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Use of Temperature Profiles Beneath Streams to Determine Rates of Vertical Ground-Water Flow and Vertical Hydraulic Conductivity

By WAYNE W. LAPHAM

U.S. GEOLOGICAL SURVEY WATER-SUPPLY PAPER 2337

DEPARTMENT OF THE INTERIOR
MANUEL LUJAN, Jr., Secretary

U.S. GEOLOGICAL SURVEY
Dallas L. Peck, Director

UNITED STATES GOVERNMENT PRINTING OFFICE: 1989

For sale by the Books and Open-File Reports Section,
U.S. Geological Survey, Federal Center,
Box 25425, Denver, CO 80225

Library of Congress Cataloging in Publication Data

Lapham, Wayne W.
Use of temperature profiles beneath streams to determine rates of vertical ground-water flow and vertical hydraulic conductivity / by Wayne W. Lapham.

p. cm.—(U.S. Geological Survey water-supply paper ; 2337)
Bibliography: p.
Supt. of Docs. No.: I 19.13:2337
1. Groundwater flow—Mathematical models. 2. Soil permeability—Mathematical models. 3. River sediments—Temperature. I. Title. II. Series.
TC176.L37 1989 89-600176
627'.042—dc20 CIP

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Use of Temperature Profiles Beneath Streams to Determine Rates of Vertical Ground-Water Flow and Vertical Hydraulic Conductivity

By Wayne W. Lapham

Abstract

The use of temperature profiles beneath streams to determine rates of vertical ground-water flow and effective vertical hydraulic conductivity of sediments was evaluated at three field sites by use of a model that numerically solves the partial differential equation governing simultaneous vertical flow of fluid and heat in the Earth. The field sites are located in Hardwick and New Braintree, Mass., and in Dover, N.J.

In New England, stream temperature varies from about 0 to 25 °C (degrees Celsius) during the year. This stream-temperature fluctuation causes ground-water temperatures beneath streams to fluctuate by more than 0.1 °C during a year to a depth of about 35 ft (feet) in fine-grained sediments and to a depth of about 50 ft in coarse-grained sediments, if ground-water velocity is 0 ft/d (foot per day). Upward flow decreases the depth affected by stream-temperature fluctuation, and downward flow increases the depth.

At the site in Hardwick, Mass., ground-water flow was upward at a rate of less than 0.01 ft/d. The maximum effective vertical hydraulic conductivity of the sediments underlying this site is 0.1 ft/d. Ground-water velocities determined at three locations at the site in New Braintree, Mass., where ground water discharges naturally from the underlying aquifer to the Ware River, ranged from 0.10 to 0.20 ft/d upward. The effective vertical hydraulic conductivity of the sediments underlying this site ranged from 2.4 to 17.1 ft/d. Ground-water velocities determined at three locations at the Dover, N.J., site, where infiltration from the Rockaway River into the underlying sediments occurs because of pumping, were 1.5 ft/d downward. The effective vertical hydraulic conductivity of the sediments underlying this site ranged from 2.2 to 2.5 ft/d. Independent estimates of velocity at two of the three sites are in general agreement with the velocities determined using temperature profiles. The estimates of velocities and conductivities derived from the temperature measurements generally fall within the ranges of expected rates of flow in, and conductivities of, the sediments encountered at the test sites.

Application of the method at the three test sites demonstrates the feasibility of using the method to determine the rate of ground-water flow between a stream and underlying sediments and the effective vertical hydraulic conductivity of the sediments.

Manuscript approved for publication March 4, 1987.

INTRODUCTION

Background

Stratified-drift deposits of Pleistocene age composed of sand and gravel form the principal aquifers in the Northeastern United States. These aquifers are limited in areal extent and thickness and, consequently, have small recharge areas and storage capacities. However, they have the greatest potential for increased development of water supplies of all aquifers in the Northeastern United States because they are traversed by streams that serve as sources of induced infiltration. One objective of the U.S. Geological Survey's RASA (Regional Aquifer-System Analysis) program in the glaciated Northeastern United States is to understand better the hydrology of these stream-aquifer systems. This objective is being met in part by investigations that evaluate alternative methods of estimating the rate of flow and effective vertical hydraulic conductivity between these aquifers and their overlying streams.

Information about the rate of flow and degree of hydraulic connection between an aquifer and an overlying stream is important in many hydrologic studies. For example, in addition to its use in estimates of potential development of aquifers, it may be required in studies of the chemical interaction between surface and ground water and in studies aimed at predicting movement of solutes in ground water.

Rates of ground-water flow to or from a stream usually are determined by measuring the change in stream discharge along a stream reach (a seepage run). However, accurate determination of the rate of ground-water flow using change in stream discharge along a reach requires a high ratio of gain or loss of stream discharge to total stream discharge. This requirement generally is met at only a few sites. In addition, seepage runs provide only the average rate of flow to or from a stream along the reach, whereas flow rates may vary from point to point along the reach.

Estimates of the effective hydraulic conductivities of aquifer and streambed materials can be obtained from sediment grain-size distributions, from aquifer tests, and from numerical and analog models. However, values of hydraulic conductivity determined by these techniques are inexact and are subject to interpretative bias. For example, hydraulic conductivities determined from sediment grain-size-distribution data may be inaccurate because the data do not adequately describe the sorting and packing of the sediment. Aquifer tests that are properly designed for accurate determination of hydraulic conductivity are usually complex, expensive, and labor intensive. Once obtained, the data can be difficult to interpret. Clearly, there is a need to develop alternative methods by which rates of flow and the effective vertical hydraulic conductivity between aquifers and streams can be determined simply and accurately.

An alternative approach to the problem of measuring ground-water velocity was proposed by Stallman (1963). He suggested that ground-water temperatures might be used to estimate ground-water movement indirectly, and that ground-water temperatures in combination with hydraulic-head data might be used to estimate sediment hydraulic conductivity. This report describes the application of Stallman's approach of using ground-water temperatures in the determination of vertical flow rates in sediments beneath streams. The measured ground-water temperatures are subsequently used, in combination with hydraulic-head data, to determine the effective vertical hydraulic conductivity of the sediments.

Purpose and Scope

This report presents the results of a study, the principal objective of which was to evaluate the use of vertical temperature profiles in sediments beneath streams to determine rates of vertical ground-water flow and the effective vertical hydraulic conductivity of the sediments. The study included development of a numerical model based on a finite-difference approximation of Stallman's (1963) equation describing the simultaneous flow of heat and fluid in porous media. The model was applied to ground-water temperatures measured beneath streams at two sites in Massachusetts and one site in New Jersey. Evaluation of the method included sensitivity analyses of terms in the flow equation.

Previous Investigations

Movement of ground water affects the distribution of ground-water temperatures. Numerous studies have applied this principle to make indirect estimates of ground-water velocity, areas of ground-water recharge and discharge, and aquifer hydraulic properties.

Suzuki (1960) proposed an analytical solution relating rates of infiltration into the earth from flooded rice fields to subsurface temperature profiles. Suzuki assumed constant velocity of water entering through the land surface and a surface temperature that varied sinusoidally.

Stallman (1965) modified Suzuki's solution and discussed the usefulness of the modified solution in field studies of ground-water flow. Stallman found that in natural media of average heat properties, percolation rates on the order of 2 cm/d (centimeters per day) or greater could be detected with ease by using temperature profiles resulting from diurnal temperature fluctuation. Percolation rates on the order of 0.1 cm/d could be detected by using temperature profiles resulting from annual temperature fluctuation.

Bredehoeft and Papadopulos (1965) presented an analytical solution for Stallman's (1963) equation by assuming vertical, steady flow of ground water and heat through an isotropic, homogeneous, fully saturated medium. They demonstrated how vertical ground-water velocity and hydraulic conductivity in a semiconfining layer could be determined by the use of temperature profiles.

Cartwright (1970) used Bredehoeft and Papadopulos' (1965) analytical solution to estimate ground-water discharge in the Illinois basin as suggested by temperature anomalies. He reasoned that temperature anomalies should be present in areas where ground water is moving vertically. To test this hypothesis, an isothermal map for a depth of 500 ft (feet) was constructed on the basis of bottom-hole temperatures in deep wells. This map was compared with a theoretical isothermal map for a depth of 500 ft that was constructed by projecting temperature gradients downward from the surface where mean annual air temperatures were known. A residual temperature map, made by subtracting the calculated temperature at a point from the observed temperature, indicated several warm and cool anomalies that were postulated to be discharge and recharge areas, respectively.

Sorey (1971) successfully applied Bredehoeft and Papadopulos' (1965) solution to determine upward movement of ground water through semiconfining beds in the San Luis valley of Colorado and the Roswell basin in New Mexico. Rates of upward movement through the semiconfining beds determined from temperature profiles were in good agreement with rates determined from aquifer-test data and water-budget calculations.

Cartwright (1974) suggested that, because circulating water is known to affect the temperature of the rock through which it flows, shallow ground-water flow should affect surface soil temperature. He reasoned that soil temperatures might be used to delineate small, shallow ground-water flow systems and to distinguish recharge and discharge areas. Temperature data collected in the field during summer and winter months supported this hypothesis, and ground-water flow patterns inferred from the temperature data agreed with flow patterns inferred from hydrologic data.

Wankiewicz (1984) used ground-temperature observations to investigate heat transfer by conduction and convection beneath two streams in the Northwest Territories. Wankiewicz found that conduction was the dominant heat-transfer process at one site, whereas convection was the dominant process at the other site. Convection at the second site increased the apparent thermal conductivity of sediments by an order of magnitude over the value attributed to pure conduction.

Methods of Investigation

This study was conducted in two phases. During the first phase, the principles governing the simultaneous flow of heat and fluid in the Earth were used to investigate the possible use of temperature profiles to determine rates of vertical ground-water flow in the sediments beneath streams. The partial differential equation that describes the simultaneous flow of fluid and heat in the Earth was solved numerically for transient, one-dimensional, vertical flow of fluid and heat in saturated sediments beneath streams. This numerical model was used to test the sensitivity of theoretical temperature profiles to variations in the thermal properties of saturated sediments and in the rate of vertical ground-water flow.

Three field test sites were selected and instrumented for measurement of their thermal and hydraulic regimes during the second phase of the study. One test site was located along a stream reach where ground-water flow is nearly zero. The second site was located along a stream reach where there is a high rate of ground-water discharge. The third site was located along a stream reach where infiltration from the stream into the underlying sediments was occurring because of nearby pumping. At each site, instrumentation was installed beneath the stream for measurement of temperatures and ground-water levels. Concurrent measurements of ground-water temperatures, ground-water levels, stream stage, and stream temperature were made at specified time intervals at each site during the study. Temperature and hydraulic data collected at these sites were used in the numerical model to estimate the rate of vertical ground-water flow and the effective vertical hydraulic conductivity of the sediments at each site. The feasibility of using ground-water temperatures to estimate ground-water velocity and hydraulic conductivity at the field sites subsequently was evaluated by comparing results obtained from the models with independent estimates.

Acknowledgments

The author would like to thank officials and employees of the towns of Hardwick, New Braintree, and Ware, Mass., and Dover, N.J., for their cooperation and help in setting up the field test sites. The author is particularly grateful to Skip Ward, superintendent of the Hardwick Department of Public Works, John P. Harszy, superintendent of the Ware Water Department, Vance Katucha, superintendent of the Ware Sewage Treatment Plant, and Andrew DuJack, superintendent of the Dover Water Department. The author also would like to thank Steven Prouty, resident of Hardwick, for permission to install a test site on his property.

ANALYTICAL AND NUMERICAL EQUATIONS DESCRIBING THE VERTICAL DISTRIBUTION OF TEMPERATURE IN SEDIMENTS BENEATH STREAMS

The general equation describing the simultaneous flow of fluid and heat in the Earth (Stallman, 1963) is

$$k\left[\frac{\partial^2 T}{\partial x^2}+\frac{\partial^2 T}{\partial y^2}+\frac{\partial^2 T}{\partial z^2}\right]-c_w\rho_w\left(v_x\frac{\partial T}{\partial x}+v_y\frac{\partial T}{\partial y}+v_z\frac{\partial T}{\partial z}\right)=c\rho\frac{\partial T}{\partial t} \quad (1)$$

where

k=thermal conductivity of the rock-fluid matrix;
T=temperature at any point x,y,z at any time t in the Earth;
c_w=volumetric heat capacity of the fluid;
ρ_w=density of the fluid;
v_x,v_y,v_z=components of darcian fluid velocity in the Earth in the x, y, and z directions, respectively;
c=volumetric heat capacity of the rock-fluid matrix; and
ρ=wet-bulk density (density of the rock-fluid matrix).

Assumptions made when deriving equation 1 are that the Earth is homogeneous and isotropic to flow of both heat and water, that fluid flow occurs under steady-state conditions, that heat flow is nonsteady, and that there is no internal heat generation or loss.

For one-dimensional, vertical, anisothermal flow of an incompressible fluid through homogeneous, porous media, equation 1 becomes (Suzuki, 1960; Stallman, 1965)

$$k\,\frac{\partial^2 T}{\partial z^2}-v_z c_w\rho_w\frac{\partial T}{\partial z}=c\rho\,\frac{\partial T}{\partial t} \quad (2)$$

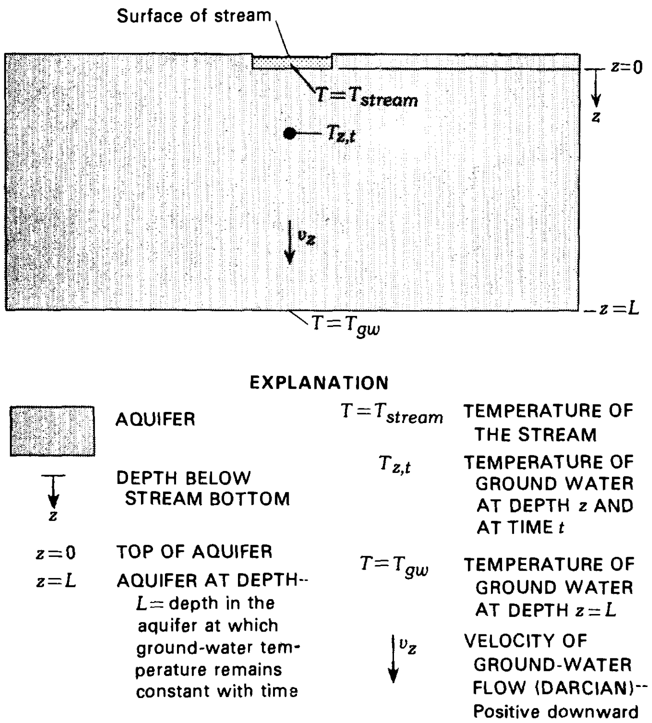


Figure 1. Conceptual model of stream-aquifer system used in describing simultaneous flow of heat and ground water in sediments beneath a stream.

Equation 2 can be used to describe ground-water temperature ($T_{z,t}$) at any depth z beneath a stream at any time t. An idealized cross section of saturated sediments underlying a stream, illustrating the conceptual model, is shown in figure 1. In the conceptual model, ground-water flow is vertically downward at velocity v_z if infiltration from the stream is occurring, and is vertically upward at velocity $-v_z$ if ground water is discharging to the stream. The upper thermal-boundary condition (T_{stream}) describes the temporal variation in temperature of the overlying stream, and the lower thermal-boundary condition (T_{gw}) describes the temporal variation in temperature of the ground water at depth $z=L$ below the stream bottom.

Diurnal and annual variation of the temperature of many streams can be described by a harmonic function (Ward, 1963; Collings, 1969; Gilroy and Steele, 1972; Collings, 1973; Tasker and Burns, 1974; Shampine, 1977). By assuming that stream temperature varies harmonically over time and that ground-water temperature remains constant over time and is equal to mean stream temperature at $z=L$, the following thermal-boundary conditions apply:

$$T=T_{stream}=T_{av}+T_{samp}(\sin((2\pi/\tau)d+ph)) \quad \text{at } z=0 \qquad (3)$$

and

$$T=T_{gw}=T_{av} \quad \text{at } z=L \qquad (4)$$

where

T_{stream}=temperature of the stream at any time d;
T_{av}=mean stream temperature during the harmonic period;
T_{samp}=semiamplitude of the stream temperature variation during the harmonic period;
π=3.14159;
τ=period of the harmonic function;
d=time;
ph=phase angle of the harmonic function;
T_{gw}=ambient ground-water temperature; and
L=depth in the aquifer at which ground-water temperature remains constant with time.

Suzuki (1960) and Stallman (1965) presented analytical solutions to equation 2 for the upper thermal-boundary condition given in equation 3 and the lower thermal-boundary condition given in equation 4 at z equals infinity.

Alternatively, an explicit finite-difference approximation to equation 2 is

$$T_i^{n+1}=\frac{k\Delta t}{\rho c\Delta z^2}\left(1+\frac{\rho_w c_w v_z \Delta z}{2k}\right)T_{i-1}^n+\frac{k\Delta t}{\rho c\Delta z^2}\left(1-\frac{\rho_w c_w v_z \Delta z}{2k}\right)T_{i+1}^n+\left(1-\frac{2k\Delta t}{\rho c\Delta z^2}\right)T_i^n \qquad (5)$$

where

T_i^{n+1}=temperature at node i at time step $n+1$;
T_{i-1}^n=temperature at node $i-1$ at time step n;
T_{i+1}^n=temperature at node $i+1$ at time step n;
Δt=time increment between time steps;
Δz=spacing between nodes; and
k, ρ, c, ρ_w, c_w, and v_z are as defined previously.

The finite-difference approximation permits solution for temperatures at z,t if there is nonsinusoidal variation in the temperature of the stream (the upper boundary condition) and (or) if ambient ground-water temperature at depth does not equal mean stream temperature.

Both the numerical stability and the accuracy of the explicit finite-difference approximation may vary depending on the values of α ($\alpha=k/c$), Δt, and Δz. The value of $\alpha\Delta t/(\Delta z)^2$ must be less than 0.5 to achieve numerical stability. In all simulations, Δt and Δz were set at values such that $\alpha\Delta t/(\Delta z)^2$ was less than 0.5. The numerical solution also was verified to be accurate by comparing temperature-depth calculations from the numerical model with those obtained using Stallman's (1965) solution, for several of the simulations in which stream-temperature variation was exclusively sinusoidal.

Application of Numerical Equation to Determine Vertical Flow Rates and Hydraulic Conductivity

The rate of vertical ground-water flow in sediments beneath a stream can be determined indirectly using equation 5 (or the analytical solution of Stallman, 1965, if the temporal variation in stream temperature is exclusively sinusoidal and ambient ground-water temperature at depth equals mean stream temperature). For this determination, the physical and thermal properties of the saturated sediments beneath the stream, the temperature variation over time at the upper thermal boundary, and the temperature at the lower thermal boundary are entered into equation 5. Then the velocity (v_z) in equation 5 is varied until the model-simulated temperature distribution with depth and over time matches the temperature distribution with depth and over time measured in the field. Finally, the effective vertical hydraulic conductivity of the sediments is calculated from Darcy's law using the model-simulated velocity and field-measured vertical hydraulic gradient across the sediments.

Physical and Thermal Properties of Saturated Fine-Grained and Coarse-Grained Sediments

The physical and thermal properties of the sediments at a field site must be determined for input into equation 5. These properties can be determined in several ways. One method is to collect cores of the sediments during test drilling and to measure all the necessary physical and thermal properties of the cores in the laboratory. A less costly method is to collect cores and to determine only a couple of the properties, such as the sediment wet- and dry-bulk densities, in the laboratory. Mathematical relations between the laboratory-determined properties and the remaining properties are then used to indirectly determine values of these remaining properties. This section of the report describes the physical and thermal properties contained in equation 5. Simple graphical relations are presented that can be used to estimate these properties at field sites for use in equation 5.

Sediments generally can be classified as either coarse grained or fine grained when describing both their physical properties (Buol and others, 1973) and their thermal properties (Kersten, 1949). In addition to differing in average grain size, coarse-grained and fine-grained sediments may differ significantly in mineralogy. Coarse-grained sediments consist mostly of sand- and gravel-sized siliceous and feldspathic minerals. Fine-grained sediments consist mostly of silt- and clay-sized particles that may be predominantly clay minerals. Differences between the thermal properties of coarse-grained and fine-grained sediments are partly attributable to differences in the mineralogy of these two sediment types, because of differences in the thermal properties of the minerals making up the sediments. Differences between the thermal properties of coarse-grained and fine-grained sediments are also attributable to differences in their structural framework (Farouki, 1981). Coarse-grained sediments have a structural framework that results from grain-to-grain contact. Fine-grained sediments have a structural framework in which individual particles are separated by films of water.

Wet-Bulk Density

Sediment wet-bulk density (total density of the solid-fluid matrix) is equal to the total mass of the fluid-sediment mixture per unit volume. The volume of the moist sediment is the volume of solids, fluid, and air.

In saturated sediments, the volume of air is negligible and the wet-bulk density equals the mass of the sediment plus water per unit volume of saturated sediment. Wet-bulk densities of saturated fine-grained sediments range from about 1.4 to 2.0 g/cm^3 (grams per cubic centimeter); wet-bulk densities of coarse-grained sediments range from about 1.7 to 2.3 g/cm^3 (Heiland, 1946; Kersten, 1949; Clark, 1966; Lunardini, 1981).

Dry-Bulk Density

Sediment dry-bulk density (dry unit weight) is equal to the mass of solids per unit volume of moist sediment. The volume of sediment is equal to the volume of solids, fluid, and air. Dry-bulk densities of fine-grained sediments generally range from 1.0 to 1.6 g/cm^3; dry-bulk densities of coarse-grained sediments generally range from 1.2 to 1.8 g/cm^3 (Heiland, 1946; Kersten, 1949; Houk, 1951; Clark, 1966; Hillel, 1971; Brady, 1974; Foth, 1978).

Thermal Conductivity

Sediment thermal conductivity, k (cal/s-cm-°C (calories per second centimeter degree Celsius)), is the quantity of heat transmitted in unit time through a unit cross-sectional area under a unit temperature gradient. Some factors that influence the thermal conductivities of sediments are sediment structure and packing (grain size and shape, the surface characteristics of the sediment particles, number and nature of grain-to-grain contacts, pore size, porosity, and density), sediment mineralogy, degree of saturation, the amount, phases, and salt content of water in the sediment, the temperature of the sediment, and the thermal conductivities of the sediment solids.

Although thermal conductivity is influenced by many factors, a general relation exists between the thermal conductivity of saturated fine-grained and coarse-grained sediments and sediment dry-bulk density, as shown in figure 2*A*. The graph of figure 2*A* summarizes experimentally derived thermal-conductivity data available in the literature (Lunardini, 1981). The soils used to determine conductivity values were reconstituted and do not represent the in situ soil matrix. Nevertheless, these data are the best available for the large range of dry-bulk densities shown in figure 2*A* (Lunardini, 1981, p. 162). The relation between thermal conductivity and dry-bulk density shown in figure 2*A* may cover a larger range of dry-bulk densities than commonly occur for in situ soils.

Figure 2*A* indicates that the thermal conductivities of both the fine-grained and coarse-grained sediments increase as the dry-bulk density of the sediment increases. The

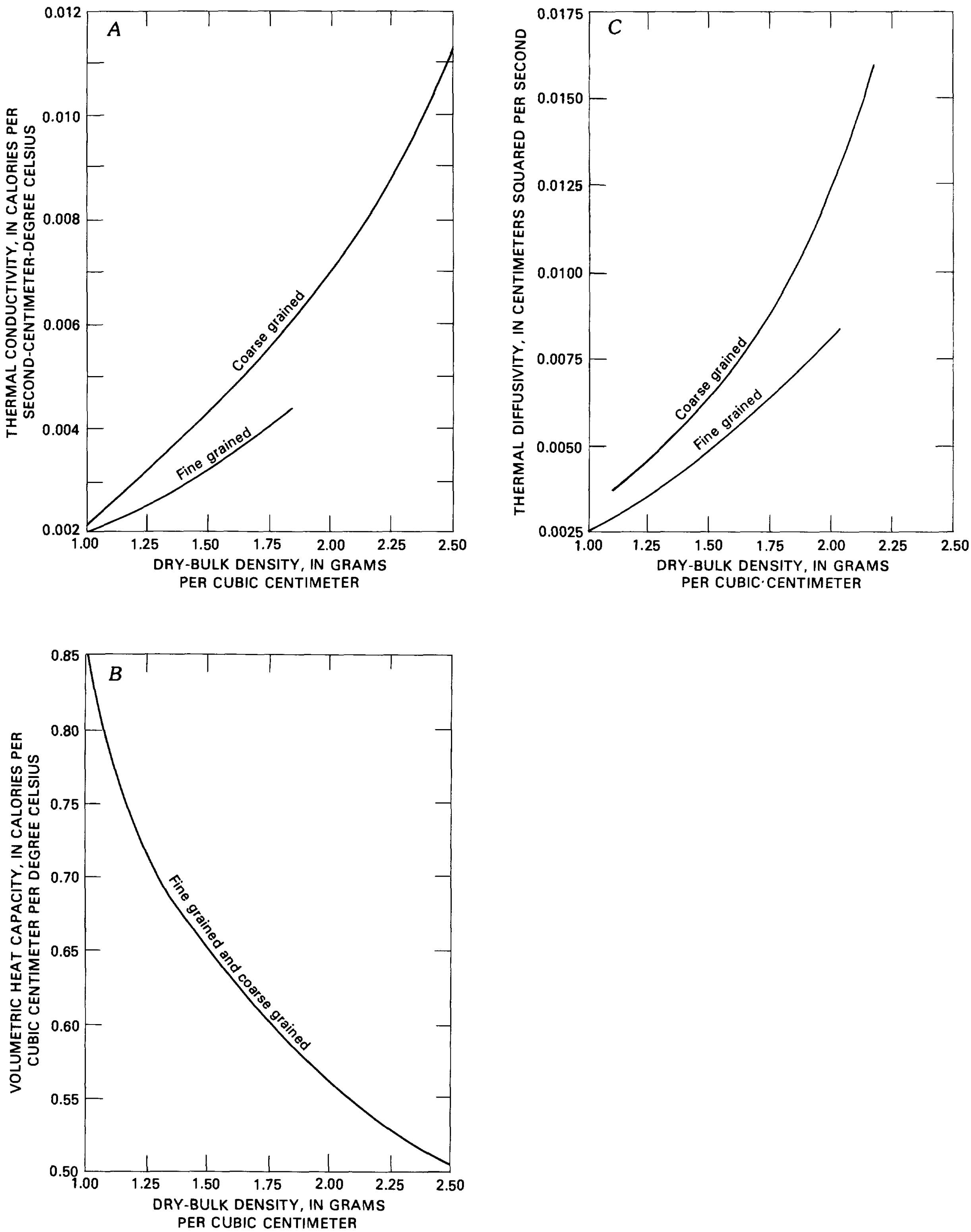


Figure 2. Relation of thermal properties of saturated fine-grained and coarse-grained sediments at 4.4 °C to sediment dry-bulk density: *A*, thermal conductivity (adapted from Lunardini, 1981); *B*, volumetric heat capacity (adapted from Lunardini, 1981); *C*, thermal diffusivity.

increase in thermal conductivity is caused by an increase in the ratio of mass of solid material to mass of pore water per unit volume of sediment and by an increase of grain-to-grain contact (Farouki, 1981). As stated previously, dry-bulk densities of naturally occurring sediments range from about 1.0 to 1.6 g/cm^3 for fine-grained sediments and from about 1.2 to 1.8 g/cm^3 for coarse-grained sediments. Therefore, on the basis of the relation between dry-bulk density and thermal conductivity (fig. 2*A*), the thermal conductivity of saturated fine-grained sediments ranges from about 0.0020 to about 0.0034 cal/s-cm-°C, and the thermal conductivity of saturated coarse-grained sediments ranges from about 0.0029 to about 0.0059 cal/s-cm-°C.

Differences in mineralogy also contribute to the differences between the two curves of figure 2*A*. Quartz, which usually is abundant in coarse-grained sediments, has a thermal conductivity about four times that of the micaceous minerals usually abundant in fine-grained sediments (Farouki, 1981). Therefore, for a given dry-bulk density, coarse-grained sediments are likely to have higher thermal conductivities than fine-grained sediments.

The thermal conductivity of saturated sediments also varies with temperature. Kersten (1949) found that thermal conductivity increased about 4 percent over a temperature range of 4 to 21 °C. Other data (Farouki, 1981) also indicate that conductivity increases with increasing temperature. However, the increase in thermal conductivity with increase in temperature is small compared with the increase in conductivity with increase in dry-bulk density.

Heat Capacity

The heat capacity per unit mass, C (cal/g-°C (calories per gram degree Celsius)), of sediment is the quantity of heat required to raise a unit mass of sediment 1 °C. The heat capacity per unit volume (volumetric heat capacity), c (cal/cm^3-°C), of a sediment is the quantity of heat required to raise a unit volume of sediment 1 °C. The volumetric heat capacity of a sediment is the product of the heat capacity per unit mass and the sediment density. The volumetric heat capacity of a sediment is related to the heat capacities and volume fractions of the individual sediment components. If the volumetric heat capacities of the solids, water, and air making up the sediment are c_s, c_w, and c_a, respectively, and the volume fractions of solids, water, and air making up the sediment are x_s, x_w, and x_a, respectively, then the volumetric heat capacity of the sediment is (Farouki, 1981)

$$c=x_s c_s+x_w c_w+x_a c_a \tag{6}$$

Heat capacity varies little among the mineral solids that make up fine-grained and coarse-grained sediments (Kersten, 1949; Lunardini, 1981). However, the heat capacities of both fine-grained and coarse-grained sediments vary with sediment dry-bulk density, as shown in figure 2*B*. Figure 2*B* is redrawn from Lunardini (1981, figs. 4.54 and 4.58). The relation between dry-bulk density and heat capacity in figure 2*B* may cover a larger range of dry-bulk densities than commonly occurs for in situ soils.

Figure 2*B* indicates that the volumetric heat capacities of both the fine-grained and coarse-grained sediment decrease as the dry-bulk density of the sediment increases. Using the relation between volumetric heat capacity and dry-bulk density (fig. 2*B*), the volumetric heat capacity of saturated fine-grained sediments ranges from about 0.85 cal/cm^3-°C for a dry-bulk density of 1.0 g/cm^3 to 0.64 cal/cm^3-°C for a dry-bulk density of 1.6 g/cm^3. The volumetric heat capacity of saturated coarse-grained sediments ranges from 0.74 cal/cm^3-°C for a dry-bulk density of 1.2 g/cm^3 to 0.60 cal/cm^3-°C for a dry-bulk density of 1.8 g/cm^3.

Thermal Diffusivity

The thermal behavior of a sediment is influenced by both the sediment's thermal conductivity and its heat capacity when unsteady thermal conditions exist. The governing parameter when unsteady thermal conditions exist is thermal diffusivity, α (cm^2/s (centimeter squared per second)). Thermal diffusivity is equal to the ratio of thermal conductivity to volumetric heat capacity. Temperatures in a sediment of high thermal diffusivity change more rapidly in response to a sudden external temperature change than do temperatures in a sediment of low thermal diffusivity (Farouki, 1981).

The thermal diffusivity of saturated fine-grained and coarse-grained sediments varies with sediment dry-bulk density, as shown in figure 2*C*. The relations between thermal diffusivity and dry-bulk density in figure 2*C* were determined by calculating the ratio of thermal conductivity (fig. 2*A*) and volumetric heat capacity (fig. 2*B*) at various dry-bulk densities. Figure 2*C* indicates that thermal diffusivity of both saturated fine-grained and coarse-grained sediments increases as dry-bulk density of the sediment increases. Thermal diffusivity of saturated fine-grained sediments ranges from about 0.0024 cm^2/s for a dry-bulk density of 1.0 g/cm^3 to 0.0053 cm^2/s for a dry-bulk density of 1.6 g/cm^3. Thermal diffusivity of saturated coarse-grained sediments ranges from about 0.0039 cm^2/s for a dry-bulk density of 1.2 g/cm^3 to 0.0098 cm^2/s for a dry-bulk density of 1.8 g/cm^3.

Characteristics of Theoretical Temperature Profiles and Profile Envelopes in Sediments Beneath Streams

An example of theoretical monthly temperature profiles (the relationship of temperature to depth) in saturated coarse-grained sediments beneath a stream is shown in figure 3. The monthly temperature profiles were calculated using equation 5. Physical and thermal properties used for this simulation were a dry-bulk density of 1.2 g/cm^3, a wet-bulk density of 1.7 g/cm^3, and a thermal diffusivity of

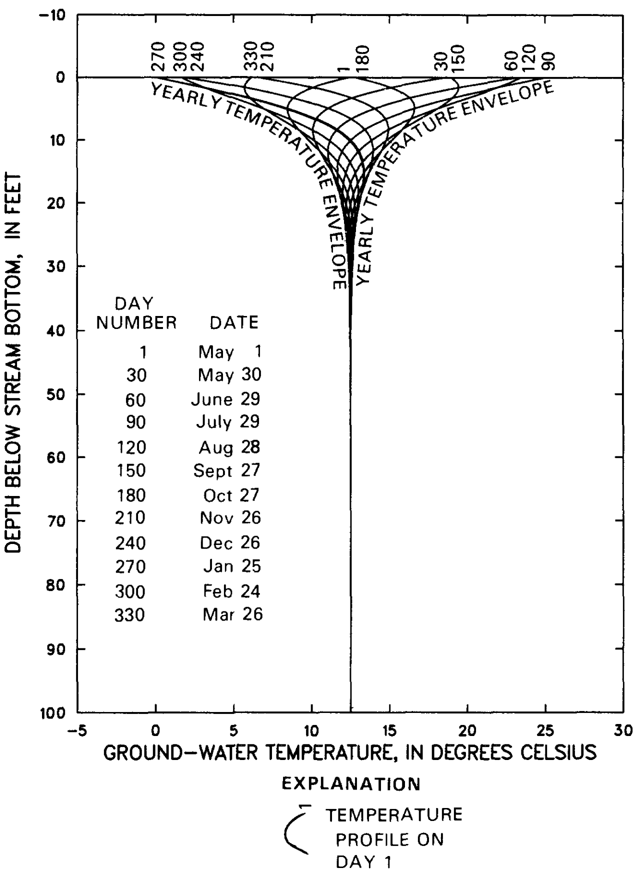


Figure 3. Simulated monthly temperature profiles, and the yearly temperature envelope described by the temperature profiles, in saturated coarse-grained sediments, with thermal diffusivity of 0.0039 centimeter squared per second and ground-water velocity of 0.0 foot per day.

0.0039 cm^2/s (from fig. 2*C* for a dry-bulk density of 1.2 g/cm^3). Stream temperature was varied between 0 and 25 °C during the year of simulation because a yearly range in temperature of 0 to 25 °C is common for New England streams. For purposes of demonstrating the characteristics of theoretical temperature profiles and profile envelopes in sediments beneath streams, stream temperature was varied sinusoidally throughout the year, although, as discussed later, most streams in New England vary sinusoidally only from early spring through late fall. During the remaining several months, stream temperature remains at 0 °C because the streams are ice-covered. The effects of stream temperatures remaining at 0 °C during winter on temperatures in sediments beneath streams are discussed in the section "Sensitivity Analysis Using the Numerical Model." Stream temperature on about May 1 is about equal to the average annual stream temperature in central New England. Therefore, day 1 (fig. 3) for this simulation is equivalent to May 1. For the simulation shown in figure 3, ambient ground-water temperature was held constant at the mean annual stream temperature of 12.5 °C at a depth of 100 ft below stream bottom. In this example, ground-water velocity (v_z) was set at 0 ft/d (foot per day). If $v_z=0$, heat is transferred only by conduction.

The monthly temperature profiles (fig. 3) indicate that temperature at depth beneath the stream lags behind stream temperature. This lag occurs because of the low thermal diffusivity of the saturated sediments. The temperature lag is particularly evident during late spring (1 and 30 d (days)) and late fall (180 and 210 d), when stream temperature changes rapidly.

Figure 3 indicates that the magnitude of annual fluctuation in temperature decreases as the depth beneath the stream increases. In figure 3, temperature fluctuation appears to be constant below a depth of about 35 ft. Actually, annual fluctuation in temperature occurs at all depths beneath the stream. Beneath a depth of about 35 ft, however, the temperature fluctuation is so small that it is not apparent in the figure, and for the purposes of this study, ground-water temperatures at all depths below 35 ft below stream bottom in figure 3 are assumed to be constant at 12.5 °C.

The yearly temperature envelope is formed by two curves. One curve is constructed by connecting the minimum temperature that occurs during the year at each depth beneath the stream. The other curve is constructed by connecting the maximum temperature that occurs during the year at each depth beneath the stream. Superposition of monthly temperature profiles defines indirectly the yearly temperature envelope. The shape of this envelope is indicative of the flow rate between the sediments and the overlying stream, as discussed later in the section on sensitivity.

Figure 3 also indicates that if stream temperature varies harmonically and ambient ground-water temperature is equal to mean stream temperature, the instantaneous temperature profiles that occur one-half year apart are mirror images of each other and the temperature envelope is symmetrical around ambient ground-water temperature.

Sensitivity Analysis Using the Numerical Model

In this section of the report, the sensitivity of monthly temperature profiles and yearly temperature envelopes to variation in sediment thermal properties, ground-water velocity, and temporal variation in stream temperature is described. For this sensitivity analysis, the theoretical temperature profiles were calculated using equation 5.

Sensitivity analysis is useful as a means of evaluating the relative importance of the terms in the flow equation, thereby providing a basis for estimating the precision of measurement required to obtain given magnitudes of probable errors in the solution.

Table 1. Minimum and maximum values of physical and thermal properties of saturated fine-grained and coarse-grained sediments

[g/cm^3, grams per cubic centimeter; cal/s-cm-°C, calorie per second centimeter degree Celsius; cal/cm^3-°C, calorie per cubic centimeter degree Celsius; cm^2/s, centimeter squared per second]

Property	Fine-grained sediments with minimum and maximum dry-bulk density		Coarse-grained sediments with minimum and maximum dry-bulk density	
	Min	Max	Min	Max
Dry-bulk density, in g/cm^3	1.0	1.6	1.2	1.8
Wet-bulk density, in g/cm^3	1.4	2.0	1.7	2.3
Thermal conductivity, in cal/s-cm-°C	.0020	.0034	.0029	.0059
Heat capacity, in cal/cm^3-°C	.85	.64	.74	.60
Thermal diffusivity, in cm^2/s	.0024	.0053	.0039	.0098

Sensitivity of Temperature Profiles and Envelopes to Thermal Diffusivity

Four simulations of monthly temperature profiles were run with the numerical model using ranges of values of physical and thermal properties that commonly occur in fine-grained and coarse-grained saturated sediments. Simulations were run for dry-bulk densities of 1.0 and 1.6 g/cm^3 for the fine-grained sediments and 1.2 and 1.8 g/cm^3 for the coarse-grained sediments (table 1). These values cover the range of values of dry-bulk density commonly found for fine-grained and coarse-grained sediments. Corresponding values of the thermal properties in table 1 for the four values of dry-bulk density were determined from the relation between dry-bulk density and thermal properties (figs. 2*A*–2*C*). Wet-bulk densities listed in table 1 represent the range of wet-bulk densities commonly found for fine-grained and coarse-grained sediments, as discussed previously.

For the four simulations, stream temperature was varied sinusoidally over a year from 0 to 25 °C, ground-water temperature was held constant at a mean stream temperature of 12.5 °C at a depth of 100 ft below stream bottom, and ground-water velocity was set at zero. The temperature profiles resulting from the two simulations for fine-grained sediments are shown in figure 4, and the temperature profiles resulting from the two simulations for coarse-grained sediments are shown in figure 5.

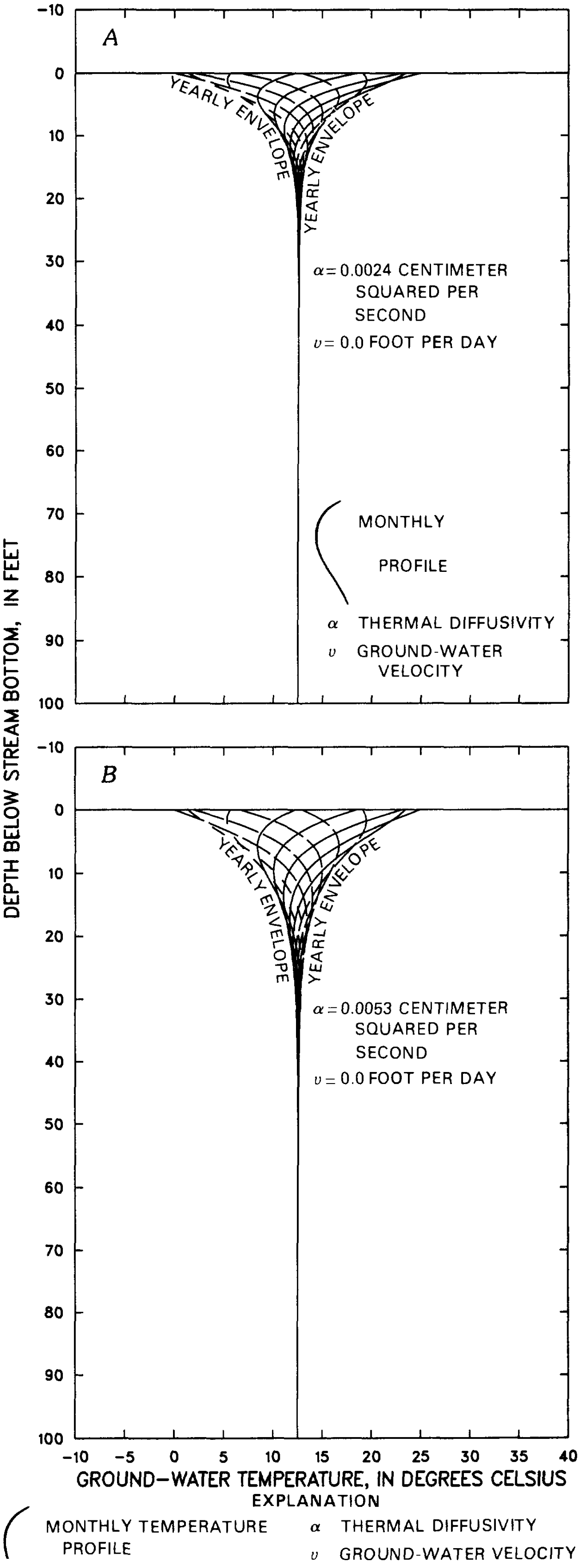
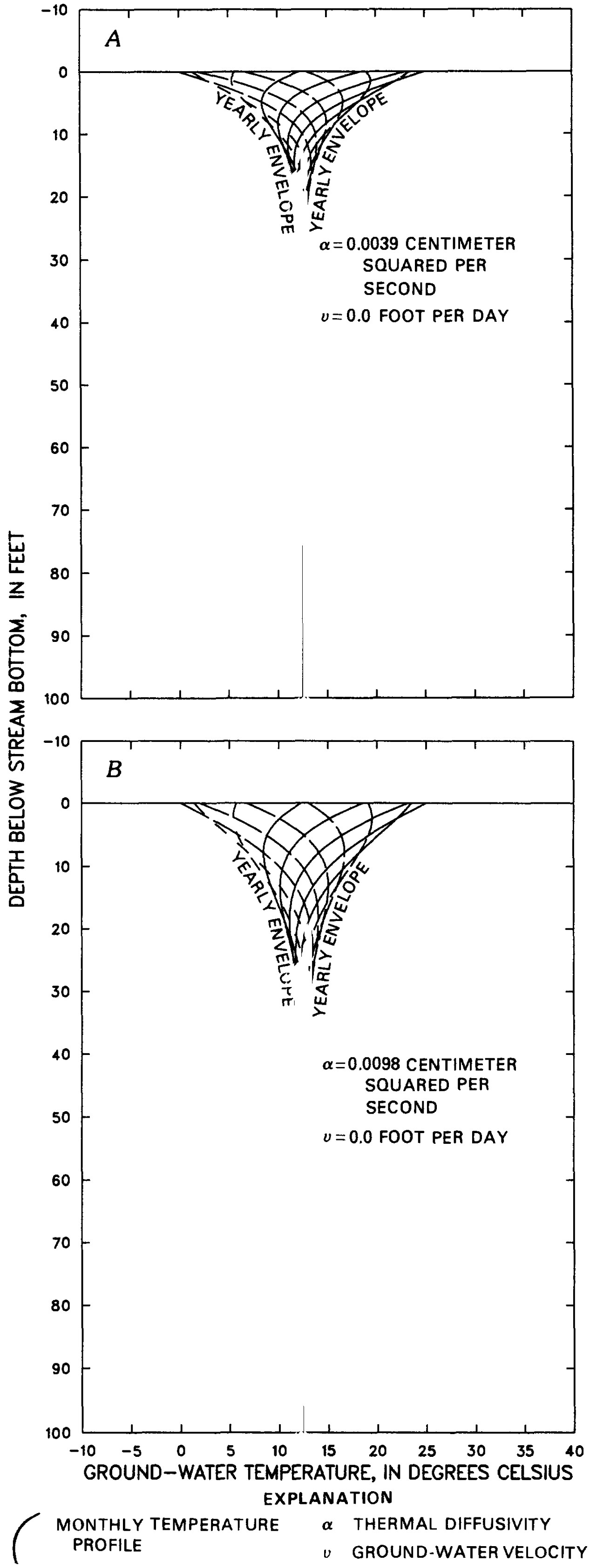


Figure 4. Relation of monthly temperature profiles, and the yearly temperature envelope described by the temperature profiles, to thermal diffusivity for fine-grained sediments in which ground-water velocity is 0.0 foot per day. ▶



◀ **Figure 5.** Relation of monthly temperature profiles, and the yearly temperature envelope described by the temperature profiles, to thermal diffusivity for coarse-grained sediments in which ground-water velocity is 0.0 foot per day.

Results of the simulations were used to evaluate the extent to which monthly temperature profiles and yearly temperature envelopes are affected by variation in the physical and thermal properties of sediments. Comparison of the monthly temperature profiles and yearly temperature envelopes resulting from simulating minimum and maximum values of thermal diffusivity of fine-grained sediments (figs. 4*A* and 4*B*) indicate that the temperature profiles and, consequently, the temperature envelopes are somewhat sensitive to variation in thermal diffusivity. Ground-water temperatures vary significantly to a depth beneath the stream of about 25 ft if thermal diffusivity is 0.0024 cm^2/s (fig. 4*A*), whereas temperatures vary significantly to a depth of about 35 ft if thermal diffusivity is 0.0053 cm^2/s (fig. 4*B*). Although variation in stream temperature is identical for all simulations, the monthly temperature profiles and yearly envelope are more compressed upward toward the stream in the simulation using the minimum value of thermal diffusivity (fig. 4*A*) than in the simulation using the maximum value of thermal diffusivity (fig. 4*B*).

Comparison of the monthly temperature profiles and yearly temperature envelopes resulting from simulating minimum and maximum values of thermal diffusivity of coarse-grained sediments (figs. 5*A*, 5*B*) indicates results similar to those for fine-grained sediments. However, temperatures vary significantly to depths of about 35 and 50 ft, respectively, for thermal diffusivities of 0.0039 cm^2/s (fig. 5*A*) and 0.0098 cm^2/s (fig. 5*B*). Both the temperature profiles and the temperature envelope are more compressed upward toward the stream in the simulation using the minimum value of thermal diffusivity than in the simulation using the maximum value of thermal diffusivity.

Sensitivity of Temperature Profiles and Envelopes to Vertical Ground-Water Velocity

The sensitivity of temperature profiles and envelopes to variation in vertical ground-water velocity was investigated only for coarse-grained sediments having a dry-bulk density of 1.2 g/cm^3. Values of wet-bulk density and the thermal properties of the sediments used in these simulations were the values corresponding to this dry-bulk density (table 1). Sensitivity testing using differing dry-bulk densities and thermal properties of both coarse-grained and fine-grained sediments was not done because it was expected that results would be similar to those obtained from the sensitivity testing for coarse-grained sediments having a dry-bulk density of 1.2 g/cm^3.

For coarse-grained sediments having a dry-bulk density of 1.2 g/cm^3, vertical upward and downward flow rates of 0.01, 0.05, 0.1, and 1 ft/d were simulated; the same yearly harmonic variation in stream temperature and the same ambient ground-water temperature were used as in the previous tests, except that for the downward flow rates (fig. 6), ambient ground-water temperatures were held constant at a depth of 250 ft beneath the stream bottom. These ranges of upward and downward flow rates probably encompass the ranges of flow rates that most commonly occur in sediments beneath streams in New England. Downward flow occurs if water is infiltrating from the stream into the underlying sediments, and upward flow occurs if ground-water is discharging from the underlying sediments into the stream.

Simulated monthly profiles and the yearly temperature envelopes for the downward velocities are shown in figure 6 and for the upward velocities in figure 7. There is almost no difference between the simulated profiles and envelopes for downward and upward velocities of 0.01 ft/d (figs. 6*D* and 7*A*, respectively). Consequently, it can be concluded that monthly profiles and yearly envelopes are insensitive to rates of vertical flow less than about 0.01 ft/d for the yearly stream-temperature fluctuation of 25 °C that typically occurs in New England. However, the profiles and envelopes in figures 6*A*, 6*B*, 6*C*, 7*B*, 7*C*, and 7*D* indicate that the temperature profiles and envelopes are sensitive to variation in ground-water velocity between 0.01 and 1 ft/d.

Convection of heat increases as vertical ground-water velocity increases. Therefore, the effect of stream-temperature fluctuations penetrates deeper beneath the stream bottom as downward ground-water velocity increases. Ground-water temperatures at depths greater than 100 ft beneath the stream bottom fluctuate over nearly the same range as stream temperature fluctuates for downward velocities greater than about 1 ft/d. Ground-water temperatures are virtually unaffected by fluctuations in stream temperature even directly beneath the stream bottom when upward ground-water velocities exceed about 1 ft/d.

The cyclic nature of past warm and cool surface-temperature fluctuations is particularly evident in each instantaneous temperature profile in figure 6*B*. These profiles show the repetitive sinusoidal fluctuation of temperature around mean surface temperature at all depths. All profiles demonstrate this repetitive sinusoidal fluctuation. However, the period of the cycle differs for different vertical flow rates.

Analysis of the sensitivity of monthly temperature profiles and yearly envelopes to variation in thermal diffusivity and ground-water velocity in saturated sediments beneath the stream (figs. 4–7) suggests that the shape of the monthly profiles and yearly envelope can be used to estimate indirectly ground-water velocity for rates of flow of less than 1 ft/d. Downward velocities exceeding 1 ft/d result in ground-water fluctuations within the top 100 ft of sediment beneath the stream that are virtually the same magnitude as the stream-temperature fluctuation. In field situations where ground-water velocity is downward at a rate of about 1 ft/d or greater, hourly temperature profiles and daily envelopes caused by diurnal harmonic variation in stream temperature may be more useful than monthly profiles and yearly envelopes in determining flow rates.

Sensitivity of Temperature Profiles and Envelopes to Nonsinusoidal Yearly Variation in Stream Temperature

The way in which stream temperature varies over time also affects the shapes of temperature profiles and envelopes. If stream temperature varies sinusoidally during the year, the temperature profiles that occur one-half year apart will be mirror images of each other and the temperature envelope will be symmetrical around mean stream temperature (figs. 3, 6, 7). Stream temperatures that vary asymmetrically throughout the year will produce asymmetric temperature profiles, and the yearly temperature envelope will be asymmetrical around mean stream temperature.

Tasker and Burns (1974) reported that stream temperatures vary harmonically over only part of the year in New England because stream temperature remains near 0 °C for up to several months during the winter. Under these climatic conditions common in New England, the following upper thermal-boundary condition replaces that given in equation 3:

$$T_{stream}=0 \qquad \text{when } 0 \leq d \leq d_0 \qquad (7)$$

$$T_{stream}=T_{av}+T_{samp}\ (\sin((2\pi/\tau)d+ph)) \text{ when } d_0 < d \leq 365 \qquad (8)$$

where

d_0=time during which temperature of stream remains at 0 °C

and all other variables are as defined previously. This temporal variation in stream temperature is typical in central and northern New England, where stream temperatures may remain at 0 °C for as much as 2 to 3 months during the winter (Tasker and Burns, 1974) and may remain near 0 °C for several additional months during mid- to late fall and early to mid-spring.

The effect of yearly nonharmonic variation in stream temperature on the shape of monthly profiles and yearly envelopes was investigated by simulating 75 continuous days during which stream temperature remained at 0 °C. During the remainder of the year stream temperature was varied harmonically between 0 and 25 °C. Physical and thermal properties used in the simulations were those for coarse-grained sediments having a dry-bulk density of 1.2 g/cm^3 (table 1). The variation in stream temperature of from

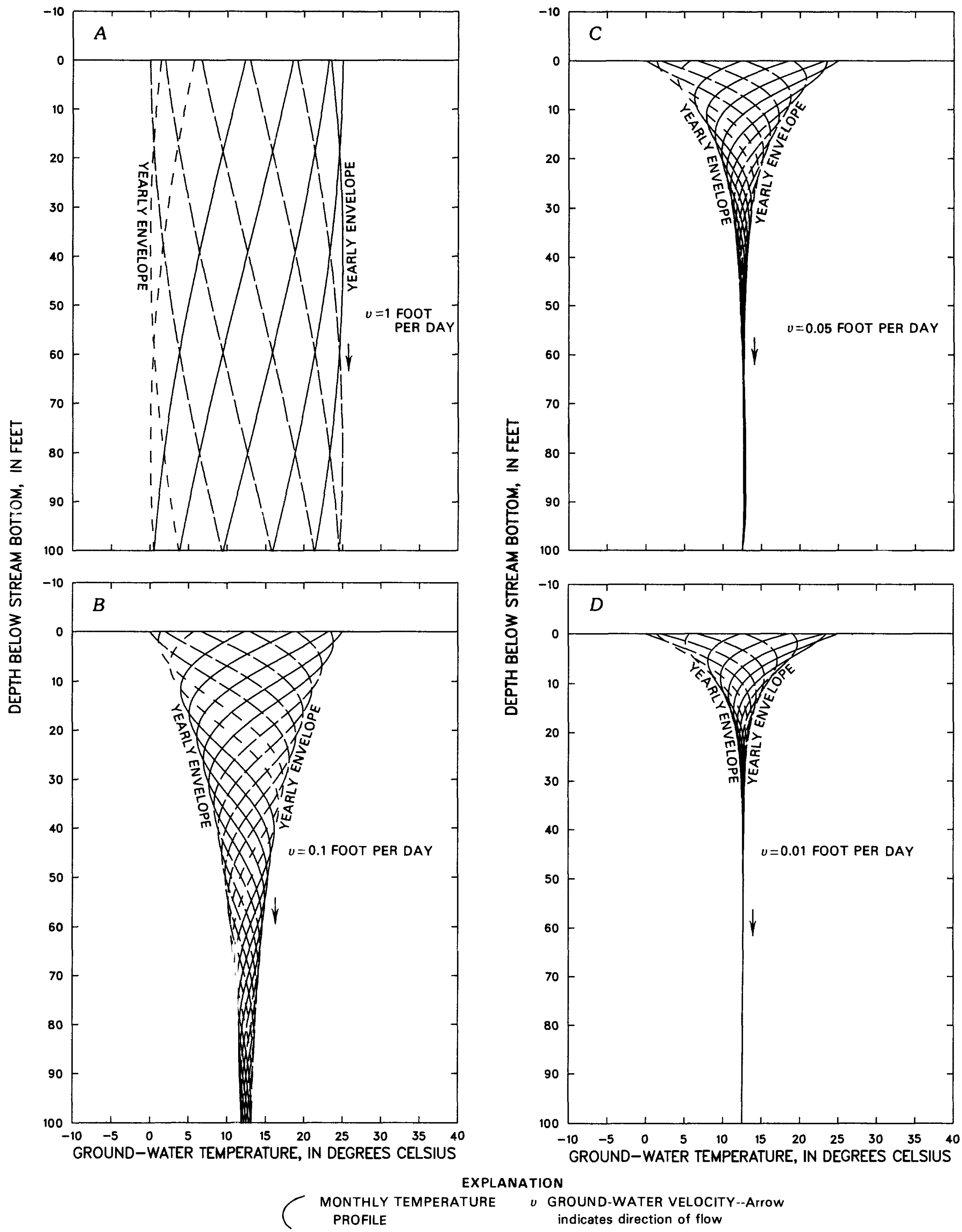


Figure 6. Simulated monthly temperature profiles, and the yearly temperature envelope described by the temperature profiles, in coarse-grained sediments for downward ground-water velocities of *A*, 1.0 foot per day, *B*, 0.1 foot per day, *C*, 0.05 foot per day, and *D*, 0.01 foot per day.

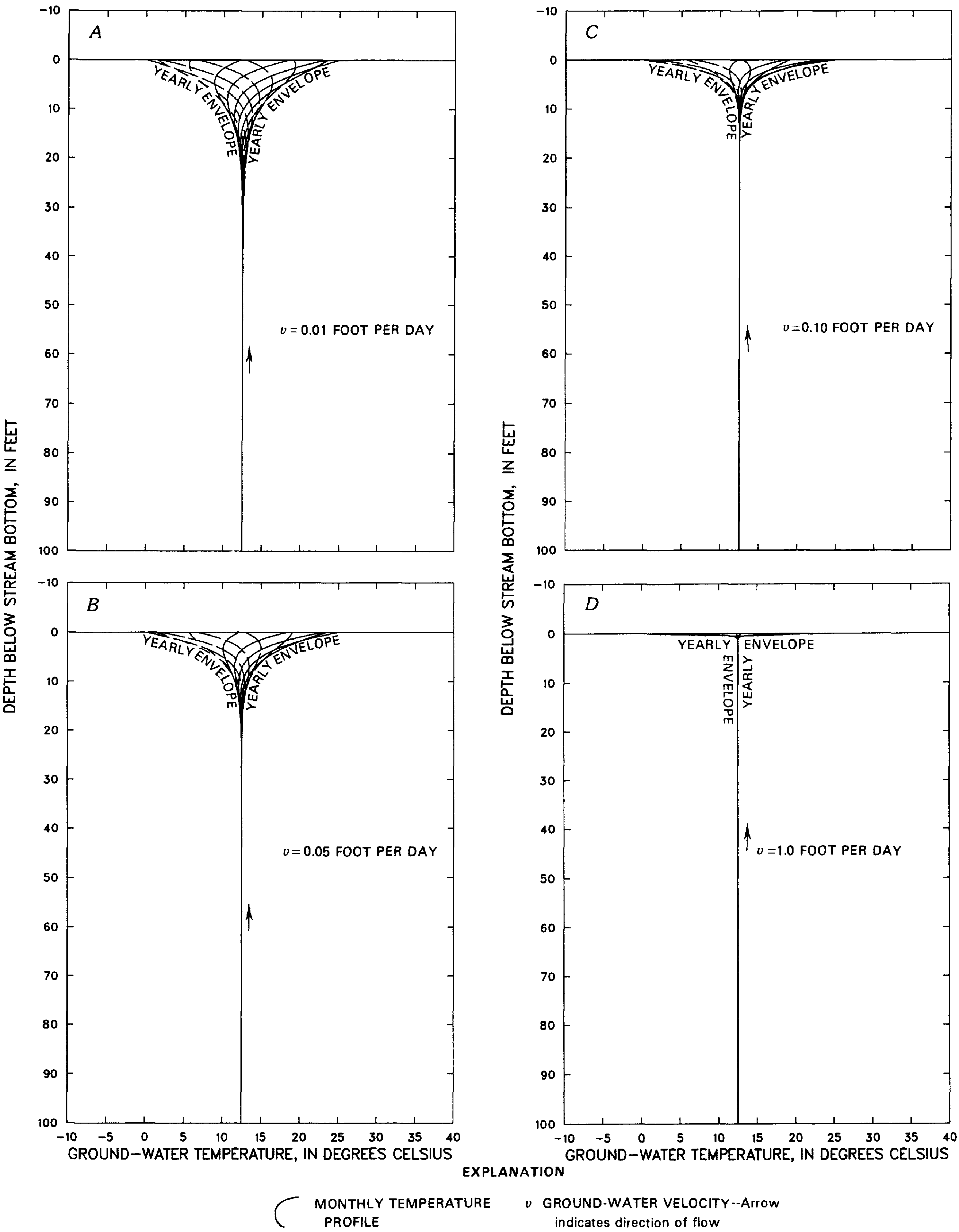


Figure 7. Simulated monthly temperature profiles, and the yearly temperature envelope described by the temperature profiles, in coarse-grained sediments for upward ground-water velocities of *A*, 0.01 foot per day, *B*, 0.05 foot per day, *C*, 0.10 foot per day, and *D*, 1.0 foot per day.

0 to 25 °C for 290 d and then constant stream temperature at 0 °C for 75 d results in an average annual stream temperature of 9.9 °C. Therefore, ambient ground-water temperature in the model was set at 9.9 °C. Ground-water velocities of 0 ft/d, 0.05 ft/d upward, and 0.05 ft/d downward were simulated.

Results of these three simulations are shown in figure 8. The asymmetry of the monthly profiles that occur one-half year apart and the asymmetry of the yearly temperature envelope around mean ground-water temperature are apparent in all three diagrams. The asymmetry of the monthly profiles and yearly envelopes shown in figure 8 demonstrates the importance of accurately simulating the nonsinusoidal nature of yearly stream-temperature variation at locations where stream temperature remains near 0 °C for more than several weeks during the winter.

FIELD TESTS OF THE TEMPERATURE-PROFILE METHOD

The use of temperature profiles and envelopes to determine vertical ground-water velocity and effective hydraulic conductivity of sediments was evaluated at three field sites (fig. 9). Two sites are located in the Ware River, in the towns of Hardwick and New Braintree, central Massachusetts. The third site is located in the Rockaway River at the Princeton Avenue well field in the town of Dover, N.J.

The Hardwick site is located in a reach of the Ware River that is underlain by silts and clays of low hydraulic conductivity. Ground-water flow at this site probably is nearly zero because of the presence of the low-conductivity clay. The New Braintree site is located in a reach of the Ware River that is underlain by a sand and gravel aquifer of high hydraulic conductivity. Ground water at this site discharges from the aquifer to the Ware River. The Dover site is located in a reach of the Rockaway River that is underlain by a sand and gravel aquifer of high hydraulic conductivity. The aquifer at this site is pumped for municipal supply. The pumping induces infiltration from the Rockaway River into the underlying aquifer.

At all three sites, hydraulic gradients and water levels were measured in piezometers installed vertically beneath the rivers. Concurrent measurements of ground-water temperatures, ground-water level, river stage, and river temperature were made monthly during the study at the Hardwick and New Braintree sites, and bihourly from June 2 to June 6, 1986, at the Dover site. Temperatures were measured in each piezometer by lowering a thermistor on a 100-ft lead through the water column. The temperature of the water column at 0.5-ft intervals in the piezometer was measured and recorded. The thermistor was held at each 0.5-ft interval for at least 30 s or until the temperature at that depth remained constant within the precision of the meter of ±0.1 °C. The temperature profile in each piezometer then was drawn using these temperature-depth data.

Hardwick, Massachusetts, Site

The Hardwick site is located in the Ware River about 0.5 mi (mile) north of the village of Wheelwright at test boring PVC145 (fig. 10). At this site, lacustrine clay, silt, and fine sand overlie glacial outwash composed of fine to coarse sand (fig. 10). These glacial sediments partly fill the bedrock valley that underlies the Ware River along this reach of the river (fig. 11). As indicated in figure 10, the fine-grained lacustrine deposits extend along the eastern half of the Ware River valley for about 1 mi north and south of the village of Wheelwright and extend south about 1 mi into the Winimusset Brook valley.

The river bottom at the site consists of a loose mixture of clay, silt, and organic material about 0.5 ft thick. Test boring PVC145 (fig. 11) indicates that the sediments in the top 10 ft beneath the bottom of the Ware River are composed of clayey, sandy silt. From 10 to 60 ft beneath the stream, the sediments are composed of uniform gray lacustrine clay. Beneath this lacustrine cap of silt and clay is a buried sand aquifer approximately 85 ft thick. The sand aquifer is composed of silty sand from a depth of 60 ft beneath the river bottom to 100 ft, where it grades into medium to coarse sand. The medium to coarse sand is 45 ft thick. The bedrock surface is at a depth of 145 ft beneath river bottom.

Piezometers with 2-ft screens at their bottoms were installed at 25 (PVC25), 60, 75, 87, 100, and 145 (PVC145) ft beneath the Ware River at the Hardwick site. Heads in these piezometers indicate that there is an upper, shallow flow system and a deep flow system, separated by a stagnation zone. Ground water above the stagnation zone flows upward and discharges to the Ware River. Ground water below the stagnation zone flows down into the buried sand aquifer and then south. The location of the stagnation zone migrates vertically during the year. During spring, when recharge is high, the zone migrates downward to depths of 60 to 75 ft beneath the river bottom. During the remainder of the year, the stagnation zone is located within the top 25 ft beneath the river bottom.

The stage of the Ware River and water levels in the piezometers at the Hardwick site were measured monthly at the same time temperatures were measured. The range of the river-stage measurements during the period October 1984 to October 1985 was only 0.18 ft. This small difference in stage is attributed to both the presence of a check dam downstream at Wheelwright (fig. 10) and the large area of wetland near the site, both of which mitigate extreme fluctuations in river stage. Water levels in the piezometers fluctuated as much as 1.2 ft.

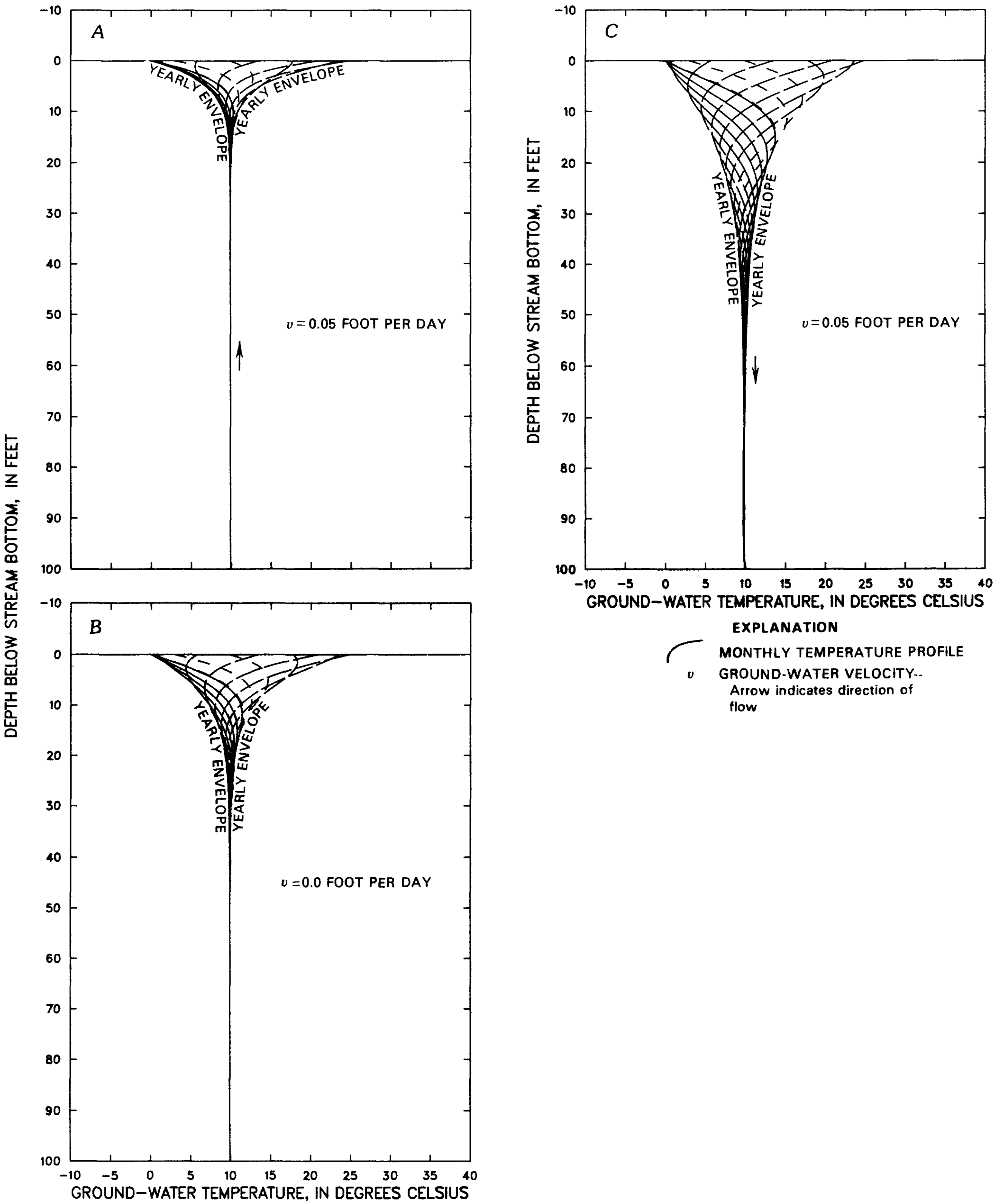


Figure 8. Simulated monthly temperature profiles, and the yearly temperature envelope described by the temperature profiles, resulting from nonsinusoidal yearly variation in stream temperature and ground-water velocities of *A*, 0.05 foot per day upward, *B*, 0.0 foot per day, and *C*, 0.05 foot per day downward.

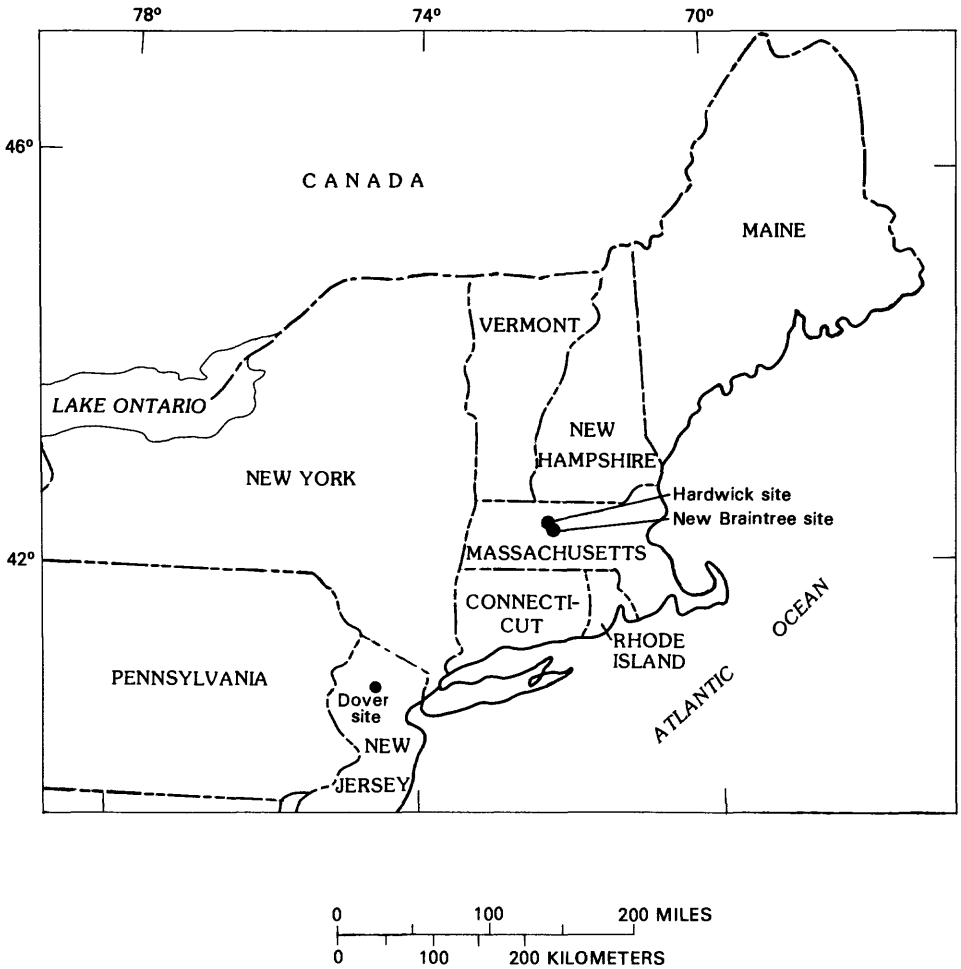


Figure 9. Locations of the Hardwick and New Braintree, Mass., and Dover, N.J., field sites, Northeastern United States.

Head gradients between the 25-ft-deep piezometer and the river varied seasonally from about 0.10 (2.57 ft/25 ft) to 0.15 (3.87 ft/25 ft) upward during the period from October 1984 to October 1985. The head in the 25-ft piezometer always was higher than the head in the river, indicating that ground-water flow is upward in the top 25 ft of sediment beneath the river bottom. Although the vertical gradients within the top 25 ft beneath the river at this site are high, vertical ground-water velocity is probably close to zero because of the low vertical hydraulic conductivity of the lacustrine silt and clay composing the sediments in the top 60 ft beneath the river bottom (fig. 11).

Physical and Thermal Properties of Sediments Underlying the Hardwick Site

Six cores of the sediments within the top 7 ft beneath the streambed of the Ware River at the Hardwick site were collected for determination of sediment wet-bulk and dry-bulk densities and thermal properties for input into the model. Some of these cores were collected using an open-ended hand auger and some by hand-driving a ½-in (inch) pipe into the sediment and extracting the cored sediment. These methods of collection undoubtedly disturbed the sediments and affect the in situ wet- and dry-bulk densities of the sediments to some extent. Average wet-bulk density of these six cores was 1.5 g/cm^3, and average dry-bulk density was 1.0 g/cm^3 (table 2). Using the average dry-bulk density of 1.0 g/cm^3 and the relation between dry-bulk density and thermal properties of fine-grained sediments (fig. 2), the thermal conductivity of the top 7 ft of the saturated sediments (clay, silt, fine sand, and organics, table 2) at the Hardwick site was estimated to be 0.0022 cal/s-cm-°C, the volumetric heat capacity was estimated to be 0.85 cal/cm^3-°C, and the thermal diffusivity was estimated to be 0.0026 cm^2/s. It was assumed that these thermal properties represent the thermal properties of the entire top 60 ft of lacustrine sediments underlying the site.

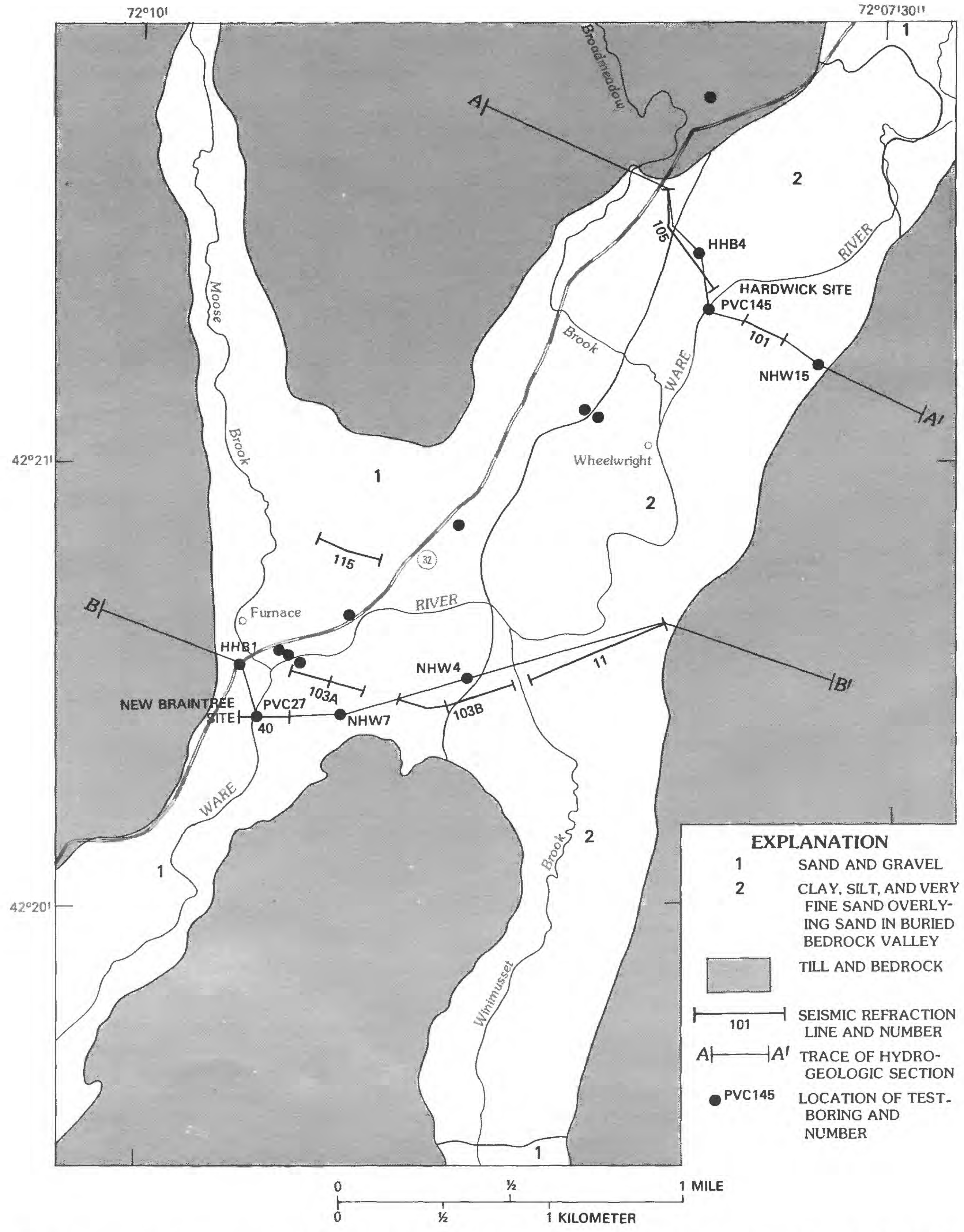


Figure 10. Unconsolidated deposits of the Ware River valley near the Hardwick and New Braintree sites, Hardwick and New Braintree, Mass.

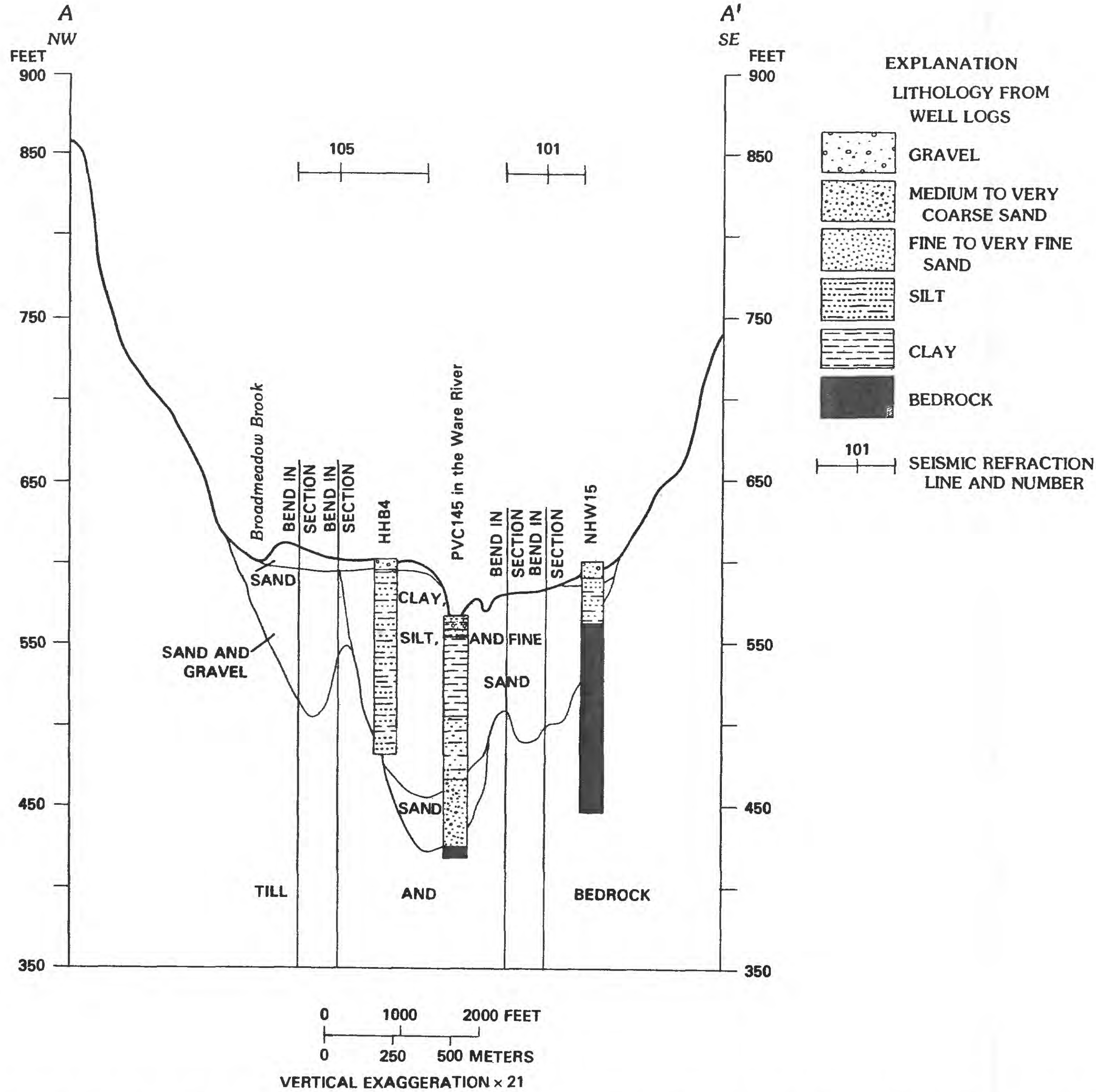


Figure 11. Hydrogeologic section *A–A'* at the Hardwick site, Hardwick, Mass. Trace of section is shown in figure 10.

Temporal Variation in Temperature of the Ware River at the Hardwick Site

The temperature of the Ware River at the Hardwick site was measured monthly from October 1984 to November 1985. In addition, the temperature of the Ware River at Hardwick Road about 1.5 mi south of the Hardwick site (fig. 10) was measured weekly beginning in July 1985. All measurements were made at 9 a.m., or as close to that time as possible. River temperature varies somewhat with depth in the river and from location to location along the river, and also diurnally. Variation of temperature with depth and from location to location generally was less than 1 °C. Diurnal river temperature varied as much as about 5 °C. These two sets of temperature measurements were used to model temporal variation in temperature of the Ware River at the Hardwick site. The program used to model this temporal variation was written by Tasker and Burns (U.S. Geological Survey, written commun., 1985). The program computes the best-fit harmonic equation describing daily stream temperature at a site, the period of which may be less than 365 d. During the remaining days, temperature is assumed to be 0 °C. A description of the mathematical

Table 2. Physical properties of fine-grained sediments at the Hardwick site, Hardwick, Mass.

[g/cm^3, grams per cubic centimeter]

Depth below river bottom, in feet	Lithologic description	Bulk density, in g/cm^3	
		Dry	Wet
1.7	Silty, sandy clay, some organics	0.8	1.2
2.1	Clayey silt, some fine sand	.9	1.2
3.3	Silt and clay, some fine sand	1.0	1.4
4.2	Silty clay	1.1	1.4
5.4	Fine to very fine sand and silt	1.2	1.8
7.1	Fine sand with silt	1.0	1.8
	Average	1.0	1.5

derivations used for the optimization of the harmonic function are described in Tasker and Burns (1974).

The "best fit" equations describing the temporal variation in the temperature of the Ware River at the Hardwick site from October 1984 to October 1985 are

$$T_{stream}=11.2+11.2(\sin(((2\pi/344)d+365)-2.2)) \text{ for } 0\leq d<14 \quad (9)$$

$$T_{stream}=0 \quad \text{for } 14\leq d\leq 35 \quad (10)$$

and

$$T_{stream}=11.2+11.2(\sin(((2\pi/344)d)-2.2)) \quad \text{for } 35<d\leq 365 \quad (11)$$

where

T_{stream}=temperature of the stream, in degrees Celsius, and

d=time, in days from January 1.

The stream temperature measurements and the best-fit curve are shown in figure 12. The temperature of the Ware River at the Hardwick site remained near zero for about one-half month during the winter of 1984–85. The stream-temperature-model results using temperature data collected from October 1984 to November 1985 indicate that the temperature of the Ware River remained at 0 °C for 21 consecutive days during the year, from day 14 to 35 (day 0=January 1). Therefore, the best-fit harmonic curve has a period of 344 d, an average stream temperature and semi-amplitude of the harmonic function of 11.2 °C, and a temperature range of 0 to 22.3 °C. The maximum temperature occurs on Julian day 207.

Determination of Vertical Ground-Water Velocity and Effective Vertical Hydraulic Conductivity

Temperature profiles in Hardwick PVC145 were measured monthly from October 1984 to October 1985 to a depth of about 100 ft (fig. 13). PVC145 is constructed of 143 ft of 1.25-in-diameter PVC pipe and has a 2-ft slotted screen at the bottom. As indicated in figure 13, ground-water temperatures equaled 9.8 °C, within ±0.1 °C, below a depth of about 28 ft beneath the river bottom throughout the period of measurement.

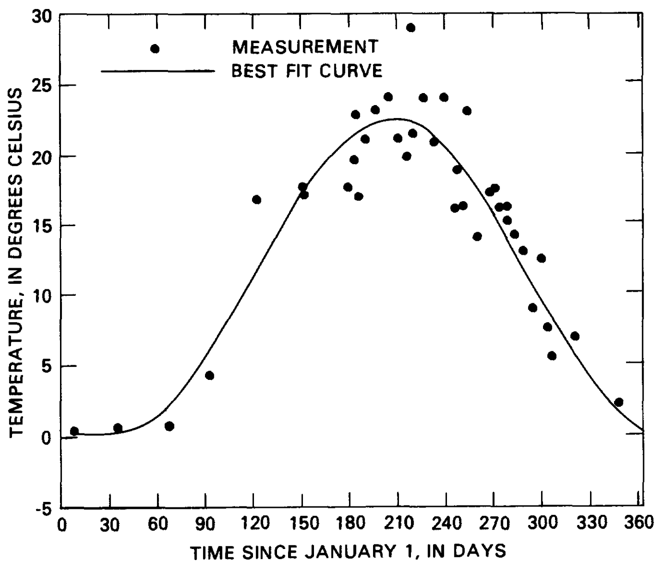


Figure 12. Temperature of the Ware River at the Hardwick and New Braintree, Mass., sites, and best-fit curve of temperature.

The temperature profiles drawn from temperatures measured in PVC145 during the winter months appear more compressed upward toward the river bottom than the profiles from temperatures measured during the summer months. The differences between winter and summer profiles are partly attributable to the difference in the method of measuring the winter and summer profiles. All piezometers, including PVC145, were cut off and capped near the river bottom. There were two reasons for cutting off and capping the piezometers near river bottom: to prevent vandalism, and to prevent the column of water in each piezometer, all of which rise above river level, from freezing during winter. Before measuring temperatures in PVC145, a 6-ft extender was attached to the piezometer at river bottom and the water column in the piezometer was allowed to rise to its equilibrium position in the extender, which was about 5 ft above river bottom (a 2-ft depth of water in the river plus a 3-ft head of water in PVC145 above the river surface). The 3-ft rise of the column of water above the river surface caused the temperature at all depths in the piezometer to differ somewhat from the temperature of ground water in the surrounding sediments. During summer, all piezometers, including PVC145, were left for several days after attaching the extenders and prior to measuring water levels and temperature profiles to allow water levels to stabilize and temperature of the water columns to equilibrate to the temperature of the surrounding sediments. However, during the winter months, the temperature profile in PVC145 had to be measured the same day the extender was attached, before the water column froze. Consequently, temperature profiles drawn from temperatures measured in PVC145 during the winter months may appear more compressed

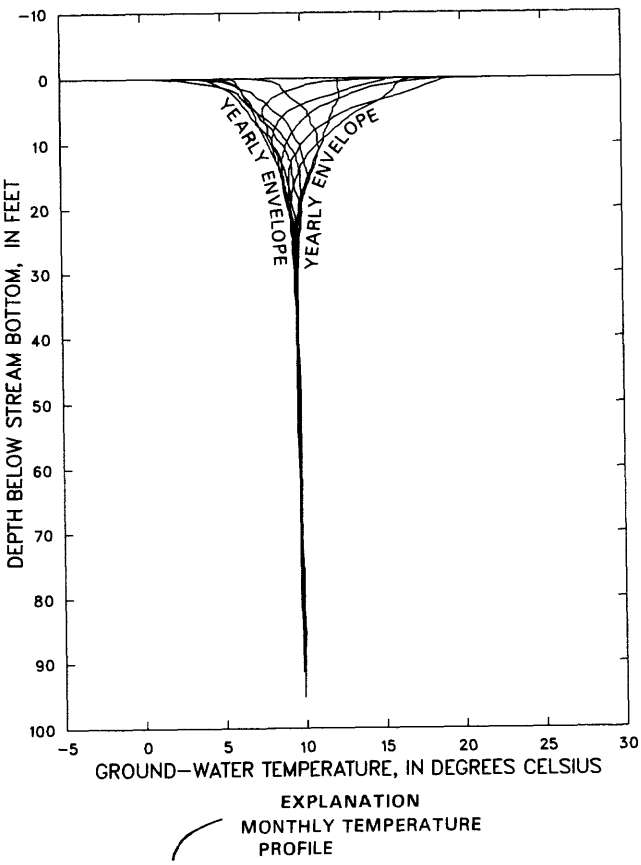


Figure 13. Temperature profiles measured approximately monthly in Hardwick piezometer PVC145 at the Hardwick site, Hardwick, Mass., and the yearly temperature envelope described by the temperature profiles.

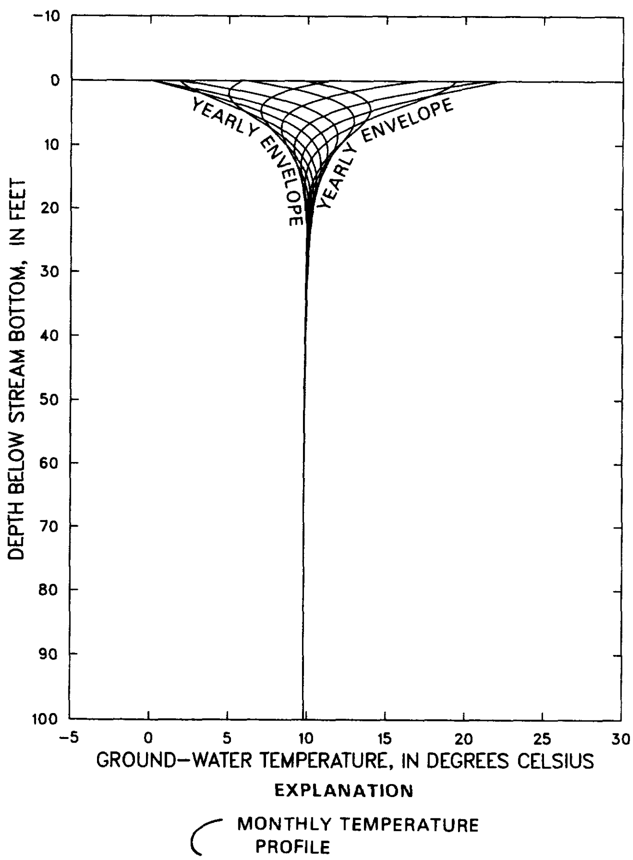


Figure 14. Simulated monthly temperature profiles, and the yearly temperature envelope described by the temperature profiles, at the Hardwick site, Hardwick, Mass., for a ground-water velocity of 0.0 foot per day.

toward the river bottom than the actual profiles in the sediments adjacent to PVC145. Additional study is required to determine how quickly temperatures in PVC145 equilibrate to temperatures in the sediments surrounding the casing. Because of uncertainties introduced by the winter measurements, the summer half of the temperature envelope in PVC145 may provide a more reliable indication of the true thermal conditions at the Hardwick site than the winter half of the temperature envelope.

The finite-difference model described by equation 5 was used to simulate theoretical monthly temperature profiles and the yearly temperature envelope at the Hardwick site. Values of the physical and thermal properties used in the model simulations at the Hardwick site were as follows: wet-bulk density, 1.5 g/cm^3; thermal conductivity, 0.0022 cal/s-cm-°C; and volumetric heat capacity, 0.85 cal/cm^3-°C. The upper thermal-boundary condition used in the model is given in equations 9–11. The lower thermal-boundary condition used in the model was the ambient ground-water temperature of 9.8 °C, held constant at a depth of 100 ft below river bottom. Ground-water velocity was varied in the model until the best visual match between the model-simulated envelope and the yearly temperature envelope determined from field-measured temperature profiles in PVC145 was achieved. Model-simulated instantaneous temperature profiles were not matched explicitly to field-measured profiles, but instead were used only to define the yearly temperature envelope.

The model-simulated yearly temperature envelope that best fits the field-measured temperature envelope indicates that ground-water velocity at the site within the top 30 ft of sediment beneath the stream is upward at a rate of less than 0.01 ft/d (figs. 14, 15). Differences in the envelopes resulting from velocities of 0 and 0.01 ft/d, as discussed in the section describing sensitivity testing, are so small that the actual velocity, for velocities less than 0.01 ft/d, is not easily determined.

The maximum measured vertical head difference across the top 25 ft of sediment beneath the river bottom between October 1984 and October 1985 was 3.87 ft, and the minimum head difference was 2.57 ft (table 3). The minimum gradient across the top 25 ft of sediment, therefore, was 0.10 ft/ft (foot per foot) upward. For a vertical upward velocity of 0.01 ft/d determined from the tempera-

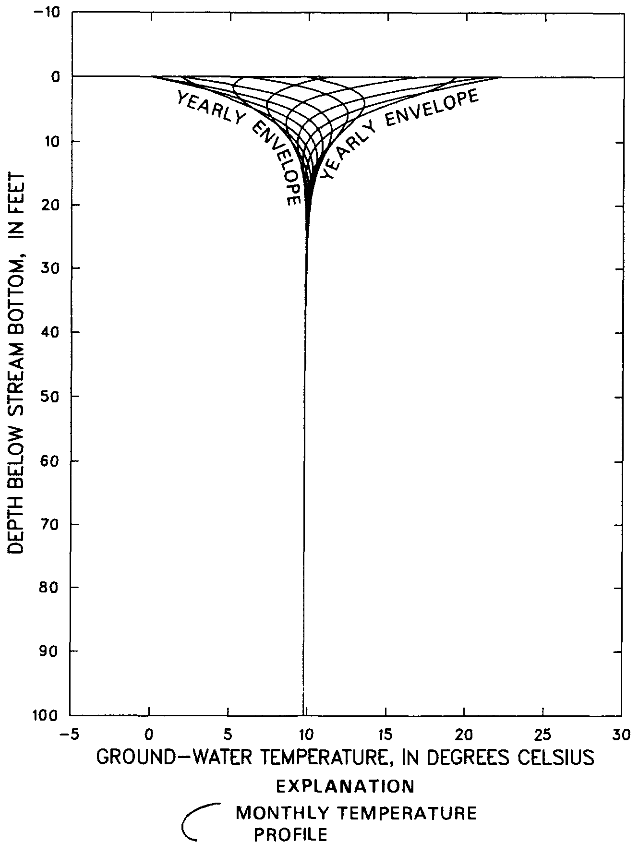


Figure 15. Simulated monthly temperature profiles, and the yearly temperature envelope described by the temperature profiles, at the Hardwick site, Hardwick, Mass., for an upward ground-water velocity of 0.01 foot per day.

ture envelope and the minimum gradient of 0.10 ft/ft, the maximum vertical effective hydraulic conductivity is calculated, using Darcy's law, to be 0.1 ft/d for the top 25 ft of lacustrine silts and clays underlying the site.

New Braintree, Massachusetts, Site

The New Braintree site is located about 1.5 mi southwest of the Hardwick site in the Ware River (fig. 10). This site was selected because it is underlain by coarse-grained sediments in a stream reach where ground water discharges to the river. Ground-water discharge at this site was anticipated to be high because the site is located just southwest of the western edge of the lacustrine clay cap that overlies the buried sand aquifer at the Hardwick site (fig. 10). As discussed previously, ground water appears to flow south-southwest in the buried sand aquifer in the vicinity of the Hardwick site, partly because it is unable to discharge upward to the Ware River through the low-conductivity lacustrine clay cap. Flow probably continues south-southwest in the buried aquifer until it reaches the western edge of this clay cap, beyond which it discharges upward to the river.

The river bottom for several miles of the river in the vicinity of the site consists of sand and gravel with numerous cobbles. Only coarse-grained sediments ranging from fine sand to sand and gravel were encountered in test boring PVC27 at the New Braintree site (figs. 10, 16). Refusal occurred at 27 ft beneath the river bottom. Data from seismic refraction lines in the area and cuttings from the b ›ttom of the bore at PVC27 indicate that refusal was on bedrock.

Two stream-discharge measurements were made on August 12, 1986, to provide an independent estimate of ground-water discharge near the New Braintree site. Stream discharge at the upstream end of the reach, just north of PVC27, was 71.01 ft^3/s (cubic feet per second), and discharge at the downstream end of the reach, 1.6 mi to the south, was 73.56 ft^3/s. The gain in discharge, 2.55 ft^3/s, divided by the area of streambed along the reach, approximately 427,000 ft^2, results in an average upward velocity along the reach of 0.52 ft/d. However, by assuming a ± 5 percent error in each measurement of discharge, the change in discharge along the reach could range from a decrease in discharge of 4.68 ft^3/s to an increase of 9.78 ft^3/s. This range of estimates results in a possible range of ground-water velocities along the reach of from 0.95 ft/d downward to 1.97 ft/d upward.

Five piezometers were installed at different locations in the river in the vicinity of the New Braintree site for measurement of temperature profiles, water levels, and hydraulic gradients. Four 1.25-in-diameter steel pipes having 1.25-in-diameter, 1.5-ft-long slotted drive points at the bottom were hand-driven to depths of 11.8, 16.8, 16.5, and

Table 3. Maximum vertical velocity and maximum effective vertical hydraulic conductivity of sediments beneath the Ware River at the Hardwick site, Hardwick, Mass.

[ft, feet; ft/ft, foot per foot; ft/d, foot per day; cm/s, centimeter per second]

Piezometer	Minimum difference in head between piezometer and river, in ft	Thickness of sediment over which head drop occurs, in ft	Minimum hydraulic gradient, in ft/ft	Vertical darcian velocity determined from temperature envelopes in PVC145, in		Maximum effective vertical hydraulic conductivity, in	
				ft/d	cm/s $\times 10^{-6}$	ft/d	cm/s
PVC25	2.57	25.0	0.10	0.01	3.5	0.10	3.5×10^{-5}

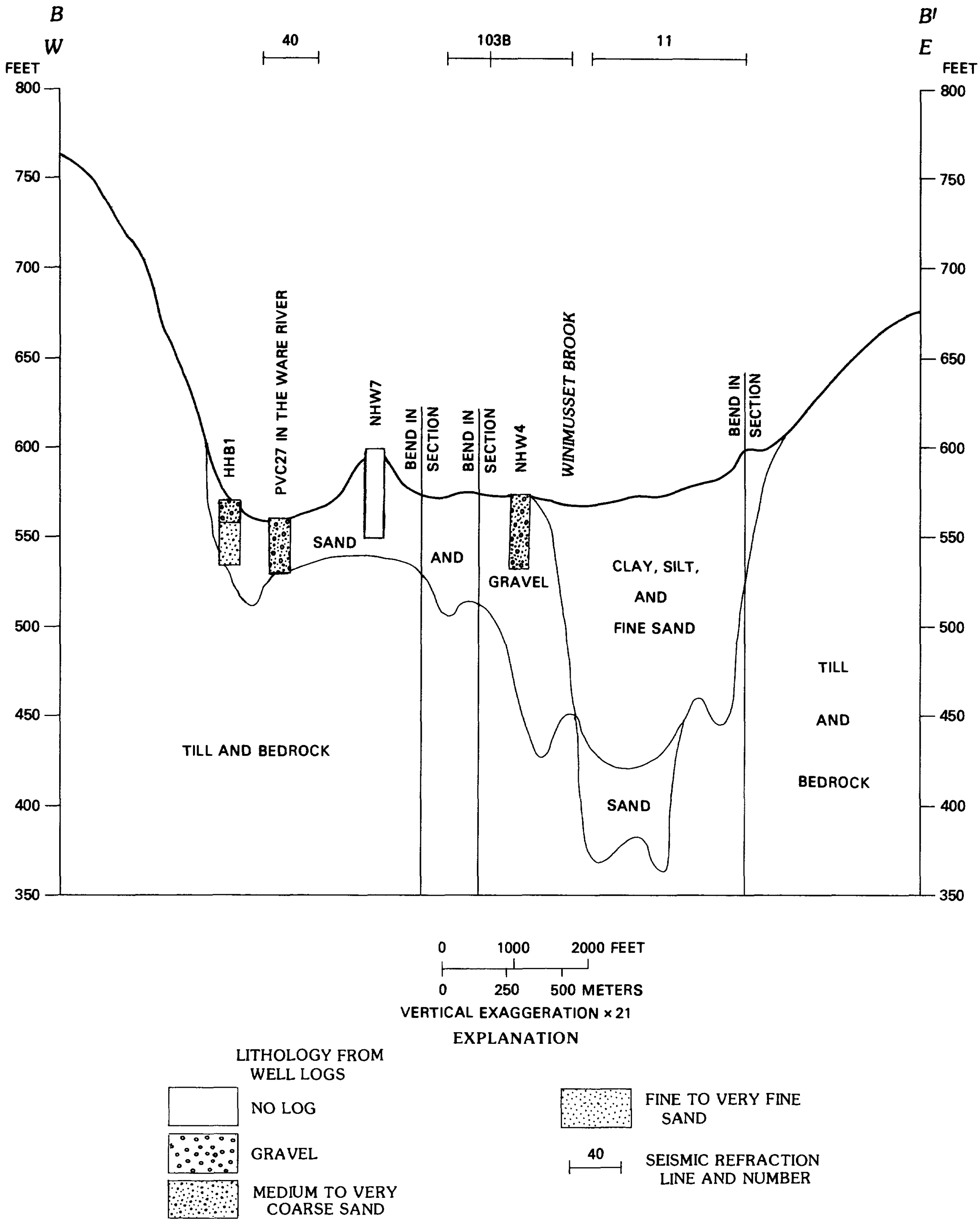


Figure 16. Hydrogeologic section *B–B'* at the New Braintree site, New Braintree, Mass. Trace of section is shown in figure 10.

16.0 ft in the sediments beneath the river at four locations (Steel piezometers 1–4, respectively) during June 1985. A fifth piezometer, PVC27 (figs. 10, 16), which consists of 25 ft of 1.25-in-diameter PVC pipe with a 2-ft-long slotted screen at the bottom, was installed beneath the river by a drilling rig in October 1985. Steel1 is located about 5 ft from the western bank of the Ware River about 6 ft downstream from PVC27. Steel2 is located in the center of the river about 37 ft from Steel1. Steel3 and Steel4 are located about 270 ft upstream from Steel1 and Steel2.

Table 4. Physical properties of coarse-grained sediments at the New Braintree site, New Braintree, Mass.
[g/cm³, grams per cubic centimeter]

Depth below river bottom, in feet	Lithologic description	Bulk density, in g/cm³	
		Dry	Wet
0–1	Fine sand with some gravel	1.6	1.9
7–9	Fine sand	1.6	1.9
	Average	1.6	1.9

Steel3 is located within about 5 ft of the western bank of the river, and Steel4 is located in the center of the river about 23 ft from Steel3.

Physical and Thermal Properties of Sediments Underlying the New Braintree Site

During the drilling of test hole PVC27, two split-spoon cores of the sediments underlying the river were taken to determine wet-bulk and dry-bulk densities (table 4) and the thermal properties for input into the model. The first sample was collected with a split-spoon sampler from 0 to 1 ft beneath the river bottom and consisted of mixed fine sand to gravel. The second sample also was collected with a split-spoon sampler at a depth of 7 to 9 ft beneath the river bottom and consisted of fine sand. Although the split-spoon samples consisted predominantly of fine sand, the wash samples from the drive-and-wash drilling indicated that fine to coarse sand and gravel is present throughout the top 27 ft of sediment beneath the stream bottom. The wet-bulk density of both samples was 1.9 g/cm³, and the dry-bulk density of both samples was 1.6 g/cm³. Using the dry-bulk density of 1.6 g/cm³ and the relationship between dry-bulk density and thermal properties of sediments (fig. 2), the thermal conductivity of the sediments at the New Braintree site was estimated to be 0.0046 cal/s-cm-°C, volumetric heat capacity was estimated to be 0.63 cal/cm³-°C, and thermal diffusivity was estimated to be 0.0073 cm²/s.

Temporal Variation in Temperature of the Ware River at the New Braintree Site

The yearly temporal variation in temperature of the Ware River at the New Braintree site for use as the upper thermal-boundary condition in the model was assumed to be the same as that estimated at the Hardwick site 1.5 mi to the north during the period October 1984 to October 1985. This temporal variation was assumed even though the temperature profiles at the New Braintree site were measured from July 1985 to July 1986. The stream temperature measurements and the best-fit curve used for both the Hardwick and New Braintree sites are shown in figure 12. The equations that describe this variation are equations 9–11.

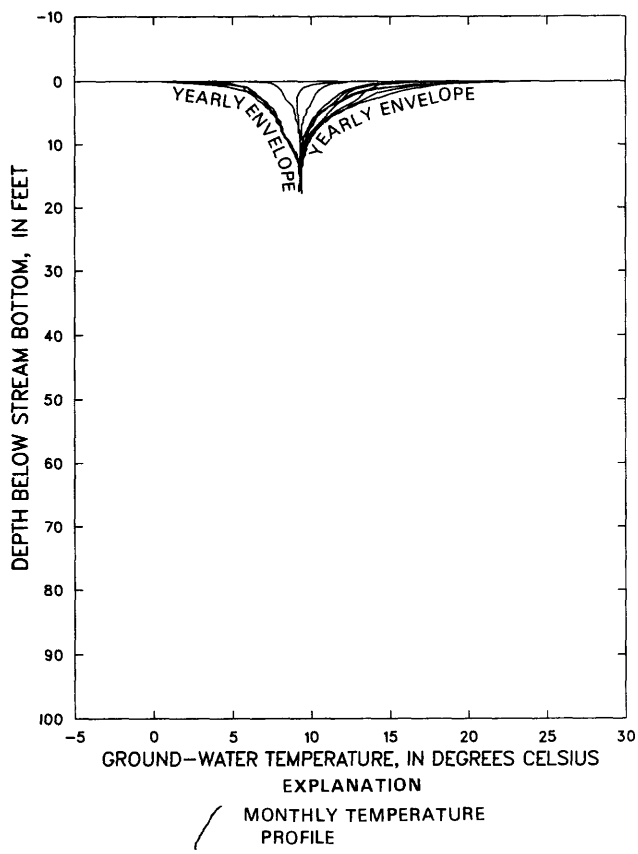


Figure 17. Temperature profiles measured approximately monthly in New Braintree piezometer Steel2 at the New Braintree site, New Braintree, Mass., and the yearly temperature envelope described by the temperature profiles.

Determination of Vertical Ground-Water Velocity and Effective Vertical Hydraulic Conductivity

Temperature profiles in piezometers Steel1–4 were measured monthly beginning in June 1985, and temperature profiles in PVC27 were measured monthly beginning in November 1985. Only temperature profiles in Steel2, 3, and 4 are shown in this report (figs. 17–19). Two conclusions can be reached regarding the characteristics of the profiles and envelopes in the three figures: (1) ambient ground-water temperature is about 9.0 °C, and (2) ground-water temperature is nearly constant at depths below the river bottom that range from about 10 ft in Steel3 to about 18 ft in Steel4. These depths to constant ground-water temperature are relatively shallow compared with the depth to constant ground-water temperature of about 28 ft in Hardwick PVC145 (fig. 13). The relatively shallow depths suggest that upward ground-water velocity at the New Braintree site is higher than at the Hardwick site. However, the differences in the thermal diffusivity of sediments at the two sites and the effects of these differences on the temperature profiles must be accounted for before this

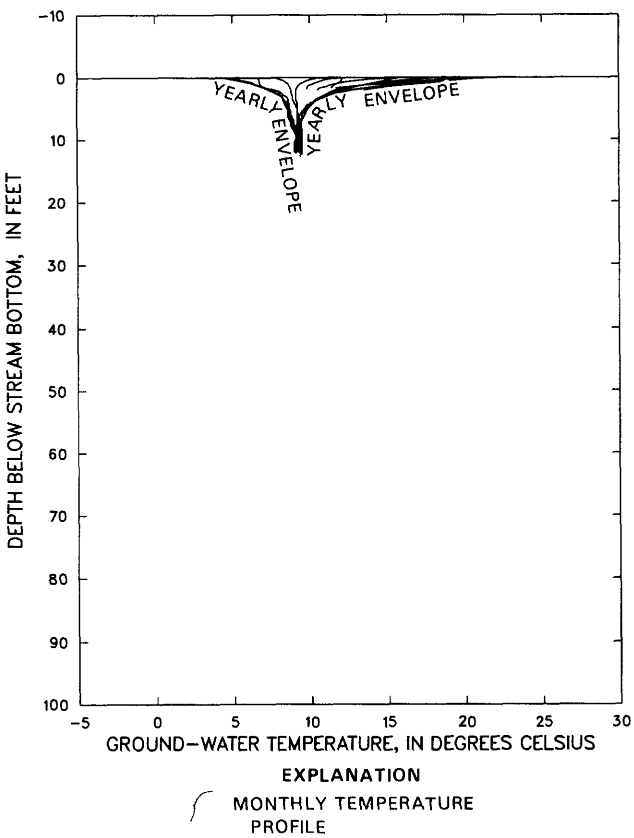


Figure 18. Temperature profiles measured approximately monthly in New Braintree piezometer Steel3 at the New Braintree site, New Braintree, Mass., and the yearly temperature envelope described by the temperature profiles.

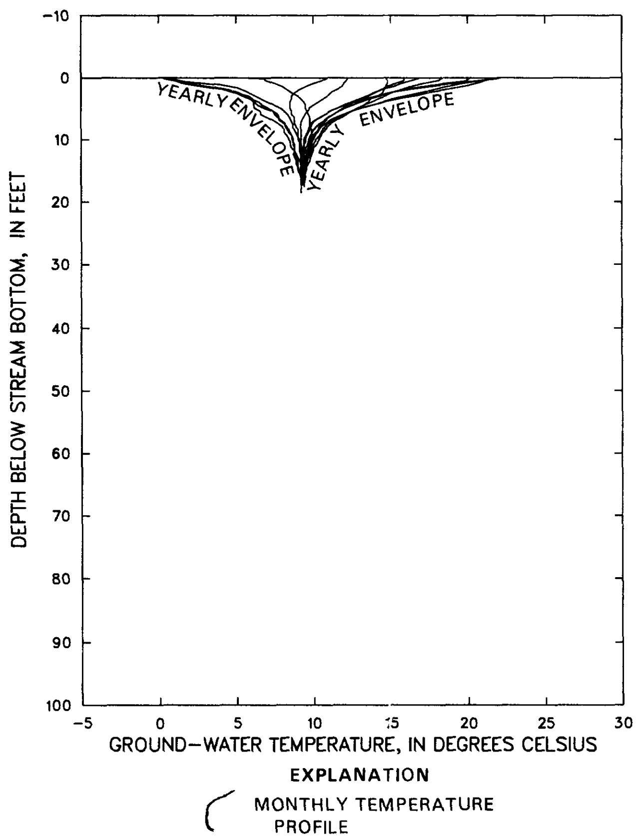


Figure 19. Temperature profiles measured approximately monthly in New Braintree piezometer Steel4 at the New Braintree site, New Braintree, Mass., and the yearly temperature envelope described by the temperature profiles.

conclusion regarding relative velocities at the two sites can be made. The differences in the depths to constant ground-water temperature in the three piezometers at the New Braintree site also suggest that the rate of ground-water discharge to the river varies somewhat from point to point along this reach of the Ware River. The differences in velocity along the reach probably are attributable to point-to-point variation in geology and streambed lithology along the reach.

The finite-difference model described by equation 5 was used to simulate theoretical monthly temperature profiles and the yearly temperature envelope at the New Braintree site. The physical and thermal properties of the sediments beneath the river (described in the section "Physical and Thermal Properties of Sediments Underlying the New Braintree Site") used in the model were as follows: wet-bulk density, 1.9 g/cm^3; thermal conductivity, 0.0046 cal/s-cm-°C; and volumetric heat capacity, 0.63 cal/cm^3-°C. The upper thermal-boundary condition, which describes the temporal variation in the temperature of the Ware River, is given in equations 9–11, and the lower thermal-boundary condition is the ambient ground-water temperature of 9.0 °C, held constant at a depth of 50 ft beneath the river bottom. Ground-water velocity in the model was varied until the best visual match between the model-simulated yearly temperature envelope and the envelope resulting from the field-measured temperature profiles in each of the three piezometers was achieved.

Model-simulated monthly temperature profiles and the yearly envelope for upward velocities of 0.075, 0.10, and 0.20 ft/d are shown in figures 20–22, respectively. Velocities that resulted in the best match of model-simulated envelopes to field-determined envelopes ranged from 0.10 to 0.20 ft/d upward in the three piezometers: 0.12 ft/d at Steel2, 0.20 ft/d at Steel3, and 0.10 ft/d at Steel4. Results are summarized in table 5.

The stage of the Ware River and water levels in the piezometers at the New Braintree site were measured monthly at the same time temperatures were measured. The range in river-stage measurements during the period July 1985 to July 1986 was 0.98 ft. Water levels in the piezometers were relatively high when river stage was high

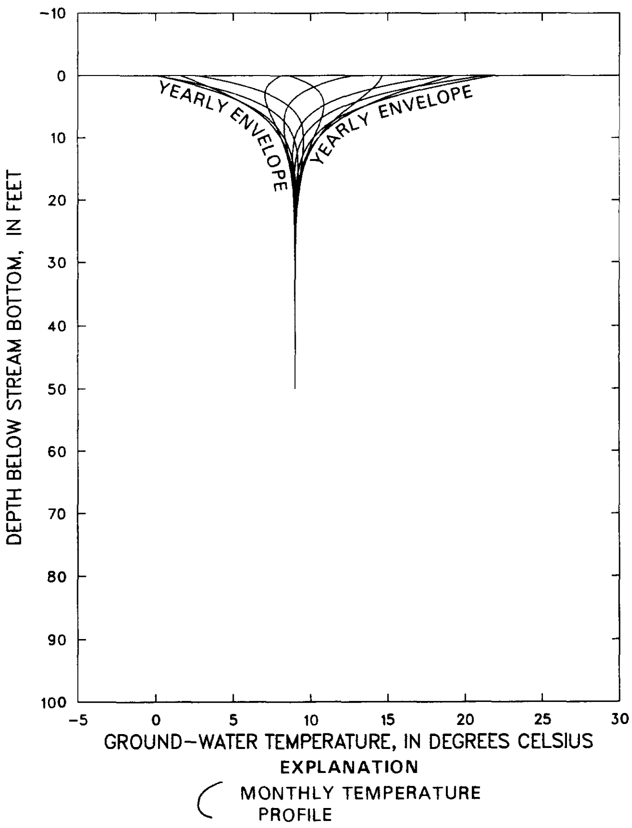


Figure 20. Simulated monthly temperature profiles, and the yearly temperature envelope described by the temperature profiles, at the New Braintree site, New Braintree, Mass., for an upward ground-water velocity of 0.075 foot per day.

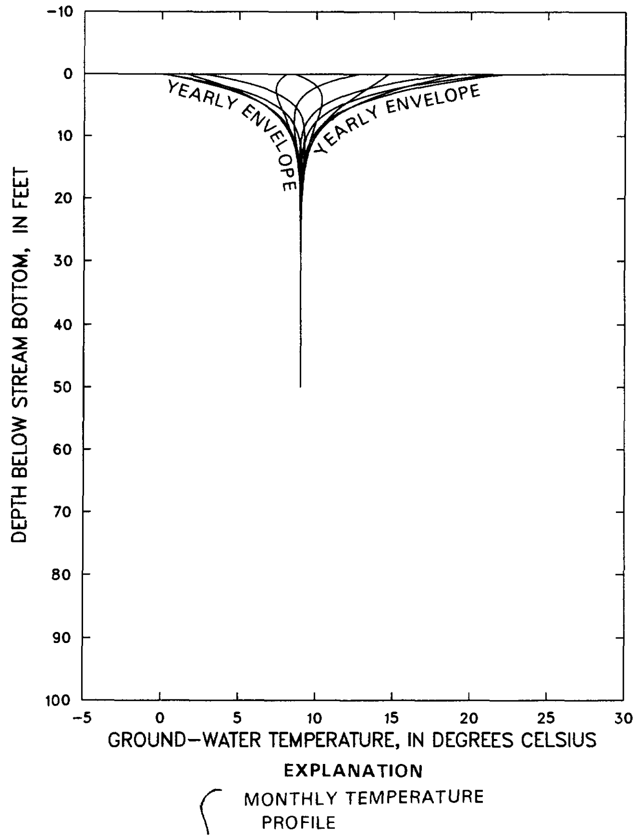


Figure 21. Simulated monthly temperature profiles, and the yearly temperature envelope described by the temperature profiles, at the New Braintree site, New Braintree, Mass., for an upward ground-water velocity of 0.10 foot per day.

and relatively low when river stage was low. However, fluctuations in water levels in the piezometers were much less than fluctuations in river stage. For example, the maximum difference in the water level in piezometer Steel4 was 0.37 ft during the period July 1985 to July 1986.

Minimum and maximum differences in head across the sediments at the three piezometers ranged from 0.12 to 0.68 ft, and hydraulic gradients ranged from 0.007 to 0.042 ft/ft (table 5). The velocity and the minimum and maximum hydraulic gradients at each piezometer were used in Darcy's law to calculate minimum and maximum effective vertical hydraulic conductivities at each piezometer (table 5). Conductivities ranged from 2.4 to 17.1 ft/d for the sand and gravel underlying the site.

Dover, New Jersey, Site

The Dover site (fig. 9) is located on the western edge of the town of Dover, N.J., along the Rockaway River at

Table 5. Vertical velocities and effective vertical hydraulic conductivities of sediments beneath the Ware River at the New Braintree site, New Braintree, Mass.

[ft, feet; ft/ft, foot per foot; ft/d, foot per day; cm/s, centimeter per second]

Piezometer	Difference in head between piezometer and river, in ft		Thickness of sediment over which head drop occurs, in ft	Hydraulic gradient, in ft/ft		Vertical darcian velocity determined from temperature envelopes, in		Effective vertical hydraulic conductivity, in			
	Min	Max		Min	Max	ft/d	cm/s×10^{-5}	ft/d Max	ft/d Min	cm/s Max ×10^{-3}	cm/s Min ×10^{-3}
Steel2	0.12	0.23	16.8	0.007	0.014	0.12	4.2	17.1	8.6	6.0	3.0
Steel3	.23	.48	16.5	.014	.029	.20	7.0	14.3	6.9	5.0	2.4
Steel4	.31	.68	16.0	.019	.042	.10	3.5	5.3	2.4	1.9	.080

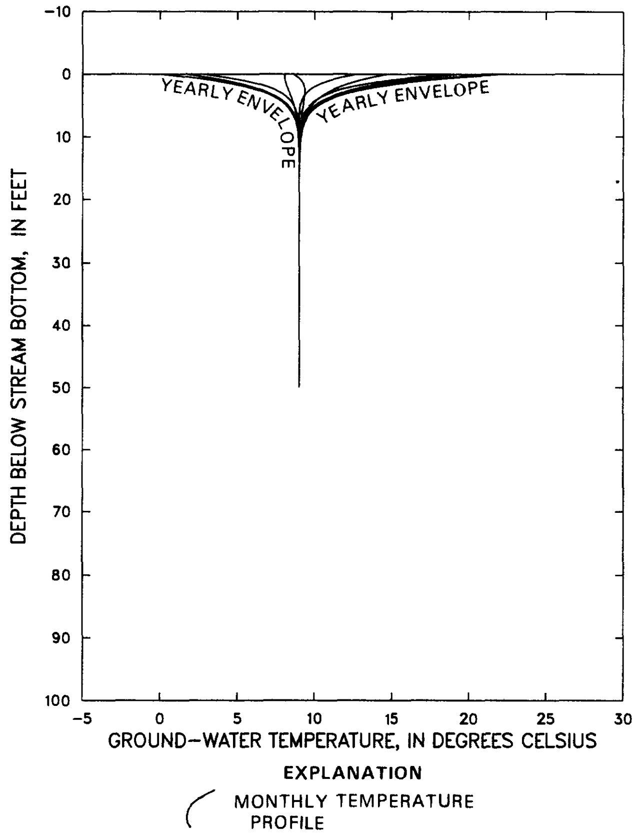


Figure 22. Simulated monthly temperature profiles, and the yearly temperature envelope described by the temperature profiles, at the New Braintree site, New Braintree, Mass., for an upward ground-water velocity of 0.20 foot per day.

the Princeton Avenue well field (fig. 23). This site was selected because municipal pumping is causing infiltration of water from the Rockaway River into the underlying aquifer.

The river bottom consists of coarse sand and gravel with numerous cobbles. Stratified drift having thicknesses as great as 100 ft underlies the site. The bottom 30 to 40 ft of drift is lacustrine silt and sand (fig. 24). The upper 60 to 70 ft of drift, in which the municipal wells are screened, is composed of sand and gravel. Rheaume and others (U.S. Geological Survey, 1987, written commun.) contains a detailed description of the hydrogeology in the vicinity of the site.

Three piezometers were installed at different locations in the river in the vicinity of the Dover site for measurement of temperature profiles and hydraulic gradients. The piezometers, which consisted of 1.25-in-diameter steel pipes with 1.25-in-diameter, 0.5-ft-long slotted drive points at the bottom, were hand-driven to depths of 3.5 ft beneath the river. P2 is located about 3 ft from the eastern bank of the river, P3 is located in the center of the river near P2, and P4 is located in the center of the river about 450 ft downstream from P2 and P3 (fig. 23).

The rate of infiltration near piezometer P2 (fig. 23) has been estimated by tracing peaks in pH and dissolved-oxygen concentrations in the river water as the river water infiltrates the aquifer (Rheaume and others, U.S. Geological Survey, 1987, written commun.). Results indicate that interstitial velocity near P2 ranged from about 7.9 to 13.2 ft/d. Assuming a porosity of 0.25, the Darcy velocity (equivalent to the velocity terms in equations 2 and 5), therefore, ranges from 2.0 to 3.3 ft/d.

Results of discharge measurements made to estimate seepage rates along the Rockaway River at the site were inconclusive because the decrease in stream discharge along the reach was a small fraction of total stream discharge. Consequently, any actual decrease in discharge along the reach could not be differentiated from probable errors in the discharge measurements.

Physical and Thermal Properties of Sediments Underlying the Dover Site

The upper strata of the aquifer at the site are composed of coarse-grained sediments whose physical properties are reasonably well known as a result of well and test-hole borings. In the absence of cores of sediment from the borings, it was believed that average values of the physical and thermal properties of coarse-grained sediments, as given in table 1, could adequately represent conditions at the site in the model simulations. The average values used were as follows: wet-bulk density, 2.0 g/cm^3; thermal conductivity, 0.0041 cal/s-cm-°C; volumetric heat capacity, 0.60 cal/cm^3-°C; and thermal diffusivity, 0.0068 cm^2/s.

Temporal Variation in Temperature of the Rockaway River at the Dover Site

The temperature of the Rockaway River at the Dover site during the period 12 noon on June 2 to 7:30 a.m. on June 6, 1986, varied in a generally sinusoidal manner each day, with an amplitude of fluctuation of about 4 °C (fig. 25). Mean river temperature gradually increased from about 18 to 21 °C during the 4 d.

Determination of Vertical Ground-Water Velocity and Effective Vertical Hydraulic Conductivity

Temperature profiles in piezometers P2, P3, and P4 were measured every 2 h (hours) from the afternoon of June 2 to the afternoon of June 6, 1986. P3 and P4 were not pumped during the 4-d period. However, P2 was pumped for about 30 min (minutes) every 2 h at a rate of 0.1 gal/min (gallon per minute) for water-quality sampling. Each temperature profile in P2 was measured just prior to a 30-min

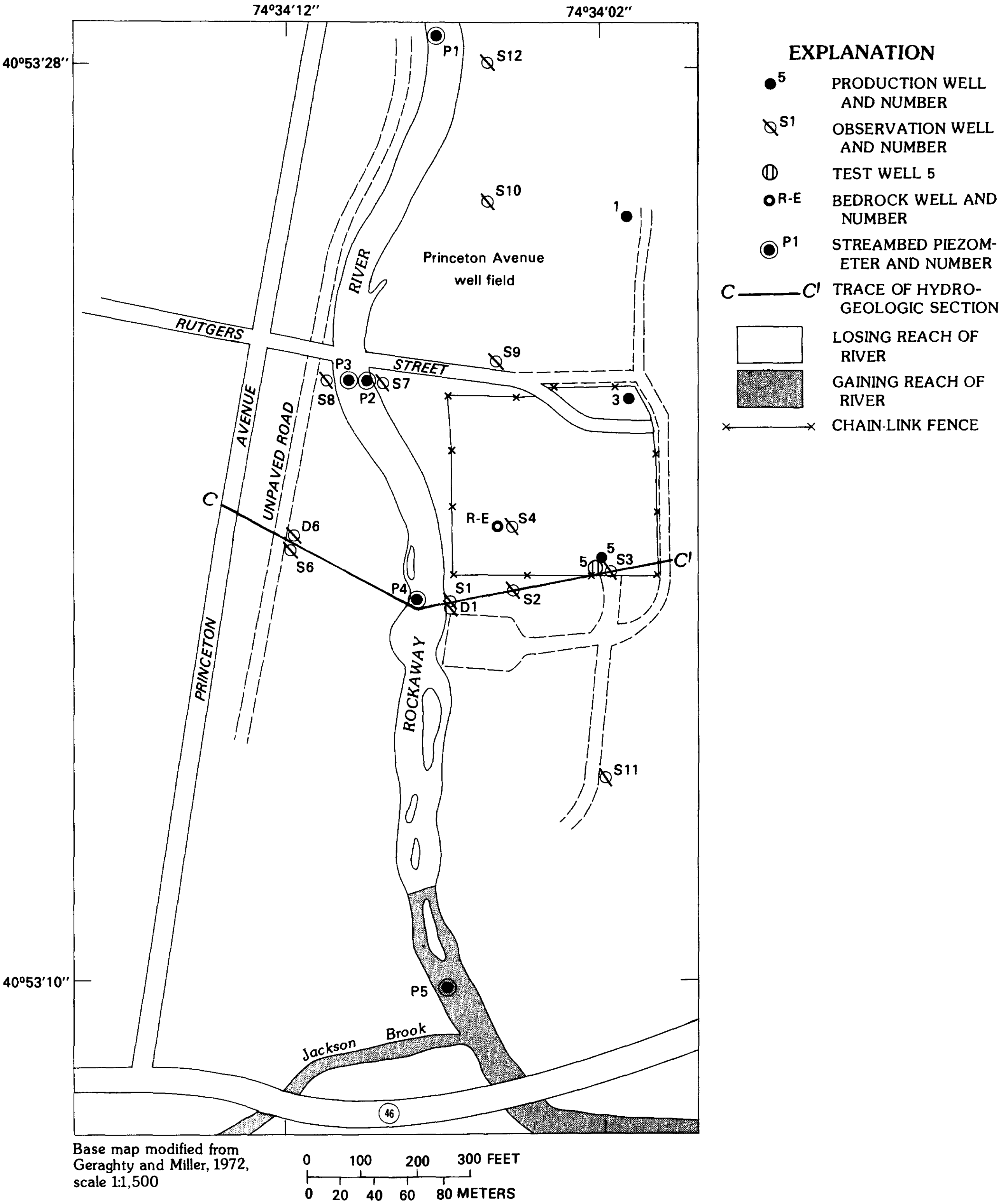


Figure 23. The Dover site in the Princeton Avenue well field, Dover, N.J.

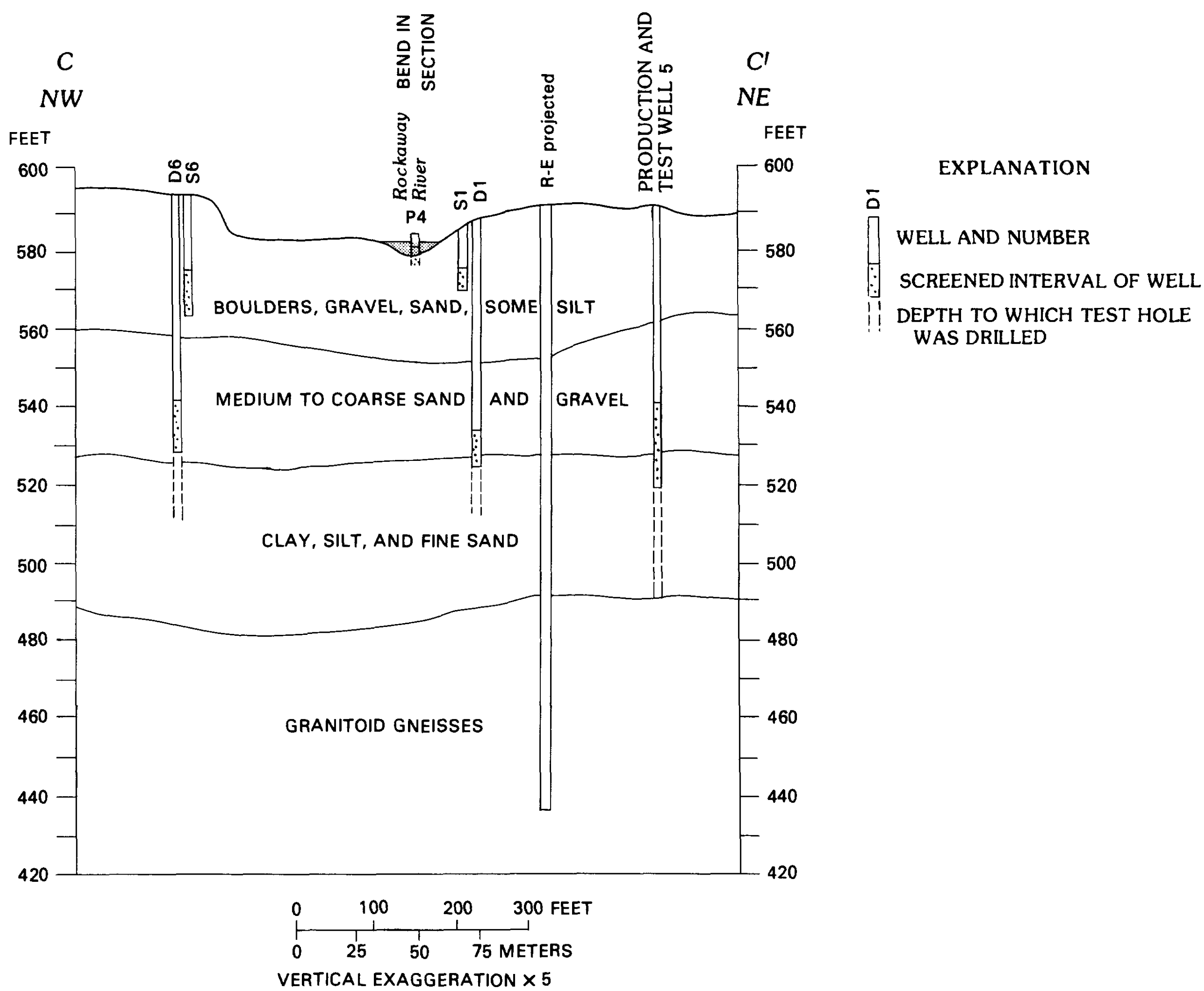


Figure 24. Hydrogeologic section C–C′ at the Dover site, Dover, N.J. Trace of section is shown in figure 23.

pumping period. The temperature profiles in P2 may be affected by the intermittent pumping.

Because the temperature of the Rockaway River did not vary in the same sinusoidal manner each day during the 4 d of measurement (fig. 25), the hourly profiles and daily envelope for each day differ somewhat from the profiles and envelopes for the other 3 d. Therefore, from this total period of 4 d of measurements, the 24-h period beginning at hour 47 (fig. 25) was selected for plots of hourly temperature profiles and daily temperature envelopes in the three piezometers and for subsequent determination of velocity and conductivity at the site. The 24-h period beginning at hour 47 was selected because the temperature profiles measured during this period were symmetrically distributed around the mean stream temperature, which was 19.5 °C. Temperature profiles in piezometers P2, P3, and P4 during this 24-h period are shown in figures 26–28, respectively.

There is no curvature to the profiles within the top 1 ft beneath the river bottom in figures 26–28; no temperature measurements were made there because water levels were about 1 ft below river bottom in the three piezometers. The plotted profiles, therefore, show a linear interpolation between river temperature and the temperature at the top of the water column in each piezometer, which may not be correct. Measurement of profiles in closed pipes filled with water and driven into the sediments beneath the river, rather than in piezometers, might eliminate the problem of missing temperature data within the top 1 ft below river bottom.

The finite-difference model described by equation 5 was used to simulate theoretical hourly temperature profiles and the daily temperature envelope at the Dover site during day 3 (hours 47 to 71) (alternatively, the analytical solution by Stallman, 1965, could be used). The physical and thermal properties of the sediments beneath the river (described in the section "Physical and Thermal Properties of Sediments Underlying the Dover Site") used in the model were as follows: wet-bulk density, 2.0 g/cm^3; thermal conductivity, 0.0041 cal/s-cm-°C; and volumetric heat capacity, 0.68 cal/cm^3-°C. The upper thermal-boundary condition used in the model, which describes the approxi-

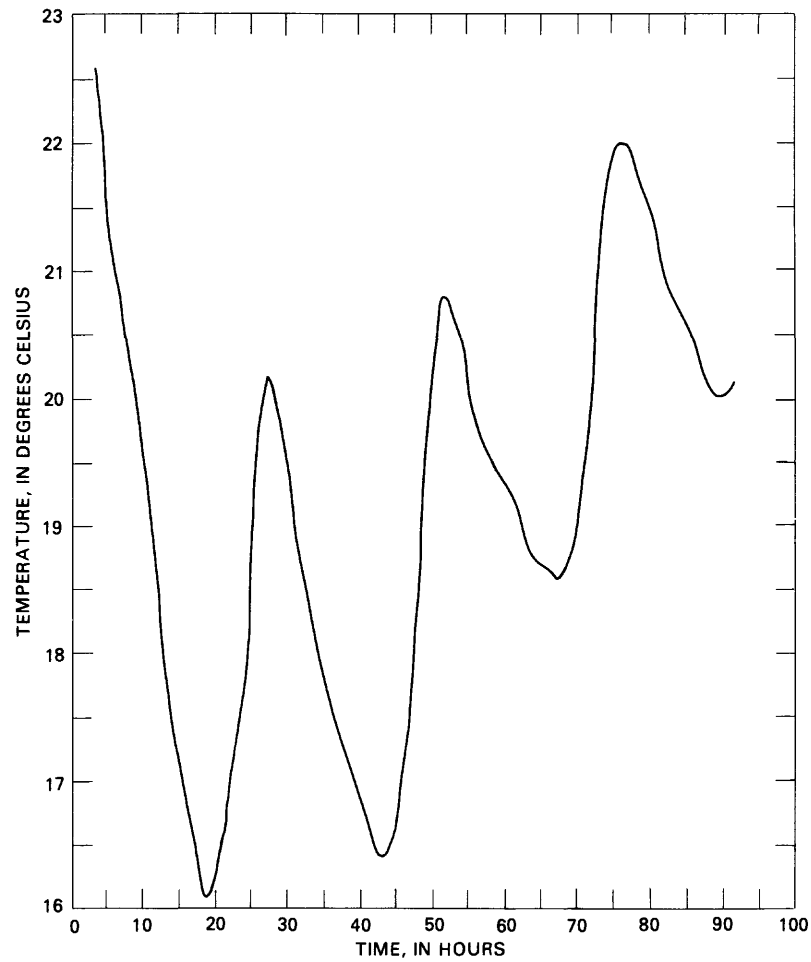


Figure 25. Temperature of the Rockaway River at Dover, N.J., from 12:00 noon on June 2, 1986 (hour 0), to 7:30 a.m. on June 6, 1986 (hour 91.5).

mate temporal variation in temperature of the Rockaway River during the 24-h period from hours 47 to 71, was varied sinusoidally with a mean stream temperature of 19.5 °C and a semiamplitude of 1.2 °C. The lower thermal boundary was held constant at the mean stream temperature of 19.5 °C at a depth of 10 ft beneath stream bottom. Ground-water velocity in the model was varied until the best visual match between the model-simulated daily temperature envelope and the envelope resulting from the field-measured temperature profiles in each of the three piezometers was achieved.

Model-simulated bihourly temperature profiles and the daily envelope for downward velocities of 0.0, 1.0, 1.5, and 2.0 ft/d are shown in figures 29–32, respectively. The velocity that resulted in the best match of model-simulated envelopes to field-determined envelopes was about 1.5 ft/d downward for all three piezometers (table 6).

The stage in the Rockaway River during the period June 2 to June 6, 1986, fluctuated 0.29 ft, but it varied by only 0.03 ft during the 24-h period from hours 47 to 71.

Average differences in head across the top 3.25 ft of sediment at P2 and P3, and the top 3.75 ft at P4 during the 24-h period were 2.17 ft for P2 and P3 and 2.23 ft for P4 (table 6). Average hydraulic gradients were 0.67 ft/ft for P2 and P3 and 0.59 ft/ft for P4 (table 6). The velocity of 1.5 ft/d and the hydraulic gradient at each piezometer were used in Darcy's law to calculate an effective vertical hydraulic conductivity of 2.2 ft/d for the top 3.25 ft of sediment beneath the river at P2 and P3, and 2.5 ft/d for the top 3.75 ft at P4 (table 6).

Heads in all three piezometers were about 1 ft below the streambed. These data are not sufficient to determine if the sediments in the top 1 ft beneath the streambed are fully saturated, close to saturation, or nearly unsaturated. However, hydraulic conductivities of unsaturated sediments are significantly lower than hydraulic conductivities of fully saturated or nearly saturated sediments. The calculated conductivity of 2.2 ft/d of the sediments beneath the stream bottom at this site suggest that the sediments within the top 1 ft beneath the stream bottom are fully or nearly saturated.

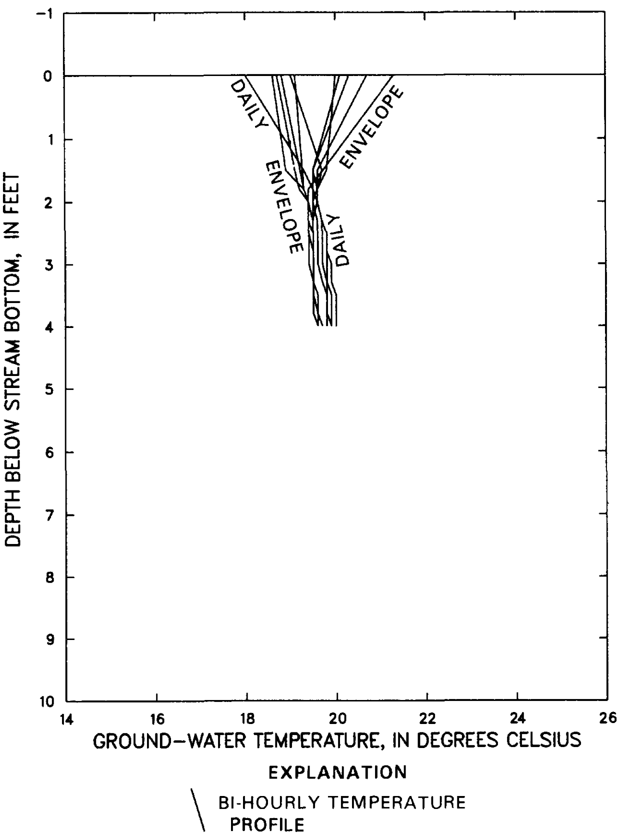


Figure 26. Temperature profiles measured approximately bihourly during day 3 in Dover observation well P2 at the Dover site, Dover, N.J., and the daily temperature envelope described by the temperature profiles.

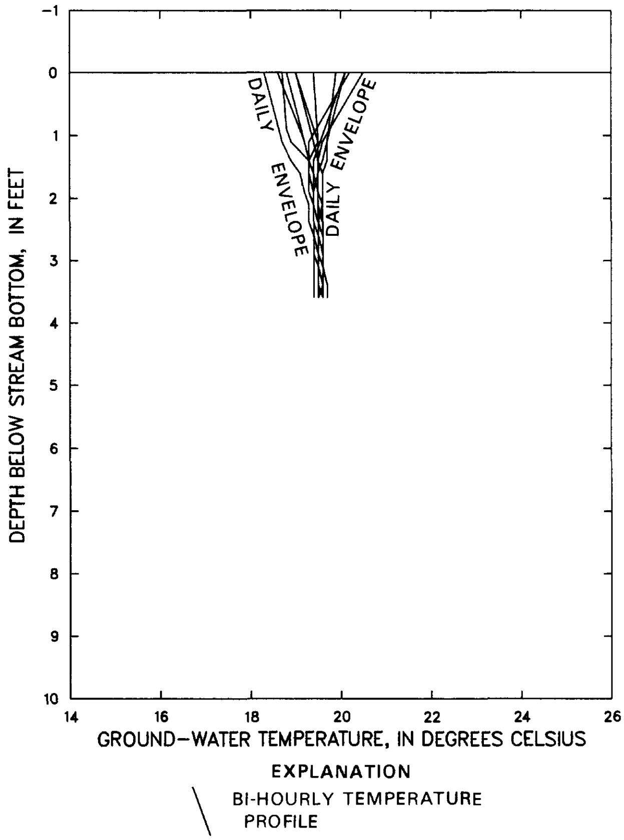


Figure 27. Temperature profiles measured approximately bihourly during day 3 in Dover observation well P3 at the Dover site, Dover, N.J., and the daily temperature envelope described by the temperature profiles.

APPRAISAL OF RESULTS AND SUGGESTIONS FOR FUTURE STUDY

Application of the method at the three field sites demonstrates the feasibility of using the method to distinguish differences in ground-water velocities and in hydraulic conductivities between sites. Independent estimates of velocity at two of the three sites are in general agreement with the velocities estimated by means of temperature envelopes. In addition, estimates of velocities and conductivities determined at the sites generally fall within the ranges of expected rates of flow in, and conductivities of, lacustrine silt and clay as well as outwash sand and gravel. The method can be used to estimate flow rates and effective hydraulic conductivities at sites where ground water discharges to streams or where infiltration occurs from a stream into underlying sediments. The method also might be useful for estimating rates of flow to and from lakes, and the rate of ground-water recharge from precipitation.

Table 6. Vertical velocities and effective vertical hydraulic conductivities of sediments beneath the Rockaway River at the Dover site, Dover, N.J.

[ft, feet; ft/ft, foot per foot; ft/d, feet per day; cm/s, centimeter per second]

Piezometer	Average difference in head between piezometer and river, in ft	Thickness of sediment over which head drop occurs, in ft	Average hydraulic gradient, in ft/ft	Vertical darcian velocity determined from temperature envelopes		Effective vertical hydraulic conductivity, in	
				ft/d	$cm/s \times 10^{-4}$	ft/d	$cm/s \times 10^{-4}$
P2	2.17	3.25	0.67	1.5	5.3	2.2	7.8
P3	2.17	3.25	0.67	1.5	5.3	2.2	7.8
P4	2.23	3.75	0.59	1.5	5.3	2.5	8.8

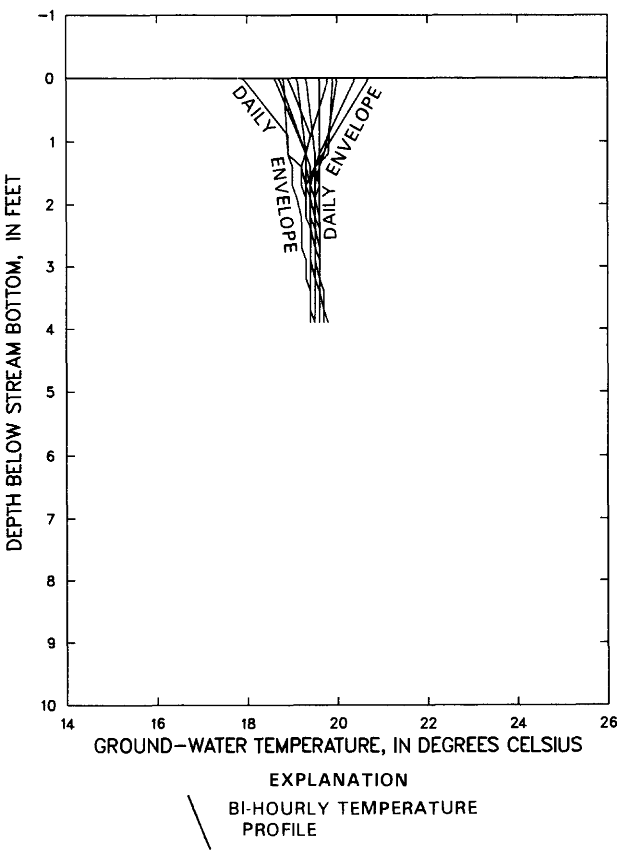


Figure 28. Temperature profiles measured approximately bihourly during day 3 in Dover observation well P4 at the Dover site, Dover, N.J., and the daily temperature envelope described by the temperature profiles.

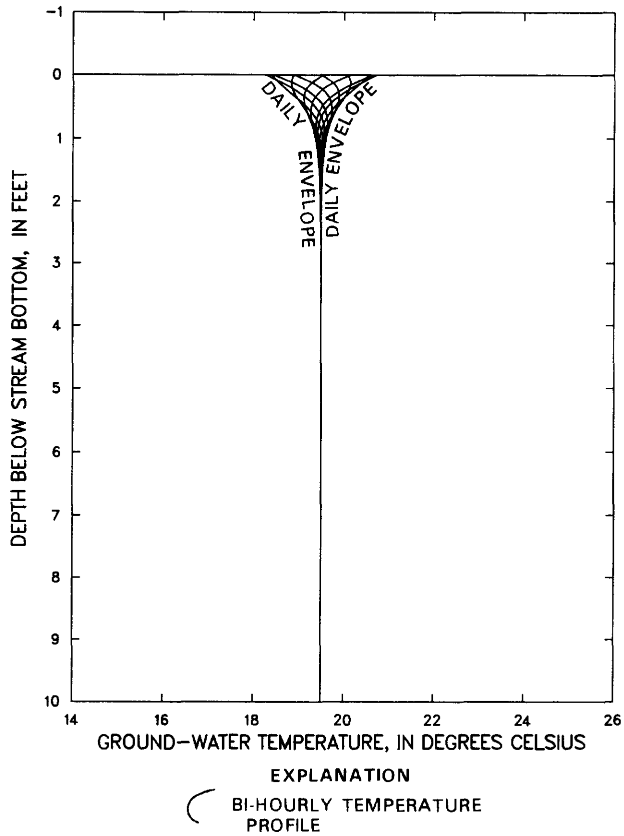


Figure 29. Simulated bihourly temperature profiles, and the daily temperature envelope described by the temperature profiles, at the Dover site, Dover, N.J., for a ground-water velocity of 0.0 foot per day.

Several assumptions made in the model described by equation 5 are probably violated to some extent at the field sites. These assumptions are that (1) ground-water velocity is constant over time, (2) ground-water velocity is constant at all depths beneath the stream, (3) ground-water flow is vertical beneath the stream and, therefore, temperatures beneath the streams are not affected by horizontal flow, and (4) the stream is infinite in width, or, if finite in width, temporal variation in the temperature at the water table near the stream is identical to temporal variation in the temperature of the stream itself.

The rate at which ground water discharges to a stream is known to change over time. This change in rate commonly is observed as a gradual decrease in stream discharge over time during prolonged base-flow conditions. Vertical velocity of the ground water probably increases somewhat in areas of ground-water discharge as the ground water flows upward toward the stream bottom. Vertical velocity probably decreases somewhat in areas of infiltration as stream water flows downward from the stream into the underlying sediments. These changes in vertical velocity occur because flow lines converge toward the stream bottom in areas of ground-water discharge and diverge away from the stream bottom in areas of infiltration. Temporal variation in the temperature at the water table away from the stream may differ significantly from temporal variation in the temperature of the stream. The effects that violations of these assumptions have on the method could be evaluated in a future study.

One limitation of the temperature-profile method described in this report is that it provides only point measurements of flow rates and conductivity. Therefore, several point measurements of velocity or conductivity should be made along a stream reach, and found to be reasonably consistent, before an average rate or conductivity for the reach is viewed with confidence. The method may be useful in determining ground-water-flow rates between aquifers and overlying streams where determination of the flow rate using seepage runs is not feasible because of the small ratio of gain or loss of stream discharge along the reach to total stream discharge.

SUMMARY AND CONCLUSIONS

Stratified-drift deposits of Pleistocene age composed of sand and gravel form the principal aquifers in the

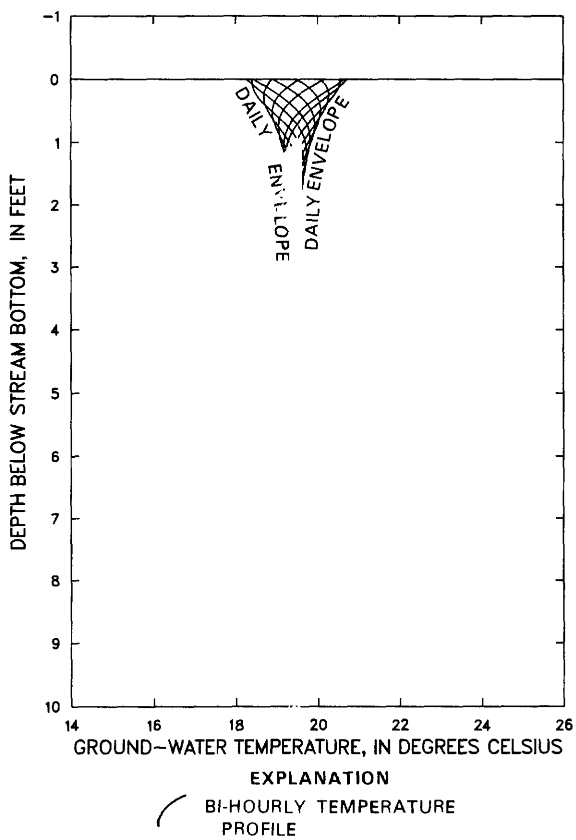


Figure 30. Simulated bihourly temperature profiles, and the daily temperature envelope described by the temperature profiles, at the Dover site, Dover, N.J., for a downward ground-water velocity of 1.0 foot per day.

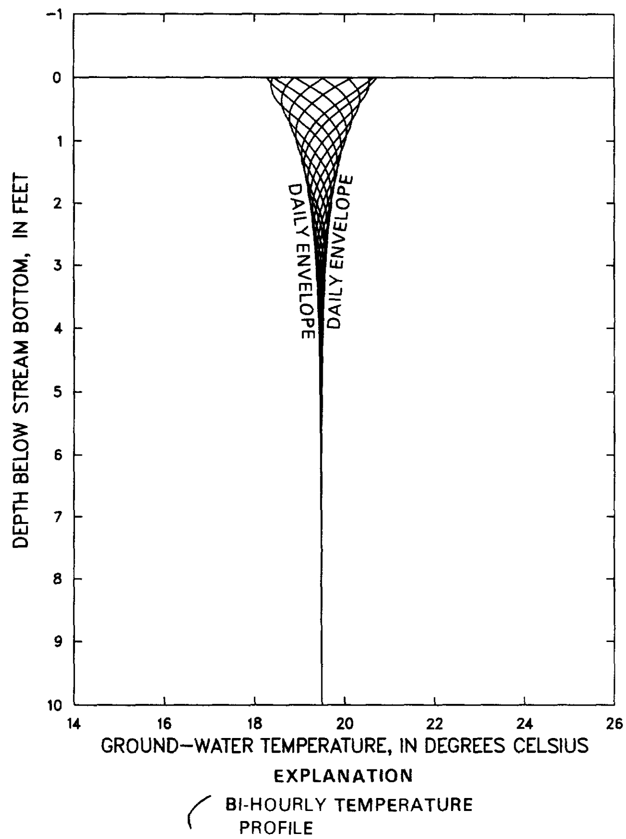


Figure 31. Simulated bihourly temperature profiles, and the daily temperature envelope described by the temperature profiles, at the Dover site, Dover, N.J., for a downward ground-water velocity of 1.5 feet per day.

Northeastern United States. These aquifers, although mostly small, have potential for development of water supplies because they are traversed by streams that serve as sources for induced infiltration. The rates of flow and degree of hydraulic connection between these aquifers and their overlying streams are necessary information for many hydrologic studies. Although several methods are currently used to estimate rates of flow and hydraulic connection, they may be difficult to apply in the field, and the results difficult to interpret. There is a need to develop alternative methodology by which rates of flow and the effective hydraulic conductivity between aquifers and streams can be determined simply and accurately.

Movement of ground water affects the distribution of ground-water temperatures. Numerous studies have applied this principle to indirectly estimate ground-water velocity, areas of ground-water recharge and discharge, and aquifer hydraulic properties. This report presents results of a study to evaluate the use of vertical monthly and bihourly temperature profiles and yearly and daily temperature envelopes in sediments beneath streams to determine rates of vertical ground-water flow and the effective vertical hydraulic conductivity of the sediments.

The partial differential equation that describes the simultaneous flow of fluid and heat in the Earth was solved numerically for transient, one-dimensional, vertical flow of fluid and heat in saturated sediments beneath streams. Temporal variation in stream temperature is the upper thermal-boundary condition, and average stream temperature held constant at depth beneath the stream is the lower thermal-boundary condition.

Sensitivity analysis showed that monthly temperature profiles and the yearly temperature envelope are somewhat sensitive to variation in the thermal properties of saturated sediments. Stream-temperature fluctuations penetrate deeper beneath the stream in coarse-grained sediments than in fine-grained sediments. This deeper heat penetration occurs because coarse-grained sediments most often have a higher thermal diffusivity than fine-grained sediments. In New England, stream temperature varies from about 0 to 25 °C during the year, which causes ground-water temperatures to fluctuate by more than 0.1 °C during a year to a

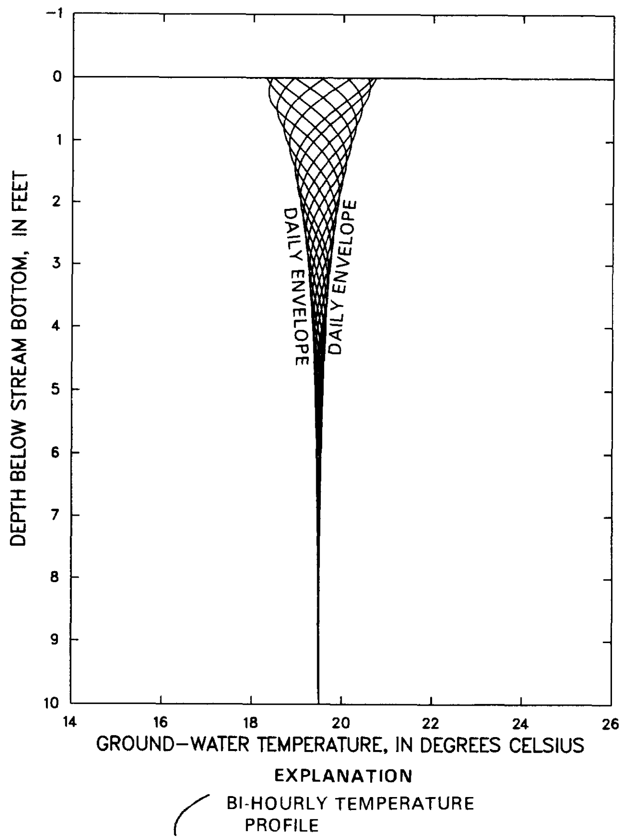


Figure 32. Simulated bihourly temperature profiles, and the daily temperature envelope described by the temperature profiles, at the Dover site, Dover, N.J., for a downward ground-water velocity of 2.0 feet per day.

depth of about 35 ft in fine-grained sediments and to a depth of about 50 ft in coarse-grained sediments, if ground-water velocity is 0 ft/d. Upward flow decreases the depth affected by stream-temperature fluctuation, and downward flow increases this depth.

Monthly temperature profiles and the yearly envelope are most sensitive to variation in ground-water velocities that exceed 0.01 ft/d. The profiles and envelope are insensitive to variation in ground-water velocity of less than about 0.01 ft/d. If stream temperature varies sinusoidally throughout the year, instantaneous temperature profiles that occur one-half year apart are mirror images and the yearly envelope is symmetric around mean stream temperature. In most of New England, stream temperature does not vary sinusoidally over the entire year. Beneath these streams, therefore, symmetrical temperature distributions would not occur.

Temperature profiles and envelopes were used to determine vertical ground-water velocity and effective hydraulic conductivity of sediments beneath streams at three field sites. The Hardwick, Mass., site is located in a reach of the Ware River where ground-water discharge is nearly 0 ft/d. The New Braintree, Mass., site is located in a reach of the Ware River where ground water is discharging from the underlying aquifer to the river. The Dover, N.J., site is located in a reach of the Rockaway River where pumping is inducing infiltration from the stream into the underlying aquifer. At all three sites, instrumentation was installed beneath the stream for determination of temperature profiles, ground-water-level fluctuations, and hydraulic gradients. Concurrent measurements of ground-water temperatures, ground-water level, stream stage, and stream temperature were made at each site during the study. Temperature and hydraulic data collected at these sites were used to determine indirectly rates of vertical ground-water flow and the effective vertical hydraulic conductivity of the sediments underlying these sites.

At the Hardwick site, ground-water flow was upward at a rate of less than 0.01 ft/d. The maximum effective vertical hydraulic conductivity of the lacustrine silts and clays underlying the site was determined to be 0.1 ft/d.

Velocity of upward ground-water flow at the New Braintree site was determined by use of temperature envelopes to range from 0.10 to 0.20 ft/d. The effective vertical hydraulic conductivity of the top 16 ft of sand and gravel underlying the site ranged from 2.4 to 17.1 ft/d. A seepage run made on August 12, 1986, along a 1.6-mi reach of the river containing the site indicated that the average ground-water velocity along the reach was 0.52 ft/d upward. A possible 5-percent error in measurement of discharge, however, results in a possible range in average velocity along the reach from 0.95 ft/d downward to 1.97 ft/d upward.

Velocity of downward ground-water flow at the Dover site was determined to be 1.5 ft/d at three piezometers. The effective vertical hydraulic conductivity of the top 3.25 ft of sediment underlying the river at P2 and P3 was 2.2 ft/d and at P4 was 2.5 ft/d. The infiltration rate at the site determined by tracing stream water as it infiltrated into the aquifer from the stream was estimated to range from 2.0 to 3.3 ft/d.

Application of the method at the three field sites demonstrates the feasibility of using the method to distinguish differences in ground-water velocities and in hydraulic conductivities between sites. Independent estimates of velocity at two of the three sites are in general agreement with the velocities determined using temperature envelopes. In addition, the estimates of velocities and conductivities determined at the sites generally fall within the ranges of expected rates of flow in, and conductivities of, lacustrine silt and clay as well as outwash sand and gravel. The method can be used to estimate flow rates and effective hydraulic conductivities at sites where ground water discharges to a stream or where water from a stream infiltrates the underlying sediments. The method also might be useful for estimating rates of flow to and from lakes and the rate of ground-water recharge from precipitation.

Several assumptions made in the model described by equation 5 are probably violated to some extent at the field sites. The effects that violations of these assumptions have on the method should be evaluated in future studies. These assumptions are that (1) ground-water velocity is constant over time, (2) ground-water velocity is constant at all depths beneath the stream, (3) ground-water flow is vertical beneath the stream and, therefore, temperatures beneath the streams are not affected by horizontal flow, and (4) only the temporal variation in stream temperature affects temperatures beneath the stream.

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METRIC CONVERSION FACTORS

For readers who wish to convert measurements from the inch-pound system of units to the metric system of units, the conversion factors are listed below:

Multiply inch-pound units	By	To obtain metric units
	Length	
inch (in)	2.54	centimeter (cm)
foot (ft)	0.3048	meter (m)
mile (mi)	1.609	kilometer (km)
	Area	
square mile (mi^2)	2.5888	square kilometer (km^2)
	Volume	
gallon (gal)	3.785	liter (L)
	Velocity	
foot per day (ft/d)	3.53×10^{-4}	centimeter per second (cm/s)
cubic foot per second (ft^3/s)	0.02832	cubic meter per second (m^3/s)
	Temperature	
Temperature in degrees Fahrenheit (°F) can be converted to degrees Celsius (°C) as follows: °C=5/9 (°F-32).		
	Density	
pound per cubic foot (lb/ft^3)	0.01602	gram per cubic centimeter (g/cm^3)
	Heat	
thermal conductivity: calorie per second centimeter degree Celsius (cal/s-cm-°C)	243.9	British thermal unit per hour foot degree Fahrenheit (BTU/h-ft-°F)
heat capacity: calorie per cubic centimeter degree Celsius (cal/cm^3-°C)	62.43	British thermal unit per cubic foot degree Fahrenheit (BTU/ft^3-°F)
specific heat: calorie per gram degree Celsius (cal/g-°C)	1.0	British thermal unit per pound degree Fahrenheit (BTU/lb-°F)
thermal diffusivity: square centimeter per second (cm^2/s)	3.875	square foot per hour (ft^2/h)

ALTITUDE DATUM

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