valley floors.

Regions At or South of the Wisconsinan Drift Border.—The Wisconsinan drift border in New Jersey approximately coincides with a terminal moraine of chiefly till that overlies a network of bedrock valleys containing gravel aquifers of Illinoian and Wisconsinan ages. From New Jersey to Ohio, valleys draining south from the drift border were aggraded with sandy valley trains many miles long, now deeply incised but still significant aquifers that extend 40 to 100 feet below stream grade in many valleys.

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GLOSSARY

Several terms related to glacial geology and used in this paper are defined below.

Ablation till: loosely consolidated, generally sandy and stony debris that had been contained within stagnant ice and remained in place when the ice was removed by melting or sublimation.

Alluvial fan: a gently sloping mass of predominantly gravelly sediment, shaped like an open fan or a segment of a cone, deposited by a stream where it issues from its own upland valley and enters a larger valley.

Alluvium: unconsolidated sediment deposited by a postglacial stream along its flood plain or in an alluvial fan; generally consists of gravel or gravelly sand, overlain by organic-rich fine sand to silt along the flood plains of larger streams.

BP: before the present; refers to age in years determined by measuring decay of carbon–14 or other radioactive isotopes.

Constructional topography: any array of mounds, ridges, hummocks, hills with rounded or level tops, or similar topographic forms whose height results from accumulation of unconsolidated sediment rather than from the form of the surface on which the sediment accumulated.

Dead-ice sink: an unusually large ice-block depression, within which progressive melting of stagnant ice provided a progressively expanding space or sink for sediment deposition.

Diamicton: any nonsorted unconsolidated sediment containing a wide range of grain sizes, regardless of genesis; includes till and various products of mudflow, debris flow, or slumping.

Distal: remote from the source of sediment; stratified drift deposited far from the active ice margin is termed distal.

Drift: all rock fragments or particles transported by glacier ice and deposited by the ice or by meltwater; includes till, stratified drift, and deposits resedimented by mass movement soon after deposition by ice or meltwater.

Drop pellets: nonlithified clasts (of diamicton, clay, or silt) dropped from floating ice into sediments accumulating on the bottom of a proglacial lake.

Esker: a narrow, steep-sided, commonly sinuous and sharp-crested ridge composed of stratified sand and gravel deposited by a meltwater stream in a tunnel at the base of an ice sheet.

Foreset: one of the inclined layers of sediment (generally sand) deposited along the advancing, relatively steep frontal slope of a delta; foresets individually grade into, and collectively prograde across, flat-lying bottomsets.

Fines: fine-grained unconsolidated sediment, including very fine sand, silt, and clay.

Glaciofluvial: pertaining to or deposited by meltwater streams flowing beside or away from a glacier.

Glaciolacustrine: pertaining to or deposited in proglacial lakes.

Grounding line: the locus of points beyond which an active ice sheet that ends in ponded water loses contact with the ground and becomes a floating ice shelf.

Gilbert-type delta: a delta that exhibits the components originally described by G.K. Gilbert, namely, a nearly level surface underlain by topset, foreset, and bottomset beds.

Ice-channel filling: a narrow, steep-sided, linear or sinuous ridge composed of stratified sand and gravel, deposited by a meltwater stream in a tunnel at the base of the ice sheet or in an ice-walled channel open to the sky; includes eskers and crevasse fillings.

Ice-contact: refers to deposits or margins of deposits that were formed against, among, or atop masses of ice, ordinarily stagnant ice, and that exhibit deformation resulting from loss of support when the ice melted.

Ice stream: a zone within an ice sheet in which the ice flows more rapidly than bordering ice; usually such rapid flow follows a broad bedrock valley directly into a large lake or ocean.

Inwash: unconsolidated sediment delivered to a proglacial lake or to the stagnant marginal zone of the ice sheet by tributary streams that are not fed by meltwater.

Jokulhlaup: a catastrophic flood resulting from the sudden draining of a proglacial lake through an outlet formerly blocked by ice.

Meltout till: till deposited by slow release of sediment from melting ice that is motionless (stagnant).

Moraine: a distinct accumulation of drift, composed mostly of till (except in valleys, where stratified drift may predominate), whose topographic form is independent of the bedrock on which it was deposited and is commonly, but not necessarily, ridgelike; moraines referred to in this paper are, strictly speaking, end moraines, deposited along the margin of active ice.

Morphosequence: a body of stratified drift that was laid down by meltwater when deposition was controlled by a specific base level such as a proglacial lake or spillway; the deposits become generally finer distally and their upper surface (where not collapsed) slopes smoothly in the same distal direction.

Moulin: a roughly cylindrical, nearly vertical, well-like hole in an ice sheet, carved by meltwater as it descends from the surface to the base of the ice.

Mudboil: a low, cone-shaped mound of sand with a central crater from which erupts muddy water that has flowed upward along a fracture, driven by artesian pressure.

Openwork: describes gravel composed entirely of well-sorted granules, pebbles, or cobbles with little or no interstitial sand, such that pore spaces between the stones are readily visible and hydraulic conductivity is extraordinarily high.

Outwash: stratified sand and gravel transported and deposited by meltwater beyond the region in which active or stagnant ice predominantly fill the valleys.

Outwash plain: a broad, outspread, gently sloping sheet of sand and gravel deposited by meltwater streams in front of or beyond an ice margin; comparable to valley train but wider and less constrained by valley walls.

Proximal: close to the source of sediment; stratified drift deposited in open water or amid stagnant ice immediately beyond the margin of active ice is termed proximal.

Radiocarbon years: the age of sediment containing organic matter, as calculated from measurements of the amount of carbon-14 relative to other carbon isotopes; ages greater than 10,000 radiocarbon years are about 87 percent of ages calculated by more precise but less widely applicable methods (Bard and others, 1990).

Resedimented: applied to sediments initially deposited by one depositional mechanism but later transported and deposited by another mechanism, generally retaining some but not all physical characteristics associated with the initial deposition; most commonly applied to glacial drift subsequently transported by solifluction or debris-flow processes.

Rhythmites: a succession of sediment layers characterized by alternation among two or more lithologies. In proglacial lakes, alternate layers of clay and silt (grading to fine sand and silt near sediment sources) are the most common type of rhythmites.

Sand-plain aquifer: an extensive, nearly level surficial sand deposit that overlies fines and is drained by streams that are commonly incised into the fines, such that induced infiltration is impossible and discharge occurs by riparian springs or seeps.

Sandur: outwash plain.

Stack units: map units that depict not only the uppermost earth material but also one or more underlying materials; for example, g/s (gravel overlying sand) or 20g/30fs (20 feet of gravel overlying 30 feet of fine sand).

Stade: a substage (of a glacial stage) that is marked by a glacial advance.

Stratified drift: sorted and layered sediment deposited by or in meltwater; includes gravel, sand, silt, and clay.

Thalweg: the line of continuous maximum descent or greatest slope that connects the lowest points along a valley (or, if so specified, along the bedrock surface beneath unconsolidated deposits in the valley).

Through valley: a valley that was cut through a bedrock ridge and floored with stratified drift deposited by through-flowing meltwater during deglaciation but that now is crossed by a watershed divide and drained only by tiny streams on each side of the divide.

Till: a nonsorted, generally nonstratified, poorly permeable unconsolidated sediment, deposited from and underneath a glacier without reworking by meltwater and consisting generally of particles ranging from clay to boulder size, some of which have been transported considerable distances.

Topset: one of the nearly horizontal layers of sediment (generally gravel) deposited slightly above lake level by streams as they cross the top of a delta and transport sediment toward the frontal delta slope; individual topsets locally are continuous with individual foresets, but collectively topsets cap and, in places, truncate foresets.

Tunnel valley: a steep-walled trench cut in drift or in bedrock by a subglacial stream.

Valley fill: all unconsolidated sediment that overlies bedrock within a valley, partly or (in rare instances) totally filling the valley.

Valley train: a long, narrow body of outwash deposited by meltwater beyond the margin of the ice sheet but confined within the walls of a valley.

Underfit stream: a stream that seems too small to have eroded the valley in which it flows; a fairly common result of drainage changes caused by glacial erosion or deposition.

Unconformity: a boundary between layers of sediment that represents a substantial period of nondeposition and may reflect loss of some sediment by erosion before renewed deposition took place.

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