Geomorphology of New England

By CHARLES S. DENNY

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Topography of crystalline rocks,
lithology of Coastal Plain sediments,
and comparisons with adjacent areas
suggest late Cenozoic uplift

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By Charles S. Denny

ABSTRACT

Widely scattered terrestrial deposits of Cretaceous or Tertiary age and extensive nearshore and fluvial Coastal Plain deposits now largely beneath the sea indicate that the New England region has been above sea level during and since the Late Cretaceous. Estimates of rates of erosion based on sediment load in rivers and on volume of sediments in the Coastal Plain suggest that if the New England highlands had not been uplifted in the Miocene, the area would now be largely a lowland. If the estimated rates of erosion and uplift are of the right order of magnitude, then it is extremely unlikely that any part of the present landscape dates back before Miocene time. The only exception would be lowlands eroded in the early Mesozoic, later buried beneath Mesozoic and Cenozoic deposits, and exhumed by stream and glacial erosion during the later Cenozoic.

Many of the rocks in the New England highlands are similar to those that underlie the Piedmont province in the central and southern Appalachians, where the relief over large areas is much less than in the highlands of New England. These comparisons suggest that the New England highlands have been upwarped in late Cenozoic time. The uplift took place in the Miocene and may have continued into the Quaternary.

The New England landscape is primarily controlled by the underlying bedrock. Erosion and deposition during the Quaternary, related in large part to glaciation, have produced only minor changes in drainage and in topography. Shale and graywacke of Ordovician, Cambrian, and Proterozoic age forming the Taconic highlands, and akalic plutonic rocks of Mesozoic age are all highland makers. Sandstone and shale of Jurassic and Triassic age, similar rocks of Carboniferous age, and dolomite, limestone, and shale of Ordovician and Cambrian age commonly underlie lowlands. High-grade metapelites are more resistant than similar schists of low metamorphic grade and form the highest mountains in New England. Feldspathic rocks tend to form lowlands. Alkalic plutonic rocks of Mesozoic age underlie a large area in the White Mountains of New Hampshire and doubtless are a factor in their location and relief.

Where the major streams flow across the regional structure of the bedrock, the location of the crossings probably is related to some other characteristic of the bedrock, such as joints or cross faults. The course of the Connecticut River is the result of the adjustment of the drainage to the bedrock geology during a long period of time. There is no ready explanation why many of the large rivers do not cross areas of calc-alkalic plutonic rock, but rather take a longer course around such areas, which tend to include segments of the divide between the streams.

The presence of coarse clastic materials in Miocene rocks of the emerged Coastal Plain of the Middle Atlantic States suggests uplift of the adjacent Piedmont and of the Adirondack Mountains at that time. The Miocene rocks of the submerged Coastal Plain in the Gulf of Maine and south of New England are fine grained and contain only small amounts of fluvial gravel. Perhaps the coarse clastic materials shed by the New England highlands in late Cenozoic time are buried by or incorporated in the Pleistocene glacial deposits.

INTRODUCTION

The landscape of New England and the surface of the adjacent Continental Shelf in large part reflect the rocks that underlie them. The character and distribution of the bedrock control the form and location of the highlands and lowlands. The adjacent Continental Shelf is underlain in part by Cretaceous and younger rocks that contain a record of the erosional history of the adjacent landmass.

The boundary of the New England States does not coincide with that of any physiographic subdivision. The study area, the New England region, includes the New England States and adjacent parts of New York and the Canadian Provinces of Quebec and New Brunswick. The Hudson-Champlain-St. Lawrence lowlands constitute the west and northwest border of the region (fig. 1). Northeast of Maine, no natural physiographic boundary exists, and the limit of the study area may be conveniently placed along the east border of the St. John River basin in New Brunswick. The New England region excluding the adjacent Continental Shelf comprises nearly 260,000 km^2; of this, the New England States make up about 173,500 km^2, Maine accounting for about half this area.

The study includes an analysis of the topography of the New England region in relation to rock types and structural fabrics. The relationships documented here reflect a long history of erosion and adjustment of surface form to bedrock character. The Coastal Plain sediments that have been derived from the New England region record times and places of intense erosional activity and suggest tectonic events not otherwise determinable.

Erosion and deposition during the Quaternary, related in large part to glaciation, have produced many changes in drainage and in topography. Hills have been formed or reshaped, valleys have been deepened, lowlands have been buried, and streams have been diverted. Examples of such changes are legion. Nevertheless, the major elements of the topography and drainage of the New England region can be explained as a response to the character and distribution of the bedrock, even where the bedrock is concealed beneath Quaternary deposits.

METHOD OF STUDY

The topographic data used in this study were derived from maps at a scale of 1:250,000, contour interval 100 ft (30.48 m), published by the U.S. Geological Survey and the Canada Department of Mines and Technical Surveys. The geographic names follow the usage of the 1:250,000-scale maps. The submarine topography of the Continental Shelf is from the map by Uchupi (1965). The topographic data for land areas were analyzed using three derivative maps at scale 1:1,000,000. The maps portray relief, summit altitudes, and stream gradients. The map of summit altitudes shows what Stearns (1967, p. 1117; Hack, 1973) has called the **envelope**; the map of stream gradients depicts the general altitude of the drainage network, the **subenvelope** of Stearns. Longitudinal profiles were drawn for many of the larger streams at scale 1:1,000,000. Only the derivative maps showing relief and stream gradients are reproduced here.

The three basic derivative maps were constructed as follows. A 1:250,000-scale sheet was overlain by a rectangular grid 6.4 km on a side. In each square was recorded the altitude of the lowest contour line minus one and of the highest contour line. For example, if in a 6.4-km square the lowest contour line is 200 ft (60.96 m) and the highest contour line is 600 ft (182.88 m), the two altitudes recorded in the square are 100 ft (30.48 m) and 600 ft (182.88 m). The data derived for the square are as follows: summit altitude, the larger number; altitude of the drainage network, the smaller number; and relief, the difference between the two numbers.

The derivative maps were constructed in the following manner: The values obtained from the topographic maps were recorded in the center of each 6.4-km square and contoured using a 200-ft (60.96-m) interval. The maps were then reduced to scale 1:1,000,000. For the maps reproduced in this paper (figs. 9, 10, 11), the contours have been redrawn using a metric interval.

The map showing bedrock lithology of land areas (fig. 4) was compiled from geologic maps of the States of New York, Vermont, New Hampshire, Maine, and Rhode Island (New York State Museum and Science Service, Geological Survey, 1962; Doll and others, 1961; Billings, 1955; Hussey and others, 1967; and Quinn, 1971), of the Provinces of Quebec and New Brunswick (Quebec Department of Natural Resources, 1969; Potter and others, 1968), and regional compilations by Dixon and Lundgren (1968), Hatch and Stanley (1973), Goldsmith (1964), Morgan (1972), White (1968), and Zen (1967, 1972). The pre-Pleistocene geology of the Continental Shelf is from maps compiled by Weed and others (1974), King and MacLean (1974), and Ballard and Uchupi (1975). The approximate offshore limit of glaciation is from Pratt and Schlee (1969).

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TOPOGRAPHIC AND GEOLOGIC FRAMEWORK

New England is largely an area of hills and mountains. The term commonly used in the past, "The New England Upland," was based in part on the belief that the area was an elevated and dissected peneplain surmounted by monadnocks. Much of the region is at altitudes of more than 200 m (fig. 2), and large areas in the central part are at altitudes of more than 500 m. Small areas in the Green Mountains of northern Vermont and in the White Mountains of northern New Hampshire, as well as Mount Katahdin in Maine and a few other isolated peaks, have altitudes of more than 1,000 m. Altitudes on the floors of the major lowlands are generally not more than 200 m, except in parts of the St. John River valley in Maine, where the altitude ranges from about 200–400 m. The valleys between the highlands are commonly narrow, except for the broad lowlands of the Hudson, Champlain, and St. Lawrence Valleys and those bordering the west and northwest shore of the Gulf of Maine.

Depths on the Continental Shelf are not more than 100 m on the largely submerged Coastal Plain south of New England and east of the Massachusetts coast. In the Gulf of Maine, a belt 20–40 km wide, where depths are less than 100 m, follows the coast from Cape Ann northeast to the Bay of Fundy. The rest of the floor of the gulf ranges in depth from 100 to 200 m, except for many large and small basins whose bottom contours are closed at depths of about 200–300 m.

To describe the landscape of the New England region and to attempt to explain its origin, it is convenient to divide both the emerged and submerged parts of the region into several physiographic units (fig. 3). The units are basically topographic and reflect differences in major bedrock units. However, some of the units are of necessity so broad and varied in character that the distinction between one unit and another is in some areas arbitrary.

HUDSON-CHAMPLAIN-ST. LAWRENCE LOWLANDS

The lowlands along the Hudson and St. Lawrence Rivers and surrounding Lake Champlain range from about 15 km to nearly 130 km in width and extend for about 700 km from Newburgh, N.Y., to Quebec City (fig. 3). Tidewater extends about as far north as Albany on the Hudson River and as far southwest as Trois Rivières on the St. Lawrence River. The lowlands are underlain largely by shale, dolomite, and limestone (fig. 4) of Ordovician and Cambrian age. The altitudes of the valley floors range from sea level to about 100 m above sea level. In the St. Lawrence Valley between Montreal and Quebec, more than three-quarters of the area has a relief of less than 60 m, in part the result of deposition on the floor of the Champlain Sea that occupied the valley at the end of the Wisconsinan glacial stade (Gadd, 1971; Elson, 1969).

Downstream from Quebec City, the St. Lawrence River estuary follows a major thrust fault, Logan's Line, between the Proterozoic rocks of the Laurentian Highlands to the northwest and the Paleozoic rocks of the Notre Dame Mountains to the southeast. In the Champlain and the Hudson Valleys, the bedrock includes sandstone that forms low hills, and the relief ranges from about 60 m to as much as 300 m. The boundary of the lowlands is in most areas easily defined on the basis of both bedrock character and topographic form (fig. 5). The Hudson-Champlain-St. Lawrence lowlands are part of a great valley system that extends the length of the Appalachian Highlands and is underlain largely by carbonate rocks of early Paleozoic age (Hack, 1980; Fenneman, 1938, p. 200).

TACONIC HIGHLANDS

The Taconic highlands, an area of low mountains and hills east of the Hudson River valley, includes the Taconic Mountains in New York (New York State Museum and Science Service, Geological Survey, 1962, text-fig. 19) and adjacent peaks in Massachusetts and southwestern Vermont (fig. 3). On the west side, the highlands gradually descend to the floor of the Hudson Valley; in many places, no sharp boundary exists between highland and lowland. To the east, however, the highlands are bordered by steep-sided narrow valleys (the Vermont Valley). In Massachusetts and New York, several peaks are more than 760 m in altitude. The highlands extend north into the southwest corner of Vermont, where several peaks rise more than 1,070 m above sea level. The highlands coincide roughly with the Taconic allochthon (Zen, 1967), which has a maximum length of about 210 km and ranges in width from about 25 to 40 km. The rocks of the allochthon are largely shale and graywacke and commonly overlie carbonate rocks of early Paleozoic age. The weathering and removal of the carbonate rocks beneath the overlying sediments of the allochthon has produced the steep-sided mountains and narrow valleys in this part of New England. For example, Dorset Mountain, south of Rutland, Vt., rises steeply almost 915 m above the adjacent flat-floored valley of Otter Creek.

VERMONT VALLEY

The lowlands east of the Taconic highlands are a long chain of narrow valleys that extend from north of Rutland south for about 200 km. Near Rutland, the lowlands are called the Vermont Valley (Stewart and MacClintock, 1969, fig. 3); the name is here used for the entire chain that includes Otter Creek, the upper reaches of Batten Kill and the Hoosic River, largely in Vermont, and the broad valley of the Housatonic River in Massachusetts (fig. 3). The valleys range in width from about 1.5 to 4.5 km, too narrow to show on the relief map (fig. 10). Altitudes on the valley floors commonly range from about 60 to 180 m. The lowlands are in a belt of rocks, largely dolomite and limestone of Ordovician and Cambrian age, that are separated by the Taconic allochthon from similar rocks in the Hudson-Champlain lowlands to the west. The lowlands are bordered by steep-sided mountains that rise 600-900 m above the valley floors. The highlands to the east include large areas of massive rocks of Proterozoic age.

HUDSON-GREEN-NOTRE DAME HIGHLANDS

The Hudson-Green-Notre Dame highlands extend across the New England region from the southwest corner to the northern border, a distance of more than 850 km (fig. 3). The west border of the highlands is in many places a sharp break between highland and lowland, whereas the east border is in many places not a conspicious topographic or lithologic break. The highlands include the Hudson Highlands in New York, the Berkshire Hills in Massachusetts, the Green Mountains in Vermont, and the Sutton and Notre Dame Mountains in Quebec. The Notre Dame Mountains continue to the northeast outside the study area to the Gaspé Peninsula. The highlands range in altitude from about 400 to 1,200 m. The highest peaks are in northern Vermont, where Mount Mansfield northwest of Montpelier has an altitude of about 1,340 m.

Rocks of Proterozoic age underlie the Hudson Highlands and much of the higher parts of the Berkshire Hills and the Green Mountains in southern Vermont. Metasedimentary and metavolcanic rocks of Paleozoic age underlie most of the rest of the Hudson-Green-Notre Dame highlands. Discontinuous belts of calc-silicate rocks are found in Massachusetts and in Vermont near the Connecticut Valley. There are only a few small areas of calc-alkalic intrusive rocks. In Connecticut, Massachusetts, and southern Vermont, the metasedimentary rocks of the highlands are largely amphibolite facies (fig. 6). In the western half of the highlands in central and northern Vermont, the metasedimentary rocks are largely greenschist facies, except in the higher parts of the Green Mountains.

The Hudson Highlands are a belt of Proterozoic rocks, which is about 20 km wide where crossed by the Hudson River estuary and which extends northeast from New York State into western Connecticut. A hornblende granite and granite gneiss form much of the highlands west of the river, including Storm King Mountain (altitude about 425 m), and are absent on the east side, except near the entrance to the gorge south of Newburgh, N.Y. (fig. 7). The dominant rock type east of the river is biotite granite gneiss (Hall and others, 1975). The rest of the highlands near the river are underlain largely by gneiss and amphibolite. Some quartz plagioclase gneiss forms ridges, which are slightly lower than adjacent ridges composed of granitic gneiss. A possible connection between the bedrock of the highlands and the course of the Hudson River will be discussed later.

The Berkshire Hills in Massachusetts and northern Connecticut are composed of Proterozoic rocks (Berkshire Massif, White, 1968) and are about 90 km long and 5–10 km wide. The Green Mountains in southern Vermont (south of lat 44° N.) are also Proterozoic rocks (Green Mountain Massif, White, 1968) and are about 150 km long and 10–20 km wide. Altitudes in the Berkshire Hills range from about 400 to 800 m, those in the Green Mountains of southern Vermont, from about 600 to 1,100 m. The Proterozoic rocks include large bodies of massive gneiss and subordinate quartzite.

In northern Vermont, the higher peaks of the Green Mountains are largely schistose metasedimentary rocks of amphibolite facies (fig. 6); the most extensive rock types are quartz-mica-schist containing abundant quartz segregations, quartzite, and amphibolite (Underhill and Hazens Notch Formations, Doll and others, 1961). The east limit of the highlands is drawn near Montpelier and separates the Green Mountains from the lower mountains and hills of northeastern Vermont (Vermont Piedmont, Stewart and MacClintock, 1969, fig. 3). Most of the calc-alkalic plutonic rocks that form low mountains in northeastern Vermont are included in the Central highlands (fig. 5).

The Sutton Mountains just north of the international boundary reach altitudes of about 750–950 m and are underlain in large part by graywacke, slate, quartzite, and volcanic rocks of Ordovician and Cambrian age (Poole and others, 1970). From the Vermont-Quebec border to the Notre Dame Mountains southeast of Quebec City, the northwest border of the highlands is a prominent topographic break. On the other hand, southeast of the highlands are hills, low mountains, and some broad valleys. The southeast border of the Hudson-Green-Notre Dame highlands is drawn along the headwater tributaries of the St. François roughly from Newport, Vt., to Thetford Mines, Quebec (fig. 5). The Notre Dame Mountains south of Rivière-du-Loup range in altitude from about 600 to 700 m and are composed largely of quartzite, sandstone, graywacke, slate, and volcanic rocks (Poole and others, 1970; Hussey and others, 1967).

CONNECTICUT VALLEY

The Connecticut Valley in Connecticut and Massachusetts ranges from about 8 to 32 km in width. The altitude of the valley floor ranges from sea level to about 100 m, except for the long narrow ridges of volcanic rock largely on the west side of the valley (fig. 8), which rise to altitudes of about 180–300 m. The relief is less than 60 m on the valley floor between Hartford, Conn., and Springfield, Mass. The bedrock is largely sandstone, shale, conglomerate, and volcanic rocks of Jurassic and Triassic age. In Connecticut and Massachusetts, the borders of the Connecticut Valley are easily defined on the basis of bedrock lithology and topographic form.

In New Hampshire and Vermont, the valley is narrow and bordered by hills that rise gradually east and west to the White and Green Mountains. As far north as the mouth of the Passumpsic River, the valley runs more or less parallel to the trend of the bedrock units, largely metasedimentary and metavolcanic rocks of Paleozoic age. The central segment of the valley, roughly from Claremont to Littleton, N.H. (fig. 9), parallels the Ammonoosuc thrust. The rocks east and west of the river differ, but the Connecticut Valley as a physiographic unit is defined solely on the basis of topography. North of the mouth of the Passumpsic River, the Connecticut Valley is slightly wider than it is to the south. Areas underlain by calc-alkalic plutonic rocks lie both east and west of the axis of the valley.

CENTRAL HIGHLANDS

The central highlands extend from eastern Connecticut to the northern limit of the study area, a distance of about 750 km (fig. 3.). The geology and topography of this unit are varied; the area includes the highest peaks in New England, as well as low mountains, hills, and some broad valleys (fig. 5). From an altitude of about 200 m at their southern end, the highlands gradually rise north to the White Mountains, where many peaks reach altitudes of more than 1,200 m; Mount Washington has an altitude of 1,886 m. The White Mountains and the adjacent peaks to the northeast, as far as Mount Katahdin, are the highest segment of the central highlands. North and northwest of the White Mountains, altitude and relief decrease; the area is one of low mountains and hills generally not more than 500 m in altitude, except near the international boundary. Several broad valleys have floors as low as 200 m, and the boundary between the central highlands and the Hudson-Green-Notre Dame highlands is arbitrary. In New Brunswick, the central highlands are the Chaleur Uplands of Bostock (1970). Along the eastern border of the central highlands, the topographic break between the highlands and the coastal lowlands is sharp, except north of Concord, N.H., where several isolated peaks (Belknap Mountains, Ossipee Mountains) rise steeply above the adjacent lowlands. Except near its southern end, the border is drawn more or less between the 100- and 200-m contour.

The central highlands are underlain largely by metasedimentary and metavolcanic rocks of Paleozoic age (fig. 4). In the White Mountains and the highlands to the south, the metamorphic rocks are largely amphibolite and granulite facies (fig. 6); they rise above areas underlain in part by calc-alkalic plutonic rocks. The medium- to high-grade metasedimentary rocks underlie Mount Washington and adjacent peaks of the Presidential Range and many of the higher peaks to the south, such as Moosilaukee, Kearsage, and Monadnock. Mount Washington is bordered to the south and southwest by resistant alkalic plutonic rocks (fig. 9).

To the northeast in Maine, the mountains decrease in altitude, and the metamorphic rocks are largely greenschist and subgreenschist facies (fig. 6). The ease of splitting, the fissility, of the low-grade metamorphic rocks makes them more susceptible to weathering and to erosion than are those of higher metamorphic grade.

The next most abundant rock type in the central highlands is calc-alkalic plutonic rocks. Such rocks underlie both mountains and valleys. At Mount Katahdin, for example, calc-alkalic plutonic rocks rise to an altitude of about 1,580 m above sea level. The rocks underlie an area roughly 50 by 30 km. In about half the area, the relief is less than 300 m; in the other half, it is more than 600 m (fig. 5). Metavolcanic rocks form lower mountains just north of Mount Katahdin. Why the plutonic rocks near Mount Katahdin have such a wide range in altitude and in relief is not clear. Perhaps the answer lies in the presence of fine-grained granitic rocks at the summit of Mount Katahdin, rocks that are more resistant to weathering than is the coarse-grained granitic rock at lower altitudes. In northeastern Vermont, on the other hand, plutonic rocks form both broad swampy valleys having a relief of only about 180 m and low mountains having a relief of as much as 670 m.

From Mount Katahdin southwest to the New Hampshire border, oval calc-alkalic intrusive bodies range from about 1 to 8 km in width and from about 15 to 65 km in length. The relief is at least 600 m, except near the State line, where several areas of plutonic rock about 16 km in diameter have a relief of less than 300 m. In the highlands from Mount Katahdin west to the longitude of Rumford, Me., some of the mountains are underlain by metamorphic rocks that form a belt surrounding the plutonic rocks and that appear to be more resistant than those of the central core. In places, a massive hornfels forms steep-sided mountains, such as Big Squaw and Prong Pound Mountains near Greenville, Me., which rise as much as 600 m above Moosehead Lake.

In Connecticut and Massachusetts, the relief on plutonic rocks is commonly less than 300 m. To the north in New Hampshire, some areas of plutonic rock are broad valleys having a relief of about 180 m; elsewhere, commonly in areas of very coarse grained plutonic rocks (Kinsman Quartz Monzonite, Billings, 1955), hills or low mountains have a relief of as much as 540 m.

From the Massachusetts line north to the White Mountains (fig. 9), many of the calc-alkalic intrusive rocks are separated from the Connecticut River by a narrow belt of resistant rocks, the massive Clough Quartzite and the Ammonoosuc Volcanics (Billings, 1955), which form a belt of hills and low mountains generally on the west side of the intrusive rocks.

UPPER ST. JOHN RIVER LOWLANDS

The upper part of the St. John River drainage basin, roughly above Grand Falls, New Brunswick, includes some broad lowlands where the relief is low. The river commonly flows in a narrow inner valley below the general level of adjacent lowlands. The bedrock is largely metasedimentary rocks of low metamorphic grade (greenschist and subgreenschist facies)(fig. 6).

COASTAL LOWLANDS

The coastal lowlands, including parts of the adjacent Continental Shelf, form a broad northeast-trending belt from south of the Rhode Island coast north and northeast to Augusta, Me., a total length of about 200 km and a width of about 60-100 km. Altitudes range from more than 100 m below sea level to about 120 m above sea level. Northeast of Augusta, the lowlands continue up the Penobscot River valley for about 240 km to the vicinity of Houlton, Me.

The bedrock of the coastal lowlands includes a wide variety of lithologic types. In eastern Massachusetts and Rhode Island are calc-alkalic plutonic rocks and volcanic rocks of Proterozoic age and sedimentary rocks largely of Carboniferous age (fig. 4). The Proterozoic rocks have a relief of less than 60 m and are cut by many faults that have a general northeast trend (Cameron and Naylor, 1976). In New Hampshire and in Maine are extensive areas of calc-alkalic plutonic rocks, the largest being northwest of Portland, where the relief ranges from 60 to 180 m. Calc-silicate rocks are found in areas from eastern Connecticut to Bangor, Me. From Portsmouth, N.H., northeastward, the submerged part of the coastal lowlands is underlain largely by rocks of pre-Triassic age (Ballard and Uchupi, 1975, fig. 3).

The coastal lowlands extend from Rhode Island west to the Hudson River (fig. 3); their southern boundary is the submerged northern edge of the Coastal Plain. However, except in the area near New Haven, Conn., which is underlain by rocks of Jurassic and Triassic age, there are no extensive lowlands in the narrow belt from Rhode Island to the Hudson. The altitude of the interfluves rises north from the shore of Long Island Sound at a constant rate, as Flint (1963) has clearly shown (fig. 15). In southern Connecticut, the bedrock includes both stratified and plutonic rocks. East of New Haven, the bedrock units run more or less parallel to the shoreline; west of New Haven they are more or less at an angle to the shoreline.

The coastal lowlands end abruptly against the central highlands (fig. 5). Altitudes rise from about 100 m in the lowlands to more than 300 m near the edge of the highlands. The relief increases from less than 120 m to more than 375 m, and stream gradients increase as one goes from the coastal lowlands into the adjacent highlands. In places, the lowlands-highlands boundary is a gently sloping escarpment, perhaps as much as 100 m high, cut by many valleys running at right angles to it. In Massachusetts, such an east-facing scarp extends from near Worcester (fig. 5) north to the New Hampshire border (west side of the Nashua River valley). In Maine, the coastal lowlands-central highlands boundary follows the belt in which altitude, relief, and stream gradients increase. Near Lake Winnipesaukee, south of the highest part of the central highlands, are several isolated mountains, and no sharp boundary exists between highland and lowland.

NEW BRUNSWICK HIGHLANDS

In Canada (fig. 3), the New Brunswick Highlands of Bostock (1970) are a belt of hills that surround the western part of a broad lowland, the Maritime Plain. The highlands rise to altitudes of more than 200 m and form a U-shaped belt about 60 km wide. The highlands extend into northeastern Maine, where, for the purposes of this paper, the boundary of the New Brunswick Highlands is drawn to include an extensive area of low mountains and broad lowlands southeast of the inland extension of the coastal lowlands (Augusta to Houlton, Me.). Some of the peaks near the shore are separated by fiords or narrow bays. Many low mountains reach altitudes of more than 300 m. The dominant bedrock type in the New Brunswick Highlands is calc-alkalic plutonic rock of Paleozoic age. Some of the peaks are in areas of contact metamorphism adjacent to the plutonic rocks. Some metavolcanic rocks are also present, expecially in areas near the coast. The division includes Mount Desert Island and extends south for about 20 km beneath the Gulf of Maine, where the bedrock is believed to be similar to that in the adjacent land areas (Ballard and Uchupi, 1975, fig. 3).

MARITIME PLAIN

Near Fredericton, New Brunswick, the St. John River crosses the western end of a broad plain of low relief—the Maritime Plain of Bostock (1970)—that slopes east, outside the study area, to the coast. Altitudes near Fredericton range from sea level to about 100 m. The bedrock is largely sandstone, siltstone, and conglomerate of Pennsylvanian and Mississippian age.

COASTAL PLAIN

The Continental Shelf off the coast of southern New England and Nova Scotia is largely submerged Coastal Plain (fig. 3), which includes Georges Bank south of the Gulf of Maine and Browns Bank south of Cape Sable (fig. 2). Georges Bank is a remnant of an extensive wedge of Coastal Plain sediments that formerly covered much of the floor of the Gulf of Maine. The landward edge of the Coastal Plain between Cape Cod and Cape Sable is a north-facing cuesta that includes the north side of Long Island and Martha's Vineyard. South of New England, water depths over the submerged Coastal Plain range from about 40 to 80 m, in a few places less than 20 m. On the west end of the Scotian Shelf, depths range from about 90 to 180 m. Browns Bank reaches a depth of only about 29 m below sea level.

From Cape Cod northeast into the Gulf of Maine are many remnants of Coastal Plain sediments; only the larger ones are shown diagrammatically on figure 4. The sediments are generally not more than about 100 m thick. (For details, see Ballard and Uchupi, 1975, figs. 3, 5, 6.)

The pre-Quaternary deposits of the Coastal Plain south of New England (Folger and others, 1978; Weed and others, 1974) are clastic sediments, in large part sand, silt, and clay of Late Cretaceous age. The deposits thicken rapidly to the south, reaching about 350 m under Nantucket Island and more than 400 m beneath the southern shore of Long Island (Folger and others, 1978). Sediments of Tertiary age are scattered and thin, including a little fluvial gravel. Beneath the northern slope of Georges Bank, the sedimentary rocks thicken rapidly to a maximum of perhaps 7.5 km, including rocks of presumed Early Cretaceous, Jurassic, and Triassic age in the Georges Bank trough (Ballard and Uchupi, 1975). At the north end of the Baltimore Canyon trough, about

150 km south of Long Island, is a thick section of upper Cenozoic clastic rocks (Scholle, 1977).

The Scotian Shelf is underlain by more than 4,250 m of Cenozoic and Mesozoic sedimentary rocks (Jansa and Wade, 1975), which overlie a basement of metamorphosed rocks of Ordovician and Cambrian age. The lower Tertiary and Cretaceous section, as much as 2,450 m thick, includes continental, deltaic, and marginal marine deposits overlain first by sand and then by shale of a transgressive sequence. The outer part of the shelf is a thick wedge of mudstone of Quaternary and later Tertiary age, thin or absent nearshore but thickening to about 1,200 m near the shelf edge.

Georges Bank forms most of the divide at the mouth of the Gulf of Maine. It has been interpreted as a remnant of the Coastal Plain dissected by two streams, one in the Northeast Channel and the other in the now-filled Great South Channel. These streams drained the interior lowland of the Gulf of Maine north of Georges Bank (Oldale and others, 1974). Northeast Channel is about 100 km long and 32 km wide, and the walls are 90–150 m high. Great South Channel is about 90 km long, 25–40 km wide, and only about 20 m deep. Work in 1975 (Lewis and Sylwester, 1976), however, suggests that the history of the bank is more complex than the earlier interpretation indicates. After Tertiary erosion of the Coastal Plain sediments by streams flowing in the two gaps, extensive Pleistocene deposition took place on and against the bank, followed by marine planation and subsequent deposition of drift of late Pleistocene age.

GULF OF MAINE BASINS

The floor of the Gulf of Maine, excluding the submerged part of the Coastal Plain and of the coastal lowlands, is about 400 km long, about 200 km wide, and has an average depth of about 150 m (Oldale and Uchupi, 1970). Numerous depressions separated by narrow ridges characterize the floor and extend to depths of about 300 m. Some are closed, the closure ranging from a few meters to about 100 m. The lithologic map (fig. 4) shows only a few of the larger remnants of Coastal Plain sediments and areas of Jurassic and Triassic rocks deposited in fault basins. Carboniferous sedimentary rocks, a northeast extension of those near Boston, may be present in the central part of the gulf. (For details, see Ballard and Uchupi, 1975, figs. 3, 5, 6.) The Jurassic and Triassic rocks are presumably similar to those in the Connecticut Valley or in Nova Scotia and New Brunswick around the Bay of Fundy (fig. 4). Pre-Triassic rocks form a high extending from Cashes Ledge east to Yarmouth, Nova Scotia. Jordan Basin is north of this high, and Crowell, Rodgers, and Wilkinson Basins are to the south. The areas of several of the basins at the sill (Emery and Uchupi, 1972) are as follows: Murray-Wilkinson Basins, 10,400 km^2; Jordan Basin, 8,070 km^2; and Georges Basin, 5,200 km^2. The floor of the Bay of Fundy rises rapidly east from Grand Manan Island.

NEWPORT-CONCORD-BOSTON LINE

A major change in topographic grain and to some extent in pattern takes place more or less along a line running from Newport, Vt., south through Concord, N.H., and Boston, Mass. (fig. 3). West of the line, the major topographic units trend slightly west of south; to the east, the trend is more nearly southwest. The change is apparent on both the topographic and lithologic maps (figs. 2, 4) and on the derivative maps, such as the relief map (fig. 10) and the subenvelope map (fig. 11). The line runs more or less along the west side of the belt of alkalic intrusive rocks.

The largest area of the alkalic rocks is in the White Mountains southwest of Mount Washington (figs. 9, 13), where intrusive rocks of Mesozoic age underlie an area about 50 km long from east to west and about 30 km wide. Over most of the area, the relief ranges from about 600 to 1,000 m. The White Mountains probably owe their altitude in part to the presence of these alkalic plutonic rocks. Similar rocks form the Ossipee and Belknap Mountains that rise about 600 m above Lake Winnipesaukee. The lowlands adjacent to the lake are underlain by calc-alkalic plutonic rocks, and the relief ranges from about 60 to 300 m. Small alkalic intrusive bodies occur in southeastern New Hampshire and adjacent Maine. Cape Ann on the coast north of Boston is underlain by alkalic plutonic rocks of Ordovician age. The cape projects east about 16 km into the Gulf of Maine. The relief is more than 60 m, and the sea bottom deepens to more than 60 m within 10 km of the shore. Another area of similar rocks is just south of Boston (Blue Hill), where the relief is more than 180 m.

The Monteregian Hills are a belt of small alkalic intrusive bodies extending from Montreal eastward toward Newport, Vt. The hills are close to the place where the north-trending Champlain Valley meets the northeast-trending St. Lawrence Valley. It is tempting to extend the Newport-Boston line to Montreal. South of Boston, a continuation of the line enters the Coastal Plain near Martha's Vineyard. The Newport-Boston line and the adjacent belt of alkalic intrusive rocks mark the approximate western limit of the Gulf of Maine. West of the line, the inner edge of the Coastal Plain trends more or less west, whereas east of the line, the inner edge of the Coastal Plain trends north for nearly 200 km.

DRAINAGE

DIVIDES AND BASINS

The rivers of the New England region flow either to the Hudson-Champlain-St. Lawrence lowlands or to the Atlantic Ocean. The major drainage divide runs from New York City on the Atlantic Coast northeast to within about 20 km of the St. Lawrence River near Rivière-du-Loup, Quebec (fig. 12).

About one-third of the New England region is northwest of the divide and about two-thirds is to the southeast. The northern and southern segments of the major drainage divide follow the Hudson-Green-Notre Dame highlands, including the Taconic Mountains, Hoosac Range, and southern Green Mountains to the south and the Notre Dame Mountains to the north. In central Vermont (about lat 44° N.), the drainage divide leaves the Green Mountains and trends east for about 50 km, where it turns north and northeast through northeast Vermont to the Boundary Mountains of Maine and the Megantic Hills of Quebec. The central segment of the major divide in northeast Vermont and along the Quebec-Maine border is about equidistant from the St. Lawrence River to the northwest and the shore of the Gulf of Maine to the southeast. The central segment is in an area of hills and low mountains. The highest part of the central highlands, including the White Mountains and Mount Katahdin, is 50–100 km southeast of the central segment of the major drainage divide.

The largest rivers in the New England region, the St. François, Chaudière, St. John, Penobscot, Kennebec, Androscoggin, and Connecticut have their headwaters in the central segment of the major drainage divide. The largest of the seven rivers, the St. John, flows southeast for perhaps half its course more or less across the north and northeast trend of the major bedrock units. From Grand Falls south to the vicinity of Houlton, Me., the river follows a belt of Carboniferous sedimentary rocks, largely unmetamorphosed. The second largest river, the Connecticut, flows parallel to the trend of the bedrock. The remaining five rivers have courses that are roughly half across the bedrock structures and half parallel with them.

The St. John River is about 725 km long and has a drainage basin of about 54,900 km². The northwest border of the drainage basin in the Notre Dame Mountains is only about 20–40 km southeast of the St. Lawrence River. The drainage basin of the St. John is about three-quarters the size of that of the Susquehanna River and nearly twice that of the Hudson River, as shown in the following table:

River	Length (km)	Drainage area (km²)
Tributary to Hudson-Champlain-St. Lawrence lowlands		
Richelieu	370	23,600
St. François	300	9,970
La Chaudière	240	6,600
Bècancour	190	2,720
Yamaska	180	4,660
du-Loup	50	960
Otter Creek	160	2,410
Missisquoi	135	2,100
Winooski	135	2,770
Lamoille	130	1,815
Hoosic	105	1,740
Batten Kill	95	1,140
Kinderhook Creek	70	1,375
Wappinger Creek	55	490
Tributary to Atlantic Ocean		
St. John	725	54,900
Connecticut	625	29,000
Kennebec	410	25,600
Penobscot	370	21,000
Androscoggin	355	8,800
Merrimack	290	12,700
Housatonic	225	4,900
Saco	185	4,560
Quinebaug	135	3,650
Presumpscot	115	1,605
Charles	105	775
Machias	95	1,320
Taunton	80	1,370
Tributary to Connecticut River		
Farmington	130	1,580
Deerfield	120	1,605
Chicopee	105	1,320
Westfield	95	1,320
White	95	1,710
Ashuelot	95	985
Ammonoosuc	90	1,035
West	80	1,190
Millers	80	985
Passumpsic	65	1,190
Sugar	50	490
Other major rivers of northeastern North America		
Hudson	490	34,700
Delaware	625	29,800
Susquehanna	715	71,500
St. Lawrence[1]		1,183,500 [2]828,800

[1] Above mouth of Rivière Saguenay, about 30 km north of Rivière-du-Loup.
[2] Below Niagara Falls, from headwaters of Ottawa River.

The second largest river in the New England region is the Connecticut, which flows south from the Boundary Mountains on the Maine-Quebec border for a total length of about 625 km, only about 100 km less than that of the St. John River. The Connecticut River drainage basin is long and narrow. The drainage area of 29,000 km² is about half that of the St. John River. The basins of these two rivers more or less enclose those of the four other large rivers that drain to the Atlantic—the Kennebec, Penobscot, Androscoggin, and Merrimack.

The rivers that flow northwest from the central segment of the major drainage divide cross the adjacent mountains – in northern Vermont, the higher parts of the Green Mountains, and in Quebec, the Sutton Mountains and the southern end of the Notre Dame Mountains. The longest of the northwest-flowing rivers is the St. François, which is about 300 km long and has a drainage basin of about 9,970 km^2. Below Sherbrooke, Quebec, the drainage basin is long and narrow where the river crosses the Hudson-Green-Notre Dame highlands and the St. Lawrence lowlands. Above Sherbrooke in the central highlands, a large headwater basin extends northeast, parallel to the bedrock structure from the vicinity of Newport, Vt., to Thetford Mines, Quebec, a distance of nearly 200 km (fig. 11). The headwater basin is underlain in part by calc-silicate metasedimentary rocks (fig. 5). The west-flowing rivers in Vermont are smaller than those in Quebec. The Winooski River, about 135 km long, heads in the central highlands and crosses the Green Mountains between Camels Hump and Mount Mansfield in a narrow valley about 700 m deep. The west-flowing tributaries of the Hudson River are short; the largest, the Hoosic River, is 105 km long and has a drainage basin of about 1,740 km^2.

STREAMS, DIVIDES, AND AREAS OF CALC-ALKALIC PLUTONIC ROCKS

The longer streams of the New England region are largely in areas underlain by sedimentary or metasedimentary rocks. Areas of calc-alkalic plutonic rocks are drained by shorter streams and, in some places, form divides (fig. 13). Why the large rivers appear to avoid granitic plutons has no ready explanation. Several possibilities are discussed in later paragraphs. Fredericton, New Brunswick, is on a segment of the St. John River that runs eastward for about 100 km and crosses a northeast-trending belt of calc-alkalic plutonic rocks. The drainage basin narrows toward the area where it is crossed by the belt of plutonic rocks, and the river's course is close to the narrowest part of the belt.

Although the drainage basins of both the Kennebec and the Penobscot Rivers include large areas of calc-alkalic granitic rocks, the main channels of the rivers are largely in metamorphic rocks. The divide on the east side of the Penobscot basin is in or near areas of plutonic rocks. The Androscoggin River, about 355 km long, flows east from Berlin, N.H., for about 80 km around the north side of a large granite pluton that is drained largely by the Presumpscot River, which is only about 115 km long.

The Merrimack River at Concord, N.H., turns slightly east to skirt a small area of plutonic rock. To the south, the river crosses a northeast-trending belt of plutonic rock that is about 13 km wide and a second narrow belt about 1.5 km wide. Near the Massachusetts State line, however, the river turns northeast and flows for about 50 km close to the north edge of a body of plutonic rock. The north edge of the granite pluton is the Clinton-Newbury fault (Cameron and Naylor, 1976). The Taunton River in southeastern Massachusetts heads in areas of plutonic rock, but most of its drainage basin is underlain largely by sandstone, graywacke, and shale of Carboniferous age.

CONNECTICUT RIVER

The Connecticut River in New Hampshire and Vermont runs more or less parallel to the trend of the bedrock units as far north as the mouth of the Passumpsic River. North of the Passumpsic, the trend of the river is northeast across the bedrock structure. The upper Connecticut River traverses calc-alkalic plutonic rocks for about 30 km. South of the mouth of the Passumpsic, the river avoids areas of calc-alkalic plutonic rocks. Two exceptions are in the highlands east of Middletown, Conn. (fig. 8), and just south of Brattleboro, Vt. (fig. 9). From Littleton, N.H., south to Claremont, the Connecticut River follows closely the trace of the Ammonoosuc thrust (fig. 9). In New Hampshire, the divide east of the river follows a wide belt of plutonic rock for about 160 km. However, the actual divide for more than half its length is in or close to narrow belts of metasedimentary rock between larger areas of plutonic rock. In northeast Vermont, the divide on the west side of the Connecticut River is also in or adjacent to areas of plutonic rock.

In southern New England, the Connecticut Valley follows a belt of sandstone, shale, and conglomerate of Jurassic and Triassic age. From Springfield to Middletown (fig. 8), the Connecticut River flows east of long narrow outcrop belts of mafic volcanic rocks that form ridges. South of Middletown, the volcanic rocks extend southeast across the area of Jurassic and Triassic rocks to abut against the older Paleozoic rocks. The river turns east at Middletown and enters the pre-Triassic rocks. It appears that the river's southerly course is blocked by the volcanic rocks and that the river enters the older rocks of the central highlands where there is a break in a belt of felsic gneiss that is close to or that forms the east border of the Jurassic and Triassic rocks. The river flows around the south end of the Great Hill syncline, where the ridge-forming Silurian quartzite plunges out (loc. 1, fig. 8). Near Long Island Sound, the Connecticut River flows across the northeast trend of the bedrock.

HUDSON RIVER

Throughout most of its course, the Hudson River follows a lowland underlain largely by dolomite, limestone, and shale of Ordovician and Cambrian age. However, the course of the river through the highlands south of Newburgh, N.Y., is not easy to explain in terms of the bedrock geology (fig. 7). South of the highlands, on the other hand, the Hudson River estuary, 15–20 km wide, follows the contact between rocks of Late Triassic age to the west and those of Paleozoic and Proterozoic age to the east. The estuary skirts the east side of a ridge of diabase (the Palisades) and, at the north end, the west side of an area of mafic and ultramafic rocks of Late Ordovician age (norite and pyroxenite of the Cortland Complex).

South of Newburgh, the Hudson River leaves the broad lowlands underlain by shale, dolomite, and limestone and traverses a deep narrow gorge for about 100 km through the Proterozoic rocks of the Hudson Highlands. Except near the northwest side of the highlands, the river follows a break in lithology. West of the river, the bedrock is chiefly hornblende granite and granite gneiss; to the east, the dominant bedrock is biotite granite gneiss. The river also tends to follow faults, both the major northeast-trending structures and shorter cross faults. Only in the first 25 km of the gorge south of Newburgh does the river flow southeast at right angles to the northeast trend of the faults and of the major lithologic units. Perhaps the course of the river in the first 25 km is related to cross joints in the hornblende granite.

STREAM GRADIENTS, THE SUBENVELOPE

The derivative map portraying the general altitude of the drainage network, the subenvelope (fig. 11) shows how stream gradients differ from one drainage basin to another or from one physiographic unit to another. The map also shows at a glance how the gradient of a river changes where the nature of the adjacent bedrock changes.

The Connecticut River, for example, from its mouth north for about 400 km (junction with Passumpsic River) rises less than 90 m. The average gradient is about 0.2 m per km. Between the mouth of the Passumpsic and Littleton, N.H., the Connecticut River turns northeast and crosses belts of metavolcanic rocks (Ammonoosuc Volcanics, Billings, 1955), where the bed rises about 90 m in a distance of about 24 km (fig. 9). The rise, formerly known as Fifteen Mile Falls, is now buried by the pools impounded behind two dams. Farther upstream, the gradient flattens where the river traverses calc-alkalic plutonic rocks for about 30 km. In Connecticut and Massachusetts, stream gradients are low in areas underlain by rocks of Jurassic and Triassic age. The tributary streams entering the Connecticut River from the west head in areas of Proterozoic rock forming the Berkshire Hills. Their gradients are steeper than those of the tributary streams that enter from the east where there are large areas of calc-alkalic plutonic rock.

In New Hampshire and Vermont about as far north as lat 44° N., the drainage basin of the Connecticut River is asymmetric; the western tributaries are longer than those entering the river from the east (fig. 9). The White River valley drains a large area of metasedimentary rocks and has a gentler gradient than that of the West River that has a large area of Proterozoic gneiss and quartzite in its headwaters. Near Claremont, the gradients of the valleys entering the Connecticut River from the west are steeper than those of either of the rivers mentioned above, probably because Proterozoic rocks and alkalic plutonic rocks are present near the river. Scattered observations suggest that streams, such as the White River, in areas of metasedimentary rocks have many rock outcrops in their beds, whereas streambeds in or near areas of Proterozoic rock or alkalic plutonic rock are covered with boulders, outcrops are scarce, and stream gradients tend to be steeper than those in areas of metasedimentary rocks.

The highlands between the Merrimack River and the Connecticut River contain large areas of calc-alkalic plutonic rock, where, in general, stream gradients are lower than those in adjacent rocks. For example, the Ashuelot River (fig. 9) south of Keene, N.H., has a lower gradient in the plutonic rocks than it does upstream. The gradients of streams east of Lebanon, N.H., are steeper near the Connecticut River than they are to the east, where the streams are largely in plutonic rocks.

A striking feature of the subenvelope map (fig. 11) is its delineation of the coastal lowlands (fig. 3). Stream gradients are gentle in the coastal lowlands and steepen rapidly along the edge of the central highlands. These features do not appear to coincide with any bedrock units. Similar rocks are found in both the coastal lowlands and adjacent parts of the central highlands. A belt of steep stream gradients extends from eastern Connecticut north and northeast to the vicinity of Mount Katahdin, Me. (fig. 5). In detail, the line where the envelope steepens extends west and northwest up the large valleys and east and southeast where valleys are small. The line varies in altitude from place to place, as follows:

Area	Altitude (m)
Worcester, Mass	100
Concord, N.H	100
South of White Mountains	150
Southeast of White Mountains	150
West of Rumford, Me	200
Rumford to Mount Katahdin, Me	150

In Maine, the beds of the major rivers rise from less than 120 m to more than 240 m above sea level where the streams cross the 8-16-km-wide belt of steep stream gradients. In the belt, both altitude and relief rise about 300 m.

The belt of steep stream gradients includes several conspicuous gaps or notches that cross the mountains (Franconia, Crawford, Grafton). At the head of the Saco River in Crawford Notch, southwest of Mount Washington (fig. 11), is a prominent step that marks the Gulf of Maine-Connecticut River divide. The riser of this step is a headwater tributary of the Saco River that has a steep gradient to the south. The tread is a headwater tributary of the Ammonoosuc River that has a gentle gradient to the northwest.

The Maine part of the St. John River basin is largely an area where stream gradients are low, although the altitude of the channel of the main stream ranges from about 150 to 300 m. The St. John River rises about 60 m from its mouth at St. John to Grand Falls, New Brunswick; the average gradient is about 0.17 m/km. In New Brunswick, many of the tributaries to the St. John have steep gradients near the river and gentle gradients near their headwaters.

From Quebec City southwest to Montreal and south up the Champlain Valley about as far as lat 44° N., the principal northwest-flowing streams have low gradients where they cross the Hudson-Green-Notre Dame highlands. In southeastern Quebec, the tributaries to the Chaudière and St. François have low gradients, whereas to the south in Vermont, the tributaries to the Lamoille and Winooski have steep gradients. The Winooski River crosses the highest part of the Green Mountains. The two Quebec rivers are roughly twice the size of the two Vermont rivers in length and drainage area. The principal tributaries that enter the Hudson River from the east, Batten Kill, Hoosic River, and Kinderhook and Wappinger Creeks, have gentle gradients in their upper reaches in areas underlain largely by dolomite, limestone, and shale. Gradients are steep where the streams are in the area of shale and graywacke of the Taconic highlands just east of the Hudson River. The Hoosic River and Batten Kill head on the west slope of the Green Mountains. The headwater streams have very steep gradients and are floored with quartzite boulders as large as 3 m in diameter derived from the adjacent Cheshire Quartzite of Cambrian age (Doll and others, 1961).

AGE OF THE LANDSCAPE

A few widely scattered deposits of Cretaceous or Tertiary age on land and extensive Coastal Plain deposits largely beneath the sea suggest that the present land areas have been above sea level throughout long periods of Cenozoic and later Mesozoic time. The volume of sediment under the Coastal Plain has been used to estimate rates of erosion and of uplift of the adjacent landmass. If the estimated rates of erosion and uplift are of the right order of magnitude, it is extremely unlikely that any part of the present landscape dates back before the late Miocene. The only exceptions are areas buried beneath Cenozoic and upper Mesozoic deposits that have been exhumed perhaps by stream and by glacier erosion during the later Cenozoic.

The New England landscape is the result of long-continued subaerial weathering and erosion induced by long-continued uplift or upwarping of the land. Such processes may have gone on since the Early Cretaceous. During this span of perhaps 120 million years or more, the drainage has been continually adjusting to changes in the lithologic and structural character of the bedrock as the ground surface has been lowered.

In the Middle Atlantic States, there have been two major periods of gravel deposition on the Coastal Plain, one in the Early Cretaceous and the other in the Miocene and perhaps the Pliocene (Owens, in press). Similarly, on the Scotian Shelf, a thick wedge of clastic deposits of Cretaceous and early Tertiary age has been derived from a source area to the north and northwest, probably in the drainage basin of an ancestral St. Lawrence River. A thick wedge of mudstone of later Cenozoic age forms the outer part of the shelf.

The base of the Upper Cretaceous sedimentary rocks (U.S. Geological Survey, 1967; Oldale and Uchupi, 1970) of the Coastal Plain beneath Long Island, Martha's Vineyard, and Nantucket slopes steeply to the south (fig. 14). The base of the sedimentary rocks–the basement surface–if projected landward to the north shore of Long Island Sound, more or less coincides with the interfluves near the coast. In a 20-km-wide coastal belt, the interfluves rise steeply to the north. The belt is called the "Fall Zone" by Flint (1963) and, as shown in figure 15, is the envelope of this report. The average slope of the basement surface shown in figure 15 is about 2.3 m/km. If extended northwest, the surface clearly passes far above the envelope, suggesting, as Flint said (1963, p. 683) "* * * recurrent (or continuous) arching beginning before late Cretaceous time." A map of the basement surface south of Rhode Island (McMaster and Ashraf, 1973) shows a smooth surface indented by subparallel channels; the whole slopes south at about 8 m/km. Clearly, the surface has been tilted south; such a slope is far too steep for the drainage as restored by McMaster and Ashraf.

Estimated rates of erosion suggest that New England has risen in later Cenozoic time. The rate at which the Appalachian Highlands have been lowered by erosion has been estimated by Mathews (1975) on the basis of the volume of sediment in the emerged and submerged Coastal Plain along the Atlantic coast and by Hack (1980) on the basis of the load carried by rivers draining the highlands (see fig. 15). The lowering of the highlands of Connecticut by about 500 m would require about 20 million years if Mathews' (1975) rate of 2.7×10^{-2} mm/yr is used, whereas about 13 million years would be required if Hack's (1980) rate of 4.0×10^{-2} mm/yr is used. The amount of vertical lowering of the landscape illustrated in figure 15 shows that in 15–20 million years, erosion would reduce Connecticut to broad lowlands unless the area was gradually upwarped during that interval. If the Pleistocene and Pliocene lasted about 5.4 million years and the Miocene 20 million years, clearly southern New England has been upwarped in later Cenozoic time. These erosion rates also show that when deposition of the Upper Cretaceous sediments beneath Long Island began, the land surface to the north was hundreds of meters above the position of the envelope shown in figure 15.

Other evidence suggests considerable erosion of the central highlands. In a study of Paleozoic regional metamorphism in New England, Thompson and Norton (1968) suggested that as much as 15 km of rock may have been removed by erosion since the late Paleozoic. Fission-track dates from apatites from igneous and metamorphic rocks (Zimmerman and Faul, 1976) in the central highlands suggest a possible uplift and erosion of 3–5 km during the last 64–110 million years.

The Gulf of Maine may be largely an erosional feature dating from Cretaceous time (Oldale and others, 1974). Oldale and Uchupi (1970, fig. 3) have prepared a reconstruction of the surface beneath the Cretaceous and younger sediments that suggests a stream-eroded landscape. They (1970, p. B–171) believe that the present floor of the gulf is the result "* * * of fluvial erosion probably during Pliocene and early Pleistocene time." However, the Gulf of Maine may be in part a structural basin (Owens, in press) modified by erosion and deposition. The western margin of the gulf follows the east side of the north-northwest-trending belt of alkalic granites (the Newport-Boston line) along which a change in topographic form and trend takes place. Perhaps the gulf was downwarped along this line.

Small and widely scattered lignitic deposits in the Appalachian region (Pierce, 1965) suggest that this region has stood above sea level since the Late Cretaceous. The lignitic deposits at Brandon, Vt., about 20 km north of Rutland (Barghoorn and Spackman, 1950), are reported to be of Oligocene age. Lignite from a clay bed in central Nova Scotia is of Early Cretaceous age (Stevenson and McGregor, 1963). Similar deposits have been found in Pennsylvania, Tennessee, Georgia, and Alabama. At Brandon, as well as in southern Pennsylvania and in Nova Scotia, the plant remains suggest a wet depression, probably a sink. The Brandon locality is in an area underlain by dolomite and limestone of Ordovician and Cambrian age. No marine fossils or any dinoflagellates or other fossils commonly associated with nearshore pollen-bearing sediments were found. Therefore, the deposits at Brandon probably accumulated in a nonmarine environment. The association of the deposits at Brandon and elsewhere with carbonate rocks and the absence of any stratification or sequence of fossils in the deposits suggests that they accumulated in a sinkhole pond and have been lowered many hundreds of meters and deformed by continued solution of the underlying bedrock. Pierce (1965) suggested that the Upper Cretaceous Pond Bank lignitic deposit in southern Pennsylvania could have been lowered more than 425 m on the basis of estimates of the average amount of material eroded from the Applachian Highlands since Late Cretaceous time. "If this amount of lowering has taken place, then the late Cretaceous position of the deposit was several hundred feet [100 m] above the highest mountains now present in the area" (Pierce, 1965, p. C155).

RELATION OF COASTAL PLAIN AND GULF OF MAINE BASINS TO THE EMERGED COASTAL PLAIN OF THE MIDDLE ATLANTIC STATES

The Coastal Plain and the Gulf of Maine basins of the New England region have some features in common with Chesapeake and Delaware Bays and with the Delmarva Peninsula that separates the two bays. However, the emerged Coastal Plain in the Middle Atlantic States has many characteristics that differ from those of the submerged Continental Shelf to the north. The upper Cenozoic sediments of the Delmarva Peninsula, Delaware, Maryland, and Virginia (Owens and Minard, 1975, 1979; Owens and Denny, 1979), include large amounts of coarse clastic deposits laid down in fluvial and deltaic environments. The Baltimore Canyon off the New Jersey coast about 150 km south of Long Island contains about 1,000 m of largely clastic sediments of Miocene and younger age (Scholle, 1977). South of southern New England, rocks of Tertiary age are of limited extent, are largely fine-grained sediments, including only a few small masses of fluvial gravel (Folger and others, 1978; Minard and others, 1974; Perry and others, 1975).

Chesapeake and Delaware Bays have a different geologic setting from that of the Gulf of Maine. The gulf is, at maximum, only about 400 m deep and contains only small amounts of Coastal Plain sediments. It is north of the line along which the surface at the base of the Cretaceous rocks is steeply downwarped to the southeast (fig. 14), the line being the submerged easterly extension of the Fall Line of the Middle Atlantic States (Mathews, 1975). If the Gulf of Maine is in part a structural feature, it represents only a gentle downwarp. Chesapeake and Delaware Bays, on the other hand, are seaward of the Fall Line and are part of the much larger Salisbury embayment (Anderson, 1948; Robbins and others, 1975; Owens, in press). The embayment extends parallel to the coast about 280 km and at right angles to the coast for about 180 km, roughly about two-thirds the area of the Gulf of Maine. However, the embayment near the Maryland coast contains about 1,800 m of Coastal Plain sediments.

In the Gulf of Maine, the surface beneath the Coastal Plain sediments, the basement surface as reconstructed by Oldale and Uchupi (1970), resembles a landscape (fig. 14) of broad valleys separated by low divides, the valleys being the seaward extension of many of those on land. Georges Bank forms most of the divide at the mouth of the gulf and is breached at two points. The eastern gap, Georges Basin and the Northeast Channel, carried the drainage from the eastern two-thirds of the floor of the gulf and from the Penobscot and St. John Rivers and the Bay of Fundy, a total drainage area of about 190,000 km^2, or nearly three times that of the Susquehanna River. The western gap, Great South Channel, has a more complex Pleistocene history (Lewis and Sylwester, 1976). Nevertheless, the channel is believed to mark the position of the stream that in the Tertiary carried drainage from the western third of the floor of the Gulf of Maine, and to have been the seaward extension of the Charles, Merrimack, Androscoggin, and Kennebec Rivers. The total area that drains to the mouth (south end) of the Great South Channel is about 99,000 km^2, roughly about one-third larger than the Susquehanna River drainage basin.

The basement surface or floor of the Gulf of Maine, as reconstructed by Oldale and Uchupi (1970), suggests an erosional surface. If the gulf is indeed a structural depression, it appears to have been extensively modified by erosion and deposition. Erosion of large masses of Coastal Plain sediments is reasonable in terms of the amount of material that may have been removed and the size of the streams that are believed to have done the work. Georges Basin and the Northeast Channel and the drainage basin upstream are comparable in size with similar valleys in the Coastal Plain south of the glaciated region, such as Chesapeake Bay, as shown below:

	Size (km^2)	Drainage area (km^2)
Georges Basin and Northeast Channel	12,200	176,000
Chesapeake Bay	10,500	167,000

North and east of the New England region, the Laurentian Channel (Canada Hydrographic Service, 1973), similar to the Northeast Channel but on a much larger scale, follows the St. Lawrence River estuary from near Rivière-du-Loup (mouth of Rivière Saguenay) for about 1,200 km through the Gulf of St. Lawrence and enters the Atlantic Ocean between Nova Scotia and Newfoundland. Where the channel crosses the submerged Coastal Plain south of Nova Scotia, it is about 80 km wide and 180–300 m deep. The Northeast Channel southwest of Novia Scotia is 32 km wide, and the walls range from 90 to 150 m in height. Above the mouth of the Saguenay, the St. Lawrence River drainage basin below Niagara Falls is about 830,000 km^2. The Laurentian Channel is presumed to be the result of glacial erosion and deposition in the ancestral St. Lawrence valley (Shepard, 1931; Emery and Uchupi, 1972; Loring, 1975).

The configuration of the Continental Shelf of northeastern North America suggests that the region has been submerged in late geologic time (Eardley, 1964). However, we do not know the position of sea level in the New England region in the Miocene when glaciers are presumed to have begun to form in the Antarctic and when glacially controlled eustatic changes in sea level became a reality. Nor do we know whether northeastern North America has returned to the same position with respect to the center of the Earth as it had before glacial loading.

Studies in the Coastal Plain of the Middle Atlantic States suggest that in the Miocene the Hudson River did not enter the ocean near its present position but that it flowed southwest from New York City near the inner edge of the Coastal Plain and entered the ocean in the vicinity of the Delmarva Peninsula. Beginning in the Miocene, large volumes of fine clastic materials were deposited in the Coastal Plain of the Middle Atlantic States. The sediments coarsen upward, culminating in deposition of coarse clastic materials in the Pliocene. The clastic deposits came from the northwest and the north. Owens and Minard (1975, 1979) have suggested the Hudson Valley as the probable source of Miocene fluvial deposits (Bridgeton Formation or "Arkose 2") near the inner edge of the New Jersey Coastal Plain.

South of the mouth of the Hudson River, no large channel crosses the Continental Shelf that is similar in size to the Northeast Channel at the mouth of the Gulf of Maine. A shallow channel crosses part of the shelf, to end near the head of the Hudson Canyon on the Continental Slope (Uchupi, 1965). The channel is about 100 km long, 1 to 10 km wide, and about 20 to 40 m deep. Perhaps the Northeast Channel is large because it was widened and deepened by an ice sheet during Wisconsinan time. The area that may have drained seaward through the Northeast Channel (176,000 km^2) is more than four times larger than that of the present Hudson River (34,700 km^2). Perhaps this difference in drainage area accounts in part for the difference in size of the two submerged channels. A buried valley in the Continental Shelf south of New York City, apparently a Pleistocene course of the Hudson River, is about the same size as the shallow channel on the shelf at present (Knebel and others, 1979). A third possibility is that the present course of the Hudson was established in post-Miocene time and that in the Miocene, as Owens and Minard have suggested, the river flowed southwest near the inner edge of the Coastal Plain and entered the Atlantic Ocean in the vicinity of the Delmarva Peninsula. Perhaps at the same time, other rivers were flowing across the floor of the Gulf of Maine and through the channels in the Coastal Plain to the south.

LANDSCAPE EVOLUTION

The landscape of the New England region is closely related to the character and distribution of the bedrock units. The relationships suggest that long-continued subaerial weathering and erosion aided by concurrent rise of the land has produced a landscape in conformity with the geologic framework of the region. These subaerial processes may have been in operation during much of post-Early Cretaceous time. The control that bedrock exerts on land form is evident throughout the area. The volcanic rocks of Jurassic and Triassic age form steep-sided narrow ridges. Shale and graywacke of Ordovician, Cambrian, and Proterozoic Z age form the Taconic highlands. Steep-sided mountains and hills are the topographic forms of alkalic plutonic rocks. Sandstone and shale of Jurassic and Triassic age, as in the Connecticut Valley, similar rocks of Carboniferous age, as near Boston, and dolomite, limestone, and shale of Ordovician and Cambrian age, as in the Vermont Valley, commonly form lowlands. The higher parts of the southern half of the Hudson-Green-Notre Dame highlands are underlain largely by resistant beds of massive gneiss and of subordinate quartzite of Proterozoic age. The northern half of the highlands is largely Paleozoic metasedimentary and metavolcanic rocks of both amphibolite and greenschist facies. Massive intermediate to high-grade metapelites are more resistant than are similar schists of low metamorphic grade. Calc-silicate metamorphic rocks and feldspathic plutonic rocks tend to form lowlands, although ease of disaggregation and differences in massiveness of the plutonic rocks influence their topographic expression.

The central highlands include not only the highest peaks in New England – the White Mountains and Mount Katahdin – but also large areas of low mountains and hills. The geology of the central highlands is diversified, and at the map scale used in this study, it has not been possible to define the lithology and structure of the central highlands adequately to permit a detailed discussion of the physiographic features of this unit.

The White Mountains, Mount Katahdin, and the higher peaks between them are 300–800 m above the coastal lowlands to the southeast. These peaks are in the center of the New England region, roughly equidistant from the St. Lawrence River to the northwest and the Gulf of Maine to the southeast. If the principal streams radiated out from these peaks in all directions, one might suppose that they are at the center of a crustal block that has been slowly rising for a long time to more or less balance the lowering of the peaks by erosion. However, such is not the case. The Connecticut and Androscoggin Rivers rise about 100 km north of Mount Washington. The former flows south and southwest around the west side of the White Mountains; the latter runs southwest to Berlin, N.H., where it turns east to enter the ocean near Portland, Me.

Although the data presented here clearly show that no part of the present landscape is probably older than the Miocene, major lowlands such as the Hudson-Champlain-St. Lawrence lowlands, the Vermont Valley, the Connecticut Valley, and the Gulf of Maine might have been in existence throughout the Tertiary Period. Streams flowing from Mount Washington and adjacent peaks west to the Connecticut River, and the river itself north of the mouth of the Passumpsic, cross a belt of resistant rocks, including the Ammonoosuc Volcanics and the Clough Conglomerate (Billings, 1955), where stream gradients are steep. Streams flowing south from the White Mountains, such as the Merrimack and the Saco, head in the largest area underlain by alkalic plutonic rocks in the New England region, rocks that clearly form steep-sided mountains. Probably these geologic relations have been causing streams to have steep gradients for tens of millions of years, and the geographic arrangement of the bedrock units has led to the formation of the White Mountains. The mountains have been buttressed to the south and west by the alkalic plutonic rocks. The metasedimentary rocks of the White Mountains and adjacent western Maine are largely amphibolite and granulite facies. Such rocks underlie the summit of Mount Washington.

In central Maine, the metasedimentary rocks in the higher mountains are greenschist facies. Many of the higher peaks are massive rocks, either metavolcanic rocks or contact-metamorphic rocks surrounding plutonic rocks. The mountains consist of somewhat isolated peaks, whereas in the White Mountains and to the south, the mountains tend to be somewhat elongate.

The coastal lowlands are a half-submerged and half-emerged piedmont at the foot of the central highlands.

The lowlands may have undergone long periods of erosion beginning in the Early Cretaceous. The northwest border of the lowlands is, in a few places, a low, gently sloping escarpment. Throughout most of its length, however, the border is a belt 15–30 km wide, where from southeast to northwest, the relief and stream gradients increase as compared with those in the lowlands. In detail, the southeast border of the highlands appears to be an erosional feature. No structural evidence for tectonic movement, perhaps related to the uplift of the White Mountains-Mount Katahdin highlands, has been reported. Many of the same rock units occur in both highlands and lowlands. A belt of calc-silicate rocks of Silurian and Ordovician age underlies areas in the coastal lowlands. These fairly nonresistant rocks may be a part of the reason for the lowlands. However, south of Worcester, Mass., the calc-silicate belt is in the central highlands, and in much of Maine, the western edge of the belt is 10–40 km east of the western edge of the coastal lowlands. It is difficult to escape the conclusion that the highlands have been raised by tectonic movements, presumably some sort of arching or doming relative to the adjacent lowlands.

Why many of the large rivers do not cross areas of calc-alkalic plutonic rock, but rather take a longer course around such areas, cannot readily be explained. Do these phenomena have a climatic explanation? Was the climate of New England warmer and wetter in the early Cenozoic than it has been since that time? Perhaps the plutonic rocks formed highlands in the early Cenozoic, which have been lowered by erosion during the cooler later Cenozoic. In spite of such a change in climate, the major streams still maintain their early Cenozoic courses around the areas of plutonic rock. Any such explanation hardly seems adequate.

Are the phenomena related to the contrast between the action of a stream flowing in coarse-grained plutonic rocks and that of one flowing in fine-grained foliated rocks? As a broad generalization, one can perhaps say that modern rivers draining areas of coarse-grained rock tend to have beds and banks of sand and gravel, whereas rivers draining areas of fine-grained rock tend to have beds and banks of silt and clay. The former rivers have shallow channels and tend to migrate from side to side; the latter have deep channels and tend to maintain their courses. Perhaps such processes acting over long periods of time (10 million years) cause rivers to move from areas of plutonic rock into areas of fine-grained rock.

Although the map of the New England region outlining physiographic units (fig. 3) shows a narrow belt of coastal lowlands in southern Connecticut, the topography in the belt is more rugged than that in the lowlands elsewhere. The coast of southern New England trends more or less at right angles to the major bedrock units, which to the south pass beneath the Coastal Plain. Southern New England has been upwarped relative to the Coastal Plain to the south in late Mesozoic and in Cenozoic time (fig. 14). Oldale and Uchupi (1970) show streams on the land extending down the slope of the basement surface. The gradient of streams on the basement surface is much steeper than that of their possible headwater reaches on land, suggesting that the basement surface has been tilted down to the south relative to the adjacent landmass. The deformation could be Late Cretaceous or younger.

TECTONISM

One of the major problems of New England geomorphology is how important tectonic movements have been in the shaping of the modern landscape. The evidence for Cenozoic tectonic movements in the Coastal Plain and Piedmont of Southeastern United States (Mixon and Newell, 1977; Prowell and O'Connor, 1978), as well as in the Adirondacks (Isachsen, 1975, 1978), raises the possibility of similar tectonic activity in the New England region.

The character of the Coastal Plain sediments tells us something about the tectonic history. The arrival of coarse clastic sediments on the Coastal Plain in the Middle Atlantic States in the late Miocene and in the Pliocene suggests the relative uplift of the adjacent Piedmont and of the Hudson River drainage basin. If the central highlands of the New England region were elevated in the late Tertiary, one might expect that coarse clastic materials should have been deposited in the Coastal Plain to the south and east. Although such sediments are not found near shore, a thick section of upper Cenozoic clastic rocks is present at the north end of the Baltimore Canyon trough. The coarse clastic sediments shed by the central highlands may have been deposited close by, that is, in the coastal lowlands, not 50–100 km to the east as part of the submerged Coastal Plain of the present day. The lowlands adjacent to the highlands contain extensive deposits of sand and subordinate gravel of Pleistocene age (National Research Council, Division of Earth Sciences, 1959). Perhaps some of these sediments were derived from older clastic deposits of Tertiary age.

If the central highlands were elevated relative to the coastal lowlands in Miocene and Pliocene time, it is interesting to note that the break between highlands and lowlands is just southeast of a belt of calc-alkalic granites. Richard Goldsmith (written commun., 1978) has suggested that, because such granites are emplaced in or near the cores of orogens, they would be the sites of former mountains. Such areas commonly are gravity lows (Kane and others, 1972). As the mountains are lowered by erosion, they tend to rise because of isostasy. In Maine, a gravity gradient that slopes down to the northwest coincides in places with the topographic break (Kane and others, 1972).

Estimated rates of erosion by streams draining to the North Atlantic Ocean (fig. 15) suggest that highlands such as the White Mountains would be reduced to hills and lowlands within about 30–40 million years. Such estimates indicate that the doming or arching of such highlands either individually or as a unit, cannot be older than middle Tertiary. No movement has been shown to have taken place along any fault planes in the New England region in later Cenozoic time. Isachsen (1975, 1978) believes that there may have been Holocene deformation in the Adirondack region, perhaps including the adjacent Hudson-Champlain lowlands.

The bedrock geology of the Piedmont in the central and southern Appalachians (King and Beikman, 1974) is similar in many respects to that of New England. The topography, on the other hand, is different. In New England, the areas of high relief are much more extensive than they are in the Piedmont to the south (Hack, 1980, fig. 5), perhaps indicative of upwarping of the highlands of New England in late Cenozoic time.

The St. Lawrence lowlands near Quebec City are seismically active, and one wonders whether the lowlands might not be the result, in part, of tectonic movements. However, "the concentrations of earthquake epicenters [in the lowlands] are apparently not controlled by near-surface geological elements" (Poole and others, 1970, p. 300 and fig. VI-32).

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