Hydrography

The NAWQA Program evaluates the quality of surface and ground waters and the aquatic biological communities found in rivers and streams. An understanding of the physical aspects of surface- and ground-water systems (the hydrography) is needed before these evaluations can be made. The following sections describe in detail the hydrography of the New England Coastal Basins study area.

Surface Water

The study are encompasses four large drainage basins and many smaller coastal drainages (fig. 7). Most of the rivers that enter the Atlantic Ocean directly are tide-affected for some distance upstream from the mouth. The large rivers are the Kennebec (5,893 mi^2), the Merrimack (5,010 mi^2), the Androscoggin (3,524 mi^2), and the Saco (1,700 mi^2) (Fontaine, 1979, 1980; New England River Basins Commission, 1978, 1980a). Small northern coastal rivers drain 2,500 mi^2 and include the Royal (143 mi^2), the Presumpscot (641 mi^2), and the Mousam (118 mi^2) Rivers in Maine (Cowing and McNelly, 1978), and the Piscataqua River (1,020 mi^2) in New Hampshire (New England River Basins Commission, 1980b); collectively these rivers form the northern coastal Basins and are grouped with the Saco River Basin in this report (fig. 7). The southern coastal rivers drain 4,243 mi^2 and include the Taunton (530 mi^2), the Charles (321 mi^2), and the Ipswich (155 mi^2) Rivers in Massachusetts, the Blackstone River (480 mi^2) in Massachusetts and Rhode Island, and the Pawcatuck River (303 mi^2) in Rhode Island, as well as many smaller coastal river basins; collectively, these rivers form the southern coastal Basins (fig. 7).

Streamflow Characteristics

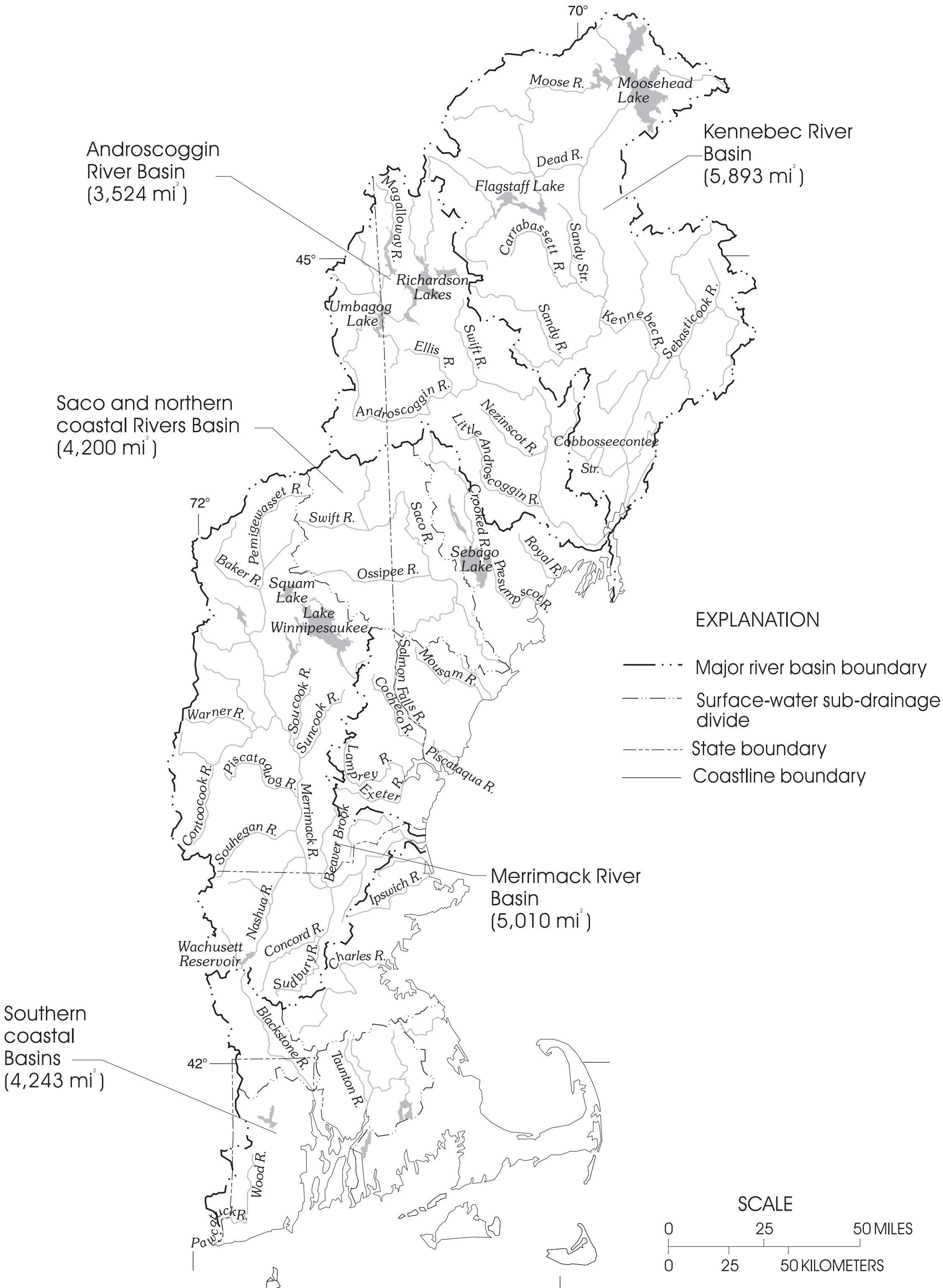
Variations in streamflow were determined on the basis of data collected at USGS streamflow-gaging stations. The USGS currently (1998) operates 90 streamflow-gaging stations in the study area. The gaged sites represent a mix of large and small rivers and their tributaries. Thirty-one of the active stations are in Massachusetts, 23 are in New Hampshire, 20 are in Maine, and 16 are in Rhode Island. Thirty-one representative streamflow-gaging stations are shown in figure 8.

The greatest amount of total streamflow is carried by the large rivers (fig. 8, table 1). Mean annual streamflow for the Kennebec River near Waterville, Maine is 7,628 ft^3/s. Mean annual flows of the other large rivers are: 6,140 ft^3/s in the Androscoggin River near Auburn, Maine; 7,632 ft^3/s in the Merrimack River near Lowell, Mass.; and 934 ft^3/s in the Saco River near Conway, Maine. The largest river in the southern coastal Basins, the Blackstone River, carries a mean annual flow of 774 ft^3/s at Woonsocket, R.I.

The highest flows in all rivers are in April as the result of spring runoff and snowmelt. Fall rains produce a secondary peak in many rivers and streams in the northern part of the study area, as shown by the monthly boxplots of streamflow for the Royal River at Yarmouth, Maine; the Saco River near Conway; N.H., the Merrimack River near Manchester, N.H.; and the Androscoggin River near Auburn, Maine (fig. 9). Low flows for the year are in July, August, and September, when high evapotranspiration rates limit the amount of precipitation that becomes available for runoff. In northern streams, winter precipitation falls as snow and is not converted to streamflow until the spring thaw, resulting in more pronounced spring streamflows as shown on the boxplots for the Royal River, the Saco River, and the Androscoggin River (fig. 9). Flow in rivers in the southern part of the study area, such as the Ipswich River near Ipswich, Mass., the Charles River at Waltham, Mass.; and the Blackstone River near Woonsocket, R.I. is more evenly distributed throughout the colder months because winter precipitation in the southern areas includes more rainfall PFPF (fig. 9). Seasonal variation of streamflow can be reduced by extreme regulation of streamflow, as shown in the boxplot for the Presumpscot River near Westbrook, Maine (fig. 9).

Runoff averages 40 in/yr in the mountainous areas of New Hampshire and averages 20 to 30 in/yr in the rest of the study area (Krug and others, 1990) (fig. 8). Most runoff in the study area is during the spring and early summer. Half of the 24.6 in. of annual runoff in the East Meadow River, a tributary to the Merrimack River in Massachusetts, occurs from March to May; the average annual runoff from August through October is less than 0.5 in/mo (Gay and Delaney, 1980).

Human activity has affected streamflow since the colonists settled New England. Some human activities, such as diverting water for municipal



Base from U.S. Geological Survey
digital data 1:2,000,000, 1972

Figure 7. Generalized hydrography of the New England Coastal Basins study area.

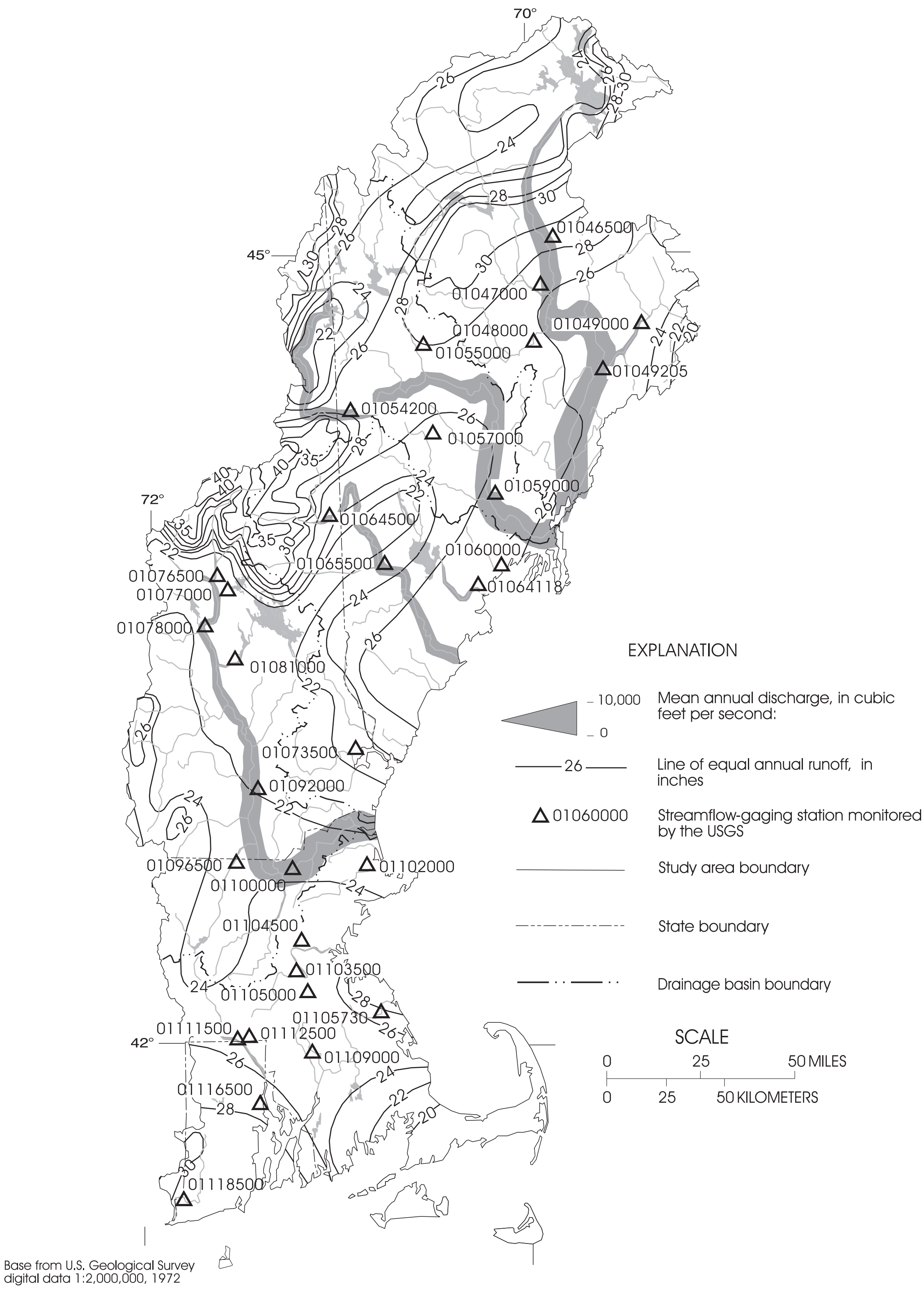


Figure 8. Mean annual discharge, mean annual runoff, and location of selected streamflow-gaging stations in the New England Coastal Basins study area.

Table 1. Streamflow characteristics for selected gaging stations in the New England Coastal Basins study area, in Maine, Massachusetts, New Hampshire, and Rhode Island

[mi^2, square miles; ft^3/s, cubic feet per second; --, no data available; Locations shown on figure 8.]

Gaging station name	Gaging station number	Drainage area, in mi^2	Period of record in water years	Mean annual stream-flow, in ft^3/s	Mean annual runoff, in inches	Flood years			Drought years		
						Annual runoff in inches, by water year					
						1984	1978	1973	1985	1965	1941
Kennebec River at Bingham, Maine	01046500	2,715	1930-96	4,445	22.2	31.6	29.3	28.5	15.7	13.5	15.2
Carrabassett River at North Anson, Maine	01047000	353	1925-96	721	27.7	42.8	39.7	36.1	17.1	14.4	12.8
Sandy River near Mercer, Maine	01048000	516	1928-79 1987-96	971	25.6	--	36.6	35.6	--	13.6	11.5
Sebasticook River near Pittsfield, Maine	01049000	572	1928-96	960	22.8	36.8	31.7	32.5	10.4	14.1	13.3
Kennebec River near Waterville, Maine	01049205	5,179	1994-96	7,628	--	--	--	--	--	--	--
Wild River at Gilead, Maine	01054200	69.6	1964-96	182	35.5	50.5	44.1	50.9	29.2	14.2	--
Androscoggin River near Auburn, Maine	01059000	3,263	1988-96	6,137	25.5	36.8	35.1	34.5	16.1	15.0	14.6
Swift River near Roxbury, Maine	01055000	96.9	1929-96	200	28.0	37.9	37.8	39.2	18.4	15.3	14.4
Little Androscoggin River near South Paris, Maine	01057000	73.5	1913-24 1931-96	138	25.5	38.7	34.3	40.4	14.0	11.6	12.2
Royal River at Yarmouth, Maine	01060000	141	1949-96	274	26.4	46.2	37.8	37.2	13.8	12.7	--
Presumpscot River near Westbrook, Maine	01064118	577	1975-95	925	22.2	40.9	31.6	--	9.5	--	--
Saco River near Conway, N.H.	01064500	385	1929-96	934	32.9	47.0	42.2	51.6	21.3	17.2	20.6
Ossippee River at Cornish, Maine	01065500	452	1916-96	879	26.4	42.8	34.9	39.0	16.7	13.4	14.9
Lamprey River near Newmarket, N.H.	01073500	183	1934-96	282	20.9	32.7	26.1	31.5	10.5	10.1	12.1
Pemigewassett River at Plymouth, N.H.	01076500	622	1903-96	1,360	29.7	42.1	37.5	43.0	20.6	16.0	20.2
Winnepesaukee River at Tilton, N.H.	01081000	471	1937-96	706	20.3	35.4	27.7	32.5	11.9	8.8	13.9
Merrimack River near Goffs Falls below Manchester, N.H.	01092000	3,092	1936-96	5,273	23.1	36.9	29.6	34.4	13.5	9.9	15.3
Nashua River at East Pepperell, Mass.	01096500	435	1935-96	576	[1]18.0	[1]29.6	[1]22.8	[1]26.6	[1]8.8	[1]6.7	[1]9.1
Merrimack River at Lowell, Mass.	01100000	4,635	1923-96	7,632	22.4	36.6	29.7	34.7	13.5	9.0	13.2
Ipswich River near Ipswich, Mass.	01102000	125	1930-96	187	[1]20.3	[1]38.1	[1]24.9	[1]29.1	[1]8.4	[1]7.6	[1]11.5
Charles River at Dover, Mass.	01103500	183	1937-96	305	[1]22.6	[1]36.8	[1]32.9	[1]29.2	[1]10.6	[1]11.0	[1]13.3
Charles River at Waltham, Mass.	01104500	251	1903-09 1931-96	306	[1]16.6	[1]30.2	[1]23.1	[1]20.3	[1]8.8	[1]9.1	[1]8.4
Neponset River at Norwood, Mass.	01105000	34.7	1939-96	55.1	[1]21.6	[1]41.6	[1]30.5	[1]27.2	[1]10.6	[1]11.3	[1]11.1
Indian Head River at Hanover, Mass.	01105730	30.3	1966-96	61.6	27.6	39.8	33.9	37.3	13.4	--	--
Wading River near Norton, Mass.	01109000	43.3	1925-96	73.3	[1]23.0	[1]38.7	[1]36.0	[1]33.5	[1]9.7	[1]11.0	[1]14.4
Branch River at Forestdale, R.I.	01111500	91.2	1912-13 1940-96	175	26.1	38.8	37.5	38.4	15.8	13.8	16.4
Blackstone River near Woonsocket, R.I.	01112500	416	1929-96	774	[1]25.3	[1]37.9	[1]34.6	[1]36.3	[1]15.4	[1]12.9	[1]15.8
Pawcatuck River at Westerly, R.I.	01118500	295	1940-96	579	26.6	39.8	36.8	40.1	17.2	16.6	--
Pawtuxet River at Cranston, R.I.	01116500	200	1939-96	349	[1]23.7	[1]35.0	[1]34.8	[1]40.4	[1]12.7	[1]12.2	[1]17.4

[1] Runoff values significantly affected by water-use activities and are not representative of natural conditions.

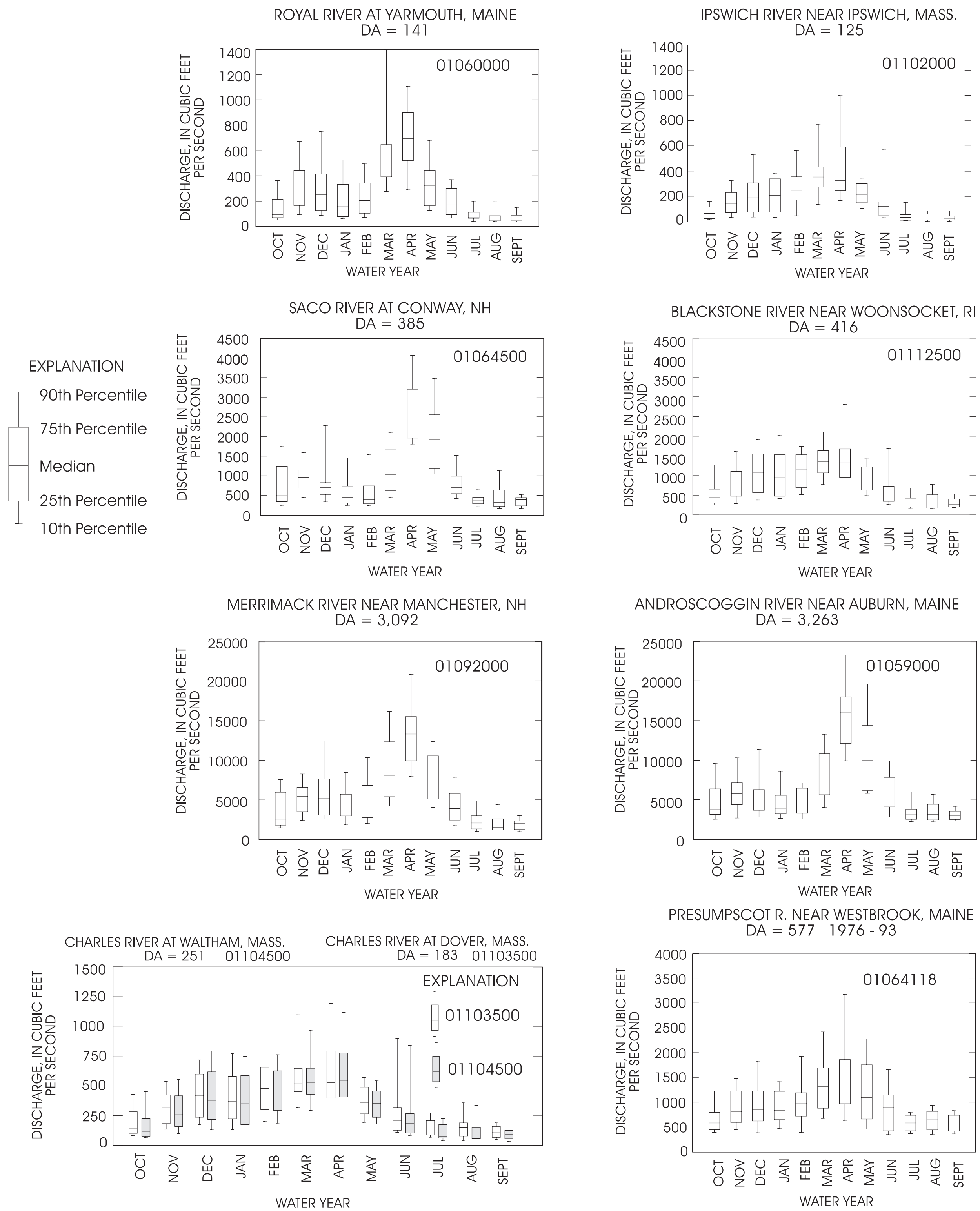


Figure 9. Distribution of monthly streamflows for selected gaging stations, for water years 1973-93 unless otherwise noted, in the New England Coastal Basins study area. Locations of stations shown on figure 8. Eight-digit number in upper corners of plot area are gaging-station numbers.

drinking-water supplies, result in reducing the amount of water flowing in a stream. Use of streamflow for hydroelectric power generation can regulate the timing and magnitude of yearly low and high flows without affecting the total flow, and often creates artificial high and low flows on a daily basis. Flood-control reservoirs diminish the magnitude of yearly high flows. Many rivers and streams in the study area are affected by more than one of these activities. Currently, most of the unregulated rivers are in the White Mountains physiographic section (fig. 2), but a few small streams in the rest of the study area remain unregulated. The largest unregulated rivers include the Sandy River (Kennebec River Basin) and the Swift River (Androscoggin River Basin) (table 1; fig. 7).

The timing and magnitude of yearly low flows (base-flow regulation) and high flows (peak-flow regulation) are regulated in most of the medium-sized rivers and all of the large rivers. Peak and baseflow is regulated by storage reservoirs and hydropower dams. There has been a shift in water-management practices in the large, northern river basins in the study area. From the mid 1800s to mid 1960s, storage was used to increase flows for driving logs downstream, but since the 1960-70’s storage is used for flood control and for providing consistent flows for power generation for industrial users and commercial electricity providers. Thus, large peaks in flow are reduced, as water is held back in the spring, and normal summer to fall low flows are increased as this stored water is released throughout the year.

There were more than 1,600 dams throughout the study area in 1996 (U.S. Army Corps of Engineers, 1996). The Androscoggin River Basin had the smallest number of dams (91), whereas the southern coastal Basins had the largest number of dams (697) (table 2). Almost 36 percent of all dams were used for recreation, 18.8 percent for water supply, 17.5 percent for hydroelectric power generation, and 7.8 percent for irrigation. The rest of the dams were used for flood control (7.3 percent), as fire ponds (1.2 percent), or other uses (10.3 percent) (table 2). Location of dams from the top three categories are shown in figure 10. In general, most of the hydroelectric power dams are in the northern part of the study area, and recreational and water-supply dams are primarily in the southern and central parts (fig. 10).

The effects of reservoirs, impoundments, and lakes on streamflows can be seen in data collected during April 1987, when widespread flooding caused record or near-record flows, including those in the Saco River near Conway, N.H. and the Pemigewassett River at Plymouth, N.H. (fig. 11). While streamflow in these and other unregulated basins increased and decreased swiftly during this flood, flows in nearby rivers of similar sizes that had significant storage (the Ossipee River at Cornish, Maine and the Winnipesaukee River at Tilton, N.H.) were relatively unaffected by the flood (fig. 11).

Large and small hydroelectric power dams also produce large shifts in streamflows on a daily basis as water is released to provide power during peak electricity-use hours and held back later in the day during low electricity use (fig. 12). The degree of daily-streamflow regulation at a location depends on its proximity to upstream hydropower dams and whether those dams are managed in such a way as to cause large shifts in flow throughout the day. Daily-streamflow regulation generally has a negative impact on aquatic habitats.

Table 2. Summary of dams, by major river basin, in the New England Coastal Basins study area, in 1995-96

[Data from U.S. Army Corps of Engineers

River basin	Number of dams by major category or purpose								
	Water supply	Fish and wildlife	Flood control	Hydro-electric	Recreation	Fire pond	Irrigation	Other	Total
Kennebec	3	5	5	54	12	9	0	9	97
Androscoggin	9	3	7	53	11	1	0	7	91
Saco and northern coastal	18	7	7	73	71	6	0	23	205
Merrimack	100	1	53	93	232	1	5	52	537
Southern coastal	176	6	48	12	255	2	122	76	697
Study area total	306	22	120	285	581	19	127	167	1,627

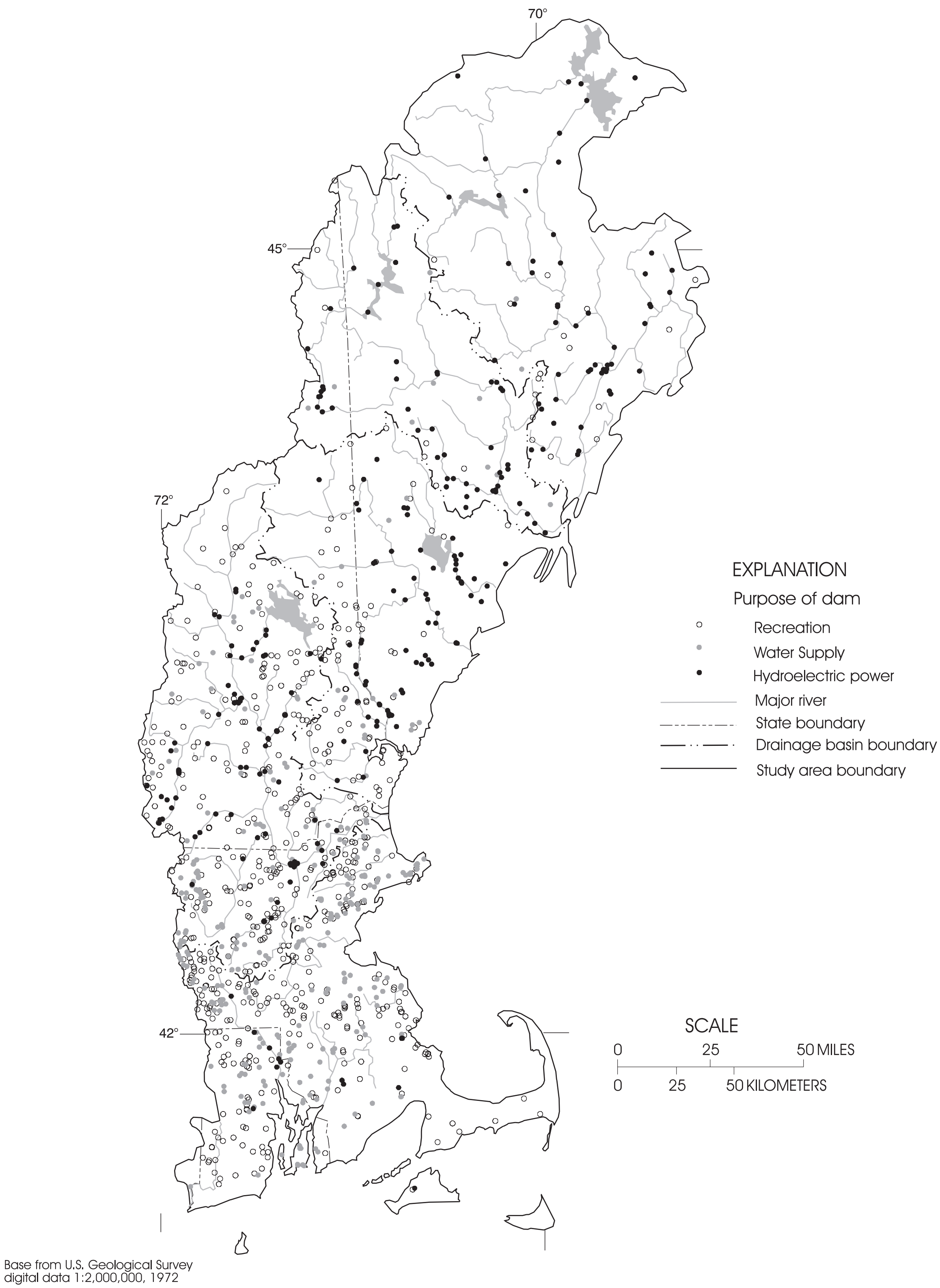


Figure 10. Location of dams used for recreation, water supply, and hydroelectric power generation in the study area. Data from the U.S. Army Corps of Engineers.

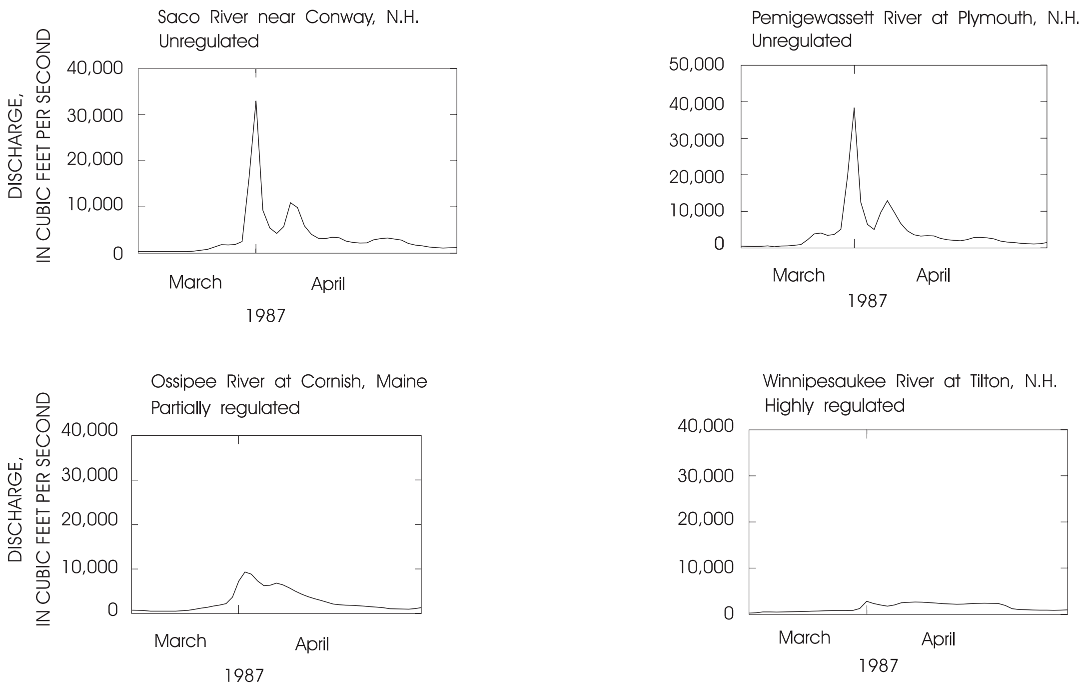


Figure 11. Selected storm hydrographs from unregulated and regulated streams and lakes for a large runoff event (March-April 1987) in the New England Coastal Basins study area in Maine, Massachusetts, Rhode Island, and New Hampshire, 1987.

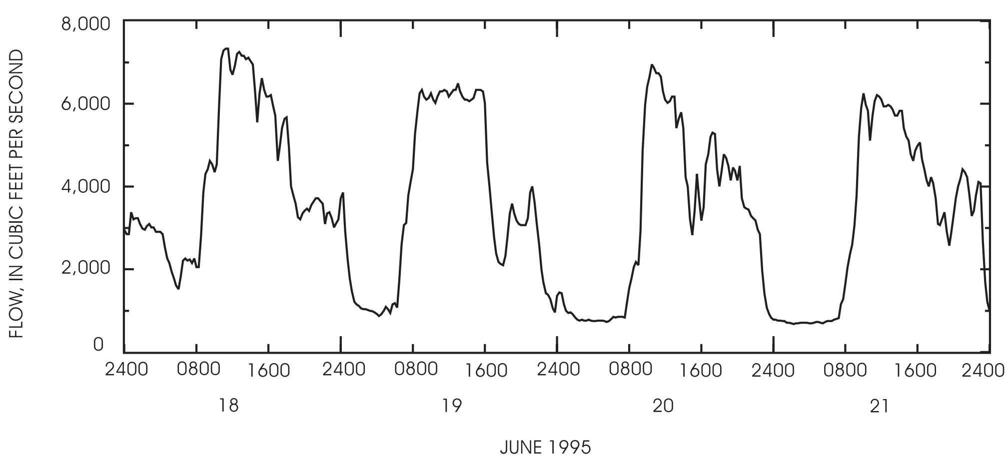


Figure 12. Daily streamflow regulation on the Kennebec River at Bingham, Maine, June 18-21, 1995.

Urbanization, which creates increased demands on water resources, has resulted in base-flow depletion in a number of rivers in eastern Massachusetts and Rhode Island. Two examples of base-flow depletion in eastern Massachusetts are in the Charles River Basin and in the Ipswich River Basin. In the Charles River Basin, withdrawals and diversions for drinking water have reduced the average runoff in the basin above Waltham, Mass., to just 16.6 in/yr (table 1). The nearby Indian Head River basin, which has fewer demands on the surface-water system, has an average annual runoff of 27.6 in/yr at Harvard, Mass. A comparison of the monthly flows in the Charles River upstream (at Dover drainage area) and downstream (at Waltham drainage area) (fig. 9) shows that the flow in the river does not increase downstream as expected because of these diversions.

A combination of surface-water withdrawals and ground-water pumping lowered baseflow in the Ipswich River Basin. Towns in the headwaters of the basin rely on wells for drinking-water supplies. The wells are completed in coarse-grained sand and gravel aquifers that are in direct hydraulic connection with the river. During the summer of 1996, 15 to 20 mi of the river ran dry because ground-water pumping had induced so much water from the river into the aquifer (D. LeVangie, Massachusetts Department of Environmental Management, Water Supply Division, oral commun., 1996). Farther downstream in the Ipswich River basin, towns rely on surface water for their drinking water. From November to May, water is diverted out of the river into reservoirs for drinking water. Also, none of the wastewater generated in the basin is returned to the Ipswich River.

Floods and Droughts

Although annual peak flows generally are in the spring, major flooding can occur at any time during the year. Typically, spring floods are caused by intense rainfall and by the melting of snow. The flood of spring 1936 was caused by intense rains and unseasonably warm weather that melted the snowpack. The severity of the flood was increased by ice jam break-ups in the rivers and a surge in runoff because the soil was frozen and impermeable. This flood resulted in a peak discharge of 161,000 ft^3/s for the Merrimack River near Lowell, Mass. (Station 01100000, fig. 8; Gadoury and others, 1994). Floods in the summer and fall are caused primarily by thunderstorms or hurricanes. The “Great Hurricane of 1938” resulted in the worst natural disaster in New England (Paulsen and others, 1940). The flood was caused by heavy rainfall followed by a hurricane and an ocean storm wave. Another major flood occurred in April 1987, primarily across the northern part of the study area. This flood caused the peak flow record of 186,000 ft^3/s for the Kennebec River at Sidney, Maine (Nielsen and others, 1994). Years with the highest amounts of total runoff (1984, 1978, and 1973) had approximately 25 to 35 percent higher runoff than the mean annual runoff for most basins (table 1).

Droughts are more difficult to define than floods because they have no distinct beginning or end. Yet they can be quantified by examining the frequency of less-than-average precipitation. A major prolonged drought in 1961-69 resulted in the drought of record in New England. This drought had a serious effect on agriculture and water supply in the region. Ground-water storage, the primary source of streamflow between rainfall events (Barksdale and others, 1966), was also severely depleted. Ordinarily in New England, ground-water storage is replenished during each non-growing season. This depletion was observed over the winter of 1964-65 and resulted in record-low streamflows during 1965, the driest year ever recorded in the study area. Other periods of drought were from 1939-44, 1947-50 (Hammond, 1991; Maloney and Bartlett, 1991; and Wandle, 1991), and from 1980-81 (Walker and Lautzenheiser, 1991). In the northern part of the study area, years with the lowest total runoff (1985, 1965, and 1941) have 50 to 65 percent of the average annual runoff (table 1). This compares to only 40 to 55 percent of the average annual runoff in the southern part of the study area, where diversions of water out of the basins leaves less of the total water budget for instream flow.

Lakes, Reservoirs, and Wetlands

There are hundreds of lakes, ponds, reservoirs, and wetland areas of various sizes throughout the study area (table 3, Anderson and others, 1976). The two largest natural lakes are Moosehead Lake (124 mi^2) in Maine and Lake Winnipesaukee (78 mi^2) in New Hampshire (fig. 7). The greater the number and total area of lakes, reservoirs, and wetlands in a basin, the greater the natural potential for storage of runoff during storms (Benson, 1962). Additional storage potential may be provided by flood-control

Table 3. Numbers and total areas of lakes and wetlands greater than 0.1 square miles in the major river basins of the New England Coastal Basins study area in Maine, Massachusetts, New Hampshire, and Rhode Island

[mi^2, square miles; data from Geographic Information and Retrieval System land cover data base (Anderson and others, 1976)]

River basin	Lakes (and reservoirs)			Wetlands	
	Number of lakes greater than 0.1 mi^2	Total area of lakes, in mi^2	Percent of basin area in lakes	Total area of wetlands, in mi^2	Percent of basin area in wetlands
Kennebec	221	335	5.6	241	4.1
Androscoggin	97	133	3.8	68	1.9
Saco and northern coastal	156	137	3.3	109	2.6
Merrimack	251	174	3.5	55	1.1
Southern coastal	269	95	2.3	151	3.6
Study area total	994	874	3.8	624	2.7

reservoirs and other reservoirs that increase the natural potential by drawing water down below the uncontrolled outlet. Based on the area of lakes, reservoirs, and wetlands, the Kennebec River Basin has almost twice the amount of natural storage potential as the other basins. The Southern coastal Basins have the smallest area of lakes and, therefore, the smallest amount of natural potential for storage of surface-water runoff during storms (table 3).

Lakes support extensive recreational activities. Lake Winnipesaukee in New Hampshire (fig. 7) and Sebago Lake in Maine (fig. 7) have evolved over the past century into popular recreational centers that support a large seasonal population and contain extensive shoreline residential developments. Moosehead Lake (fig. 7), and other large, remote lakes in Maine and New Hampshire are still relatively undeveloped. Recently, the U.S. Fish and Wildlife Service (1991) established the Lake Umbagog National Wildlife Refuge in the headwater region of the Androscoggin River Basin in Coos County, N.H. and Oxford County, Maine. The Refuge was established primarily to protect the pristine lake from future shoreline development and to protect endangered wildlife and plant species and their habitats.

Ground Water

Ground water is an important source of water for streams and lakes and is used for domestic, public, commercial, and industrial water supplies. It is available in highly variable quantities in all the geologic units of the study area (table 4).

Aquifers

There are three main types of aquifers in the study area—stratified drift, till, and bedrock. The highly permeable, relatively shallow (less than 100 ft), discontinuous stratified-drift aquifers that occupy most river valleys in New England are the principal source of drinking water for many communities that use ground water. Glacial-till aquifers are characterized by low permeability but can supply sufficient water to shallow wells for households and farm use. Fractured-bedrock aquifers are the primary source of drinking water to rural households and are an important source of water to a few public-supplied, commercial and industrial users.

A digital compilation of the sand and gravel resources for the States of Maine, New Hampshire, Massachusetts, and Rhode Island (Marvinney and Walters, 1993; Koteff, 1993; and Stone and Beinikis, 1993; Cain and Hamidzada, 1993) was used to determine the extent of stratified-drift deposits (fig. 5b). Stratified-drift deposits cover 21 percent of the study area and range from 3.7 percent of the Kennebec River Basin to 53 percent of the southern coastal Basins (fig. 5b).

Stratified-drift aquifers consist mainly of sand and gravel deposited in layers by meltwater streams flowing from the retreating glacial ice. The areal distribution, thickness, and hydraulic properties of stratified-drift aquifers are directly related to their mode of deposition. Most stratified-drift aquifers in the study area formed in a glacial-lake environment in inland and upland areas, in a marine environment near coastal areas north of Boston, Mass., and as large outwash plains in the Cape Cod region. Stratified-drift aquifers that contain significant amounts of saturated, coarse-grained ice-contact and outwash deposits can yield high quantities of water (table 4).

Table 4. Geologic units, hydraulic properties, and general water-bearing characteristics in the New England Coastal Basins study area, in Maine, Massachusetts, New Hampshire, and Rhode Island

[ft/d, foot per day; gal/min, gallons per minute; ft^2/d, feet squared per day]

Geologic unit and occurrence	Range in thickness, in feet	Hydraulic properties	General water-bearing characteristics
Outwash deposits--Stratified deposits of sand and gravel in outwash plains and valley trains. Below 400 feet altitude in Maine and New Hampshire, outwash can overlie marine or lacustrine deposits.	0 - 200	Hydraulic conductivity (k) ranges from 35 - 1,000 ft/d for outwash and ice-contact deposits. A detailed study of stratified-drift aquifers in New Hampshire showed that the calculated hydraulic conductivity for well sorted, very-fine to fine sand is about 12 ft/d; well sorted medium sand is about 51 ft/d; well sorted coarse to very-coarse sand is about 970 ft/d; and granules (or gravel) can exceed 1,000 ft/d.[1] Specific yield is approximately 0.2. Porosity of coarse gravel ranges from 24 to 36 percent.	Yields small to moderate amounts of water to dug, drilled, or driven wells and to springs. Largest yields are from wells in areas where the saturated thickness is large, the deposits are coarse grained, and in hydraulic contact with a nearby surface-water body as a source of induced recharge.
Ice-contact deposits--Well to poorly sorted, stratified deposits of sand, gravel, and cobbles, with some silt and boulders. Land forms include eskers, kames, kame deltas, and kame terraces. Not an extensive deposit in the study area.	0 - 150		These deposits are the best potential source of large supplies of ground water in the study area, especially in areas where the saturated thickness is large, the deposits are coarse grained, well sorted, and in hydraulic contact with a nearby surface-water body as a source of induced recharge. Under the most favorable conditions, from 500 to more than 1,500 gal/min of water can be obtained from individual, gravel-packed wells screened in these deposits.
Glacial-lake (lacustrine) deposits--Deposits consist of blue-gray silt, clay, and fine sand; may contain lenses of medium sand. These deposits are similar in composition to marine deposits (see fig. 5a for general locations).	0 - 280	Hydraulic conductivities of silt and clay are generally less than 1 ft/d; very fine sand is about 4 ft/d. Hydraulic conductivity is as low as 2.3×10^{-7} ft/d in marine clays.[2] Specific yield is negligible. Porosity of fine sand ranges from 24 to 36 percent. In general, porosity increases and hydraulic conductivity decreases with decreasing particle size for glacial sediments.	These deposits are too fine grained to yield significant water to wells, but can contain lenses of sand and gravel from which wells of moderate yield could be developed. Glacial-lake deposits, where found, act as a confining unit to more permeable deposits buried beneath or overlying them.
Marine clays deposits--Dark-blue to gray silt, clay and fine sand; tan where weathered. Contain layers of sand and gravel. Underlie outwash deposits and may crop out in stream valleys up to about 400 feet above sea level (see fig. 5a for general locations).	0 - 100		Marine clays have low permeability and release water slowly. Marine clays are not a significant aquifer and can act as a confining unit to more permeable deposits buried beneath or overlying them.
End moraine deposits--Composed of complex, glacially deformed sediments of sand, gravel, silt, clay, and till which locally overlie sand and gravel deposits at the heads of large promorainal outwash plains.[3] These deposits are in the southernmost part of the study area (see fig. 5a).	0 - 1,000	A study of morainal aquifers on Block Island, R.I., reported hydraulic conductivity values ranging from 3 - 2,100 ft/d, with a median of 27 ft/d.[4]	Yields depend greatly on the lithologic unit within the moraine deposit that a well is screened in. Transmissivity from 114 wells screened in differing morainal units on Block Island, R.I., ranged from 15 - 17,5000 ft^2/d, with a median of only 200 ft^2/d.[4] Wells are commonly used to supply domestic drinking water.

Table 4. Geologic units, hydraulic properties, and general water-bearing characteristics in the New England Coastal Basins study area, in Maine, Massachusetts, New Hampshire, and Rhode Island—Continued

[ft/d, foot per day; gal/min, gallons per minute; ft^2/d, feet squared per day]

Geologic unit and occurrence	Range in thickness, in feet	Hydraulic properties	General water-bearing characteristics
Till deposits--Till, the most extensive glacial deposit in the study area, is an unsorted, unstratified mixture of clay, silt, sand, gravel, angular cobbles, and boulders. Below the first few feet, particularly where thick, till is commonly clay-rich and very dense. Till covers the bedrock in upland areas in varying thicknesses and commonly underlies stratified-drift deposits in valleys and other lowland areas.	0 - 325	A study of till in southern New England showed that the horizontal hydraulic conductivity of tills derived from metamorphic and igneous (crystalline) rocks ranged from 0.004 - 65 ft/d.[5] The porosities and specific yields of these tills ranged from 22.1 to 40.6 percent and from 3.9 to 31.2 percent, respectively.[5]	Till can be a source of water to dug wells. Till transmits water slowly and the yield of dug wells is small once water stored in the well casing is pumped out. Dug wells in till are likely to go dry during periods of little or no precipitation. Yields of wells in till typically range from 1 to 5 gal/min.
Crystalline Bedrock--Bedrock formations consist of a variety of igneous and metamorphic rocks. Igneous rocks include granite, pegmatite, and granodiorite with smaller amounts of basic volcanic or intrusive rocks. Metamorphic rocks consist largely of metamorphosed sedimentary rocks and include schist, gneiss, phyllite, quartzite, and slate. Bedrock formations outcrop on hills, ridges, and steep valley walls.	Average depth of drilled wells is 309 feet, with greater than 75 percent of wells less than 400 feet in depth.[6]	Single-hole hydraulic testing done by Hsieh and others (1993) indicated that the hydraulic conductivity in fractured crystalline bedrock range from $2.8\text{x}10^{-4}$ to 2.8 ft/d. Cross-sectional models over a scale of several miles in fractured crystalline bedrock of moderate to high relief in the White Mountain section of the New England physiographic province indicate that the hydraulic conductivity is less than 0.086 ft/d.[7] Numeric modelling of ground-water flow near Mirror Lake, N.H., indicates that hydraulic conductivities are about 0.09 ft/d for bedrock aquifers beneath upper hillsides and hilltops.[8] Many crystalline rocks have a high number of fractures but few are connected. Therefore, the effective porosity, which is defined as the percentage of interconnected pore space, is generally much less than the total porosity of crystalline rocks and ranges from $5.0\text{x}10^{-5}$ to 0.01 percent.[2]	Bedrock formations are dense, relatively impermeable, and have low porosity. They contain recoverable water in secondary openings such as joints, fractures, and bedding or cleavage planes. The water in fractured bedrock aquifers is generally confined. Yields in bedrock wells depend on the number, size, and degree of interconnection of water-bearing fractures. The average yield of inventoried drilled wells (over 13,700) is 14.3 gal/min.[6] High-yielding wells in fractured bedrock are relatively rare; only 2 percent have yields greater than 100 gal/min. Yields of wells that tap bedrock aquifers in Rhode Island commonly range from 1 to 20 gal/min.[9] In Massachusetts, sedimentary-rock aquifers have higher median yields (12 gal/min) than crystalline-rock aquifers (6 gal/min) and commercial or industrial wells have median yields of 30 gal/min.[10]

[1]Moore and others, 1994.
[2]Domenico and Schwartz, 1990.
[3]Stone and Sirkin, *in* Veeger and Johnston, 1996.
[4]Veeger and Johnston, 1996.
[5]Melvin and others, 1992.
[6]Data from the U.S. Geological Survey's Ground-Water Site Inventory data base.
[7]Harte, 1992.
[8]Tiedeman and others, 1997.
[9]Johnston, 1985, p. 374.
[10]Hansen and Simcox, 1994.

The largest sole-source aquifer in the study area is the Cape Cod glacial aquifer, which covers 440 mi^2 in southeastern Massachusetts (fig. 5b). The aquifer is composed of extensive outwash deposits overlying and interbedded with layers of lacustrine clay, silt, very fine sand and till. Combined, the glacial deposits range in thickness from 100 ft at the western end of the peninsula near the Cape Cod Canal to about 1,000 ft at the northern end near the town of Truro (Leblanc and others, 1986). The water table of the aquifer is dominated by six hydraulically independent flow cells or mounds. Ground water flows radially from the center of the flow cells towards the Atlantic Ocean or ocean inlets. Analysis of the aquifer system shows that approximately 270 million gallons of water flows through the six cells on a daily basis (Olcott, 1995). In 1985, the aquifer provided water to 128 municipal wells and more than 20,000 private wells (Persky, 1986).

The USGS and the State of New Hampshire conducted a cooperative program from 1983 to 1995 to assess the State's stratified-drift aquifers, which included a description of the areal extent, geohydrology, potential yield, and quality of water in these aquifers (Medalie and Moore, 1995). The largest stratified-drift aquifer in the State is the Ossipee Lake aquifer in the upper Saco River Basin in east-central New Hampshire (Moore and Medalie, 1995) and west-central Maine (Tepper and others, 1990). This aquifer is an example of a valley-fill system formed in a glacial-lake environment. Stratified-drift landforms in the Ossipee Lake aquifer include eskers, kames, outwash, fine-grained lacustrine, and alluvial deposits.

The USGS, in cooperation with the Maine Department of Conservation, Maine Geological Survey, and the Maine Department of Environmental Protection, has been mapping Maine's "significant" sand and gravel aquifers since 1981. One well-studied aquifer, the Little Androscoggin aquifer in southeastern Oxford County, Maine, is an example of a valley-fill aquifer system that formed in a lowland bedrock valley near the coast and was either partly or completely inundated by the ocean during the last glaciation (Morrissey, 1983). The surficial deposits consist of highly permeable sand and gravel that were deposited in contact with glacial ice at the northern part of the valley and outwash sand deposited in front of the ice that overlies a thick layer of impermeable marine silt, clay, and fine sand at the southern part of the valley.

The Rhode Island Water Resources Board has identified 21 high-yielding stratified-drift aquifers, termed ground-water reservoirs, with transmissivity equal to or exceeding 4,000 ft^2/day and saturated thicknesses equal to or exceeding 40 ft (Trench, 1991). These ground-water reservoirs represent only a small fraction of the total area underlain by stratified drift in Rhode Island. High-yielding public-supply wells tap many of these ground-water reservoirs. A typical public-supply well may yield 700 gal/min in the deeply saturated, permeable areas (Johnston, 1985).

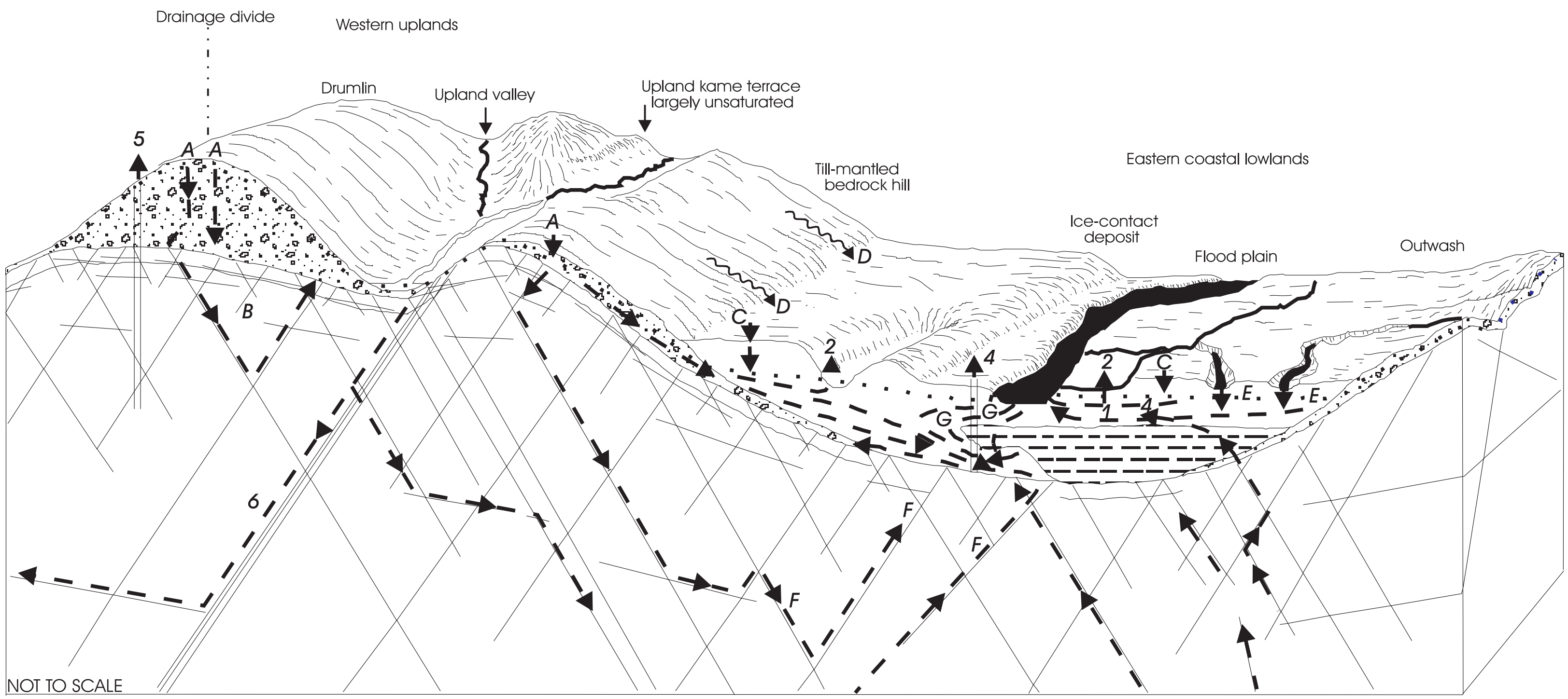
Fractured-bedrock aquifers primarily store and transmit water through intersecting fractures (table 4; fig. 13). These fractures were formed as a result of cooling stresses in magma, tectonic activity, erosion of overlying rock, and the freeze-thaw activities of glacial ice sheets that once covered New England (Hansen and Simcox, 1994). Brittle and coarser-grained rocks, such as granite and basalt, generally have wider and more continuous fractures than those of finer-grained rocks such as schist and gneiss. Most bedrock ground-water supply wells in the study area are less than 700 ft deep. Generalized hydraulic properties and aquifer characteristics of fractured-bedrock aquifers are summarized in table 4.

Recharge, Discharge, and Ground-Water Levels

Ground water is recharged by precipitation that infiltrates the land surface and percolates through the soil layer to the water table, which is the upper surface of the saturated layer (or saturated zone). The water contained in the saturated zone moves in the direction of decreasing head from recharge areas to areas of ground-water discharge. This usually corresponds to areas of topographic uplands to areas of topographic lowlands (fig. 13). The rate of ground-water flow is dependent on the hydraulic conductivity of aquifer materials and the hydraulic gradient.

Most of the recharge in New England is in late autumn and early spring, when precipitation is greatest and evapotranspiration is lowest. Vegetation, soil permeability, climatic conditions, and impervious urban areas can affect the rate of recharge to aquifers.

Average annual recharge to stratified-drift aquifers from precipitation is approximately half of the annual precipitation. This equals about 20 to 24 in/yr in glaciated areas of eastern Massachusetts (Knott and Olimpio, 1986; Leblanc and others, 1986; Bent, 1995; and Hansen and Lapham, 1992), in



EXPLANATION

RECHARGE TO BEDROCK

A Infiltration of precipitation through till in uplands

B Inflow to the basin from circulation through bedrock

RECHARGE TO STRATIFIED DRIFT

C Precipitation on valley floor, which infiltrates to water table, unless diverted as evapotranspiration or storm runoff from pavement or saturated soil

D Runoff from adjacent hillsides at shallow depth Through sandy till, through soil horizons and (or) as surface rivulets

E Natural seepage losses from small tributaries not incised to the water table

F Lateral and upward flow from bedrock

G Induced infiltration from rivers near large-capacity wells where the water table is lowered by pumping

DISCHARGE FROM STRATIFIED DRIFT

1 Seepage to river

2 Ground-water evapotranspiration where the water table is shallow

3 Underflow downvalley through stratified drift (not shown)

4 Pumpage from well screened in stratified drift

DISCHARGE FROM BEDROCK

5 Pumpage from well that intersects fractures

LITHOLOGIC UNIT

Till

River and streams

Stratified drift

Fractured bedrock

Silt and clay

Water table

Direction of ground-water flow

Fault

Figure 13. Idealized geohydrologic section in the glaciated Northeast (modified from Randall and others, 1988).

southern Maine (Morrissey, 1983), in east-central New Hampshire (Tepper and others, 1990), and in southeastern New Hampshire (Stekl and Flanagan, 1992; Mack and Lawlor, 1992; Harte and Mack, 1992). In southern Rhode Island, the recharge rate is estimated to be 25 to 31 in/yr (Dickerman and Ozbilgin, 1985; Dickerman and others, 1990; Dickerman and Bell, 1993; Barlow, 1997), 14 in/yr in south-central New Hampshire (Harte and Johnson, 1995), and about 9 in/yr in areas underlain by till (Trench, 1991).

Runoff from till uplands (fig. 13) also provides large amounts of recharge to adjacent stratified-drift aquifers. A study of seepage losses to surface water along a 4-mi reach of the Saco River near North Conway, N.H., in a region of high topographic relief, indicates that adjacent upland areas account for nearly 60 percent of the recharge to stratified drift in this valley (Morrissey and others, 1988).

Factors that affect recharge to crystalline bedrock aquifers in a mountainous setting are described by Harte and Winter (1996). Bedrock recharge is controlled by relief of land and bedrock surface above ground-water sinks (such as lakes) and glacial-drift stratigraphy. Recharge to crystalline bedrock aquifers is estimated to be about 3 to 5 in/yr (Harte and Winter, 1995).

Recharge to an aquifer from surface-water bodies is induced when withdrawal wells reverse natural flow directions and induce flow into the aquifer (fig. 13). Induced infiltration is an important source of water for many high-yielding municipal wells in the study area, especially for those wells within a few hundred feet of lakes, rivers, and wetlands.

Artificial recharge of aquifers is a minor component of total recharge, but does occur in areas where farmland is irrigated, from municipal leach fields in urban areas, and from domestic leach fields in rural areas. Tepper and others (1990) reported that approximately 80 percent of the water pumped from municipal wells in the Saco River Valley aquifer between Bartlett, N.H., and Fryeburg, Maine, is returned to the aquifer through septic systems.

Ground water discharges from aquifers through seepage into streams, lakes, and wetlands; evapotranspiration; and withdrawal from wells (fig. 13). Ground-water discharge, primarily from stratified-drift aquifers, sustains streamflow during dry periods, usually during late summer or early autumn and during droughts. The rate of discharge per square mile of drainage area from coarse-grained stratified drift is greater than that from till (Wandle and Randall, 1994). Ground-water evapotranspiration is another source of discharge from aquifers and is greatest during the April to October growing season (Stekl and Flanagan, 1992). Ground-water evapotranspiration has been estimated to range from 1 to 9 in/yr in the northeastern United States (Lyford and others, 1984).

Ground-water-level monitoring shows that water-level fluctuations are usually greater in till and bedrock uplands than in stratified-drift deposits because of differences in their specific yields, which are a function of lithology (fig. 14). Median water levels for six wells show recharge occurring from approximately January through April as rainfall and snowmelt infiltrate the ground (fig. 14). Little recharge occurs during the summer growing season, however, and ground-water levels decline from April through October. Recharge begins again in the late fall after the growing season, continues into December, and ends when the ground freezes. After the ground thaws, the annual recharge cycle begins again.

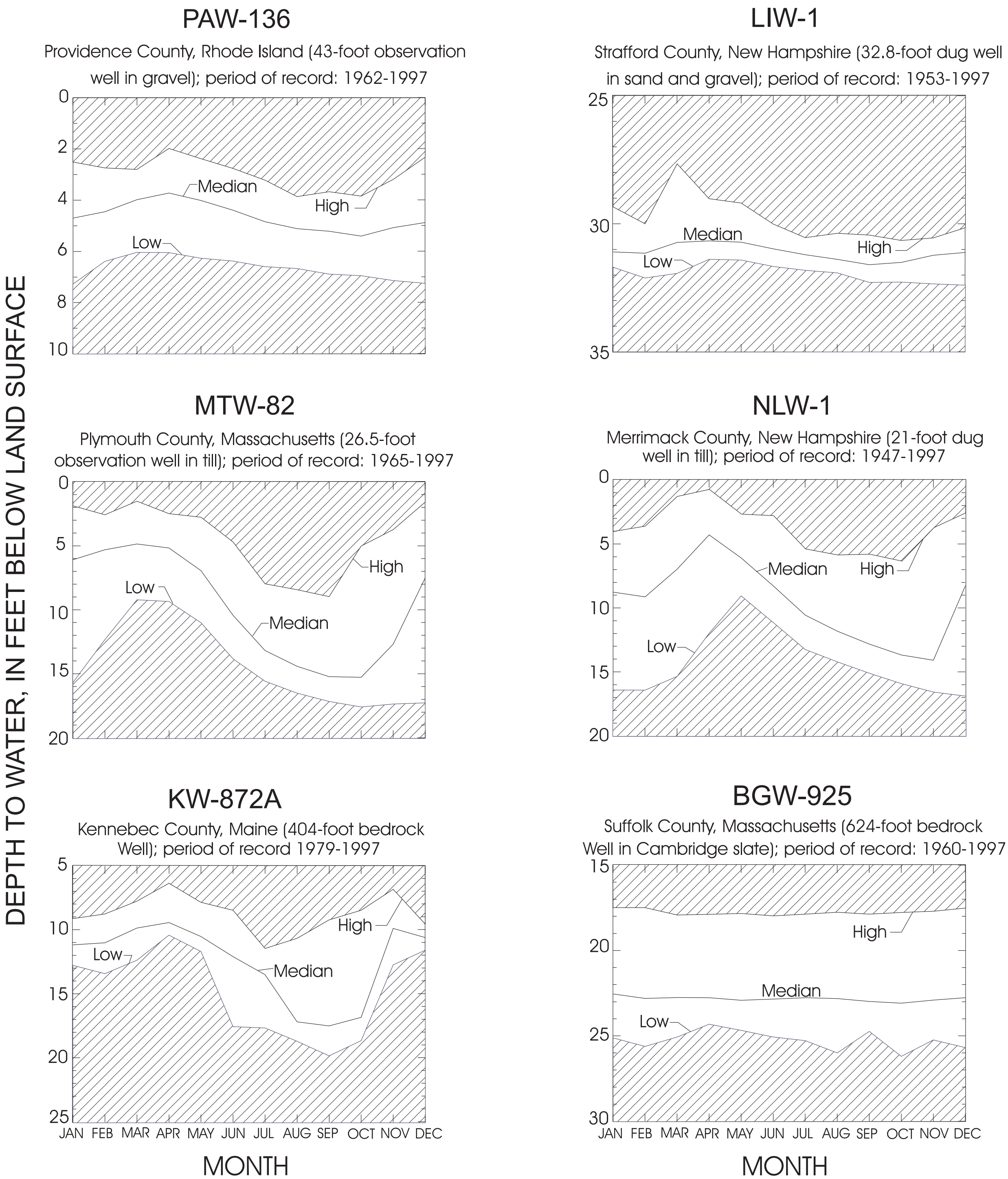


Figure 14. Comparison of monthly median and ranges of water levels in selected observation wells during the 1994 water year in the New England Coastal Basins study area. Unshaded area shows the range between highest and lowest monthly water level for periods of record. (Locations of wells shown on figure 5b).