GROUND-WATER RESOURCES AND SIMULATION OF FLOW IN AQUIFERS CONTAINING FRESHWATER AND SEAWATER, ISLAND COUNTY, WASHINGTON

By D. B. Sapik, G. C. Bortleson,
B. W. Drost, M. A. Jones, and E. A. Prych

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[Plates in back cover pocket]

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CONVERSION FACTORS

For the convenience of readers who may prefer to use metric (International System) units rather than the inch-pound units used in this report, values may be converted by using the following factors:

Multiply inch-pound unit	By	To obtain metric unit
inch (in.)	25.4	millimeter (mm)
foot (ft)	0.3048	meter (m)
mile (mi)	1.609	kilometer (km)
acre	4,047	square meter (m^2)
	0.4047	hectare
square foot (ft^2)	0.09290	square meter (m^2)
square mile (mi^2)	2.590	square kilometer (km^2)
gallon (gal)	3.785	liter (L)
	0.003785	cubic meter (m^3)
gallon per minute per foot [(gal/min)/ft]	0.2070	liter per second per meter [(L/s)/m]
cubic foot (ft^3)	0.02832	cubic meter (m^3)
foot per second (ft/s)	0.3048	meter per second (m/s)
cubic foot per second (ft^3/s)	0.02832	cubic meter per second (m^3/s)

Temperature conversion
degree Fahrenheit (°F) to degree Celsius (°C): °C = 5/9 (°F - 32)

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 "Mean Sea Level of 1929."

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ABSTRACT

Increased ground-water use associated with population growth in recent years has caused concern about deterioration of the fresh ground-water supply by intrusion of Puget Sound seawater into aquifers underlying Island County, Washington. Whidbey and Camano Islands, the two largest islands in the county, are the subject of this study and are located about 25 miles northwest of Seattle.

Ground water occurs in glacial deposits having a maximum thickness of about 3,000 feet; these deposits were divided into five aquifers (designated as "A" through "E") and five confining units (designated as "A" through "E") extending to 900 feet below sea level. Fresh ground water moves from recharge areas on land surface, downward through aquifers and confining units, to discharge as springs above and below sea level and to discharge from pumped wells. Most wells pump from aquifers C and D.

A chloride concentration of 100 mg/L (milligrams per liter) was used to distinguish fresh ground water from seawater in the mixing zone. In August 1981, chloride concentrations exceeded 100 mg/L in 24 percent of the sampled wells that were drilled below sea level. The largest chloride concentration was 14,000 mg/L, but chloride concentrations did not exceed 1,000 mg/L in most of the sampled wells. Chloride concentrations were usually large in wells drilled in low-altitude areas along the coastline, and chlorides generally increased with depth below sea level.

Four models were constructed to simulate steady flow of fresh ground water in multiple aquifers and confining units, and to locate a sharp interface between moving freshwater and stationary seawater. Three of the models were constructed for Whidbey Island and one for Camano Island. Each model was calibrated using time-averaged data. Model computations show that most of the recharge is discharged from aquifers C and D as springs below sea level, and only a small fraction of the recharge moves downward below aquifer C. Also, model computations of interface depths and hydraulic head indicate that aquifers in all areas except northeast Camano Island are not recharged by

ground water moving from the mainland through aquifers beneath Puget Sound. Beneath northeast Camano Island, gradients of observed hydraulic heads indicate that aquifers may be recharged by ground water from the mainland. The computed interface was deepest beneath the central parts of the islands and sloped upward near the coastlines to intersect the bottom of Puget Sound. The maximum computed depth of the interface exceeded 900 feet below sea level, beneath southern Whidbey Island. In most of the areas separating Whidbey and Camano Islands the interface intersects the bottom of Puget Sound, indicating that there is no movement of fresh ground water between the two islands.

INTRODUCTION

Island County, in western Washington State, has had a rapid increase in population in recent years, especially since 1975 (Cline and others, 1982). Most of the population increase has occurred in rural areas where ground water is the source of water supply. A large number of wells, located near the coast, have larger chloride concentrations than do wells located farther inland, because of seawater intrusion from Puget Sound. There is concern that additional ground-water development accompanying a continued population growth will cause extensive seawater intrusion, with a corresponding reduction in the amount of fresh ground water. To plan for future growth in the county, it is important to determine the quantity and quality of available ground water and to develop methods for evaluating the consequences of additional ground-water development.

Purpose and Scope

In December 1979, the U.S. Geological Survey, in cooperation with Island County Board of Commissioners and the State of Washington Department of Ecology, began a study of Whidbey and Camano Islands. The purpose of this study was to (1) define the geohydrologic setting of the islands, (2) describe the chemical quality of ground water and delineate areas where degradation of water quality is occurring or is likely to occur, (3) identify areas that have existing seawater-intrusion problems and areas where ground-water overdrafts occur, (4) learn more about the ground-water-flow system by using a model to simulate ground-water flow, and (5) evaluate the effects of increased ground-water withdrawals on water levels and chloride concentrations (associated with seawater intrusion) by using a ground-water model.

Parts of the first three objectives were completed and presented in earlier reports (Cline and others, 1982; Jones, 1985). This report presents the results of the five objectives outlined above. To complete the fourth objective, a ground-water model was constructed and calibrated to simulate three-dimensional steady flow of fresh ground water in a multiple-aquifer system containing freshwater and seawater separated by a sharp interface. To complete the fifth objective, a high-yield well was added to the calibrated model to determine the effects of additional pumping on the position of the interface.

Field data were collected to aid in describing the ground-water-flow system and to calibrate the ground-water-flow model. Data collected by Cline and others (1982) included water-level measurements, chloride concentrations, and pumpage records. After Cline's study, additional data were collected to define time-varying changes for water levels, chloride concentrations, and pumpage. Also, 10 test holes were drilled below the depths of existing wells to provide data on vertical variations in geology, water quality, and hydraulic head (or water levels). Geophysical logs (natural gamma) were obtained for 92 existing wells and eight of the test holes. Drillers' logs provided data on both geology and specific capacity (used to determine aquifer transmissivity) for existing wells.

Description of the Study Area

Island County consists of two major islands, Whidbey and Camano, and several smaller islands situated in Puget Sound, in Washington State. The smaller islands in the county are not included in this study. Whidbey and Camano Islands lie close to the mainland of Washington and are about 25 miles northwest of Seattle (fig. 1). Whidbey, the larger of the two islands, is about 40 miles long, up to 10 miles wide, and has an area of approximately 165 mi^2. Camano Island is about 15 miles long, up to 7 miles wide, and has an area of approximately 45 mi^2. Both islands are long and narrow and contain many indentations in their shorelines; no point on either island is more than 2½ miles from shore.

Island County has a temperate marine climate characterized by warm, dry summers and cool, wet winters. Marine waters and breezes from Puget Sound, which surrounds the islands, provide mild temperatures throughout most of the year. Mean annual temperature is about 50 °F. The coolest month of the year is January, with an average temperature of 38 °F. The warmest months are July and August, with average temperatures of 61 °F. Mean annual precipitation on the islands ranges from 18 inches in the northwestern part of the county to 35 inches in the southeastern part. Rainstorms are not intense, and the rainfall rate is usually less than 0.5 inch per day.

The topography of Whidbey and Camano Islands is characterized by rolling uplands 100 to 300 feet above sea level with steep bluffs along the coast. In a few areas, such as the northeastern and the southeastern parts of Whidbey Island, land-surface altitudes exceed 500 feet above sea level.

The largest percentage of land on the islands, especially the inland areas, is covered by forest. Most of the remaining land is used for urban and agricultural purposes. A small percentage of the land consists of range land, barren land, wetland, and lakes.

Water is used in the county primarily for domestic purposes; only minor amounts are used for agriculture and industry. Ground water is the primary source of water in the county. Surface water is not a source of water except where residents have captured water from a few perennial streams that receive spring discharge from the ground-water system. Most streams are less than 2 miles long, have drainage-basin areas less than 5 mi^2, have small discharges, and only flow intermittently, in direct response to precipitation. Some water is imported from the Skagit River (located on the mainland) to serve residents at Whidbey Island Naval Air Station and Oak Harbor (fig. 1).

About 84 percent of the pumped ground water is distributed to county residents through three municipal and 229 community water systems (as inventoried in 1979 and 1983). The municipal water systems serve residents and small businesses in Oak Harbor, Coupeville, and Langley (fig. 1), and each community water system serves four or more households or serves small businesses in the rural areas. In 1983, the community water systems served 40,694 of the 47,000 county residents. The remaining residents were served by individual wells and springs. Ground water used for irrigation is pumped from 19 wells and accounts for the 16 percent of the total pumped ground water.

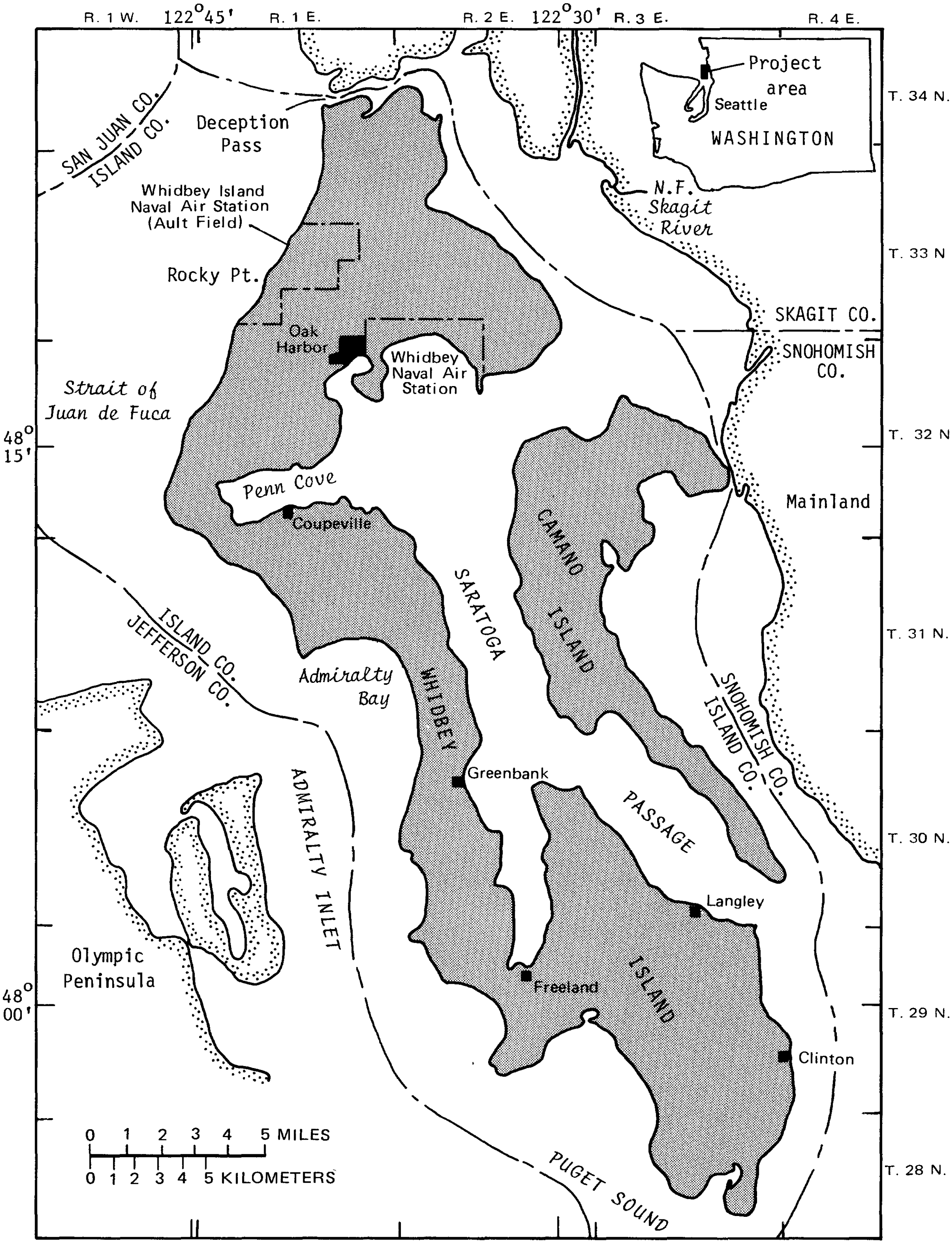


Figure 1.--Location and principal geographic features of Island County (modified from Anderson, 1968).

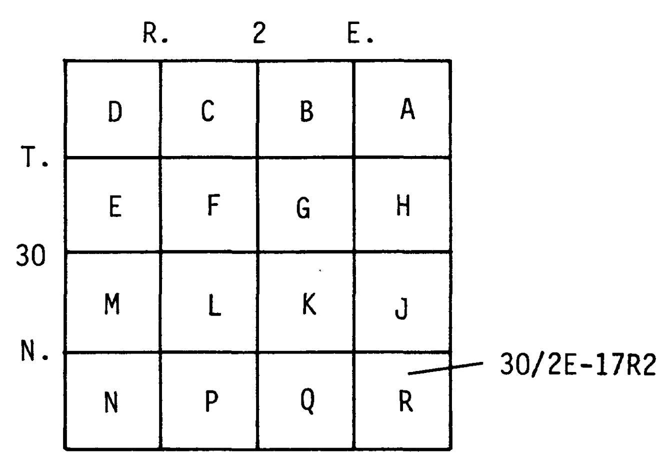
The population of the county increased from 19,638 in 1960 to 47,000 in 1983 (Washington State Office of Program Planning and Fiscal Management, 1971; Washington State Office of Financial Management, 1984). The population growth during 1960-75 (Washington State Office of Program Planning and Fiscal Management, 1971 and 1976) occurred mainly in Oak Harbor, Coupeville, and Langley; after 1975, growth was mainly in the rural areas.

Previous Investigations

Previous work in the study area that applies to the current study includes an interpretation of the Pleistocene stratigraphy of Island County by Easterbrook (1968), a description of the ground-water resources by Anderson (1968), and a description of the chemical quality of ground water by Van Denburgh (1968). Reports by Walters (1971) and Dion and Sumioka (1984) discusses the occurrence of seawater intrusion in coastal areas of Washington, and summarized the concentrations of chloride in water collected from wells drilled within 1 mile of the coast. A report prepared by Cline and others (1982) describes the geohydrology of the ground-water-flow system and presents maps showing the extent of seawater intrusion. Jones (1985) presents maps and cross sections showing the extent and thickness of aquifers and confining units, and maps showing the extent of seawater intrusion.

Numbering System for Wells

Wells in Washington are assigned numbers that identify their location in a township, range, and section. Well number 30/2E-17R2 indicates, successively, the township (T.30 N.) and range (R.2 E.) north and east of the Willamette base line and meridian; the letter indicating north for township is omitted because all wells in Island County are north of the base line and, for range, the letter "W" is used for wells located west of the Willamette Meridian. The first number following the hyphen indicates the section (17) within the township, and the letter following the section gives the 40-acre subdivision of the section, as shown below. The number following the letter is a sequential number assigned to wells within the 40-acre subdivision.



Section 30

Acknowledgments

We gratefully acknowledge the cooperation of the many well drillers and well owners who provided information and allowed access to wells. B and W Well Drilling, Hayes Well Drilling, Kounkel Well Drilling, and Whidbey Well Drillers provided access to newly drilled wells for geophysical logging. Ten deep test holes were drilled by the U.S. Geological Survey on land owned by Martin Boon, Ryan Fakkema, Fred Frei, Dan Garrison, Clifford Hagstrom, James Meyer, Ernest Youderian, the estate of Dale and Irma Scholz, and Island County. Well records and water rights information were made available through the State of Washington Department of Ecology. Additional well records and well locations were provided by county resident Harry Wilbert. The State of Washington Department of Social and Health Services provided laboratory analyses of ground-water samples collected in April and August of 1980. Precipitation records were provided by the following county residents: Ted Bradley, Alberta Finley, Marshall Hartley, William Hatch, Galen Haven, Elmer Newell, Mabel Nielsen, Warren Taylor, and Hope Walker. William Hatch, Galen Haven, and Hope Walker also provided air temperature records. Additional precipitation and temperature records for Whidbey Island Naval Air Station were provided by the U.S. Navy.

GROUND-WATER RESOURCES

Geohydrology

Geologic Setting

Island County lies within the Puget Sound lowland, a topographic and structural depression between the Cascade Range on the east and the Olympic Mountains on the west. As noted by Easterbrook (1968), Whidbey and Camano Islands are generally composed of unconsolidated Pleistocene glacial and interglacial deposits that overlie Tertiary and older bedrock.

Bedrock is exposed on Whidbey Island at Deception Pass and 5 miles farther south at Rocky Point (see fig. 2); the altitude of bedrock is shown in figure 2. Bedrock materials consist of fine-grained sedimentary, metasedimentary, and igneous rocks. Because bedrock materials transmit water at much slower rates than the unconsolidated deposits, bedrock was not considered an important hydrogeologic unit and was assumed to be impermeable throughout the study area.

Surficial deposits of the islands consist of glacial till, clay, sand, and gravel. Older glacial and interglacial deposits occur at depth and are exposed locally in coastal bluffs.

Glacial sediments of Island County were deposited by repeated advances and retreats of continental glaciers which originated in Canada. The most recent of the glaciers, about 14,000 years ago, extended southward to a point about 15 miles south of Olympia and covered Island County with about 4,000 feet of ice.

In aggregate, the unconsolidated deposits of Island County range in thickness from a few hundred feet to about 3,000 feet. This large variation in thickness may be due to faulting of the underlying bedrock, erosion of surficial deposits, or to a large amount of erosional relief on the bedrock.

The large differences observed in physical characteristics of the glacial deposits are due mainly to differences in mode of deposition. Advancing glaciers typically deposited a compact mixture of clay, silt, sand, gravel, and boulders beneath them; this mixture is generally referred to as till. Glaciers advancing into the Puget Sound lowland from the north dammed several large north-flowing streams, forming numerous lakes. Sediment-laden streams flowing into these lakes deposited thick layers of clays and silts. These till and lake deposits constitute the confining units between the aquifers in the county.

Meltwater streams that issued from retreating glaciers typically deposited coarse-grained sands and gravels; these materials are usually referred to as outwash. Some of the outwash deposits of Island County may also have originated from alpine glaciers in mountains that border the Puget Sound lowland. These outwash deposits constitute the aquifers in the county.

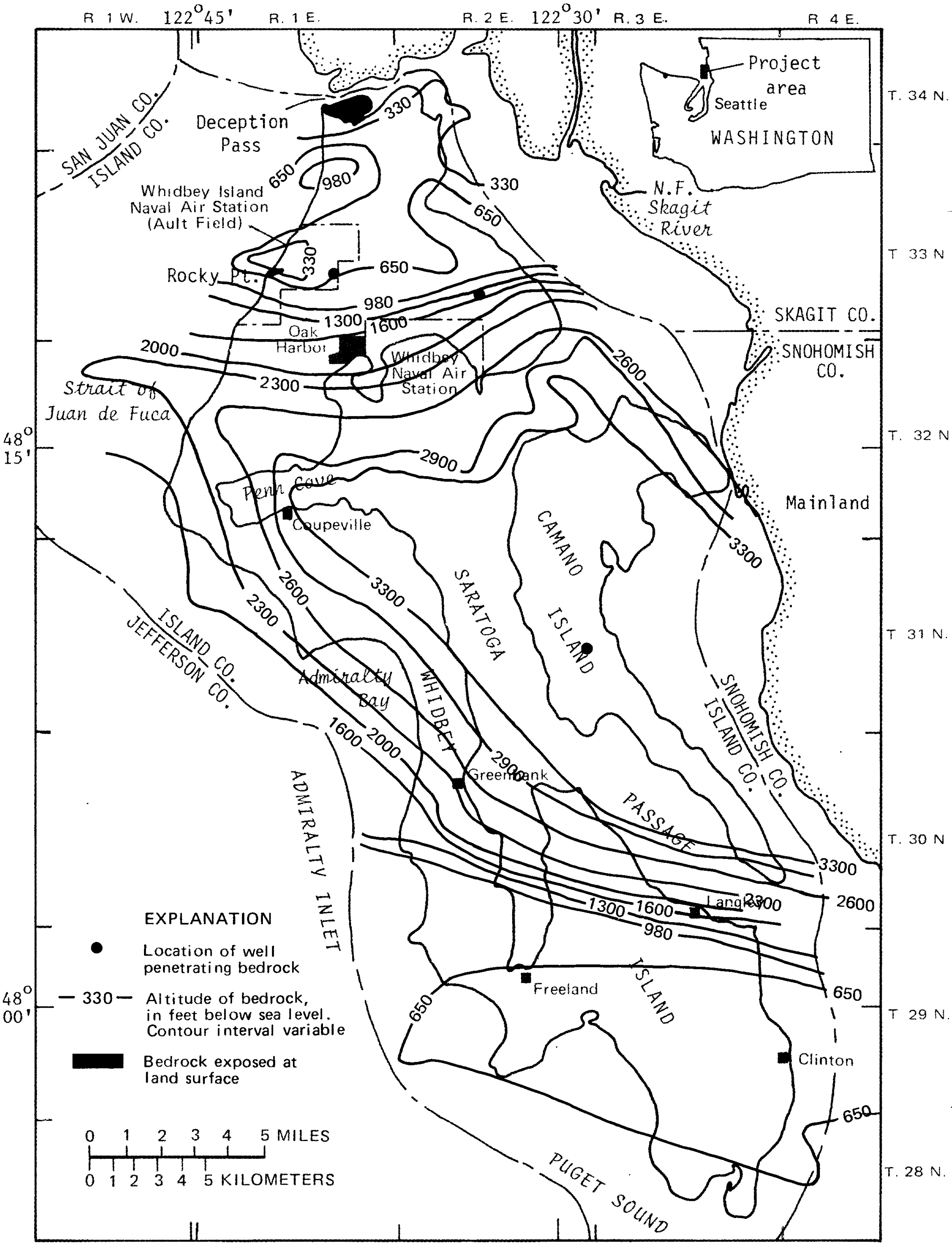


Figure 2.--Altitude of top of bedrock (modified from Fred Pessl, U.S. Geological Survey, written commun., 1982).

Definition of the Aquifer System

For purposes of modeling, the unconsolidated deposits were divided into five aquifers of sand and (or) gravel, and four interstratified confining units (fig. 3). Aquifers and confining units are designated, from oldest to youngest, by the letters "A" through "E" on the hydrogeologic section in figure 3, to conform with the nomenclature used by Jones (1985) for the same area. As used in this report, confining units are layers of fine-grained materials with hydraulic conductivities that are at least an order of magnitude lower than conductivities of aquifer materials. Data pertaining to materials beneath the oldest aquifer (A) are sparse; therefore, all materials below that aquifer were assumed to be a fifth confining unit. As discussed later in this report (see "Sensitivity Analysis for the Calibrated Model"), this lower confining unit was replaced by an aquifer to determine how differences in geology affect model results.

The delineation of aquifers and confining units was based on data from about 350 drillers' logs, 92 geophysical logs (natural gamma), and 10 deep test holes (fig. 4) drilled as part of this study. Some of the layers, especially the younger ones, do not extend laterally over the entire study area because they are truncated by land surface, the sea floor, or bedrock. Where an aquifer or confining unit was missing from a well log, but the unit was found in logs for surrounding wells, the unit was assumed to be laterally continuous. This assumption of continuous layers was used to reduce the differences between drillers' logs. At locations where an aquifer was missing, a hydraulic conductivity was assigned for modeling purposes that was lower than that of surrounding aquifer materials, but higher than the underlying or overlying confining units. Offshore from the islands and below sea level, individual layers were extended to either the sea floor or a model boundary (whichever came first). Probable correlations between the identified layers and geologic units described by Easterbrook (1968) are shown in figure 5.

The altitudes of the tops of the five aquifers identified are shown in plate 1. The thicknesses of aquifers B through E were determined as the difference between top and bottom altitudes of those layers, and are shown in plate 2. A summary of available geohydrologic data for aquifers A through E is in table 1.

Coarse-grained materials were found below aquifer A in some of the deep test holes, but the available data were not sufficient to map these units. Therefore, for the ground-water model constructed in this study, aquifer A was assigned a thickness of 50 feet (the average thickness of aquifer B), and the confining unit below aquifer A was extended to bedrock or to 900 feet below sea level, whichever was shallower. The 900-foot level was chosen because materials below this level were shown to be totally intruded with seawater during trial model runs that were made before calibration of the model described in "Simulation of Ground-Water Flow."

Aquifer E, the top layer in the model, is both confined and unconfined in the county. For modeling purposes, aquifer E is assumed to be confined throughout the county.

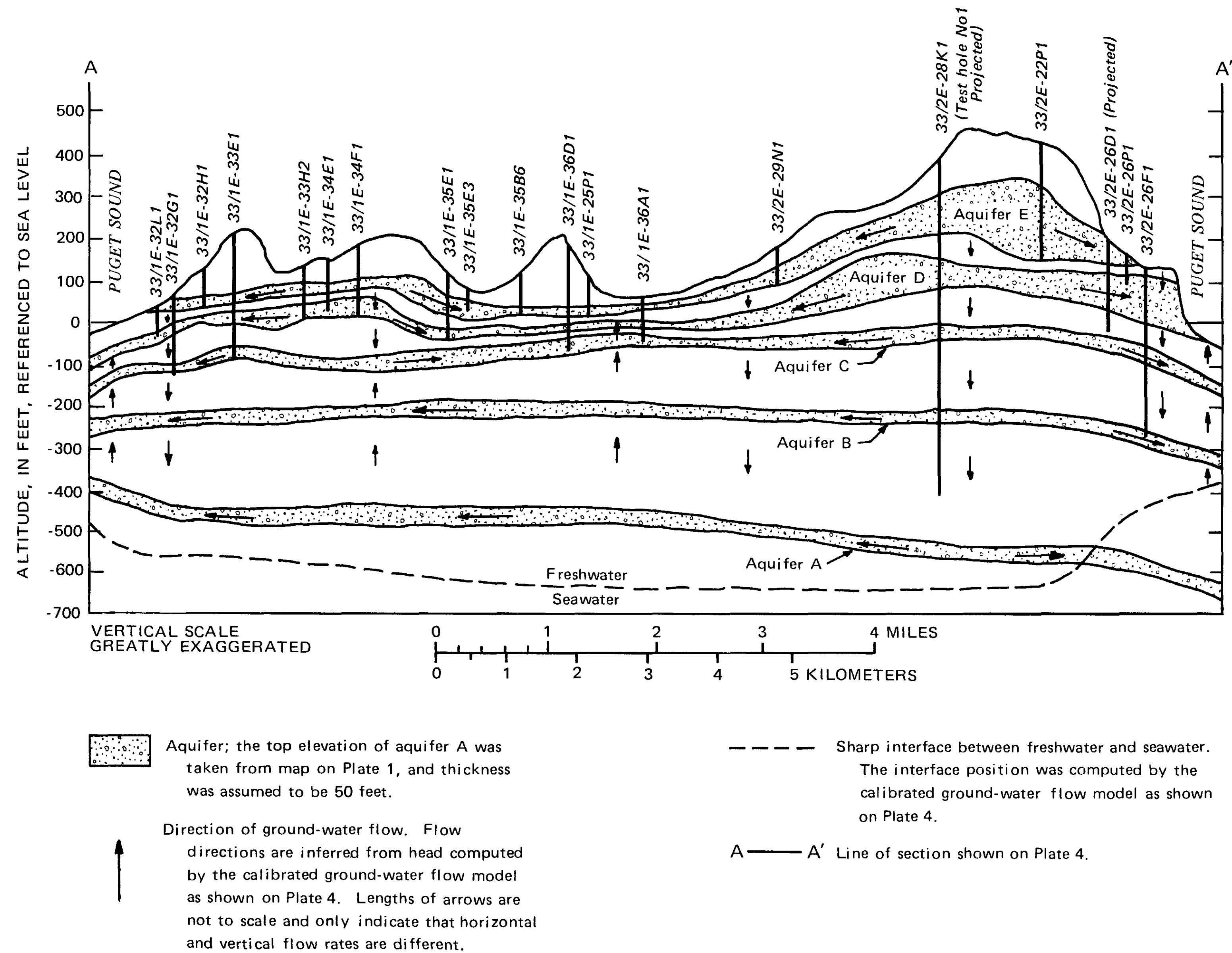


Figure 3.--Schematic hydrogeologic section for the ground-water system in northern Whidbey Island, Island County, Washington.

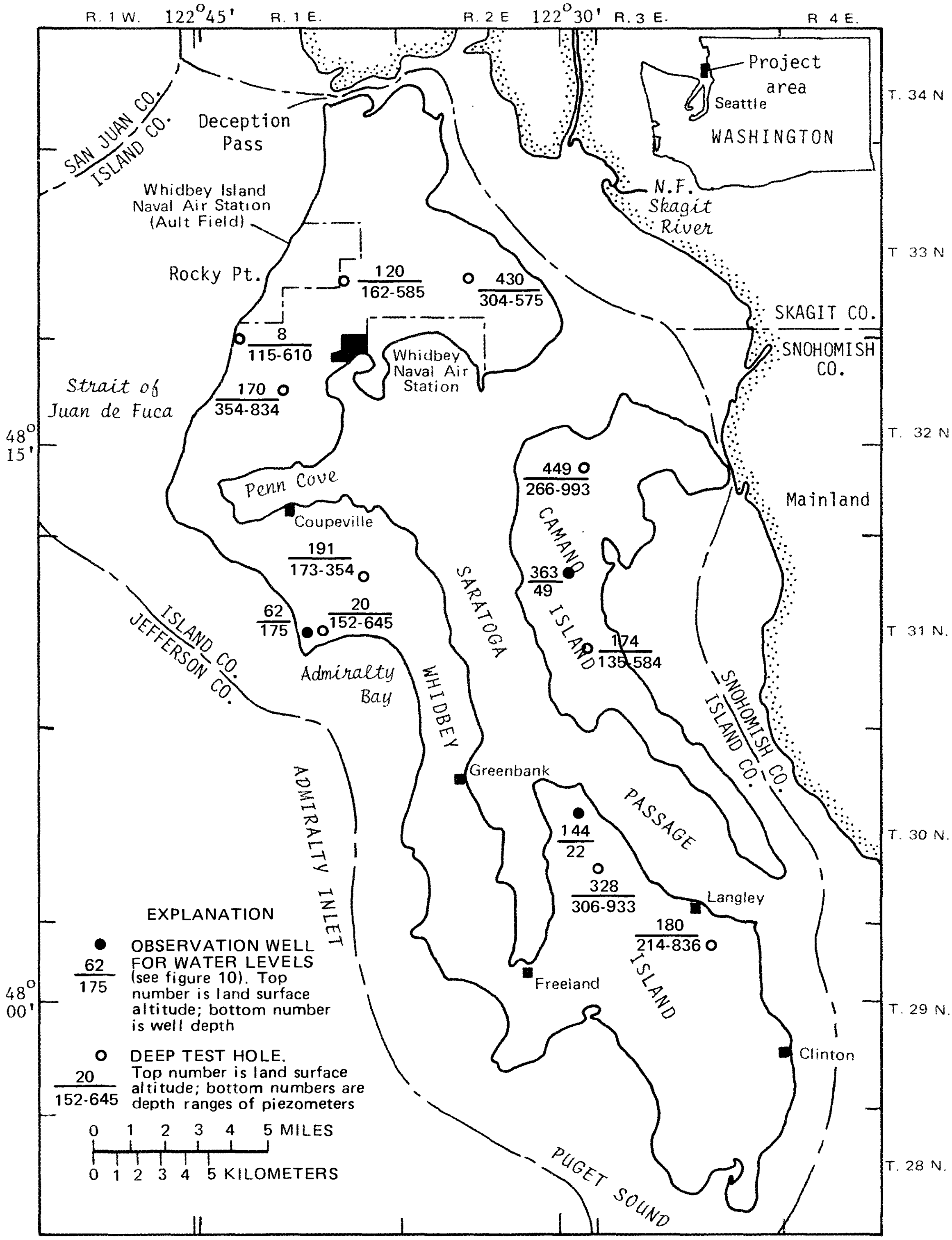


Figure 4.--Locations of deep test holes drilled in 1983 and 1984, and locations of selected observation wells.

<table>
<tr><th>Erathem</th><th>System</th><th>Series</th><th colspan="2">GEOLOGIC CLIMATE UNITS</th><th colspan="2">STRATIGRAPHIC UNITS</th><th>AQUIFERS AND CONFINING UNITS</th></tr>
<tr><td rowspan="8">CENOZOIC</td><td rowspan="8">QUATERNARY</td><td rowspan="8">PLEISTOCENE</td><td rowspan="4">Fraser
Glaciation</td><td rowspan="2">Everson Interstade</td><td colspan="2">Glaciomarine Drift of Everson Age</td><td></td></tr>
<tr><td colspan="2">Partridge Gravel of Easterbrook (1968)</td><td>Aquifer E</td></tr>
<tr><td rowspan="2">Vashon Stade</td><td rowspan="2">VASHON DRIFT</td><td>Till and Associated Drift of Vashon Age</td><td>Confining Unit E</td></tr>
<tr><td>Esperance Sand Member</td><td rowspan="2">Aquifer D</td></tr>
<tr><td colspan="2">Olympia Interglaciation</td><td colspan="2">Quadra Formation (Canadian usage)</td></tr>
<tr><td colspan="2">Possession Glaciation</td><td colspan="2">Possession Drift</td><td>Confining Unit D</td></tr>
<tr><td colspan="2">Whidbey Interglaciation</td><td colspan="2">Whidbey Formation</td><td>Aquifer C
Confining Unit C</td></tr>
<tr><td colspan="2">Double Bluff Glaciation</td><td colspan="2">Double Bluff Drift</td><td>?</td></tr>
</table>

Figure 5.--Probable correlation between aquifers, confining units, and geologic units (modified from Easterbrook, 1968).

Table 1.--Summary of geohydrologic data available for aquifers A through E

Aquifer	Predominant composition	Maximum thickness (feet)	Altitude of top of aquifer[1] (feet)	Altitude of water level[1] (feet)	Specific capacity (gallons per minute per foot)	Number of wells with specific capacity data	Remarks
E	Sand + gravel	160	20 to 490	24 to 421	10.5 to 135	3	Locally discontinuous; occurs in upland.
D	Sand	220	-80 to 335	-51 to 330	0.04 to 27	37	Very productive; widely used.
C	Sand	440	-150 to 270	-140 to 235	0.03 to 42	134	Do.
B	Sand	180	-345 to 20	-80 to 73	0.30 to 25	20	Locally contains saline water.
A	Sand + gravel	50	-600 to -300	-34	0.75 to 22	2	Do.

[1] Altitudes are referenced to sea level.

Hydraulic Characteristics of the Aquifer System

A knowledge of the hydraulic conductivities of aquifers and confining units is necessary to quantify ground-water movement, and to estimate the effects of stresses on the ground-water-flow system. Hydraulic-conductivity values are used to compute transmissivities of the aquifers and confining units, vertical leakage coefficients between units, and spring discharge coefficients.

Horizontal hydraulic conductivity of an aquifer is the rate at which water is transmitted horizontally through a unit cross-sectional area under a unit hydraulic gradient. Estimates of horizontal hydraulic conductivity for aquifer materials were computed from transmissivity values that were based on pump-test data reported by well drillers. The equation used in the computation is:

$$K_x = \frac{T_w}{M_w} , \qquad (1)$$

where

K_x = horizontal hydraulic conductivity, in feet per second;

M_w = length of the open interval in a well, in feet; and

T_w = transmissivity of the open interval, in feet squared per second.

The value for transmissivity (T_w) was estimated using the modified Theis nonequilibrium formula as presented by Ferris and others (1962, p. 99). In solving for T_w, a value of 1.5×10^{-3} was used for the coefficient of storage that was obtained from a formula derived by Jacob (Ferris and others, 1962, p. 88). The coefficient of storage for aquifers was estimated using the following values: (1) 50 feet for thickness, (2) 0.35 for porosity, (3) 7.0×10^{-5} in.^{2}/lb for vertical compressibility of solid materials, (4) 62.4 lb/ft^3 for specific weight of water at 50 °F, and (5) 3.3×10^{-6} in.^{2}/lb for compressibility of water at 50 °F.

Estimated values of horizontal hydraulic conductivity for aquifers and confining units are summarized in table 2. Conductivities for the aquifers are based on 196 well tests reported on drillers' logs. Median aquifer values are in the range of values reported by Walton (1970) for permeable glacial deposits in Illinois. Conductivity for the confining units was estimated to be 1.0×10^{-7} ft/s, the median value given by Norris (1963) for glacial till underlying Illinois and Ohio.

Table 2.--Estimates of horizontal hydraulic conductivity for aquifers and confining units underlying Whidbey and Camano Islands

[Values for aquifers estimated from well tests; value for confining units is median given by Norris (1963) for glacial till]

Aquifer	Confin-ing unit	Hydraulic conductivity, in feet per second: Estimate[1]	Hydraulic conductivity, in feet per second: Range	No. well tests
E		5.7×10^{-3}	5.0×10^{-3} to 4.5×10^{-2}	3
	E	1.0×10^{-7}	--	0
D		9.4×10^{-4}	5.2×10^{-5} to 8.1×10^{-2}	37
	D	1.0×10^{-7}	--	0
C		7.9×10^{-4}	1.7×10^{-5} to 2.5×10^{-2}	134
	C	1.0×10^{-7}	--	0
B		4.9×10^{-4}	3.7×10^{-5} to 5.8×10^{-3}	20
	B	1.0×10^{-7}	--	0
A		4.8×10^{-4}	9.6×10^{-5} to 2.4×10^{-3}	2
	A	1.0×10^{-7}	--	0

[1] Median values are used for all aquifers except aquifer A where the geometric mean was used.

Vertical hydraulic conductivities for the materials in Island County are unknown, but were estimated to be 0.01 times the horizontal hydraulic conductivities using data from well tests in glacial deposits underlying Wisconsin and Illinois (Weeks, 1964; Walton, 1970).

Recharge to the Aquifer System

Recharge to the ground-water system comes mostly from precipitation that falls on the islands, infiltrates into the ground, and percolates downward to the water table. In addition, some of the pumped well water and water imported into the area from the Skagit River percolates below ground surface after being used, and recharges the ground-water system. Most recharge occurs during the winter and spring, when most of the precipitation occurs. Recharge varies throughout the county and is a function of precipitation, temperature, land use, soil type, and vegetation type.

Average annual recharge from precipitation was computed for a 20-year period using a daily soil-moisture accounting method (eq. 2).

$$\text{Recharge} = \text{Precipitation} - \text{Increase In Soil Moisture} - \text{Actual Evapotranspiration} \qquad (2)$$

When precipitation exceeds actual evapotranspiration, and when field capacity is exceeded, the excess water becomes ground-water recharge. Surface-water runoff was assumed to be zero during this study, because surface runoff represents only a small percentage of the total precipitation and is considered insignificant in the overall water budget of the islands.

Daily precipitation in equation 2 was computed as follows:

$$P_d = P_{cd} \frac{P_a}{P_{ca}} , \qquad (3)$$

where

P_d = daily precipitation at any location, in inches;

P_{cd} = daily precipitation at the Coupeville weather station, in inches;

P_a = average annual precipitation at any location, in inches; and

P_{ca} = average annual precipitation at Coupeville, in inches.

Average annual precipitation, P_a in equation 3, was estimated by using a regression analysis on data for 10 weather stations (Coupeville and Whidbey Island Naval Air Station weather stations, and eight backyard gages maintained by county residents, fig. 6). The period of record for these sites ranged from 4 to 42 years. Regression analysis related precipitation to north-south location, east-west location, and land-surface altitude. The only statistically significant independent variable in the relation was east-west location; the correlation coefficient for precipitation and east-west location was 0.95. The resulting equation was found to be:

$$P_a = 23.2 + 0.919E , \qquad (4)$$

where

P_a = average annual precipitation at any location, in inches, and

E = distance, in miles, east of the north-south line between Ranges 1 and 2 East (E is negative for locations west of this line).

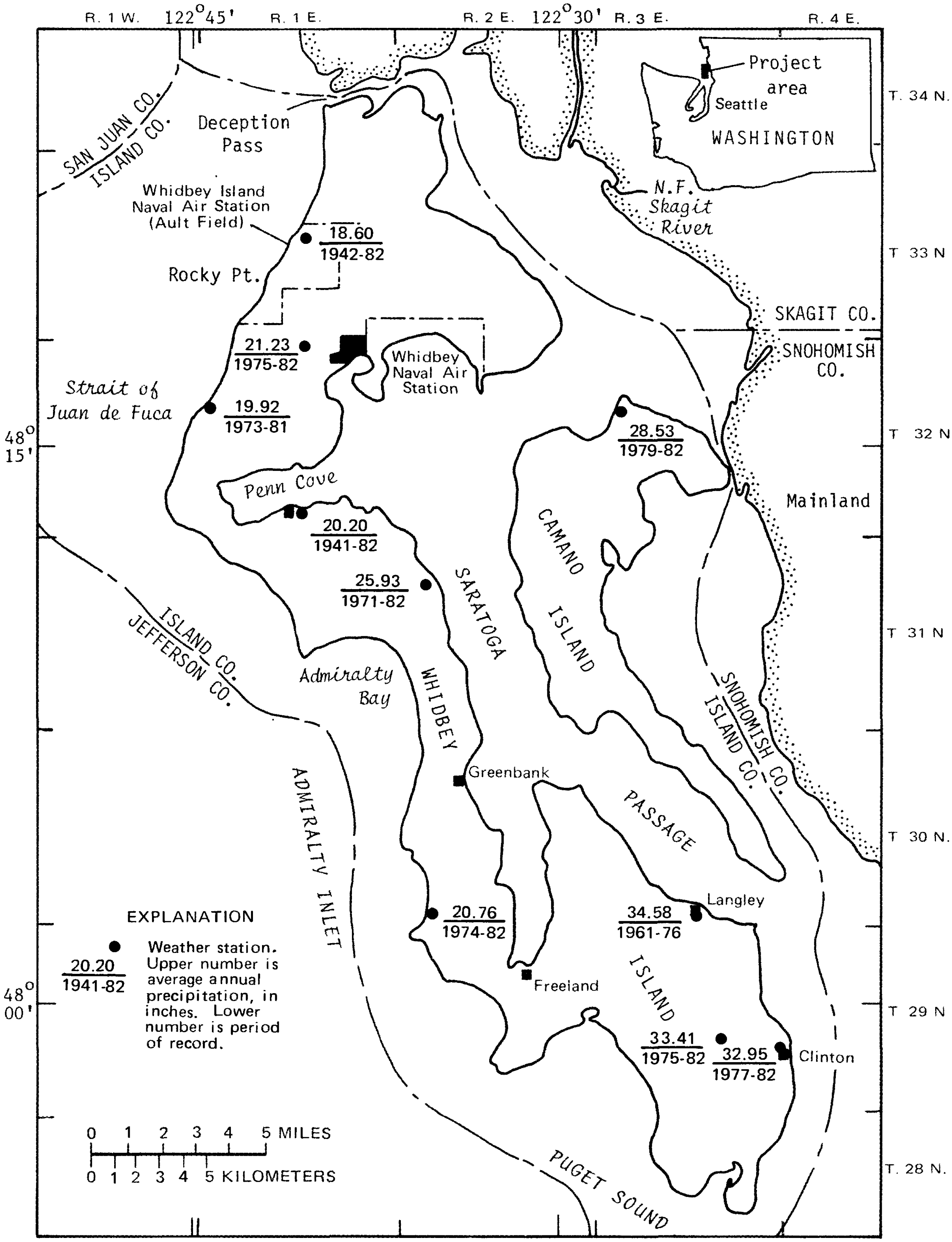


Figure 6.--Average annual precipitation at 10 weather stations.

Figure 7 shows a plot of average annual precipitation, P_a, for the 10 weather stations, as a function of E (above).

Field capacity (inches of water) for any particular soil was computed by multiplying moisture-holding capacity (inches of water per foot of soil) by root-zone depth (feet). Moisture-holding capacity was assumed to be 1 inch of water per foot of soil (Ogrosky and Mockus, 1964). Root-zone depths were assumed to be a function of land use (table 3).

Potential evapotranspiration, the amount of evapotranspiration that would occur if an unlimited supply of water was available, was calculated using a modified Blaney-Criddle formula (U.S. Department of Agriculture, 1970) and the daily average air temperature at the Coupeville weather station. This formula was further modified by using crop growth-stage curves (U.S. Department of Agriculture, 1970). These curves allow potential evapotranspiration to be seasonally adjusted to account for the type and growth stage of the vegetation. The crop-growth curve for forest land was not available and was obtained by taking a weighted average of the coniferous forest curve (for 75 percent of forest land) and the deciduous orchard curve (for 25 percent of

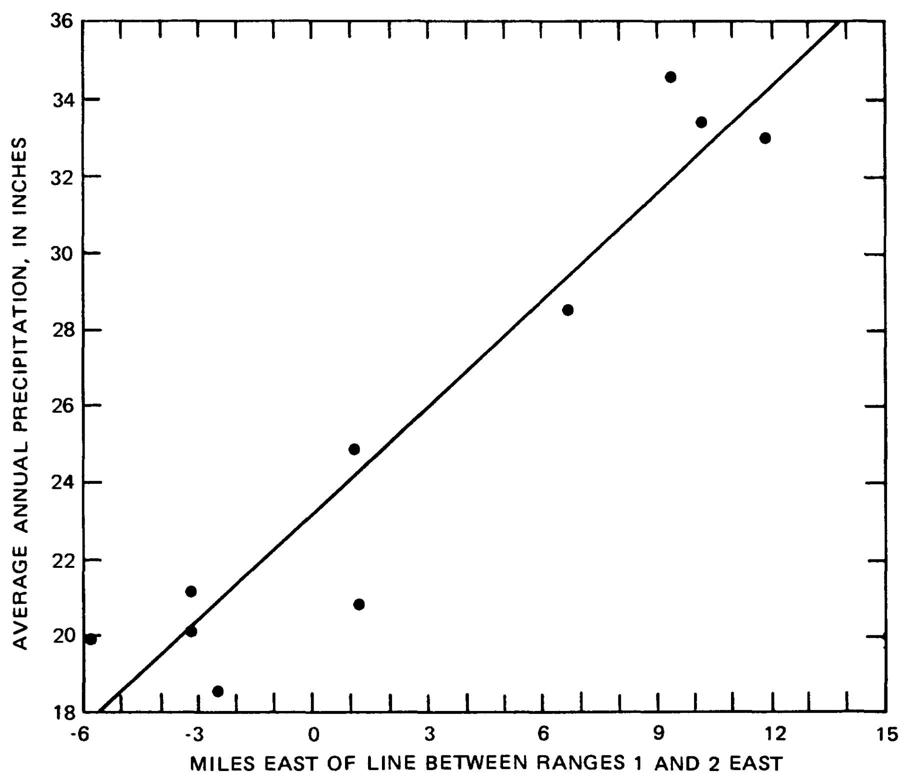


Figure 7.--Average annual precipitation as a function of east-west location in Island County.

forest land). The coniferous forest curve was obtained by multiplying the pasture grass curve by 0.64 (C. R. Cole, Pacific Northwest Laboratory, written commun., 1980). Actual evapotranspiration was assumed to be equal to potential evapotranspiration times the ratio of actual soil moisture to the moisture content at field capacity.

Using the above equations and assumptions, recharge was calculated for midpoints of selected ranges of precipitation and for different land-use types (table 3). The maximum calculated recharge is 25.54 in./yr (inches per year) for suburban, range, or barren land receiving an annual precipitation of 35 to 40 inches. The minimum calculated nonzero recharge is 3.85 in./yr for forest lands receiving an annual precipitation of 15 to 20 inches.

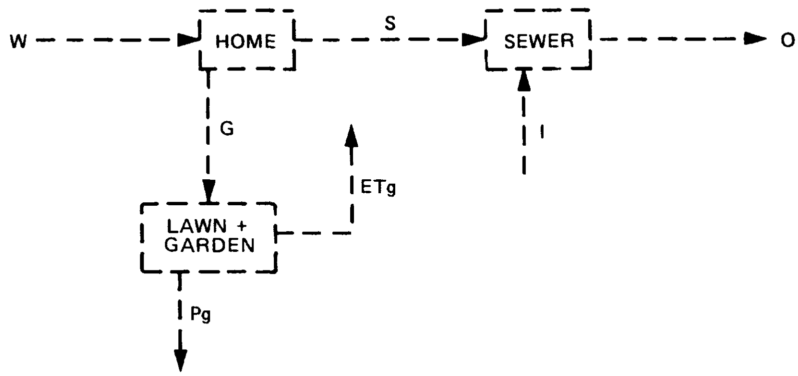
The ground-water system is also recharged by percolation of water from septic-tank drainfields and from irrigation (farm and domestic). Most of this water originates as pumpage from the ground-water system (see "Discharge from the Aquifer System"). Ground water pumped from wells, together with water imported from the Skagit River (609 million gallons in 1981, or 2.6 ft^3/s), is disposed of in three major ways: (1) as wastewater to sewers, (2) as wastewater to septic-tank drainfields, and (3) as irrigation (both farm and domestic). Water discharged to sewers is eventually discharged to Puget Sound and is therefore completely lost from the ground-water system. In sewered areas, the rate of percolation from domestic irrigation is assumed to be approximately equal to the rate of ground-water infiltration to sewers, therefore net recharge is nearly zero (fig. 8). In non-sewered areas, some of the water discharged to septic drainfields or used for domestic irrigation percolates downward and becomes ground-water recharge. Recharge in non-sewered areas is estimated to be 70 percent of the ground water pumped (fig. 9).

Total recharge from precipitation and pumped water was estimated to be 186 ft^3/s: 144 ft^3/s for Whidbey Island and 42 ft^3/s for Camano Island.

Table 3.--Average annual recharge for selected ranges of precipitation and land-use types, and root-zone depths for land-use types

Range of annual precipitation (inches)	Recharge for a land-use type, in inches per year[1]			
	Freshwater or wetland	Forest	Agricultural	Suburban, range land, or barren
15 to <20	--	3.85	5.87	7.78
20 to <25	--	8.02	9.84	11.94
25 to <30	6.27	12.27	13.99	16.27
30 to <35	11.19	16.67	18.35	20.74
35 to <40	16.33	21.41	23.03	25.54
Root-zone depth (feet)	--	6	3	1

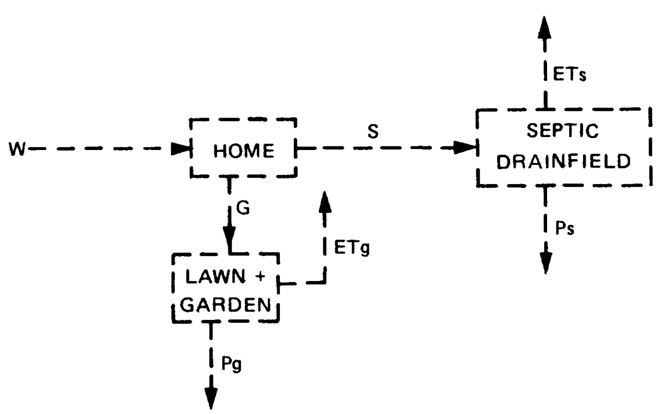
[1]Recharge at midpoint of precipitation range.



W = Water suppied
G = Irrigation of lawn and garden = 1/3 W
ETg = Evapotranspiration from lawn and garden = 2/3 G = 2/9 W
(from Ogrosky and Mockus, 1964)
Pg = Recharge from lawn and garden irrigation = 1/3 G = 1/9 W
S = Water to sewer = 2/3 W (from Metcalf and Eddy, 1979)
I = Ground-water infiltration to sewer is at least 3/20 (or 15 percent) of total flow in sewer (from Metcalf and Eddy, 1979) = 3/20 (S + I) = 3/20 (2/3 W + I), or I = 2/17 W
O = Outflow from sewer to Puget Sound

Net Recharge = Pg - I = 1/9 W - 2/17 W
(The net recharge is so small that it is assumed to be zero.)

Figure 8.--Method of calculating recharge from water used in sewered areas.



W = Water supplied
G = Irrigation of lawn and garden = 1/3 W
ETg = Evapotranspiration from lawn and garden = 2/3 G = 2/9 W
(from Ogrosky and Mockus, 1964)
Pg = Recharge from lawn and garden irrigation = 1/3 G = 1/9 W
S = Water to septic drainfield = 2/3 W (from Metcalf and Eddy, 1979)
ETs = Evapotranspiration from septic drainfield = 1/10 S (assumed) = 2/30 W
Ps = Recharge from septic drainfield = 9/10 S ≐ 18/30 W

Net recharge = Pg + Ps = 1/9 W + 18/30 W = 0.7 W

Figure 9.--Method of calculating recharge from water used in non-sewered areas.

Discharge from the Aquifer System

Ground water is discharged from springs and is pumped from wells. Springs occur both above sea level and along the bottom of Puget Sound. Springs above sea level are visible at hundreds of locations both inland and along the coast of each island. Discharge from these springs occurs as seeps that cannot be measured. Discharge was measured for 38 springs on Whidbey Island and 6 springs on Camano Island, and the discharge of 11 springs reported by Anderson (1968) was added to the discharge measured for Camano Island. Total estimated spring flows above sea level for Whidbey and Camano Islands are 4 and 2 ft^3/s, respectively. Estimates are not available for spring flows below sea level.

About 84 percent of the ground water pumped from wells is used for household purposes and 16 percent is used for irrigation and industrial purposes. The major public water systems are located at Oak Harbor, Coupeville, and Langley (fig. 1). The remaining water systems are scattered over the islands, but most of these systems are within a half mile of the coast. Pumpage was estimated from a field inventory of the 229 public water systems serving four or more households, and pumpage for the smaller water systems was estimated using a 1981 population census and a per capita consumption rate of 100 gallons per day per person. Pumpage for irrigation was estimated for 19 wells. Ground-water pumpage in 1981 was estimated to be 4 ft^3/s for Whidbey Island and 1 ft^3/s for Camano Island.

Ground-Water Levels and Movement

Fresh ground water moves from recharge areas, through the ground-water system, to discharge areas. Ground water moves both horizontally and vertically, from areas of high hydraulic head (the altitude to which water rises in a well) to areas of low head. The conceptual pattern of steady-state ground-water movement, shown in figure 3, is horizontal and toward Puget Sound in aquifers, downward in confining units that underlie high-altitude areas, and upward in confining units that underlie low-altitude areas and in areas offshore from the coastlines. The flow pattern shown in figure 3 is based on the assumption of no flow in bedrock below the Pleistocene deposits and no flow in the seawater zone. The conceptual pattern of movement can be altered, as shown in figure 3, by topography and by pumping wells (not all pumping wells near the line of section are shown in figure 3). There are some discrepancies in horizontal flow directions shown in figure 3, possibly because the line of section is not oriented parallel to the principal directions of horizontal flow in all aquifers. The flow directions in figure 3 are inferred from head computed by a calibrated steady-state ground-water-flow model (see "Calibration of the Steady-State Flow Model"); the model assumes no movement in the seawater zone.

Differences in the directions of ground-water movement in aquifers and confining units can be explained by using Darcy's law (Hubbert, 1940) to compute an apparent velocity as the product of hydraulic conductivity and hydraulic gradient. In aquifers, the primary direction of movement is horizontal, because the Darcy velocity is greater in the horizontal direction than in the vertical; in confining units, the primary direction of movement is vertical because the Darcy velocity is greatest in the vertical direction.

The pattern of horizontal ground-water movement in aquifers could be inferred from maps of hydraulic head published by Cline and others (1982) and by Anderson (1968); however, after the aquifer system was defined (see "Definition of the Aquifer System"), it was found that most of the wells used to construct the published maps were not completed in the aquifers shown on the maps. There were not sufficient data available to construct head maps for each aquifer; therefore, maps of hydraulic head were constructed by using head computed in calibrated ground-water-flow models (see "Simulation of Ground-Water Flow"). These maps are shown on plate 4, along with observed heads for selected wells.

Water-level altitudes measured in wells vary with the completed depths of the wells. Where ground water moves downward (fig. 3), water-level altitudes decrease with depth; conversely, where ground water moves upward, water-level altitudes increase with depth. For the aquifer nearest to land surface, depths to water in wells are generally greater in areas of high land-surface altitudes than in areas of low altitudes.

Water levels fluctuate in response to changes in recharge and pumping. Seasonally, water levels decline in summer, when recharge from precipitation is low and pumping is greatest; conversely, water levels rise in winter when recharge is high and pumping is low (fig. 10). Water levels in many wells drilled along the coastline also change daily in response to tides.

Long-term fluctuations of water levels for selected wells that were measured in 1963 and 1980-83 are shown in figure 10. These trends and fluctuations for wells shown in figure 10 are typical for wells located inland and away from high-rate pumping wells. Near pumping wells, water-level fluctuations are greater than those shown in figure 10, but long-term trends are about the same. Because these water-level data do not indicate any significant long-term trend in ground-water levels, ground-water recharge and discharge are equal, and the ground-water system is assumed to be in steady state for the period 1963-83.

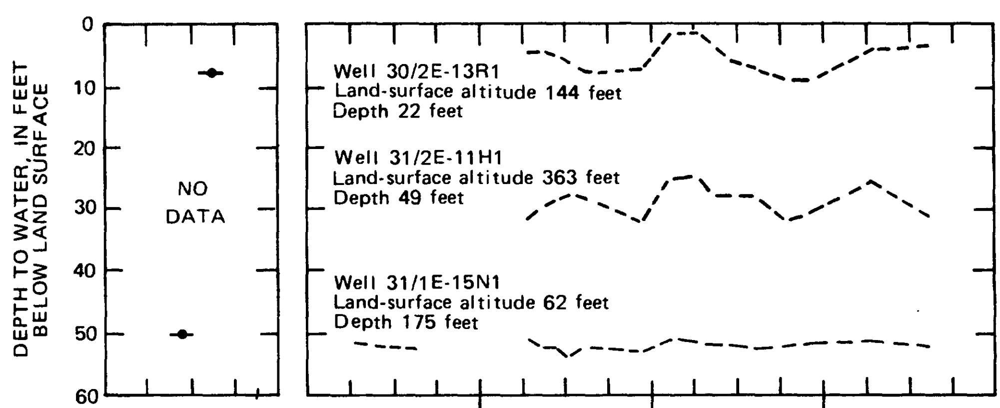


Figure 10.--Water levels in selected wells in Island County.
See figure 4 for locations of wells.

Chemical Quality of Ground Water

Some of the water-quality information that is needed to make effective water-management decisions includes: (1) description of the chemical quality of ground water, (2) delineation of areas where water-quality degradation is occurring or is likely to occur in the future, and (3) definition of geohydrologic conditions that may influence ground-water quality.

General Chemical Characteristics of Ground Water

The chemical composition of naturally occurring ground water is controlled by several factors: (1) the composition of recharge water when it enters the aquifer systems; (2) the minerals that are (or have been) in contact with the water, and their solubilities; and (3) the amount of time the water has been in contact with aquifer materials. These controls are often complex and difficult to evaluate. Some variation in the chemical composition of the water can be explained, however, by determining water chemistry of individual aquifers.

Complete chemical analyses were made for ground-water samples collected during 1981-83 from 38 wells and 10 test holes (fig. 11). These samples were analyzed for major chemical ions dissolved in the water. One sample was collected from each of the 38 wells. Twenty-five samples were obtained during the drilling of 10 deep test holes (see fig. 11). These samples were collected at different depths that were isolated during drilling.

The chemical variations of ground water in Island County are shown graphically in figure 12, on a trilinear plot that displays major ionic constituents in milliequivalents per liter. Plotted values are expressed as percentages of the total milliequivalents per liter of cations and anions. Along the plot axis for sodium plus potassium, sodium is the predominant constituent. Aquifers D and E yielded predominately a calcium magnesium bicarbonate water. Deeper aquifers A and B yielded predominately sodium chloride and sodium bicarbonate waters. Aquifer C contained a mixture of waters found in aquifers B and D. Some of the ground water collected from aquifers intruded by seawater approached the chemical composition of ocean water (fig. 12), and the predominant ions in solution were sodium and chloride. Dissolved-solids concentrations (represented by circle diameters in figure 12) were generally larger in aquifers A and B than in aquifers C, D, and E, and the largest dissolved-solids concentrations were usually associated with waters dominated by sodium and chloride.

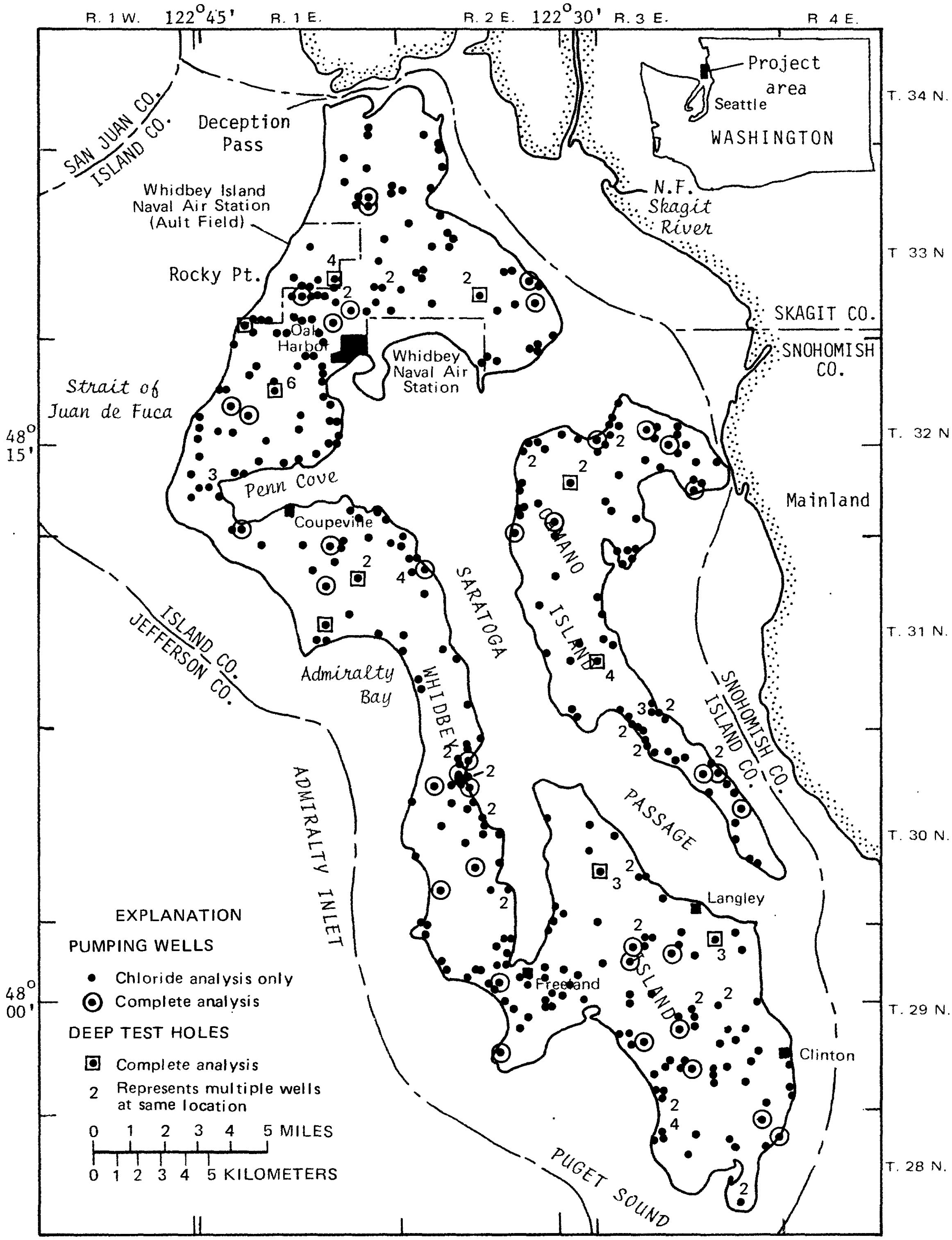


Figure 11.--Location of sampled wells and types of analyses.

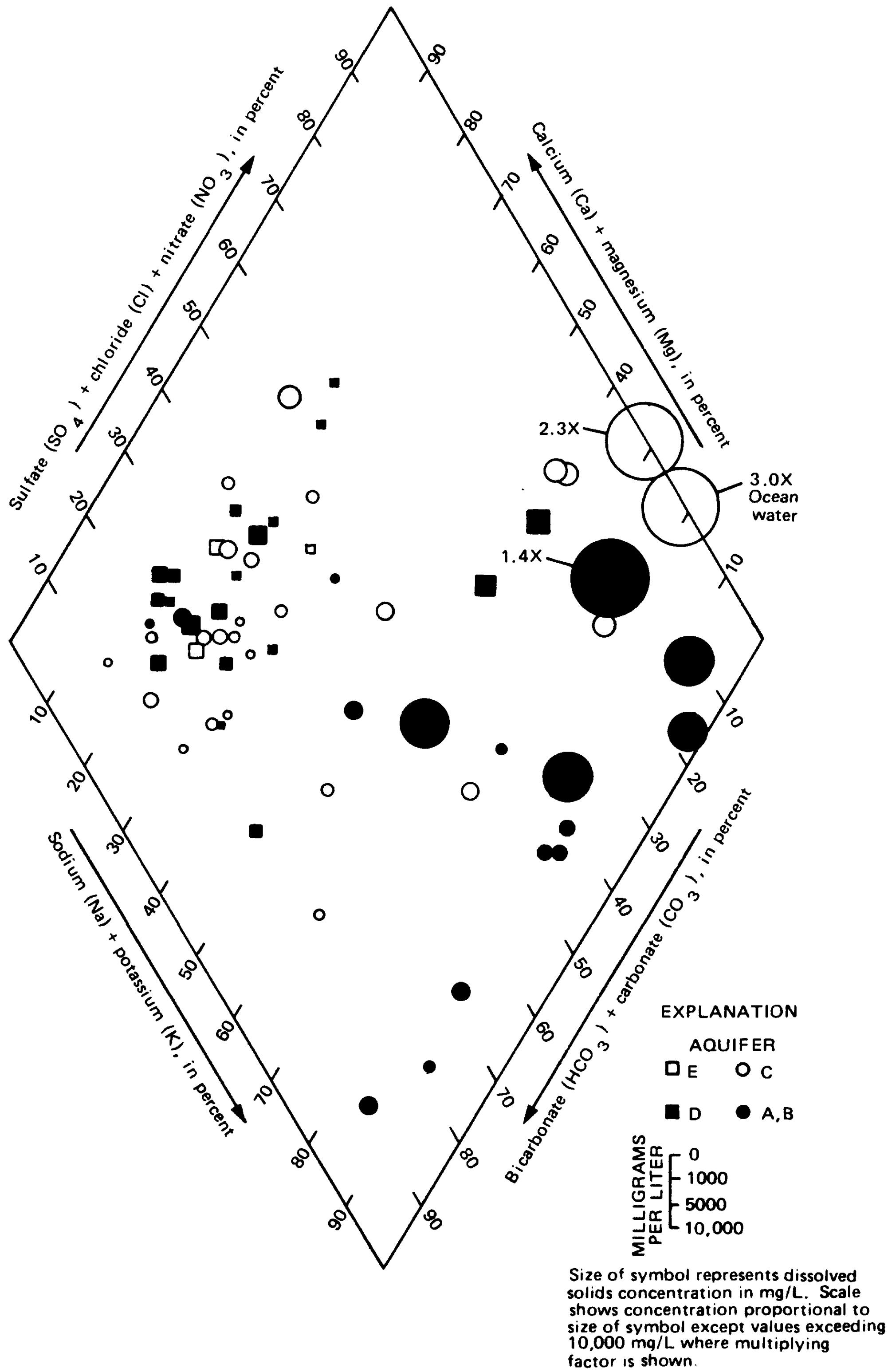


Figure 12.--Percentage composition of ions for Island County ground-water quality sites in aquifers A through E.

Dissolved-Solids Concentrations

Dissolved-solids concentrations in ground waters are an indicator of the degree of mineralization or the amount of dissolved substances in the water. For water samples without complete chemical analyses, dissolved-solids concentrations were estimated from specific-conductance values. In Island County, the dissolved-solids concentration for ground water, in mg/L (milligrams per liter), is approximately 60 percent of the specific conductance, in μS/cm (microsiemens per centimeter). Thus, a specific conductance of 800 μS/cm corresponds to about 500 mg/L dissolved solids. Concentrations above 500 mg/L are considered high in terms of drinking-water standards.

Areal variations in dissolved-solids concentrations are shown in figure 13. To eliminate the possible bias of high dissolved solids resulting from seawater contamination, only well waters containing less than 100 mg/L of chloride are shown. Dissolved-solids concentrations were generally larger in aquifers A and B than in C, D, and E, but the general patterns shown in figure 13 do not vary with depth. About two-thirds of the sampled wells contained moderate (200 to 500 mg/L) levels of dissolved solids. Dissolved-solids concentrations were generally small (less than 200 mg/L) in the southern part of Whidbey Island, and were generally large (greater than 500 mg/L) in the part of Whidbey Island between Oak Harbor and Coupeville. Ground waters had moderate levels of dissolved solids throughout Camano Island and the remainder of Whidbey Island.

Suitability of Ground Water For Drinking Purposes

One of the primary objectives of water-quality management is the protection and enhancement of drinking water. A comparison of the observed chemical concentrations to drinking-water standards is especially pertinent because the principal use of ground water in Island County is for public and domestic supplies. Ancillary uses include irrigation, stock watering, and industrial and commercial supplies.

The public-supply standards adopted by the State of Washington Department of Social and Health Services (1978, p. 29, 35) for constituents analyzed in this study are shown in table 4. Samples were not analyzed for all constituents for which criteria are established, such as trace metals or pesticides, nor were the constituents summarized by aquifer because there were not enough samples. The primary constituents (fluoride and nitrate) in table 4 relate to health-risk considerations, and the secondary constituents (dissolved solids, chloride, iron, manganese, and sulfate) relate to odor, taste, and other esthetic considerations. Median values for primary and secondary constituents, and the percentage and number of sampled wells exceeding water-quality criteria, are also shown in table 4. The median values for dissolved solids and chloride were based on 193 wells sampled in August 1981, and median values for the other constituents were based on complete analyses for the 63 samples described in "General Chemical Characteristics of Ground Water."

The primary criterion for fluoride was not exceeded, and the criterion for nitrate was exceeded only once. The median nitrate concentration was 0.1 mg/L, well below the drinking-water criterion of 10 mg/L.

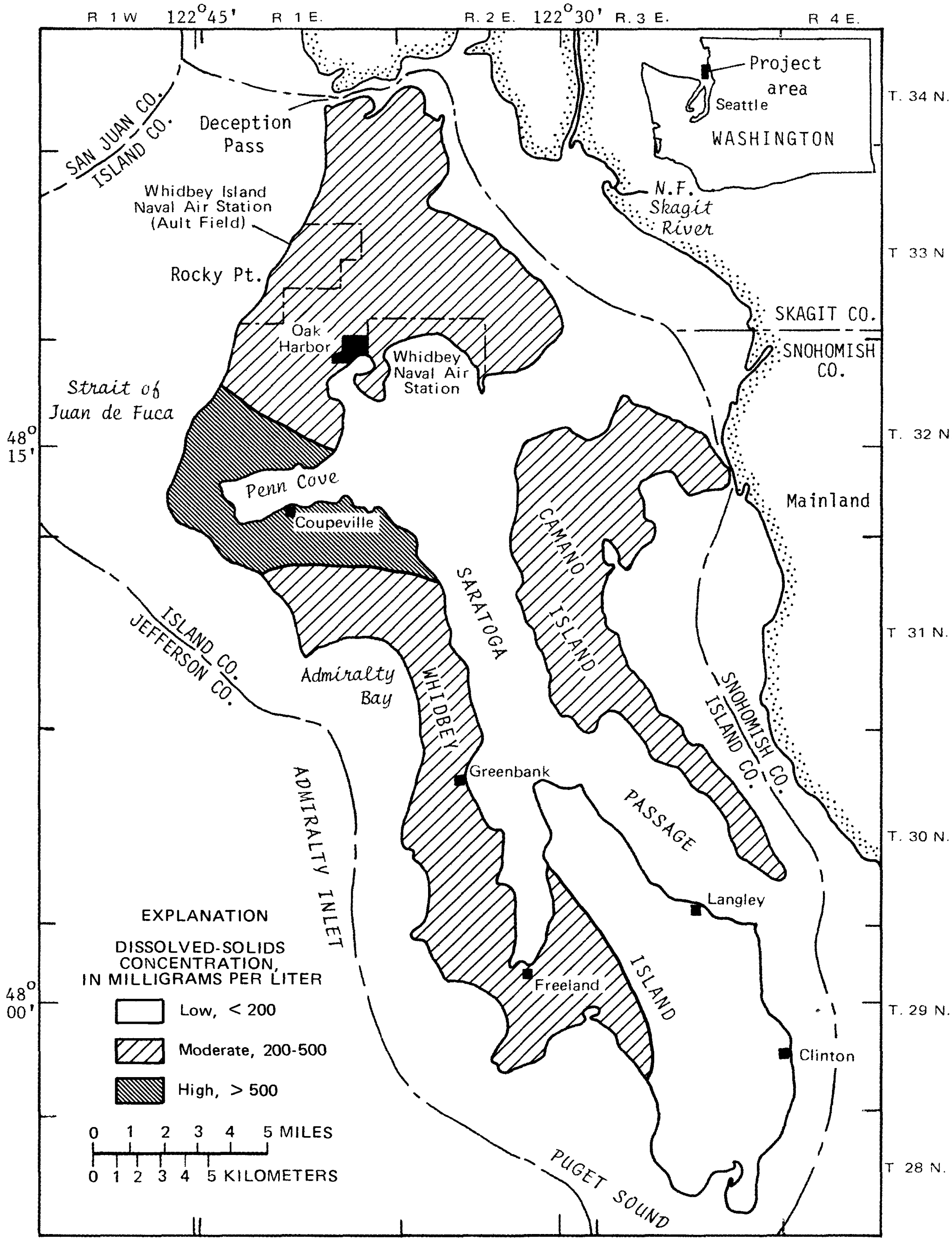


Figure 13.--Areal variation in dissolved-solids concentration. Estimated from specific conductance of waters in August, 1980 to1983.

Table 4.--Median values for selected primary and secondary constituents and number of sampled wells exceeding criteria for drinking water

[All constituents in dissolved phase; constituents expressed as milligrams per liter unless shown otherwise.]

Constituent	Number of wells sampled	Water-quality criteria	Median concen-tration	Wells exceeding criteria	
				Number	Percent
Primary					
Fluoride	62	2.0	0.2	0	0
Nitrate plus nitrite as nitrogen[1]	56	10	.1	1	1.8
Secondary					
Dissolved solids[2] [3]	193	500	250	11	5.7
Chloride	193	250	30	15	7.8
Iron, micrograms per liter	57	300	62	24	42
Manganese, micrograms per liter	57	50	82	34	60
Sulfate	63	250	15	1	1.6

[1] Analytical determination as nitrate plus nitrite; water-quality criteria as nitrate only.
[2] Number of wells used to determine median is from large data set of August 1981.
[3] Concentration estimated from specific-conductance value.

The secondary criteria were exceeded in 6 percent of wells sampled for dissolved solids, 8 percent for chloride, 42 percent for iron, 60 percent for manganese, and 2 percent for sulfate (table 4). The large iron and manganese concentrations are probably derived naturally from the weathering of rocks and minerals. It is common to find high iron and manganese in the ground waters in contact with glacial sediments in the Puget Sound region (Turney, 1986).

Hardness is not a standard applicable to drinking-water supplies, but because of its relation to taste and the formation of deposits in distribution systems, hardness is of general concern. Public acceptability of the degree of hardness of water may vary considerably depending on local conditions, and in some cases even excessively hard water is tolerated. In Island County, waters with high dissolved solids generally have high hardness values. The table below lists the percentage of the 63 samples with complete analyses that occurred within each of four hardness ranges (Durfor and Becker, 1964, p. 27).

Range of hardness as $CaCO_3$ (mg/L)	Degree of hardness	Percentage of 63 samples within each range
0-60	Soft	6
61-120	Moderately hard	34
121-180	Hard	22
Greater than 180	Very hard	38

Seawater Intrusion into Aquifers

Freshwater in the aquifers and confining units below sea level is surrounded by seawater (fig. 3). The interface between freshwater and seawater is rarely sharp. Instead, freshwater is separated from seawater by a mixing zone within which chloride concentrations increase with seaward distance from shore and with depth. The absence of a sharp interface is evidenced by the relatively small chloride concentrations (generally 100 to 1,000 mg/L) in samples obtained from pumping wells near the coast that intercept the mixing zone. These small concentrations are in contrast to large concentrations of 16,000 mg/L for water in Puget Sound. Chloride is a good indicator of seawater intrusion because (1) it is a major constituent in seawater, (2) it is chemically stable (not volatile, not adsorbed, and not precipitated), and (3) it moves at the same rate as the encroaching seawater.

The dimensions and position of the mixing zone, and consequently chloride concentrations in pumped wells, are influenced by hydraulic heads, hydraulic conductivities of aquifers and confining units, and rates of recharge and pumpage. The mixing zone moves both laterally and vertically in response to changes in aquifer head caused by changes in recharge and pumpage, and by tides near the coast. When head is lowered in a pumped aquifer, the mixing zone can move landward a sufficient distance to cause an increase in chloride concentration of the pumped water. An example of chloride fluctuations caused by seawater intrusion into a pumping well was presented by Anderson (1968) and is shown in figure 14. This well, located at Whidbey Island Naval Air Station (fig. 1) and about 1 mile from the shoreline, was pumped continuously at a rate of 0.4 ft^3/s for 10 days in September and October 1964.

The variation in chloride concentrations with depth below sea level is shown in table 5 for samples obtained from 10 deep test holes (fig. 11). Chloride concentrations increased significantly with depth in all test holes except 31/1E-11H1, 32/1E-5C1, and 32/2E-25K1. The highest chloride concentration observed in the test holes was 7,200 mg/L in aquifer B (test hole 29/3E-3J3) beneath southern Whidbey Island.

To detect seawater intrusion in an aquifer, a background level of chloride was determined by sampling wells completed above sea level that are generally not susceptible to intrusion. In a cumulative frequency distribution of chloride concentrations (fig. 15), significant differences are observed in waters pumped from wells completed above and below sea level. For wells completed above sea level the chloride concentrations did not exceed 83 mg/L. Consequently, a concentration of 100 mg/L was used as the background chloride value for this study. The value of 100 mg/L was assumed to represent ground water along the freshwater side of the mixing zone. It was further assumed that wells pumping water with a chloride concentration greater than 100 mg/L were probably intruded by seawater.

Large chloride concentrations do not always indicate seawater intrusion; large concentrations also can result from contamination introduced by disposal of manmade wastes or from the presence of connate water that has been in the aquifer since its deposition. Large chloride concentrations associated with manmade wastes were not detected in this study; however, the sampling program was not oriented toward waste sites. In well waters where large chloride concentrations occur near the coast and where heads are at or below sea level, seawater intrusion appears to be the most probable cause.

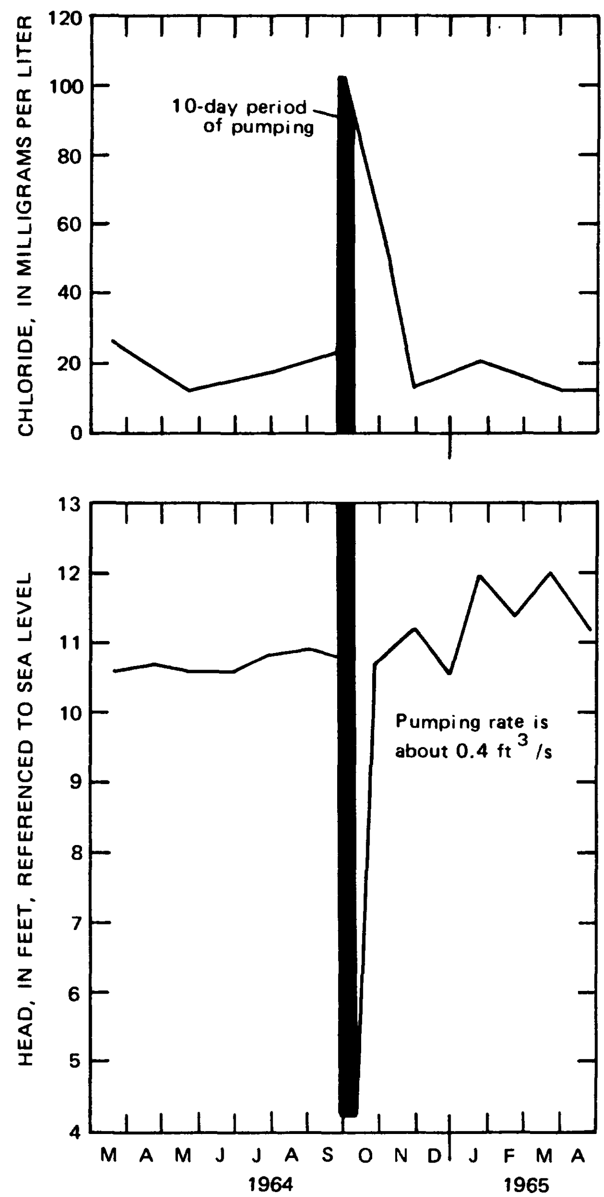


Figure 14.--Fluctuation in chloride concentrations and water levels with pumping of well 33/1E-22C1 (from Anderson, 1968).

Table 5.--Chloride concentrations in water from 10 deep test holes. See figure 11 for locations of test holes

Location	Altitude of land surface[1] (feet)	Altitude of sampling interval[1] (feet)	Aquifer	Chloride concentration, in milligrams per liter
33/2E-28K1	430	150 to 153	D	8.9
		-210 to -370	A,B	1,600
33/1E-26D1	120	-75 to -80	D	56
		-222 to -228	C	66
		-310 to -316	B	2,200
32/1E-9M1	170	-127 to -130	C	67
		-140 to -150	C	30
		-255 to -258	B,C	35
		-263 to -266	B	18
		-575 to -578	A	170
31/1E-11H1	191	-162 to -172	B	1,600
		-322 to -327	A	74
30/3E-30M1	328	-136 to -163	C	13
		-212 to -215	B	130
		-612 to -615	A	2,900
29/3E-3J3	180	142 to 144	D	11
		-204 to -207	B	110
		-325 to -340	B	7,200
32/2E-25K1	449	171 to 177	D	13
		-539 to -545	A	22
31/3E-30D1	174	6 to 0	E	15
		-62 to -68	C	19
		-331 to -377	A,B	230
		-425 to -431	A	210
31/1E-15H1	20	-112 to -117	C	77
		-293 to -298	A	240
32/1E-5C1	8	-104 to -109	D	62
		-182 to -186	C	57
		-304 to -310	C	51
		-382 to -402	B	66

[1] Altitudes are in feet, referenced to sea level.

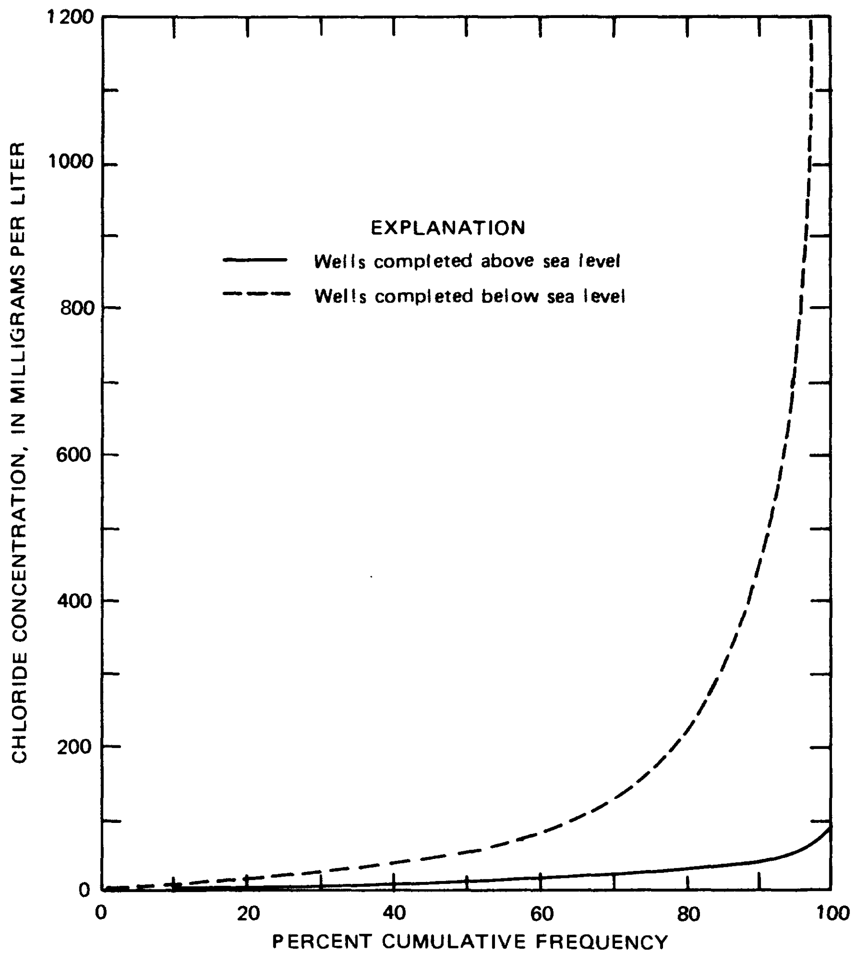


Figure 15.--Cumulative frequency distribution of chloride concentrations in aquifers B through E in the month of August, 1980 to 1983. Maximum chloride concentrations of 12,000 to 14,000 milligrams per liter are not shown.

Areal Variations in Chloride Concentrations

Chloride concentrations were determined for 425 wells sampled over a period of 4 years (1980 to 1983). Most of the samples were collected in April and August of each year. The highest chloride concentrations occurred in August (or September) of each year. Areal variations in chloride concentrations for aquifers B through E, shown in figure 16, are based on 193 samples collected in August 1981. Maximum chloride concentrations, observed in the month of August of years 1980 through 1983, are shown in figure 17. The areal variations shown in these figures are summarized as follows:

Aquifer E. Chloride concentrations were less than 100 mg/L in all wells.

Aquifer D. Chloride concentrations were usually large in wells drilled in low-altitude areas and along the coastline. In August 1981, chloride exceeded 100 mg/L in seven wells located in northern Camano Island and in two wells near the coast of Whidbey Island.

Aquifer C. This aquifer had the largest number of intruded wells; that is, wells with chloride concentrations greater than 100 mg/L. In August 1981, chloride exceeded 100 mg/L in 16 wells on Whidbey Island and 10 wells on southern Camano Island. Chloride concentrations commonly exceeded 250 mg/L in the southern part of Camano Island and the central part of Whidbey Island where these islands are narrow. In five of the wells, chloride concentrations exceeded 750 mg/L. One well, 32/1E-32N1, had chloride concentrations as large as 14,000 mg/L, which is nearly as large as water in Puget Sound.

Aquifer B. Chloride concentrations were generally less than 100 mg/L, but this aquifer has relatively few wells. In August 1981, chloride exceeded 100 mg/L in one well near Coupeville on Whidbey Island.

The overall magnitude and distribution of chloride concentrations is shown by the frequency distribution in figure 18. At least 76 percent of the wells sampled in August 1981 had chloride concentrations below 100 mg/L. Chloride concentrations exceeded the secondary criterion of 250 mg/L (table 4) in 15 wells (6 percent) penetrating aquifer D and 23 wells (12 percent) penetrating aquifer C. Chloride did not exceed 250 mg/L for any samples obtained from aquifers B and E. Of the 193 wells sampled in August 1981, 143 were completed below sea level into aquifers B, C, and D. Thirty-four (24 percent) of the wells completed below sea level had chloride concentrations equal to or greater than 100 mg/L, and 16 wells (11 percent) had chloride concentrations exceeding 250 mg/L.

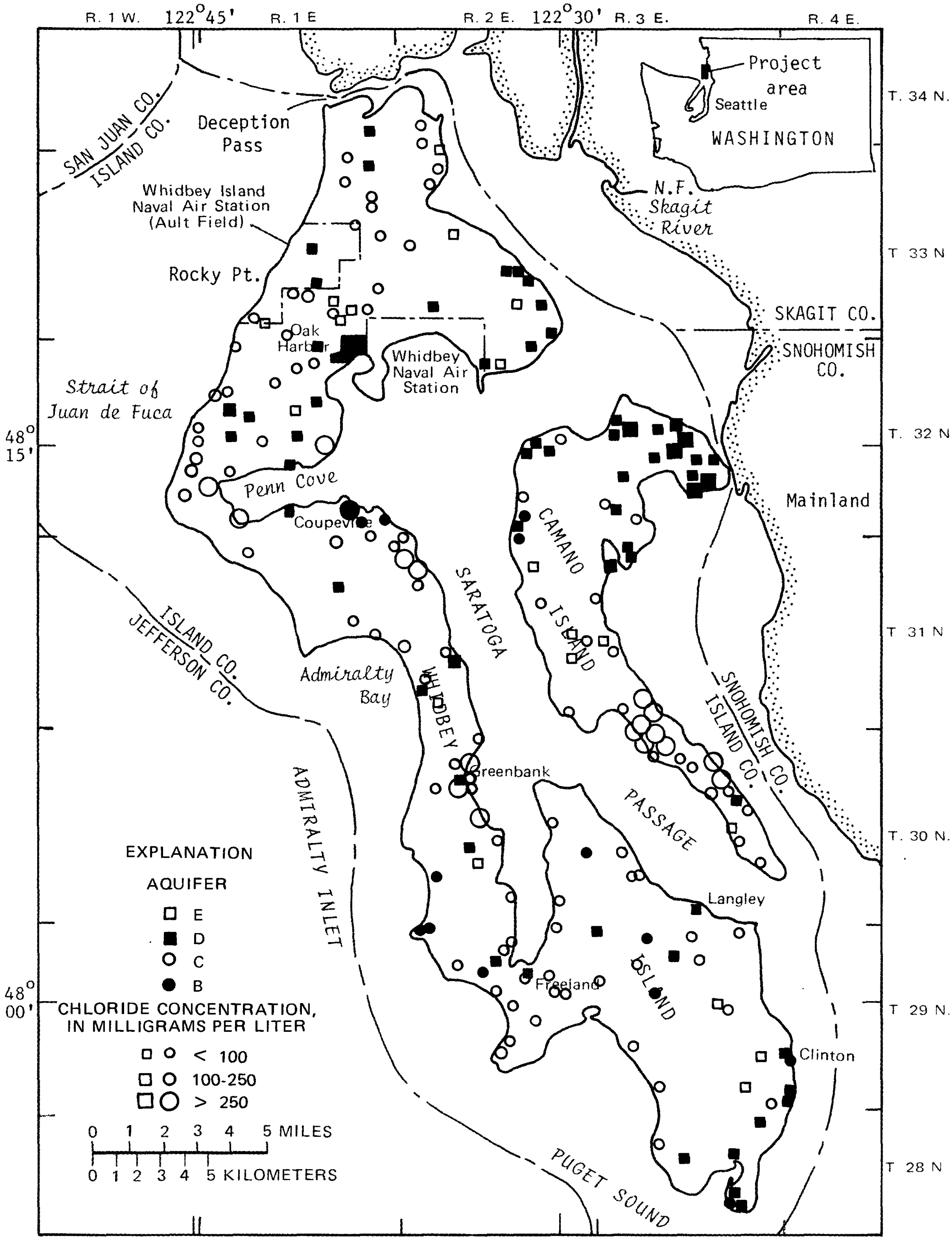


Figure 16.--Chloride concentrations for aquifers B through E in August 1981.

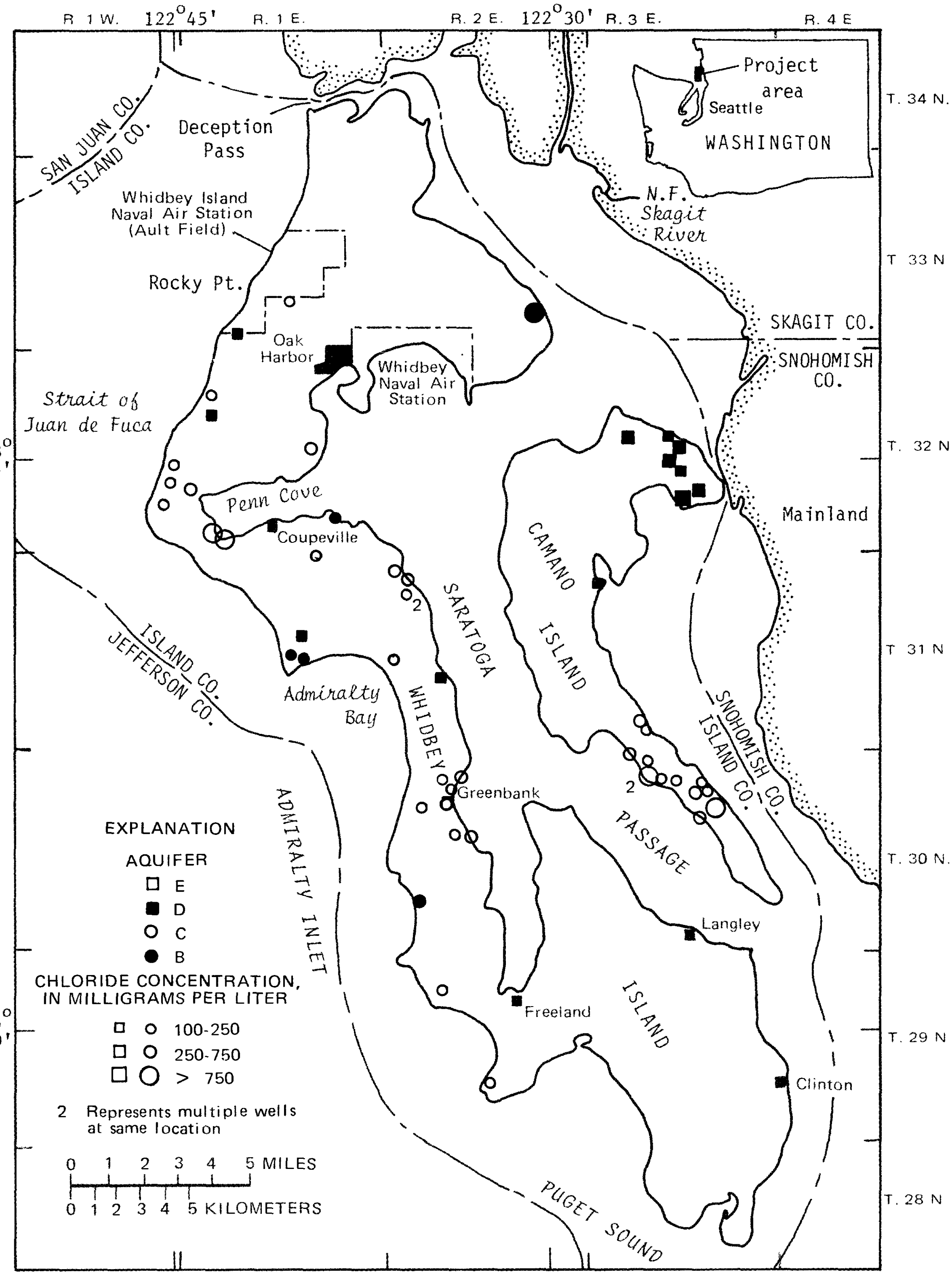


Figure 17.--Maximum chloride concentations for aquifers B through E in intruded wells for the month of August, 1980 to 1983.

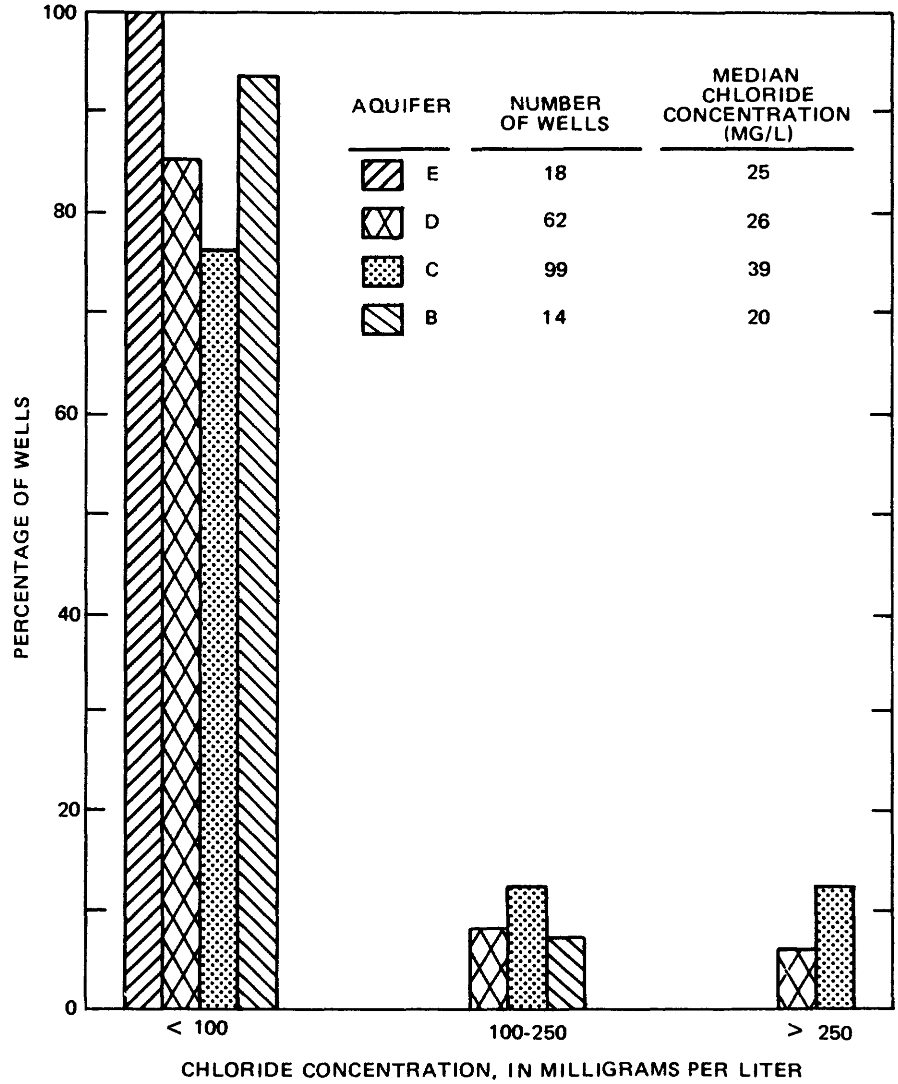


Figure 18.--Frequency distribution of chloride concentrations for aquifers B through E in August 1981.

Seasonal and Long-Term Variations in Chloride Concentrations

In general, the smallest concentrations of chloride would be expected in late winter and early spring months when hydraulic heads in the aquifers are high (see "Ground-Water Levels and Movement") and the mixing zone is at its deepest and most seaward position. Conversely, the largest concentrations of chloride would be expected in the late summer and early fall months when heads in aquifers are low and the mixing zone is at its shallowest and most inland position.

Monthly and bimonthly chloride concentrations were determined for 51 wells from April 1981 to August 1983; concentrations for two of the wells are shown in figure 19. Well 30/3E-10J1 was intruded by seawater, and chloride concentrations fluctuated seasonally by almost an order of magnitude. In contrast, seasonal chloride fluctuations were slight in well 28/3E-1E2, which was not intruded and had chloride concentrations less than 100 mg/L throughout the period. Of 51 sampled wells, about 60 percent had small chloride concentrations and showed little or no fluctuation seasonally, about 20 percent showed a trend of increasing chloride concentration from spring to fall, and about 20 percent showed random fluctuations from spring to fall (examples not shown).

Data for studying long-term trends were obtained by resampling 129 wells that were sampled during previous studies in the 1960's and 1970's. Most of the resampled wells had changes in chloride concentrations within a range of $\pm$25 mg/L. Variations of this magnitude were not considered large enough to indicate a long-term trend. Twenty-five of the wells showed a concentration increase or decrease of 25 mg/L or more over a period of at least 4 years, and these samples were examined in detail. Only 5 of 25 wells examined showed a steady increase or decrease, indicating an apparent long-term trend. Plots of chloride concentration over time for those five wells are shown in figure 20. Some physical and hydrologic characteristics of the wells are shown in table 6.

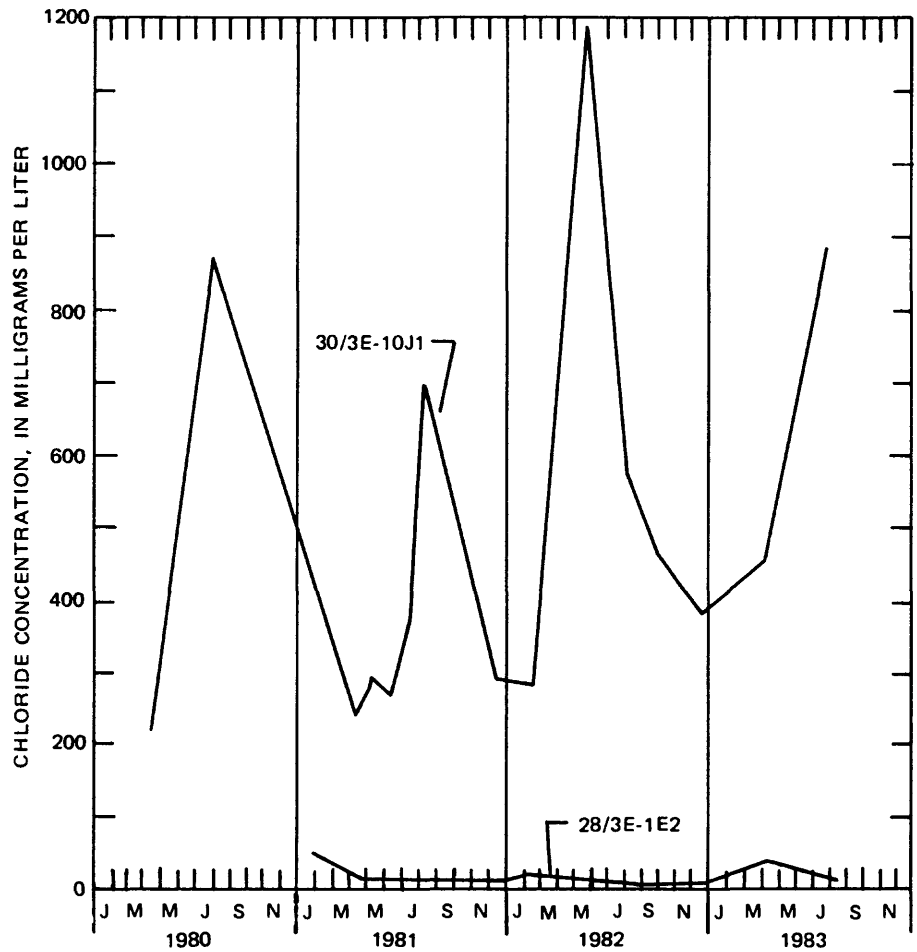


Figure 19.--Seasonal variations in chloride concentations for selected wells.

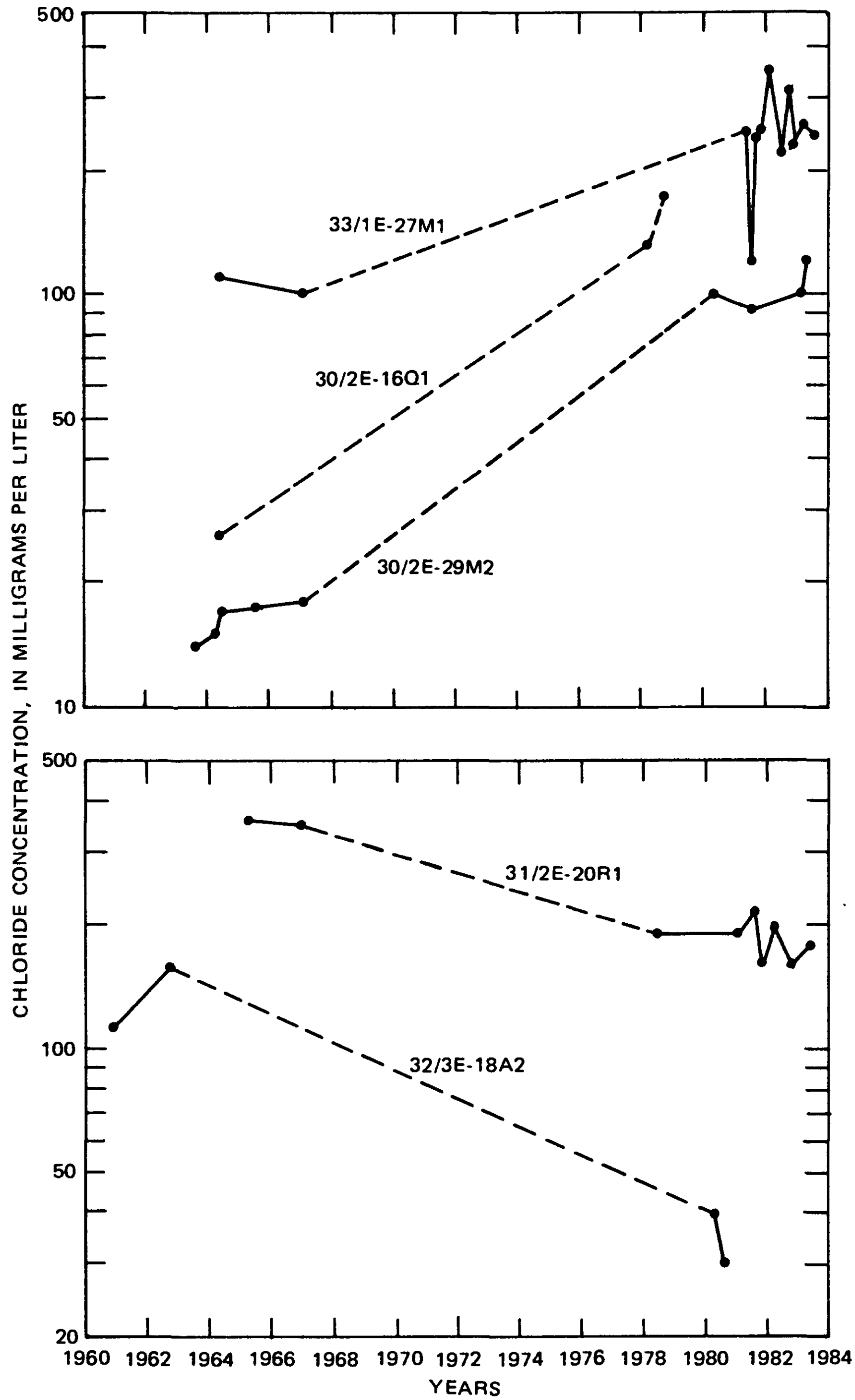


Figure 20.--Long-term variations in chloride concentations for selected wells.

Table 6.--Selected characteristics of wells showing long-term trends in chloride concentrations. See figure 20 for graphs

Well number	Use	Aquifer	Hydraulic head range for 1980-83 (feet)	Range
30/2E-16Q1	Domestic	C	+1 to -2	Located near coast south of Greenbank; low chloride concentrations in early 1980's.
30/2E-29M2	Public[1]	B	+2 to -67	South Whidbey State Park; low chloride concentrations in early 1960's.
31/2E-20R1	Domestic	D	+3.8 to -0.1	Located near coast north of Greenbank; pumpage decrease apparently has resulted in decrease in chloride concentration.
32/3E-18A2	Public	D	None available	Located near the coast on southern tip of Camano Island; presumably a decrease in pumpage resulted in a decrease in chloride concentration.
33/1E-27M1	Public	C	+8 to +2	Located inland north of Oak Harbor; chloride concentrations increased 2- to 3-fold since mid- to late-1960's.

[1] Public use is defined as two or more households using the same well.

Relation between Seawater Intrusion and Hydraulic Head

The relation between chloride concentrations and measured heads for aquifers B through E are shown in figures 21 and 22. As mentioned previously, chloride concentrations were less than 100 mg/L in all samples from aquifer E, and were less than 100 mg/L for most samples from aquifers B, C, and D. However, concentrations exceeded 100 mg/L in aquifers B, C, and D where these aquifers had low heads. Wells with low heads usually were located near the coast. The larger chloride concentrations were observed in wells tapping aquifers C and D where heads were between sea level and 10 feet above sea level. Chloride concentrations generally were small for wells with heads higher than 10 feet above sea level. Chloride concentrations in aquifer B generally were independent of head; this observation may be due to the small number of wells sampled in aquifer B. The outliers in figures 21 and 22 are probably wells that had not recovered to static conditions after recent pumping.

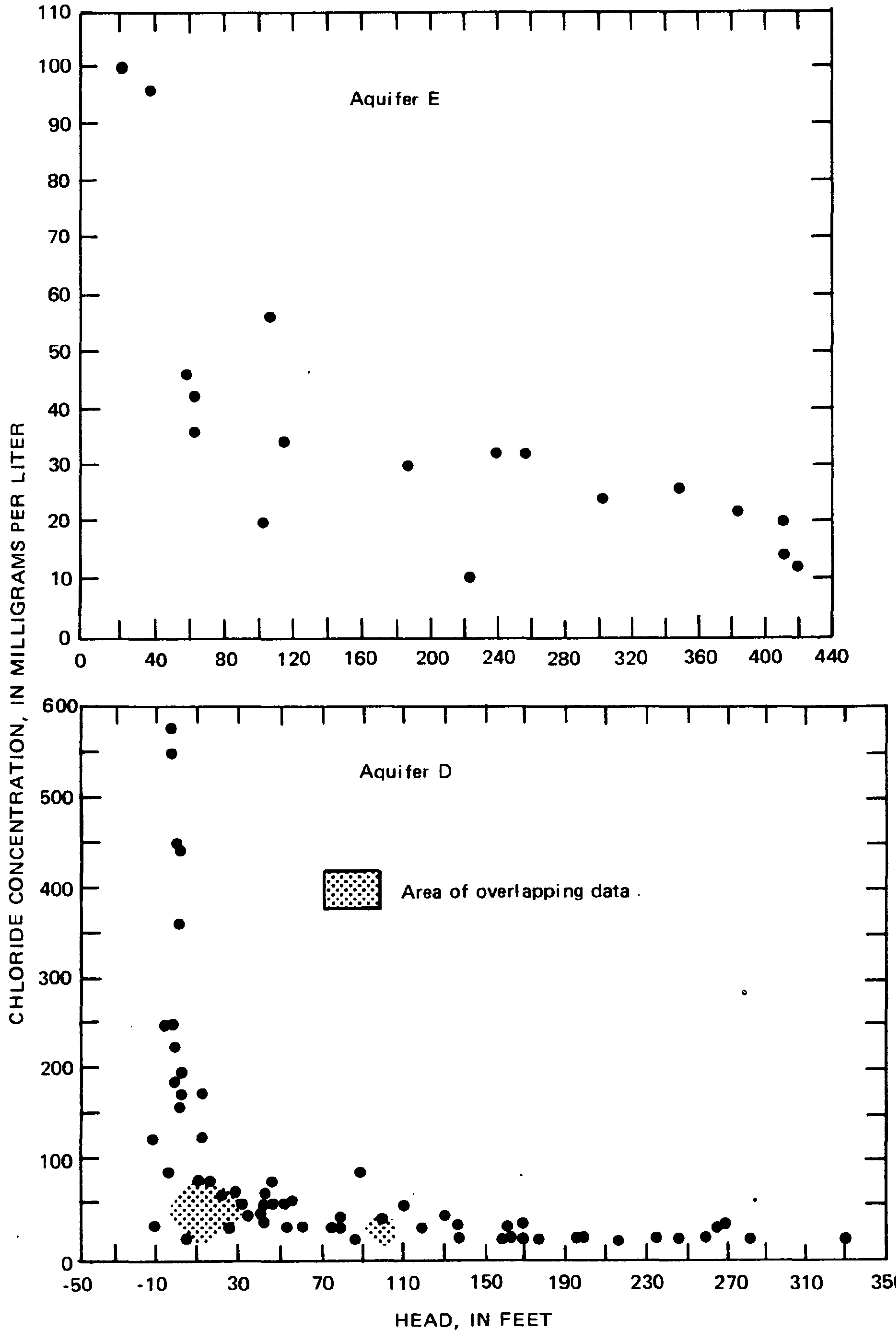


Figure 21.--Relation of chloride concentrations and heads for aquifers D and E in the month of August, 1980 to 1983.

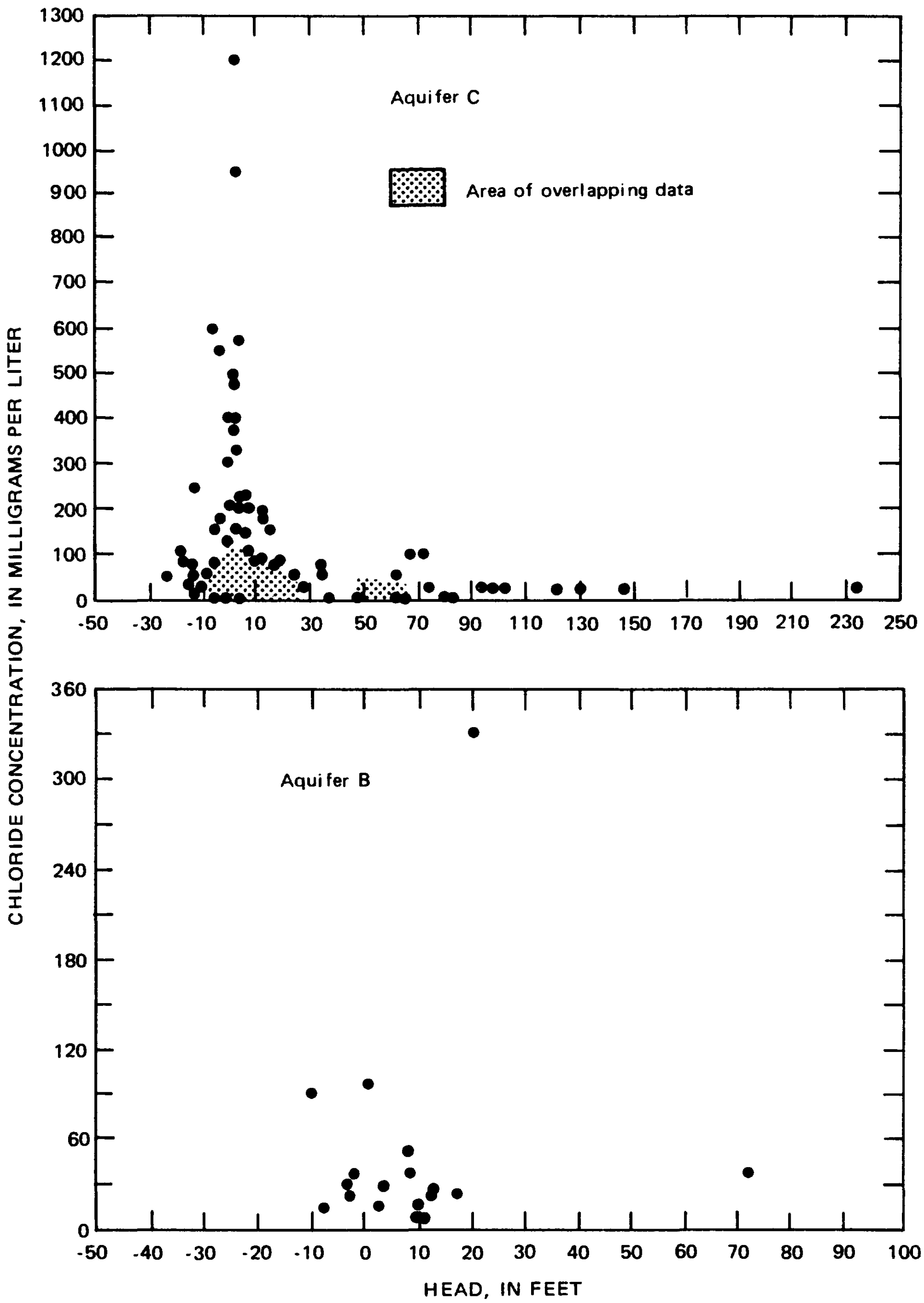


Figure 22.--Relation of chloride concentrations and heads for aquifers B and C in the month of August, 1980 to 1983.

The relation between head and chloride concentration in a ground-water system that includes a mixing zone cannot be expressed in simple mathematical terms. However, by assuming a sharp interface between freshwater and seawater, Hubbert (1940) developed the following equation to compute the depth below sea level to the interface:

$$D^i = \frac{\rho_f}{\rho_s - \rho_f} H_f^i - \frac{\rho_s}{\rho_s - \rho_f} H_s^i \quad , \tag{5}$$

where

D^i = depth below sea level to a sharp interface [L];
H_f^i = hydraulic head in the freshwater zone at the interface, in terms of freshwater, referenced to sea level [L];
H_s^i = hydraulic head in the seawater zone at the interface, in terms of seawater, referenced to sea level [L];
ρ_s = density of seawater [M/L^3]; and
ρ_f = density of freshwater [M/L^3].

If seawater is assumed to be stationary, then hydraulic head in the seawater zone H_s^i , is zero and equation 5 becomes

$$D^i = \frac{\rho_f}{\rho_s - \rho_f} H_f^i \tag{6}$$

Equation 6 is used in the ground-water-flow model described in "Simulation of Ground-Water Flow."

If head distribution in the freshwater zone is assumed to be hydrostatic, equation 6 becomes the Ghyben-Herzberg equation (Hubbert, 1940) where H_f^i is replaced by the altitude of the water table. The Ghyben-Herzberg equation is commonly used by hydrologists to estimate depth to interface, but it was not used in this study because the equation overestimates the interface depth in a region of downward ground-water movement, and underestimates the depth in a region of upward movement. These errors occur because of the vertical head gradients that occur in regions of vertical ground-water movement.

SIMULATION OF GROUND-WATER FLOW

Description of the Model

A numerical model was used to simulate the steady flow of fresh ground water in a multilayered hydrogeologic system that is bounded by seawater. An existing numerical model developed by Trescott (1975) to simulate three-dimensional flow of a constant-density fluid was modified by Sapik (1988) to compute the position of a stationary sharp interface between flowing freshwater and nonflowing seawater. The model computes heads in the freshwater zone by solving the finite-difference equations obtained from the following differential equation (Sapik, 1988) that describes ground-water flow in each layer of a multiple-aquifer system:

$$\frac{\partial}{\partial x}\left(T_x \frac{\partial H}{\partial x}\right) + \frac{\partial}{\partial y}\left(T_y \frac{\partial H}{\partial y}\right) + (C_z \Delta_z H)_a - (C_z \Delta_z H)_b = - Q_{re} + Q_w + Q_{sa} + Q_{sb}, \qquad (7)$$

where

x, y, z = directions in a rectangular coordinate system (x and y are in a horizontal plane and z is vertical) [L];

a, b = adjacent layers above and below, respectively;

T_x, T_y = hydraulic transmissivities in the x- and y-directions [L^2/t];

C_z = discharge coefficient for vertical leakage between adjacent layers [1/t];

H = hydraulic head, in terms of freshwater, referenced to sea level [L];

$\Delta_z H$ = vertical difference in head between adjacent layers [L];

Q_{re} = rate of freshwater recharge per unit horizontal area [L/t];

Q_w = rate of freshwater discharge from pumping wells per unit horizontal area [L/t];

Q_{sa} = rate of freshwater discharge from springs above sea level per unit horizontal area [L/t]; and

Q_{sb} = rate of freshwater discharge from springs below sea level per unit horizontal area [L/t].

Discharges Q_{sa} and Q_{sb} are computed using the equations

$$Q_{sa} = C_{sa}\ (H-Z_{sa}) \text{ for } H>Z_{sa}\ ,$$
$$Q_{sa} = 0 \qquad \text{ for } H\leq Z_{sa}\ , \tag{8}$$

and

$$Q_{sb} = C_{sb}\ (H-H_{sb}) \text{ for } H>H_{sb}\ ,$$
$$Q_{sb} = 0 \qquad \text{ for } H\leq H_{sb}\ , \tag{9}$$

where

C_{sa} and C_{sb} = discharge coefficients [1/t],

H_{sb} = freshwater head in the seawater zone at the top of a layer [L], and

Z_{sa} = altitude at the bottom of a layer [L].

All hydraulic parameters and discharges in equations 7, 8, and 9 can vary within a layer and between layers.

The model computes the depth to the interface, so that along the interface the fluid pressures in the freshwater and seawater zones are equal. This condition is satisfied by equation 6, which was rewritten by Sapik (1988) as shown below:

$$D^i = \frac{\rho_f}{\rho_s - \rho_f} H^i \quad , \tag{10}$$

where

D^i = depth below sea level to the freshwater-seawater interface [L];

H^i = hydraulic head (computed from eq. 7) in the freshwater zone at the interface, referenced to sea level [L];

ρ_s = density of seawater [M/L^3]; and

ρ_f = density of freshwater [M/L^3].

The model solves equations 7 and 10 using an iterative method in which equation 7 is solved for head, equation 10 is used to compute the interface depth, and these two computations are repeated until the computed interface depth changes only a small amount in two successive iterations (Sapik, 1988). When a solution is obtained, the parts of layers that are fully intruded by seawater are deleted from the freshwater flow system, and discharges, such as well and spring discharges, associated with these intruded areas are not included in mass-balance computations. Where layers are partly intruded by seawater, the freshwater thickness is reduced and the transmissivities, spring discharge coefficients, and well discharges are reduced by an amount proportionate to the reduction in freshwater thickness. This reduction in freshwater outflow from springs and wells is needed to balance the reduction in freshwater inflow caused by lowering the transmissivity. The reason for reducing well discharge in intruded cells is to stabilize the computational procedure used by the model for locating the interface between freshwater and seawater (Sapik, 1988). Reductions in well discharge will affect the results of model calibration; therefore, any calibration must be repeated until well discharges in the calibrated model agree with discharges specified as model input.

The interface-location procedure described above was derived from one used by Guswa and Le Blanc (1981), but their procedure was not used in the model described by Sapik (1988) because the procedure would not converge to the correct solution for equations 7 and 10.

Specifications for the Model

The model described in the preceding section was used to simulate ground-water flow in the aquifer system underlying Whidbey and Camano Islands. Each aquifer and confining unit in the aquifer system was represented as a layer in the model. Each layer was subdivided into cubelike cells by superimposing a common rectangular grid on the layers (pl. 3). The spacings between grid lines ranged from 1,000 to 2,700 feet.

To solve equations 7 and 10, model-input data for each cell included specifications for boundaries that surround the model and estimates of the numerous parameters in equations 7 through 10. These parameters included hydraulic transmissivities, vertical leakage coefficients, spring discharge coefficients, well pumping and surface-recharge rates, fluid densities, dimensions of the rectangular grid, and altitudes at the bottom of each cell. Initial estimates made for some model parameters were changed during model calibration (see "Calibration of the Steady-State Flow Model")

Boundaries

The entire modeled area could not be included into a single model because of limitations on available computer memory. Consequently, four submodels were constructed (pl. 3)--three on Whidbey Island and one on Camano Island. The lateral boundaries of each submodel were placed about 2 miles offshore in order to include the entire mixing zone that separates freshwater and seawater. Each of the submodels overlaps a sufficient distance so that the types of boundaries surrounding a submodel have little effect on heads computed in the non-overlap area of a submodel (see pl. 3). A condition of model calibration (see "Calibration of the Steady-State Flow Model") is that differences in head computed for adjacent submodels must be less than 1 foot in the common overlap areas.

Boundaries that surround the submodels can be specified as flows or heads. At the base of the model, either where the layers are in contact with bedrock or at the bottom of confining unit A, the assumed boundary condition is that there is no flow across this surface; this condition is satisfied by default within the model. The boundary condition at the top of the model, where the uppermost layer is above sea level, is satisfied by specifying the recharge rate, Q_{re} in equation 7, that was determined by methods described in "Recharge to the Aquifer System." Where a model layer is below sea level and in contact with the bottom of Puget Sound, the boundary condition is specified as a spring below sea level (Q_{sb} in eqs. 7 and 9). Where a model layer intersects land surface (above sea level), the boundary condition is specified as a spring above sea level (Q_{sa} in eqs. 7 and 8). For layers that extend to submodel boundaries shown on plate 3, lateral no-flow boundaries were specified. Boundaries around the submodels were oriented, using available field data for head in the aquifers, so that ground water would not flow across these boundaries. The validity of these no-flow boundaries is discussed in "Calibration of a Steady-State Flow Model."

Geometry and Hydraulic Characteristics

The Island County model contains 10 layers (see "Definition of the Aquifer System"), with aquifer E at the top and the confining unit underlying aquifer A at the bottom (fig. 3). Some of the layers extended to the model boundaries shown on plate 3, and other layers were limited in their extent, as shown on plate 2.

Thicknesses of aquifers B through E and confining units B through E were obtained from well drillers' logs, logs of test holes drilled during this study, and geophysical logs obtained during this study as described in "Definition of the Aquifer System." These thicknesses vary areally as shown in plate 2. The thickness of aquifer A was unknown because few wells completely penetrated this aquifer; therefore, the average thickness of aquifer B, about 50 feet, was used for aquifer A. The type and distribution of water-bearing materials below aquifer A are unknown; therefore, confining unit A was assumed to extend to the bottom of the model, which is 900 feet below sea level. As discussed in "Sensitivity Analysis for the Calibrated Model", this confining unit was replaced by an aquifer of the same thickness to determine how differences in geology affect the model computations. Placing the bottom of the model at 900 feet below sea level was done after making trial runs with the model before calibration to determine the maximum depth of the freshwater-seawater interface.

Hydraulic properties of the aquifers and confining units were specified for each model cell, and these properties were used to estimate the numerous parameters in equations 7 through 10. Hydraulic transmissivity was computed as the product of horizontal hydraulic conductivity and saturated thickness for each model cell. Initial estimates of horizontal hydraulic conductivities (table 2) for model calibration were obtained as described in "Hydraulic

Characteristics of the Aquifer System." The horizontal hydraulic conductivities were assumed to be independent of direction, and a common transmissivity, T, was used for both T_x and T_y in equation 7. For most aquifer layers, two different horizontal hydraulic conductivities were used: one value for areas where the material making up the layer was typical of aquifers, and a smaller value where the material was finer grained.

Vertical leakage coefficients (C_z in eq. 7) are used by the model to compute fluxes between layers. These leakage coefficients are part of the model-input data and were computed as

$$C_z = \frac{2\, K_z^a\, K_z}{K_z^a\, \Delta z + K_z\, \Delta^a z} \tag{11}$$

where

a = adjacent layer (above or below),
Δz = layer thickness [L], and
K_z = vertical hydraulic conductivity for model layer [L/t].

This equation was derived using Darcy's law to express vertical flow between cells having different hydraulic conductivities and thicknesses (Sapik, 1988). Initial estimates of vertical hydraulic conductivities used to compute C_z were 0.01 times the horizontal hydraulic conductivities (see "Hydraulic Characteristics of the Aquifer System" and table 2).

Fluid densities in equation 10 are constant for all model cells and were specified as 1.0 for freshwater and 1.021 for seawater in Puget Sound (reference temperature is 50 °F for both densities). The density for seawater was computed from average densities of Puget Sound water near Anacortes, Port Townsend, and Everett, Wash. (locations are north, south, and east of Island County, respectively). Observed densities ranged from 1.0148 to 1.0231 at 59 °F (U.S. Department of Commerce, 1954).

Recharge and Discharge

Recharge (Q_{re} in eq. 7) from precipitation and pumped ground water was estimated as outlined in "Recharge to the Aquifer System." In the model, recharge was applied to the uppermost active cell in each column of cells above sea level. Recharge data used for input to the calibrated ground-water-flow models are stored on computer files described in Appendix I.

Discharge from wells (Q_w in eq. 7) was estimated as outlined in "Discharge from the Aquifer System." Well discharges are in units of cubic feet per second, and these data are converted within the model to the desired units of feet per second after dividing by the horizontal areas of cells. Well-discharge data used for input to the calibrated ground-water-flow models are stored on computer files described in Appendix I.

Discharge from springs above sea level (Q_{sa} in eq. 7) and from springs below sea level (Q_{sb} in eq. 7) are computed by the model using equations 8 and 9, respectively. Initial estimates of the spring discharge coefficients C_{sa} and C_{sb}, used in equations 8 and 9, are part of the model-input data and were computed as

$$
\begin{aligned}
C'_{sa} &= 2T\ \Delta y/\Delta x, \text{ for discharge in the x-direction,} \\
&= 2T\ \Delta x/\Delta y, \text{ for discharge in the y-direction,}
\end{aligned} \tag{12}
$$

and

$$
\begin{aligned}
C'_{sb} &= 2K_z\ \Delta x\Delta y/\Delta z, \text{ for discharge in the z-direction,} \\
&= 2T\ \Delta y/\Delta x, \text{ for discharge in the x-direction, and} \\
&= 2T\ \Delta x/\Delta y, \text{ for discharge in the y-direction,}
\end{aligned} \tag{13}
$$

where Δx, Δy are cell dimensions in the x- and y-directions. These equations were derived from Darcy's law (Sapik, 1988). To obtain the coefficients C_{sa} and C_{sb} in equations 8 and 9, the model divides the above coefficients by the horizontal areas of cells. Springs above sea level were assumed to discharge horizontally, whereas springs below sea level could discharge horizontally or vertically. The direction of spring discharge was determined before computing the discharge coefficient, and only one coefficient was computed for each model cell containing a spring. Data used by the calibrated ground-water-flow models to compute discharge from springs above and below sea level are stored on computer files described in Appendix I.

Calibration of the Steady-State Flow Model

Procedure

The calibration process consisted of running each of the four submodels numerous times using different values of hydraulic parameters until there was agreement between computed and observed values for aquifer heads, total spring discharge above sea level, and, to a lesser extent, for depths to the freshwater-seawater interface. Because the model reduces freshwater discharge from wells intruded by seawater (see "Description of the Model"), a condition of calibration was that well discharges in the submodels agree with discharges specified as model input. Another condition of calibration was that differences in computed head be less than 1 foot in the common overlap areas of adjacent submodels. A final condition of calibration was that ground water should not move across the no-flow boundaries specified around each submodel.

The hydraulic parameters that were varied during calibration were the horizontal hydraulic conductivities used to compute transmissivities of the aquifer layers, the vertical hydraulic conductivities of the confining layers used to compute vertical leakage coefficients between layers, and the discharge coefficients for springs above and below sea level. Horizontal hydraulic conductivities used to compute transmissivities of confining layers and vertical hydraulic conductivities of aquifers used to compute vertical leakage coefficients were not varied during calibration because model results were insensitive to changes in these parameters. During calibration, the discharge coefficients for springs above sea level were varied independently from the coefficients for springs below sea level, and both were varied independently from transmissivities and vertical leakage coefficients, even though all are functions of the same hydraulic conductivities. This is justified because equations 12 and 13 were derived for ideal geometric and flow conditions that only approximate field conditions, and the initial estimates of these coefficients would probably require adjustments even if the hydraulic conductivities were known precisely.

The observed heads used for calibration were time-averaged values of observations made during the 1981 calendar year. That year was chosen because of the abundance and areal distribution of available data, and because average head in 1981 was not much different than in other years (fig. 10).

Most of the observed chloride concentrations in water samples from wells were less than 1,000 mg/L and could not be used to determine the position of the freshwater-seawater interface. Consequently, the usefulness of these data were limited to providing upper limits for the altitude of the interface.

The first computations with the models, using initial estimates of all parameters, resulted in computed discharges from springs above sea level that were much higher than those observed. Therefore, the discharge coefficients for these springs were lowered and those for springs below sea level were raised until computed and observed discharges agreed for springs above sea level. Next, horizontal hydraulic conductivities of aquifer layers and vertical hydraulic conductivities of confining layers were varied to obtain agreement between computed and observed values of heads and depths to the interface. Varying these conductivities did not have a large effect on discharges from springs above sea level; therefore, only small additional adjustments of the spring discharge coefficients were necessary.

During calibration the values of a hydraulic parameter for all cells within a model layer were varied by using a single multiplier whenever possible. Only in a few instances were different multipliers used in different parts of a layer to get the desired degree of agreement between computed and observed values. In the overlapping areas of adjacent submodels, the adjusted hydraulic parameters were the same in both submodels.

Results

Hydraulic parameters

Horizontal hydraulic conductivities used for computing transmissivities of aquifer layers in the calibrated model are given on plate 3 and table 7. The horizontal hydraulic conductivities for the confining layers are also given in table 7. The maximum adjustments made to the initial estimates of horizontal hydraulic conductivity during calibration were to use multipliers of 0.1 for aquifer E and 10 for aquifer D on Whidbey Island, and 0.01 for aquifer E on Camano Island. Multipliers used for other aquifers ranged from 0.1 to 6.7. The adjustments for aquifer E are not unreasonable because the initial estimate for this aquifer was about 10 times greater than the maximum estimate for other aquifers, and was based on data from only three wells.

Table 7.--Horizontal hydraulic conductivity for aquifers and confining units in calibrated models

Aquifer	Confining unit	Hydraulic conductivity, in feet per second: Whidbey Island	Camano Island	Remarks
E		5.7×10^{-4}	5.7×10^{-5} to 3.8×10^{-5}	See plate 3
	E	1.0×10^{-7}	1.0×10^{-7}	
D		9.4×10^{-5} to 9.4×10^{-3}	9.4×10^{-5} to 6.3×10^{-3}	See plate 3
	D	1.0×10^{-7}	1.0×10^{-7}	
C		1.6×10^{-3}	7.9×10^{-5} to 5.3×10^{-3}	See plate 3
	C	1.0×10^{-7}	1.0×10^{-7}	
B		2.4×10^{-4}	4.9×10^{-5}	
	B	1.0×10^{-7}	1.0×10^{-7}	
A		2.4×10^{-4}	4.9×10^{-5}	
	A	1.0×10^{-7}	1.0×10^{-7}	

The vertical hydraulic conductivities of confining layers used for computing vertical leakage coefficients between layers in the calibrated model are given in table 8. The maximum adjustments made to vertical hydraulic conductivity during calibration were to use multipliers of 15 for confining layers A, B, C, and E on Whidbey Island, and 10 for confining layers D and E on Camano Island. A single multiplier was used for each confining layer except for confining layer D on Whidbey Island, and for this layer the multipliers ranged from 3 to 10. The vertical hydraulic conductivity of confining layer D is shown on plate 3.

The initial discharge coefficients for springs above sea level on both islands were multiplied by 0.0002 for both aquifer E and confining layer E, and by 0.002 for the other aquifers and confining layers. The initial discharge coefficients for springs below sea level on both islands were multiplied by 5 for all aquifers and confining layers. Before these multipliers were used, spring discharge computed by the model (eq. 8) was greater than observed discharge. The large adjustment in the coefficient for springs above sea level was probably required because equation 8 overestimates discharge where aquifers and confining layers are not fully saturated and where there is a big difference between the average altitude of a seepage face and the bottom of a layer.

Table 8.--Vertical hydraulic conductivity for confining units in calibrated models

Confining unit	Hydraulic conductivity, in feet per second		Remarks
	Whidbey Island	Camano Island	
E	1.5×10^{-8}	1.0×10^{-8}	
D	3.0×10^{-9} to 1.0×10^{-8}	1.0×10^{-8}	See plate 3
C	1.5×10^{-8}	1.0×10^{-9}	
B	1.5×10^{-8}	1.0×10^{-9}	
A	1.5×10^{-8}	1.0×10^{-9}	

Comparison of model results with observations

Model-computed heads for aquifers A through E are shown on plate 4. These maps were prepared by combining head maps for the individual submodels. The general pattern of lateral ground-water movement, inferred from heads, is from the center of each island outward toward Puget Sound. However, on northeast Camano Island, hydraulic gradients for model-computed head in aquifer D are very small and to the east, but observed heads indicate that ground water could be moving westward from the mainland. There are several irrigation wells on northeast Camano Island that pump from aquifer D, and this pumpage is apparently the cause of the westward hydraulic gradients for observed heads. Directions of vertical flow inferred from model-computed heads are downward beneath high-altitude areas of the islands and upward beneath low-altitude areas and offshore from the coastline (fig. 3).

One of the conditions of calibration was that ground water should not move across submodel boundaries. This condition did not hold for aquifers A, B, and C along the north side of the Camano Island submodel. Hydraulic gradients in aquifers A and B are very small, indicating a small amount of flow across the submodel boundary. However, larger hydraulic gradients in aquifer C indicate a larger amount of flow across the submodel boundary; therefore, there is some error in this part of the model.

Observed heads also are shown on plate 4. Differences between computed and observed heads are as much as 172 feet, but half the differences are less than 8 feet. There are various reasons for the differences. Probably the most important are inaccuracies in the model caused by (1) the assumption that the aquifer layers are continuous over the modeled area, (2) insufficient data to completely define the magnitudes and variations of layer thicknesses and hydraulic conductivities, and (3) the lack of a direct method for determining recharge from precipitation. Differences between computed and observed heads exist also because the model computes heads at nodes which are located at the centers of cells, whereas most of the observation wells are not drilled at locations corresponding to the nodes. Therefore, significant differences can exist wherever an observation well is located near a cell boundary and in areas where there are large head gradients (vertical or horizontal) across a cell.

Computed depths below sea level to the freshwater-seawater interface are shown on plate 4. The greatest depth exceeds 900 feet below sea level, on southern Whidbey Island. The interface has a gentle slope beneath the central parts of the islands, but offshore the interface slopes steeply upward to intersect the bottom of Puget Sound. In most areas separating Whidbey and Camano Islands, the interface intersects the bottom of the Sound, indicating that there is no movement of fresh ground water between the two islands. However, the computed interface does not intersect the bottom of the Sound in the following areas: north Whidbey Island, where freshwater extends to bedrock; along the east side of north Whidbey Island, where the interface is truncated by the model boundary; beneath the section of Saratoga Passage that separates northeast Whidbey and north Camano Islands; and along the east side of northeast Camano Island, where the island is separated from the mainland (see fig. 1) by sloughs and marshes that are generally less than 10 feet deep. This latter area is the only location where observed heads indicate that ground water could be moving westward from the mainland to the study area in sections 16, 21, 22, and 27 of T.32 N., R.3 E. (see pl. 4). Elsewhere, the ground-water systems of the islands are separated from the mainland and from each other.

Seawater intrusion by upconing into aquifer A has left pockets of freshwater beneath the southern half of Whidbey Island (pl. 4). These freshwater pockets occur where the freshwater-seawater interface is below the tops of model cells; elsewhere, the interface is above the tops of model cells. The position of the interface along the bottom of the model is located close to aquifer A, and very small changes in head caused by changes in model parameters during calibration will cause intrusion into the aquifer. Errors in the location and extent of the freshwater pockets shown on plate 4 could be due partly to errors in model calibration and partly to the lack of data available for defining the aquifer (see "Definition of the Aquifer System").

Depths below sea level and observed chloride concentrations are shown on plate 4 for (1) wells with observed chloride concentrations of 100 mg/L or more, and (2) wells that extend 100 feet or more below sea level. Chloride concentrations greater than 100 mg/L are indicative of ground water in the mixing zone that separates freshwater from Puget Sound water, which has a chloride concentration of about 16,000 mg/L (see "Seawater Intrusion into Aquifers"). The computed depths to the interface are greater than the depths of nearly all the wells and test holes shown on plate 4, as they should be, because most of the wells had chloride concentrations significantly less than that of Puget Sound water. One well, located at 32/1E-32N1, had a chloride concentration of 14,000 mg/L; however, this well is located in a low-elevation area along Penn Cove and is subject to lateral intrusion.

Discharges from springs above sea level computed by the calibrated model were 3.98 and 1.82 ft^3/s on Whidbey and Camano Islands, respectively (tables 9 and 10). The corresponding observed discharges were 4 and 2 ft^3/s. Discharges charges for springs below sea level could not be observed, but model-computed discharges were 135.53 ft^3/s for Whidbey Island and 39.40 ft^3/s for Camano Island.

Table 9.--Water budget for calibrated model of Whidbey Island

Aquifer	Confining unit	Inflow, in cubic feet per second: Surface recharge: Precipitation	Surface recharge: Wastewater[1]	From layer above	From layer below
E		54.31	1.22	0	0.15
	E	47.27	.60	54.45	.63
D		11.55	.09	102.15	.37
	D	25.20	.32	75.70	1.31
C		2.93	.09	99.91	1.11
	C	.10	<.01	3.16	1.25
B		0	0	3.23	.22
	B	0	0	.22	.21
A		0	0	.21	.01
	A	0	0	.01	0
	Totals	141.36	2.32	--	--

Aquifer	Confining unit	Outflow, in cubic feet per second: Springs above sea level	Springs below sea level	To layer above	To layer below	Pumping wells
E		0.44	0.71	0	54.45	0.08
	E	<.01	.65	.15	102.15	0
D		3.54	33.47	.63	75.70	.82
	D	<.01	2.25	.37	99.91	0
C		<.01	96.55	1.31	3.16	3.02
	C	0	.17	1.11	3.23	0
B		0	1.73	1.25	.22	.25
	B	0	<.01	.22	.21	0
A		0	0	.21	.01	0
	A	0	0	.01	0	
	Totals	3.98	135.53	--	--	4.17

[1] Includes percolating water from irrigation and septic tanks.

Table 10.--Water budget for calibrated model of Camano Island

		Inflow, in cubic feet per second			
		Surface recharge			
Aquifer	Confining unit	Precipitation	Wastewater[1]	From layer above	From layer below
E		15.75	0.21	0	0
	E	20.92	.41	15.22	.05
D		3.66	.02	36.46	2.54
	D	1.22	.02	11.36	2.75
C		0	0	12.47	.10
	C	0	0	.59	.10
B		0	0	.58	.02
	B	0	0	.02	.01
A		0	0	.01	<.01
	A	0	0	<.01	0
	Totals	41.55	0.66	--	--

		Outflow, in cubic feet per second				
Aquifer	Confining unit	Springs above sea level	Springs below sea level	To layer above	To layer below	Pumping wells
E		0.64	0	0	15.22	0.10
	E	<.01	.14	0	36.46	0
D		1.18	29.66	.05	11.36	.43
	D	<.01	.34	2.54	12.47	0
C		0	8.88	2.75	.59	.35
	C	0	.01	.10	.58	0
B		0	.37	.10	.02	.11
	B	0	<.01	.02	.01	0
A		0	0	.01	<.01	0
	A	0	0	<.01	0	0
	Totals	1.82	39.40	--	--	0.99

[1] Includes percolating water from irrigation and septic tanks.

Sensitivity Analysis for the Calibrated Model

The sensitivity of model-computed parameters to model-input parameters was investigated by running the calibrated model with changes in input parameters. The model-computed parameters used for these tests were depth to the freshwater-seawater interface and spring discharge (above and below sea level); the model-input parameters were transmissivities of all layers, vertical leakage coefficients, discharge coefficients for springs, and recharge to layers above sea level. These sensitivity tests also included changing confining layer A to an aquifer in order to evaluate the consequences of the lack of hydrogeologic data for the lower layers. Model-computed head was not used as a parameter in the tests because the effects of head changes are reflected in changing interface depths and spring discharges, and because head changes are difficult to summarize when there are large variations (in magnitude and direction) between layers. All tests were made using only the submodel for north Whidbey Island; however, the results are probably representative of the entire modeled area because materials underlying each submodel area have similar hydraulic characteristics. The results of sensitivity tests are summarized in table 11. Sensitivity was measured by changes in average depth to the interface and changes in total discharge of springs above and below sea level.

The sensitivity tests (table 11) showed that simulated changes in recharge and transmissivity caused the largest changes in model-computed depth to the interface. Computed discharge from springs above and below sea level was most sensitive to changes in recharge. Changing confining layer A to an aquifer, with the same hydraulic characteristics as aquifer A, had no measurable effect on model-computed spring discharge and very little effect on the vertical position of the freshwater-seawater interface; therefore, not knowing the geology below aquifer A does not significantly affect model results.

Table 11. Changes in model-computed parameters caused by changes in model-input parameters

Model-input parameter	Change in model-input parameter	Change in model-computed parameter, in percent: Average depth to interface	Spring discharge above sea level	Spring discharge below sea level
Layer transmissivities	20-percent increase	-14.5	-8.2	0.1
	20-percent decrease	20.1	11.0	-.2
Vertical leakage coefficients between layers	20-percent increase	4.3	-18.5	.3
	20-percent decrease	-5.6	27.4	-.4
Discharge coefficients for springs above sea level	20-percent increase	-.3	18.0	-.2
	20-percent decrease	.5	-18.6	.3
Discharge coefficients for springs below sea level	20-percent increase	-.6	-.6	.2
	20-percent decrease	2.2	1.0	.1
Change confining bed A to an aquifer	--	.7	0	0
Recharge to layers above sea level	20-percent increase	15.6	35.3	20.3
	20-percent decrease	-16.6	-32.0	-20.3
Density of seawater	From 1.021 to 1.025	-5.7	.9	0
	From 1.021 to 1.017	13.8	-1.5	.2

Limitations and Use of the Model

The usefulness of the model calibrated in this study is limited by the assumptions used in developing the numerical model. The assumptions used by Sapik (1988) to develop the numerical model are: (1) freshwater and seawater are separated by a sharp interface and the two fluids do not mix; (2) the two fluids have constant but different densities; (3) seawater is static; (4) the flow of freshwater is steady (meaning that hydraulic head and interface position do not change with time); and (5) there is no ground-water movement across the artificial boundaries that surround the model. These assumptions rarely correspond to conditions in natural ground-water systems, therefore the effects of these assumptions on model results are discussed below.

The assumption of a sharp interface means that model results will show freshwater in cells adjacent to intruded cells. However, corresponding freshwater cells in the natural system may contain water with chloride concentrations exceeding 1,000 mg/L because the mixing zone may be several hundred feet wide (see "Seawater Intrusion into Aquifers"). The model does not simulate vertical leakage of freshwater into cells that are totally intruded because of the assumption that freshwater and seawater do not mix. However, the model does simulate vertical leakage into cells that are partly intruded.

An average density was used for each fluid (freshwater and seawater), and each density was constant throughout the model (see "Specifications for the Model"). However, fluid densities in natural systems vary both vertically and laterally. The effects of varying fluid densities on model results are examined in "Sensitivity Analysis for the Calibrated Model."

Because seawater is assumed to be static in the model, head in the seawater zone is greater than head that would be computed by a model that simulates moving seawater. Therefore, the interface position computed by this model is shallower and farther inland than it would be in a model that simulates moving seawater.

Using a steady-state model to simulate a natural system that is transient can affect model results. Although there were short-term fluctuations in ground-water levels, there were no significant trends in water levels for the period 1963-83 (see "Ground-Water Levels and Movement"); therefore, the ground-water-flow system was assumed to be in a steady state for 1963-83. Because the system appears to be in steady state for the long term, time-averaged heads were used to calibrate the model.

Because the model simulates only steady-state conditions, computed responses to changing stresses represent long-term average responses. The model does not determine how long it takes the ground-water-flow system to reach equilibrium under newly imposed stresses, which are simulated as being applied continuously and forever. For example, if model computations indicate that the interface moves 1,000 feet horizontally inland in response to a new well pumping at a specified rate, the actual time for the interface to move the 1,000 feet is not known.

The submodels calibrated in this study could be used to simulate responses to changing stresses on the ground-water-flow system in Island County, provided that these stress changes do not cause movement of ground water across the artificial no-flow boundaries that surround each submodel. Stresses can be in the form of ground-water pumpage or recharge. Changes in pumpage could include the addition of new wells, the abandonment of existing wells, or changing the pumping rates of existing wells. Changes in recharge could include covering an existing recharge site or developing new artificial recharge sites. The responses to changing stresses on a ground-water system are seen as changes in hydraulic heads (or water-level altitudes) which, in turn, cause the freshwater-seawater interface to move vertically and horizontally, and cause changes in discharges from springs above and below sea level. If hydraulic heads are lowered because of pumping, seawater can intrude aquifers and this will cause an increase in the chloride concentration of water pumped from wells. It is likely that chloride in a well will increase gradually because the freshwater side of the mixing zone precedes seawater when intrusion occurs. The model described herein will simulate areal responses to changes in stress, but it cannot be used to simulate responses at a point because each model cell has a volume that is larger than that of a point. A typical aquifer cell has a horizontal area of 4,000,000 ft^2 and is 120 feet thick.

As a demonstration case, the steady-state model calibrated in this study was used to simulate how the freshwater-seawater interface would respond to additional pumping from a new well drilled into either aquifer B or C. The well was located near Oak Harbor (pl. 4), the largest population center in the county, but away from the coastline to minimize the lateral intrusion of seawater. (The selection of this site was for demonstration purposes only and should <u>not</u> be interpreted as a recommendation for development of a well at this site or as information about a new development planned for the site.) The new well was pumped from aquifer C at 0.50 ft^3/s in one test, and at 0.13 ft^3/s from aquifer B in another test. The lower pumping rate for aquifer B was selected because, in an earlier test, pumping at 0.50 ft^3/s caused total intrusion into the model cell being pumped. The results of the tests (fig. 23) show that pumping at 0.13 ft^3/s from aquifer B at this site would cause upconing of seawater into this aquifer; however, pumping at 0.5 ft^3/s from aquifer C would not cause significant intrusion of seawater into either aquifer B or C. Pumping from aquifer C did not change the position of the freshwater-seawater interface where the aquifer intersects the bottom of Puget Sound (shown in fig. 23), indicating that the radius of influence for the pumping well did not extend to the Sound. The time required for the interface to move the distance shown in figure 23 is not known because of reasons explained previously.

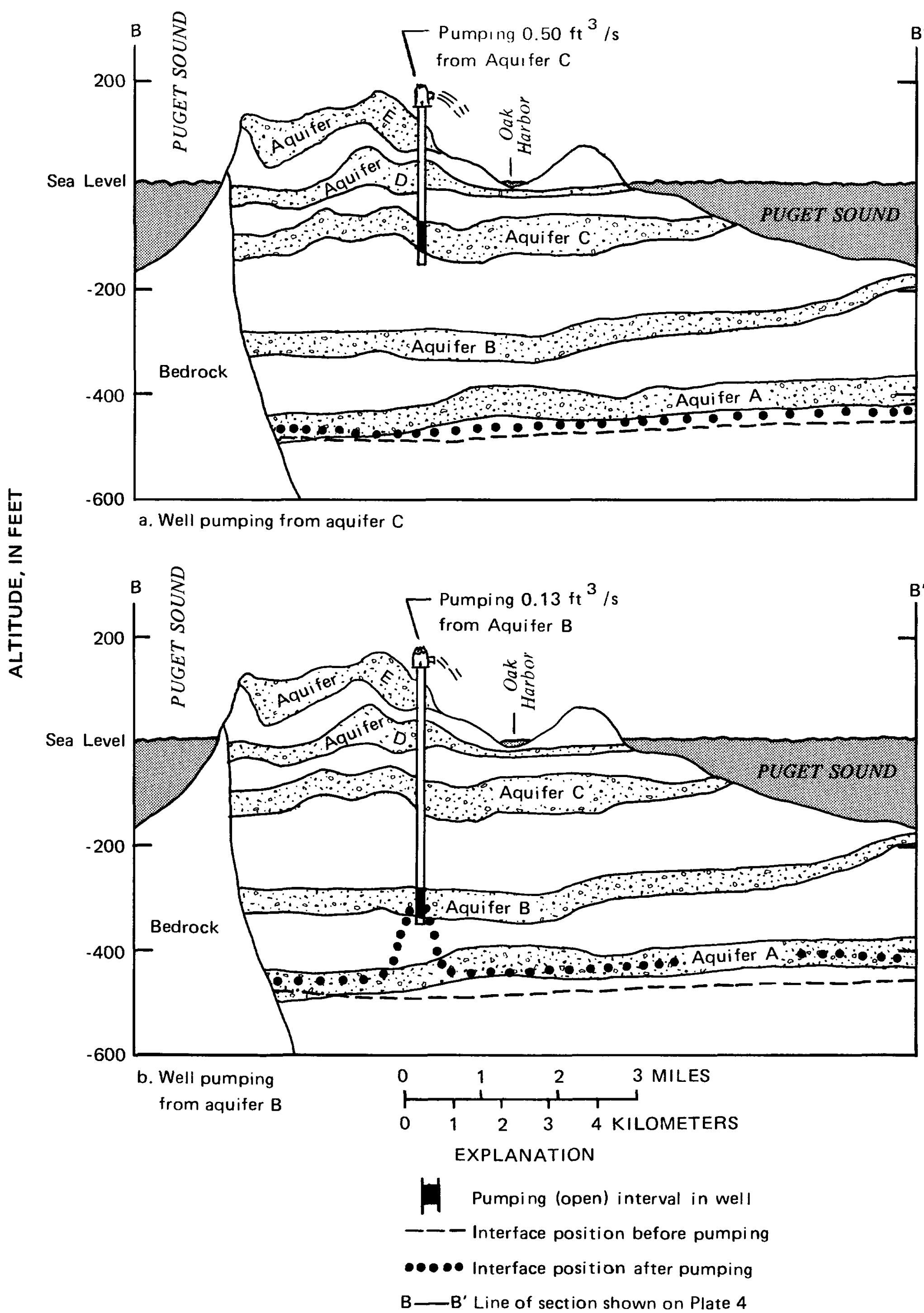


Figure 23.–Schematic hydrogeologic sections showing effects of a new well pumping from (a.) aquifer C and (b.) aquifer B.

Needs for Further Study

Seawater intrusion could increase in magnitude and become more widespread with additional ground-water development. Some of the tasks that could be performed to monitor and simulate the effects of additional pumping on seawater intrusion are outlined below:

1. Continue measurements of water levels and sampling for chloride concentrations in April and August of each year for selected wells drilled below sea level and for the piezometers in the 10 deep test holes. Wells with high chloride concentrations at the present time could be measured and sampled quarterly. Chloride and specific conductance could be determined for all samples, and complete chemical analyses could be made for selected samples.

2. Update data on ground-water pumpage every 5 years.

3. Calibrate a transient three-dimensional model that will simulate the flow of variable-density ground water, and use this model to determine the rate of seawater intrusion in response to pumping. This task would require the development of a new model and may require additional fieldwork to estimate the width of the mixing zone. Additional data required for model calibration would include coefficients of storage for aquifers and confining units and time-dependent data for (1) discharge from wells and springs, (2) recharge, (3) water-level altitudes, and (4) chloride concentrations.

The extent of high nitrate, bacteria, and other chemicals in aquifer E is unknown, and with continued use of septic-tank drainfields and landfills, bacteria and chemicals could spread throughout aquifer E and down into aquifers D and C, which presently supply most of the ground water pumped from wells. To determine if bacteria and chemical constituents other than chlorides are a problem now or could be in the future, water samples could be collected and analyzed from a dense network of wells in aquifers C, D, and E every 5 years. In between the 5-year sampling efforts, water samples could be collected from a less-dense network of wells and analyzed for the same constituents. If a problem area is detected from the sampling program, a detailed study could be conducted.

SUMMARY AND CONCLUSIONS

Ground water in Island County occurs in Quaternary glacial deposits that have a maximum thickness of about 3,000 feet. These deposits are underlain by bedrock of Tertiary age (or older) that is exposed at Deception Pass and at Rocky Point. The unconsolidated glacial materials lying above bedrock were divided into five aquifers and five confining units on the basis of available data. The thickness of the deepest aquifer (A) could not be estimated from available data and very little was known about the geology below aquifer A. However, for the ground-water model constructed in this study, aquifer A was assumed to be 50 feet thick, and a confining unit below aquifer A was extended to 900 feet below sea level.

Ground water moves from recharge areas, through the aquifers and confining units, to discharge as springs above and below sea level and to discharge from pumping wells. The lateral flow pattern in aquifers is outward from the center of each island toward Puget Sound. There may be some ground water moving from the mainland to the aquifers underlying northeast Camano Island. The vertical component of flow is downward in the central parts of the islands and upward near the coastlines and beneath Puget Sound. In confining units, the primary direction of flow is vertical.

Recharge to the ground-water system comes from precipitation, irrigation (farm and domestic), and septic-tank drainfields. Total recharge is estimated to be 186 ft^3/s.

Ground water is used primarily for household purposes, and smaller amounts are used for farm irrigation and industrial purposes. Total pumpage of ground water was estimated to be 5 ft^3/s in 1981, and most of this water was pumped from aquifers C and D. Total discharge from springs above sea level was estimated to be 6 ft^3/s. Discharge from springs below sea level could not be estimated, but a discharge of 175 ft^3/s was computed by the ground-water-flow model.

Ground-water levels were measured in wells to observe seasonal and long-term trends in response to changes in recharge and pumping. Seasonally, water levels are high in late winter and low in late summer. Long-term records of water levels do not indicate any significant trends during the period 1963-83; therefore, average annual ground-water discharge for this period is assumed equal to recharge.

The chemical composition of ground water in aquifers D and E is predominately calcium magnesium bicarbonate, and water in aquifers A and B is predominately sodium chloride or sodium bicarbonate. Ground water in aquifer C is a mixture of waters from aquifers B and D. Hardness values for sampled waters ranged from moderate to very hard.

The criteria for drinking water in Washington were exceeded for nitrate in 2 percent of the wells sampled. Criteria not related to health-risk considerations for drinking water were exceeded in 6 percent of the wells sampled for dissolved solids, 8 percent for chloride, 42 percent for iron, 60 percent for manganese, and 2 percent for sulfate.

A chloride concentration of 100 mg/L is assumed to represent ground water along the freshwater side of the mixing zone that separates freshwater and seawater. In August 1981, chloride concentrations exceeded 100 mg/L in 34 of the 143 wells (24 percent) drilled below sea level. Seawater intrusion comes from Puget Sound, which has an average chloride concentration of about 16,000 mg/L in the study area. A chloride concentration of 14,000 mg/L was found in one well, but most of the concentrations above 100 mg/L did not exceed 1,000 mg/L. Chloride concentrations were usually high (greater than 100 mg/L) in wells drilled in low-altitude areas along the coastline. In aquifer D, chloride concentrations exceeded 100 mg/L in seven wells located in the northern part of Camano Island, and at scattered locations along the coastline of central and northern Whidbey Island. In aquifer C, chloride concentrations exceeded 100 mg/L in 10 wells located in southern Camano Island and in 16 wells on Whidbey Island. In aquifer B, chloride concentrations were generally less than 100 mg/L, except for one well located near Coupeville on Whidbey Island. In piezometers that were installed in 10 test holes drilled several hundred feet below sea level during 1983-84, chloride concentration generally increased with depth, and the highest concentration was 7,200 mg/L in aquifer B beneath southern Whidbey Island.

Both seasonal and long-term trends were observed for chloride in wells. Seasonal fluctuations of chloride were most significant in pumping wells affected by seawater intrusion. Wells not affected by seawater intrusion showed little seasonal change in chloride concentrations. Long-term trends were significant in 5 of 129 wells (3.8 percent) sampled in the early 1960's and during this study.

A simulation model was developed to simulate three-dimensional steady flow of fresh ground water in a multiple-aquifer system containing freshwater and seawater. The model simulates freshwater head and the position of an assumed sharp interface between moving freshwater and stationary seawater. This sharp interface replaces the mixing zone in the real hydrologic system. One submodel was constructed for Camano Island, and Whidbey Island was divided into three submodels. Each model had 10 layers, one layer for each aquifer and confining unit. The four models were calibrated using pumping, recharge, and head data that were time-averaged for the period 1980-83, and using maximum chloride concentrations for wells sampled during 1980-83. For the calibrated models, values of horizontal hydraulic conductivity for aquifers ranged from 4.9×10^{-5} to 9.4×10^{-3}, and values of vertical hydraulic conductivity for confining layers ranged from 1.0×10^{-9} to 1.5×10^{-8} ft/s.

Computations made by the calibrated models show that most of the water recharged to the hydrologic system discharges from aquifers C and D as springs below sea level, and only a small fraction of the recharge water moves downward below aquifer C. Model computations of interface depths and hydraulic head indicate that aquifers in all areas except northeast Camano Island are not recharged by ground water moving from the mainland through aquifers beneath Puget Sound. Observed heads for northeast Camano Island indicate that aquifer D could be recharged in this area by ground water moving westward from the mainland.

The freshwater-seawater interface computed by the model exceeds 900 feet below sea level at its deepest point, beneath southern Whidbey Island. In general, the depth of the interface is greatest beneath the central parts of the islands, and the interface slopes upward near the coastlines to intersect the bottom of Puget Sound. The intersection of the interface with the bottom of Puget Sound indicates that there is no movement of fresh ground water in most of the areas separating Whidbey and Camano Islands and between the islands and the mainland, except on northeastern Camano Island.

A sensitivity analysis for the submodel covering north Whidbey Island indicated that simulated changes in recharge caused the largest changes in model-computed discharge from springs, and simulated changes in recharge and transmissivity had the greatest effect on the vertical position of the interface. To evaluate the effects of unknown geology below aquifer A, confining layer A was changed to an aquifer; this change had very little effect on the vertical position of the interface.

The calibrated model was used to examine the effects of additional pumping by adding a new well near Oak Harbor to the submodel for north Whidbey Island. Pumping at a rate of 0.50 ft^3/s from aquifer C had little effect on the interface position, but pumping at a rate of 0.13 ft^3/s from aquifer B caused the interface, previously in aquifer A, to cone upward through confining layer B into the well. Pumping this new well from either aquifer caused no noticeable lateral movement of the interface.

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Appendix I. -- Files Used for Calibration of the Ground-Water-Flow Models

This appendix contains the names of archive files that were used to calibrate each of the four ground-water-flow models described in the section "Simulation of Ground-Water Flow." The files were archived on magnetic tape at the U.S. Geological Survey's computer center in Reston, Virginia. These archived files contain FORTRAN code for the flow model, job control language, and data on hydraulic and geometric parameters for the ground-water-flow system. A listing of the FORTRAN source code for the model and instructions for preparing both job control language and model-input data are contained in documentation for the model (Sapik, 1988). The contents of all files used for calibrating the models were not included in this appendix because a complete listing would exceed 5,000 pages. The files listed in the following table are in the same sequence as the magnetic tape files, and the data groups (for example, group V data) referenced in the table are described in the model documentation.

File name	Description
Source code for the model:	
AG40XML.SSIM3D	FORTRAN source code for the model.
Files for each model:	
AG40XML.ISLANDn.JCLMOD1	Job control language.
AG40XML.ISLANDn.MODATA	Parameters for group I thru group IV data.
AG40XML.ISLANDn.MODARY	Parameters for group V data excluding the array data sets.
AG40XML.ISLANDn.HEADCALX	Starting head (group V array data set STRT)
AG40XML.ISLANDn.TRTKFINX	Transmissivity and vertical leakage coefficients (group V array data sets T and TK).
AG40XML.ISLANDn.RECHX	Recharge rates (group V array data set QRE).
AG40XML.ISLANDn.DXDYDZ	Dimensions of model cells (group V array data sets DELX, DELY, and DELZ).
AG40XML.ISLANDn.PUMPX	Well discharge rates (group V array data set WELL).
AG40XML.ISLANDn.SPGFINX	Data used to compute discharge from springs above sea level (group V array data sets IDN, DRCF, and DREL).
AG40XML.ISLANDn.RIVFINX	Data used to compute discharge from springs below sea level (group V array data sets IDR, RC, RB,and RH).
AG40XML.ISLANDn.DBOTX	Depths below sea level to bottoms of model layers (group V array data set DBOT).
Files for all models:	
AG40XML.ISLAND.MSQDATA	Records 1 and 2 for group VI data.
AG40XML.ISLAND.HEADOBS	Records 3 and 4 for group VI data.

The above data sets containing the identifier ISLANDn (n = 1, 2, 3, or 4) are repeated for each of the four models. The group V array data sets S, TXTY, and LAYRCH (described in the model documentation but not shown above) are initialized by using parameter values specified in the file AG40XML.ISLANDn.MODARY.