In cooperation with the Connecticut Department of Public Health

Simulation of Ground-Water Flow to Assess Geohydrologic Factors and their Effect on Source-Water Areas for Bedrock Wells in Connecticut

Scientific Investigations Report 2004-5132

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U.S. Department of the Interior
U.S. Geological Survey

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Conversion Factors and Datum

Multiply	By	To obtain
	Length	
foot (ft)	0.3048	meter (m)
mile (mi)	1.609	kilometer (km)
	Area	
square mile (mi^2)	2.590	square kilometer (km^2)
	Flow rate	
inch per year (in/yr)	2.54	centimeter per year (cm/yr)
cubic foot per day (ft^3/d)	0.02832	cubic meter per day (m^3/d)
gallon per minute (gal/min)	0.06309	liter per second (L/s)
	Hydraulic conductivity	
foot per day (ft/d)	0.3048	meter per day (m/d)
	Transmissivity*	
foot squared per day (ft^2/d)	0.09290	meter squared per day (m^2/d)

Temperature in degrees Celsius (°C) may be converted to degrees Fahrenheit (°F) as follows:

°F = (1.8 x °C) + 32

Temperature in degrees Fahrenheit (°F) may be converted to degrees Celsius (°C) as follows:

°C = (°F - 32) / 1.8

Vertical coordinate information is referenced to the *North American Vertical Datum of 1988 (NAVD 88).*

Horizontal coordinate information is referenced to the *North American Datum of 1983 (NAD 83).*

Altitude, as used in this report, refers to distance above the vertical datum.

*Transmissivity: The standard unit for transmissivity is cubic foot per day per square foot times foot of aquifer thickness [(ft^3/d)/ft^2]ft. In this report, the mathematically reduced form, foot squared per day (ft^2/d), is used for convenience.

Specific conductance is given in microsiemens per centimeter at 25 degrees Celsius (µS/cm at 25°C).

Concentrations of chemical constituents in water are given either in milligrams per liter (mg/L) or micrograms per liter (µg/L).

Simulation of Ground-Water Flow to Assess Geohydrologic Factors and their Effect on Source-Water Areas for Bedrock Wells in Connecticut

By J. Jeffrey Starn and Janet Radway Stone

Abstract

Generic ground-water-flow simulation models show that geohydrologic factors—fracture types, fracture geometry, and surficial materials—affect the size, shape, and location of source-water areas for bedrock wells. In this study, conducted by the U.S. Geological Survey in cooperation with the Connecticut Department of Public Health, ground-water flow was simulated to bedrock wells in three settings—on hilltops and hillsides with no surficial aquifer, in a narrow valley with a surficial aquifer, and in a broad valley with a surficial aquifer—to show how different combinations of geohydrologic factors in different topographic settings affect the dimensions and locations of source-water areas in Connecticut.

Three principal types of fractures are present in bedrock in Connecticut—(1) Layer-parallel fractures, which developed as partings along bedding in sedimentary rock and compositional layering or foliation in metamorphic rock (dips of these fractures can be gentle or steep); (2) unroofing joints, which developed as strain-release fractures parallel to the land surface as overlying rock was removed by erosion through geologic time; and (3) cross fractures and joints, which developed as a result of tectonically generated stresses that produced typically near-vertical or steeply dipping fractures.

Fracture geometry is defined primarily by the presence or absence of layering in the rock unit, and, if layered, by the angle of dip in the layering. Where layered rocks dip steeply, layer-parallel fracturing generally is dominant; unroofing joints also are typically well developed. Where layered rocks dip gently, layer-parallel fracturing also is dominant, and connections among these fractures are provided only by the cross fractures. In gently dipping rocks, unroofing joints generally do not form as a separate fracture set; instead, strain release from unroofing has occurred along gently dipping layer-parallel fractures, enhancing their aperture. In nonlayered and variably layered rocks, layer-parallel fracturing is absent or poorly developed; fracturing is dominated by well-developed subhorizontal unroofing joints and steeply dipping, tectonically generated fractures and (or) cooling joints. Cross fractures (or cooling joints) in nonlayered and variably layered rocks have more random orientations than in layered rocks. Overall, nonlayered or variably layered rocks do not have a strongly developed fracture direction.

Generic ground-water-flow simulation models showed that fracture geometry and other geohydrologic factors affect the dimensions and locations of source-water areas for bedrock wells. In general, source-water areas to wells reflect the direction of ground-water flow, which mimics the land-surface topography. Source-water areas to wells in a hilltop setting were not affected greatly by simulated fracture zones, except for an extensive vertical fracture zone. Source-water areas to wells in a hillside setting were not affected greatly by simulated fracture zones, except for the combination of a subhorizontal fracture zone and low bedrock vertical hydraulic conductivity, as might be the case where an extensive subhorizontal fracture zone is not connected or is poorly connected to the surface through vertical fractures.

Source-water areas to wells in a narrow valley setting reflect complex ground-water-flow paths. The typical flow path originates in the uplands and passes through either till or bedrock into the surficial aquifer, although only a small area of the surficial aquifer actually contributes water to the well. Source-water areas in uplands can include substantial areas on both sides of a river. Source-water areas for wells in this setting are affected mainly by the rate of ground-water recharge and by the degree of anisotropy.

Source-water areas to wells in a broad valley setting (bedrock with a low angle of dip) are affected greatly by fracture properties. The effect of a given fracture is to channel the water downward from the surficial aquifer toward the open borehole. If leakage occurs through the vertical fractures, the source-water area is less affected by the fracture geometry. In one simulation, a fracture near the top of bedrock in a well allowed water to come from closer to the well, as in the case where there was not a good seal between the surficial aquifer and the borehole.

Ground-water-flow simulation models in Old Lyme, Connecticut, showed source-area results similar to results from the generic simulation models. Four simulation models at Old Lyme were used to test the effect of different assumptions about fracture-related properties: (A) a homogeneous medium, (B) discrete zones of high hydraulic conductivity that represent zones of closely spaced vertical fractures, (C) homogeneous rock with fracturing in one direction, and (D) a combination of (B) and (C). The four models fit the available data equally well. Source-water areas for the four different conceptual models were very similar to one another, except when the aquifer was simulated with a preferred orientation of hydraulic conductivity

(conceptual model C). When a preferred orientation of hydraulic conductivity was assumed, modeled source-water areas were elongated in the direction of the principal hydraulic conductivity. Model C is reasonable because that model estimates a direction of higher hydraulic conductivity in the strike direction of layering, as expected, and the available hydrologic data support it. Model C also calculated the most difference in the estimated source-water areas; therefore, it is important to know if an aquifer has a preferred orientation of hydraulic conductivity.

Introduction

Ground water is an important source of drinking water in Connecticut. About 3,000 public water-supply systems in Connecticut serve water to about 2.7 million people. Most of these systems have surface-water sources, but about 300,000 people have publicly supplied ground water as their source of drinking water. In addition, about 700,000 people drink ground water from private wells, and about 190,000 people drink ground water from non-community systems, such as schools and businesses. There are 403 public water-supply systems, which serve about 132,000 people in total, with at least one bedrock well.

Most contaminants derived from humans in ground water are related to activities at the land surface and enter the ground-water-flow system at the water table. One approach to protection of ground-water supplies is to delineate source-water areas and then implement ground-water protection practices on the overlying land surface. Contaminant sources in the source-water areas determine the potential vulnerability of the source water. The Connecticut Aquifer Protection Program uses extensive data collection and three-dimensional numerical ground-water modeling to provide an estimate of the source-water area for large (serving more than 1,000 people) community water-supply systems in Connecticut. For smaller systems, many of which rely on water in bedrock, the Connecticut Departments of Public Health (DPH) and Environmental Protection (DEP) adopted a two-phase approach. The first phase entailed an initial estimate of the source-water area using a calculated fixed-radius method. In this application, all bedrock properties are generalized. The initial estimates will be revised in the second phase, based on information assembled and analyzed in this study, which allows the spatially varying properties of bedrock to be considered.

Understanding ground-water flow in fractured bedrock is important for delineating source-water areas for bedrock wells. In the complex geologic terrain of Connecticut, fractures and fracture zones are the principal conduits for ground-water flow. Characterization of the types of fractures in the bedrock units in Connecticut and the geometry of those fracture systems at a regional scale will allow a better understanding of the source of water to wells tapping bedrock. To provide information about these issues, the U.S. Geological Survey (USGS) began a cooperative study in 1999 with the Connecticut Department of Public Health (CTDPH) to investigate ground-water flow in bedrock as part of the Connecticut Source-Water Assessment Program (SWAP).

Purpose and Scope

This report describes the geohydrologic factors that affect the size, location, and shape of source-water areas to bedrock wells in Connecticut. It includes information on fracture characteristics and their relation to bedrock geology and on the role of surficial geology in the storage and conveyance of water to wells. Source-water areas are delineated using analytical models. More complex numerical simulation models were constructed to illustrate how geohydrologic factors can be used to delineate the location and shape of source-water areas. The geohydrologic factors are grouped into settings that consist of
(1) wells on hilltops and hillsides with no surficial aquifer,
(2) wells in a narrow valley with a surficial aquifer, and
(3) wells in a broad valley with a surficial aquifer. Source-water areas from a numerical simulation model in the Old Lyme quadrangle (the "case study") are compared to the results of the analytical models. This report also contains a synthesis of previous investigations on fractured bedrock in the Northeast (appendix 1) and a description of bedrock units in the State (appendix 2).

Previous Investigations

E.E. Ellis (1909) described the relation of geologic structure to source-water areas in Connecticut. By carefully examining fracture patterns in outcrops and relating them to hydrologic characteristics observed in water wells, Ellis was able to make preliminary statements about the possible sources of water to wells. He noted that "the contributing area to a granite well should occupy a space with an approximately uniform radius around the well." In schist and gneiss, "a single well, instead of drawing water from an area surrounding it on all sides, will draw from long distances through the feeding fractures and vertical fractures connecting with them."

Fractured bedrock in the northeastern United States has been studied extensively. The extension and modification of Ellis' ideas presented in this report are possible through the advancements in geologic mapping, analytical techniques, and water-well reporting over the last century. A synthesis of previous investigations on regional ground-water flow and areas contributing recharge to wells in crystalline and sedimentary bedrock, studies of well yield, and studies of fracture-domain mapping is provided in appendix 1.

Methods

An important premise of this study is that regional geologic characterization can be used to predict physical properties of bedrock, including fracture geometry, and that this information can be used to improve models of ground-water flow in fractured bedrock. This study involved (1) grouping map units from two statewide geologic maps (bedrock and surficial materials) into groups (factors) that are thought to affect the size, location, and shape of source-water areas; (2) simulating ground-water flow in settings based on characteristics of the combined factors; and (3) conducting a case study in the Old Lyme quadrangle, including a ground-water-flow simulation, geophysical logging of boreholes, an aquifer test, and a study of bedrock outcrops.

Grouping of Geologic Units

A classification scheme for the bedrock geologic units used in this report is based on the Bedrock Geological Map of Connecticut[1] (Rodgers, 1985).The scheme was developed in conjunction with geologists from the Connecticut Geological and Natural History Survey and is based on lithologic and structural factors that produce different fracture characteristics. Approximately 5,000 strike (the directional orientation of bedding and foliation) and dip (the angle of inclination) measurements from the Bedrock Geological Map of Connecticut (Rodgers, 1985) were digitized from 1:50,000-scale compilation sheets to show the orientation of primary geologic structure (bedding and foliation) in the bedrock.

The bedrock classification scheme also was based, in part, on studies of fractures at road cuts and recent geologic mapping. A study of fractures at road cuts along Rt. 9 in southeastern Connecticut (Zinsser, 2002) and an investigation in a 6-quadrangle area in southeastern Connecticut (Zeitlhofler, 2003) have demonstrated that systematic fracture geometry is present in various rock units and in various geologic terrains, and that this fracture geometry commonly can be observed as linear aspects of the land-surface topography near the outcrop. This information, in conjunction with other previous observations (for example, Mabee and others, 1994; Mabee, 1998; Newell and Wise, 1964), has contributed to the development of the statewide bedrock classification scheme discussed in this report.

Geologic mapping was conducted in two quadrangles representing different crystalline (metamorphic) bedrock terrains in Connecticut: the New Milford quadrangle in western Connecticut, and the Old Lyme quadrangle in the southeastern coastal area (fig. 1). The geologic mapping included focused data collection of brittle fracture characteristics. The geologic map for the New Milford quadrangle was recently published (Walsh, 2004), and the geologic map for the Old Lyme quadrangle is being prepared (G.J. Walsh, U.S. Geological Survey, written commun., 2004); only the map for the Old Lyme quadrangle is discussed in this report. Geologic data gathered during mapping of the Old Lyme quadrangle were incorporated into the interpretation of the borehole logging, aquifer testing, and simulation of ground-water flow in this quadrangle. Knowledge gained about fracture systems during mapping of the New Milford and Old Lyme quadrangles was incorporated into the statewide classification of bedrock units.

A second classification scheme for the surficial geologic units shown on the Surficial Materials Map of Connecticut (Stone and others, 1992) was developed on the basis of the character of the material directly overlying bedrock.

Simulation of Ground-Water Flow

The effect of presumed fracture patterns on ground-water flow was assessed by constructing simulation models of ground-water flow in three settings based on the conceptual models developed in this study. Other factors that affect the location and shape of the source-water area were incorporated into these models, such as topographic position, ground-water recharge, presence of a surficial aquifer, and details of well construction.

Ground-water flow was simulated using the computer program MODFLOW-2000 (Harbaugh and others, 2000; Hill and others, 2000), which is based on MODFLOW, a computer program that simulates three-dimensional ground-water flow through a porous medium by using a finite-difference method (McDonald and Harbaugh, 1988). MODFLOW-2000 has the capability to solve a MODFLOW calibration problem by calculating values of selected input data that result in the best match between measured and model-calculated values. The partial-differential equation of ground-water flow used in MODFLOW is (McDonald and Harbaugh, 1988, p. 2-1):

$$\frac{\partial}{\partial_x}\left(K_{xx}\frac{\partial h}{\partial x}\right) + \frac{\partial}{\partial_y}\left(K_{yy}\frac{\partial h}{\partial y}\right) + \frac{\partial}{\partial_z}\left(K_{zz}\frac{\partial h}{\partial z}\right) + W = S_s\frac{\partial h}{\partial t}, \qquad (1)$$

where

K_{xx}, K_{yy}, *and* K_{zz} are values of hydraulic conductivity along the x, y, and z coordinate axes, which are assumed to be parallel to the major axes of hydraulic conductivity (L/T);

h is the potentiometric head (L);

W is a volumetric flux per unit volume representing sources and/or sinks of water, with W<0.0 for flow out of the ground-water system, and W>0.0 for flow in (T^{-1});

S_S is the specific storage of the porous material (L^{-1}); and

t is time (T).

Equation 1, when combined with boundary and initial conditions, describes transient three-dimensional ground-water flow in a heterogeneous and anisotropic medium, provided that the principal axes of hydraulic conductivity are aligned with the coordinate directions. In this study, only steady-state ground-water flow was simulated, so the term on the right-hand side of equations was equal to zero. Source-water areas were delineated using the particle-tracking computer program MODPATH (Pollack, 1994). The computer simulation of ground-

[1]Names of geologic formations used in this report correspond to those used on the Bedrock Geological Map of Connecticut (Rodgers, 1985) and may not match the geologic names used by the U.S. Geological Survey.

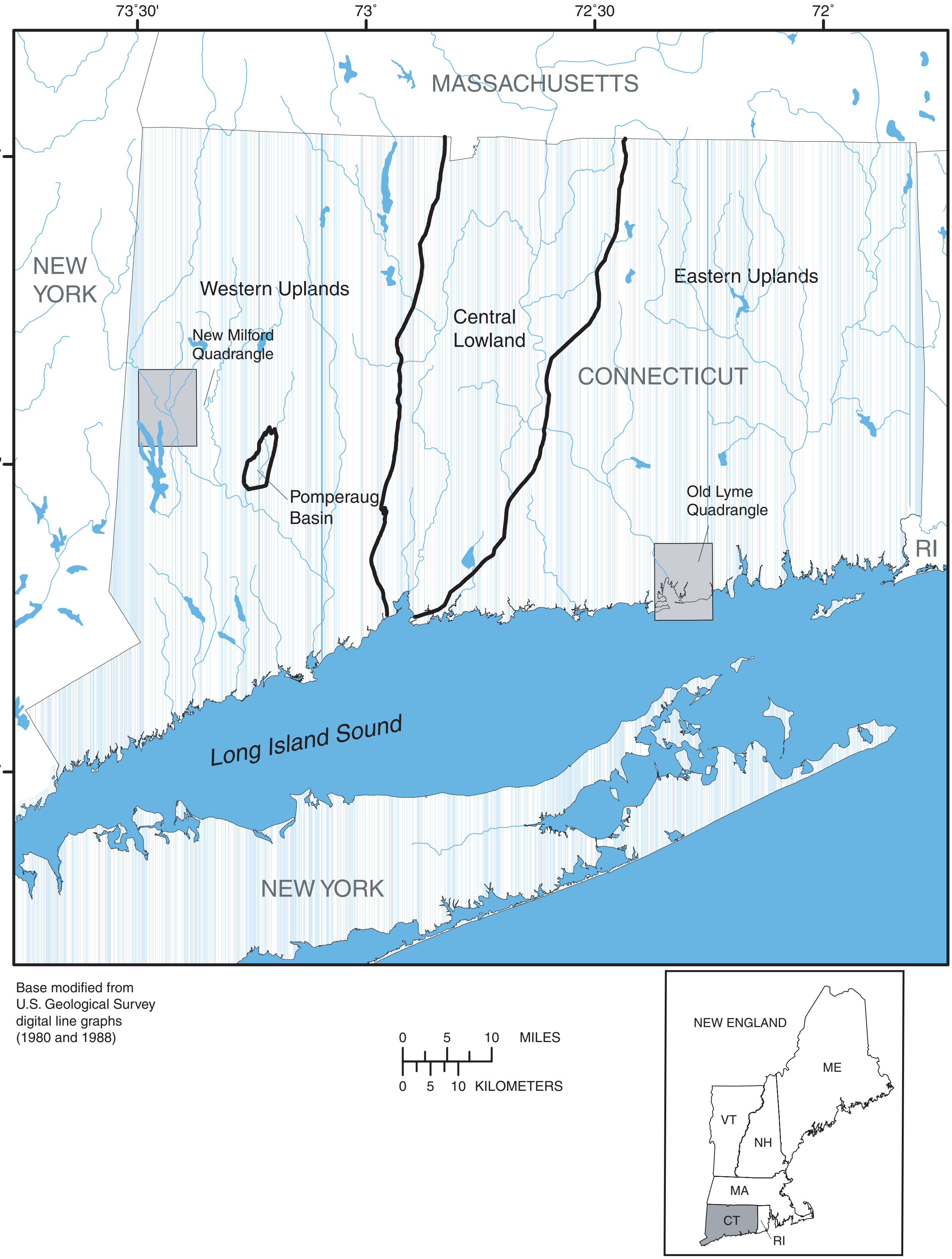


Figure 1. Location of the study area, Connecticut.

water flow in the Old Lyme quadrangle was calibrated using ground-water levels measured by well drillers, and estimated streamflow. Different assumptions of homogeneous, heterogeneous, isotropic, and anisotropic aquifer conditions were tested with the simulation model. The calibrated simulation model was used to delineate source-water areas to hypothetical wells in different settings.

Case Study

To test concepts about the relation between geologic structure and the delineation of source-water areas in an actual geohydrologic setting, ground-water flow was simulated and source-water areas were delineated for hypothetical well locations in the Old Lyme quadrangle. The investigation in Old Lyme included borehole-geophysical logging and a 20-day aquifer test at a community well field. Borehole-geophysical analysis was conducted to define important water-bearing fractures in the bedrock at the well field and to relate fracture types and fracture geometry to fractures observed in nearby outcrops. The aquifer test was conducted, in part, to assess how the preferred direction of fracturing in the bedrock affected ground-water flow during pumping at the well field. Observations at highway road cuts and rock quarries were made to characterize the types of fracturing in various rock units.

Geohydrologic Factors

Lithologic units, fracture types, and fracture geometry are important factors in assessing source-water areas to bedrock wells, because water flows through and is stored in fractures in the bedrock as it moves to water-supply wells. The character and thickness of unconsolidated (predominantly glacial and postglacial) sediments overlying bedrock also are important factors in assessing source-water areas, because these materials store ground water that can recharge bedrock.

The landscape of Connecticut is made up of three physiographic regions—the Eastern Uplands, the Western Uplands, and a broad Central Lowland (fig. 1). Many small- to medium-sized valleys cut the uplands, and a line of narrow ridges separates the Central Lowland into two basins. This broad-scale physiography of Connecticut is a result of differences in the character of the underlying bedrock—harder, more erosion-resistant rocks underlie the hills and ridges; less resistant rocks underlie the broader valleys. Glacial and postglacial processes produced the finer details of the landscape. Flood plains are a result of downcutting of rivers and streams; swamps and marshes developed in poorly drained areas in postglacial time. In the uplands, the position of many smaller valleys is controlled by primary geologic structure and crosscutting fracture systems in the bedrock. Nearly all the bedrock-controlled valleys are partially filled with coarse-grained and fine-grained glacial stratified deposits, and bedrock hills have been sculpted by glacial ice and mantled by a discontinuous blanket of till. Erosion and weathering through geologic time have produced broad valleys in the Central Lowland because the sedimentary rocks are less resistant to erosion; by comparison, erosion and weathering in the uplands have produced narrower valleys because the crystalline rocks are more resistant to erosion. Thick, permeable glacial stratified deposits, which can provide recharge to the bedrock, cover much of the area in the broad valleys of the state.

Bedrock Geology

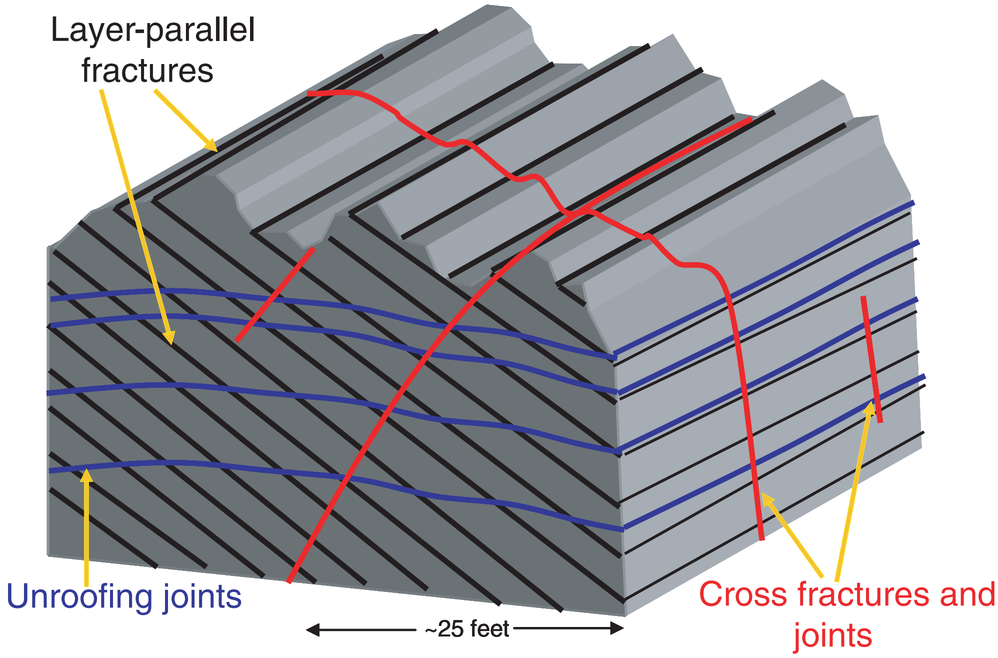
The bedrock geology of Connecticut has been previously mapped at 1:24,000-scale for most areas and is compiled at 1:125,000-scale on the Bedrock Geological Map of Connecticut (Rodgers, 1985). For the most part, the description of the bedrock on published geologic maps includes information on the lithology and primary structure of the rock units. Little information is given about fractures and fracture orientation on existing maps.

Bedrock in Connecticut is highly variable in lithology and structure and includes Proterozoic through mid-Paleozoic metamorphic rocks, late Paleozoic and early Mesozoic igneous rocks, and early Mesozoic sedimentary rocks. The hydrologic properties of the various rock types are similar in that ground water moves primarily through fractures and fracture zones and not through the rock matrix; however, the degree of fracturing and the geometry of fracture systems differs among rock types and from place to place within rock types. The primary structure of the bedrock units—foliation or layering in metamorphic rocks and bedding in sedimentary rocks—is a controlling factor in the type of fractures found in different rock units.

Types of Fractures

Three principal types of fractures—layer-parallel fractures, unroofing joints, and cross fractures—are present in the bedrock of Connecticut (fig. 2). These fractures provide pathways for the flow of ground water. Layer-parallel fractures develop as partings along bedding in sedimentary rock and along layering or foliation in metamorphic rock. Dips of these fractures range from horizontal (gentle) to vertical (steep). Unroofing joints develop as strain-release fractures parallel to the land surface as overlying rock is removed by erosion through geologic time; this vertical stress produces generally subhorizontal fractures. Cross fractures and joints are a result of tectonically generated stresses (horizontally directed extension and compression) on the bedrock; this produces typically near-vertical or steeply dipping fractures.

When layering is present in a rock unit, either as bedding in sedimentary rock or as foliation and (or) layering in metamorphic rocks, layer-parallel fracturing usually dominates the fracture geometry because it is pervasive throughout the rock. Layer-parallel fractures are present in many rock types in Connecticut. The orientation and dip angle of layering, where layering is present, are important for understanding ground-water



Layer-parallel fractures — developed as partings along bedding in sedimentary rock and compositional layering or foliation in metamorphic rock; dips of these fractures can be can be gentle or steep (horizontal through vertical).

Unroofing joints — developed as strain-release fractures parallel to the land surface as overlying rock was removed by erosion through geologic time; vertical stress produced generally subhorizontal fractures.

Cross fractures and joints — developed as a result of tectonically generated stresses (horizontally directed extension or compression) on the bedrock producing typically near-vertical or steeply dipping fractures.

Figure 2. General types of fractures in bedrock units in Connecticut.

flow in bedrock. As used in this report, steeply dipping rocks dip at angles of more than or equal to 50^{o} and gently dipping rocks dip at angles of less than 50^{o}.

Unroofing joints provide lateral connections between steeply dipping fractures. They occur as separate fracture sets predominantly in nonlayered rocks and in layered rocks that have steeply dipping foliation (>50^{o}). Based on observations in road cuts and quarry exposures, unroofing joints are closely spaced only in the upper 30 ft or so of the bedrock, become more widely spaced with depth, and are less prevalent below about 200 ft.

Cross fractures and joints generally are steeply dipping to vertical and potentially provide vertical connection with overlying surficial materials and between major subhorizontal water-bearing zones. Cross fractures and joints generally are more widely spaced than layer-parallel fractures, are concentrated in zones, and have an orientation that commonly is related in some way to the primary structure in layered rocks.

Lithogroups

Rock type is a major factor in the classification of fracture characteristics, particularly in reference to whether the rock is layered or nonlayered. Bedrock formations from the Bedrock Geologic Map of Connecticut (Rodgers, 1985) have been categorized into nine groups, termed lithogroups, based primarily on rock type (lithology) (fig. 3; table 1; appendix 2). Seven lithogroups include metamorphic rocks of Paleozoic and Proterozoic age, and two lithogroups include sedimentary and igneous rocks of Mesozoic age. As previously noted, names of geologic formations used in this report correspond to those used on the Bedrock Geological Map of Connecticut (Rodgers, 1985) and may not match the geologic names used by the U.S. Geological Survey.

Paleozoic and Proterozoic rocks (lithogroups GN, SCH, MIX, MBL, GR, GRL, and MF; fig. 3) underlie eastern and western Connecticut and are the oldest rock types in Connecticut (800 million years to 250 million years). They are predominantly metasedimentary and meta-igneous rocks that have undergone extensive metamorphism and ductile deformation during at least three Paleozoic orogenic events, as well as brittle deformation in the more recent geologic past.

Mesozoic rocks (lithogroups SED and BAS) underlie the broad Central Lowland and the smaller Pomperaug Basin in western Connecticut, and are the youngest rock types in Connecticut (about 200 million years old). They include sedimentary and igneous rocks that have not undergone metamorphism or ductile deformation. Only brittle faults, cross fractures, and layer-parallel partings are present.

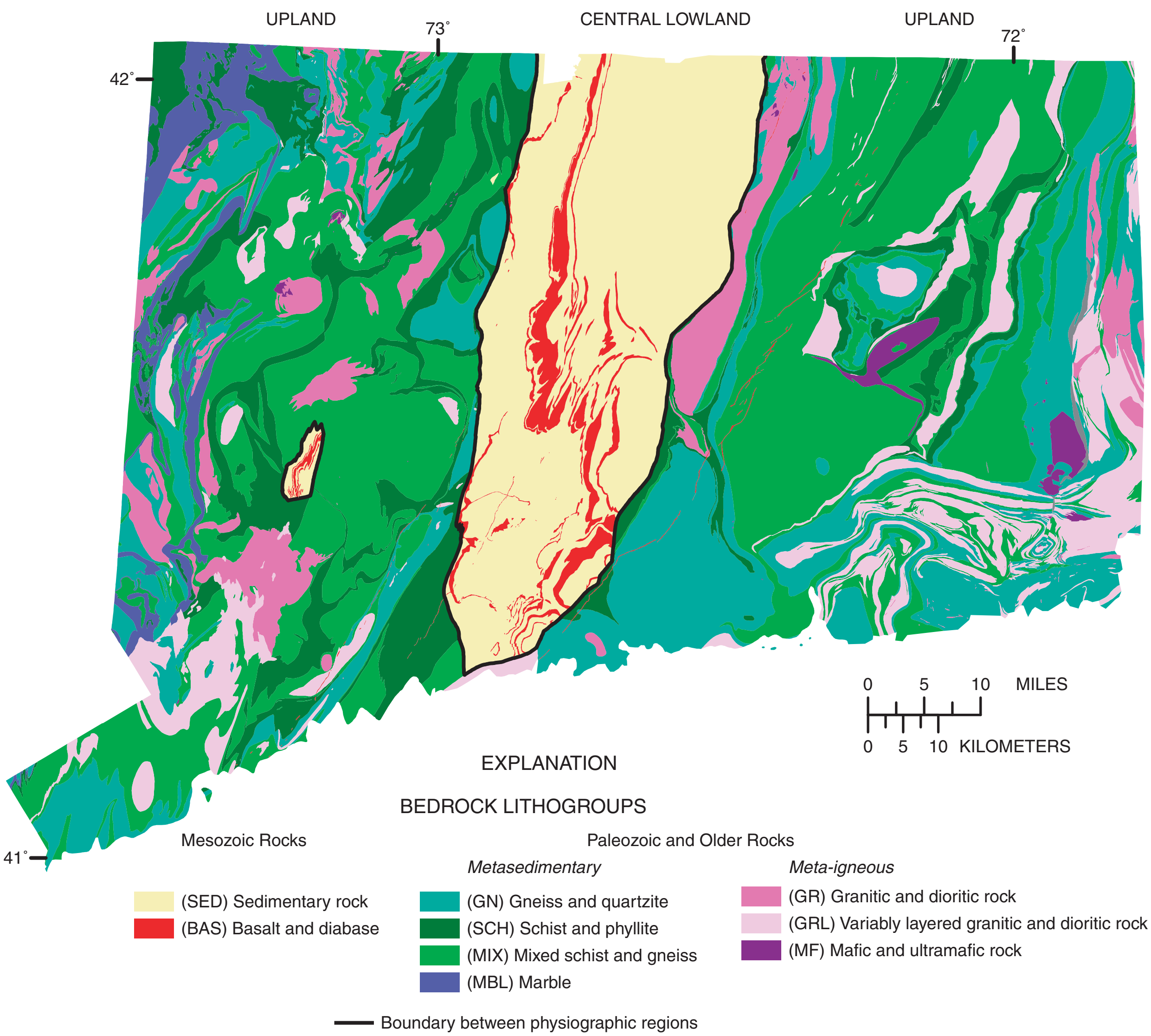


Figure 3. Lithogroups in Connecticut based on general types of fractures in the rock.

Table 1. Description of lithogroups and fracture geometry in Connecticut.

[SED, sedimentary; BAS, basalt; GN, gneiss; SCH, schist; MIX, mixed; MBL, marble; GR, granite; GRL, layered granite; MF, mafic; LAY, layered; NON, non-layered; VAR, variably layered]

Lithogroup	Fracture geometry	Description	Hyperlink
SED	Layered	Layered sedimentary rock including sandstone, siltstone, shale, and conglomerate	SED.pdf
GN		Well-layered metamorphic rock including gneiss and quartzite	GN.pdf
SCH		Well-layered metamorphic rock including schist and phyllite	SCH.pdf
MIX		Well-layered metamorphic rock unit that includes belts of both gneiss and schist, and local quartzite layers	MIX.pdf
BAS	Nonlayered	Massive (nonlayered) basaltic lava flows and shallowly intruded diabase dikes and sills that are interlayered with sedimentary rocks	BAS.pdf
MBL		Poorly layered, mostly massive marble, locally schistose, metamorphosed limestone and calcareous siltstone and sandstone	MBL.pdf
GR		Poorly layered meta-igneous rock including granitic and dioritic gneisses and pegmatite	GR.pdf
MF		Nonlayered meta-igneous rock including gabbro and ultramafic rock	MF.pdf
GRL	Variably layered	Variably layered granitic and dioritic rock	GRL.pdf

Fracture Geometry

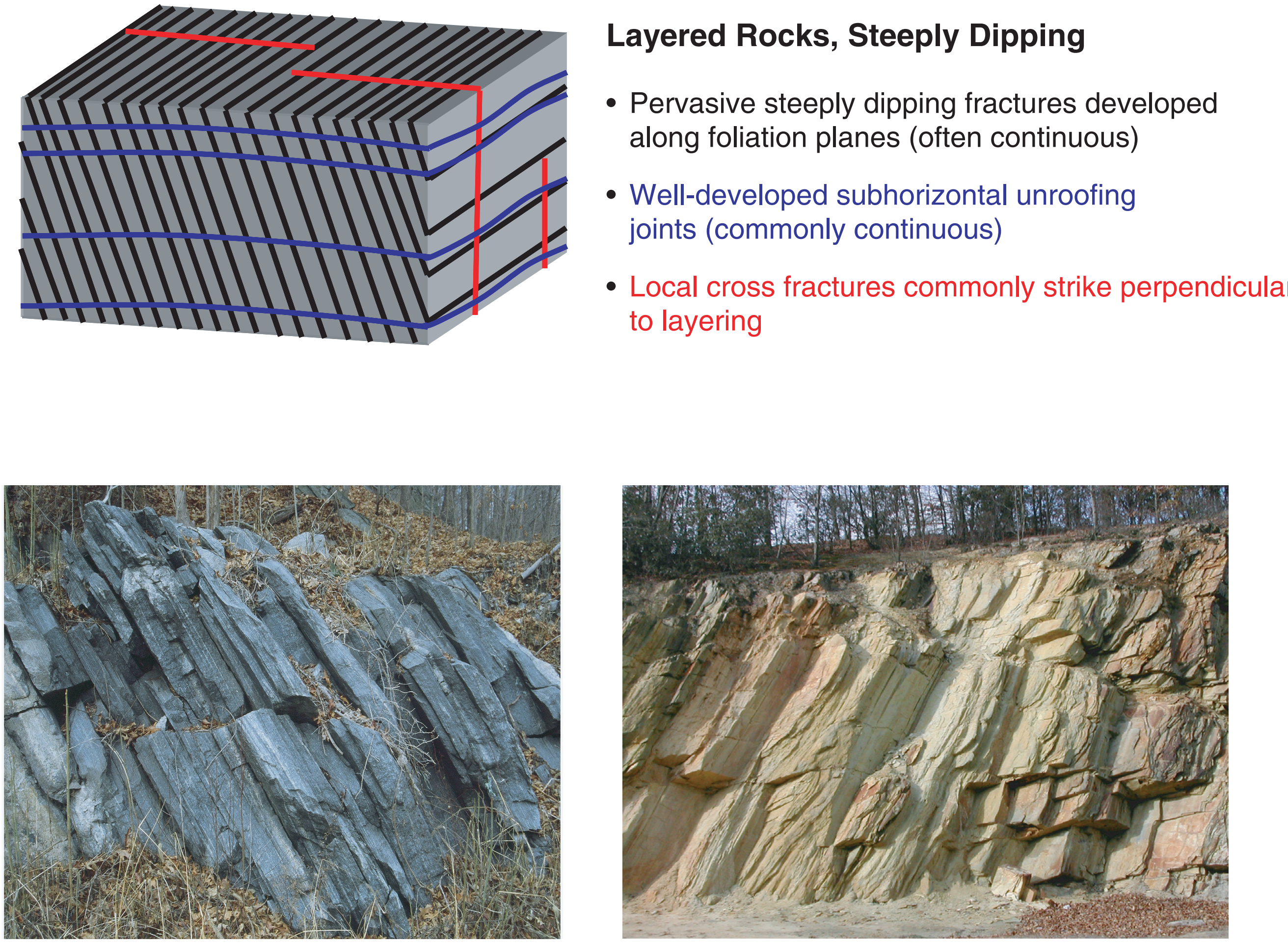
Basic fracture geometry in the nine lithogroups is defined primarily by the presence or absence of layering in the rock unit, and, in layered rock, by the angle of dip in the layering. Layered rocks underlie approximately 75 percent of the total area of Connecticut and comprise four lithogroups. Nonlayered and variably layered rocks underlie 25 percent of the state and comprise five lithogroups.

Layered rocks (table 1; SED, GN, SCH, and MIX) include well-foliated and layered gneisses and schists in eastern and western Connecticut and sedimentary rocks of the Central Lowland. Where layered rocks dip steeply (fig. 4), layer-parallel fracturing generally is dominant. Unroofing joints, which provide continuous lateral connections between steeply dipping layer-parallel fractures, typically also are well developed. In addition, in many places, at least one dominant set of cross fractures (or joints) strikes perpendicular (or nearly so) to the strike of layering. Other sets of cross fractures also may be present, but these generally are less frequent and less continuous than the dominant set of cross fractures.

Where layered rocks dip gently (fig. 5), layer-parallel fracturing also is the dominant fracture type. In addition, several sets of cross fracture commonly are present. Typically, the main set of cross fractures has a similar strike as the layering, but dips perpendicular to the layering. As in steeply dipping rocks, cross fracturing in gently dipping rocks generally is less pervasive and less continuous than the layer-parallel fracturing. Cross fractures provide vertical connections between layer-parallel fractures. Unroofing joints generally do not form as a separate fracture set; instead, strain release from unroofing has occurred along gently dipping layer-parallel fractures, enhancing their aperture.

Nonlayered rocks (BAS, MF, GR, and MBL) include extrusive igneous rocks, poorly foliated meta-igneous rocks, and massive marble. In these rocks (fig. 6), layer-parallel fracturing is absent or poorly developed; fracturing is dominated by well-developed subhorizontal unroofing joints and steeply dipping, tectonically generated fractures, or (in the case of basalt) cooling joints. Cross fractures (or cooling joints) in these rocks have more random orientations than in layered rocks. In basalt (BAS), cooling joints are pervasive throughout the rock, resulting in a highly fractured rock type. In the other groups (MBL, GR, MF), fractures are much more sporadic and localized and the fracture network is not strongly oriented. Variably layered rocks display characteristics of either layered or nonlayered rocks.

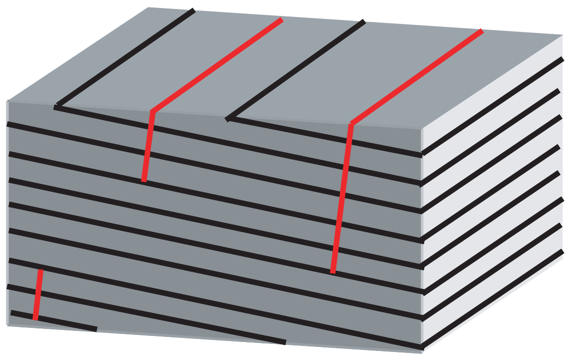
The strike and dip—the directional orientation of layering and foliation—is particularly important in steeply dipping layered rocks because it may have a strong effect on the direction of ground-water flow. Structural symbols (strike and dip measurements) from the Bedrock Geological Map of Connecticut (Rodgers, 1985) show the orientation of primary geologic structures (bedding and foliation) in the bedrock (figs. 7 and 8). This information potentially is useful for hydrologic analysis in two ways: (1) the strike indicates the trend of layer-parallel fracturing, and (2) the dip angle indicates whether layer-parallel fracturing is steep or gentle. In many areas, fractures developed along foliation and layering will have strongly preferred orientations that may affect the direction of ground-water flow in bedrock.



Rope Ferry Gneiss, Essex St., Deep River

Collins Hill Formation, Rt. 9, Haddam

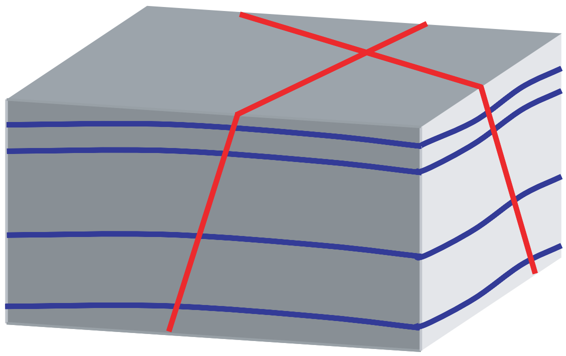
Figure 4. Typical fracture geometry in steeply dipping layered rocks in Connecticut.



East Berlin Formation, Rt. 9, Berlin

Potter Hill Formation, Old Lyme

Figure 5. Typical fracture geometry in gently dipping layered rocks in Connecticut.



Nonlayered Rocks

- Pervasive gently dipping unroofing joints in upper zones becoming less frequent with depth (continuous)
- Local steeply dipping cross fractures with variable orientations

Stony Creek Granite

Stockbridge Marble

Figure 6. Typical fracture geometry in nonlayered rocks in Connecticut.

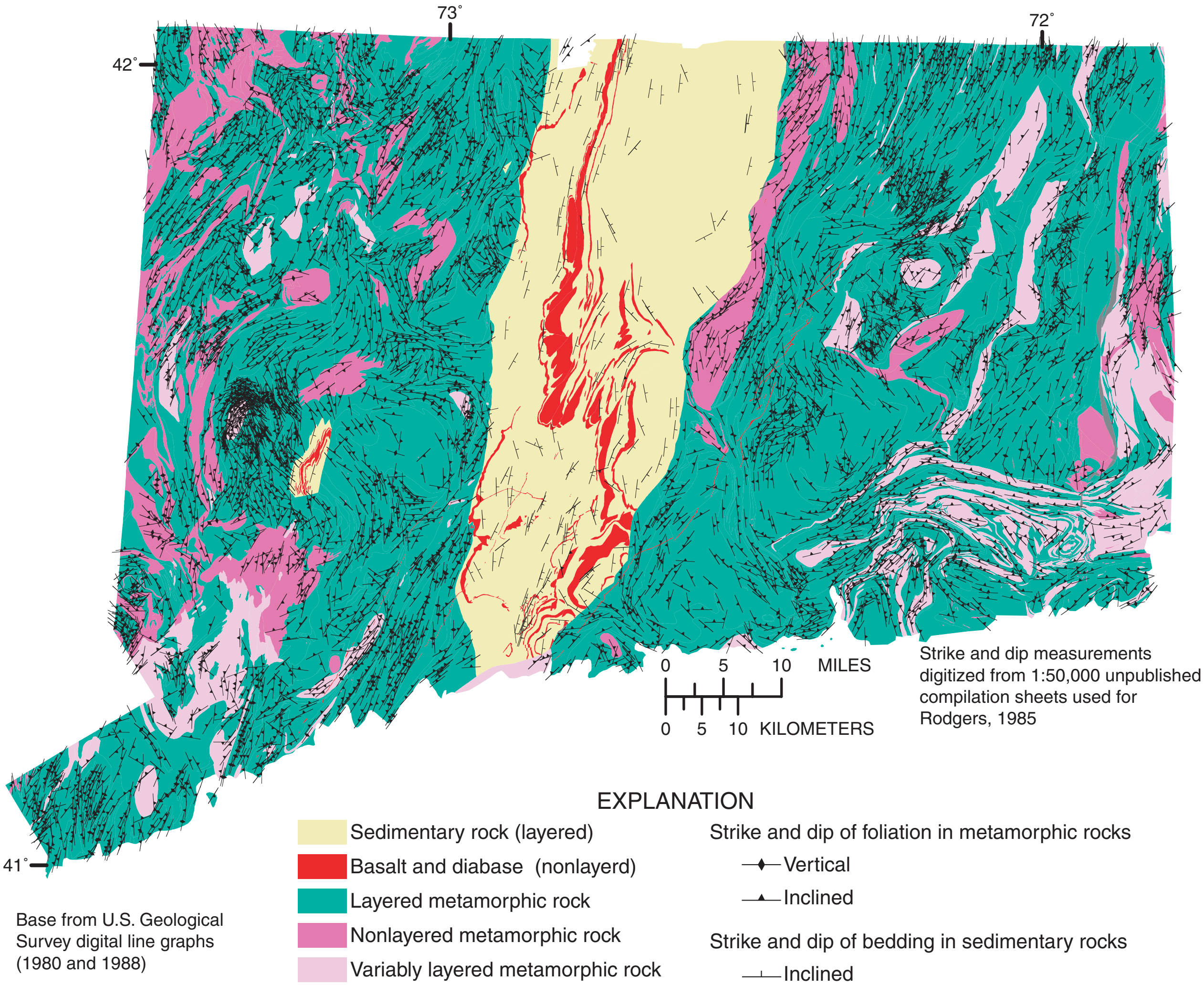


Figure 7. Strike and dip directions of foliation in metamorphic rocks and bedding in sedimentary rocks in Connecticut.

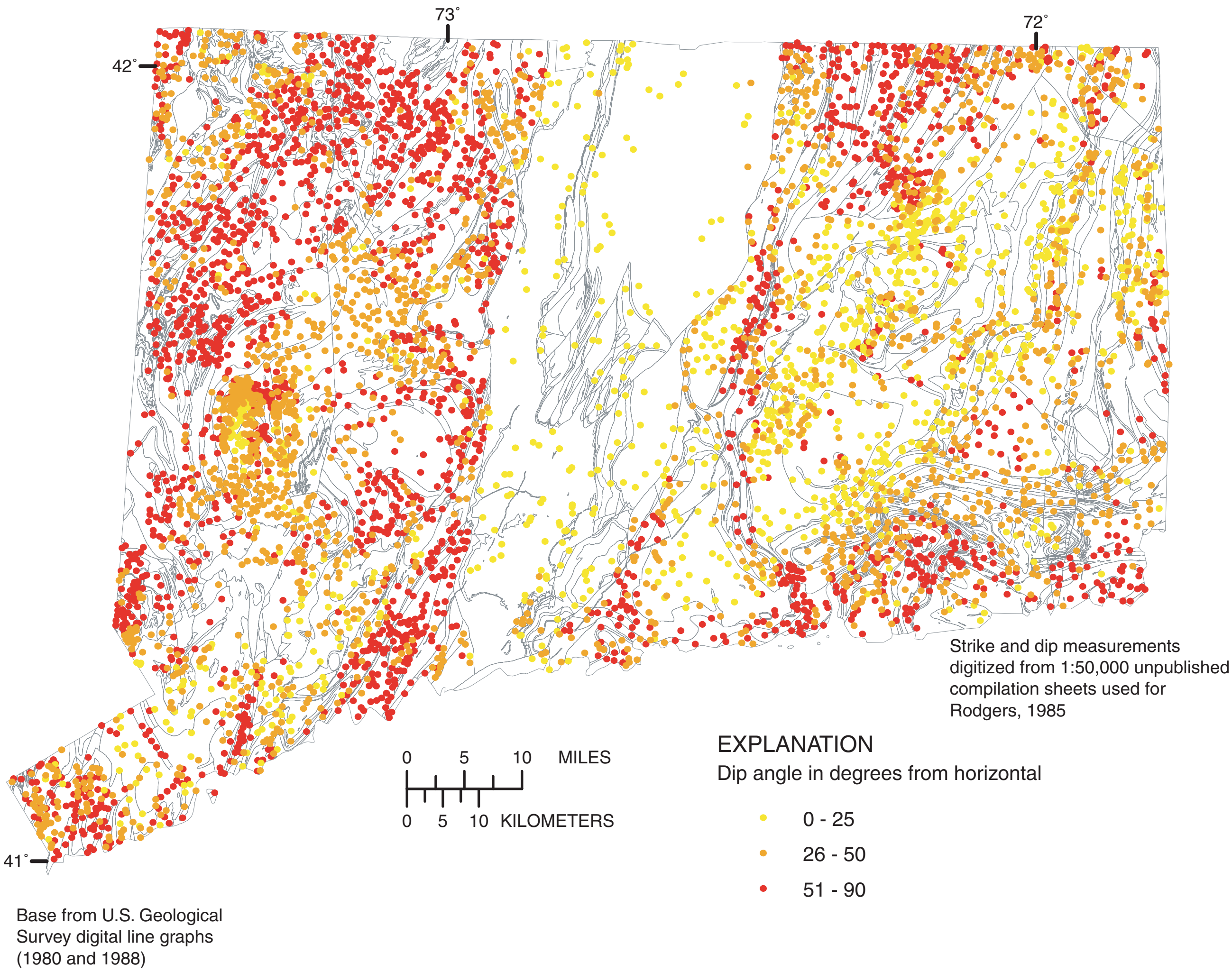


Figure 8. Dip angle of foliation in metamorphic rocks and bedding in sedimentary rocks in Connecticut.

Surficial Geology

The Surficial Materials Map of Connecticut (Stone and others, 1992) was used to describe the character of the material directly overlying the bedrock and the presence or absence of thick unconsolidated sediments (table 2). Unconsolidated materials (overburden) are classified on the Surficial Materials Map of Connecticut as three basic types—glacial ice-laid deposits, glacial meltwater deposits, and postglacial deposits. These are grouped into 49 map units that characterize the texture (grain size) and, to a certain extent, the thickness of these sediments. Glacial ice-laid deposits include end moraine deposits and two map units of glacial till—thick till and thin till. Glacial meltwater deposits (also called glacial stratified deposits) consist of four basic textural units—gravel deposits, sand and gravel deposits, sand deposits, and fine deposits (very fine sand, silt, and clay)—and 17 combinations of these 4 basic units. The combinations of the four basic units, in various order of superposition, are known as "stack units." Postglacial deposits include floodplain alluvium, swamp deposits, salt-marsh deposits, and 19 stack units where these materials overlie various units of glacial stratified deposits; also included are beach deposits, talus, and artificial fill. An additional factor has been added that simplifies the 49 units into 4 units that describe the character of surficial material directly overlying the bedrock and its ability to store water (fig. 9). The simpler grouping of surficial materials was made to assess the effect of thick surficial materials on the shape and location of source-water areas. In this simplified classification, two major categories represent the presence or absence of thick unconsolidated sediments. The surficial aquifer (SA) category includes units CS, FS, and TT; the no-surficial aquifer (NSA) category is the unit T (table 2, appendix 2).

Table 2. Description of surficial aquifer units in Connecticut.

[SA, surficial aquifer present; NSA, no surficial aquifer; CS, coarse stratified deposits; FS, fine stratified deposits; TT, thick till; T, thin till; ft, feet]

Surficial aquifer	Surficial unit	Description	Hyperlink
SA	CS	Includes areas of gravel deposits, sand and gravel deposits, sand deposits, and floodplain alluvium where these units make up the entire thickness of surficial materials, and 10 stack units of coarse-grained units (such as sand and gravel overlying sand). Also includes eight stack units in which fine-grained deposits overlie coarse-grained units (such and fines overlying sand), and four stack units in which swamp deposits overlie coarse-grained deposits (such as swamp deposits overlying sand and gravel). Also included in this category are beach deposits along the coast and artificial fill that consists of large areas of "made land." Surficial materials in this category range in thickness from a few feet near the edges of these map unit areas to several hundred feet in the thickest sections; an average thickness of 46 ft is indicated in records of approximately 1,900 inventoried bedrock wells penetrating this unit across the state.	COARSE.pdf
	FS	Includes areas of very fine sand, silt, and clay deposits, swamp deposits, and salt-marsh deposits where these units make up the entire thickness of surficial materials. Also includes 10 stack units where coarse-grained stratified deposit and (or) floodplain alluvium overlie fine-grained deposits (such as sand and gravel overlying sand overlying fines). Surficial materials in this category range in thickness from a few feet near the edges of the map-unit area to several hundred feet in the thickest sections; an average thickness of 76 ft is indicated in records of 470 inventoried bedrock wells penetrating this unit across the state.	FINE.pdf
	TT	Includes areas of till (a nonsorted, generally nonstratified mixture of grain sizes from clay to large boulders) where this material is greater than 15 ft in thickness, typically in drumlins and on the northern or northwestern sides of bedrock hills; an average thickness of 53 ft is indicated in records of 973 inventoried bedrock wells penetrating this unit across the state.	THICK TILL.pdf
NSA	T	Includes areas where till is less than 15 ft thick and areas where till is absent and bedrock is present at land surface.	TILL.pdf

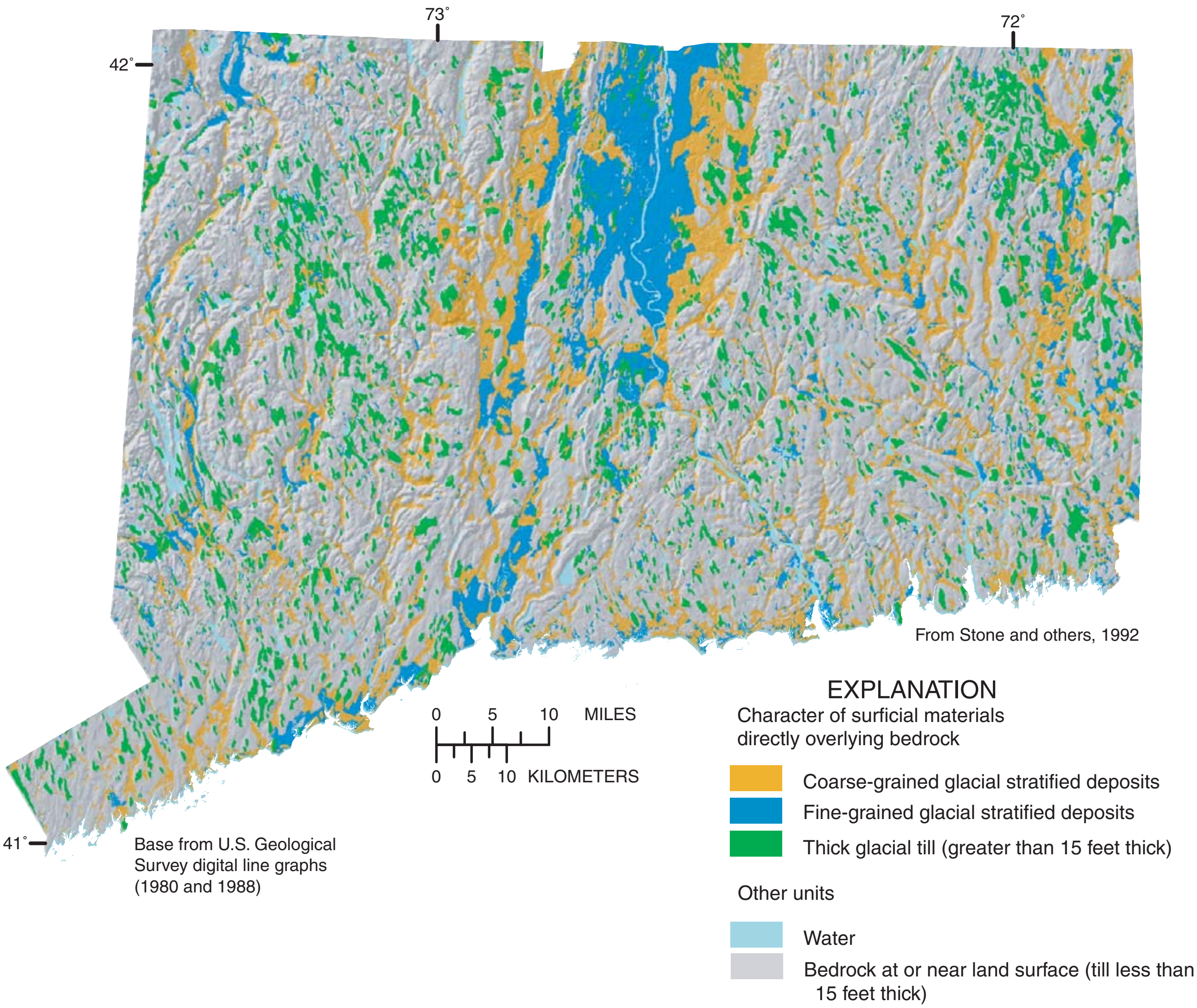


Figure 9. Character of surficial materials directly overlying bedrock in Connecticut.

Simulation of Ground-Water Flow to Assess Geohydrologic Factors and Their Effect on Source-Water Areas in Connecticut

The first step in estimating source-water areas to bedrock wells is to develop a conceptual model of ground-water flow (Barton and others, 1999). Subsequent refinement of the conceptual model can be based on different levels of effort and accuracy depending on the intended purpose. Methods that have been used to delineate source-water areas include relatively simple analytical models and more complex numerical models. The accuracy of the method increases with the efforts involved in determining aquifer characteristics and the level of detail applied to characterize hydrologic boundaries.

Source-water areas can be described by their size, location, and shape. The size of a source-water area is primarily a function of the water balance—the amount of water pumped from a well is balanced by sources of ground water, which include direct recharge from precipitation, natural recharge from losing streams, induced recharge from surface-water bodies, and possibly, removal of water from storage in the aquifer. The location of a source-water area refers (in this report) to where the source-water area is in relation to the well. The location of the source-water area is, in part, a function of the extent of the ground-water-flow system, which can range in size from local flow systems that discharge to small brooks and may only be seasonal, to regional flow systems that discharge to major water bodies. The lateral and vertical extents of local and regional flow systems are a function of the hydraulic properties of the aquifer and the location of recharge and discharge areas. The shape of a source-water area refers to the ratio of the dimensions of the source-water area—for example, long and narrow as opposed to short and broad. The shape of the source-water area is affected, in part, by the hydraulic properties of the bedrock. Hydraulic properties commonly are assumed to be spatially uniform (homogeneous) and the same in all directions (isotropic). Fracturing in a preferred direction could produce hydraulic properties that vary by direction (anisotropic) or by position (heterogeneous). In this report, an anisotropic aquifer is simulated by varying hydraulic conductivity in different directions to represent the combined effect of many fractures, and a heterogeneous aquifer is simulated by including zones of different hydraulic conductivity that represent individual fracture zones.

Analytical Models

Source-water areas in bedrock can be delineated in several ways, some of which were evaluated by Barton and others (1999). Two analytical methods have been used in this study to determine the size of the source-water area: (1) the calculated fixed-radius method—based on the volume of water stored within the aquifer for a given time period, and (2) the water-balance method—based on the recharge rate of the aquifer and the pumping rate of the well. These two analytical methods are simpler and easier to apply than the numerical methods described later in this report. The numerical methods allow the inclusion of more complex factors at a site and produce source-water areas that generally are not circular areas around the well; the location and shape of these areas are controlled by hydraulic properties and location of recharge and discharge areas in the aquifer.

A calculated fixed-radius method that is currently used by the Connecticut Department of Environmental Protection (Corinne Fitting, Connecticut Department of Environmental Protection, written commun., 2001) assumes a circular source-water area with the radius determined by

$$r = (Qt/(\pi n H))^{0.5}, \qquad (2)$$

where

r is the radius of the source-water area, in feet,
Q is the well discharge, in cubic feet per day,
t is the time of pumping, in days (assumed to be 180 days),
π is 3.1415926,
n is porosity (assumed to be 0.0022), and
H is saturated thickness, in feet (assumed to be 200 ft).

This equation can be simplified to $(130Q)^{0.5}$. This method yields a source-area radius of 806 ft for a well pumping 5,000 ft^3/d and 1,140 ft for a well pumping 10,000 ft^3/d.

The second analytical method for calculating the size of the source-water area is the water-balance method (Risser and Barton, 1995). This method requires an estimate of the ground-water recharge rate and the pumping rate of the well with the area determined by

$$A = Q/R, \qquad (3)$$

where

A is the area of the source-water area, in square feet,
Q is the well discharge, in cubic feet per day, and
R is the rate of ground-water recharge, in feet per day.

The rate of ground-water recharge can be calculated using the formula derived by Mazzaferro and others (1979):

$$R = (35 + 0.6 * CS) * MAR, \qquad (4)$$

where

R is ground-water recharge, in feet per day,
CS is the percentage of the basin having coarse-grained glacial stratified deposits at the surface, and
MAR is mean annual runoff, in feet per day.

Using GIS and equation 4, recharge was calculated for each of the approximately 2,800 small drainage basins in the state (fig. 10). Recharge rates range from 7 in/yr in basins underlain entirely by glacial till and bedrock to about 25 in/yr in basins underlain entirely by glacial sand and gravel. In areas of thin till and bedrock outcrop, most of the recharge (7 in/yr) could reach the water table. In sand and gravel areas, only part of the recharge (up to 25 in/yr) reaches the water table, because much of this recharge moves through the sand and gravel to surface-water bodies. Sand and gravel also has high porosity and a high storage capacity for providing water to bedrock wells that penetrate it. Equation 4, which is based on hydrograph separation from many studies in New England (Mazzaferro and others, 1979), may underestimate the amount of water that infiltrates because some water may discharge after traveling a short distance to ephemeral streams and topographic lows. The recharge rate in this report, therefore, refers to deep recharge, that is, only that portion of infiltration that reaches the average depth of the water table.

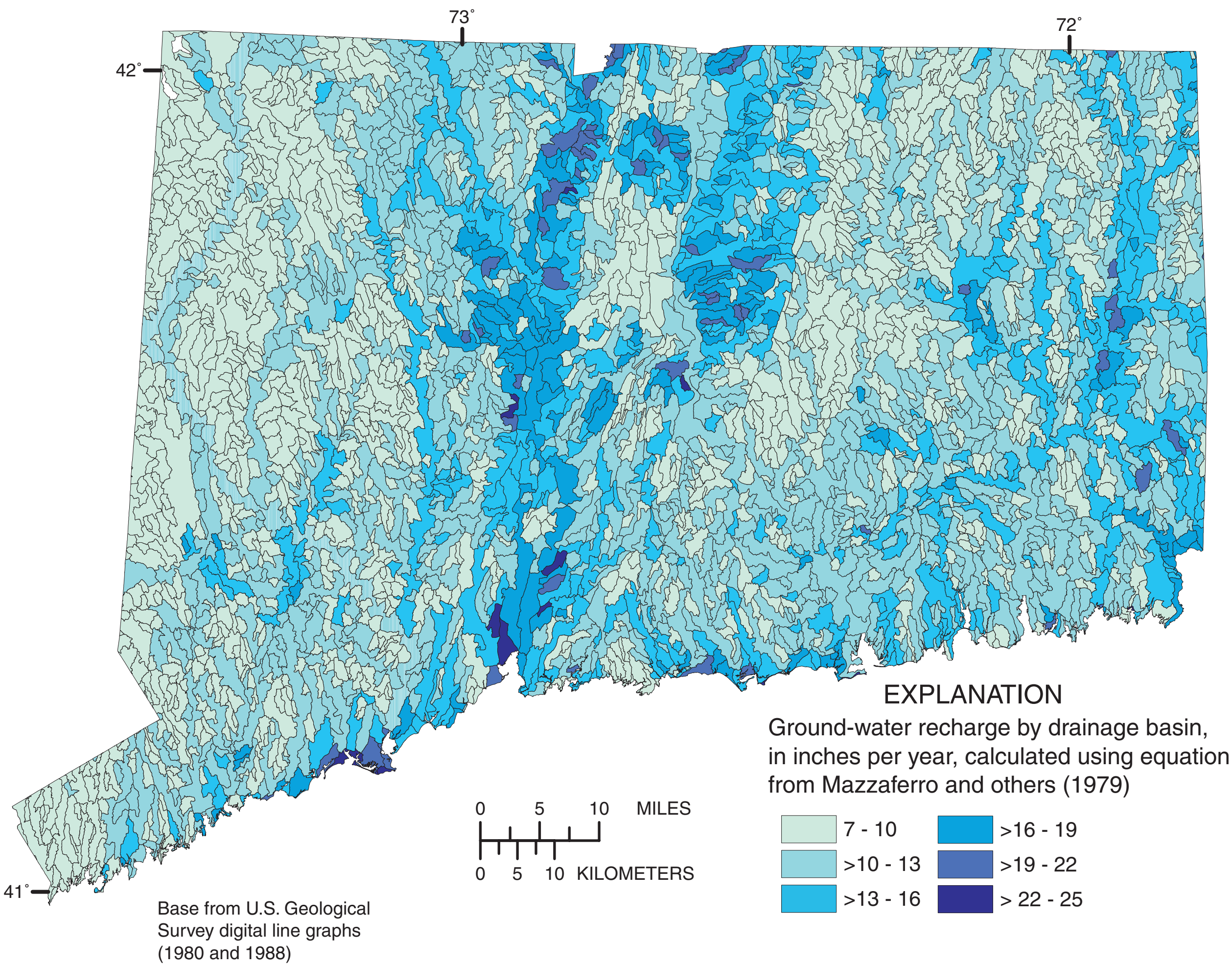


Figure 10. Ground-water recharge by drainage basin in Connecticut.

Numerical Simulation Models in Three Settings

Numerical models of ground-water flow that use bulk properties of the bedrock can provide estimates of source-water areas to bedrock wells (Shapiro, 2002). In this study, numerical models were used to simulate the steady-state source-water areas based on the geohydrologic factors previously discussed (fracture geometry, angle of dip, and surficial aquifer). Additionally, hydrologic factors (recharge and pumping rate) and a topographic factor (position) were simulated to test their effect on source-water areas. Factors related to well construction, such as the depth of penetration of the well in the aquifer and the integrity of the seal between the well bore and the surficial aquifer, also were simulated in the models. The combined factors were grouped into three settings—wells on hilltops and hillsides with no surficial aquifer, wells in a narrow valley with a surficial aquifer, and wells in a broad valley with a surficial aquifer (table 3). The first two settings represent the Eastern and (or) Western Upland physiographic regions; the third setting represents the Central Lowland. For each setting, a basic model was constructed, and various factors were assessed by comparing simulations to the basic model. The same model area was used for all models and is described below. Maps and sections of all the models tried are not presented in this report because they were too numerous, and there was much similarity among them. Only the maps and (or) sections that are significant or that illustrate an important point are presented.

An infinite number of models are possible, and not all reasonable models have been tried; however, the models developed represent different geohydrologic settings in Connecticut. The models presented here illustrate some concepts about source-water areas in fractured rock, but many more studies could be done before all important factors are understood. No calibration of these models was done, and the parameter values used are general and do not relate to a specific location. The numerical models in this study were constructed using simplified representations of a fractured-rock aquifer system to isolate the effect of single variables, to the extent possible. These models are based on model designs and parameter values used in previous studies, with an emphasis on the fractured-rock aquifer study done by the USGS at Mirror Lake, New Hampshire (Harte and Winter, 1995; Tiedeman and others, 1997; and Tiedeman and others, 1998).

The dimensions of the models were based on the characteristics of subregional surface-water drainage basins in Connecticut. Subregional basins that are completely within the state boundaries have drainage areas between 1 and 78 mi^2 and contain the major river valleys in Connecticut. The shape of the basins was characterized by computing the maximum distance from any point in the basin to the basin divide. The median value for all subregional basins is 5,000 ft, and this value is taken to represent a typical distance from a ground-water-flow divide to a ground-water discharge area in a river. This value is sufficient for the purpose of constructing a representative model; however, in general, this approach is too simplistic for a quantitative analysis of basin characteristics. The area of interest in the simulation models is the area between a ground-water basin divide and a river. To minimize the effects of the hydrologic boundaries on the simulations, the model area included two rivers and two ground-water divides over a distance of 15,000 ft. A separate set of simulations was done with a basin width of 30,000 ft to test the effects of model domain size. The results of that set of simulations were essentially identical to the results presented here. Tiedeman and others (1998) estimated the depth of active ground-water flow in the Mirror Lake study area to be about 500 ft under natural, nonstressed conditions. For lack of other data, the model area in this study also extends to a depth of 500 ft. Mirror Lake is in crystalline bedrock with high relief; however, bedrock in Connecticut is both crystalline and sedimentary and of lower relief and may have a different depth of active ground-water flow.

A moveable water-table boundary is simulated in the uppermost model layer, meaning that the saturated thickness of the upper layer is not uniform. Water is added to the model through recharge at the water table. The remaining layers are simulated as having constant thickness. The top of the model grid was estimated by running the basic model and using the elevation of the water table as the top of the model grid in subsequent simulations. The uniformity of depth to water in Connecticut supports simulating the top of the model in this way. Of the 7,493 water-level altitude records compiled for this study, the median depth to water is 20 ft below land surface (lower quartile—15 ft; upper quartile—30 ft).

Table 3. Relation of generic simulation models to geohydrologic, hydrologic, and topographic factors.

[Fracture geometry: LAY, layered; NON, nonlayered; VAR, variably layered. Angle of dip: HIGH, dip greater than 50 degrees; LOW, dip less than or equal to 50 degrees; na, angle of dip not applicable in nonlayered rocks. Surficial: SA, surficial aquifer; NSA, no surficial aquifer. Recharge: Normal, 0.002 feet per day in till and 0.005 feet per day in glacial stratified deposits; Low, 0.0002 feet per day in till and 0.005 feet per day in glacial stratified deposits. Pumping rate: Normal, 50 gallons per minute; High, 250 gallons per minute]

Model	Geohydrologic factor			Other factors			
	Fracture geometry	Angle of dip	Surficial aquifer	Recharge	Topographic position	Simulated model layers penetrated	Pumping rate
1	NON	na	NSA	Normal	Hilltop	layers 2-5	Normal
2	NON, LAY	LOW	NSA	Normal	Hilltop	layers 2-5	Normal
3	NON, LAY	LOW	NSA	Normal	Hilltop	layers 2-5	Normal
4	LAY	HIGH	NSA	Normal	Hilltop	layers 2-5	Normal
5	LAY	HIGH	NSA	Normal	Hilltop	layers 2-5	Normal
6	LAY	HIGH	NSA	Normal	Hilltop	layers 2-5	Normal
7	LAY	HIGH	NSA	Normal	Hilltop	layers 2-5	High
8	LAY	HIGH	NSA	Normal	Hilltop	layers 2-5	High
9	NON	na	NSA	Normal	Hillside	layers 2-5	Normal
10	NON, LAY	LOW	NSA	Normal	Hillside	layers 2-5	Normal
11	LAY	LOW	NSA	Normal	Hillside	layers 2-5	Normal
12	LAY	LOW	NSA	Normal	Hillside	layers 2-5	Normal
13	NON, LAY	LOW	NSA	Normal	Hillside	layers 2-5	Normal
14	NON, LAY	LOW	NSA	Normal	Hillside	layers 2-5	Normal
15	LAY	HIGH	NSA	Normal	Hillside	layers 2-5	Normal
16	LAY	HIGH	NSA	Normal	Hillside	layers 2-5	Normal
17	LAY	HIGH	NSA	Normal	Hillside	layers 2-5	Normal
18	LAY	HIGH	NSA	Normal	Hillside	layers 2-5	High
19	NON	na	SA	Normal	Valley	layers 3-10	Normal
20	NON	na	SA	Normal	Valley	layers 3-10	Normal
21	NON, LAY	LOW	SA	Normal	Valley	layers 3-10	Normal
22	NON, LAY	LOW	SA	Normal	Valley	layers 3-10	Normal
23	LAY	HIGH	SA	Normal	Valley	layers 3-10	Normal
24	LAY	HIGH	SA	Low	Valley	layers 3-10	Normal
25	LAY	HIGH	SA	Normal	Valley	layers 3-10	Normal
26	LAY	HIGH	SA	Normal	Valley	layers 3-10	Normal
27	LAY	HIGH	SA	Normal	Valley	layers 3-10	High
28	NON	na	SA	Normal	Valley	layers 3-10	Normal
29	NON	na	SA	Normal	Valley	layers 3-10	Normal
30	NON	na	SA	Low	Valley	layers 3-10	Normal
31	NON	na	SA	Normal	Valley	layers 3-10	Normal
32	NON	na	SA	Normal	Valley	layers 3-10	Normal

Table 3. Relation of generic simulation models to geohydrologic, hydrologic, and topographic factors.—Continued

[Fracture geometry: LAY, layered; NON, nonlayered; VAR, variably layered. Angle of dip: HIGH, dip greater than 50 degrees; LOW, dip less than or equal to 50 degrees; na, angle of dip not applicable in nonlayered rocks. Surficial: SA, surficial aquifer; NSA, no surficial aquifer. Recharge: Normal, 0.002 feet per day in till and 0.005 feet per day in glacial stratified deposits; Low, 0.0002 feet per day in till and 0.005 feet per day in glacial stratified deposits. Pumping rate: Normal, 50 gallons per minute; High, 250 gallons per minute]

Model	Geohydrologic factor			Other factors			
	Fracture geometry	Angle of dip	Surficial aquifer	Recharge	Topographic position	Simulated model layers penetrated	Pumping rate
33	NON	na	SA	Normal	Valley	layers 3-10	Normal
34	NON	na	SA	Normal	Valley	layers 3-10	Normal
35	LAY	HIGH	SA	Normal	Valley	layers 3-10	Normal
36	LAY	HIGH	SA	Normal	Valley	layers 3-10	Normal
37	NON	na	SA	Normal	Valley	[1]layers 3-6	Normal
38	NON	na	SA	Normal	Valley	[2]layers 2-10	Normal
39	LAY	HIGH	SA	Normal	Valley	[2]layers 2-10	Normal
40	LAY	HIGH	SA	Normal	Valley	layers 3-10	Normal
41	LAY	HIGH	SA	Normal	Valley	layers 3-10	Normal
42	LAY	HIGH	SA	Normal	Valley	layers 3-10	High
43	LAY	LOW	SA	Normal	Valley	layers 3-11	Normal
44	LAY	LOW	SA	Normal	Valley	layers 3-11	Normal
45	LAY	LOW	SA	Normal	Valley	[2]layers 2-11	Normal
46	LAY	LOW	SA	Normal	Valley	layers 3-11	Normal
47	LAY	LOW	SA	Normal	Valley	layers 3-11	Normal

[1]This model simulates a shallow well.

[2]This model simulates a poor seal between bedrock and the surficial aquifer by simulating the open well bore as present in layer 2.

Bedrock Wells on Hilltops and Hillsides with No Surficial Aquifer

Upland areas with no surficial aquifer (primarily hilltops and hillsides) are underlain primarily by metamorphic rock with thin soils developed on glacial till. The water table can be in till, but commonly is in the bedrock. The metamorphic rock can be layered or not; where layered, the rock can have shallow or steep angles of dip. This setting is found primarily in ground-water recharge areas on hilltops and hillsides and in some discharge areas in narrow valleys.

Model Design

The model grid and boundary conditions were designed to simulate source-water areas to a well on a hilltop and on a hillside within the model area (table 4); both well locations are outside the area of the simulated glacial stratified deposits (fig. 11). A uniformly spaced model grid with 200-by-200-ft model cells was used, with five layers of different thicknesses. In this geohydrologic setting, rivers are simulated as drains. If the simulated water level falls below the drain, no more water is removed, and, more importantly, no water is added to the model. In the real world, rivers can be sources of water to pumped wells through induced infiltration. In these simulations, induced infiltration was not simulated so that the effects of fractures and topography could be isolated.

Hydraulic properties were assigned to the model grid according to the type of geologic material represented (table 4). All model cells represented bedrock, a surficial aquifer, or in some cases, a fracture zone or till. The surficial aquifer was assumed to be present in the top two layers in the valley bottom; bedrock was assumed to be present in all other areas and layers unless otherwise noted. The hydraulic properties from the Mirror Lake study (Harte and Winter, 1995; Tiedeman and others, 1997; and Tiedeman and others, 1998) were primarily used in this study because they result from analysis of one of the most thorough and consistent sets of data available on ground-water flow in fractures. Hydraulic properties from these and other studies (table 5) were used as guidelines for reasonable ranges within which to vary properties.

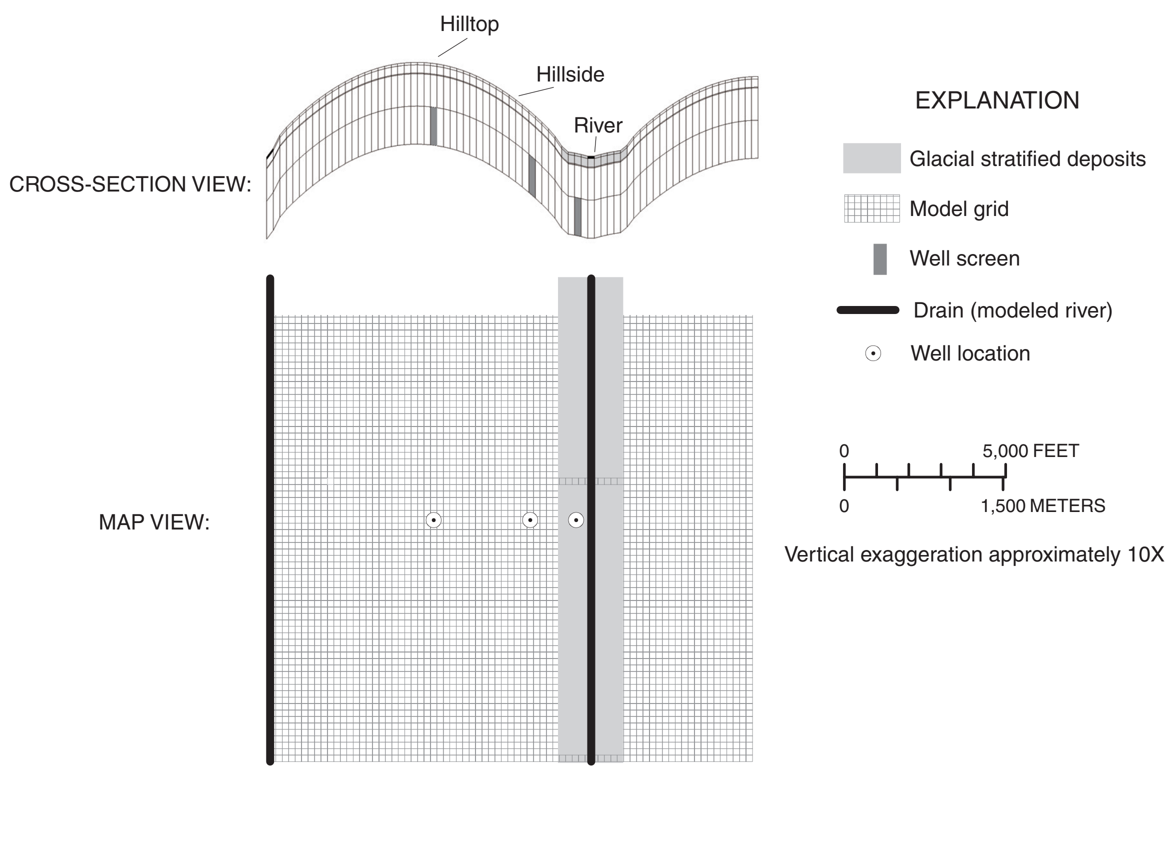


Figure 11. Finite-difference grid and boundary conditions for the basic simulation model for wells on hilltops and hillsides with no surficial aquifer.

Table 4. Model grid, boundary conditions, and hydraulic properties of the basic simulation model for wells on hilltops and hillsides with no surficial aquifer.

[ft, feet; ft/d, feet per day; in/yr, inches per year; ft^3/d, cubic feet per day; gal/min, gallons per minute]

Model grid	
Grid	200-by-200-ft model cells in a 15,000-by-15,000-ft grid
Layers	Five layers, from top to bottom, 15, 50, 5, 197.5, and 232.5 ft thick. Total thickness 500 ft.
Boundary conditions	
Base and lateral boundaries of model	No-flow boundaries surround the model domain
Rivers	Rivers were simulated as drains. The bottom elevation of the drain was set to 0 ft and the conductance of the drain as set to 100,000 ft/d. The high conductance allowed water to flow freely into the drain, and no water could flow from the river to the aquifer.
Recharge	Recharge was applied to the top layer of the model at the rate of 0.002 ft/d (about 9 in/yr) to upland areas and 0.005 ft/d (about 22 in/yr) to glacial stratified deposits.
Well	A well was simulated in layer 5 pumping at about 50 gal/min (10,000 ft^3/d). The cells in layers 2 through 5 in which the well was simulated were assigned a high vertical hydraulic conductivity (10,000 ft/d) to approximate the effects of an open borehole.
Hydraulic properties	
Horizontal hydraulic conductivity	14 ft/d, coarse-grained glacial stratified deposits; 0.08 ft/d, bedrock; 17 ft/d, fracture. Hydraulic conductivity was isotropic horizontally unless otherwise noted.
Vertical hydraulic conductivity	Vertical hydraulic conductivity was equal to horizontal hydraulic conductivity unless otherwise noted.

Table 5. Hydraulic and hydrologic properties used to simulate ground-water flow in upland areas in the northeastern United States.

K, hydraulic conductivity; subscript of K indicates grid direction where x is along rows, y is along columns, and z is along layers; --, not applicable or not determined]

Material	Values from previous studies							
	Tiedeman and others (1997)	Barton and others (1999)	Lyford and others (1999)	Mack and Dudley (2001)	Lyford and others (2003)	Lyford and others (2003)	Lyford and others (2003)	Mack (2003)
Hydraulic conductivity of glacial deposits, in feet per day								
Coarse	14	--	30	--	--	10	--	10-150
Fine	0.0003	--	0.001 - 0.01	--	--	--	--	--
Till	.5 - .76	--	0.01 - 0.5	1	--	--	0.2	1
Conductive till	--	--	1	10	--	--	--	--
Hydraulic conductivity of bedrock, in feet per day								
Undifferentiated material	.083	--	--	.1	--	.4	.2 - .5	.72-1.5
K_x	--	--	--	1.0	--	--	--	--
K_y	--	--	--	.1	--	--	--	--
K_z	--	--	--	--	0.45 - 9.0	.3 - .45	~.001	--
Valley	.09	--	--	--	--	--	--	--
Hilltop	.02	--	--	--	--	--	--	--
Shallow fractured	--	8	.5 - 1.5	1	5.0 - 10.0	--	--	--
Shallow unfractured	--	1	.003	--	.4	--	--	--
Deep fractured	--	8	40 - 120	--	2.5 - 5	14	1	--
Deep unfractured	--	.1	.24	--	.2	--	.02	--
Fracture	17	--	--	1	--	--	--	--
Ground-water recharge, in inches per year								
Areawide	11	9	--	13	--	24	--	19
Till	--	--	10	--	15	--	4.4	--
Coarse	--	--	20	--	--	--	--	--
Fine	--	--	.5	--	.04	--	--	--
Thick till	--	--	--	--	3	--	--	--

Simulated Factors

Factors related to wells on hilltops and hillsides with no surficial aquifer were simulated using variations on the basic simulation model (table 6). In all simulations, the pumped well was assumed to be at the center of the fracture zone. Fracture geometry was based on descriptions of fractures by Tiedeman and others (1998) and by Moore and others (2002). Tiedeman and others (1998) reported that, at Mirror Lake, fractures of various orientations are in near-horizontal zones about 5 ft thick and 150 ft in horizontal extent. They also report that the hydraulic conductivity of the fracture zones is about 17 ft/d; that value is used in this study in all simulated fractures. Their near-horizontal zones were simulated in this study using a fracture hydraulic-conductivity zone that was parallel to land surface, 600-by-600 ft (3 model cells by 3 model cells) and 5 ft thick. This large fracture zone was used here because of the coarseness of the model grid (200-by-200 ft); a fracture zone 150-by-150 ft would only occupy one model cell and would not affect the flow system greatly. Models also were run with a fracture hydraulic-conductivity zone that was parallel to land surface throughout layer 3 (termed the "extensive subhorizontal fracture"). Moore and others (2002) reported that high well yields are correlated with near-vertical fracture zones. Models were run that had a vertical fracture zone 1,000 ft long parallel to the river, 200 ft wide (one model cell), and 485 ft thick, extending from the bottom of the upper layer to the base of the model. Models also were run with an extensive vertical fracture 15,000 ft long, 200 ft wide, and 485 ft thick.

Other variations of the models were run based on comments received during various reviews of this study. Additional simulations include a higher pumping rate, fractures that did not intersect the well, lower recharge rates, vertical and horizontal anisotropy, and a loose surface till in the uplands (in which the hydraulic conductivity of the top model layer was equal to that of a surficial aquifer). No attempt was made to simulate all combinations of factors.

Table 6. Simulation models of wells on hilltops and hillsides with no surficial aquifer.

[gal/min, gallons per minute; ft, feet; K_h, horizontal hydraulic conductivity; K_v, vertical hydraulic conductivity]

Model	Characteristics
1	Well on hilltop basic model
2	Well on hilltop and subhorizontal fracture
3	Well on hilltop and extensive subhorizontal fracture
4	Well on hilltop and vertical fracture
5	Well on hilltop and extensive vertical fracture
6	Well on hilltop and extensive vertical fracture not intersecting the well
7	Well on hilltop and extensive vertical fracture not intersecting the well and a higher pumping rate (250 gal/min)
8	Well on hilltop and two extensive vertical fractures not intersecting the well and a higher pumping rate (250 gal/min)
9	Well on hillside basic model
10	Well on hillside and subhorizontal fracture
11	Well on hillside and subhorizontal fracture, K_h/K_v = 100
12	Well on hillside and subhorizontal fracture, K_h/K_v = 0.01
13	Well on hillside and larger (1,000-by-1,000 ft) subhorizontal fracture
14	Well on hillside and extensive subhorizontal fracture
15	Well on hillside and vertical fracture
16	Well on hillside and vertical fracture and highly permeable top layer
17	Well on hillside and extensive vertical fracture
18	Well on hillside and extensive vertical fracture and higher pump rate (250 gal/min)

Effect of Simulated Factors on Source-Water Areas

Source-water areas to wells reflect the general ground-water-flow direction (fig. 12). The source-water area to the hilltop well was not greatly affected by simulated fracture zones, except for the extensive vertical fracture zone (model 5 in table 6). The flow paths to wells were affected by the subhorizontal fracture zone (ground water was collected by the fracture zone and did not pass below it); however, the surface expression of the source-water area was not affected (fig. 13). Interestingly, the fracture zones that did not intersect the well seemed to have the strongest effect on the source-water area (fig. 14). At a low pumping rate, the fracture zone seems to be a barrier to flow; this is because the regional flow system makes use of the fracture to transmit water down the slope, and the pumped well is not strong enough to overcome the regional flow path. At the higher pumping rate, the well is strong enough that the fracture zone is not a barrier to flow. The simulation with two fracture zones shows how complicated the real situation could be (fig. 15).

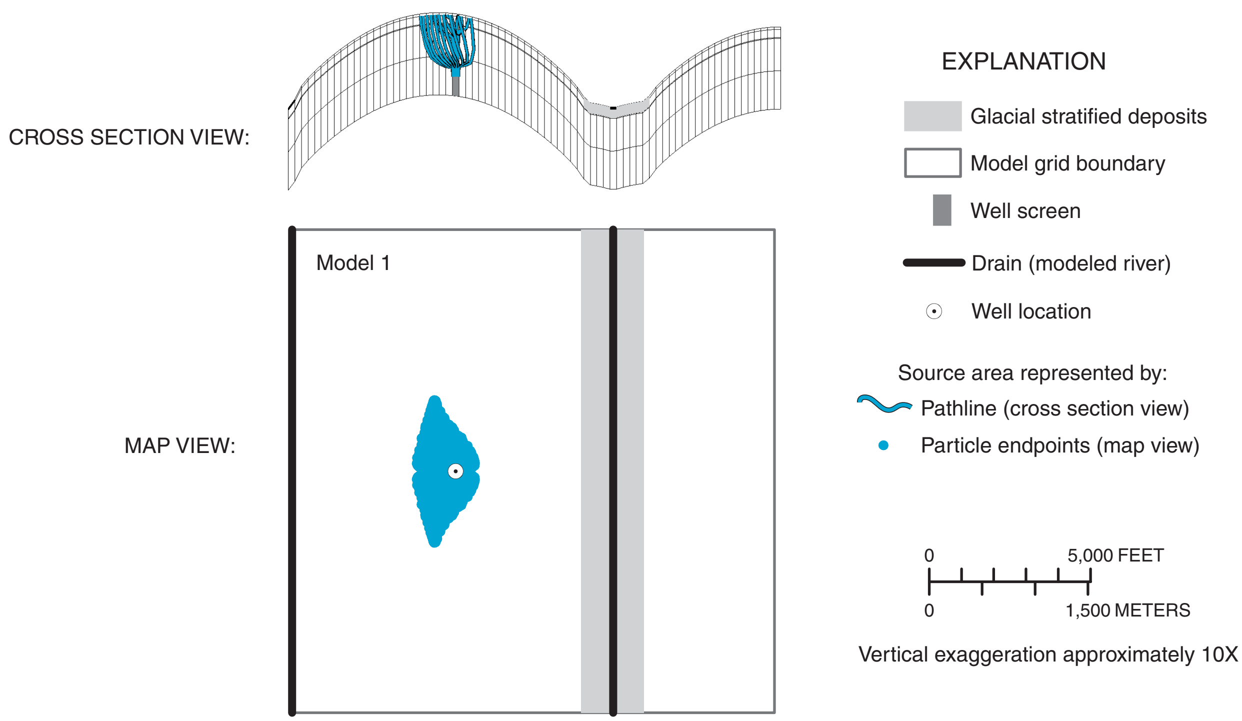


Figure 12. Source-water areas to a hilltop well with no surficial aquifer.

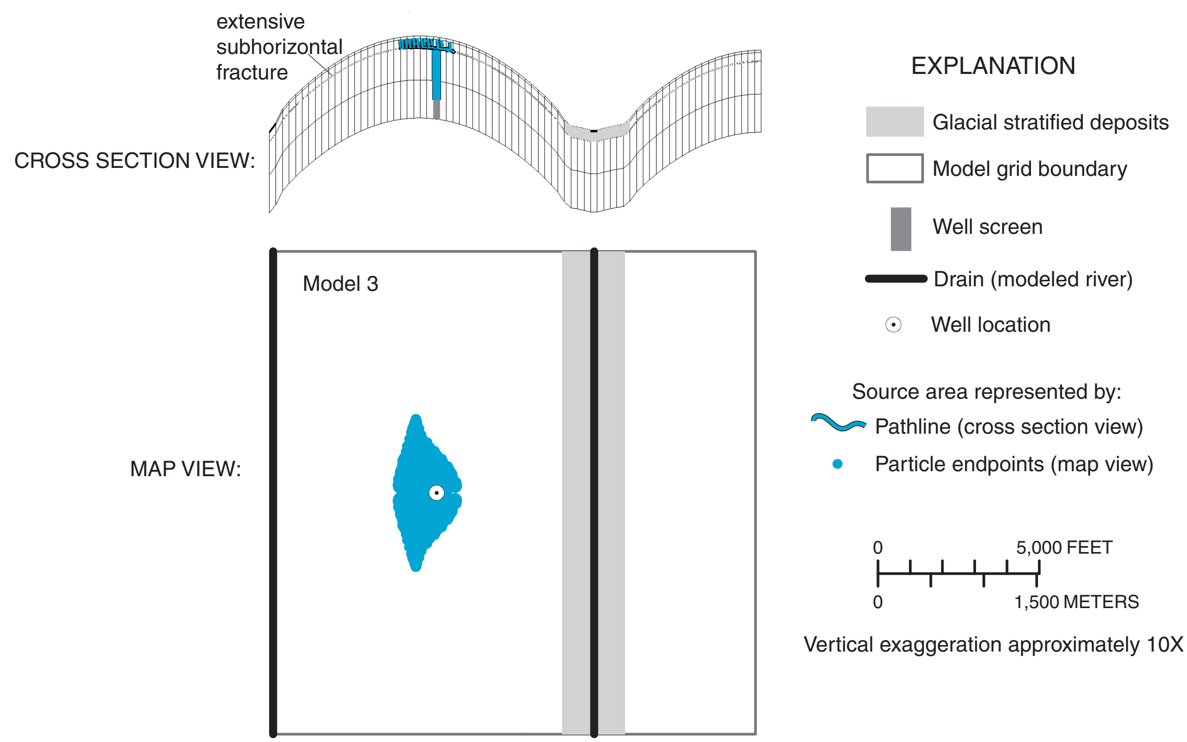


Figure 13. Source-water areas to a hilltop well with no surficial aquifer and an extensive subhorizontal fracture.

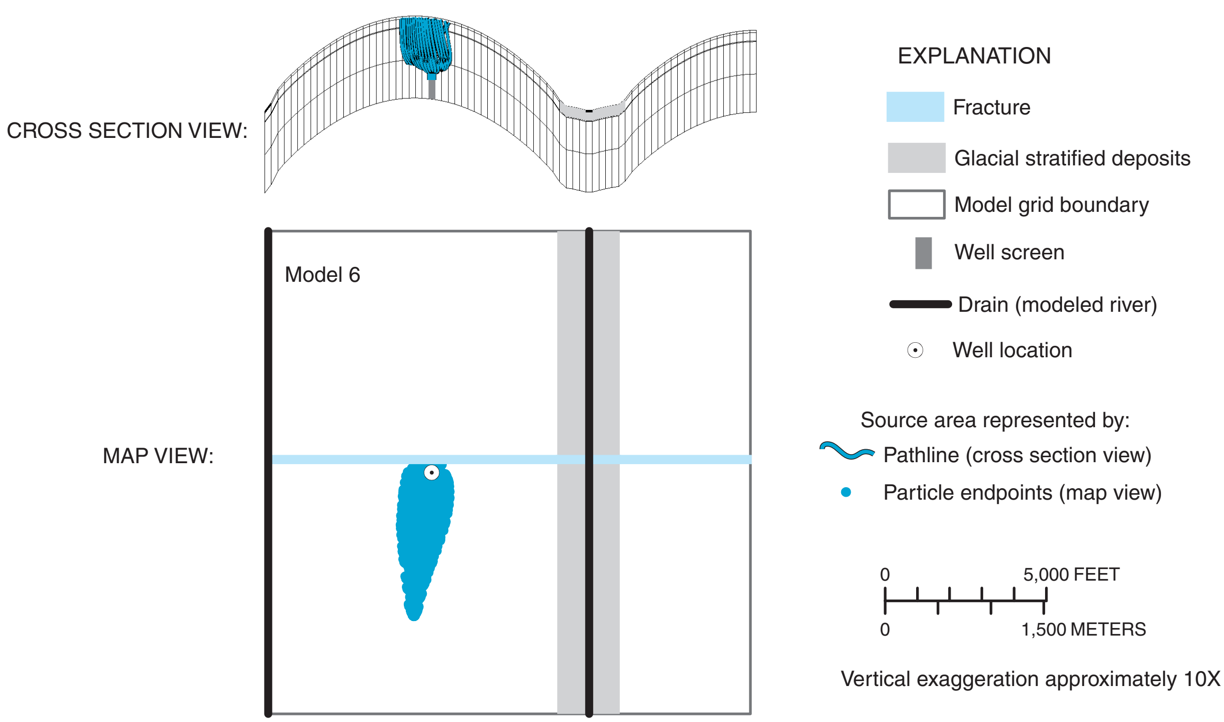


Figure 14. Source-water areas to a hilltop well near a fracture not intersected by the well.

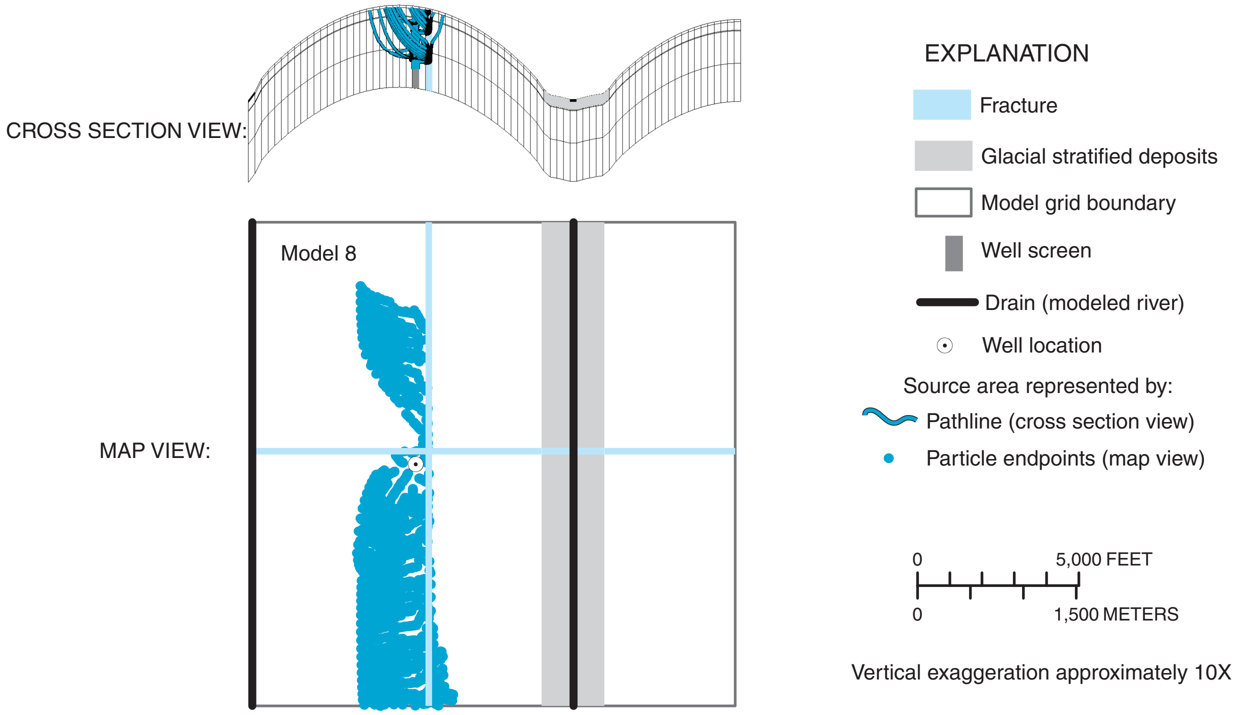


Figure 15. Source-water areas to a hilltop well with no surficial aquifer, two fractures, and a higher pumping rate (250 gallons per minute).

The source-water area to the hillside well (fig. 16) was not greatly affected by the fracture zones (with one exception); even the extensive vertical fracture zone was not as influential as it was on the hilltop source-water area. The one exception is the combination of a subhorizontal fracture zone and low bedrock vertical hydraulic conductivity, as might be the case where an extensive subhorizontal fracture zone was not connected to the surface through vertical fractures. In this case, the source-water area is long and narrow and extends upgradient past the ground-water-flow divide (fig. 17). Part of the source-water area is elongated parallel to the river, but the simulated travel times from this area are many orders of magnitude greater than for the rest of the source-water area.

Although these simulations show that topography could be used to delineate source-water areas in simplified settings, the real world is not so simple. These simulations are intended to estimate source-water areas for inventorying potential contaminants that might affect the quality of the source water if they were released to the environment. The source-water areas are similar for many of the simulations, but the travel times, which are not reflected in the source-water areas, can vary greatly. In general, ground-water travel times in fractures can be short because of the small volume of aquifer (the fracture) through which the water must flow.

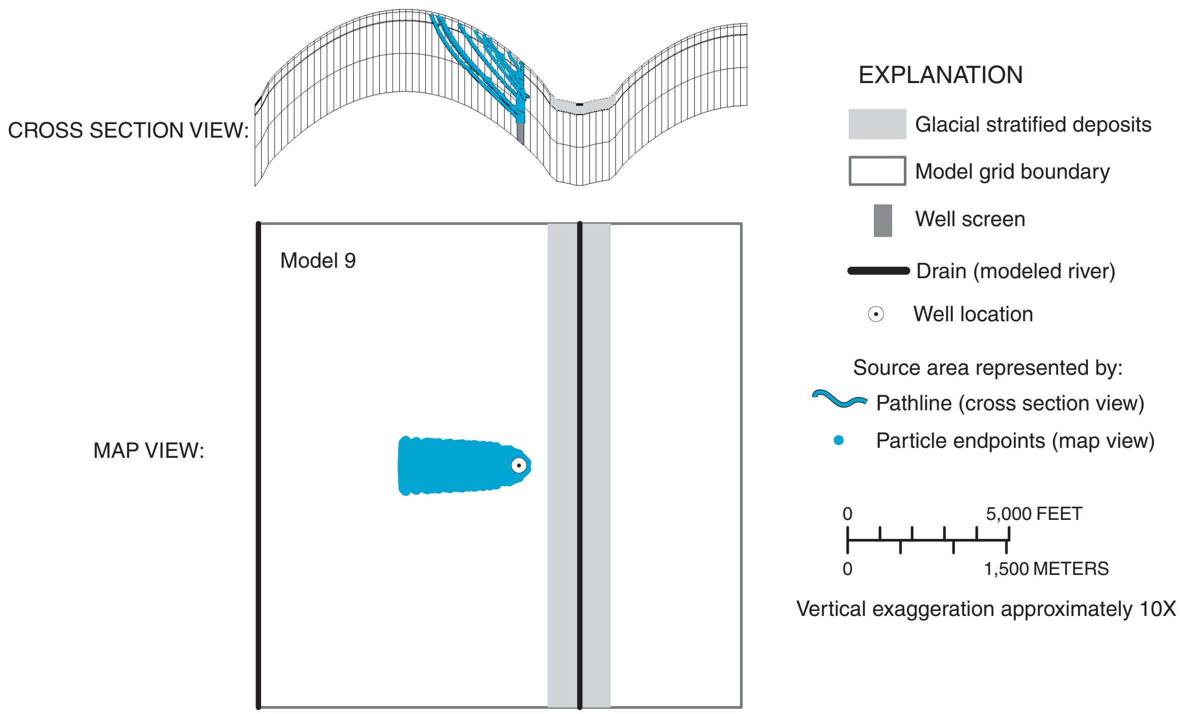


Figure 16. Source-water areas to a hillside well with no surficial aquifer.

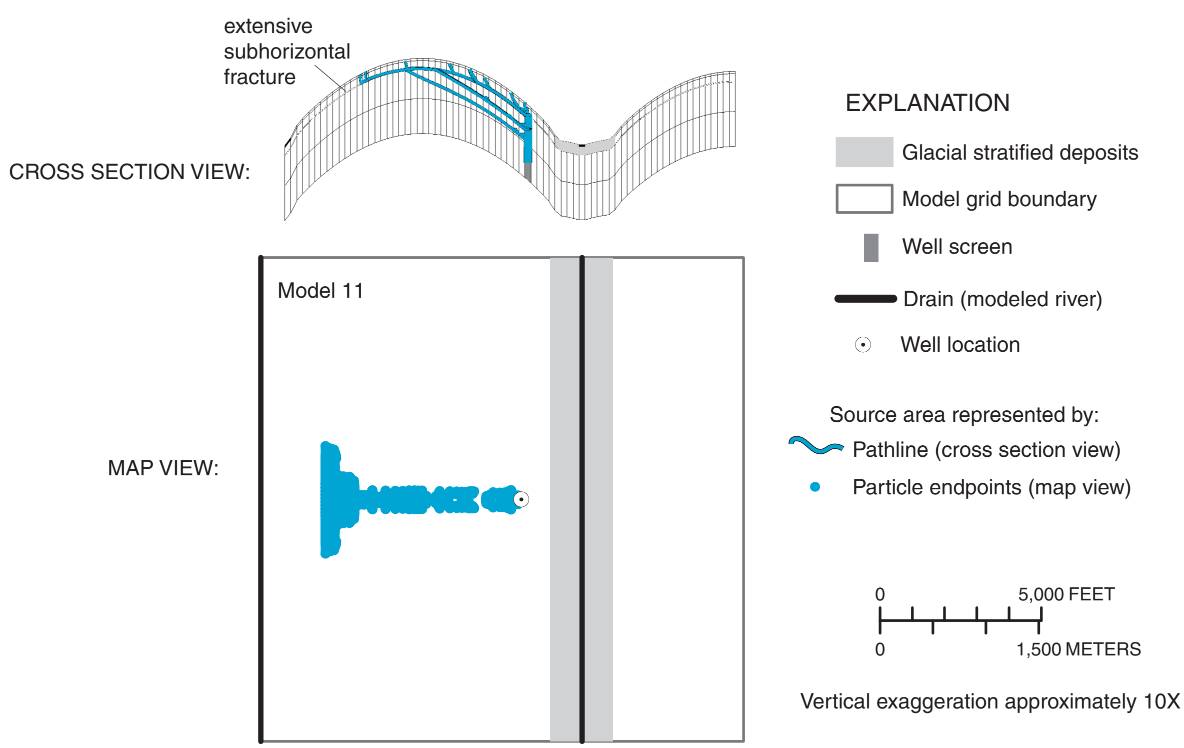


Figure 17. Source-water areas to a hillside well with no surficial aquifer and an extensive subhorizontal fracture.

Bedrock Wells in a Narrow Valley with a Surficial Aquifer

Upland areas with a surficial aquifer in narrow valleys are underlain primarily by metamorphic rock with soils developed on fine- or coarse-grained glacial stratified deposits. The water table can be in bedrock, but commonly is in the surficial aquifer. The metamorphic rock can be layered or not; where layered, the rock can have shallow or steep angles of dip.

Model Design

Two sets of simulations were done for wells in narrow valleys with a surficial aquifer. The first set (models 19 to 27, table 7) was similar to those of the previous section (table 4), except that the well was in a valley rather than in the uplands. The second set (models 28 to 42, table 7) used a variably spaced grid to simulate the high hydraulic gradients near the well (table 8; fig. 18). The model cell containing the well was 2-ft-by-2-ft, and the grid was expanded outward from the well by a factor of approximately 1.5. Except for the top two layers, which were identical in dimension to those previously used (table 4), a uniform thickness of 54.375 ft was used in each of the remaining eight layers (fig. 18). The area of the model domain was slightly larger than the previous models because of the variable grid spacing. The hydraulic properties were identical to those used in the previous models. In the second set of simulations, the rivers were simulated as constant-head boundaries. This type of boundary, in contrast to the drain boundary used previously, allowed water to move from the river to the aquifer. In the real world, rivers can be sources of water to pumped wells through induced infiltration.

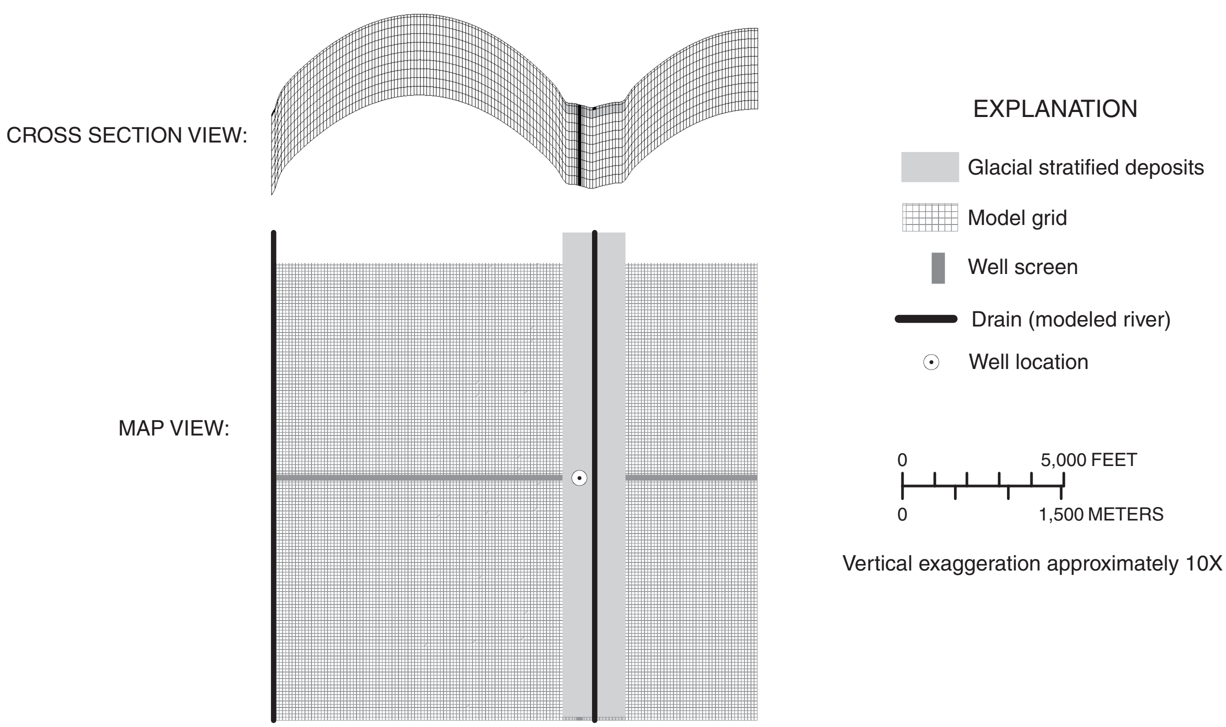


Figure 18. Finite-difference grid and boundary conditions for the basic simulation model for a well in a narrow valley with a surficial aquifer.

Simulated Factors

Simulations with fracture-zone orientations similar to models 1 to 18 were run, except that the simulated well was in the valley bottom in the area of the surficial aquifer. One additional simulation was run to check the effects of grid-cell size and layering. In this model, the grid-cell size was reduced to 50-ft-by-50-ft and the layer thickness was reduced to about 50 ft. Harte and Winter (1995) found that natural recharge (under non-pumping conditions) from till in uplands to bedrock is commonly between 1 and 3 in/yr. Higher rates were simulated in this study because the till was included in the simulation. To see the effects of a lower recharge rate, a simulation was run with recharge to uplands equal to one-tenth the normal rate, or about 0.9 in/yr. Other simulations with a higher pumping rate and a highly permeable upper layer, which represents a highly permeable glacial till, also were run.

Simulations 28 to 42 used a variable grid spacing (table 7) and a constant-head boundary to represent rivers. One simulation was run using a drain boundary to represent rivers to test the effect of boundary conditions on the source-water area. The purpose of the rest of these simulations was to test the effects of the surficial aquifer on source-water areas. Variations on the basic model include lower recharge, highly permeable upper layer, and higher well pumpage, as in previous simulations. Additional variations of surficial aquifer properties include a wider surficial aquifer (4,000 ft wide rather than 2,000 ft wide), a thinner surficial aquifer (present only in the uppermost layer and thus 15 ft thick rather than 65 ft thick), and a surficial aquifer of lower permeability (hydraulic conductivity of the surficial aquifer equal to that of bedrock).

Additional variations of bedrock properties include a horizontal anisotropy where the hydraulic conductivity parallel to the river is 100 times that perpendicular to the river, and the inverse, where the hydraulic conductivity parallel to the river is 0.01 times that parallel to the river. These models simulate the condition where there are enough fractures to impart a bulk property to the rock that is directionally dependent. Additional models were used to simulate variations in well construction, including a partially penetrating well and a well in which the open borehole was well-connected to the surficial aquifer. The latter situation was simulated by including the high vertical conductivity that simulates the open well bore as being present in layer 2 (in other simulations, the well bore is simulated in layers 3 through 11 only, thus creating a separation distance between the surficial aquifer and the open borehole in bedrock). The effect of an extensive vertical fracture zone also was simulated in the upland source-water area to a well in the valley, in the valley itself, with a higher pumping rate, and in the valley with a good connection between the open borehole and the surficial aquifer.

Table 7. Simulation models of wells in a narrow valley with a surficial aquifer.

[ft, feet; ft/d, feet per day; gal/min, gallons per minute; K, hydraulic conductivity; subscript of K indicates grid direction where x is along rows and y is along columns]

Model	Characteristics
19	Well in narrow valley basic model
20	Well in narrow valley basic model with model grid cells reduced to 50-by-50 ft and about 50 ft thick
21	Well in narrow valley and subhorizontal fracture
22	Well in narrow valley and extensive subhorizontal fracture
23	Well in narrow valley and vertical fracture
24	Well in narrow valley and vertical fracture and lower recharge (recharge to uplands = 0.0002 ft/d)
25	Well in narrow valley and vertical fracture and highly permeable top layer
26	Well in narrow valley and extensive vertical fracture
27	Well in narrow valley and extensive vertical fracture and higher pump rate (250 gal/min)
28	Well in narrow valley, variable grid, basic model
29	Well in narrow valley, variable grid, with river constant heads replaced with drains
30	Well in narrow valley, variable grid, and lower recharge (recharge to uplands = 0.0002 ft/d)
31	Well in narrow valley, variable grid, and highly permeable top layer
32	Well in narrow valley, variable grid, surficial aquifer 4,000 ft wide

Table 7. Simulation models of wells in a narrow valley with a surficial aquifer.—Continued

[ft, feet; ft/d, feet per day; gal/min, gallons per minute; K, hydraulic conductivity; subscript of K indicates grid direction where x is along rows and y is along columns]

Model	Characteristics
33	Well in narrow valley, variable grid, surficial aquifer 15 ft thick
34	Well in narrow valley, variable grid, low permeability surficial aquifer
35	Well in narrow valley, variable grid, $K_y = 100 * K_x$
36	Well in narrow valley, variable grid, $K_y = 0.01 * K_x$
37	Well in narrow valley, variable grid, partially penetrating well (well in layers 3 through 6)
38	Well in narrow valley, variable grid, no seal between bedrock and surficial aquifer
39	Well in narrow valley, variable grid, fracture in valley and no seal between bedrock and surficial aquifer
40	Well in narrow valley, variable grid, fracture in uplands
41	Well in narrow valley, variable grid, fracture in valley
42	Well in narrow valley, variable grid, fracture in valley, and higher pump rate (250 gal/min)

Table 8. Model grid, boundary conditions, and hydraulic properties of the variable-grid simulation model for wells in a narrow valley with a surficial aquifer.

[ft, feet; ft/d, feet per day; in/yr, inches per year; ft^3/d, cubic feet per day; gal/min, gallons per minute]

Model geometry	
Grid	Variably spaced; 15,042 ft along columns and 15,142 ft along rows. The grid cell in which the well was located was 2-by-2 ft. The grid was expanded from this cell by a factor of about 1.5 up to a maximum cell size of 100-by-100 ft.
Layers	Ten layers, from top to bottom, 15, 50, and the lower eight layers were each 54.375 ft thick. Total thickness 500 ft.
Boundary conditions	
Base and lateral boundaries of model	No-flow boundaries surround the model domain.
Rivers	Rivers were simulated as constant-head boundaries that would allow an unlimited supply of water to be induced to flow from the river.
Recharge	Recharge was applied to the top layer of the model at the rate of 0.002 ft/d (about 9 in/yr) to upland areas and 0.005 ft/d (about 22 in/yr) to glacial stratified deposits.
Well	A well pumping at about 50 gal/min (10,000 ft^3/d) was simulated as being distributed in layer 3 through 10. The cell in layers 3 through 10 in which the well was simulated was assigned a high vertical hydraulic conductivity (10,000 ft/d) to approximate the effects of an open borehole.
Hydraulic properties	
Horizontal hydraulic conductivity	14 ft/d, coarse-grained glacial stratified deposits; 0.08 ft/d, bedrock; 17 ft/d, fracture. Hydraulic conductivity was isotropic horizontally unless otherwise noted.
Vertical hydraulic conductivity	Vertical hydraulic conductivity was equal to horizontal hydraulic conductivity unless otherwise noted.

Effect of Simulated Factors on Source-Water Areas

The source-water areas lie primarily in the till-covered hilltop and hillside (fig. 19), with less area in the surficial aquifer; however, 2.5 times the amount of water per unit area recharges the glacial stratified deposits as till (2.5 is the ratio of recharge rates, 0.005/0.002 ft/d) (Mazzaferro and others, 1979). For simulations with the 200-ft-by-200-ft grid, most source-water areas are very similar with one important exception. Refining the grid does not seem to change the source-water area.

Ground-water flow paths in this geohydrologic setting are complex (fig. 19). The typical flow path originates in the uplands and passes through either the till or bedrock into the surficial aquifer. Only a small area of the surficial aquifer actually contributes water to the well. Most flow paths pass through the surficial aquifer, although they do not originate there, and pass vertically downward from the surficial aquifer into the well.

The simulation with the lower recharge rate has a very different source-water area as compared to the other simulations in this setting (fig. 20). The size of the source-water area was expected to be larger because the size of the area is proportional to the recharge rate and the pumping rate, which was constant. As expected, with the lower recharge rate, the size of the source-water area increased in both the glacial stratified deposits and in the upland areas, and the source-water area includes substantial areas on both sides of the river. With the lower recharge rate, however, a larger percentage of the water came from the glacial stratified deposits than from the uplands; with the higher recharge rate, most of the water came from the uplands. This result underscores the need to have better information on rates of ground-water flow between bedrock and glacial deposits.

Simulations from models with a variably spaced grid (fig. 21) cannot be compared directly with those using the 200-ft-by-200-ft grid (uniformly spaced grid) because particles were tracked from a smaller point in the flow system and because there is more detail in the system represented by the finer grid spacing near the well. The source-water areas from the variably spaced grid show, in general, the same characteristics as the simulation with the uniformly spaced grid (figs. 20 and 21).

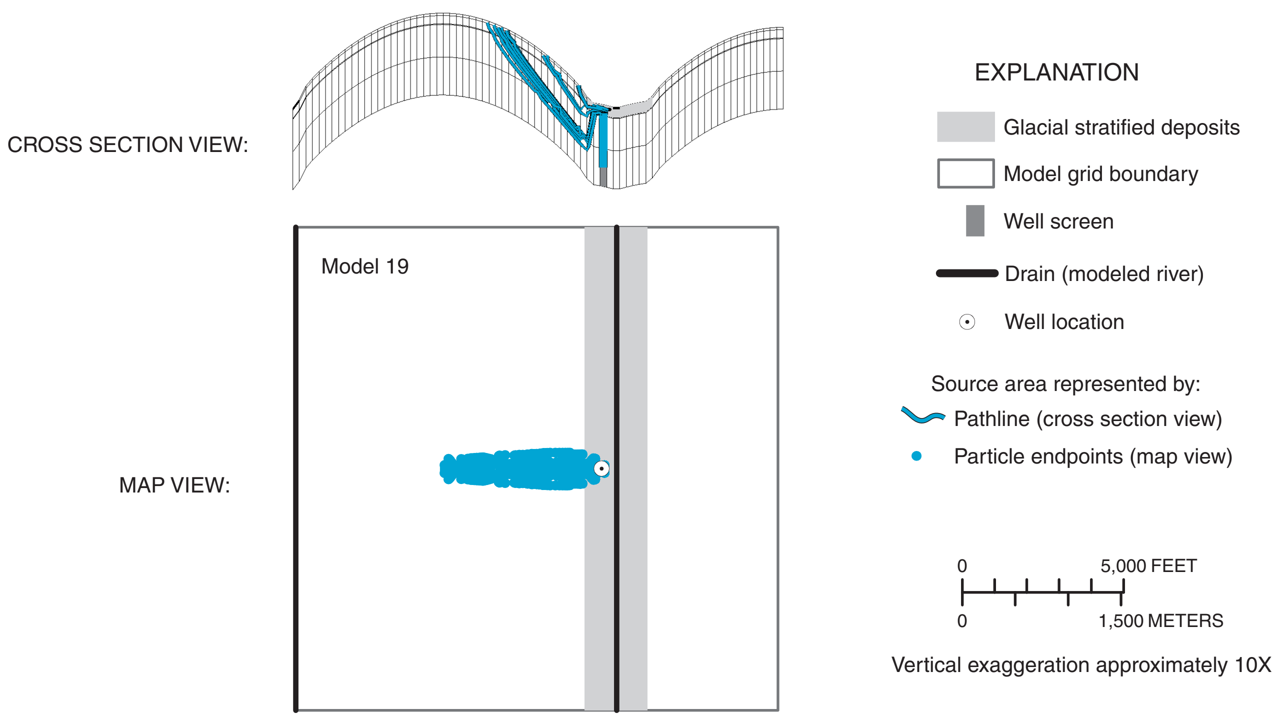


Figure 19. Source-water areas to a well in a narrow valley with a surficial aquifer.

The simulations in this setting are very similar to one another, except for the simulation of anisotropy (fig. 22). This model produced a source-water area that was elongated in the direction of the principal hydraulic conductivity (parallel to the valley). The main differences among other simulations were the width of the source-water area and the amount of the source-water area in the surficial aquifer. Conditions that led to narrow source-water areas were those with thin surficial aquifer, no surficial aquifer, greater hydraulic conductivity perpendicular to the river than parallel to it, and good connection between the surficial aquifer and the bedrock. Conditions that led to the largest source-water area in the surficial aquifer were low recharge, wide surficial aquifer, thin surficial aquifer, no surficial aquifer, and greater hydraulic conductivity perpendicular to the river than parallel to it.

The small grid-cell size near the pumped well allowed refined drawdown levels to be calculated using the models. Drawdown in the surficial aquifer (model layer 2) was generally less than 10 ft. Drawdown was similar in pattern but greater in magnitude in the bedrock, with the maximum drawdown in the pumped well at about 290 ft. In comparison, the model in which anisotropic bedrock was simulated produced an elliptical pattern of drawdown beneath the surficial aquifer that is typical of anisotropic bedrock; however, the source-water area was not located in the surficial aquifer, but was elongated in the direction of greatest hydraulic conductivity (fig. 22).

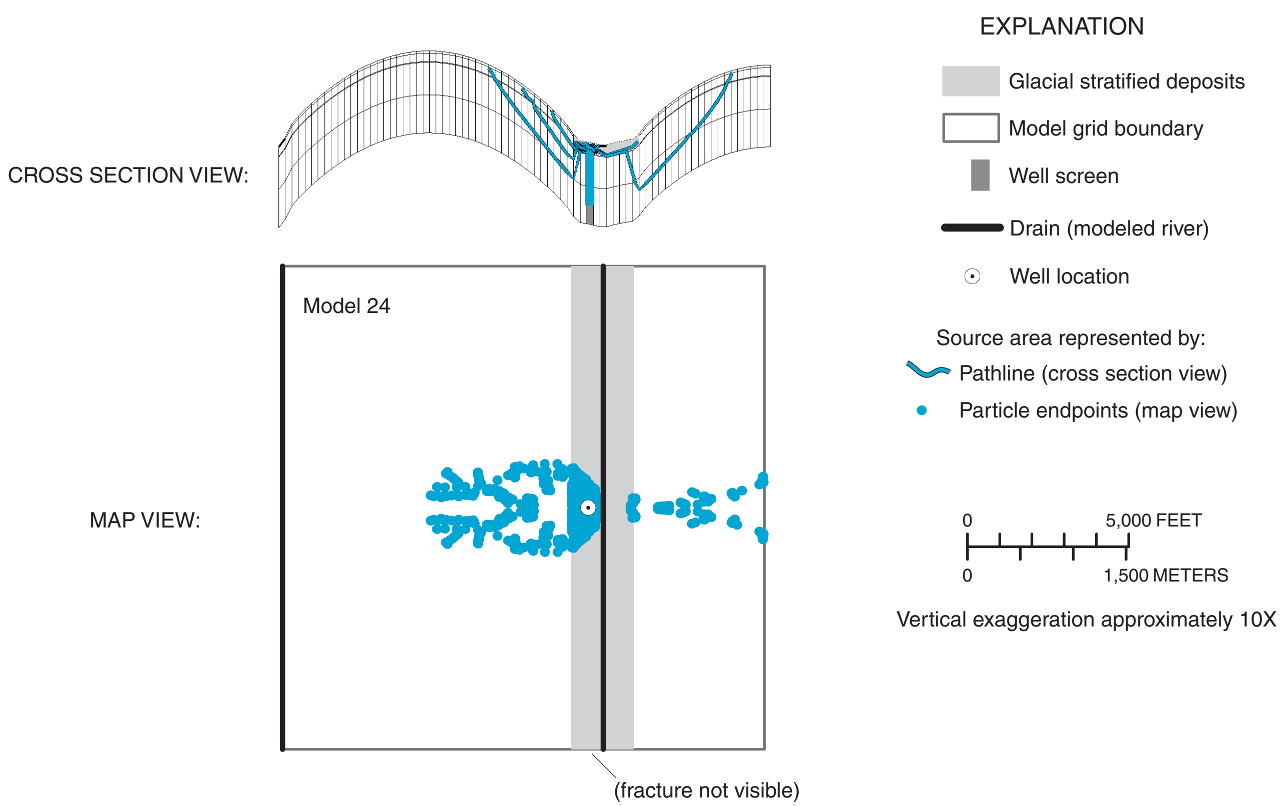


Figure 20. Source-water areas to a well in a narrow valley with a surficial aquifer and low recharge with a uniformly spaced grid.

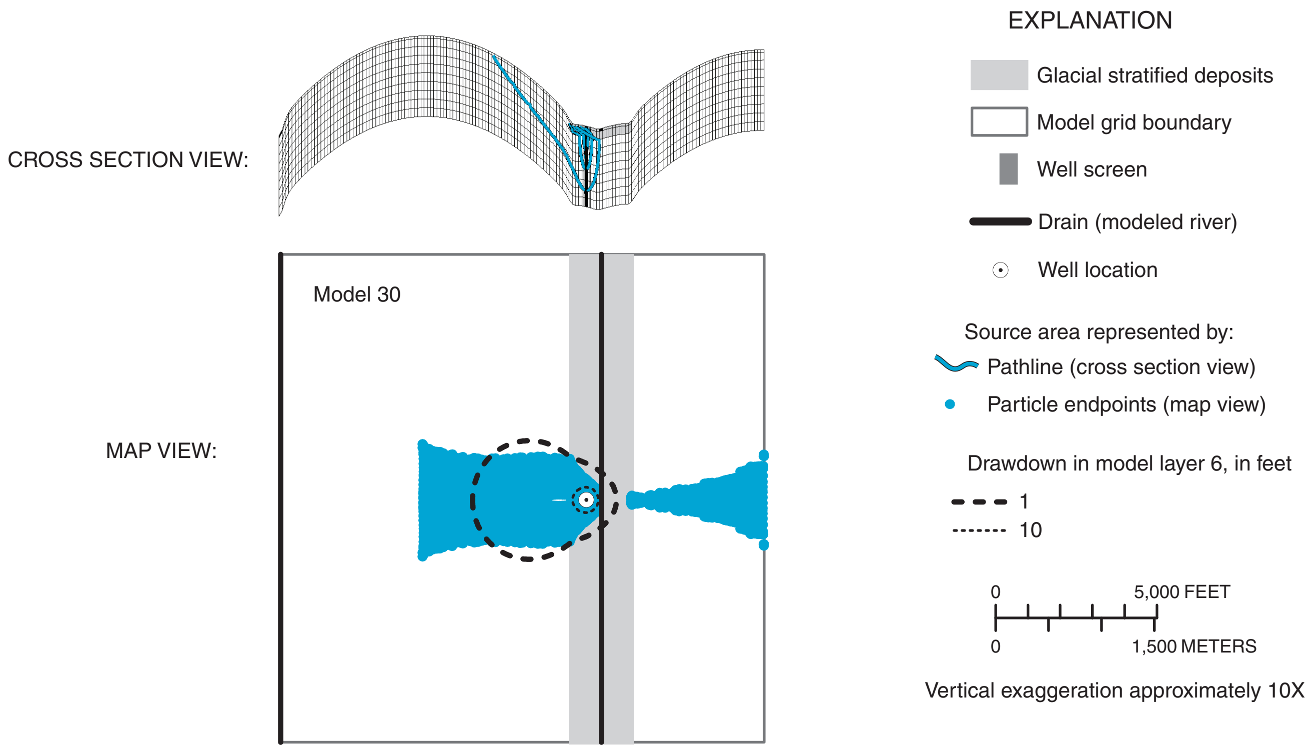


Figure 21. Source-water areas to a well and drawdown in a narrow valley with a surficial aquifer and low recharge, with a variably spaced grid. [Note: Fewer number of particles were used to generate the pathlines shown in the section view than were used to generate the end points in the map view, because using the same number particles in both views would cause there to be too many pathlines to be distinguishable. As a result, no pathlines are shown on the right side of the river.]

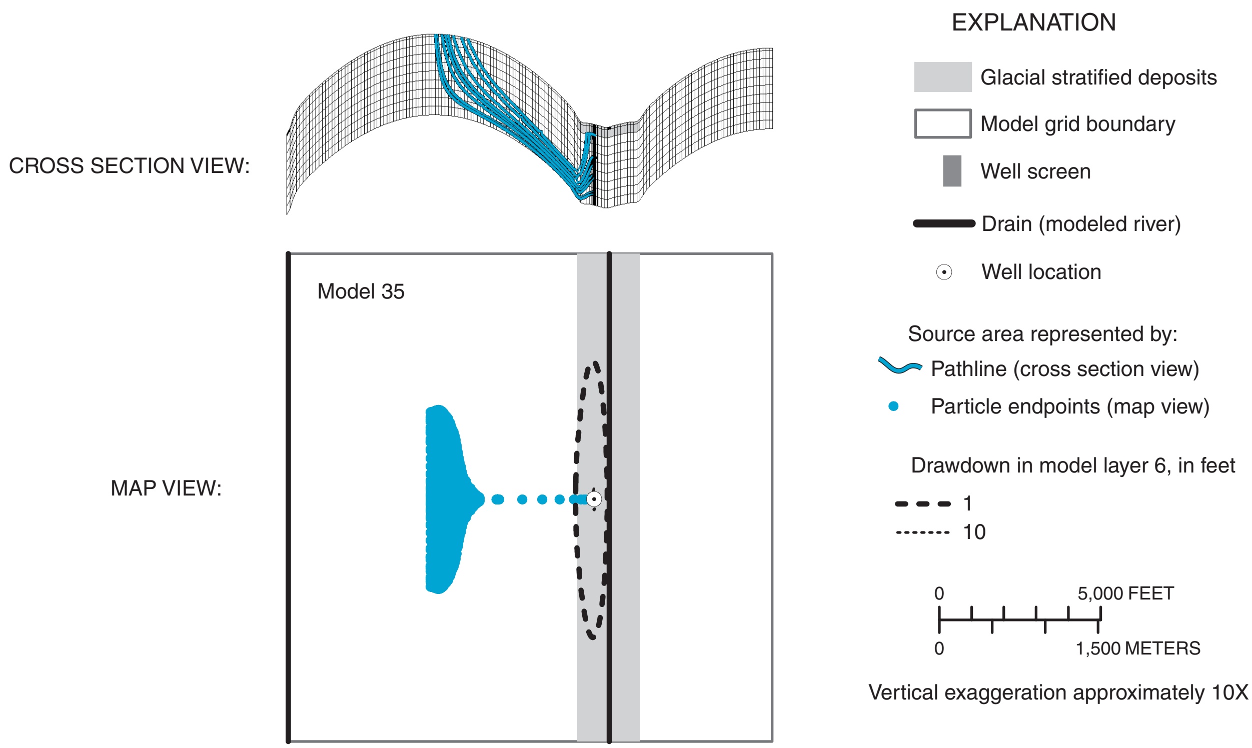


Figure 22. Source-water areas to a well and drawdown in a narrow valley with a surficial aquifer and anisotropic bedrock.

Bedrock Wells in a Broad Valley with a Surficial Aquifer

Broad valleys with surficial aquifers are typical of the Central Lowland in Connecticut. These simulations include fractures with a low angle of dip, which is typical of the Central Lowland, but which also applies to other bedrock that has a low angle of dip. The anisotropic nature of these rocks may be enhanced because the layer-parallel fractures and the cross fractures tend to strike in a similar direction. In the Central Lowland, the surficial aquifer is underlain primarily by sedimentary rock that has soils developed on fine- or coarse-grained glacial stratified deposits. The water table commonly is in the glacial deposits. The sedimentary rock in the Central Lowland is layered with a shallow dip to the east of about 10^{o}. The angle of dip increases near the edges of the lowland. Broad valleys can contain both ground-water discharge and recharge areas.

Model Design

The model grid and boundary conditions are different than in previously discussed models. Topographic relief generally is less in broad valleys than in the uplands, and the surficial aquifer tends to be thicker and more uniform. The model grid and boundary conditions were designed to simulate source-water areas to wells in a flat, broad valley containing a surficial aquifer overlying bedrock with a low angle of dip (table 9). The models used to describe this setting were constructed using 50-by-50 ft model cells in a 15,000-by-15,000-ft grid (fig. 23). The model consisted of 11 layers—1 layer representing the surficial aquifer (upper layer) and 10 layers representing the bedrock (lower 10 layers). For this set of simulations, rivers were simulated as drains. If the simulated water level falls below the drain, no more water is removed, and more importantly, no water is added to the model. In the real world, rivers can be sources of water to pumped wells through induced infiltration. In these simulations, induced infiltration was not simulated so that the effects of fractures and topography could be isolated.

Fractures were simulated using ideas discussed by Stone and others (1996) and Goode and Senior (2000). Stone and others (1996) conceptualized flow in the fractured bedrock of the Central Lowland as primarily in bedding-plane partings, which are connected by near-vertical fractures. The vertical fractures tend to terminate against more competent beds, and the result is a stair-step pattern of flow through vertical fractures, across bedding planes, and so on toward discharge areas. Although this conceptual model applies strictly to the Central Lowland, the concepts apply in a general way to metamorphic bedrock with a low angle of dip. Goode and Senior (2000) simulated ground-water flow in such a system by explicitly including a single bedding plane parting in the model as a high hydraulic conductivity zone. Hydraulic properties were assigned as in previous models, although there is limited evidence that hydraulic conductivity may be higher in sedimentary bedrock (table 10).

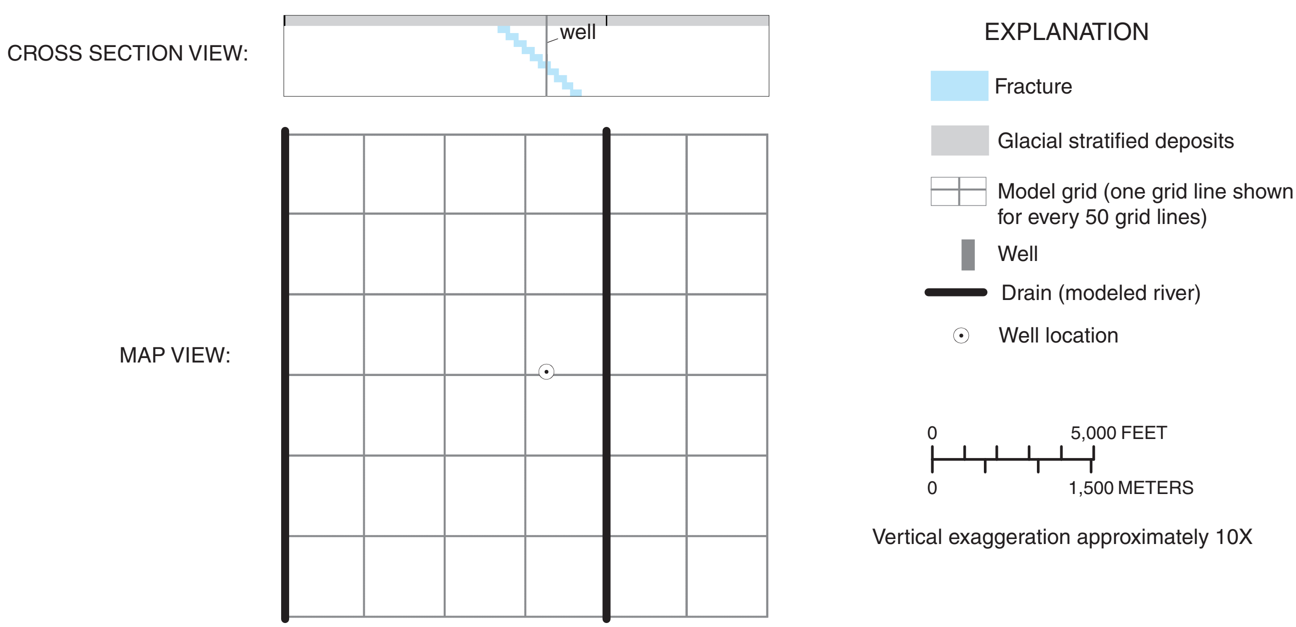


Figure 23. Finite-difference grid and boundary conditions for simulation model of a well in a broad valley with a surficial aquifer.

Table 9. Model grid, boundary conditions, and hydraulic properties of simulation models for wells in a broad valley with a surficial aquifer.

[ft, feet; ft/d, feet per day; in/yr, inches per year; ft^3/d, cubic feet per day; gal/min, gallons per minute]

Model geometry	
Grid	50-by-50-ft model cells in a 15,000-by-15,000-ft grid.
Layers	Eleven layers were used. Upper layer 65 ft thick representing the surficial aquifer. Ten lower layers 43.5 ft thick representing bedrock.
Boundary conditions	
Base and lateral boundaries of model	No-flow boundaries surround the model domain.
Streams	Rivers were simulated as drains. The bottom elevation of the drain was set to 0 ft and the conductance of the drain was set to 100,000 ft/d. The high conductance allowed water to flow freely into the drain, and no water could flow from the river to the aquifer.
Recharge	Recharge was applied to the top layer of the model at the rate of 0.005 ft/d (about 22 in/yr).
Well	A well was simulated in layer 6 pumping at about 50 gal/min (10,000 ft^3/d). The cells in layers 3 through 11 in which the well was simulated were assigned a high vertical hydraulic conductivity (10,000 ft/d) to approximate the effects of an open borehole.
Hydraulic properties	
Horizontal hydraulic conductivity	14 ft/d, coarse-grained glacial stratified deposits; 0.08 ft/d, bedrock; 17 ft/d, fracture. Hydraulic conductivity was isotropic horizontally, unless otherwise noted.
Vertical hydraulic conductivity	Vertical hydraulic conductivity was equal to horizontal hydraulic conductivity, unless otherwise noted.

Table 10. Hydraulic and hydrologic properties used to simulate ground-water flow in sedimentary bedrock in the northeastern United States.

[K, hydraulic conductivity; subscript of K indicates grid direction where x is along rows, y is along columns]

Material	Values from previous studies		
	Stone and others (1996)	Senior and Goode (1999)	Goode and Senior (2000)
Hydraulic conductivity of bedrock, in feet per day			
Fracture K_x	92		7.2 to 36
K_y/ K_x		0.04 to 0.09	
Unfractured rock			9.6 x 10^{-4}
Bulk rock K_x	0.35 to 0.60	.19 to 11.4	.05

Simulated Factors

The basic model and four variations were simulated for a well in a broad valley with a surficial aquifer (table 11). The variations include the basic model with a higher vertical hydraulic conductivity, such as might be found where vertical fractures are numerous. Senior and Goode (1999) used a similar approach in their model, a low ratio of horizontal hydraulic conductivity to vertical hydraulic conductivity (table 10). Another variation included simulating a poor seal between bedrock and the surficial aquifer by including the high vertical conductivity that simulates the open well bore as being present in layer 2 (in other simulations, the well bore is simulated in layers 3 through 11 only, thus creating a separation distance between the surficial aquifer and the open borehole in bedrock). Two other variations simulated the effect on the location and number of fractures.

Effect of Simulated Factors on Source-Water Areas

The source-water areas to wells are affected by the ground-water-flow direction in the surficial aquifer and by the fracture properties (fig. 24). The effect of the fracture is to channel the water downward from the surficial aquifer toward the open borehole. If leakage takes place through the vertical fractures, the source-water area is less affected by the fracture geometry. The case where there was not a good seal between the surficial aquifer and the borehole allowed water to come from closer to the well (fig. 25). The presence of a fracture near the top of bedrock in the well could similarly allow water to come from near the borehole.

Table 11. Simulation models of wells in a broad valley with a surficial aquifer.

[K_h, horizontal hydraulic conductivity, K_v, vertical hydraulic conductivity]

Model	Characteristics
43	Well in broad valley basic model
44	Well in broad valley with $K_h/K_v = 0.01$
45	Well in broad valley, no seal between bedrock and surficial aquifer
46	Well in broad valley, with fracture near well
47	Well in broad valley, with two fractures

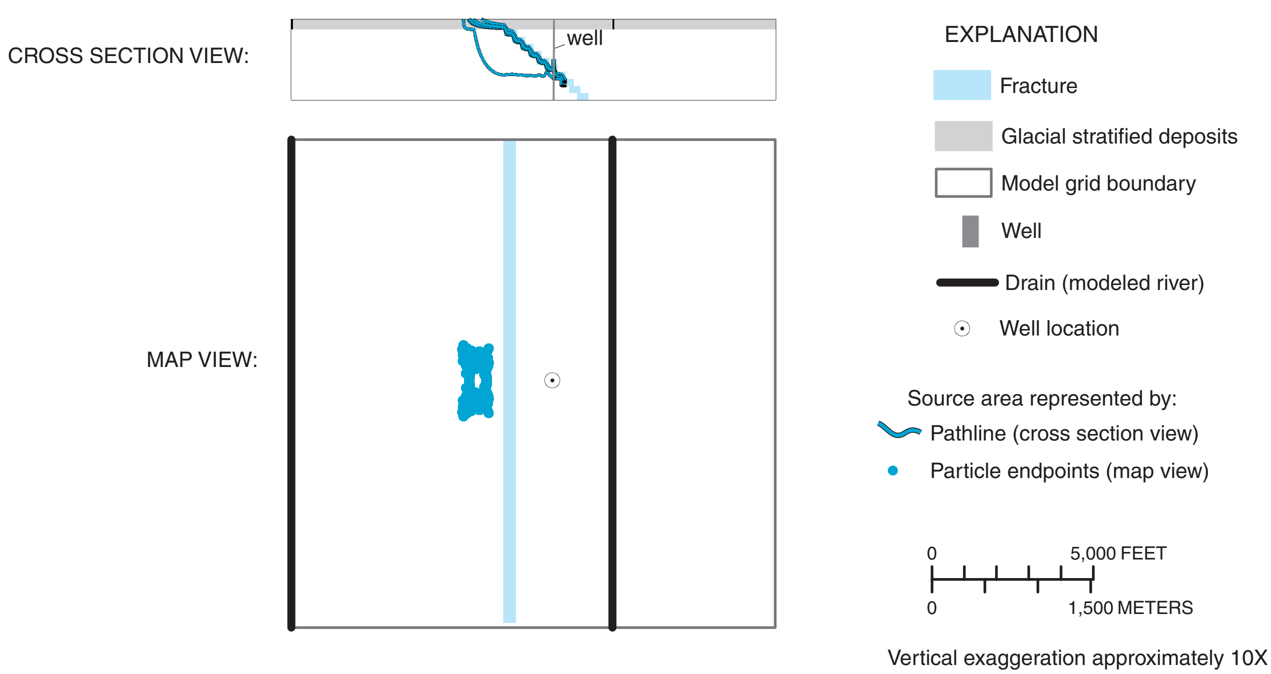


Figure 24. Source-water areas to a well in a broad valley with a surficial aquifer.

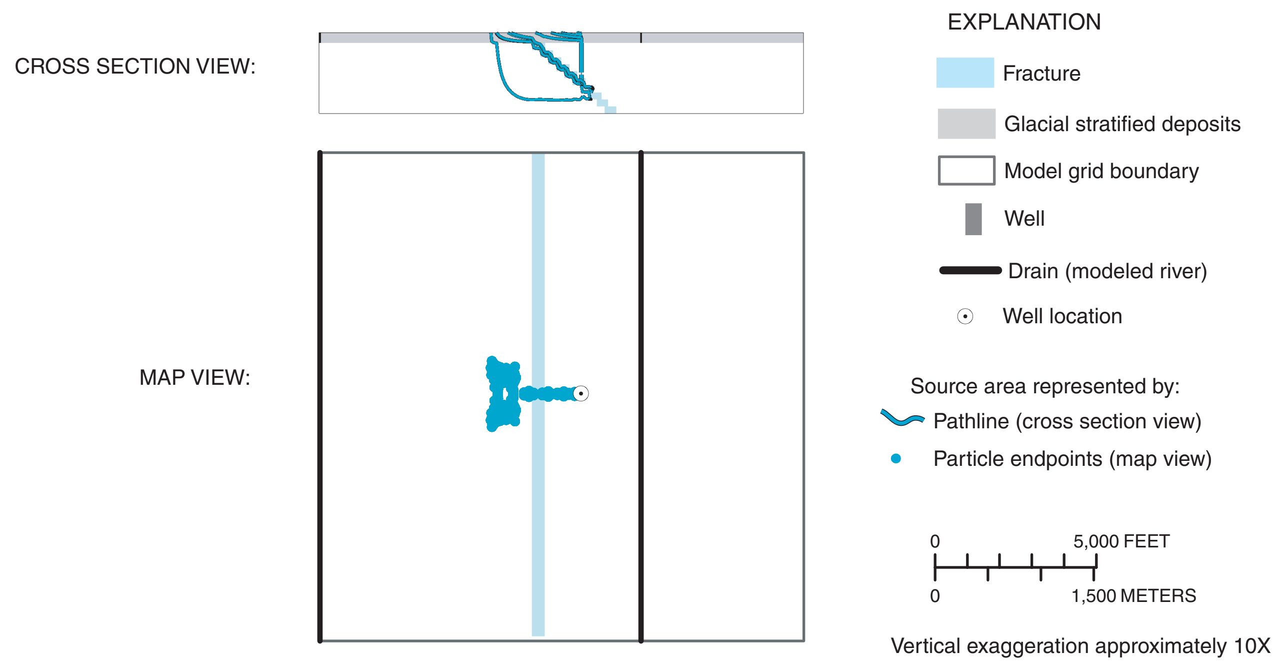


Figure 25. Source-water area to a well in a broad valley with a poor seal between the surficial aquifer and the bedrock.

Geohydrologic Investigation in Old Lyme, Connecticut: A Case Study

A geohydrologic investigation was conducted in the Old Lyme, Connecticut, quadrangle to help understand the relation between geologic structure and the hydrology of fractured-rock aquifers, in particular, the delineation of source-water areas. A numerical simulation model in Old Lyme was constructed and used to delineate source-water areas to four hypothetical wells, given a realistic set of geologic, hydrologic, and topographic factors. The source-water areas delineated with the Old Lyme numerical model were evaluated with respect to the source-water areas delineated from the analytical methods and simulation models of the three settings. This part of the study focused on whether the steeply dipping bedrock results in a hydraulic anisotropy that affects the shape of source-water areas, and if so, how the hydraulic anisotropy could be quantified. Another goal was to assess the effects of the presence or absence of the surficial aquifer on the shape and location of source-water areas.

The Old Lyme quadrangle (fig. 26) is in the Western Uplands physiographic region and borders Long Island Sound and the eastern side of the Connecticut River. The area is underlain primarily by steeply dipping metamorphic rocks. Surficial aquifers are present in the river valleys, but much of the area has no surficial aquifer. Methods for this investigation included measurements of fracture orientations at outcrops, borehole-geophysical logging, and an aquifer test near at the Sound View community well field in Old Lyme. Fractures were measured in outcrops to see if the local geologic structure was similar to the regional geologic structure and to local fractures observed in the boreholes at Sound View. Borehole logging was conducted at the well field to determine the position and orientation of fractures in the boreholes. The aquifer test was used to determine the response of ground-water levels and streamflow to pumping in and near the well field and to test whether the preferred fracture orientation observed in the bedrock would affect ground-water-flow during pumping at the site.

The investigation also included detailed geologic mapping at the 1:24,000 scale, an inventory of about 1,000 well drillers' reports, geostatistical analysis of well yields, and simulation of ground-water flow in a 24-mi^2 area in Old Lyme. Detailed geologic mapping was done to assess the regional distribution of types and orientations of fractures. The well drillers' reports were used in a geostatistical analysis to see if well yields had spatial correlations in certain directions. Ground-water flow was simulated to assess the effect of hydraulic anisotropy on the shape of source-water areas and to delineate source-water areas for comparison with analytical methods.

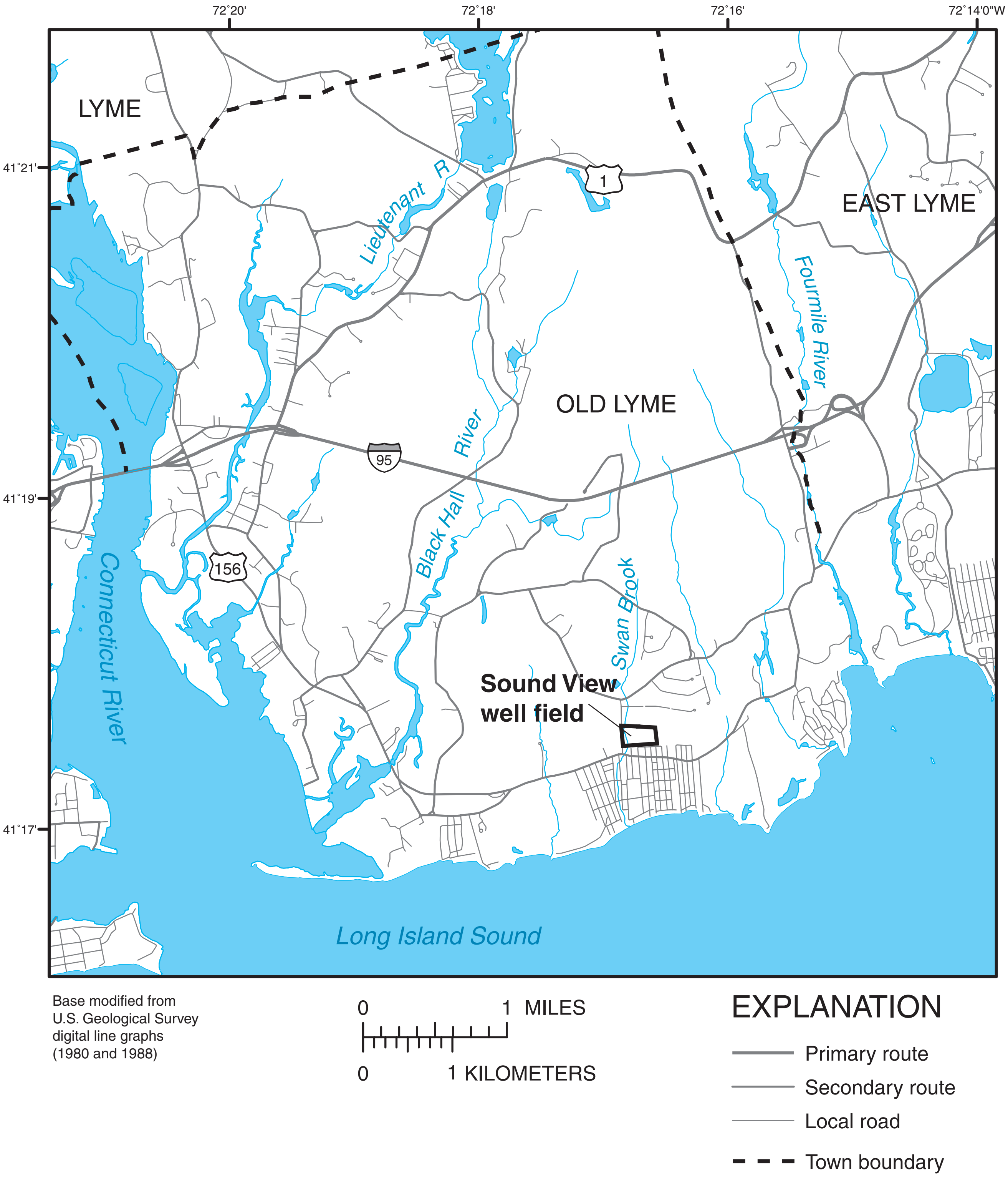


Figure 26. Old Lyme study area, Connecticut. (See fig. 1 for location of Old Lyme.)

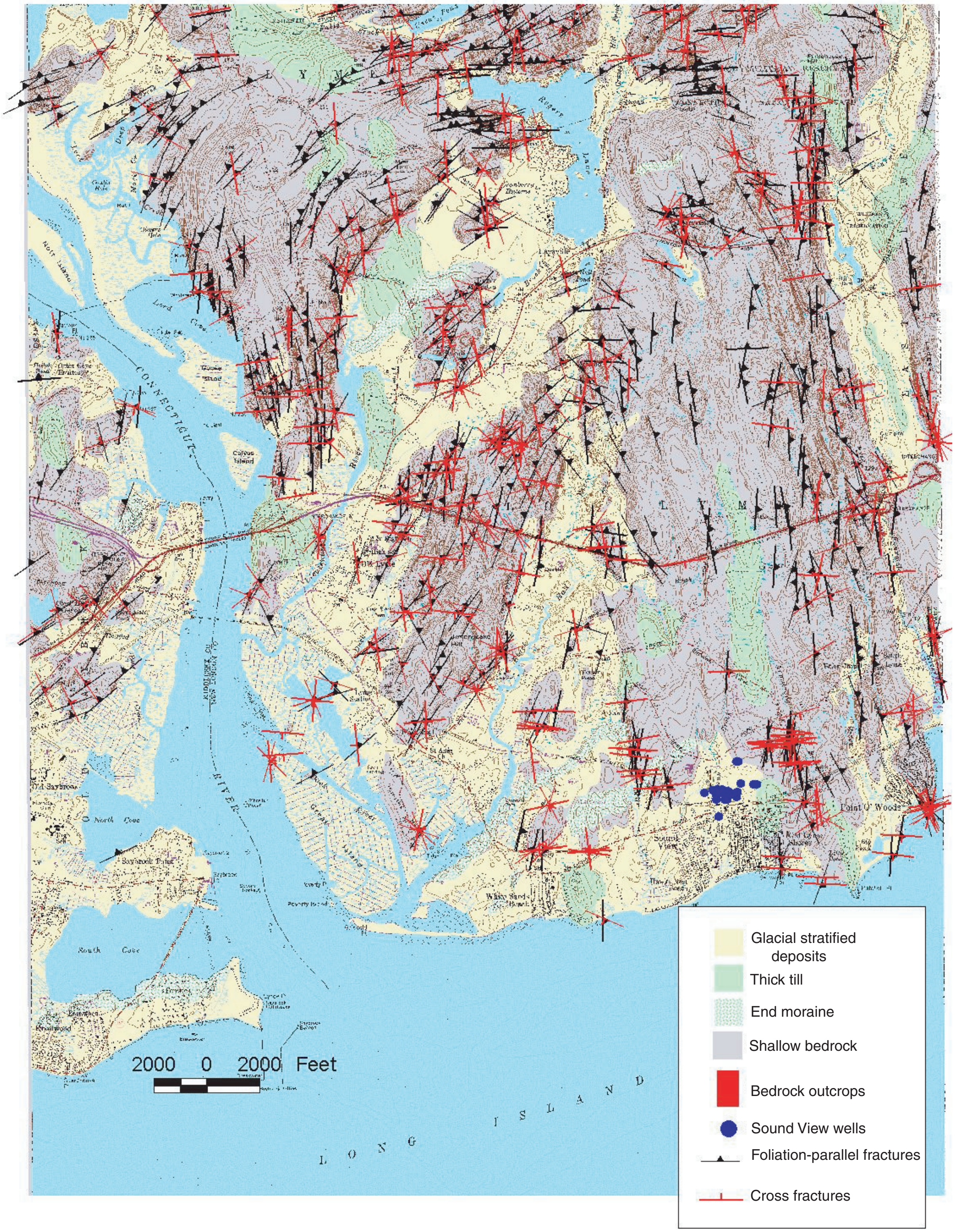


Figure 27. Geology of Old Lyme study area, Connecticut.

Geology and Hydrology of the Sound View Well Field

The well field is underlain by mixed gneiss, schist, and thin quartzite layers of the Plainfield Formation. The well field is on the eastern side of the Old Lyme structural dome, where foliation and layering in the rock strike northerly and dip steeply eastward (60-80°) (fig. 27). Glacial stratified deposits, consisting of very coarse gravel and sand, overlie bedrock at the well field and extend south to Long Island Sound. These deposits are 50 ft thick on the eastern edge of the site and taper to 10 ft thick on the western edge. Two streams flow across the well field—Swan Brook in the west and an unnamed brook in the east. Several seasonal wetlands were present at the time of the aquifer test.

The orientation of fractures was measured in bedrock outcrops near the well field. These fractures fall into three types: (1) foliation-parallel fractures that strike N-S, dip steeply eastward, and typically are continuous at the outcrop, (2) subhorizontal unroofing joints, and (3) two sets of cross-fractures—one near-vertical set that strikes E-W, and another set that strikes N-S and dips steeply west; the cross fractures are typically not continuous. Fracture geometry near the well field is dominated by continuous foliation-parallel fracturing and well-developed unroofing joints. The foliation-parallel fractures provide a strongly N-S-oriented vertical connection to subhorizontal unroofing joint zones (fig. 28).

Borehole-geophysical logging was conducted in 5 of the 14 bedrock wells at the Sound View well field (wells SV-11, SV-13, SV-10, SV-9, and SV-2 on fig. 29). Logs included acoustic and optical televiewer, caliper, gamma, fluid conductivity, and fluid temperature. Detailed driller's logs were available for most of the wells, and depths of water-bearing zones were noted on the logs.

Orientations of foliation and fractures identified from optical televiewer logs in the wells corresponded to measurements at nearby outcrops (fig. 30 stereonet). Major fractures observed on optical and acoustic televiewer logs included both subhorizontal unroofing joints and N-S striking, east-dipping, steep foliation-parallel fractures. Well-developed subhorizontal zones at 50 to 75 ft in depth and (or) at 100 to 130 ft in depth were observed in wells SV-11, SV-10, SV-13, and SV-2. Foliation-parallel fractures that appear to be water-bearing in well SV-11 at 130 to 140 ft in depth strike northward to the vicinity of well SV-12 and likely provide a strong connection between these two wells. A large water-bearing unroofing fracture is present at 76 ft in depth in well SV-11, but a water-bearing zone at that depth was not noted in the driller's log for this well.

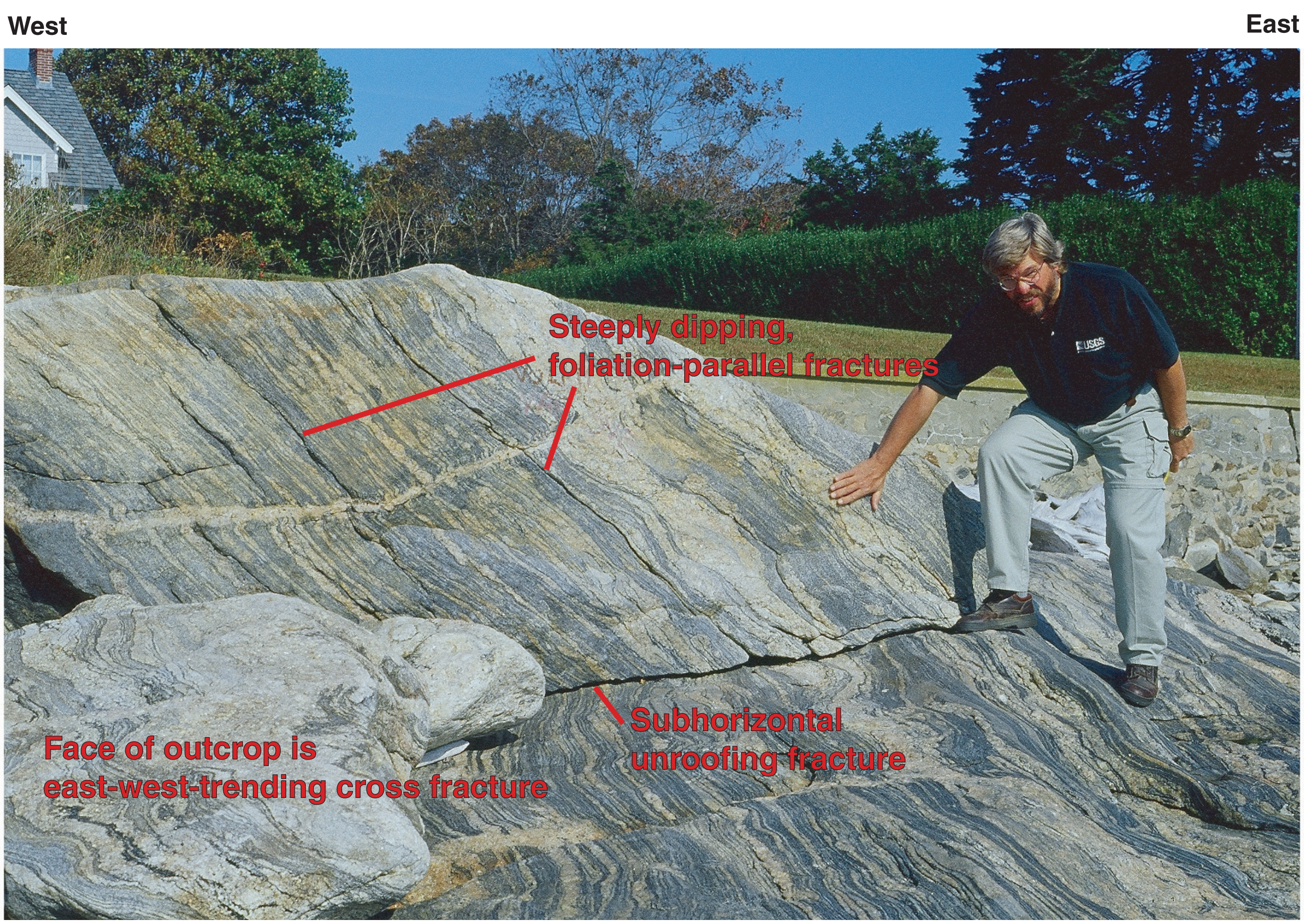


Figure 28. Fracture patterns at Salt Works Point, Old Lyme, Connecticut. (Photograph by J.R. Stone, U.S. Geological Survey.)

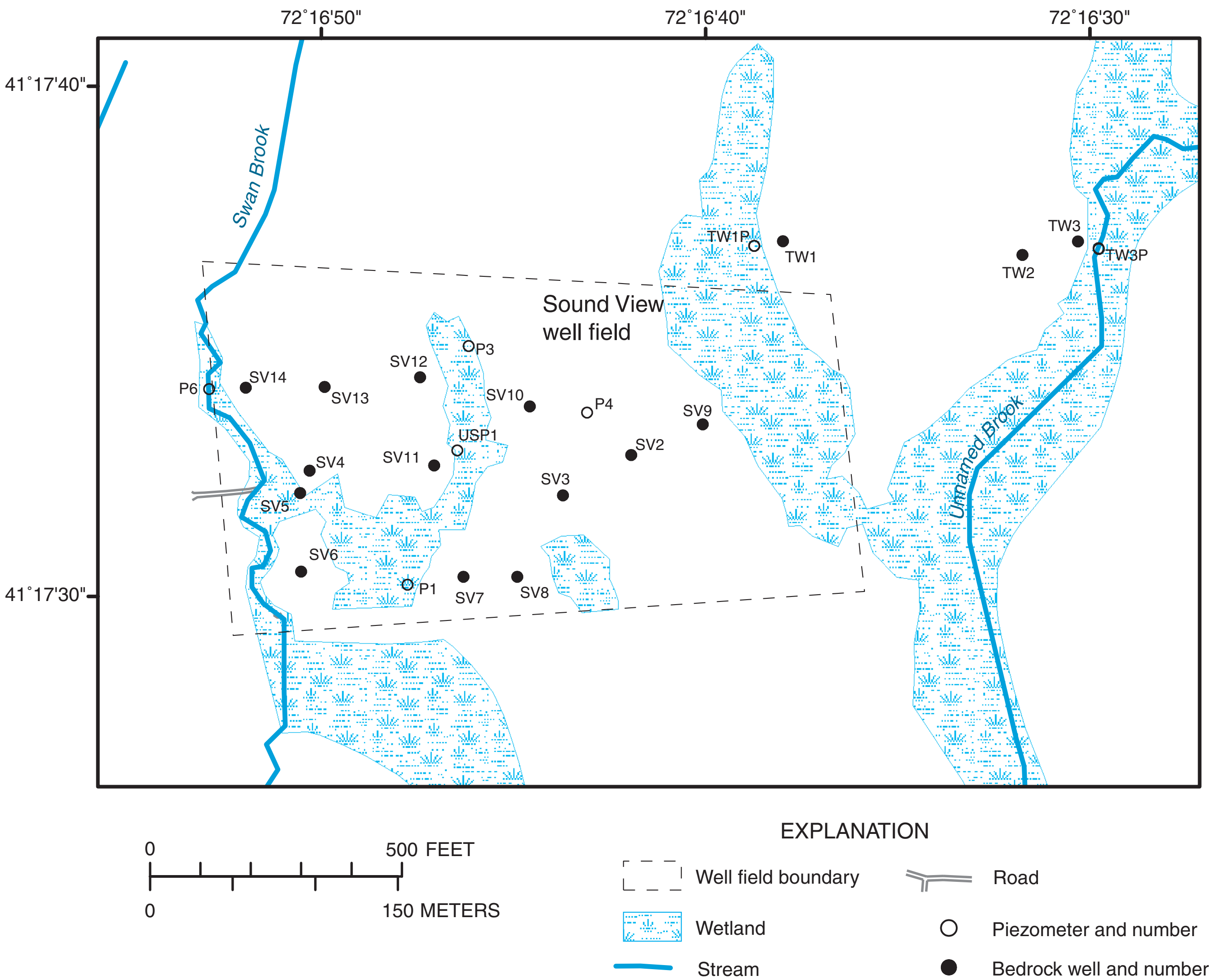


Figure 29. Sound View well field, Old Lyme, Connecticut.

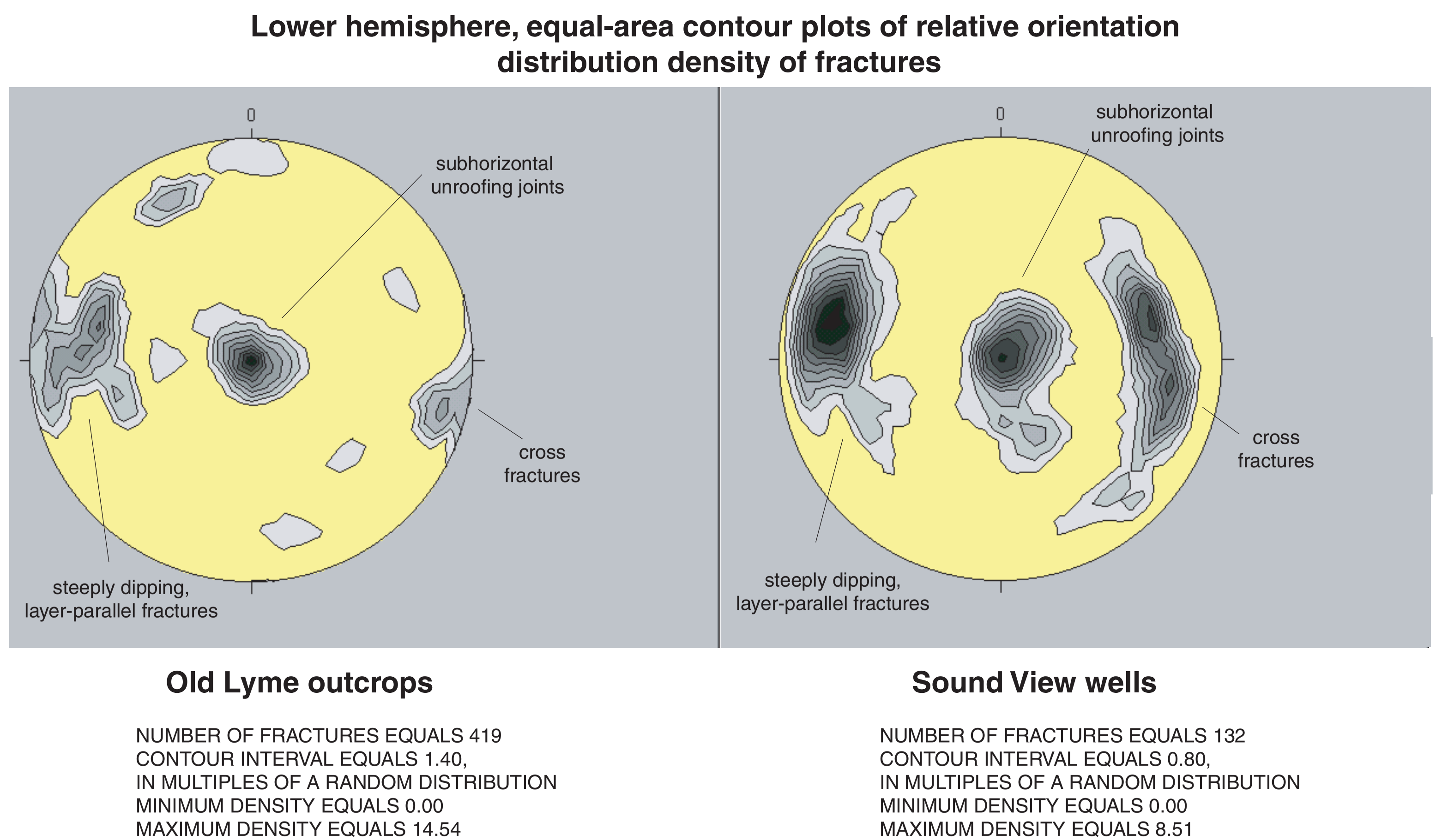


Figure 30. Fracture orientations in outcrops and Sound View wells, Old Lyme, Connecticut.

Aquifer Test at the Sound View Well Field

An aquifer test was conducted at the Sound View public-supply well field from May 2 to 23, 2001 to determine (1) the hydraulic properties of the bedrock and (2) if drawdown responses were affected by the geologic structure. Well SV-11 was pumped at 20 gal/min for 20 days. Drawdown was measured in 15 wells that are cased through the surficial materials and open-hole in the bedrock, and 9 piezometers that are open to the overlying glacial deposits. An analytical method was used to estimate aquifer properties from the data that were collected. A more detailed numerical model would be needed to analyze aquifer properties in relation to the complex fracture system that is evident from borehole-geophysical surveys onsite and from regional geologic structure at area outcrops, but that was beyond the scope of this study. Data from the aquifer test are available from the Connecticut District of the USGS.

Drawdown data in fractured-rock aquifers can be analyzed by fitting the Theis curve to the data at early and late times (Kruseman and deRidder, 1994). Drawdown data from an open-hole well in a fractured-rock aquifer can be used to determine aquifer transmissivity, but are less useful in determining aquifer-storage properties (Tiedeman and Hsieh, 2001). Drawdown initially (early time) occurs at a rate determined by fracture properties very close to the well bore. Drawdown stabilizes at some level in each well because the well discharge is balanced by the flow of water into the subhorizontal fracture zone from a source. Once the capacity of that source to supply water to the larger fractures is reached, drawdown takes place at a rate determined by the bulk rock matrix (late time).

Water-level changes in wells during the aquifer test show three basic patterns. Data presented by Hydrodynamic (John Sima, written commun., 1989) show that wells SV-11 and SV-12 have identical drawdown when both wells are pumped, and early-time drawdown in SV-12 follows a linear slope somewhat less than 1:2. This suggests that a single fracture or a single flow path within a fracture plane connects these two wells (Kruseman and deRidder, 1994). Data from SV-7, SV-8, SV-10, SV-12, and SV-13 are indicative of a source of water to the aquifer during the test because of the S-shape of the curves (fig. 31). The source of water could be delayed yield from the glacial deposits or from the matrix of relatively unfractured rock. Data from wells SV-1, SV-2, SV-3, SV-6, SV-9, and SV-14 responded in a manner consistent with radial flow at late time.

Early-time data did not yield consistent results, although the match to the Theis curve was good in many cases. Late-time data yielded results that are more consistent. Late-time transmissivity ranged from 61 to 243 ft^2/d (table 12). Well SV-3 is known to be a poorly producing well (David Radka, Connecticut Water Co., oral commun., 2001) and it had the lowest transmissivity of all the wells. The transmissivity, exclusive of SV-3, ranged from 127 to 243 ft^2/d. This range is small, considering the transmissivity in fractured rock often ranges over orders of magnitude.

Late-time storage was lowest in SV-12 (10^{-7}), which is consistent with a water-level response being rapidly transmitted through a fracture from SV-11 (the pumped well) to SV-12. The wells that responded to a source of water had the next lowest storage coefficients, which ranged from 2×10^{-4} in SV-10 to 1×10^{-2} in SV-13. The remaining wells had storage coefficients that ranged from 2×10^{-2} to 5×10^{-2}.

Table 12. Aquifer hydraulic properties, Sound View well field, Old Lyme, Connecticut.

Well	Aquifer hydraulic properties	
	Late-time transmissivity (in feet squared per day)	Late-time storage coefficient (dimensionless)
SV-01	146	1.64E-02
SV-02	127	2.64E-02
SV-03	61	5.30E-02
SV-06	243	3.98E-02
SV-07	162	2.73E-03
SV-08	162	1.62E-03
SV-09	153	2.21E-02
SV-10	187	1.87E-04
SV-12	170	1.19E-07
SV-13	188	1.31E-02
SV-14	208	2.66E-02

The wells clearly differ in their response to pumping although the transmissivities were remarkably consistent. The speed at which drawdown progresses through an aquifer is governed by the aquifer diffusivity (transmissivity divided by storage coefficient): if an aquifer has a very low aquifer diffusivity, drawdown is propagated through the aquifer more quickly than in an aquifer that has a high aquifer diffusivity. The aquifer diffusivities differ among the wells, with the wells having the most drawdown, and therefore presumably the best connections through the fracture network, having the lowest aquifer diffusivities. Well SV-12 has, by far, the lowest aquifer diffusivity, and well SV-3 has the highest.

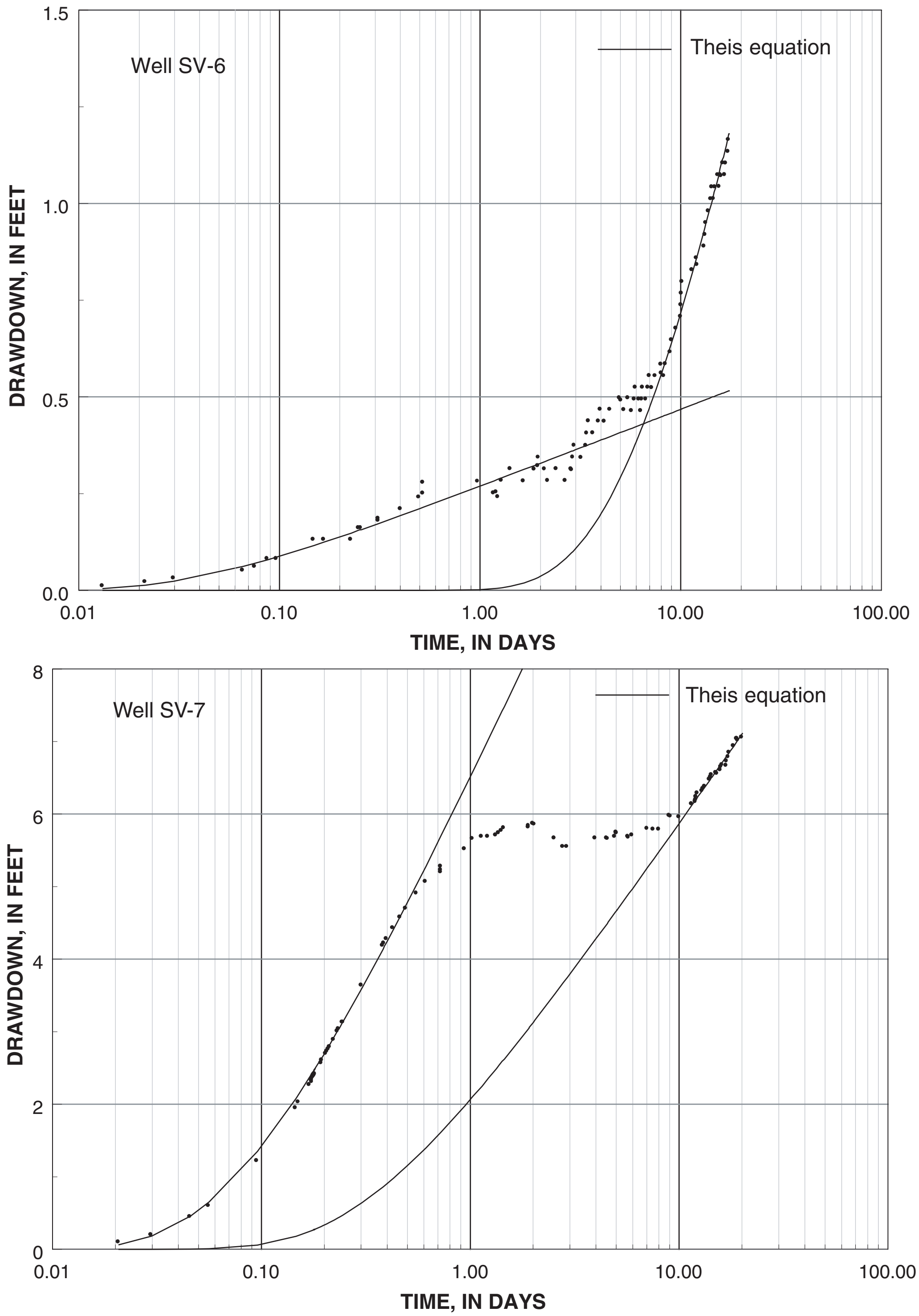


Figure 31. Drawdown in observation wells SV-6 and SV-7 during an aquifer test at well SV-11, Sound View Well Field, Old Lyme, Connecticut.

The aquifer test itself may bias the conceptual model because multiple vertical wells are likely to penetrate the same subhorizontal fracture zone but not the same near-vertical fracture zone. Water-level responses in aquifer tests commonly show that some wells are strongly connected to the pumped well(s), whereas other wells nearby are weakly connected (Shapiro and Hsieh, 2001). Detailed tracer and hydraulic testing of discrete intervals (Shapiro and Hsieh, 2001) shows that strongly connected wells commonly penetrate a common subhorizontal zone that comprises fractures of various orientations. Even though aquifer tests show that subhorizontal fractures are important in well-field hydraulics, water must enter subhorizontal zones through vertical connections, which may be near-vertical fractures, to the land surface. The nature of the vertical connection is important in this study because the connection with land surface determines the location and shape of source-water areas. Aquifer tests with a single pumped well also may bias the conceptual model because they tend to look at fracture connections to a single well. Water-level responses to the aquifer test in Old Lyme and to other hydraulic stresses in the area show that there are many connected zones, possibly even overlapping zones, that exist in and near the well field.

Borehole, geologic, and water-level data collected at the site were used to form a conceptual model of the aquifer, similar to the method used by Risser and Barton (1995). The patterns of drawdown at the well field during the aquifer test are consistent with the interpretation of the geologic structure in and around the well field. The conceptual model is similar to that proposed by Shapiro and Hsieh (2001) for fractured–rock aquifers in general, where the aquifer consists of subhorizontal zones made up of fractures of various orientations embedded within a matrix of relatively unfractured rock. Borehole geophysical and outcrop data indicate four populations of fractures—subhorizontal unroofing joints, foliation-parallel partings (that strike N-S and dip east), cross fractures (that strike N-S and dip west), and cross fractures (that strike E-W). Outcrop data indicate that foliation-parallel fractures are generally continuous, but that E-W fractures are not. The conceptual model formed from these data is that wells SV-7, SV-8, SV-10, SV-11, SV-12, and SV-13 are in the same subhorizontal zone, and the remaining wells are in the relatively unfractured zone. It also is possible that wells SV-11 and SV-12 penetrate the same N-S steeply dipping fracture.

Spatial Correlation (Variography) of Well Yields in Old Lyme

Drew and others (1999; 2001) have shown a spatial relation between geologic structure and well yield in the Pinardville, New Hampshire, quadrangle using variogram analysis. They found that yields from low-yield wells (yield less than 40 gal/min) were spatially correlated, and that the correlation was consistent with geologic structure. In that study, high-yield wells were thought to be caused by single, high permeability fractures and thus were excluded from the analysis. Individual bedrock types were found to vary in a complex manner, ranging from rocks that have no spatial structure to rocks that have a strong local and (or) regional preferred orientation of correlations.

The variogram statistic was computed for well yields in the Old Lyme study area. The variogram is “half the average squared difference between all possible pairs of data values whose locations are separated by a certain distance in a particular direction” (Isaaks and Srivastava, 1989, p. 60 and p. 52). This analysis reveals the spatial structure of the variable of interest. In this study, the same approach as Drew and others was used; well yields greater than 40 gal/min were excluded because these well yields may be the result of single high permeability fractures. The variogram can be computed for each range of distance and angle of separation to produce a two-dimensional map that reveals the spatial structure of well yield. One complication of variogram analysis is that well yields may be related to topographic highs and lows, not because of fractures, but because of the higher hydraulic gradient in the source-water areas to wells in topographically low areas. The apparent spatial correlation of well yields with geologic structure may be strongly affected by a correlation of geologic structure with topographic highs and lows.

The variogram for the Old Lyme quadrangle has a strong northerly orientation (fig. 32). Variogram values at h_x=0 are lower than at other h_x positions (left boxplot on fig. 32). Variogram values along the y-axis do not show significant variation (right boxplot on fig. 32). This spatial distribution of variogram values parallels the dominant north-south structure of the fractures in the bedrock, and the yields seem to be correlated to a distance of about 4,000 ft.

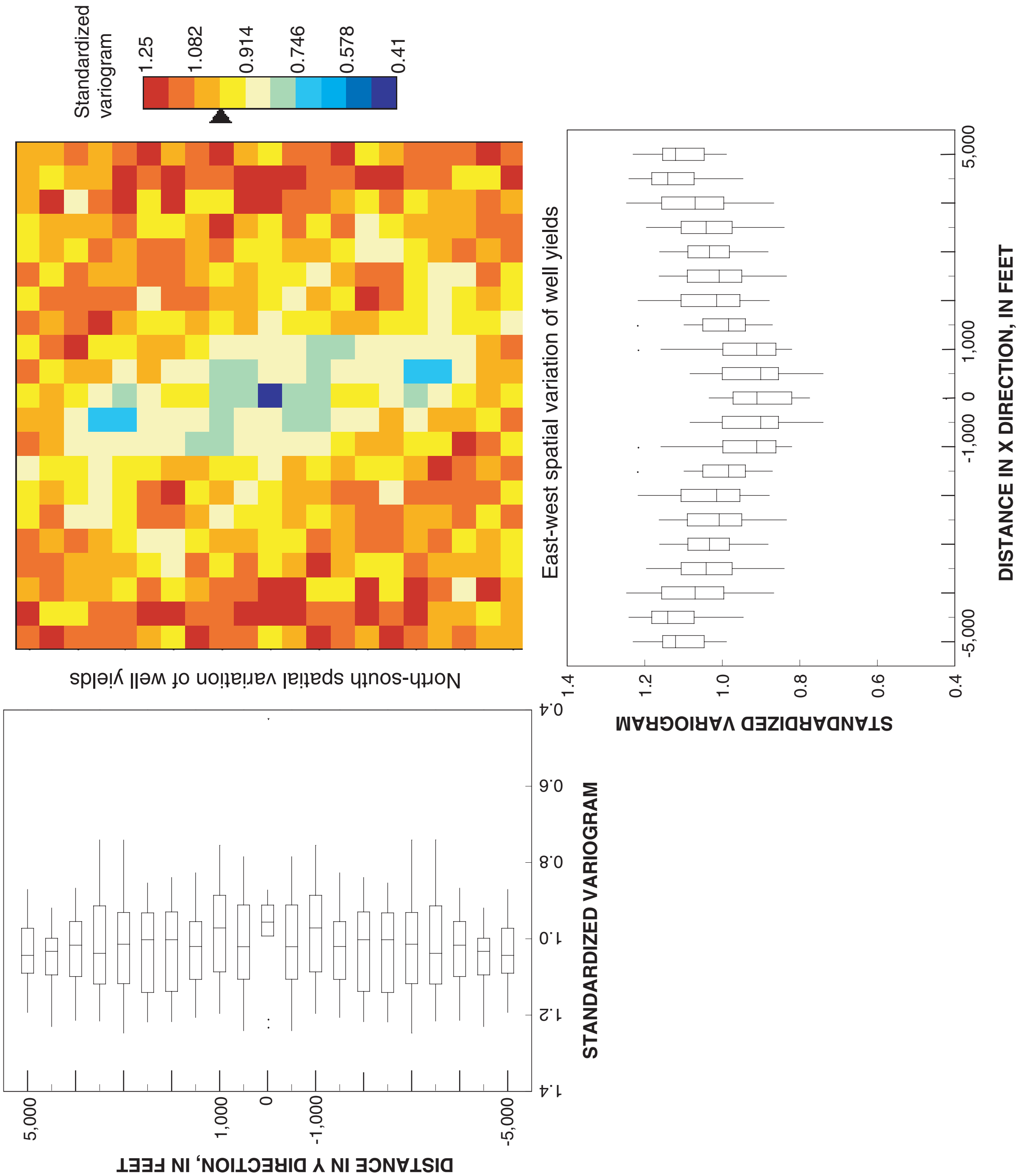


Figure 32. Variogram surface for well yields, Old Lyme study area, Connecticut.

Simulation of Ground-Water Flow in the Old Lyme Study Area

Numerical simulation models of ground-water flow were constructed for the Old Lyme area to test concepts about source-water areas in different geohydrologic settings. The model design was based on (1) accurate representation of the surface topography and surficial geology, (2) generalized concepts of subsurface geology that have been demonstrated by other investigators to be useful, and (3) a nonlinear regression calibration to existing data such that there was a minimum of bias in the model. The models should not be used to design wells, well fields, or to predict drawdown at any well or well field. The source-water areas shown in this section could be approximated more accurately by collecting (1) additional data on the rate of recharge to the bedrock, (2) additional streamflow and water-level measurements, and (3) additional aquifer-test data to define spatial distribution and variability of aquifer properties.

Ground-water flow was simulated using a finite-difference ground-water-flow model and parameter estimation in a 24-mi^2 area predominantly in the town of Old Lyme, Connecticut (fig. 33). Ground-water flow was simulated using the computer program MODFLOW-2000 (Harbaugh and others, 2000; Hill and others, 2000), which is based on MODFLOW, a computer program that simulates three-dimensional ground-water flow through a porous medium by using a finite-difference method (McDonald and Harbaugh, 1988). MODFLOW-2000 has the capability to solve a MODFLOW calibration problem by calculating values of selected input data that result in the best match between measured and model calculated values.

The calibration data consisted of estimated streamflow and water levels measured by drillers in 108 wells completed in surficial materials and 714 wells completed in bedrock. Median annual streamflow for the Lieutenant, Black Hall, and Four Mile Rivers (fig. 26) was calculated using an equation developed by Ries and Friesz (2000) for streams in Massachusetts. Massachusetts has similar geology, climate, and topography to Connecticut, and the equation is considered valid for estimating streamflow for this model. The median annual flow was considered to be representative of the average streamflow conditions corresponding to average ground-water discharge.

The primary area of interest in the modeled area was the Black Hall River Basin and several smaller coastal basins. To minimize the effects of boundaries at the edges of the modeled area, the modeled area was defined by topographic boundaries not contiguous to the Black Hall River Basin. The eastern and western boundaries were drainage divides between the Four Mile River (eastern boundary) and the Lieutenant River (western boundary) and drainage basins outside the modeled area. Because these boundaries were far from the area of interest and because ground-water divides can be approximated by topographic divides in many places, they were treated as no-flow boundaries. The modeled area is bounded on the south by Long Island Sound, which was treated as a constant-head boundary (head equal to zero; sea level). The northern boundary was defined by the drainage basin of Rogers Lake, which is part of the Black Hall River Basin. Rogers Lake was treated as a constant-head boundary (head equal to 36 ft); the drainage basin boundary up to Rogers Lake was treated as a no-flow boundary.

The modeled area was divided into grid cells for simulation; the grid was oriented in the direction of geologic structure and presumed direction of principal anisotropy (grid columns oriented north-south). The cells of the grid were 1,000 ft by 1,000 ft. The grid had five layers numbered from 1 (the surface layer) to 5 (the deepest layer). Layer 1 was simulated as a water-table layer—the top of layer 1 was the altitude of land surface, but the saturated thickness of the layer was determined by the simulated water level within layer 1. The bottom of layer 1 was at the top of the bedrock surface, as determined from driller's logs. The thicknesses of layers 2 to 5 were 25, 100, 200, and 200 ft, respectively.

Hydraulic properties were assigned to the model grid in five zones that corresponded to surficial and bedrock geology (fig. 33). Layer 1 contained zones that represented the surficial units till, thick till, coarse-grained glacial stratified deposits, and fine-grained glacial stratified deposits. Layers 2, 3, 4, and 5 represented layered bedrock with a high angle of dip. Horizontal and vertical hydraulic conductivity were assigned to each zone based on previous studies (table 13).

Ground-water flow into and out of the modeled area was governed by boundary conditions, which have already been discussed, and sources and (or) sinks of water in the model. Sources and (or) sinks that were simulated were recharge from precipitation, streamflow gains and (or) losses, and well pumpage. Recharge from precipitation was assigned using zones corresponding to the extent of till or glacial stratified deposits. Streamflow was simulated by specifying the altitude of the stream stage, altitude of top of streambed, altitude of bottom of streambed, and hydraulic conductivity of the streambed. The computer code calculated the amount of streamflow gain and (or) loss and accounted for streamflow in each reach of the stream, so that the amount of water that could leak from the stream was limited by the amount of water flowing in the stream. The simulation model was calibrated without considering domestic well pumpage, so that recharge estimated by the model is an effective recharge inclusive of the effects of domestic pumpage. Most of the water pumped by domestic wells is returned to the ground (layer 1) through septic systems. Although many domestic wells are in the modeled area, the amount of water pumped is small compared to the volume of water in each grid cell, and the combined effect of the domestic wells on the ground-water-flow system was considered minimal.

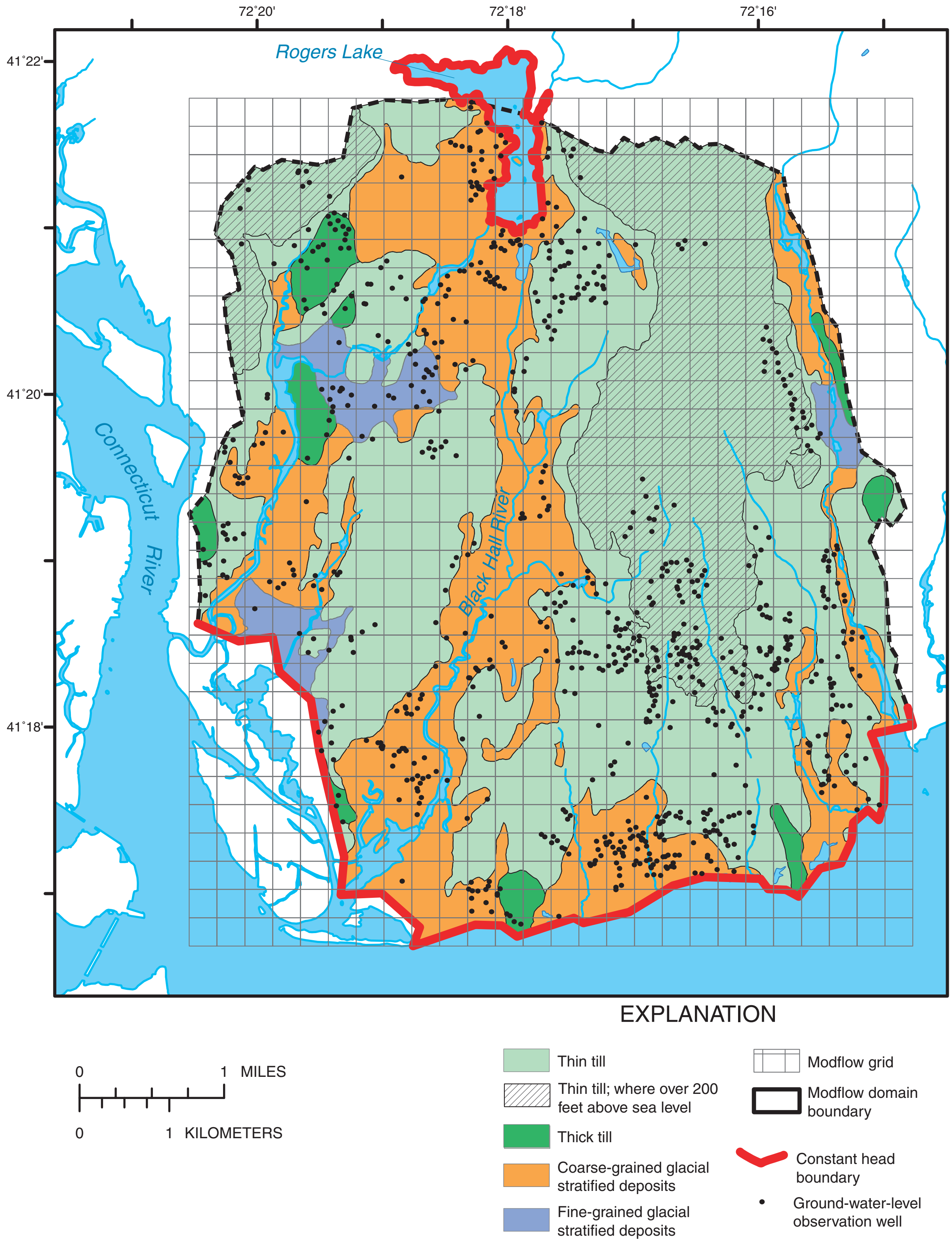


Figure 33. Model grid, boundary conditions, and calibration data locations for simulation model in Old Lyme, Connecticut.

Even with extensive field studies, data commonly are inadequate to depict all fracture connections in a numerical model. Fractured-rock aquifer systems have been simulated with discrete fracture-zone orientation and size determined from water-level data and aquifer-test data (Barton and others, 1999; Goode and Senior, 2000; Lyford and others, 2003). Other investigators have simulated a preferred orientation of fractures in which the hydraulic conductivity in the fracture direction is greater than the hydraulic conductivity perpendicular to fractures (Senior and Goode, 1999; Lipfert and others, 2001). Tiedeman and others (1997) used parameter estimation to test models that simulated a general hydraulic conductivity distribution and a preferred orientation of fractures.

To assess the sensitivity of the simulation to different ways of treating the bedrock, four alternative aquifer treatments were posed—(A) a homogeneous and isotropic aquifer, (B) an isotropic and heterogeneous aquifer having higher hydraulic conductivity under valleys than under hilltops, as has been suggested by several previous studies (for example, Daniel and others, 1997; Tiedeman and others, 1997; Moore and others, 2002), (C) an anisotropic and homogeneous aquifer having different hydraulic conductivities in the direction of strike than perpendicular to strike, and (D) an anisotropic and heterogeneous aquifer (a combination of B and C). Heterogeneity refers to a difference of hydraulic conductivity in the model by location rather than by direction. Tiedeman and others (1997) posed an alternative model by hypothesizing that hydraulic conductivity under hills was lower than hydraulic conductivity under valleys. A similar approach was used in this study by creating two zones for hydraulic conductivity—one zone where land surface is above 200 ft altitude and one zone where it is below. The 200-ft altitude cutoff was chosen because the relation between observed heads and simulated heads seemed to be different for wells above and below 200 ft. Anisotropy is represented in the model by a difference in hydraulic conductivity along the rows of the model grid relative to the columns of the model grid. Anisotropy might be imparted to the rocks, for example, by a prevalence of steeply dipping foliation-parallel partings. Parameter estimates for each model are realistic (table 13), and the residuals of the regression appear equally unbiased (fig. 34).

If inclusion of anisotropy is a reasonable model characteristic for this aquifer, then estimated parameter values for hydraulic conductivity in the row and column directions should be different as long as the model grid is parallel to the directions of maximum and minimum hydraulic conductivity. In this model, the grid is oriented parallel N-S and E-W based on a preliminary analysis of the geologic structure in the Old Lyme area. The estimated anisotropy was 0.15 (K_x/K_y; see table 13). To check that the estimated anisotropy was reasonable, another simulation was done with the model grid rotated 45°. In this grid orientation, the parameter-estimated hydraulic conductivity is about equal in each direction (table 13), indicating that anisotropy is a reasonable way of modeling the aquifer.

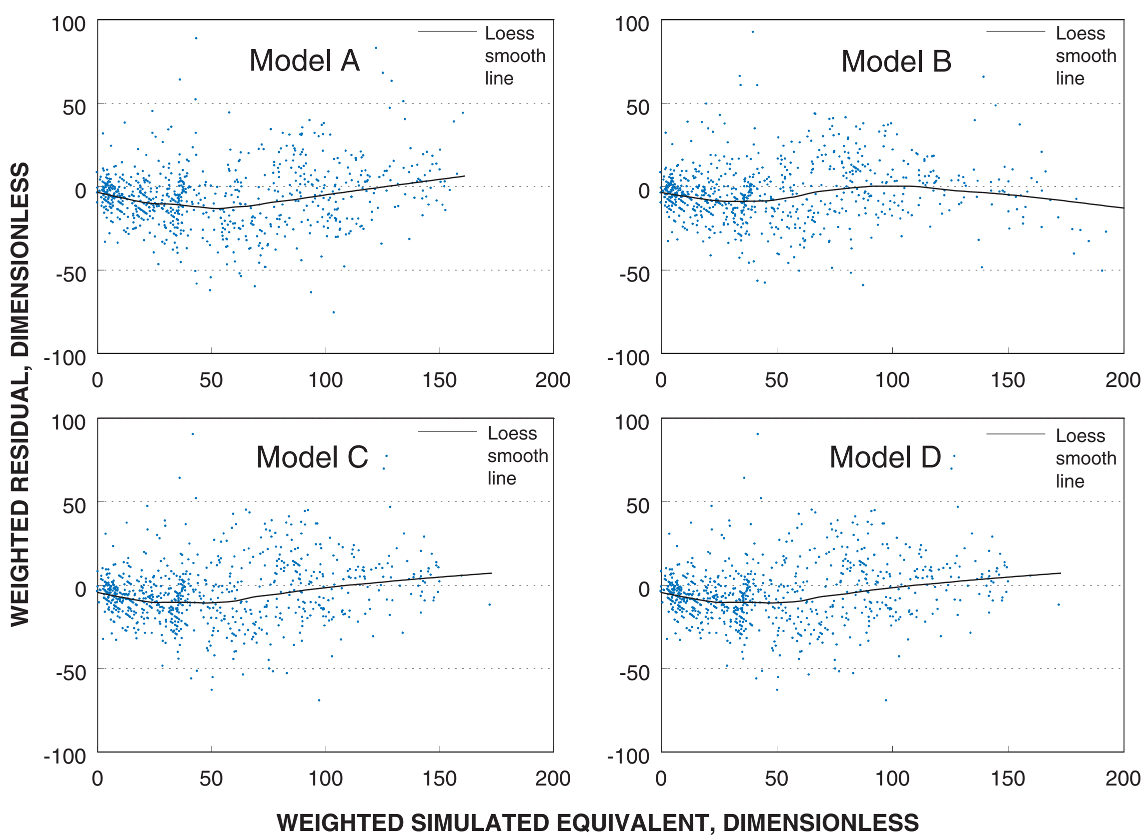


Figure 34. Weighted model residuals and simulated equivalents.

Table 13. Model parameter estimates for a simulation model of ground-water flow, Old Lyme, Connecticut.

[ft/d, feet per day; --, not estimated in the aquifer treatment; K, hydraulic conductivity; subscript of K indicates grid direction where x is along rows, y is along columns, and z is along layers; in/yr, inches per year; shaded cell indicates optimal parameter estimate from model]

Model parameter	Aquifer treatment				
	A	B	C		D
	Isotropic	Isotropic	Anisotropic	Anisotropic[1]	Anisotropic
	Homogeneous	Heterogeneous	Homogeneous	Homogeneous	Heterogeneous
Hydraulic properties, in feet per day					
K_x, rock	0.23		0.13	0.22	--
K_x valley, rock	--	0.33	--	--	0.30
K_x hilltop, rock	--	0.10	--	--	0.088
K_y, rock	--	--	0.85	0.17	--
K_y valley, rock	--	--	--	--	0.54
K_y hilltop, rock	--	--	--	--	0.16
K_z, rock	0.23		0.13	0.22	--
K_z valley, rock	--	0.33	--	--	0.30
K_z hilltop, rock	--	0.10	--	--	0.088
K_x, glacial stratified deposits	180	180	180	180	180
K_z, coarse glacial deposits	1.8	1.8	1.8	1.8	1.8
K_x, thick till	0.003	0.003	0.003	0.003	0.003
K_z, thick till	0.003	0.003	0.003	0.003	0.003
K_z, fine glacial deposits	0.0018	0.0018	0.0018	0.0018	0.0018
K, streambed	1	1	1	1	1
Ground-water recharge, in inches per year					
Recharge, till	7.4	7.2	8.1	7.6	7.7
Recharge, glacial stratified deposits	22.0	22.0	22.0	22.0	22.0

[1]Grid rotated 45 degrees.

Data Requirements to Characterize Hydraulic Conductivity

The aquifer test in Old Lyme indicated a highly heterogeneous aquifer, but the simulation model indicated that the aquifer could possibly be simulated in several reasonable ways. Although groups of fractures in outcrops and in boreholes can have a well-defined dominant orientation in some areas, individual fractures commonly are continuous only over a scale of hundreds of feet. In a source-water area model, in which regional ground-water flow is simulated, the observed site-specific detail may not be significant. The model could be more realistic if the known detail could be generalized in some way. Based on the spatial structure of well yield discussed previously, it is reasonable to assume that the hydraulic properties of the aquifer can be generalized using knowledge of the geology of an area.

To test whether anisotropy or heterogeneity or both might be detectable in regional water-level data, a model was used to generate four sets of perfect water-level data, with and without anisotropy, and with and without the same heterogeneity as discussed previously (Starn and Stone, 2002; Starn and others, 2002a; Starn and others, 2002b). These data were perturbed using normally distributed random numbers such that the measurements were accurate to within 2 ft at the 95-percent confidence interval. A randomly selected subset of 50 values from each data set was chosen to use as observations in a parameter-estimation model. In both the anisotropic and heterogeneous data sets, the parameter-estimation model correctly estimated the true parameters. A test also was done in which the perfect heterogeneous data were estimated with an anisotropic model. In this case, the parameter-estimation model successfully estimated anisotropy when the true situation was heterogeneous, but could not estimate a heterogeneous distribution of hydraulic conductivity when the true situation was anisotropic.

The result of the simulation of ideal data is that it may be possible to estimate either anisotropy or heterogeneity using parameter estimation, at least in some cases, with a small number (50) of accurate head measurements. It also is possible, however, to mistake true heterogeneity for anisotropy. The converse, that a true anisotropic system would be mistaken for a heterogeneous system, seems to be less likely.

Summary of Effects of Geohydrologic Factors on Source-Water Areas in Connecticut

The first step in delineating source-water areas is to form a site-specific conceptual model of ground-water flow to a well; the concepts discussed below can be used to form this conceptual model. The concepts also are applicable, in a general sense, to estimating source-water areas for conducting aquifer vulnerability assessments. The delineation of areas to be regulated for land use, however, requires site-specific studies beyond the conceptual model, similar to the Level-A mapping requirements in the Connecticut Aquifer Protection Program. Source-water areas for land-use regulation can be effectively delineated using site-specific numerical simulation models; however, the drawback to using simulation models is that they are expensive, time-consuming, and require considerable expertise to construct. Numerical simulation models can use data from many sources, including hydrogeologic mapping, water-level and streamflow measurements, geochemistry, geophysics, aquifer testing, and tracer testing (Risser and Barton, 1995).

Based on the generic models described in this report, factors that affect the dimensions and locations of source-water areas to wells in bedrock can be used to evaluate and modify the conceptual model used with an analytical model. Use of this knowledge requires considerable judgment on the part of the analyst, and there is no unique solution to delineating the source-water area. The bedrock and surficial geology, hydrology, and topography of an area, as well as well construction details, affect the shape and (or) location of source-water areas. Subhorizontal fractures in layered bedrock with a high angle of dip and in nonlayered bedrock do not typically have a great affect on source-water areas. Vertical fracture zones appear to cause elongation of the source-water area in the direction of the fracturing. Vertical fracture zones can be represented in simulation models either explicitly as zones of high hydraulic conductivity or as a bulk property of the bedrock using an anisotropy factor. The presence of a surficial aquifer can cause highly complex ground-water-flow paths.

The hydrology of an area, particularly the ground-water recharge rate, has a great effect on the size of the source-water area. The size of the source-water area for a given well can be estimated using the pumping rate for the well and the recharge rate, as discussed previously. Although information exists on estimating recharge rates, there is some question about how to represent recharge to bedrock rather than to the water table. The recharge rate to bedrock may depend on pumping conditions and may be different under stressed and nonstressed conditions.

Topographic position has an effect on the shape and location of the source-water area. The source-water areas will tend to be uphill from the well. Although this is not a quantitative statement, the idea can be used to approximate source-water areas more accurately in some settings than can be done by assuming a circular radius, as is done in the analytical models presented in this report.

Evaluation of Source-Water Areas to Bedrock Wells in Upland Areas

Topography plays a key role in estimating ground-water-flow patterns in bedrock. Water in upland systems comes from the direction of the ground-water-flow divide. The difficulty, and where the judgment of the analyst is critical, is in deciding where the ground-water basin boundaries are. In this study, it was assumed that ground-water basin divides exist at the scale of subregional drainage basins. This may not be true in areas where there is a large amount of topographic relief, a highly permeable fracture near a basin divide, or in small basins with low relief. These cases can only be found through site-specific studies. Continued research in this area would help to refine this conceptual model.

The source-water area also may be affected by predominant fractures in the bedrock. Although specific fracture properties can only be obtained through site-specific studies, the general fracture direction can be estimated using information presented in figure 7. Wells in areas that have a predominantly high angle of dip (fig. 4) may be directly connected to sources of water at the land surface. Wells in areas that have a predominantly low angle of dip (fig. 5), and where the vertical fractures are poorly developed, may be poorly connected to the land surface, and the source of water to these wells may be diffuse and therefore originate at some distance from the well. For example, Lyford and others (2003) found that where subhorizontal fracturing is coincident with foliation, source-water areas can be some distance from the well and can be outside the topographic boundaries of the overlying surface-water basin. The effect of steeply dipping fractures, whether considered as individual fractures or as a bulk property of the bedrock, is to elongate the source-water area in the direction of fracturing. The elongation seems to be most pronounced where the fracture direction is coincident with the hydraulic gradient and where the fractures are extensive.

To test these concepts in a “real” setting, a numerical simulation model in Old Lyme was constructed and used to delineate source-water areas to four hypothetical wells (fig. 35). Details of the Old Lyme study are provided in the section “Geohydrologic Investigation in Old Lyme, Connecticut: A Case Study.” The source-water areas from the simulation model are evaluated with respect to source-water areas from the analytical methods and simulation models discussed above. The model also is used to investigate the effect of hydraulic anisotropy on source-water area delineation, and if present, how it might be quantified. The Old Lyme model area includes steeply dipping layered bedrock overlain by some areas having no surficial aquifer and some areas having a surficial aquifer. The simulation model shows that it may be possible to estimate anisotropy with a small number of accurate head measurements. It is also possible, however, to mistake true heterogeneity for anisotropy.

For comparison with the source-water areas delineated using conceptual models, circular source -water areas were calculated using equations (2) and (3) (fig. 35). The ground-water recharge using equation (3) for the four hypothetical well locations in Old Lyme ranges from 9.4 to 15.4 in/yr. The recharge rates estimated using parameter-estimation models in Old Lyme ranged from 7.2 in/yr to 8.1 in/yr. The lowest estimated recharge rate (7.2 in/yr) yields a source-area radius (assuming a circular shape) of 982 ft, and the highest recharge rate (15.4 in/yr) yields a source-area radius of 673 ft. Circles of 500 and 1,000-ft radii, which bracket the analytical results, are shown on figure 35.

The source-water area to the well on a hilltop (green well on fig. 35) is on the hilltop near the well. Although the true shape of the source-water area is not circular, most of the particles lie within the outer circle calculated using the analytical methods discussed previously. Because there is little area uphill from the well, the well must receive water from the general area around the well. With an appropriate increase in radius to account for uncertainty in the estimation approach, a circular area could be used to approximate the source-water area for this hypothetical well. The uncertainty inherent in this approach is unknown, however, and the extent to which the radius is increased depends on the judgment of the analyst. The shape of the source-water area could be affected by the layering and dip of the bedrock in the manner discussed in the previous section. The shape also could be affected by upland wetlands and other sources of water. Where upland sources of water are in the possible source-water area, consideration should be given to including these surface-water bodies in the source-water area.

Source-water areas to wells on hillsides (red and orange wells on fig. 35) are not centered at the well, but are uphill from the wells. These source areas are not delineated accurately using the analytical methods discussed previously. These wells capture ground water that flows down the hydraulic gradient, which in this case mimics the slope of land surface, from the recharge areas on the hilltop to discharge areas in the valley bottom. The source-water area is at the hilltop and the width of the source-water area is inversely proportional to the hydraulic conductivity of the bedrock and the slope of the hydraulic gradient.

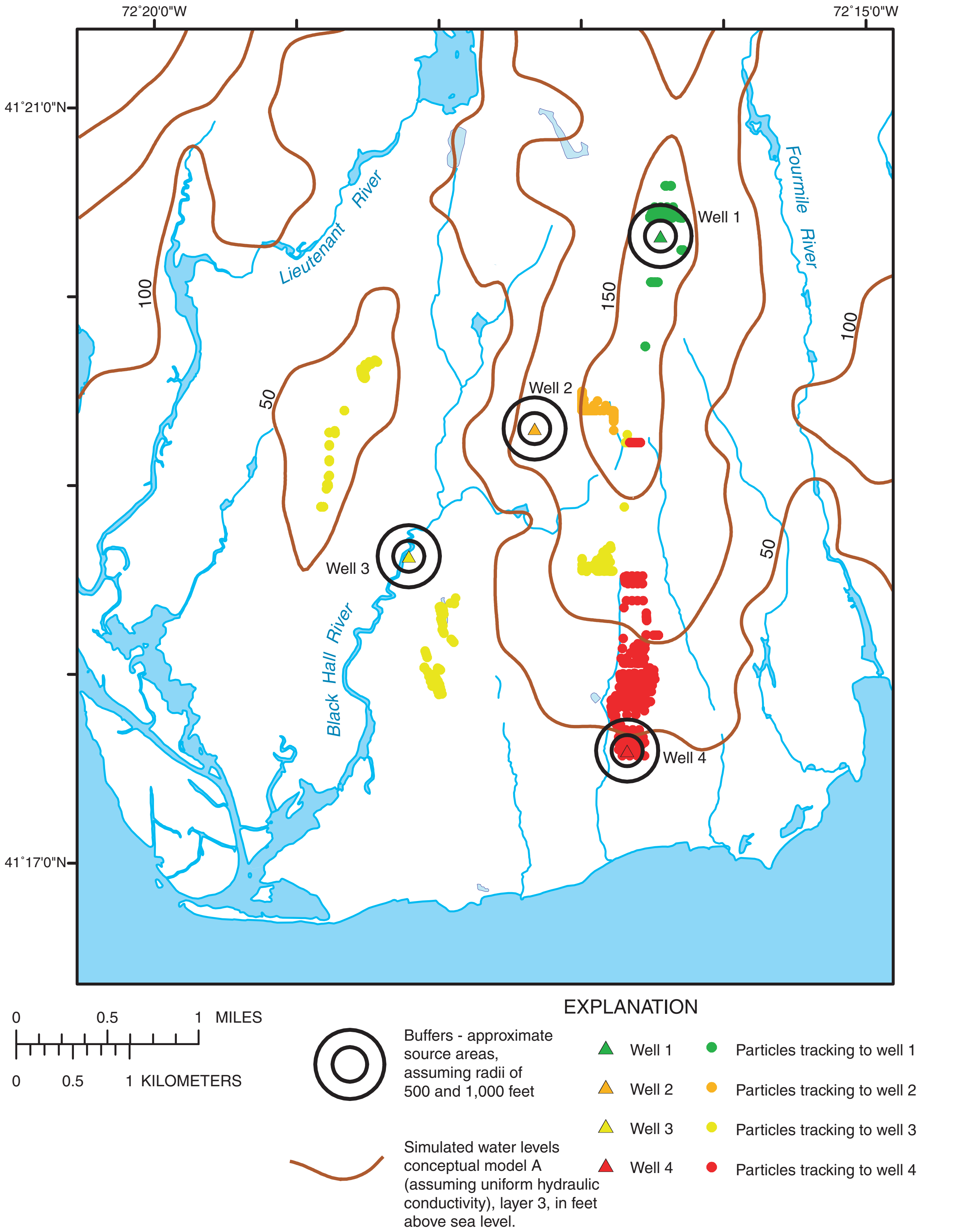


Figure 35. Source-water areas for four hypothetical wells determined using analytical models and a numerical simulation model, Old Lyme, Connecticut. (See fig. 1 for location of Old Lyme.)

As previously discussed, the subregional drainage divide is hypothesized to be the ground-water-flow divide. The extent of ground-water-flow systems in fractured rock needs further study to confirm or reject this hypothesis.

The source-water area to the hypothetical well in the valley bottom (yellow well on fig. 35) is more complex than in the preceding cases. As in the generic simulation models, source-water areas to wells in valley bottoms can extend to drainage divides on opposite sides of the basin. Complex surface-water drainage patterns, which may include wetlands, lakes, reservoirs, and (or) coastal water bodies, can complicate the map pattern of these source-water areas. In these situations, more site-specific numerical simulation modeling based on the types of data discussed in this report is advisable.

The simulation model at Old Lyme also was used to test the effect of different assumptions about fracture-related properties. Based on the analysis of geohydrologic factors in this report, the fractured-rock aquifer at Old Lyme could be simulated with (A) uniform hydraulic conductivity; (B) higher hydraulic conductivity under valleys than under hilltops, as has been suggested by several previous studies (for example, Daniel and others, 1997; Tiedeman and others, 1997; Moore and others, 2002); (C) different hydraulic conductivity in the strike direction of steeply dipping layering; and (D) combination of (B) and (C). Measures of model calibration indicated that all four models were unbiased and were equally likely to be true given the calibration data that were used.

Source-water areas for each topographic setting using the four different conceptual models were very similar to one another, except when the aquifer is simulated as having a preferred orientation of hydraulic conductivity (conceptual model C). When a preferred orientation of hydraulic conductivity is assumed, simulated source-water areas were elongated in the direction of the principal hydraulic conductivity. Model C is reasonable because it estimates a direction of higher hydraulic conductivity in the strike direction of layering as expected and because the hydrologic data available support it. Model C also makes the most difference in the estimated source-water areas; therefore, it is important to know if and when to use a conceptual model like model C.

Evaluation of Source-Water Areas to Bedrock Wells in the Central Lowland

In much of the Central Lowland, thick surficial aquifer materials generally overlie gently dipping sedimentary rocks. Prominent bedding-plane partings form the dominant horizontal connection between a well bore and its source-water area (Stone and others, 1996; Senior and Goode, 1999; Goode and Senior, 2000). Vertical connection between bedding-plane partings is provided by steeply dipping fractures that dip perpendicular to bedding planes. This combination of fractures results in a stair-step type of flow pattern in which water flows downward through near vertical fractures and horizontally through bedding-plane partings. The degree of connection among bedding-plane partings probably varies locally.

Estimating source-water areas to wells in sedimentary bedrock can be done using the dip of bedding-plane partings and the general direction of regional ground-water flow. Where the regional flow is in the direction of dip, ground water most likely flows down-dip along bedding-plane fractures to the pumped well. Ground water can flow horizontally in the surficial aquifer before entering the fracture. Where ground-water flow is opposite the direction of dip, the source is likely to be closer to the well. In both cases, the source-water area is likely to be somewhat elongated along the strike of the bedrock.

The plane of a fracture can be projected from its intersection with the well bore to the bedrock surface by

$$PD = DF/\tan(dip)\,, \tag{5}$$

where

PD is the projected distance from the wellhead to the intersection of the fracture and the bedrock surface, in feet,

DF is the depth of the fracture in the well bore, in feet, and

Dip is the dip, in degrees, of the fracture.

This formula assumes that the fracture remains in a plane throughout its extent, which it may not. Only a general approximation of the distance from where the fracture intersects the well to where the fracture intersects the surficial aquifer can be estimated this way. Sedimentary rocks in Connecticut typically dip from 5-15^{o}. On the basis of equation 5, if a fracture at the bottom of a 400 ft deep well were projected to the bedrock surface, it would intersect the bottom of the surficial aquifer approximately 1,500 to 4,600 ft from the wellhead.

Summary and Conclusions

Understanding ground-water flow in fractured bedrock is important for delineating source-water areas for bedrock wells. In this study, conducted by the U.S. Geological Survey in cooperation with the Connecticut Department of Public Health, the effect of geohydrologic factors on ground-water flow in metamorphic and sedimentary bedrock in Connecticut was assessed. The geohydrologic factors included fracture types and fracture geometry, which are important factors in assessing source-water areas to bedrock wells because water moves through and is stored in fractures in bedrock as it moves to water-supply wells. The character and thickness of unconsolidated (predominantly glacial and postglacial) sediments overlying bedrock also are important factors in assessing source-water areas, because these materials store ground water that can recharge the bedrock. Other factors, such as well construction details, recharge rate, and topographic position also affect the dimensions and locations of source-water areas.

Three principal types of fractures are present in the bedrock of Connecticut—layer- parallel fractures, unroofing joints, and cross fractures. When layering is present in a rock unit, either as bedding in sedimentary rock or as foliation and (or) layering in metamorphic rocks, layer-parallel fracturing usually dominates the fracture geometry because it is pervasive throughout the rock. The orientation and dip of layering, where layering is present, is important for understanding ground-water flow in bedrock. Unroofing joints provide lateral connections between steeply dipping fractures and occur as separate fracture sets predominantly in nonlayered rocks and in layered rocks that are steeply dipping. Cross fractures and joints provide vertical connection with overlying surficial materials and between major subhorizontal water-bearing zones. Cross fractures and joints generally are more widely spaced than layer-parallel fractures and their orientation commonly is related in some way to the primary structure in layered rocks.

Ground-water flow was simulated to bedrock wells in three settings to show how different combinations of geohydrologic factors affect size, shape, and location of source-water areas. Based on generic simulation models, geohydrologic factors, together with ground-water recharge, topographic position, and well- construction details, can be used to make inferences about the size, shape, and location of source-water areas. Source-water areas to wells reflect the general ground-water-flow direction, which generally mimics the land-surface topography.

The three settings were (1) hilltops and hillsides with no surficial aquifer, (2) a narrow valley with a surficial aquifer, and (3) a broad valley with a surficial aquifer. The source-water area to the hilltop well was not greatly affected by simulated fracture zones, except for the extensive vertical fracture zone. The source-water area to the hillside well was not greatly affected by the fracture zones except for the combination of a subhorizontal fracture zone and low bedrock vertical hydraulic conductivity, as might be the case where an extensive subhorizontal fracture zone was not connected to the surface through vertical fractures.

Source-water areas for a well in a narrow valley with a surficial aquifer have complex flow paths. The typical flow path originates in the uplands and passes through either till or bedrock into the surficial aquifer. Although only a small area of the surficial aquifer actually contributes water to the well, most of the water may originate in the surficial aquifer. The source-water area in the uplands can include substantial area on both sides of a river. Source-water areas for wells in this setting were affected mainly by ground-water recharge rate and by the degree of anisotropy. A reduction in the rate of ground-water recharge causes a larger percentage of the source-water area to be in the glacial stratified deposits than in the uplands. This result underscores the need to have better information on the rate of recharge to bedrock from till under both pumping and non-pumping conditions. The simulation of anisotropy produced a source-water area that was elongated in the direction of principal hydraulic conductivity.

Source-water areas for wells in a broad valley with a surficial aquifer (bedrock with a low angle of dip) are greatly affected by the fracture properties. The effect of a fracture is to channel the water downward from the surficial aquifer toward the open borehole. If leakage takes place through the vertical fractures, the source-water area is less affected by the fracture geometry. A fracture near the top of bedrock in the well allowed water to flow from the water table to near the well, as in the case that simulated lack of a good seal between the surficial aquifer and the borehole.

A numerical simulation model in Old Lyme, Connecticut, was constructed and was used to delineate source-water areas to four hypothetical wells, given a realistic set of geologic, hydrologic, and topographic factors. The source-water areas delineated with the Old Lyme numerical model were evaluated with respect to the source-water areas delineated from the analytical methods and simulation models of the three settings. The source-water area to a hypothetical well on a hilltop was roughly circular. Because there is little area uphill from the well, the well must receive water from the general area around the well. The source-water areas to hypothetical wells on hillsides could not be delineated by a circular shape, although the ground-water-flow paths generally follow the land-surface topography. The source-water area to the hypothetical well in the valley bottom is more complex than in the preceding cases. As in the simplified simulation models and the Old Lyme simulation models, source-water areas to wells in valley bottoms can extend to drainage divides on opposite sides of the basin. Complex surface-water drainage patterns, which may include wetlands, lakes, reservoirs, and (or) coastal water bodies, can complicate the map pattern of source-water areas. In these situations, more site-specific numerical simulation modeling, based on the types of data discussed in this report, would provide greater confidence in the results.

The simulation model at Old Lyme also was used to test the effect of different assumptions about fracture –related properties. Based on the analysis of geohydrologic factors in this report, the fractured-rock aquifer at Old Lyme could be simulated as (A) a homogeneous medium; (B) with higher hydraulic conductivity under valleys than under hilltops, as has been suggested by several previous studies; (C) with different hydraulic conductivity in the strike direction of layering; and (D) a combination of (B) and (C). Source-water areas for each topographic setting using conceptual models A–D were very similar to one another, except when the aquifer is simulated as having a preferred orientation of hydraulic conductivity (conceptual model C). When a preferred orientation of hydraulic conductivity is assumed, modeled source-water areas were elongated in the direction of the principal hydraulic conductivity. Model C is reasonable because it estimates a direction of higher hydraulic conductivity in the strike direction of layering, as expected, and because the hydrologic data available support the conceptual model. Model C also makes the most difference in the estimated source-water areas; therefore, it is important to know if an aquifer has a preferred orientation of hydraulic conductivity.

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Appendixes 1 - 2

Appendix 1. Synthesis of previous investigations.

Appendix 2. Description of lithogroups and surficial units.

Appendix 1. Synthesis of Previous Investigations

Previous investigations have yielded many results, sometimes seemingly contradictory, that were used in this study to develop conceptual models. The following paragraphs describe (1) simulation of regional ground-water flow in crystalline bedrock, (2) simulation of source-water areas to wells in crystalline bedrock, (3) simulation of ground-water flow in sedimentary bedrock, (4) studies of well yields, and (5) geohydrologic studies that focused on fracture-domain mapping.

Simulation of Regional Ground-Water Flow in Crystalline Bedrock

Studies at the USGS Toxic Substances Hydrology fractured-rock site in Mirror Lake, New Hampshire, have produced many useful techniques and concepts for ground-water flow in crystalline rocks (Hsieh and others, 1993; Shapiro, 1993). Tiedeman and others (1997) used a ground-water-flow model to test different hypotheses about the fractured-rock aquifer at Mirror Lake. They found that model fit was not improved by introducing vertical anisotropy or variation of hydraulic conductivity with depth. Model fit might have been improved by hypothesizing a lower hydraulic conductivity beneath hillsides and hill slopes than in valleys, but more data were needed to confirm this result. The study by Tiedeman and others (1997) also showed that the ground-water basin was significantly larger than the overlying surface-water basin and that about half the flow from the surficial glacial deposits was downward through the bedrock.

Lyford and others (1998; 1999) and Hansen and others (1999) extensively studied a site in Meddybemps, Maine. The hydrogeology of this site was characterized and ground-water flow was simulated by explicitly including a known fracture in the model. Leakage from surficial glacial deposits and from surface-water infiltration were cited as important factors in the ground-water-flow system.

Daniel and others (1997) conducted a modeling study of fractured crystalline rock in North Carolina that is relevant to New England. They used statistical relations among well yields, rock type, and topography to develop a conceptual model and a numerical simulation model of their study area. They found that local flow systems, which lie between adjacent topographic divides that range from a few thousand feet to a few miles apart, contain 95 percent of the ground-water flow through the bedrock. A base-flow study showed a narrow range of unit discharges, indicating a relatively uniform contribution of ground water to streamflow. This was assumed to mean that the hydrologic properties of the aquifer materials were areally uniform. A chemical analysis of base flow showed that most base flow is derived from shallow bedrock, follows short flow paths, and has low concentrations of chemical constituents. They used a technique similar to that used in the Mirror Lake model. In their model, hydraulic conductivity decreased with depth and varied horizontally by topographic position. The assignment of different hydraulic conductivities according to topography resulted in an apparent anisotropy in the model because hills and ridges in the area have a preferred orientation.

Based on experience at Mirror Lake and other sites, Shapiro (2002) concluded that bulk properties of the bedrock can be used to simulate regional ground-water flow and that regional models of ground-water flow in bedrock can provide likely scenarios of source-water areas to bedrock wells. Near water-supply wells, the spatial distribution of highly permeable fractures in the bedrock affects potential sources of water. Farther away from water-supply wells, highly permeable fractures act like collectors of water, rather than the source of the water. To obtain sufficient yields for water supply, water must be withdrawn from a fracture that is connected to a surface-water body or from a large volume of rock.

Simulation of Source-Water Areas To Wells in Crystalline Bedrock

The area that is the source of water to wells is the product of many complex processes, many of which occur in the subsurface; therefore, source-water areas cannot be directly observed, and they must be estimated. Estimating source-water areas can be an iterative process, with the most defensible and accurate delineations based on integrated data from a variety of sources. Risser and Barton (1995) outlined a strategy for estimating source-water areas in which an initial conceptual model is developed using published information about the geology and hydrology of the area. Information that can be used to develop the initial conceptual models for wells is available in Risser and Barton (1995), including the type of bedrock, the strike and dip of the primary structure of the bedrock, estimates of yield and (or) hydraulic conductivity, and ground-water recharge rates. In some cases, well-completion reports at individual wells also provide estimates of the depths of water-producing fractures. The strategy developed by Risser and Barton (1995) is to refine the initial conceptual model with collection of hydrogeologic data, including (1) geologic maps, (2) water-level and streamflow measurements, (3) geochemical analyses, (4) borehole-geophysical logs, (5) aquifer tests, and (6) tracer tests. Items (1), (2), (4), and (5) were used in a case study in Old Lyme, Connecticut and are discussed in the earlier section "Geohydrologic Investigation in Old Lyme, Connecticut: A Case Study." Barton and others (1999) used this strategy to delineate the source-water area to a water-supply well in crystalline bedrock in Pennsylvania and concluded that numerical flow modeling allows the most accurate representation of the fractured bedrock. Lipfert and others (2001) developed numerical models of ground-water flow using a strategy similar to Risser and Barton (1995). They formed a preliminary conceptual model using existing data, including photo-lineament analysis and fracture measurements at outcrops.

Numerical simulation models of source-water areas have been documented in several studies in New England. Two recent studies used numerical models specifically for estimating source-water areas in fractured crystalline rock. Lyford and others (2003) used hydrogeologic data as discussed above to calibrate a numerical ground-water-flow model and estimate source-water areas to three public-water supply systems in Massachusetts. Two of the systems are in bedrock with prevalent steeply dipping fractures that provide a close hydraulic connection between ground water in the glacial deposits that overlie bedrock and streams and wetlands. In these wells, with well-field yields of 250 to 780 gal/min, the source-water areas were contained within ground-water-flow divides. In both bedrock systems, zones of highly fractured rock parallel to the primary structure of the bedrock were identified by analysis of water-level data and an aquifer test. At one well field, discrete sets of subhorizontal and near-vertical fractures were identified, whereas at the other system, a dense network of fractures was identified. Simulation indicated that pumping at one system (with a combined yield of 780 gal/min) lowered water levels in a wetland up to 2,000 ft from the wells. The third site studied by Lyford and others (2003) differed from the first two in that two wells at the site received water from fractures at depths greater than 500 ft below land surface. Bedrock at this site had a low angle of dip such that water-bearing fractures intersected land surface outside the surface-water basin in which the wells were located. The lack of vertical fractures at this site make the vertical hydraulic conductivity in the model, which controls the amount of recharge that reaches the water-bearing zones, a critical piece of information for accurate source-area estimation.

Some techniques that may be useful in defining source-water areas in complex hydrogeologic settings are regional ground-water-flow simulation models calibrated using parameter estimation with water-level and streamflow data, and variogram estimation. Some investigators have used numerical simulation models to estimate uncertainty in the estimated source-water area. Lipfert and others (2001) combined the results of simulations that made a variety of assumptions to delineate areas of low-, medium-, and high-confidence zones within the source-water area. Lyford and others (2003) used sensitivity analyses to choose the best estimate of source-water area from among many alternatives. Starn and others (2000) used parameter estimation to evaluate the ability of aquifer-test data in a glacial aquifer to distinguish among models that hypothesized a permeable fracture in the underlying bedrock. Starn and others (2000) also used parameter estimation with a Monte Carlo simulation of source-area uncertainty caused by uncertainty in estimated model parameter values. All these techniques are useful and should be considered when applying numerical simulation models to source-water area problems.

Simulation of Ground-Water Flow in Sedimentary Bedrock

Sedimentary bedrock in the Northeast is structurally less complex than crystalline bedrock, but conditions are similar in that ground-water flow takes places primarily within fractures. Several investigations in the Mesozoic-aged rocks of the Newark Basin of New Jersey and Pennsylvania and the Hartford Basin in Connecticut have yielded useful insights. A series of studies done in Pennsylvania illustrates how fractured rocks are modeled as a function of the type of data available. Goode and Senior (1998) analyzed aquifer-test data and concluded that, "some evidence of well-field-scale horizontal anisotropy exists, with maximum transmissivity aligned with the regional northeast strike of bedding, but this evidence is weak because of the small number of observation wells, particularly wells screened in isolated depth intervals." In a subsequent study, Senior and Goode (1999) constructed a regional ground-water-flow model of the same area and concluded that, "the regional anisotropy ratio for the sedimentary rock aquifer is about 11 to 1, with permeability greatest along strike." In that study, fractures were not explicitly simulated. Goode and Senior refined their model of the area (2000) and explicitly included high-permeability dipping beds. Although "this model structure yields ground-water-flow patterns characteristic of anisotropic aquifers and preferred flow is in the strike direction," the contributing area simulated by the regional model was more elongate along strike than in the refined model. The contributing area simulated by the more refined model is more detailed in shape and extends in a different upgradient direction than in the regional model. The difference in the contributing areas is caused by the treatment of preferential flow directions in each model and by the differences in model grid resolution.

Stone and others (1996) characterized fracture geometry in interlayered sandstone and siltstone of the New Haven Arkose in Cheshire, Conn. Borehole-geophysical logging in four bedrock wells and observations at outcrops in the area indicated that water-bearing fractures included N-S striking, east-dipping (about 20^{o}) layer-parallel partings and N-S striking, west-dipping (about 70^{o}) cross fractures. An idealized block of layered sedimentary rock was simulated that included discrete high-permeability fractures at spacing observed in wells and outcrop. The model demonstrated that ground-water-flow paths in fractured rock are complex, three-dimensional shapes.

Bradbury and Muldoon (1993) used a discrete fracture model to investigate the effects of fracture density and orientation on the shape of the contributing area of a well. The discrete fracture model uses the statistical approach to generate realizations of fractures that are simulated with the Monte Carlo technique. Advective particle tracking defines the contributing area. A network of dense, orthogonal fractures produced a contributing area that is similar to the porous-media approach except that the contributing area for the fracture network is wider at the upgradient boundary than is predicted by the porous-media model. In another simulation, a set of NE-SW fractures is intersected at right angles by a less-dense, NW-SE fracture set. The contributing area in this simulation is very large and occupies considerable areas outside the porous-media-derived contributing area. As the scale of the problem becomes larger (beginning at 50-by-50 m), the contributing areas are more closely approximated using a homogeneous, anisotropic medium. The authors conclude that discrete fracture models have limited use in well-head-protection studies, but that they are useful in providing insights into the use of porous-media models. In general, they found that porous-media models would predict contributing areas that are too small. Fractures cause an apparent dispersion in ground-water-flow paths, particularly in anisotropic situations. Significant spreading occurs in the direction of predominant fracture sets.

Studies of Well Yield

Johnson (1999) analyzed fracture measurements made in boreholes at the USGS Mirror Lake research site that corroborate the larger-scale field conclusions. According to that study, the density of fractures is high in shallow zones and drops off rapidly with depth, and there seems to be a higher density of fractures on hilltops and hillsides than in valley bottoms, possibly because of glacial scouring of the valleys. There is a higher fracture density in granitoids than in schist and gneissic rocks, and other igneous rocks are relatively unfractured. In metamorphic rock, fractures were observed to be parallel to foliation. Hydraulic conductivity seemed to be somewhat independent of fracture characteristics such as fracture density. Hydraulic conductivity ranged over 6 orders of magnitude and did not correlate with altitude or depth. The magnitude of hydraulic conductivity did not correlate with lithology or with direction of fracturing. Because there is a higher probability of finding a fracture in granitic rocks, the probability of finding a high hydraulic conductivity fracture also is higher, compared to other rock types studied.

Daniel and others (1997) created hydrogeologic units based on the hypothesis that the origin, composition, and texture (in other words, the qualities that make up traditional geologic map units and formations) are related to the susceptibility of the rocks to develop secondary porosity. The hydrogeologic characteristics of hydrogeologic units were determined from well data compiled from various sources. The characteristics that were used were yield, yield per foot of depth, specific capacity, and depth. The well data show little variation among hydrogeologic units. Differences in well yield among topographic position, however, are pronounced: the average yield of wells in valleys is 2.4 times the yield of wells on hills and ridges. The yield for wells on hillsides is intermediate between hilltop and valley bottom wells, and this pattern is consistent across all hydrogeologic units. Well yield and depth data indicated that the maximum depth of the flow system was about 850 ft below land surface and that open interconnected fractures were more abundant and persisted to greater depths beneath valleys than under hills.

Knopman and Hollyday (1993) reported that lithology explains about 24 percent of the variation in specific capacity when considered alone. When combined with other factors, lithology can explain more of the variation. Hansen and Simcox (1994) found results similar to those at Mirror Lake, but through analysis of well reports rather than modeling. They found that thickness of overburden played a large role in increased yields in valleys, but not on hilltops. This may be the result of the combination of factors that affect yield in valleys, or because the composition of the overburden is different (till with small saturated thickness on hilltops compared to coarse-grained glacial stratified deposits with large saturated thickness in valleys). An analysis in which wells were segregated by major lithologic units indicated a similar set of relations to the overall data set, except that well yields in crystalline rocks decreased with depth to 500 ft and then increased slightly, but well yields in the Connecticut valley in Massachusetts increased substantially below 400 ft.

Moore and others (2002) developed a multi-factor equation to estimate the probability of locating a high-yield well for the purpose of exploring high-yield wells in New Hampshire. The predicted variable they used was the natural log of well yield as reported by well drillers. They found an apparent decrease of yield with deeper wells, a relation that does not make physical sense, and, indeed, was found to be a function of water demand. If the bedrock yield was high, drilling was stopped at a shallow depth. If the bedrock yield was low, drilling continued to a greater depth to provide more well-bore storage. Depth was related to water use (commercial wells have higher yields than domestic wells because of the greater demand), and that yield also was correlated with year drilled (water demands have been increasing over time), well driller, and median household income. A regression model was used to determine the relation among other variables and well yield. The other factors that were significant in the regression included topographic factors, major lithologic groups, lineaments, detailed quadrangle-scale lithologic groups, and less detailed state-map-scale lithologic groups. The results are largely in accordance with previous studies. Well yields are higher for lower (flatter) slopes, lower altitudes, shorter distances to water bodies, and larger uphill drainage areas. The combined topographic/lithologic variable was significant only in one category—valley bottoms in two specific gneiss and granite units (Massabesic Gneiss Complex and Breakfast Hill Granite). Wells on concave downward topography had higher yields than wells on concave upward topography, the rational being that concave downward areas are more easily eroded and indicate fracture zones or areas that collect more water, whereas concave upward areas indicate hills that are resistant to erosion. Other significant variables were indicator variables that can be compared based on the magnitude of the coefficient, but the effects are additive because a given location may be described by several different rock types at different scales. Of seven major lithologic groups, only one was significant—foliated plutons—which were associated with lower well yields. This group covers about one-quarter of the area of the state. Of the many lineament groupings tested, only three were significant, the most significant being fracture domain-correlated lineaments identified on 1:80,000 high-altitude aerial photographs. Discrete fracture-correlated lineaments in plutons and statewide lineaments from N0^{o}W to N40^{o}W were also significant. Although most of the major lithologic groups were not significant, many of the individual formation and member scale units identified on the state geologic map and the detailed quadrangle scale maps in two quadrangles were significant.

Drew and others (2001) also developed a conceptual model of well yield as a function of depth, elevation, and rock type in the Pinardville quadrangle, New Hampshire. They defined a high-yield well as one with a yield more than 40 gal/min, and a low-yield well as one with a yield less than 0.5 gal/min. The percentage of high- and low-yield wells increased from 1984 to 1998, possibly as a function of the trend in housing development. The development trend in their study area was for more wells to be drilled at higher elevations. During the same time period, well depths increased. As wells were drilled deeper, the chance of encountering a high-yielding fracture increased, and the percentage of high-yield wells increased. They did not know why the percentage of low-yield wells increased, but speculated that it may have been because of differences in geology at higher elevations. The relation between depth and elevation is significant; however, neither factor is significant with respect to yield.

Studies of Fracture-Domain Mapping

Mabee and Hardcastle (1994; 1997) used the fracture-domain mapping technique to show that outcrop data could be used to predict the orientation of fractures found in boreholes using acoustic televiewer data. The orientation of the fracture families was shown to be hydraulically significant. The orientation, trace length, spacing, planarity, roughness, and nature of terminations of fractures were measured at 79 outcrops in a 100-m radius from the well field. Fracture families were plotted on a map of the site and areas with similar orientations were combined into fracture-domain map units. Fracture domains were shown to be continuous, overlapping, and fairly homogeneous. The most significant aspect of the fractures, hydraulically, was the mineralization that led to a much lower hydraulic conductivity. The rocks in this study were mainly granitic rocks cut with quartzo-feldspathic veins (veins were included in the fracture population studied); many of the most productive water-bearing zones were subhorizontal unroofing fractures. The results of this study indicated that fracture-domain mapping may be a promising technique for analyzing flow in fractured rock, but there are many potential limitations. It is not really known, for example, how extensive or continuous fracture domains are with depth, nor how much variability can be expected within fracture domains.

Walsh and Clark (2000) and Drew and others (1999) have combined ideas from the fracture-domain mapping with regional geologic mapping. Their work shows that there are large-scale geologic units that seem to have a relation to well yield.

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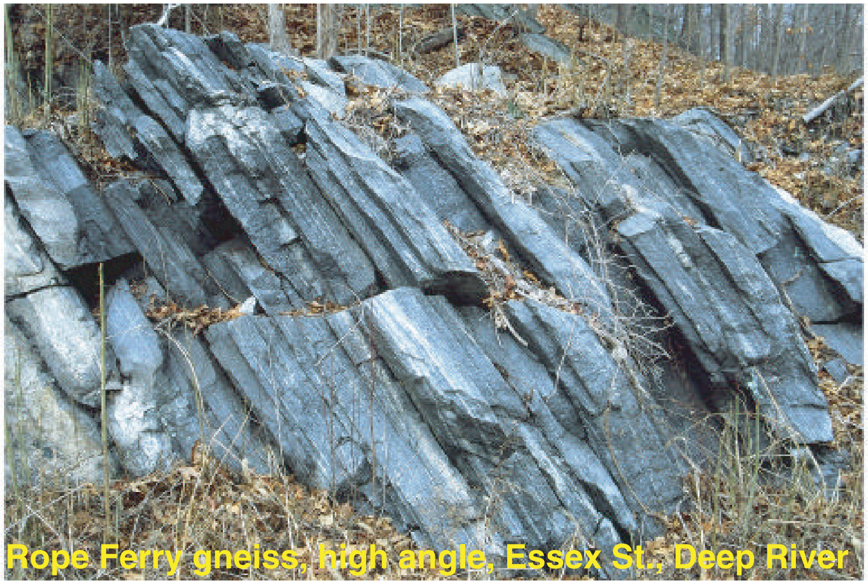
Appendix 2. Description of Lithogroups and Surficial Units

Bedrock formations from the Bedrock Geologic Map of Connecticut (Rodgers, 1985) have been categorized into nine lithogroups, based primarily on rock type (lithology). Seven lithogroups include metamorphic rocks of Paleozoic and Proterozoic age (lithogroups GN, SCH, MIX, MBL, GR, GRL, and MF), and two lithogroups include sedimentary and igneous rocks of Mesozoic age (lithogroups SED and BAS).

Surficial materials from the Surficial Materials Map of Connecticut (Stone and others, 1992) were grouped into two major categories that represent the presence or absence of thick unconsolidated sediments. The surficial aquifer category includes units CS, FS, and TT; the no-surficial aquifer category is the unit T. Photographs and descriptions of the bedrock lithogroups and surficial units are presented in this appendix.

As previously noted, names of geologic formations used in this report correspond to those used on the Bedrock Geological Map of Connecticut (Rodgers, 1985) and may not match the geologic names used by the U.S. Geological Survey.

GN (Gneiss and Quartzite)
SCH (Schist and Phyllite)
MIX (Mixed Gneiss and Schist)
MBL (Marble)
GR (Granitic and Dioritic Rocks)
GRL (Variably layered Granitic Rock)
MF (Mafic and Ultramafic Rocks)
SED (Sedimentary rock)
BAS (Basalt and Diabase)
CS (Coarse Stratified Deposits)
FS (Fine Stratified Deposits)
TT (Thick Till)
T (Thin Till)



Rope Ferry gneiss, high angle, Essex St., Deep River

Waterford Group, low angle, Rt. 32, Willimantic

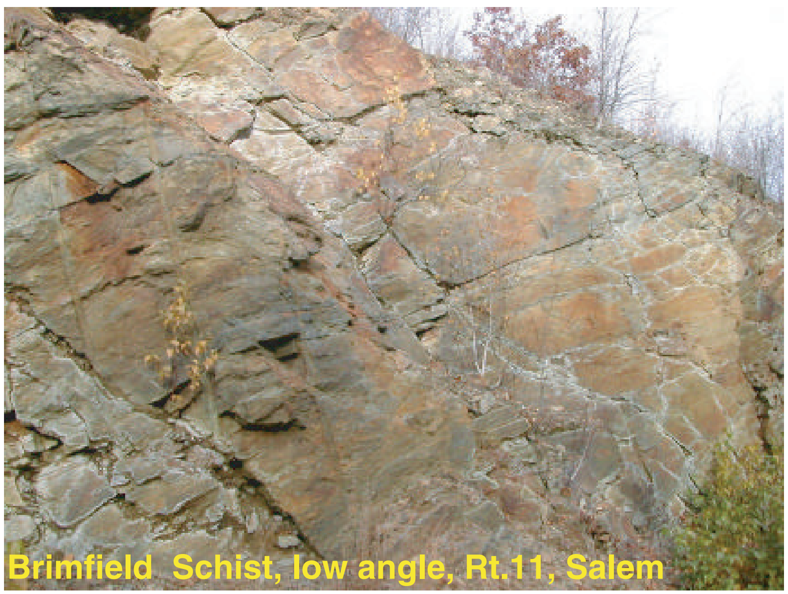
GNEISS AND QUARTZITE (lithogroup GN)

Resistant, well-foliated metamorphic rocks including orthogneiss and paragneiss, and including some hornblende gneisses and thin amphibolite units. More or less distinct banding is commonly caused by contrasting mineral segregations. Dominance of equidimensional grains results in strong physical bonds between individual folia, which produces a relatively strong rock with well-developed brittle fracturing, both along the layering and across it. Fracturing is generally dominated by foliation-parallel parting. Where foliation dips steeply, foliation-parallel fracturing produces ubiquitous high-angle pathways for water, strongly oriented along the strike direction; in these places, subhorizontal unroofing joints are also well developed (see photo). Where foliation is less steep, foliation-parallel fracturing is well developed and produces low-angle pathways; these low-angle fractures have been utilized by unroofing stress as well. High-angle cross fractures generated by tectonic stresses occur less frequently than foliation-parallel fractures. These rock units generally form ridges that trend along the strike of layering in the rock.

Includes Cheshire Quartzite, Bristol Gneiss, Harrison Gneiss, Middletown Formation, Monson Gneiss, Quinebaug Formation, Clough Quartzite, gneiss of the Highlands massifs, layered gneiss, Ponaganset Gneiss, Waterford Group, Mamacoke Formation, New London Gneiss, Rope Ferry Gneiss, hornblende gneiss and and amphibolite, and gneissic members of Canaan Mountain Schist, Collinsville Formation, Ratlum Mountain Schist, Southbridge Formation.

(Names of geologic formations used here correspond to those used on the Bedrock Geological Map of Connecticut (Rodgers, 1985)).


Collins Hill Fm., high angle, Rt. 9, Haddam


Brimfield Schist, low angle, Rt.11, Salem

SCHIST AND PHYLLITE (lithogroup SCH)

Less resistant, well-foliated, predominantly metasedimentary rock, and including some metavolcanic greenschist units. Strong foliation is caused by subparallel orientation of platy minerals such as mica and chlorite, which are tabular and flaky. Grain foliation is frequently interrupted by equidimensional minerals such as quartz and garnet; mechanical bonding between individual folia is weak because of the platy mineral cleavage; this results in less well-developed brittle fracturing. These rock units are also less resistant to weathering than the gneissic units because of their mineral composition. Although compositional layering and foliation in schist is well developed, fracturing along the layering is often less frequent than in gneiss, and cross fractures tend to be less through-going. Where foliation dips steeply, foliation-parallel fracturing produces ubiquitous high-angle pathways for water, strongly oriented along the strike direction; in these places, subhorizontal unroofing joints are also well developed (see photo). Where foliation is less steep, foliation-parallel fracturing is well developed and produces low-angle pathways; these low-angle fractures have been utilized by unroofing stress as well. High-angle cross fractures generated by tectonic stresses occur less frequently than foliation-parallel fractures. These rock units commonly underlie valleys and swales that trend along the strike of layering in the rock.

Includes Everett Schist, Hoosac Schist, amphibolite-bearing unit of the Manhattan Schist, Canaan Mountain Schist, Scotland Schist, The Straits Schist, Wepawaug Schist, Rowe Schist, Allingtown Metavolcanics, Collins Hill Formation, Maltby Lakes Metavolcanics, Trap Falls Formation, Walloomsac Schist, Sweetheart Mountain Member of Collinsville Formation, Yantic Member of Tatnic Hill Formation, Scranton Mountain Member of Taine Mountain Formation, upper member of Bigelow Brook Formation, Brimfield Schist.

(Names of geologic formations used here correspond to those used on the Bedrock Geological Map of Connecticut (Rodgers, 1985)).

Hebron Gneiss, low angle, School Rd., East Haddam

Plainfield Formation, high angle, Salt Island, Clinton

MIXED GNEISS AND SCHIST (lithogroup MIX)

Predominantly metasedimentary rock units that include belts of gneiss and schist, and local quartzite layers. Many of the metamorphic rock units on the Bedrock Geological Map of Connecticut include areas of gneiss, schist, and quartzite either as alternating layers or as alternating broad belts. These mixed units are predominantly metasedimentary rocks; schistose units were metamorphosed from fine-grained sedimentary rock (shale and siltstone) and gneissic and quartzitic units were metamorphosed from coarser-grained sandstone and conglomerate. Rock units in the MIX lithogroup display physical properties and fracture characteristics as described for GN and SCH lithogroups; these characteristics are variable from place to place within the lithogroup. Where foliation dips steeply, foliation-parallel fracturing produces ubiquitous high-angle pathways for water strongly oriented along the strike direction; at these sites, subhorizontal unroofing joints are also well developed (see photo). Where foliation is less steep, foliation-parallel fracturing is well developed and produces low-angle pathways; these low-angle fractures have been utilized by unroofing stress as well. High-angle cross fractures generated by tectonic stresses occur less frequently than foliation-parallel fractures. -

Includes Dalton Formation, Manhattan Schist, Waterbury Gneiss, Littleton Formation, Erving Formation, Brimfield Schist, Collinsville Formation, Cobble Mountain Formation, Golden Hill Schist, Oronooque Schist, Black Hill member of Quinebaug Formation, Ratlum Mountain Schist, Taine Mountain Formation, Tatnic Hill Formation, Hebron Gneiss, Southbridge Formation, Fitch Formation, Plainfield Formation, and mixed unit members of Scotland Schist, The Straits Schist, Rowe Schist, Hawley Formation, Trap Falls Formation, Bigelow Brook Formation, and rusty mica schist and gneiss (Ygs).

(Names of geologic formations used here correspond to those used on the Bedrock Geological Map of Connecticut (Rodgers, 1985)).

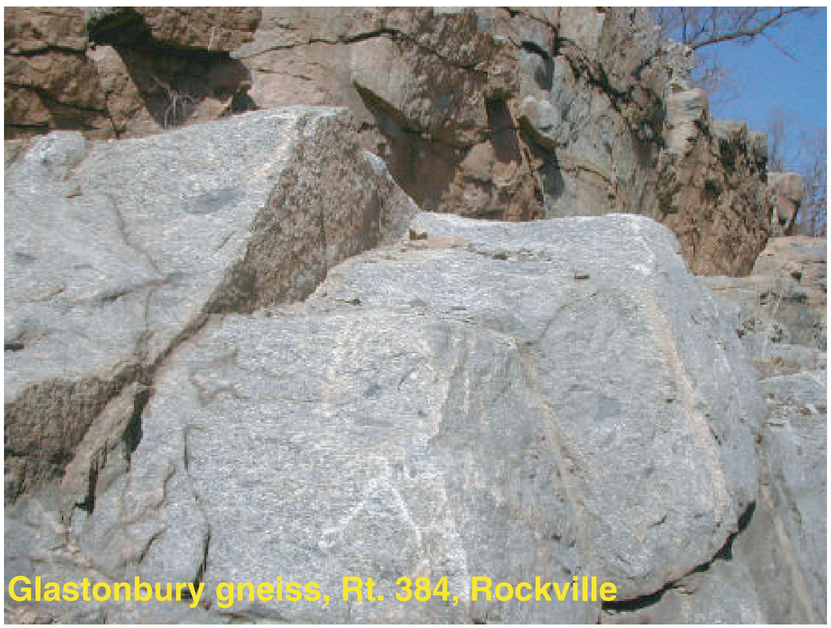
Stockbridge Marble, North Canaan

MARBLE (lithogroup MBL)

Mostly massive marble, locally layered and schistose, metamorphosed limestone and calcareous siltstone and sandstone. Because of its high calcite content, marble is geochemically less resistant than other metamorphic rocks and predominantly underlies the Housatonic Valley and several of its tributary valleys. Locally, thick weathered rock (saprolite) zones overlie the competent bedrock. Fracturing is dominated by subhorizontal unroofing joints, which provide continuous, low-angle pathways for water. High-angle pathways are restricted to tectonically generated fractures and joints that are localized and in some places are rather widely spaced. Where bedding is preserved, it is generally high-angle, but very little fracturing occurs along the layering.

Includes units a, c, d, e, f, and g of Stockbridge Marble and basal marble member of Walloomsac Schist.

(Names of geologic formations used here correspond to those used on the Bedrock Geological Map of Connecticut (Rodgers, 1985)).



GRANITIC AND DIORITIC ROCKS (lithogroup GR)

Intrusive, medium to coarse-grained nonfoliated igneous and foliated metaigneous rocks. Nonfoliated igneous rocks (Narragansett Pier and Westerly Granites) are relatively small intrusive bodies in Connecticut and occupy a very small percentage of the map area. Although meta-igneous units are foliated--that is, they exhibit alignment of mineral grains achieved during metamorphism, these rocks have little compositional layering and hence fracturing along the foliation is rare. These rock units are resistant and generally form highs in the topography. Fracturing is dominated by subhorizontal unroofing joints, which provide continuous low-angle pathways for water High -angle pathways are restricted to tectonically generated fractures and joints that are localized and in some places are rather widely spaced.

Includes Maromas Granite Gneiss, Nonewaug Granite, Brookfield Gneiss, Glastonbury Gneiss, Narragansett Pier Granite, porphyry, Pinewood Adamellite, syenite, Westerly Granite, Augen gneiss, pick granitic gneiss, Stony Creek Granite Gneiss, and Scituate Granite Gneiss.

(Names of geologic formations used here correspond to those used on the Bedrock Geological Map of Connecticut (Rodgers, 1985)).

VARIABLY LAYERED GRANITIC ROCK (lithogroup GRL)

Granitic units in which fracturing developed along foliation planes are variable from place to place within the unit.

Includes Canterbury Gneiss, foliated quartz diorite, foliated granitic gneiss, Ordovician? granitic gneiss, Light House Gneiss, Hope Valley Alaskite Gneiss, Potter Hill Granite Gneiss, and Shelton Member of Trap Falls Formation)

Potter Hill Granite Gneiss, poorly layered, Rt. 12, Old Mystic

Potter Hill Granite Gneiss, well layered, Jericho Rd., Old Lyme

Hope Valley Alaskite Gneiss, New London

Preston Gabbro, Ledyard

MAFIC AND ULTRAMAFIC ROCKS (Lithogroup MF)

Intrusive, nonlayered, fine- to medium-grained igneous and meta-igneous rock units that are dark gray in color due to mafic mineral content. Compositional layering is absent and, therefore no layer - parallel fracturing occurs. Rock is relatively resistant, and forms topographic highs in most areas. Fracturing is dominated by subhorizontal unroofing joints, which provide through-going low-angle pathways for water. High-angle pathways are restricted to tectonically generated fractures and joints that are localized and in some places are rather widely spaced.

Includes Preston Gabbro, Lebanon Gabbro, Litchfield Norite, massive mafic rock in Middletown Formation, lamprophyre, hornblende norite, ultramafic rock, mafic phase of Narragansett Pier Granite.

(Names of geologic formations used here correspond to those used on the Bedrock Geological Map of Connecticut (Rodgers, 1985)).

SEDIMENTARY ROCK (lithogroup SED)

Well-layered arkosic sandstone, siltstone, and conglomerate, generally red-brown in color; locally gray sandstone, siltstone, and black shale. These rock units are easily erodible and underlie the broad Central Lowland of Connecticut and the smaller Pomperaug Basin to the west. Coarse-grained layers (sandstone and conglomerate) have thicker beds (1-3 m) and are generally more resistant; fine-grained units (siltstone and shale) are generally more thinly bedded and less-resistant. In most of the sedimentary rock area, layering dips gently (5-25^0) eastward (northward or southward locally along broad, gentle fold axes). Layer-parallel partings and fissile zones provide through-going low-angle fractures that are major water-bearing pathways in these rocks. Cross-fracturing is typically dominated by paleovertical jointing that strikes N-S (parallel to bedding strike) and dips steeply westward (perpendicular to bedding); less prominent orientations of high-angle cross fractures may also be present. In areas affected by faulting, high-angle fractures are dominated by orientations associated with the faults. Because layering in these rocks is low-angle, stress-relief due to unroofing has occurred along the layering and separate unroofing joints are generally absent.

Includes New Haven Arkose, Shuttle Meadow Formation, East Berlin Formation, and Portland Arkose.

(Names of geologic formations used here correspond to those used on the Bedrock Geological Map of Connecticut (Rodgers, 1985)).

Holyoke Basalt Rt. 66

BASALT AND DIABASE (lithogroup BAS)

Massive basaltic lava flows and shallow diabase dikes and sills that in most places are interlayered with sedimentary rocks and dip gently eastward. These rocks are highly resistant and form narrow topographic highs (traprock ridges) within the broad Central Lowland of the State and in the Pomperaug Basin in western Connecticut. These rock units are highly fractured by cooling joints (columnar jointing). Cooling joints formed perpendicular to tops and bottoms of flows and sills, and are steeply dipping polygonal arrays of high-angle fractures. More widely spaced subhorizontal fracturing is also present, which parallels the top and bottom of flows and provides some low-angle pathways. These rock units are more highly fractured than other rock types in Connecticut because of the pervasive cooling joints, but their areal extent at land surface is relatively small due to the fact that they are thin units (100 - 500 ft) within the much thicker stack (6,000 –10,000 ft) of sedimentary units. Outcrop areas of these rock units are high narrow ridges beneath which the water table is relatively low; this is perhaps indicative of the higher permeability of this rock type. Where basalt occurs at depth below the water table and below sedimentary rock units, it may be a highly productive bedrock aquifer.

Includes Talcott Basalt, Holyoke Basalt, Hampden Basalt, Buttress Dolerite, and West Rock Dolerite.

(Names of geologic formations used here correspond to those used on the Bedrock Geological Map of Connecticut (Rodgers, 1985)).

Coarse Stratified Deposits

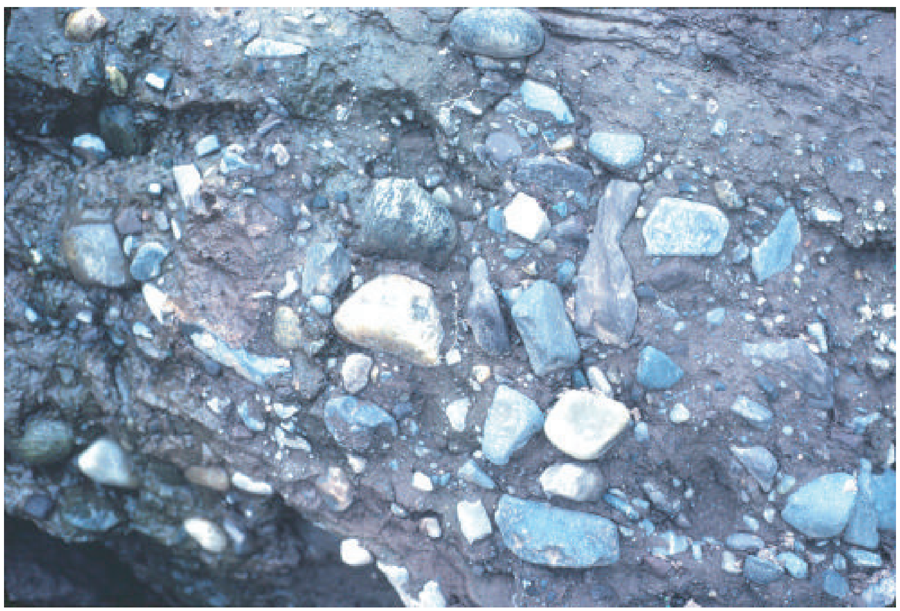
Includes areas of gravel deposits, sand and gravel deposits, sand deposits, and floodplain alluvium where these units make up the entire thickness of surficial materials, and 10 stacked-unit combinations of coarse-grained units (such as sand and gravel overlying sand). Also includes eight stacked-unit combinations in which fine-grained deposits overlie coarse grained units (such as fines overlying sand), and four stack units in which swamp deposits over lie coarse-grained deposits (such as swamp deposits overlying sand and gravel). Also included in this category are beach deposits along the coast and artificial fill that consists of large areas of "made land". Surficial materials in this category range in thickness from a few feet near the edges of these map unit areas to several hundred feet in the thickest sections. An average thickness of 46 ft is indicated in records of approximately 1,900 inventoried bedrock wells penetrating this unit across the State.

Fine Stratified Deposits

Includes areas of very fine sand, silt, and clay deposits, swamp deposits, and salt-marsh deposits where these units make up the entire thickness of surficial materials. Also includes 10 stacked-unit combinations where coarse-grained stratified deposits and (or) floodplain alluvium overlie fine grained deposits (such as sand and gravel overlying sand overlying fines). Surficial materials in this category range in thickness from a few feet near the edges of the map unit area to several hundred feet in the thickest sections; an average thickness of 76 ft is indicated in records of 470 inventoried bedrock wells penetrating this unit across the State.

THICK TILL

Includes areas of till (a nonsorted, generally nonstratified mixture of grain sizes from clay to large boulders) where this material is greater than 15 feet in thickness, typically in drumlins and on the northern or northwestern sides of bedrock hills. An average thickness of 53 feet is indicated in records of 973 inventoried bedrock wells penetrating this unit across the State.



THIN TILL

Includes areas where till is less than 15 feet thick and areas where till is absent and bedrock is at land surface.