THERMAL ANOMALIES AND THE GROUND-WATER FLOW SYSTEM SOUTH OF THE NARROWS, UPPER SAN PEDRO VALLEY, ARIZONA

by

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A Thesis Submitted to the Faculty of the

DEPARTMENT OF HYDROLOGY AND WATER RESOURCES

In Partial Fulfillment of the Requirements
For the Degree of

MASTER OF SCIENCE
WITH A MAJOR IN HYDROLOGY

In the Graduate College

THE UNIVERSITY OF ARIZONA

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STATEMENT BY AUTHOR

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ACKNOWLEDGMENTS

This study could not have been undertaken without the help, advice and generosity of many people. I wish to thank Dr. Stanley N. Davis for his many insightful ideas and direction with this thesis. Dr. Daniel D. Evans, Dr. Eugene S. Simpson, the Department of Hydrology and Water Resources, and the Tucson office of the United States Geological Survey helped by providing their expertise and the use of their facilities and equipment. Dr. C. Brent Cluff and Dr. Mark Kartchner, M.D., and the staff of the Water Resources Research Center have been most helpful. I also wish to thank the many residents of the San Pedro Valley who were so cooperative, helpful and hospitable. Lastly, I dedicate this thesis to my parents, who have always given me their love, understanding, and support, and who shall always garner mine.

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ABSTRACT

Hydrologic conditions in the Upper San Pedro Valley are diverse and vary greatly in quantity and quality. The hydrogeologic system includes a permeable unconfined upper aquifer seperated from a lower aquifer by confining beds in the middle of the valley. Flowing wells are found in the St. David-Benson-Pomerene artesian area. Anomalously high ground-water temperatures down-gradient of this artesian area indicate a possible upward flowing portion of the confined aquifer.

Temperature and head data were collected at 31 wells, mostly in the unconfined aquifer. Thermal gradients were measured at 20 wells, and areas of elevated heat flow were found to correspond with a rise in the underlying bedrock elevation. These data tend to support the hypothesis of upward advection of heat in the lower aquifer.

CHAPTER 1

INTRODUCTION

Statement of Problem

The objective of this study was to gain a better understanding of the ground-water flow system of the Upper San Pedro basin near The Narrows of the San Pedro River. The importance of the effect of the granitic rock barrier present at The Narrows with its imposed "bottle-necking" of the ground-water flow system was of particular interest.

The importance of The Narrows became particularly evident after preliminary field investigations revealed the presence of anomalously high ground-water temperatures from certain wells in an area about two km up-gradient from The Narrows. Further analysis of the thermal anomalies coupled with selected chemical and hydraulic investigations constitutes the body of this thesis.

The remainder of this chapter details the areas studied and presents a brief review of previous hydrologic investigations in the Upper San Pedro Valley. The methodology of the work undertaken in this study is detailed in Chapter 2. The hydrogeologic system is described in Chapter 3, and its thermal regime is described in Chapter 4. Analysis of the hydrochemical data is presented in Chapter 5, and Chapter 6 reviews and concludes the thesis.

Location and Drainage of the Study Areas

This thesis covers two study areas. The regional study area covers an area of about 1000 km^2 from just south of St. David to north of The Narrows (Figure 1). Major geologic, hydrologic and topographic features are most easily studied on this relatively small scale. The local study area covers about 18 km^2 within the area bounded by the regional study area (Figure 2). The larger scale allows a more detailed analysis of the anomalous behavior exhibited by the hydrogeologic system near The Narrows, as well as a comparison of how the system behaves at two different scales.

The Narrows of the San Pedro River is defined by the Water Resources Division of the United States Geological Survey as the boundary between the Upper and Lower San Pedro Basins. The northward-flowing river drains 6470 km^2 above The Narrows, of which about 1680 km^2 are in Mexico. The regional study area is bounded on the west by the Whetstone and Rincon Mountains, and on the north and east by the Little Dragoon Mountains. Numerous washes drain into the San Pedro within the regional study area, noteably the Dragoon Wash and the Tres Alamos Wash, which drain the Little Dragoon Mountains, and Cornfield Canyon, which drains the eastern slopes of the Rincon and Whetstone Mountains. The northern part of the local study area contains what Montgomery (1963) described as a barrier rock of Precambrian granite at The Narrows. Population centers in the regional study area are Benson, St. David, and Pomerine; the valley is sparsely settled elsewhere. No towns lie within the local study area, although approximately 70 people currently live there.

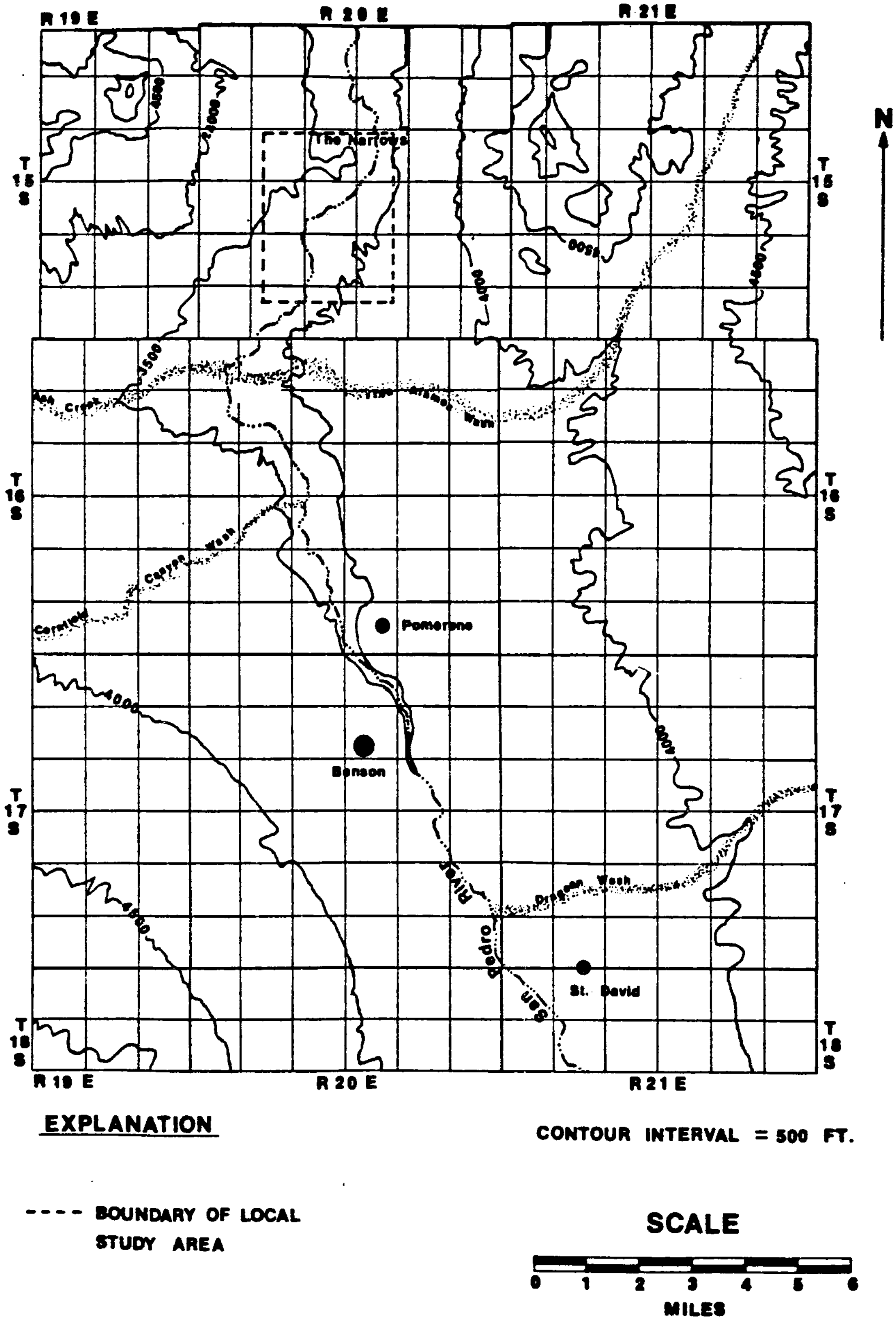


Figure 1. Location of the regional study area.

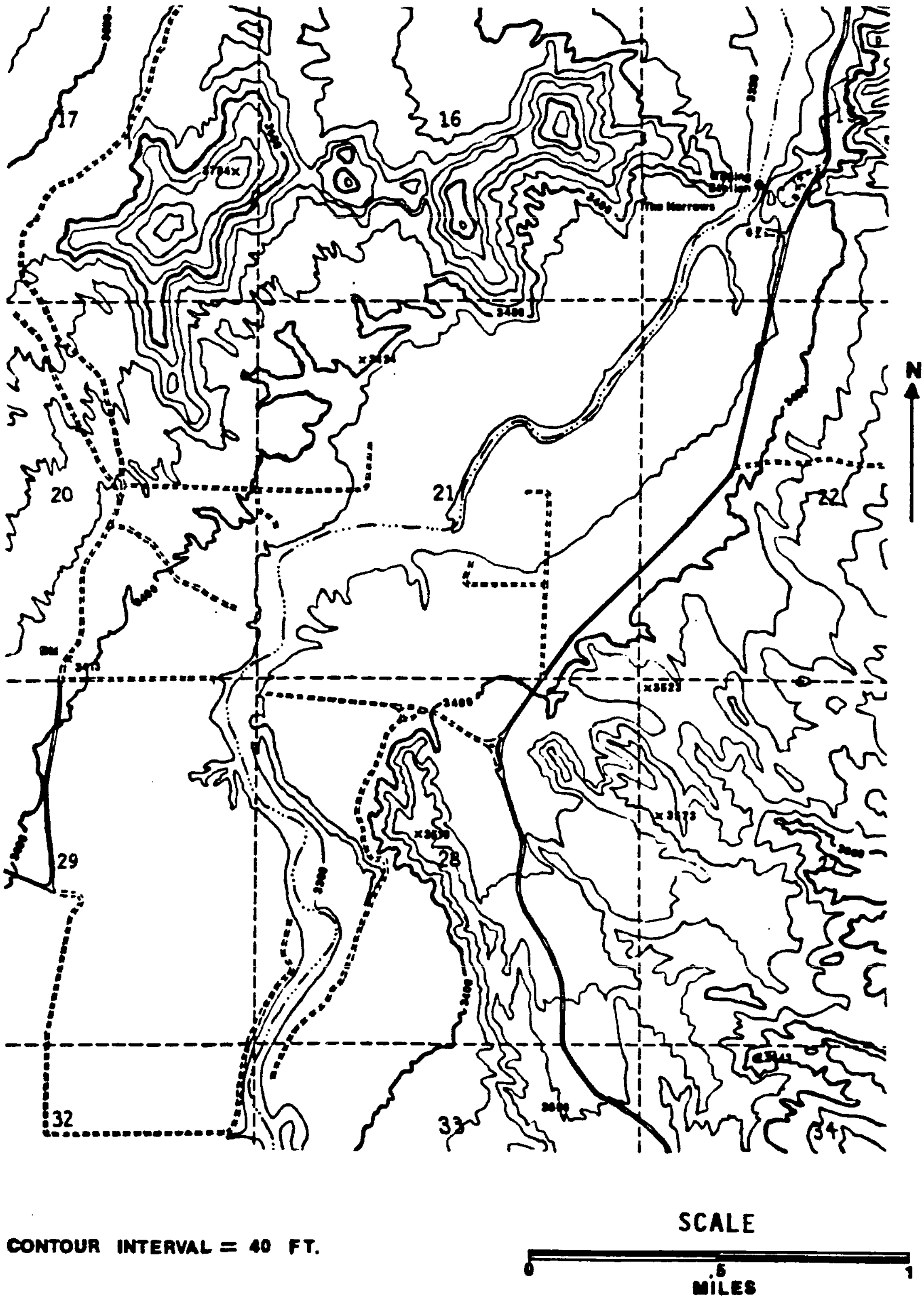


Figure 2. Location of the local study area.

Previous Studies

The first written report on hydrologic conditions in the San Pedro Valley was by Lee (1905) who described the early development of artesian ground water in the Benson-St. David area, which began shortly after a severe earthquake near Cananea, Mexico in 1887. The earthquake opened a fissure in the ground from which water flowed for several hours, which led to the supposition that artesian conditions existed in the area. Bryan, Smith, and Waring (1934) investigated the water resources of the entire San Pedro Valley in an unpublished open-file report. Heindl (in Halpenny, 1952) reported on the geology, hydrology, and water quality of the San Pedro Valley, including much general information about the hydrogeology of the upper basin. Montgomery (1963) studied the geology and ground water of an area around and north of The Narrows. Roeske and Werrell (1973) published a comprehensive report on hydrologic conditions in the San Pedro Valley, including maps of irrigated acreage and depth-to-water. More hydrogeologic maps of the Upper San Pedro Basin were prepared by Konieczki (1980). Freethey (1982) conducted a hydrologic analysis of the southern part of the Upper San Pedro Valley using a finite differnce model, and was able to simulate pre-development ground-water levels in the modeled area. Halverson and Sumner (1983) conducted a gravity survey and prepared depth to bedrock maps for the San Pedro Valley. Usunoff (1984) investigated the water quality of an area from St. David to The Narrows, with special emphasis on fluoride in the lower aquifer.

Well Identification System

As each well was visited, it was assigned an identification number; the first well visited was well 1, the last was well 25. When two or more wells were owned by the same owner, each was identified by the same number followed with a letter suffix. For example, wells 10a and 10b are owned by the same person but are on different parts of the owner's property and different data were collected from each. Appendix A contains a data summary table which includes the location of each well visited, as defined by the USGS location system. This location system divides the state of Arizona into four quadrants, designated A, B, C, and D counterclockwise about the confluence of the Gila and Salt Rivers, with A the northeast quadrant (Figure 3). The first digit identifies the township, the second the range, and the third digit indicates the section that contains the well. The following three letters locate the well within the section. The first letter denotes a 160-acre tract, the second a 40-acre tract, and the third a 10-acre tract. These letters are also assigned in a counterclockwise fashion, with a in the northeast. In the example shown in figure 3, the location (D-4-5)19caa designates that the well is in the northeast quarter of the northeast quarter of the southwest quarter of section 19, township 4 south, range 5 east. Where more than one well is present in a 10-acre tract, consecutive numbers, starting with 1, are added as suffixes.

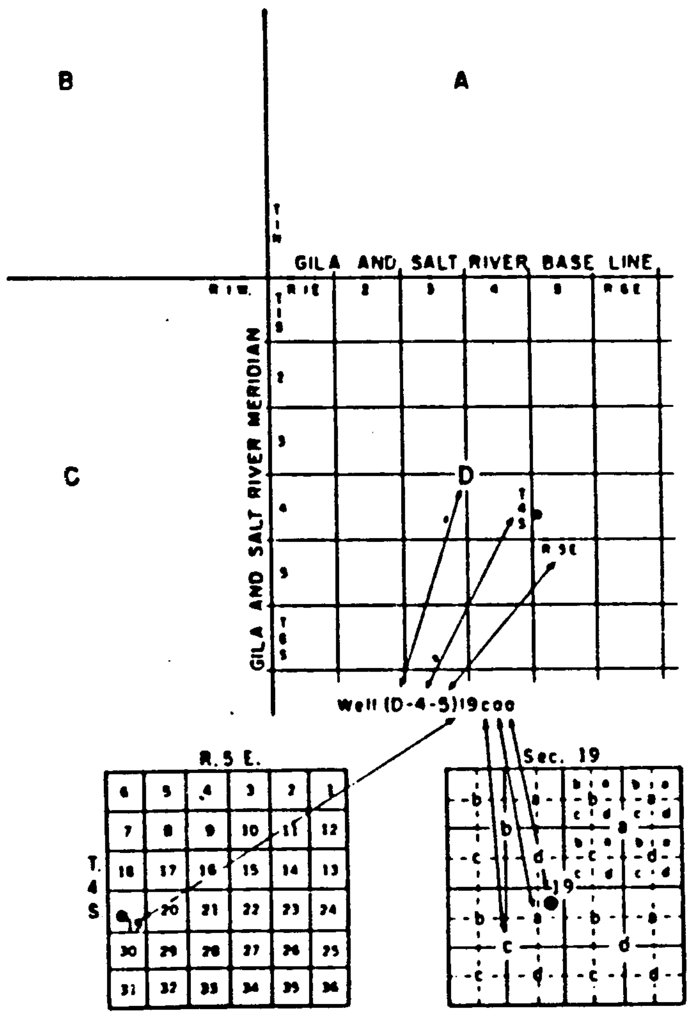


Figure 3. Well-location system used. (From Roeske and Werrell, 1973.)

CHAPTER 2

METHODOLOGY

In this study, a ground-water flow system is investigated from a hydrogeological standpoint, with the aid of thermal and chemical analysis. This thesis represents an outgrowth of a previous, less detailed hydrologic study. During the course of that study, a well south of The Narrows was found to have a temperature near 30 ^{o}C at a depth of less than 65 m. Other wells in the area are typically almost 10 ^{o}C cooler at similar depths. The discovery of this anomalous water temperature stimulated further and more detailed hydrogeologic study of the area just south of The Narrows.

The Hypothesis Explaining the Thermal Anomaly

Several explanations are possible for the observed thermal anomaly. A local magmatic body could be heating the ground water in that area. A fracture zone or a lithology change in the confining layer could allow advection of deeper, warmer water to a shallower depth. The well could be tapping a shallow portion of a deeper aquifer where ground water is flowing in an upward direction. The above possible explanations of the anomalous thermal behavior observed are addressed in Chapter 6.

After examining the physical boundaries of the ground-water flow system in the vicinity of The Narrows, it was hypothesized that the granitic rocks that crop out at The Narrows, assumed impermeable, provide a barrier to ground-water flow at depth. Anomalously high temperatures in ground water up-gradient of the barrier could thus be explained by possible upward flow of deep, warmer ground water to a shallower level. Due to the proximity of the warm well to The Narrows, it was also hypothesized that if underflow of the lower aquifer takes place only at The Narrows, then subsurface temperatures would increase in the direction of The Narrows due to advection of the deeper ground water. The increased temperature in the subsurface would be reflected in the upper aquifer, which would be heated by conduction through the confining layers, and possibly by advection through permeable portions of the confining layers.

Test of Hypothesis

Field work was required to test the hypothesis. Temperature in the subsurface was measured at various wells in both the regional and local study areas. Because the local study area just south (up-gradient) of The Narrows contains numerous wells that tap the upper aquifer, temperature profiles of several wells could be obtained. An analysis of the data showing a rise in ground-water temperature of the upper aquifer might indicate increased heat flow in the subsurface. Furthermore, chemical analysis of ground water, under favorable circumstances, could be used to determine if advection between aquifers is substantial. Knowledge of hydraulic head was essential to the determination of direction of leakage between aquifers, so depth-to water data and well elevations were recorded where possible.

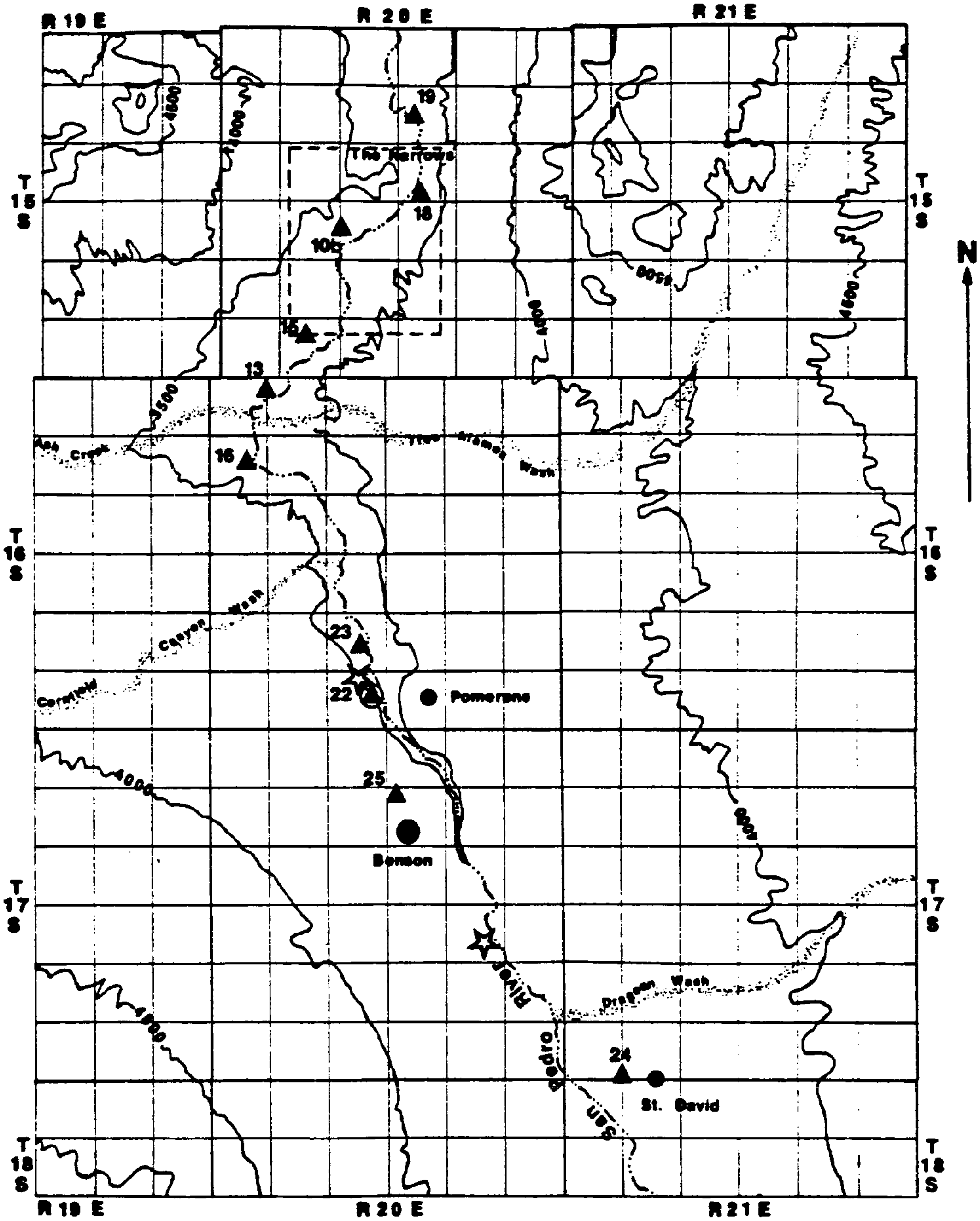
Location of Wells

Figure 4 shows the location of the wells visited in the regional study area outside of the local study area. Figure 5 shows the location of wells visited within the boundaries of the local study area. Appendix A contains a data summary of the wells visited and includes location, depth and diameter of the wells, as well as the type of data collected from each well.

Collection of Temperature Data

Temperature data were collected in the field using a waterproofed thermister on a 150 m reel of shielded cable. A Polycorder, loaned by the USGS, gave a direct digital readout of temperature in degrees Celsius. This equipment is precise to at least 0.01 ^{o}C, and probably to 0.005 ^{o}C at any particular location; its absolute accuracy is believed to be better than 1.0 ^{o}C. The probe was initially immersed in an ice bath, where it read a temperature of 0.1 ^{o}C.

Temperatures in wells were measured at intervals of approximately 3 m wherever possible. In order to obtain a representative thermal profile of a well, it was necessary to make certain the well had not been pumped for at least one day prior to testing, so that thermal conditions would be near equilibrium in the borehole. Some wells visited had no access for the thermister. Such wells were either not tested for temperature, or were allowed to fill a basin with water, from which temperature was measured. Temperatures thus recorded are thought to be indicative of the temperature of the aquifer over the water-bearing interval, and do not necessarily represent the mean temperature in the



EXPLANATION

---- BOUNDARY OF LOCAL STUDY AREA

○ UPPER AQUIFER PUMP TEST

☆ LOWER AQUIFER PUMP TEST

CONTOUR INTERVAL = 500 FT.

SCALE

0 1 2 3 4 5 6

MILES

Figure 4. Location of wells visited in the regional study area.

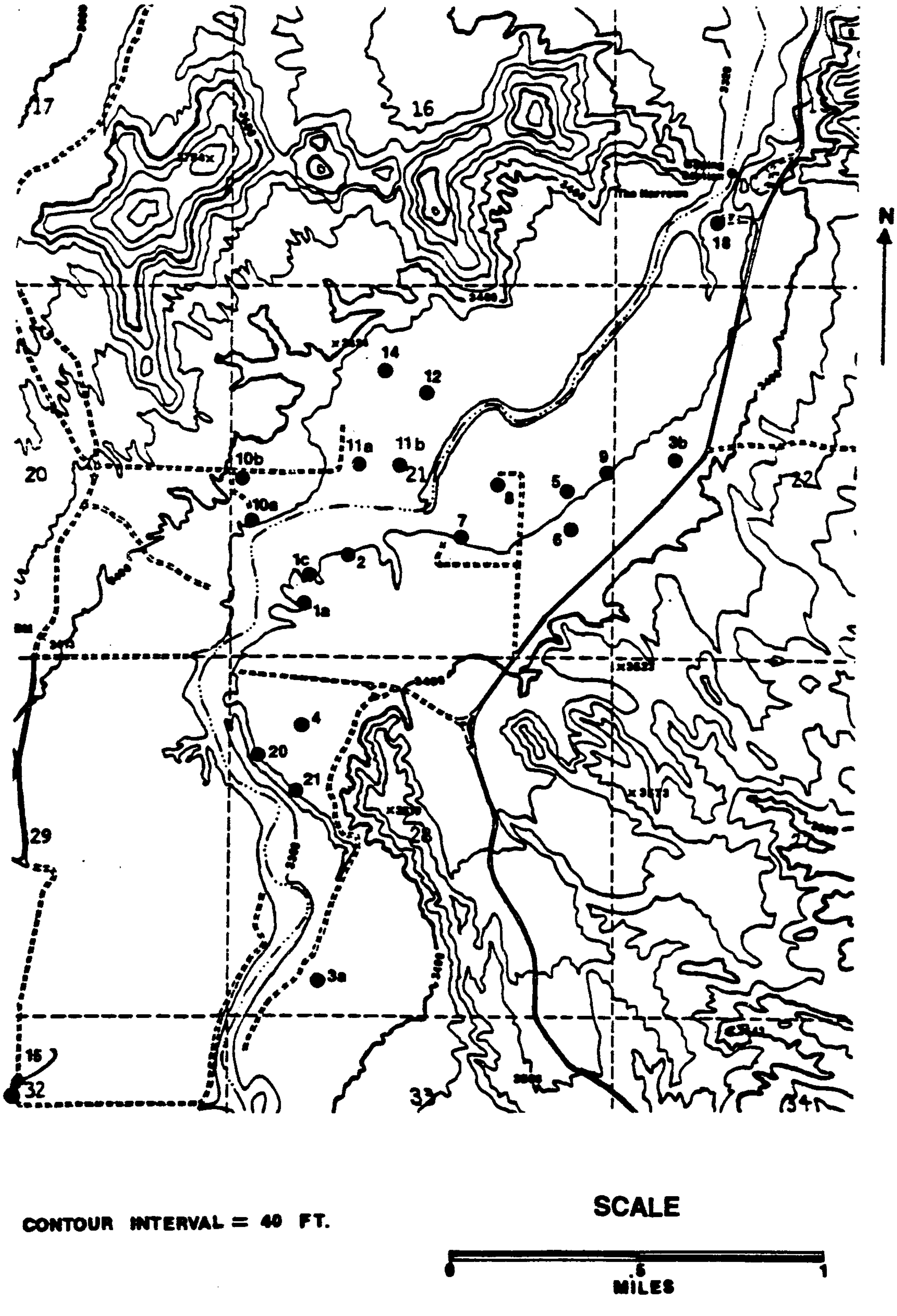


Figure 5. Location of wells visited in the local study area.

penetrated formation. Raw temperature data are included in Appendix A. Graphs of depth versus temperature are presented in Appendix B. For each one of these graphs, the data were fitted by the least squares method to obtain values of average geothermal gradient.

Collection of Hydraulic Head Data

In order to determine the hydraulic head in the wells visited, depth-to-water had to be measured and elevation of the top of the casing determined. The depth-to-water was measured directly when it was possible to drop a weighted measuring tape into the well. Determination of the elevation was more difficult. An altimeter was used to determine elevation differences between successive wells visited, and between a benchmark and wells. However, accessible benchmarks in the area are sparse, and weather conditions proved too unstable to rely on barometric readings from the altimeter. Thus, elevation was estimated in most cases from a topographic map having a 20 ft contour interval, so a fairly large (2-4 m) error can be expected in the head data. These data are also included in Appendix A. Contour maps of the water table are in Appendix C.

Chemical Sampling and Analysis

Water samples were taken from 15 wells in the regional and local study areas. Samples were stored in plastic bottles until they were analyzed using the Hach Chemical Co. Model DR-EL/4 testing kit. Samples were tested for sulfate, nitrate, and silica, as well as pH and specific conductivity. These data are presented in Chapter 5.

CHAPTER 3

HYDROGEOLOGY

The hydrogeologic system within the study area includes unconfined and confined alluvial aquifers that receive recharge from two chemically distinct sources. Thus, the two aquifers may be readily differentiated on the basis of water chemistry. Also, the aquifers are subject to two different thermal regimes. Analysis of the thermal regimes, as partly revealed by anomalously high temperatures in wells, can lead to inferences as to the behavior of certain components of the hydrogeologic system.

Introduction

The general hydrogeology of the regional study areas is illustrated by a generalized geologic cross section in Figure 6, which shows 4 hypothetical wells (Heindl in Halpenny, 1952). The upper aquifer consists of recent floodplain deposits along the San Pedro River. Well A taps this aquifer under water table conditions. Wells B, C, and D penetrate confining layers to tap the lower aquifer. Well B penetrates several interbedded sand and clay units under confined conditions. Well C is drilled mainly through clay, and encounters the confined lower aquifer at depth. The land surface is below the piezometric surface there, so well C flows. Well D is drilled in coarser sediments on the valley flank, and encounters ground water below a thin clay bed.

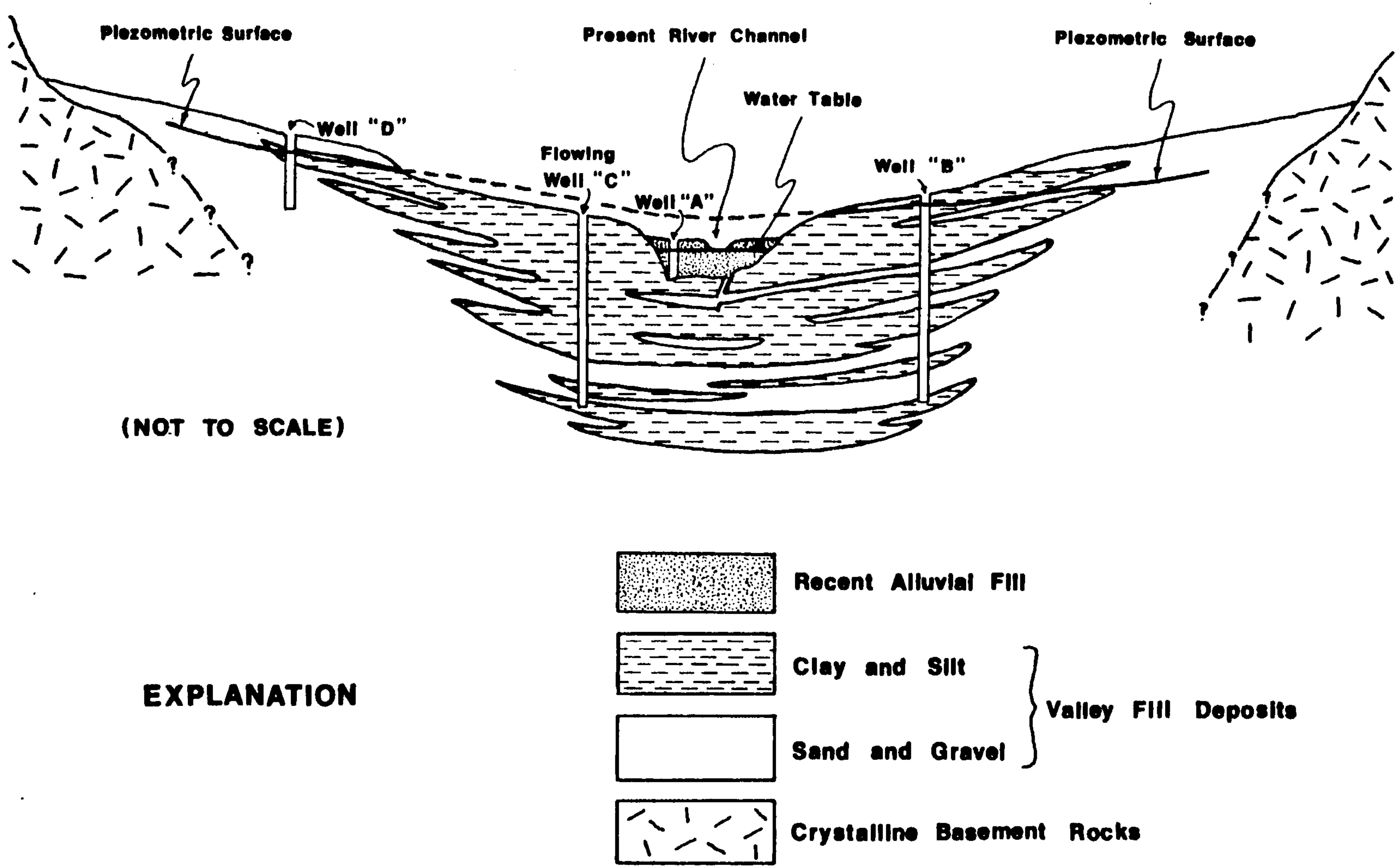


Figure 6. Generalized transverse geologic cross section of the San Pedro Basin near Benson. (After Heindl, in Halpenny, 1952.)

Upper Aquifer

The upper, unconfined aquifer provides most of the water used for irrigation in the study area. It also provides domestic water to most residents who live outside the St. David-Benson-Pomerine artesian area. This aquifer yields water that is generally of poorer quality than the lower, confined aquifer.

Geology of the Upper Aquifer

The upper aquifer is the youngest geologic formation in the San Pedro Valley. The Recent alluvial fill is a unit which occupies channels incised into older rock units, generally the underlying older valley fill deposits (Heindl, 1952). These deposits usually are 30 to 40 m thick near the river, and pinch-out within a transverse distance of 3 km from the San Pedro River (Figure 7). Sediments of this unit are flat-bedded, range in size from clay to boulders, and contain lenses of predominantly sand-sized material varying widely in thickness (Montgomery, 1963). The Recent alluvium is coarsest at depth and along the river and becomes finer-grained as the unit pinches out.

Boundaries of the Upper Aquifer

The upper aquifer is unconfined, therefore, its upper boundary is, by definition, the water table (Freeze and Cherry, 1979). Its lower boundary is the confining clay bed that makes up the top of the valley fill deposits. Drillers in the area typically stop drilling in the Recent alluvium when they encounter a characteristic red clay underlying permeable sand and gravel at depths of 25 to 40 meters. Lateral boundaries of the upper aquifer are believed to be no-flow boundaries, as the

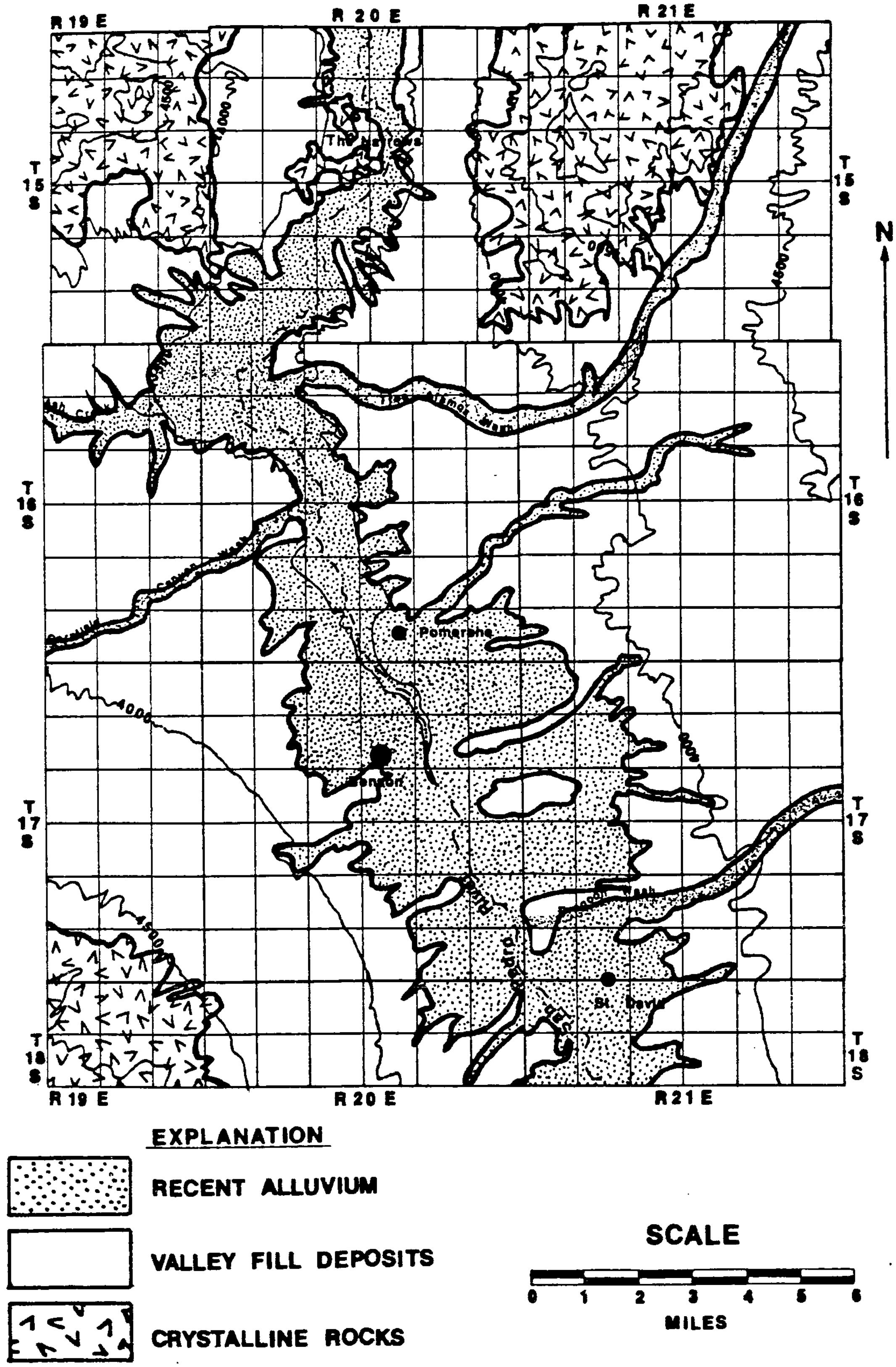


Figure 7. Major geologic units in the regional study area. (After Montgomery, 1963, and Roeske and Werrell, 1973)

Recent alluvial fill is also in contact with the relatively impermeable silt and clay beds that make up the upper unit of the valley fill deposits. The Recent alluvial fill at The Narrows is pinched to a lateral width of about 100 m by the granitic rocks that outcrop there (Figure 8). Additionally, a constant head boundary is imposed on the aquifer directly below the wetted river channel when the San Pedro River flows.

Hydrologic Stresses on the Upper Aquifer

The unconfined aquifer receives nearly all its recharge through the channel of the San Pedro River. Flux of ground water between the confined and unconfined aquifers is considered negligible in most places in the study area because of the low permeability of the clay beds that generally underlie the Recent alluvium. Discharges from the unconfined aquifer are to wells, phreatophyte transpiration, and to the river channel.

Recharge to the Upper Aquifer. The San Pedro River flows mainly in direct response to precipitation, which aversges about 28 cm/yr in the Benson area (Green and Sellers, 1964). The river has low flow during much of the year, and usually has zero flow during May, June, and early July. A reach of the river in T15S R20E-32 is perennial, probably due to a rise in the clay layer that forms the lower boundary of the unconfined aquifer. This phenomenon effectively decreases the thickness of the aquifer and causes the water table to intersect the bottom of the river channel. Several other small perennial reaches are present. The water table seldom lies more than one meter below the riverbed,

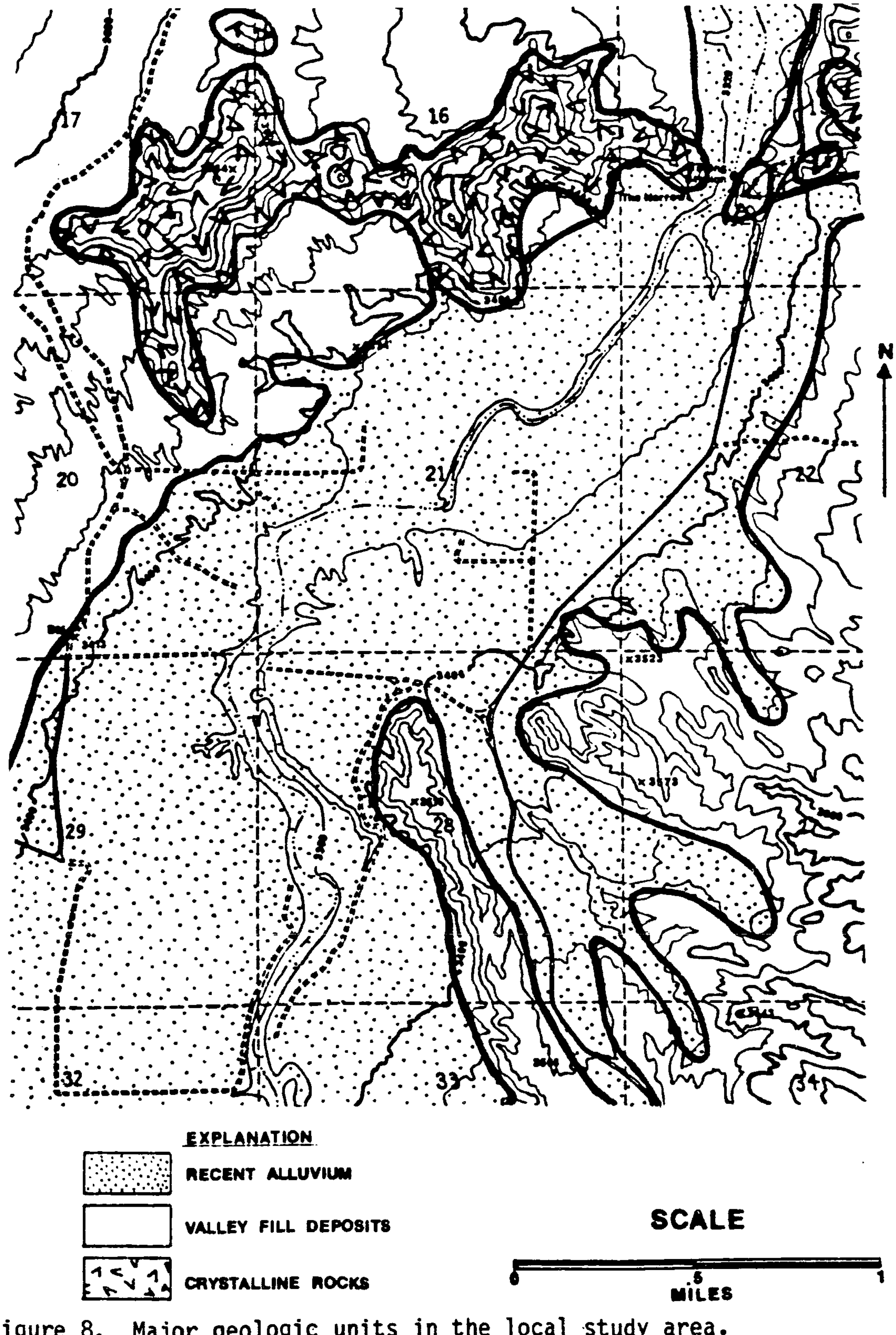


Figure 8. Major geologic units in the local study area. (After Montgomery, 1963, and Roeske and Werrell, 1973.)

which is often wetted by the capillary fringe immediately above the water table. Because of the shallow depth to water and the high permeability of the river channel, recharge from the river to the unconfined aquifer is rapid and substantial during periods of moderate to high flow. Irrigators in the area report well-water levels rising over one meter in a day in response to the first flow event in the river after a dry period. Recharge from washes is also substantial when they flow, which, however, is seldom. Well hydrographs show no long-term decline of the water table in the area, indicating that recharge has been able to replenish all ground water withdrawn from the unconfined aquifer thus far.

Although recharge to the upper aquifer is mainly through the river channel and adjacent washes, some water from agricultural activities percolates through the soil and reaches the water table. Representing as much as 15 to 30% of the water withdrawn from the aquifer and applied to crops, this recharge water is high in salts and dissolved solids as a result of leaching fertilizers, pesticides and animal wastes.

Discharges from the Upper Aquifer. Before the exploitation of the water resources in the study area, the natural discharges from the upper aquifer included phreatophyte transpiration and discharge to the river through the channel bottom. No springs, other than the perennial and gaining reaches of the San Pedro River, are observed in the upper aquifer in the regional or local study area. Phreatophyte transpiration, however, occurs much more rapidly during the summer months. With the increased phreatophyte transpiration in the summer, vegetated areas became areas of discharge, while the riverbed became an area of recharge. It is uncertain whether the transpiration rate was or is in excess of the recharge rate during the summer.

Development of agriculture in the valley has imposed increased stresses on the hydrologic system. Pumping demands on the upper aquifer are greatest during the summer growing season. Irrigators typically apply 60 cm of water to pasture lands and 90 cm to alfalfa, the two most common crops in the area. Irrigated acreage in the San Pedro Valley was 5100 hectare (12,500 acres) in 1966 (Roeske and Werrell, 1973). The ratio of irrigated acreage in the regional study area to that of the San Pedro Valley was calculated to be 0.34 using planimetry and maps from Roeske and Werrell's (1973) report which show the location and extent of irrigated areas in the San Pedro Valley as of 1966. This ratio is assumed constant in time. Using these data and assuming an average of 75 cm of water was applied to the crops annually, ground-water pumping in the regional study area accounted for about 13 million cubic meters in 1966. The trend of pumping over time in the upper basin is assumed to be indicative of trends in the study area. Using these assumptions, Figure 9 details the ground-water pumping as a function of time.

Water-Bearing Characteristics of the Upper Aquifer

Well Yields. The unconfined upper aquifer is the most permeable unit in the study area. Nearly all agricultural pumping is done in the upper aquifer, with most wells near the river, where the sediments are coarsest and thickest. These wells yield up to 125 L/sec (2000 gpm), although most irrigation wells produce between 30 and 75 L/sec (500 and 1200 gpm). Saturated thicknesses in these wells range from 10 to 30 m.

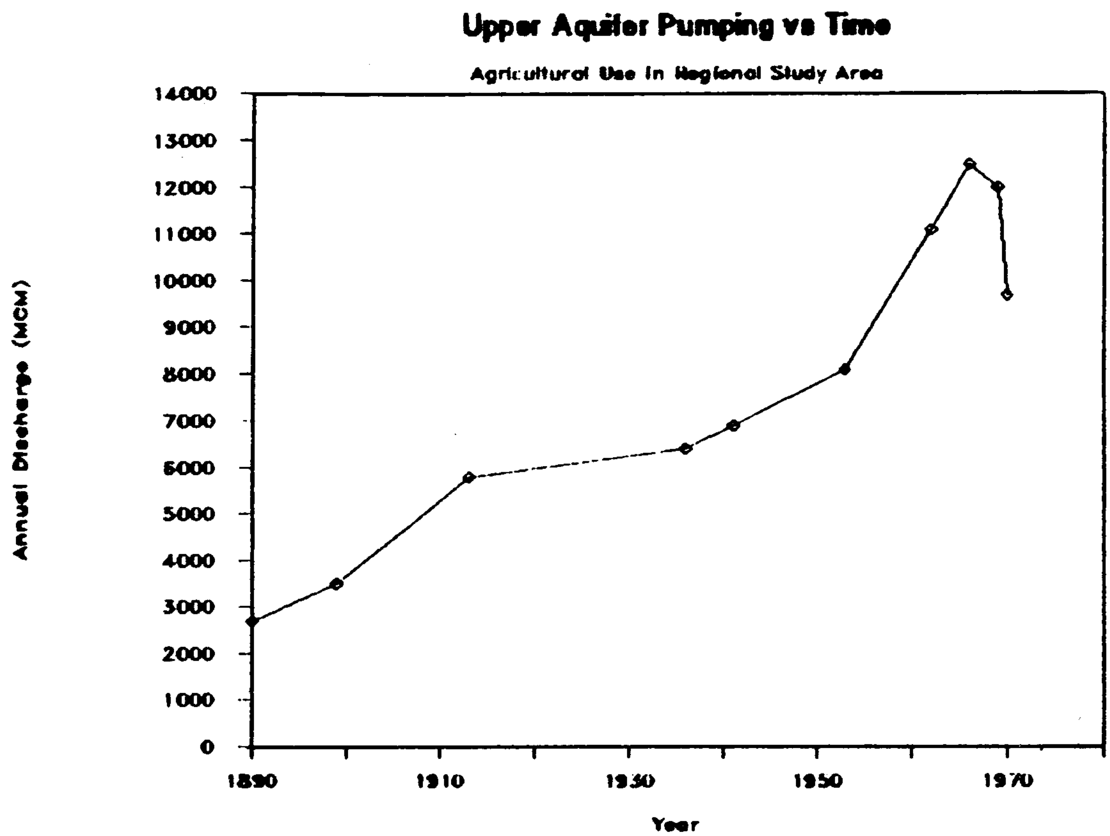


Figure 9. Ground-water withdrawals from the upper aquifer in the regional study area as a function of time.

Hydraulic Properties of the Upper Aquifer. An aquifer test performed by the author in October of 1985 found the transmissivity to be 1,277 m^2/day. The tested well is located near the west bank of the San Pedro River at (D-16-20)33adc, and fully penetrates the upper aquifer. The saturated thickness at the tested well is about 28 m, thus the calculated hydraulic conductivity is 46 m/day. The tested well was discharging at a rate of 32.5 L/sec, and an observation well at a distance of 2.6 m was drawn down 0.484 m after 3 hours of pumping. Water was discharged into the San Pedro River about 70 m from the pumped well, and may have affected the aquifer test by imposing a recharge boundary, in which case the calculated transmissivity may be high. Montgomery (1963) calculated transmissivities of other wells in the area by multiplying specific capacity (gpm/ft), minus well losses, by 2,000 to obtain transmissivity in units of gpd/ft. This approximation is also used in this paper. Numerical values thus obtained were 450 and 1120 m^2/day with corresponding hydraulic conductivities of about 41 and 82 m/day. Heindl (in Halpenny, 1952) reported that hydraulic conductivities in the upper aquifer range between about 40 and 200 m/day. Specific yield values are not available in the regional study area, but Freethey (1982) used values ranging from 0.05 to 0.18 for a numerical model of the unconfined upper valley fill sediments near Sierra Vista.

Due to the well-bedded, alluvial nature of the upper aquifer, permeability is assumed to be isotropic in the horizontal plane, with lower permeability in a vertical direction.

Lower Aquifer

The lower aquifer, which is confined in most of the regional study area and all of the local study area, accounts for the domestic water supplies for the towns of St. David, Benson, and Pomerine. Numerous flowing wells tap the lower aquifer at depths from 100 to 400 meters in the St. David-Pomerine artesian area (Figure 6). The chemical quality of water from this aquifer is generally better than that of the upper aquifer.

Geology of the Lower Aquifer

Water in the lower aquifer flows in an alluvial formation which Roeske and Werrell (1973) referred to as the valley fill deposits. These deposits rest on an igneous and metamorphic basement complex of Precambrian age. Halverson's (1984) gravity survey of the San Pedro Valley provided a map of the depth to bedrock, defined as rock having a density greater than 2.67 gm/cm^3 (Figure 10). The depth to this basement rock is as much as 1000 m near Benson, and less than 250 m near The Narrows, where the bedrock forms an east-west trending ridge across the valley beneath the overlying alluvium.

The valley fill formation above the bedrock stores most of the ground water in the Upper San Pedro Valley. It consists of a permeable lower unit which comprises the lower aquifer, and a fine-grained upper unit that acts as a confining bed near the center of the valley. Both units are coarse grained and poorly sorted near the mountains, and become increasingly fine grained and sorted toward the axis of the valley. The lower portion of the valley fill grades from a conglomeritic facies near the mountains to a sand in the center of the valley. The upper unit grades from a poorly sorted silt and gravel near the mountains to confining beds of clay and silt ranging in thickness from 80m to more than 300 m near the center of the valley. The beds generally dip from 10 to 15 ^{o} toward the center of the valley, where they are mostly horizontal. Heindl (<u>in</u> Halpenny, 1952) illustrated the lenticular ind interfingering nature of the beds that make up the lower aquifer and the confining clay unit in the St. David-Pomerine artesian area (Figure 6).

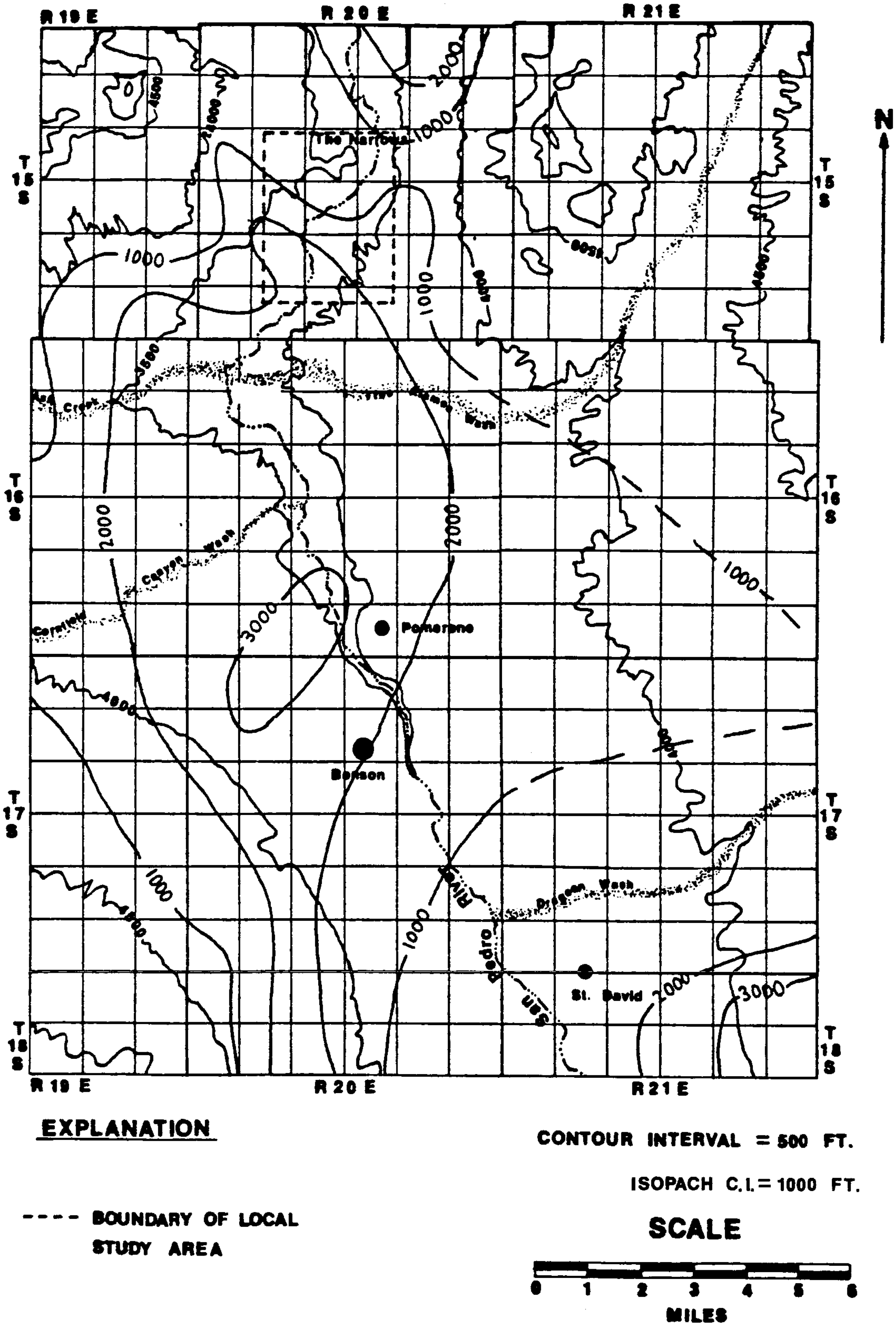


Figure 10. Depth-to-bedrock map of the regional study area. (After Halverson, 1984.)

Boundaries of the Lower Aquifer

The lower aquifer is both confined and unconfined in the regional study area; it is completely confined throughout the local study area. The confining beds of clay and silt are thickest near the center of the valley. As Figure 6 illustrates, the confining layers pinch-out towards the mountains, and the valley fill deposits are an unconfined aquifer on the high piedmont slopes of the valley margins. The confining beds are assumed nearly impermeable in most areas. The lower boundary is the igneous-metamorphic basement complex, which is also assumed impermeable. The lateral boundaries of the lower aquifer are the basement rocks at the basin margins and the interfingering clay and silt deposits of the upper valley fill sequence (Figure 6). An important boundary on the lower aquifer's ground-water flow system is the igneous rocks at The Narrows. The presence of a large outcrop of massive granite at The Narrows (Figure 8) indicates a possible obstacle to underflow at depth; analysis of the depth-to bedrock map (Figure 10) suggests that ground water in the lower aquifer could flow around the rock barrier, both under The Narrows and to the west of the granitic outcrop.

Montgomery (1963) suggested that the lower portion of the confining unit may be substantially coarser and more permeable near The Narrows due to increased stream velocity in the ancient San Pedro River around the granitic barrier. He also suggested that the basement rocks may be highly permeable near The Narrows due to large fracture zones associated with the block faulting that formed the Basin and Range Province. He postulated that both possibilities could result in ground water flowing through The Narrows within the confined aquifer, which would explain the occurrence of confined ground-water north of The Narrows that he documented.

Hydrologic Stresses on the Lower Aquifer

Recharge to the lower aquifer takes place through the coarse sediments along mountain fronts, and is of better quality than the San Pedro River water that recharges the upper aquifer. Discharge from the lower aquifer in the study area is to wells and springs and slow upward leakage through the confining beds in areas where the piezometric surface of the lower aquifer is above the water table of the upper aquifer.

Recharge to the Lower Aquifer. Mountain front recharge is the primary mechanism for recharge to the lower aquifer. The mountains in the area receive more precipitation than the valley does, and much of the runoff from this precipitation infiltrates through the washes and coarse sediments of the valley fill deposits at the base of the mountains. Some recharge takes place by direct infiltration of water through the piedmont slope of the valley, although the amount is considered negligible in comparison to mountain front recharge. Exchange of ground water between the upper and lower aquifers is assumed to be small in the regional study area due to the low permeability of the confining clay layers that characteristically separates the two aquifers and to the small head gradient usually present across the confining layer.

Discharges from the Lower Aquifer. The primary mechanism of ground-water discharge from the lower aquifer in the area today is through wells. All municipalities in the area obtain their water from the lower aquifer. Many small domestic, irrigation and stock wells tap the artesian strata and are allowed to flow year-round, and have been flowing for decades. Artesian wells in the area generally flow at a lower rate in the summer when demand is greatest. When two flowing wells near (D-17-20)36bb were first pumped, several wells up-gradient stopped flowing. Discharge from the lower aquifer to phreatophyte transpiration does not take place in either of the study areas. Some springs present in the regional study area discharge small amounts of ground water from the lower aquifer along highly permeable zones.

Water Bearing Characteristics of the Lower Aquifer

Well Yields. Wells in the Upper San Pedro Valley that tap the lower aquifer will yield from 6 to 175 L/sec (100 to 2800 gpm) and average 37 L/sec (590 gpm) according to data published by Roeske and Werrell (1973). In the artesian area, however, many of the flowing wells are never pumped, and values obtained from recovery tests of artesian strata indicate that some flowing wells may not be capable of yielding more than about 3 L/sec (50 gpm). Heindl (in Halpenny, 1952) reported that flowing wells in the Benson-St. David area discharged an average of 0.38 L/sec (6 gpm). Early reports of some of the first artesian wells drilled in the Benson area described flow rates as high as 12.6 L/sec (200 gpm) soon after well completion. Many flowing wells in the regional syudy area flow at less than 0.25 L/sec (4 gpm). The flowing well at (D17-20)10bbc in Benson, well 25, was flowing at a rate of 0.013 L/sec (0.2 gpm) when its thermal profile was measured in October, 1986.

Hydraulic Properties of the Lower Aquifer. Two aquifer tests on flowing wells in the regional study area were conducted by the author in October 1985 and January 1986 at (D-17-20)23dbd south of Benson and at (D-16-20)33adb north of Benson (Figure 4). Transmissivities of the tested wells were computed to be 4.2 and 13.3 m^2/day, respectively. Transmissivity was calculated from data obtained by observing recovery of a flowing well that was connected to a mercury-filled U-tube manometer.

Other values of transmissivity in the lower aquifer include Montgomery's (1963) calculation of 223 m^2/day based on a specific capacity of 9.0 gpm/ft at a well penetrating the confined aquifer north of The Narrows at (D-14-20)34dbc. Roeske and Werrell (1973) gave a range for specific capacity of wells in the lower aquifer in the Upper San Pedro Valley of from 1 to 40 gpm/ft (0.002 to 0.08 cm^2/s), with an average value of 13 gpm/ft (0.03 cm^2/s). Relating these to transmissivity yields values between 25 and 1000 m^2/day with an average of 325 m^2/day. A coefficient of storage for the lower aquifer has not been determined within the regional study area; however, Freethey (1982) used coefficient of storage values near 10^{-3} in his numerical model of the San Pedro Basin near Sierra Vista.

Movement of Ground Water

The general pattern of ground-water flow in the San Pedro basin is from the mountain fronts toward the axis of the valley, then northward in the direction of the San Pedro River. Appendix C contains contour plots of water table and piezometric surface elevations in the regional study area compiled from various sources, including the water-table mapping performed by the author in the summers of 1985 and 1986. Because the upper aquifer does not extend very far laterally, contours of the water table generally indicate a flow of ground water along the axis of the valley in the direction of the river. Contour bends point gently upstrean, indicating that the San Pedro River is a losing stream in most reaches in the regional study area. No noticeable depressions have appeared in either the piezometric surface or the water table, although piezometric surface data is scarce and the long-term drop in discharge from flowing wells in the Benson-St. David area indicates that withdrawls from artesian strata currently are slightly in excess of recharge. The hydraulic gradient in the upper aquifer generally ranges between 4.0×10^{-3} and 6.0×10^{-3}.

CHAPTER 4

TEMPERATURE AND HEAT FLOW

The observation of anomalous thermal behavior in well 10b at (D-15-20)21cbb constituted the basis for further investigation of the thermal regime of the subsurface near The Narrows. Thermal gradients were measured in 20 different wells in the area, and resultant data were analyzed. Raw field data are reproduced in Appendix A, and graphs of well-depth versus temperature are presented in Appendix B.

Introduction

The three mechanisms of heat transfer are radiation, conduction, and advection or convection. The terms convection and advection are often used synonymously to describe the transport of heat by mass flow, however, convection implies a cyclic flow of heat. Radiation is considered an insignificant means of heat transfer in aquifers. The interior of the earth is heated by radioactive decay, and its surface is cooler than its interior; thus, by the first law of thermodynamics, heat flows outward from the earth. The flux of heat may be approximated by measuring thermal gradients and estimating thermal conductivity. Observation and analysis of the effects of subsurface heat flow may indirectly reveal some information about hydrogeologic conditions. Areas of aquifer recharge are characterized by relatively cooler temperatures; elevated temperatures are often indicative of areas of discharge (Cartwright, 1970).

Conduction of Heat in the Subsurface

Conduction is a mode of heat transfer whereby heat propogates within a body due to thermal motion on a molecular scale. Fourier's Law governs conduction in the saturated zone, and states that the flux of heat in a homogeneous body is in the direction of and proportional to the temperature gradient:

$$\bar{q} = - k \nabla T \qquad [1]$$

which in the vertical dimension is :

$$q_z = - k \frac{dT}{dz} \qquad [2]$$

where,

q = magnitude of geothermal heat flux [cal/cm^2/sec].

k = thermal conductivity of porous medium [cal/sec/cm/^{o}C].

T = temperature [^{o}C].

z = vertical axis [cm].

The minus sign signifies that heat flows from areas of high temperature to areas of low temperature.

Advection of Heat in the Subsurface

Advection is the transport of heat by mass flow. Stallman (1960) described the simultaneous flow of heat and water through isotropic, homogeneous, fully saturated porous media with the differential equation:

$$\nabla^2 T - \frac{c_f \rho_f}{k} [\nabla(\bar{v}T)] = \frac{c\rho \partial T}{k \partial t} \qquad [3]$$

where,

c_f = specific heat of fluid [cal/cm^3°C].

ρ_f = density of fluid [gm/cm^3].

c = specific heat of solid-fluid complex [cal/cm^3°C].

ρ = density of solid-fluid complex [gm/cm^3].

$\bar{v}$ = vector velocity of fluid [cm/sec].

t = time since flow started [sec].

Sammel (1968) studied the convection of borehole fluid due to thermally induced fluid density differences. He showed that wells having diameters greater than 5 cm will be unstable under thermal gradients greater than 5.0 °C/km, but that temperature oscillation amplitudes will be only a few hundredths of a °C for geothermal gradients between 10 and 100 °C/km.

Factors Controlling Heat Flow in the Subsurface

A brief examination of the factors that affect conduction and advection is in order. The three components of Fourier's Law govern conduction. Of these, heat flux is determined by regional geology and advective flow, and thermal conductivity is determined mainly by physical properties of the conducting medium. Thermal gradients can indicate elevated heat flux if it can be assumed that observed increases in thermal gradient are not due to variations in thermal conductivity. The magnitude of advective heat flux in aquifers is dependent on the vertical velocity components of the local flow system.

Factors Affecting Geothermal Heat Flux. Geothermal heat flux at depth can be nearly constant over relatively large areas. Anomalously high values of flux nearer the surface of the earth in small areas are generally indicated by the presence of elevated temperatures and thermal gradients in an area, such as those observed at wells 10a and 10b (Figure 5). Such anomalously high temperature values at a given depth may be associated with intrusive magmatic bodies or caused by advection of warmer fluids to shallow levels. The presence of magmatic bodies is not believed to be the case in the study area, mainly because the amplitude of temperature anomalies is relatively small (10 $^{\circ}C$), and evidence of hot thermal springs or late Cenozoic volcanism in the region is lacking. Advective transfer of heat to shallow depths by the movement of ground water is believed to take place near The Narrows. For the flow to produce anomalously high temperatures at shallow depths, the water must have risen through the subsurface faster than it was able to dissipate heat into the surrounding alluvium.

Factors Affecting Thermal Conductivity. The thermal conductivity of saturated media will depend upon its mineral composition as well as its porosity and pore-grain packing geometry. To a lesser extent, it is also affected by temperature itself. Thermal conductivity of a saturated porous medium is affected by the same factors that affect volumetric heat capacity; the composition of the media's solid phase, bulk density, and volume fractions of minerals and water. The thermal conductivity of quartz is 15 times that of water; clay minerals are about 5 times more conductive than water. Composite thermal conductivity of a two phase system will be intermediate between the thermal conductivity of the solid and fluid phases. It is difficult to quantitatively characterize composite thermal conductivity because of its dependence on the packing geometry of the media.

Factors Affecting Advection. Because advective heat transfer in aquifers results from dispersive fluid flow, the velocity of the fluid, particularly its vertical component, will be a factor in the heat flow regime of a ground-water system. Increased vertical velocities in an aquifer will consequently increase advective heat flow. Simpson and McEligot (1983) noted that random dispersive flow of ground water has a vertical component; thus, even horizontal flow in an aquifer will involve advective heat transfer.

Geothermal Gradients

Thermal gradients can be accurately measured in non-pumping wells. From Fourier's Law, then, a first approximation of thermal flux due to conduction can be calculated using assumed thermal conductivities. Temperature profiles in 20 wells were recorded in July, 1986. Most were taken in wells that tap the upper aquifer. The data were collected near the end of the dry season to minimize the effects of temperature differences between the aquifer and its recharge. The San Pedro River had not flowed for over one month, so it was assumed that the thermal mass of the aquifer had absorbed any anomalously cold or warm recharge and heat was flowing at steady state. The effect on the thermal regime of wells pumping in the aquifer was neglected. Because the depth to water in measured wells was greater than 10 m, measurements were assumed to have been made made at depths well below the range of diurnal temperature fluctuations, and near the limits of measurable seasonal effects. The most significant datum point for each well is the bottom-hole temperature, because advection due to ground-water flow around and in the well is poorly known.

Range of Geothermal Gradient Values in the Study Areas

Geothermal gradients were calculated from the data by 2 methods. First, for each well, a line was fitted to temperature data by the least squares method, as illustrated in Appendix B. What appeared to be questionable surface or bottom-hole effects were removed from the data before they were fit. In most cases, the lines plotted using least squares are similar to those that would have been fitted by hand. Fitting the data in such a way should provide a uniform, unbiased and reliable method for treating the data. These gradients were extrapolated to intersect the zero-depth line, where a corresponding temperature was read. The average observed gradient was 51 ^{o}C/km, and the average surface intercept temperature was 18.7 ^{o}C, about 1.6 ^{o}C greater than mean annual air temperature (Table 1).

Second, thermal gradients were computed by interpolating between two points on the graph of depth versus temperature: the mean annual temperature near the surface, and the temperature at the bottom of the borehole. The temperature and depth of the bottom of the well were measured. Three different methods of characterizing the average surface temperature were attempted.

The first method used the mean annual air temperature in Benson, 17.1 ^{o}C (Green and Sellers, 1964), as the average surface temperature. Gradients thus calculated were considerably greater than those measured. These steep temperature gradients are present because the mean annual

Table 1. Calculated and measured thermal gradients and fluxes.

WELL #	LOCATION	DEPTH (m)	MEASURED GEOTHERMAL GRADIENT (deg. C/km)	SURFACE INTERCEPT (deg. C)	BOTTOM HOLE TEMP (deg. C)	BOTTOM HOLE DEPTH (m)	CALCULATED GEOTHERMAL GRADIENT (deg. C/km)	MEASURED TO CALCULATED RATIO	CONDUCTIVE HEAT FLUX (measured) HFU	CONDUCTIVE HEAT FLUX (calculated) HFU
1a	(D-15-20)21dcc-1	40	11	19.24	19.66	39.6	24.92	0.44	0.55	1.25
1c	(D-15-20)21dcc-3	26	20	18.92	19.40	24.4	29.73	0.67	1.00	1.49
3a	(D-15-20)28ccd	35	40	18.76	19.56	22.9	39.27	1.02	2.00	1.96
7	(D-15-20)21dbc	37	14	19.45	19.82	33.5	34.55	0.41	0.73	1.73
8	(D-15-20)21dba	37	7	19.59	19.64	24.4	40.30	0.17	0.35	2.02
10a-1	(D-15-20)21cbb-1	34	207	15.38	22.00	30.5	111.90	1.85	10.35	5.60
10a-2	(D-15-20)21cbb-1	34	100	15.38	22.00	30.5	111.90	0.89	5.00	5.60
10b	(D-15-20)21cbb-2	64	69	25.76	29.85	61.0	185.90	0.37	3.45	9.30
11a	(D-15-20)21bdc-1	27	20	19.97	20.35	18.3	95.20	0.21	1.00	4.76
11b	(D-15-20)21bdc-2	30	69	18.26	20.06	27.4	51.55	1.34	3.45	2.58
12	(D-15-20)21acb	14	51	17.95	18.65	13.7	-3.86	-13.21	2.55	-0.19
13	(D-16-20)05bbc	34	19	17.83	18.37	33.5	-10.03	-1.89	0.95	-0.50
14	(D-15-20)21bdb	40	54	18.62	20.68	39.6	51.17	1.06	2.70	2.56
15	(D-15-20)32bad	24	9	18.45	18.69	24.4	-0.26	-34.62	0.45	-0.01
17	(D-15-20)21dbd	32	14	19.95	20.25	30.5	52.54	0.27	0.70	2.63
18	(D-15-20)15cca	18	-32	20.05	19.54	18.3	48.50	-0.66	-1.60	2.43
19	(D-15-20)10cbb	40	167	13.12	19.56	39.6	22.23	7.51	8.35	1.11
20	(D-15-20)28bca	46	81	18.97	21.32	30.5	88.64	0.91	4.05	4.43
22	(D-16-20)33add	38	39	18.69	19.62	24.4	39.40	0.99	1.95	1.97
23	(D-16-20)28dbb	27	67	17.64	19.23	27.4	20.08	3.34	3.35	1.00
25	(D-17-20)10bbc	262	44	21.59	26.43	109.7	71.10	0.62	2.20	3.56
***** AVERAGES		44.7	51	18.7	20.7	33.5	52.6	-1.3	2.5	2.6
***** STD. DEVIATIONS		49.6	53.5	2.4	2.6	19.6	44.9	8.2	2.7	2.2

air temperature is usually less than mean annual soil temperature because direct solar radiation warms the upper layers of the soil more than the air. Based on data gathered in Safford, Arizona, the mean temperature at 1 m below the surface is normally about 3 ^{o}C greater than mean annual air temperature (Matthias,1986).

Thus, the second method of estimating average surface temperature used a mean temperature of 20.1 ^{o}C at a depth of 1 m, and assumed that the thermal regime of the upper soil horizon in the study area is similar to that near Safford. Gradients calculated in this manner were usually smaller than measured gradients, and often negative. If advective heat transfer were taking place within the aquifer, the calculated gradients would be expected to be larger than those observed. Because advection is believed to account for some of the heat flow, this method of calculating gradient is not believed to give reasonable results, possibly because the registration temperature of the thermister was not the same from well to well.

The third method of estimating the average surface temperature used the mean surface temperature as calculated from the extrapolation of the measured gradients. Gradients calculated in this way generally agreed well with observed gradients, and averaged slightly greater than observed gradients. Thermal gradients calculated using this method are given in Table 1.

Measured thermal gradients in the wells ranged from 7 ^{o}C/km at well 8 to 207 ^{o}C/km at well 10a, and included a negative value at well 18, located about 200 m south of the granitic outcrop at The Narrows. The mean of the measured gradients is 51.0 ^{o}C/km with a standard deviation of 53.5 $^{o}C/km$. The values of geothermal gradient calculated using the interpolation between mean surface temperature intercept and well-bottom temperature ranged from -10 $^{o}C/km$ at well 13 to 185 $^{o}C/km$ at well 10b, with a mean of 52.6 $^{o}C/km$ and a standard deviation of $45^{o}C/km$. Most of these calculated gradients equaled or exceeded measured gradients.

An attempt was made to construct a contour plot of the measured gradient and to contour isotherms at different depths based on the temperature data, but the data were too sparse and ranged too much. Isothermograds and isotherms plotted from well temperature data should closely represent lines of equal heat flow because there is no reason to believe that thermal conductivity varies greatly between wells.

Thermal Conductivities

In order to calculate heat flux, representative values of thermal conductivity of alluvium must be estimated. Hillel (1980, after van Wijk and de Vries (1963)) gives values of thermal conductivity for saturated sand and clay soils having 40% porosity as being 5.2×10^{-3} and 3.8×10^{-3} cal/cm/sec/^{o}C, respectively. Most gradients were measured in wells tapping the upper aquifer, which is composed mostly of sand and silt with a porosity estimated near 25%. Simpson and McEligot (1983) estimated the thermal conductivity of saturated alluvium in the Tucson basin at between 5.0×10^{-3} and 5.9×10^{-3} cal/cm/sec/^{o}C. A conservative value of 5.0×10^{-3} cal/cm/sec/^{o}C is used for the thermal conductivity of the upper aquifer in this report.

Based on available flux and conductivity data, Simpson and McEligot (1983) were able to calculate thermal gradients that would arise from pure conduction by applying Fourier's Law based on estimated thermal conductivities and reliable geothermal heat flux data in Tucson. They found that these calculated gradients usually were greater than the thermal gradients measured in the area. The difference was attributed to the convection of heat by ground-water advection. The mass transport mechanism that causes this advection results from the vertical component of dispersive flow through porous media.

Geothermal Heat Flux

The flux of heat from the earth is thought to average about 1.2 HFU (Simmons, 1966). The units of geothermal heat flux are commonly expressed in HFU, or heat flow units. One HFU equals 10^{-6} cal/cm^2/sec. Values of geothermal flux in the Basin and Range physiographic province are generally above the world-wide average. Data are not available regarding geothermal heat flux in the upper San Pedro Valley. Sass, et al. (1971) published flux values for the western United States, which include values in the Tucson basin of 2.56, 2.14, 1.98, 2.10, 1.88 and 1.56 HFU. The nearest measurements to the regional study area are 2.03 HFU in the Santa Rita Mountains and 1.54 HFU at San Manuel, each some 75 kilometers from The Narrows.

A rough estimate of heat flux in the regional and local study areas was calculated by assuming pure conduction. The estimated thermal conductivity in saturated alluvium of 5.0×10^{-3} cal/sec/cm/^{o}C was used with both measured and calculated geothermal gradients to compute thermal flux. These data are presented in Table 1. The generally elevated flux of heat, as reflected by the increased gradients near The Narrows, is not believed to result from a hot spot in the basement rocks. Halversen's (1983) depth-to-bedrock maps indicate that the geometry of the flow system's lower boundary favors upward flow of warmer ground water. Thus, the heat flux calculations presented in Table 1 are an approximation of conductive heat flow near the surface, and do not represent geothermal heat flux at a deep (1000 m) horizon, which is assumed nearly uniform over the regional study area.

Measured heat flux averaged 2.5 HFU and calculated flux averaged 2.6 HFU. Figure 11 is the temperature profile of well 25, which was flowing at a rate of 0.013L/sec (0.2 gpm) in October, 1986. Located in Benson at (D-17-20)10bbc, the well is far enough up-gradient from The Narrows to represent assumed "normal" or background geothermal conditions in the regional study area. Temperatures were measured at 1.5m (5 ft) intervals from the surface to a depth of 110 m (360 ft) where an obstacle in the well prevented further lowering of the thermister. Well 25 probably penetrates the fine silts and clays of the upper valley fill deposits at depths greater than 30 m, and its reported total depth is 262 m. By using the value of geothermal gradient displayed by this well at depth (39 ^{o}C/km), heat flux at this location was calculated to be 1.64 HFU. Birch (1947) described a method for calculating geothermal heat flux from a flowing well, provided the well is accurately logged, displays the proper temperature profile, and it is known how long and at what rate the well has flowed. It was not possible to apply Birch's method to the flowing well described here.

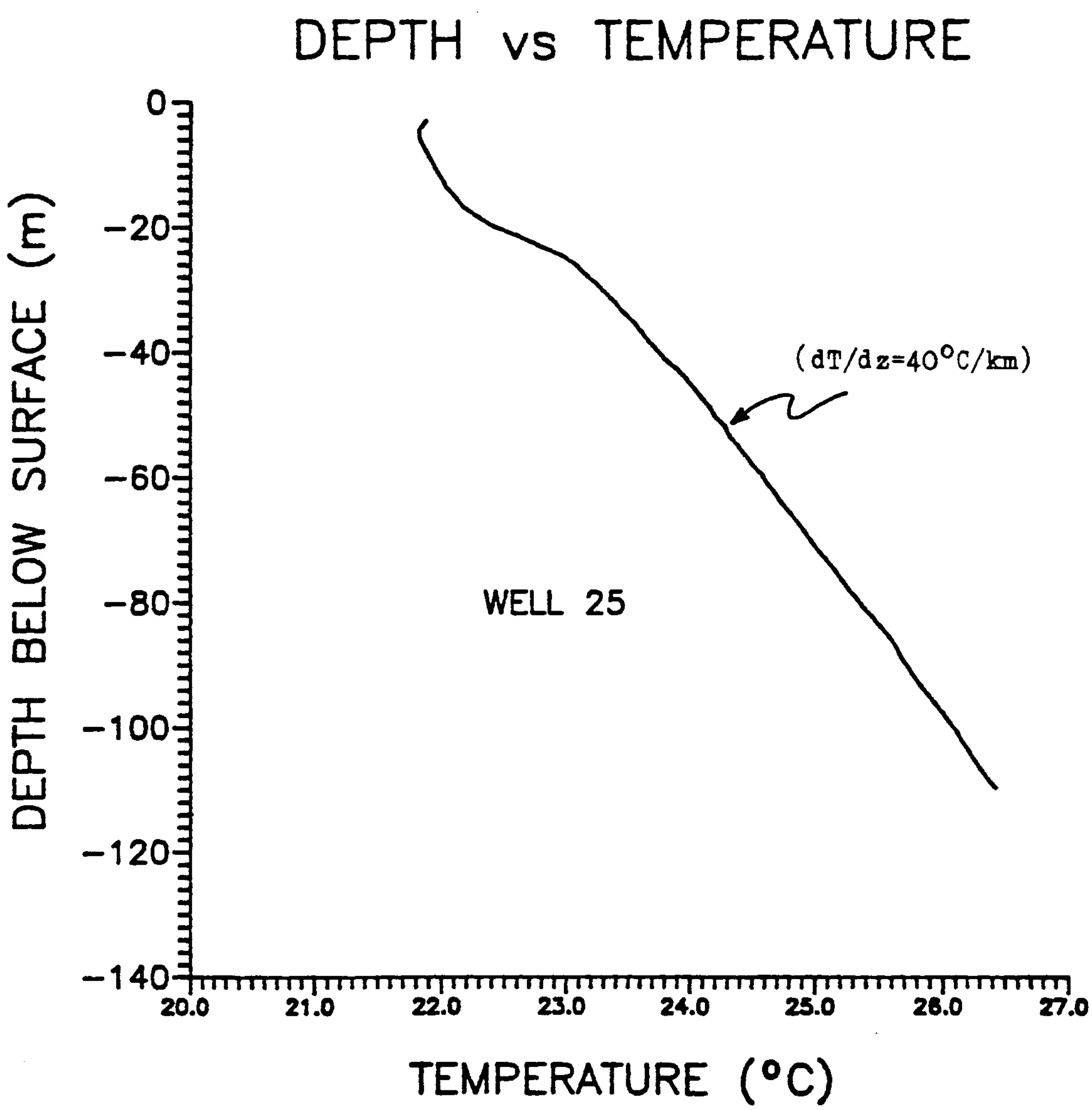


Figure 11. Thermal profile of a flowing well at (D-17-20)10bbc.

Anomalous Thermal Behavior Near The Narrows

Data gathered from wells 10a and 10b, and from several other wells near The Narrows, are anomalous in various ways. Firstly, the temperatures of the above-mentioned wells at (D-15-20)21cbb are anomalously high, indicating elevated heat flux. Second, given that the heat flow is high in the area, and assuming spatial variations of flux are not extraordinary, the wells still behave anomalously by displaying vastly different geothermal gradients. The difference in the gradients may be due to greater advection in the lower aquifer.

Description of Anomalous Behavior

A description of the physical, geologic, and hydrologic conditions encountered at (D-15-20)21cbb is in order. Figure 12 is a north-south cross section showing wells 10a and 10b. Well 10a, on the west bank of the San Pedro River, taps the upper aquifer to a depth of 34 m, where the driller reported encountering the characteristic red clay bed that defines the bottom of the unconfined aquifer and the the top of the confining beds. The depth to water in well 10a was 15.5 m. Well 10b is 4 m above and less than 150 m north of 10a. The depth to the piezometric surface in well 10b was 29 m, thus the head difference across the confining layers in this area was 9.5 m, the greater head being in the upper aquifer at that location. Well 10b had not been pumped in 5 days. Well 10b penetrates the top of the confined aquifer at a depth of about 58 m, then taps 6 m of coarse, water-bearing sand and gravel. The drilling was stopped at a depth of 64 m, where massive granite was encountered. The same granite probably crops out about 400

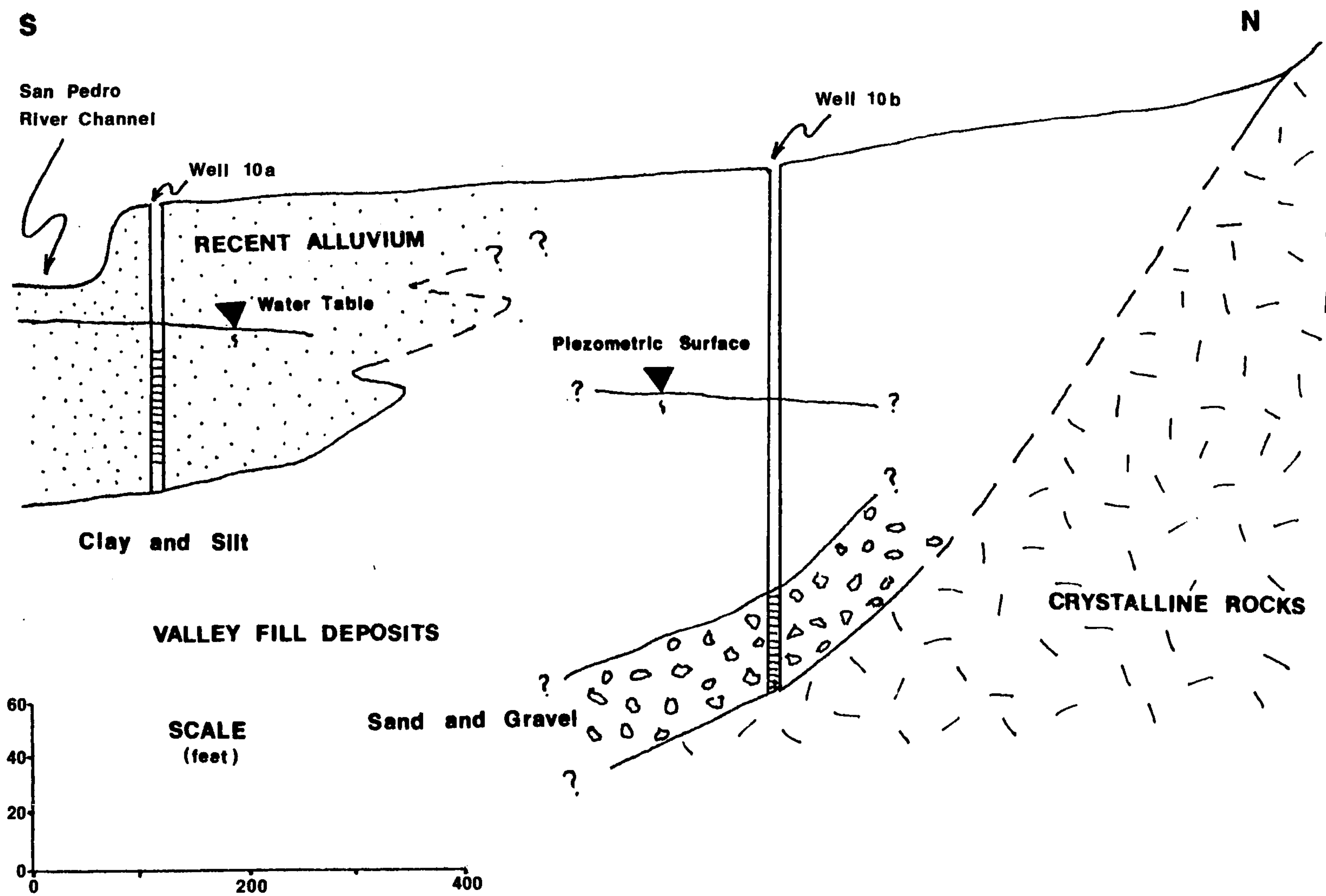


Figure 12. Generalized north-south geologic cross section between wells 10a and 10b.

m northwest of the well (Figure 8), and is thought to be representative of the basement complex near The Narrows (Figure 10).

Based on an "average" value of geothermal gradient of about 40 ^{o}C/km, as obtained from the linear portion of the temperature profile of well 25 (Figure 11), and the mean annual air temperature at Benson, water as warm as that measured at the bottom of well 10b would have to come from a depth of about 300 m. Well 10b is only 64 m deep. Using the estimated value of thermal conductivity and the measured temperature gradient in well 10a, the heat flux due to conduction through the upper aquifer was calculated to be 10.35 HFU. This is about 5 times higher than most fluxes in the Basin and Range Province, and certainly constitutes anomalous behavior. However, the least-squares-plotted value of gradient is made steeper by a the datum point at the bottom of the well (Figure 13). The other datum points in well 10a define a straight line, so the temperature gradient for well 10a was also fitted by hand to give a gradient of 100 ^{o}C/km, with a resulting heat flux of 5.0 HFU. The conductive flow of heat in well 10b was similarly calculated to be 3.45 HFU. Spatial changes in thermal conductivity may not account for all of the large difference in the measured geothermal gradient. In fact, thermal conductivity of the confining beds tapped at well 10b are presumed to have a lower value than 5.0 cal/cm/sec/^{o}C due to the fine-grained nature of the sediments, thus partially offsetting the effect of a decrease in heat flux between wells 10a and 10b.

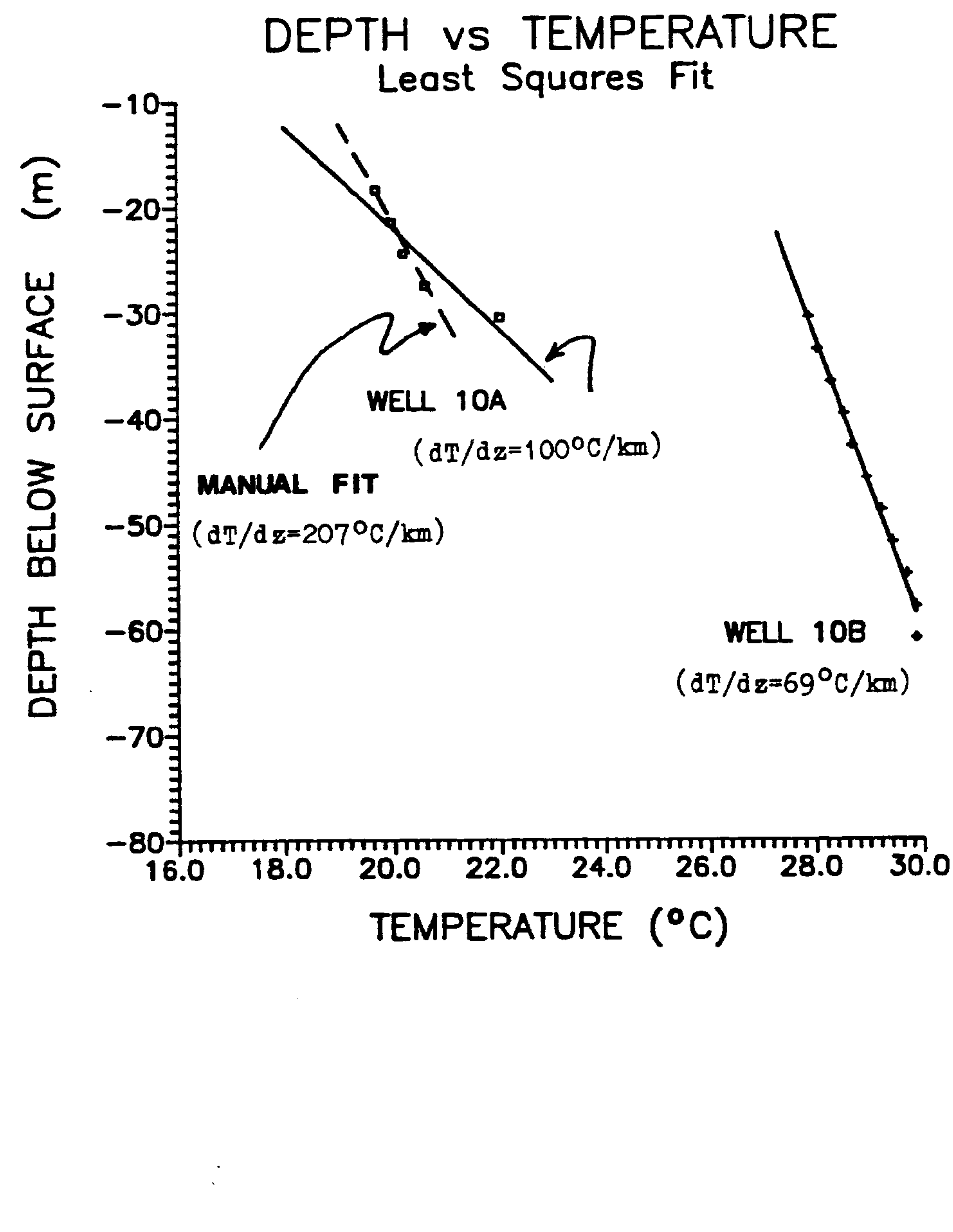


Figure 13. Thermal profiles of wells 10a and 10b at (D-15-20)21cbb.

Conduction Versus Advection

If the heat flow at well 10a is similar to that at well 10b, then the difference in temperature gradient observed could be attributed to greater advective heat flow in the lower aquifer, as reflected by the smaller gradient present in well 10b. Simpson and McEligot (1983) calculated temperature gradients that would result from pure heat conduction based on knowledge of thermal flux and conductivity. Because measured geothermal gradients in wells generally were found to be smal ler than calculated values, they proposed that ground-water advection is responsible for a portion of heat transfer in aquifers. The data presented here also suggest that advection may account for part of the heat transported in the subsurface, because observed geothermal gradients were often smaller than the calculated gradients predicted by assuming pure conduction between points at the bottom of a borehole and the surface. However, this is dependent upon the manner in which average surface temperature is determined. The ratios between measured and calculated gradients are presented in Table 1.

Other Anomalies and Interpretations

Rigorously speaking, the hypothesized trend of a general temperature increase toward The Narrows was not detected. However, many temperatures in the area south of The Narrows seem to be somewhat greater than those further upstream toward Benson. The larger thermal anomalies observed were not present over a large area. This leads to the conclusion that underflow of the lower aquifer at The Narrows takes place at greater depths than expected. Halverson's (1983) maps suggest that underflow at The Narrows would take place at levels above the ridge in the basement rock there, at depths shallower than about 250 m.

Other wells in the area near The Narrows exhibited unusual thermal behavior. Well 18, located at (D-15-20)15cca showed a negative thermal gradient. This behavior can not easily be explained. The proximity of this well to The Narrows (200 m) makes odd behavior all the more interesting. The negative profile may be due to strong surface effects, or to high convection within the borehole. The large calculated gradient indicates high heat flow in the area. This is especially noteworthy since the bottom-hole temperature that the gradient was calculated with was the coolest temperature measured in the well. Well 19, located about 2 km north of The Narrows at (D-15-20)10cbb had the second highest measured thermal gradient in the area, and the coolest ground water. Temperatures measured near the well's surface were the lowest recorded in the regional study area. This could be attributed to recharge arriving from significantly higher elevations or a difference in temperature registration of the thermister. Well 14 had a very straight measured thermal gradient curve, and it is an uncased and unused well. This well showed good agreement between measured and calculated thermal gradients.

Although the hypothesized increase of ground-water temperature toward The Narrows was not rigorously proven correct, it does seem reasonable to conclude that the upward advection of deep ground water causes unusually high surface heat flow in places near The Narrows, as well as some thermal phenomena that are difficult to explain.

CHAPTER 5

GROUND-WATER CHEMISTRY

The upper and lower aquifers in the study areas can be readily differentiated on the basis of water quality. Ground water in the upper aquifer generally contains more dissolved-solids than water from the lower aquifer. Analysis of well water can reveal which aquifer is being pumped without knowledge of the well's log or depth. Under certain conditions, careful analysis of the concentration of certain chemical constituents may lead to an estimation of ground-water flux between aquifers.

Roeske and Werrell (1973) found no indication that water quality conditions had changed much since the 1950's, and tabulated much data regarding chemical analyses in the San Pedro Valley.

Hydrochemistry of the Upper Aquifer

The chemical nature of the ground-water in the upper aquifer is determined primarily by the chemical quality of the San Pedro River that provides its recharge. Heindl (*in* Halpenny, 1952) described the chemical quality of the upper aquifer's ground water as usually having less than 300 mg/L of dissolved-solids. As measured by the author in July 1985, the upper aquifer in the regional and local study areas usually contains sulfate concentrations alone greater than 300 mg/L (Table 2). High specific conductivities measured in the areas are believed to be due to the quality of water in the San Pedro River, the recharge of

Table 2. Results of chemical analyses.

SAMPLE #	SPECIFIC CONDUCTIVITY (micromhos)	pH	NITRATE (mg/L)	SULFATE (mg/L)	SILICA (mg/L)	FLUORIDE (mg/L)
1	785	8.19	3.74	380	26.0	
2	805	8.06	7.04	415	28.0	
3A	700	7.81	8.36	315	25.8	0.95
3B	830	7.70	5.72	440	29.8	
4	785	8.41	3.08	354	25.8	
5B	925	7.96	5.94	535	27.8	
6	915	7.90	6.60	600	28.3	
7	835	7.95	7.92	465	28.3	0.80
9	855	8.13	7.26	460	28.0	
10A	670	8.20	7.04	315	21.5	1.20
10B	292	8.00	3.30	40	27.8	1.80
12	785	7.96	4.18	365	20.5	
16	280	8.00	5.28	35	25.8	
20	710	7.82	7.04	310	25.8	
22	1900	7.64	15.84	1200	23.5	

irrigation water high in leached salts, and possibly to gypsiferous beds in the fine-grained lenses in the aquifer.

Figure 14 is a scattergram based on data in Table 2, and shows the very good correlation (R^2 = 0.98) between sulfate and specific conductivity. This indicates that sulfate makes up the bulk of the dissolved solids in this region of the upper aquifer. The sulfate may be present due to gypsum beds in the finer-grained parts of the alluvium, although gypsum beds are usually only found in the upper unit of the valley fill deposits.

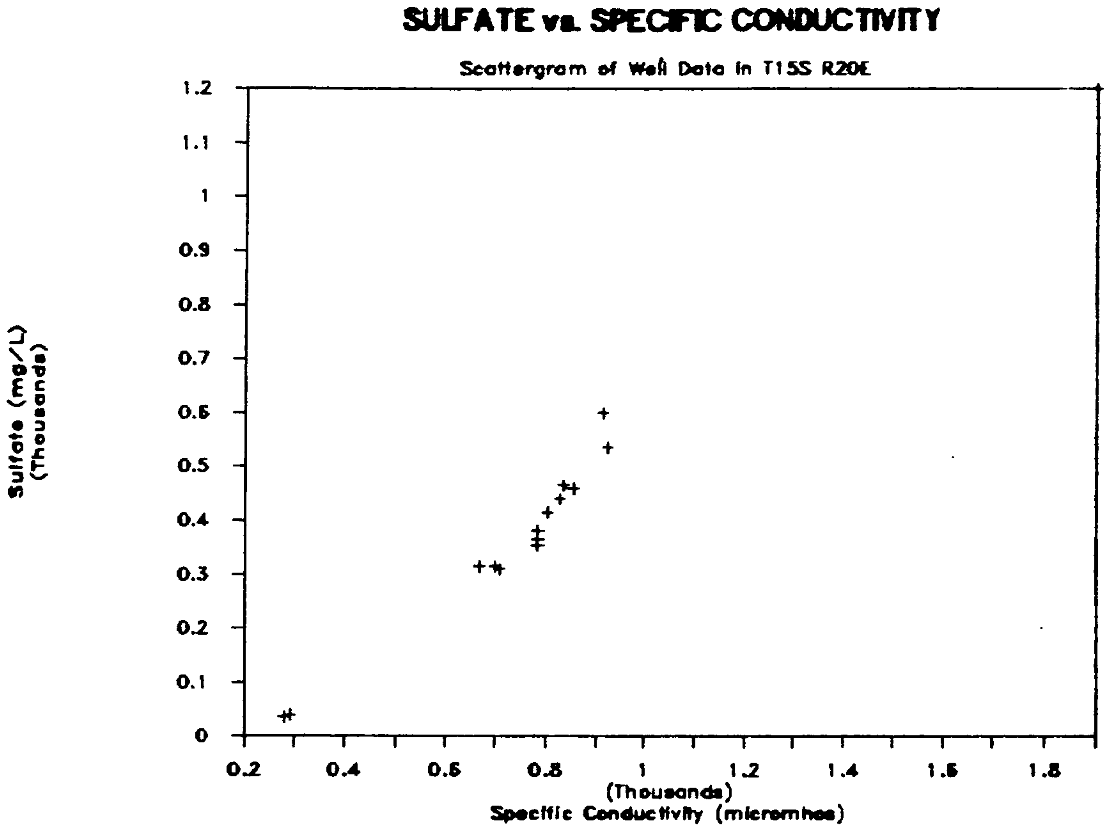


Figure 14. Scattergram of sulfate and specific conductivity.

Hydrochemistry of the Lower Aquifer

The lower aquifer yields ground water that is generally of higher quality than the upper aquifer, with the exception of occasional spotty areas of high fluoride or sulfate concentration. The mountain front recharge that replenishes the lower aquifer has low concentrations of dissolved solids, hence, ground water that is obtained from the lower aquifer can reveal the chemical nature of the sediments through which it flowed. Municipalities in the regional study area pump ground water from the lower aquifer because of its better quality.

High concentrations of fluoride are found scattered throughout the lower aquifer in the regional study area. Gypsum is assumed to be the major source of sulfate and calcium ions in the lower aquifer (Usunoff, 1984). Fluoride and sulfate concentrations seem to decrease with depth, indicating that fluoride and sulfate originate in the fine-grained clays and silts in the upper valley fill deposits.

From Table 2, the specific conductivities of the two samples obtained from the lower aquifer are about an order of magnitude less than specific conductivities of the upper aquifer samples. Concentrations of silica and nitrate were similar to those in the upper aquifer.

Mixing of Aquifers

Water samples were collected in the field to determine if chemical analysis could indicate leakage between the aquifers near The Narrows, and possibly help explain the high thermal gradients observed in some wells. Ideally, leakage could be indicated by detecting an increase in concentration of a particular index chemical constituent in an aquifer containing low background concentrations of that constituent, due to leakage from an aquifer with a relatively high concentration of the index chemical species. It was thought that either silica or fluoride could be such a characteristic chemical species in the lower aquifer. Thus, if silica or fluoride was found to be abundant in the lower aquifer, but at far lower concentrations in the upper aquifer, then an increasing concentration of silica and fluoride near The Narrows could explain the observed thermal anomalies by convection of deep, warmer ground water through the confining beds. As it turns out, though, neither fluoride or silica was present in sufficient quantities in the lower aquifer at well 10b to use this method.

Silica in ground water can be use as a geothermometer, provided that pH is not so basic that silica is precipitated. Silica anomalies have been reported in areas with thermal anomalies, and can be used to trace some geothermal systems. Hence, the warm water in well 10b might have shown a silica anomaly. Silica concentration in well 10b was slightly elevated from other values (Table 2), but no conclusions could be drawn regarding the thermal regime or ground-water flow through the confining beds on the basis of hydrochemistry. Fluoride was not found in great enough concentrations in well 10b to encourage further use of it as an index constituent for the lower aquifer.

CHAPTER 6

CONCLUSIONS

The investigation into the complex coupled system of simultaneous flow of fluid and heat near The Narrows was accomplished by analyzing hydrologic, geologic, and geothermal data as interrelated parts of a whole. Each discipline imposes constraints on the others.

Temperature Anomalies

A well less than 65 m deep that yields ground water near 30 °C constituted the thermal anomaly that initiated this study. This observation south of The Narrows probably indicates an area of increased geothermal flux, as there is no reason to believe that a sudden and drastic change in thermal conductivity accounts for such behavior. The increased thermal flux in the local study area near The Narrows can be attributed to the advective transfer of heat in the lower aquifer by upward-flowing ground water. A supposed silica anomaly was not found to correspond with the temperature anomaly at well 10b. Data from other wells further up-gradient from the region of elevated heat flux indicate the temperature of the water found at 60 m in well 10b corresponds to a depth of at least 300 m. Analysis of the geologic boundaries present at The Narrows suggests the upward flow of deep ground water in coarse, permeable sediments immediately overlying the crystalline basement rocks.

Observed temperature gradient data were fitted using the method of least squares. Geothermal gradients were calculated by interpolating between the temperature at the well-bottom and the mean temperature at the surface. Calculated gradients were in best agreement with observed gradients when the average surface temperature was characterized by the mean zero-depth temperature intercept of the observed data. The method of characterizing this mean surface temperature greatly affects the calculated thermal gradients. Advection accounts for a portion of heat transfer in the subsurface, thus, calculated thermal gradients are expected to be greater than those observed.

Ground-water Flow System

Analysis of available geologic and morphologic data leads to the conclusion that ground water from the lower aquifer flows into the lower San Pedro Basin beneath and west of The Narrows, probably in permeable sediments deposited by the ancient San Pedro River overlying the crystalline basement rocks. At The Narrows, the ridge of impermeable basement rock is buried under a minimum thickness of 250 m of alluvium, mostly valley fill sediments. This relatively shallow elevation of the bedrock is believed responsible for directing deep, normally heated ground water to higher elevations, which manifests its presence by causing anomalous thermal behavior near the surface. Magmatic intrusive bodies are probably not responsible for the anomalies because the temperature anomalies are fairly small and there is no geologic evidence of recent volcanism. The temperature anomalies south of The Narrows cannot be caused by leakage from the lower aquifer into the upper aquifer, as the head in the upper aquifer is 9.5 m greater than the lower aquifer head in that area.

APPENDIX A

DATA SUMMARY

WELL #	LOCATION	DEPTH (ft)	DEPTH (m)	TEMPERATURE DATA	MEASURED PUMPING TEMPERATURE (deg. C)	MEASURED BOTTOM-HOLE TEMPERATURE (deg. C)	HEAD DATA (Y/N)	DEPTH TO WATER (m)	CHEMICAL ANALYSIS
1a	(D-15-20)21dcc-1	130	40	Profile		19.48	Yes	15.54	No
1b	(D-15-20)21dcc-2	85	26	No			No		Yes
1c	(D-15-20)21dcc-3	84	26	Profile		19.40	Yes	14.02	No
2	(D-15-20)21ccb	100	30	Yes	19.76		Yes	14.78	Yes
3a	(D-15-20)28ccd	115	35	Profile			Yes	9.14	Yes
3b	(D-15-20)22cbb	168	51	Yes	21.22		No		Yes
4	(D-15-20)28bac	140	43	Yes	21.32		Yes	18.90	Yes
5a	(D-15-20)21dab-1	53	16	Yes	19.93		Yes	15.54	No
5b	(D-15-20)21dab-2	150	46	No			No		Yes
6	(D-15-20)21dac	130	40	No			Yes	15.24	Yes
7	(D-15-20)21dbc	110	34	Profile		19.82	Yes	13.41	Yes
8	(D-15-20)21dba	120	37	Profile			Yes	11.89	No
9	(D-15-20)21add	100	30	Yes	20.58		Yes	15.85	Yes
10a	(D-15-20)21cbb-1	100	30	Profile		22.00	Yes	15.54	Yes
10b	(D-15-20)21cbb-2	200	61	Profile		29.89	Yes	28.96	Yes
11a	(D-15-20)21bdc-1	88	27	Profile			Yes	10.67	No
11b	(D-15-20)21bdc-2	90	27	Profile		20.06	Yes	10.36	No
12	(D-15-20)21acb	45	14	Profile		18.65	Yes	9.75	Yes
13	(D-16-20)05bbc	110	34	Profile		18.37	Yes	9.75	No
14	(D-15-20)21bdb	130	40	Profile		20.68	Yes	14.63	No
15	(D-15-20)32bad	80	24	Profile		18.69	Yes	9.75	No
16	(D-16-20)07acc	95	29	No			Yes	11.89	Yes
17	(D-15-20)21dbd	104	32	Profile		20.25	Yes	16.46	No
18	(D-15-20)15cca	60	18	Profile		19.54	Yes	13.41	No
19	(D-15-20)10cbb	130	40	Profile		19.56	Yes	12.50	No
20	(D-15-20)28bca	100	30	Profile		21.32	Yes	9.45	Yes
21	(D-15-20)28bcd	125	38	No			Yes	9.14	No
22	(D-16-20)33add	80	24	Profile		19.62	Yes	10.36	Yes
23	(D-16-20)28dbb	90	27	Profile		19.23	Yes	9.45	No
24	(D-17-20)31ddd	35	11	Yes		17.75	Yes	7.62	No
25	(D-17-20)10bbc	860	262	Profile			Yes	FLOWS	No

WELL 1A

DEPTH (ft)	TEMP (deg. C)
60	19.525
70	19.529
80	19.540
90	19.531
100	19.531
110	19.510
120	19.540
130	19.662
95	19.546

WELL 1C

DEPTH (ft)	TEMP (deg. C)
50	19.225
60	19.335
70	19.362
80	19.395
65	19.329

WELL 3A

DEPTH (ft)	TEMP (deg. C)
35	19.199
45	19.220
55	19.435
65	19.632
75	19.560
55	19.409

WELL 7

DEPTH (ft)	TEMP (deg. C)
50	19.698
60	19.709
70	19.733
80	19.780
90	19.827
100	19.861
110	19.823
80	19.776

WELL 8

DEPTH (ft)	TEMP (deg. C)
40	19.599
50	19.665
60	19.651
70	19.641
80	19.643
60	19.640

WELL 10A

DEPTH (ft)	TEMP (deg. C)
60	19.677
70	19.966
80	20.201
90	20.602
100	22.001
80	20.489

WELL 10B

DEPTH (ft)	TEMP (deg. C)
100	27.889
110	28.047
120	28.289
130	28.529
140	28.678
150	28.948
160	29.209
170	29.413
180	29.689
190	29.846
200	29.854
150	28.945

WELL 11A

DEPTH (ft)	TEMP (deg. C)
40	20.216
50	20.276
60	20.347
50	20.280

WELL 11B

DEPTH (ft)	TEMP (deg. C)
40	19.185
50	19.321
60	19.436
70	19.898
80	19.973
90	20.061
65	19.646

WELL 12

DEPTH (ft)	TEMP (deg. C)
35	18.501
40	18.584
45	18.651
40	18.579

WELL 13

DEPTH (ft)	TEMP (deg. C)
40	18.311
50	18.155
60	18.143
70	18.172
80	18.260
90	18.311
100	18.346
110	18.374
75	18.259

WELL 14

DEPTH (ft)	TEMP (deg. C)
50	19.433
60	19.601
70	19.804
80	20.002
90	20.186
100	20.344
110	20.474
120	20.562
130	20.675
90	20.120

WELL 15

DEPTH (ft)	TEMP (deg. C)
40	18.580
50	18.592
60	18.607
70	18.623
80	18.694
60	18.619

WELL 17

DEPTH (ft)	TEMP (deg. C)
60	20.171
70	20.186
80	20.309
90	20.177
100	20.250
80	20.219

WELL 18

DEPTH (ft)	TEMP (deg. C)
45	19.675
50	19.568
55	19.520
60	19.539
52.5	19.576

WELL 19

DEPTH (ft)	TEMP (deg. C)
50	15.547
60	15.926
70	17.056
80	17.792
90	18.061
100	18.041
110	13.320
120	19.107
130	19.558
90	17.712

WELL 20

DEPTH (ft)	TEMP (deg. C)
40	19.985
50	20.150
60	20.511
70	20.730
80	20.962
90	21.297
100	21.315
70	20.707

WELL 22

DEPTH (ft)	TEMP (deg. C)
35	19.154
40	19.159
45	19.183
50	19.218
55	19.362
60	19.454
65	19.506
70	19.516
75	19.540
80	19.586
57.5	19.368

WELL 22

DEPTH (ft)	TEMP (deg. C)
35	19.146
40	19.164
45	19.182
50	19.212
55	19.364
60	19.459
65	19.55
70	19.572
75	19.599
80	19.622
57.5	19.387

WELL 23

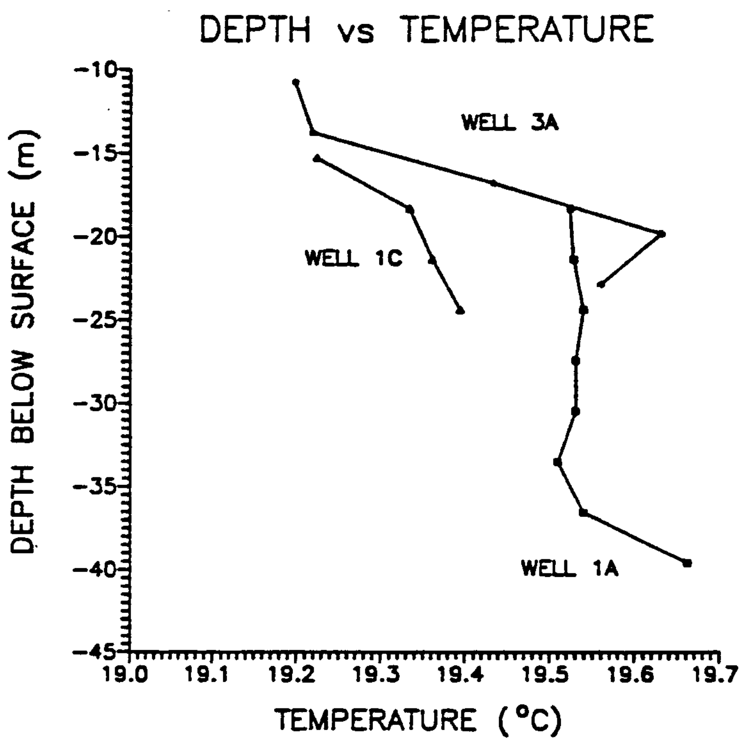
DEPTH (ft)	TEMP (deg. C)
35	19.033
40	18.977
45	18.868
50	18.798
55	18.795
60	18.802
65	18.813
70	18.812
75	18.844
80	19.055
85	19.134
90	19.23
62.5	18.93008

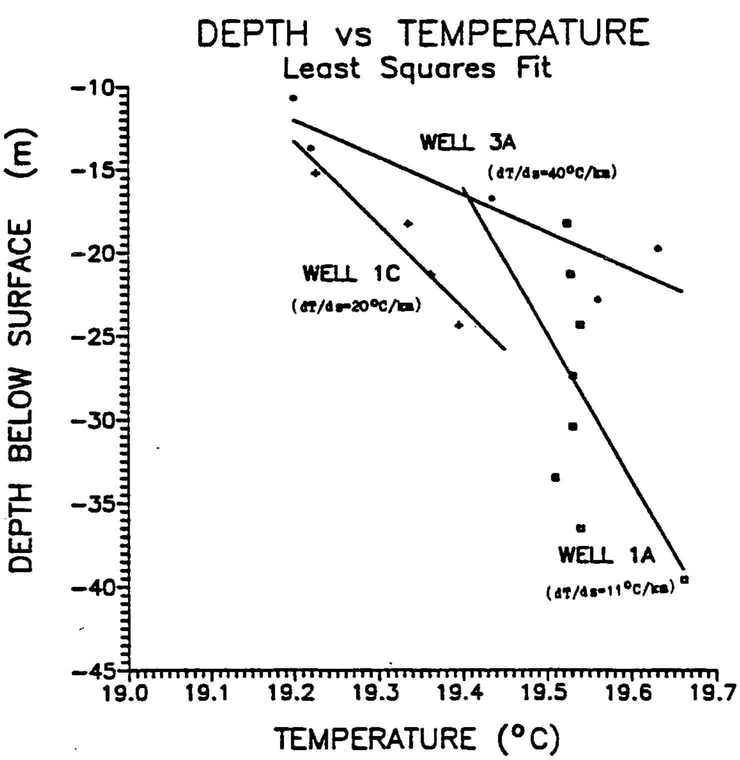
WELL 25

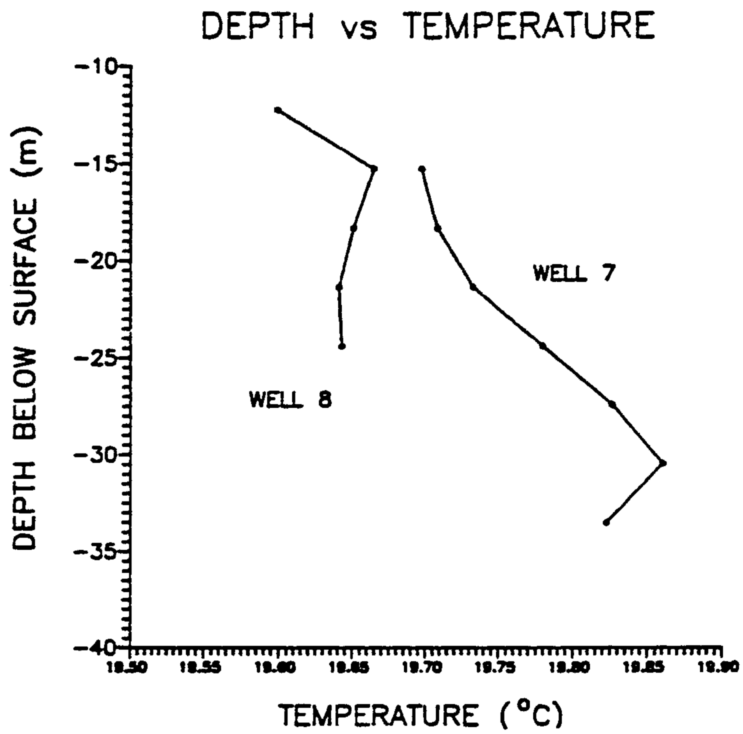
DEPTH (ft)	TEMP (deg. C)	DEPTH (ft)	TEMP (deg. C)
5	21.941	185	24.447
10	21.883	190	24.505
15	21.823	195	24.579
20	21.835	200	24.625
25	21.876	205	24.688
30	21.918	210	24.734
35	21.956	215	24.807
40	22.004	220	24.866
45	22.048	225	24.922
50	22.115	230	24.970
55	22.177	235	25.030
60	22.293	240	25.093
65	22.413	245	25.154
70	22.606	250	25.209
75	22.772	255	25.263
80	22.955	260	25.330
85	23.078	265	25.387
90	23.152	270	25.454
95	23.245	275	25.514
100	23.323	280	25.576
105	23.397	285	25.630
110	23.460	290	25.665
115	23.544	295	25.716
120	23.608	300	25.767
125	23.669	305	25.821
130	23.743	310	25.880
135	23.811	315	25.940
140	23.903	320	26.000
145	23.975	325	26.050
150	24.038	330	26.112
155	24.099	335	26.154
160	24.160	340	26.210
165	24.207	345	26.257
170	24.284	350	26.310
175	24.320	355	26.365
180	24.391	360	26.429
	Averages:	182.5	24.257

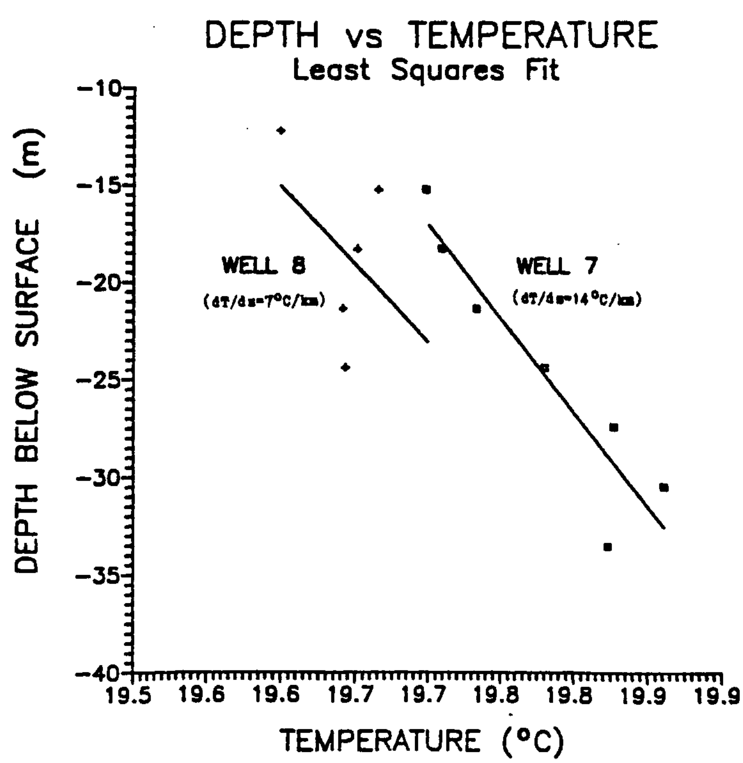
APPENDIX B

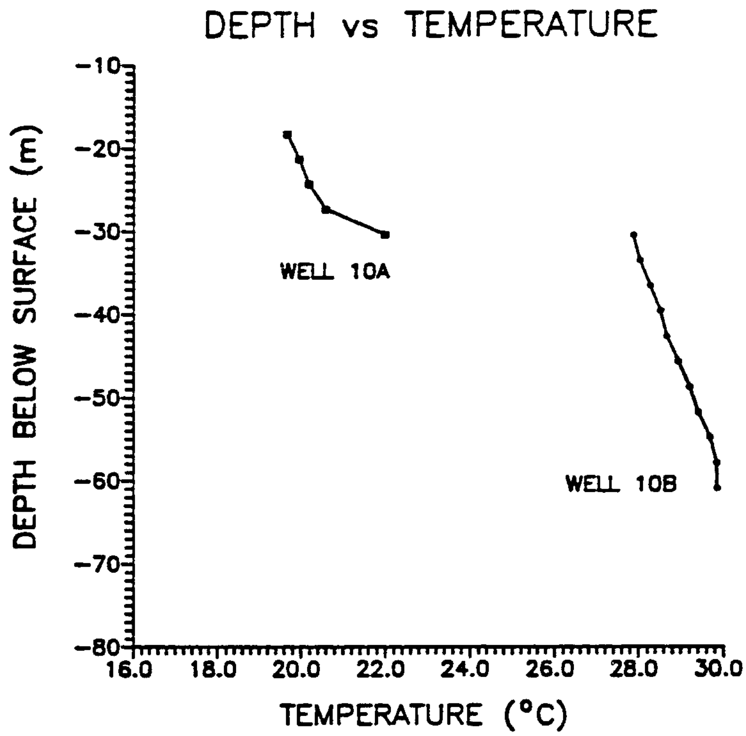
TEMPERATURE PROFILE PLOTS

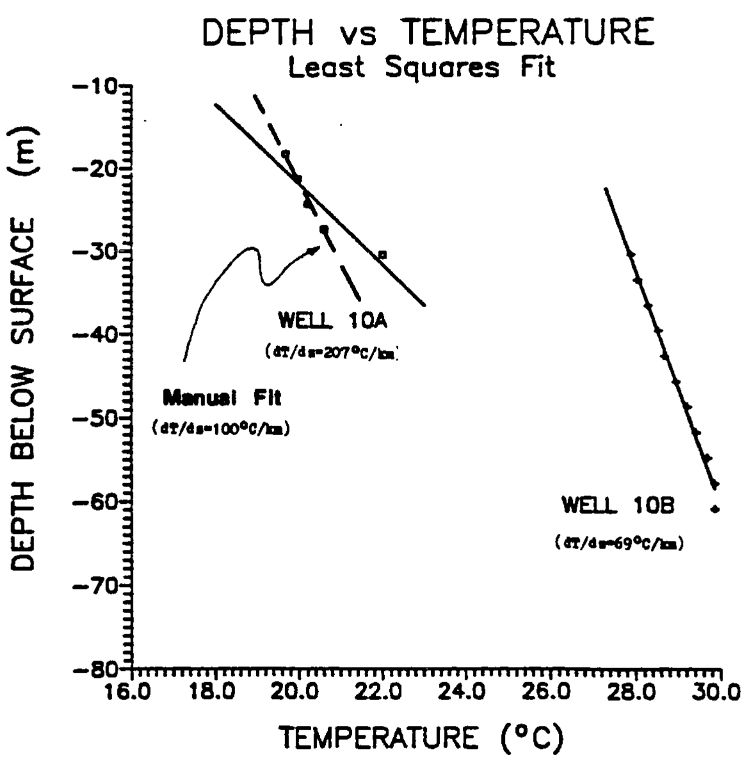

DEPTH vs TEMPERATURE
WELL 3A
WELL 1C
WELL 1A
DEPTH BELOW SURFACE (m)
-10
-15
-20
-25
-30
-35
-40
-45
19.0 19.1 19.2 19.3 19.4 19.5 19.6 19.7
TEMPERATURE (°C)

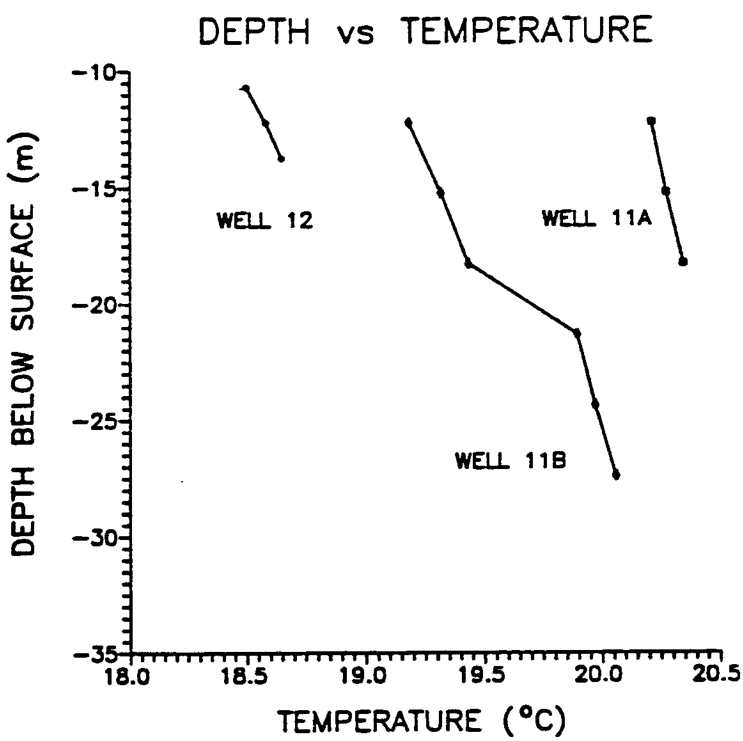

DEPTH vs TEMPERATURE
Least Squares Fit
WELL 3A
(dT/ds=40°C/km)
WELL 1C
(dT/ds=20°C/km)
WELL 1A
(dT/ds=11°C/km)
DEPTH BELOW SURFACE (m)
-10
-15
-20
-25
-30
-35
-40
-45
19.0 19.1 19.2 19.3 19.4 19.5 19.6 19.7
TEMPERATURE (°C)

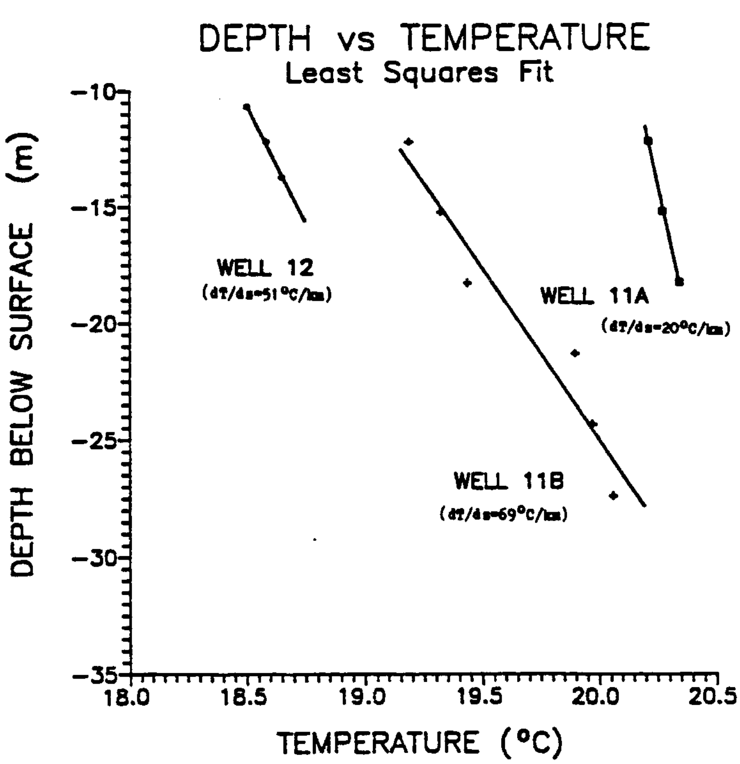

DEPTH vs TEMPERATURE
DEPTH BELOW SURFACE (m)
-10
-15
-20
-25
-30
-35
-40
WELL 7
WELL 8
19.50
19.55
19.60
19.65
19.70
19.75
19.80
19.85
19.90
TEMPERATURE (°C)

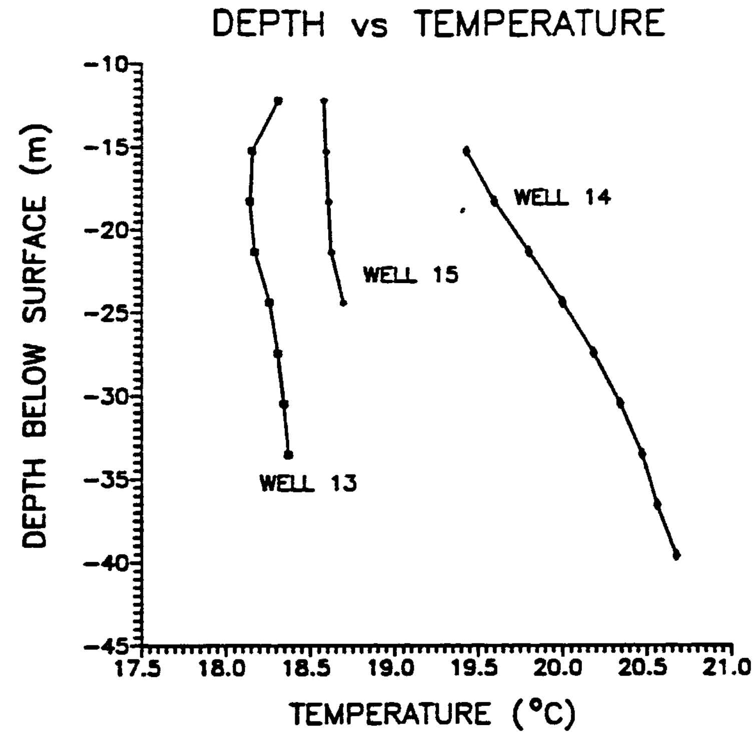

DEPTH vs TEMPERATURE
Least Squares Fit
DEPTH BELOW SURFACE (m)
-10
-15
-20
-25
-30
-35
-40
WELL 8
WELL 7
19.5
19.6
19.6
19.7
19.7
19.8
19.8
19.9
19.9
TEMPERATURE (°C)

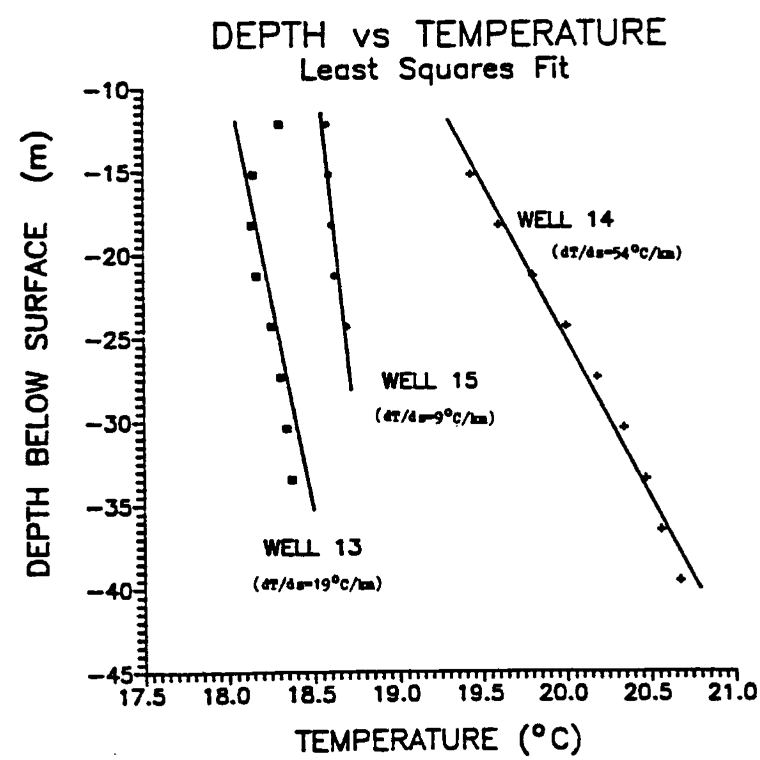

DEPTH vs TEMPERATURE
DEPTH BELOW SURFACE (m)
-10
-20
-30
-40
-50
-60
-70
-80
16.0
18.0
20.0
22.0
24.0
26.0
28.0
30.0
TEMPERATURE (°C)
WELL 10A
WELL 10B

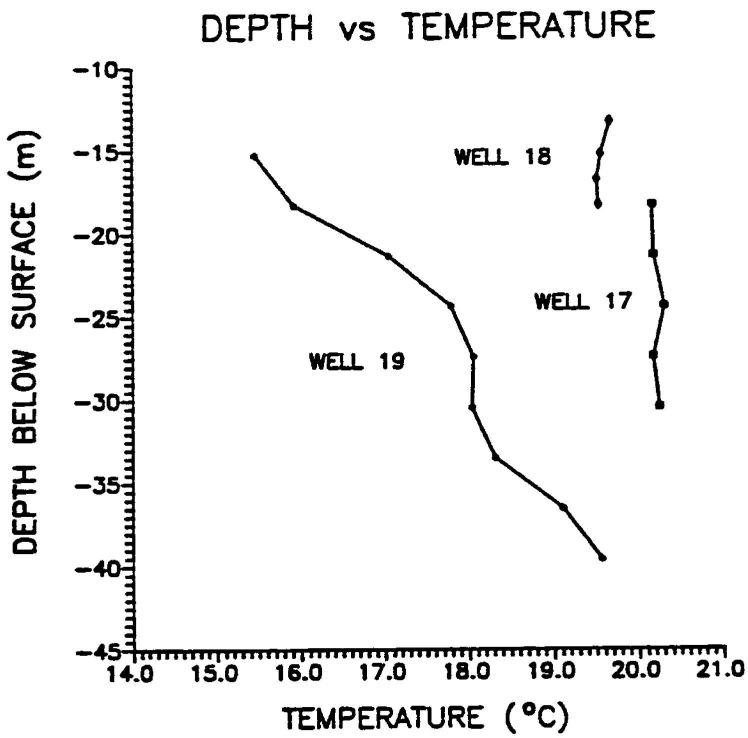

DEPTH vs TEMPERATURE
Least Squares Fit
DEPTH BELOW SURFACE (m)
-10
-20
-30
-40
-50
-60
-70
-80
16.0
18.0
20.0
22.0
24.0
26.0
28.0
30.0
TEMPERATURE (°C)
WELL 10A
(dT/dz=207°C/km)
Manual Fit
(dT/dz=100°C/km)
WELL 10B
(dT/dz=69°C/km)

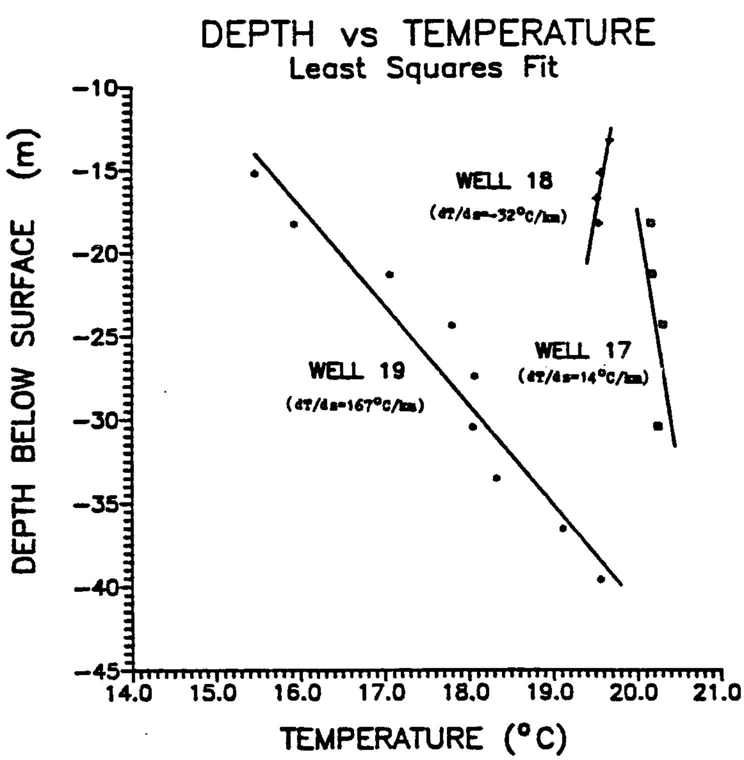

DEPTH vs TEMPERATURE
DEPTH BELOW SURFACE (m)
-10
-15
-20
-25
-30
-35
WELL 12
WELL 11A
WELL 11B
18.0
18.5
19.0
19.5
20.0
20.5
TEMPERATURE (°C)

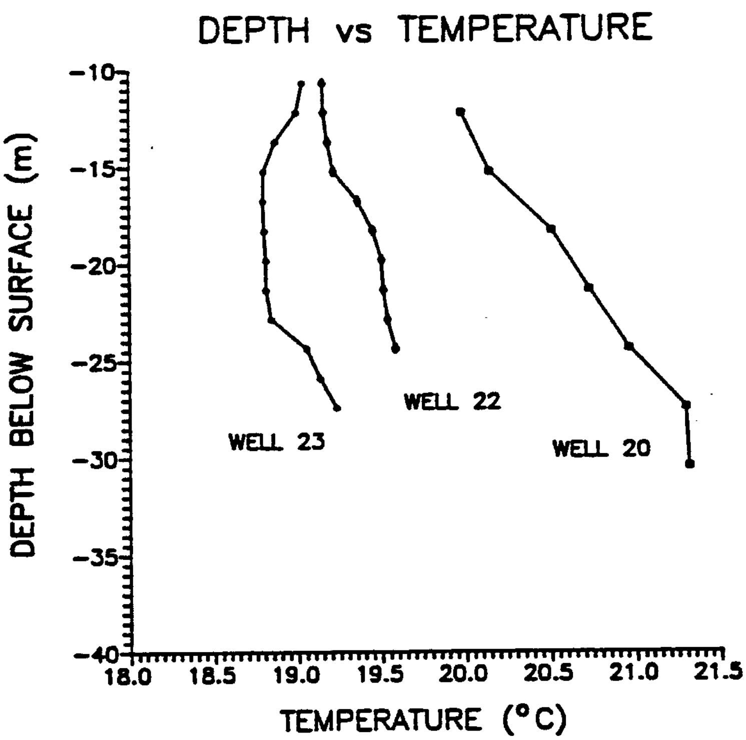

DEPTH vs TEMPERATURE
Least Squares Fit
DEPTH BELOW SURFACE (m)
-10
-15
-20
-25
-30
-35
WELL 12
(dT/ds=51°C/km)
WELL 11A
(dT/ds=20°C/km)
WELL 11B
(dT/ds=69°C/km)
18.0
18.5
19.0
19.5
20.0
20.5
TEMPERATURE (°C)

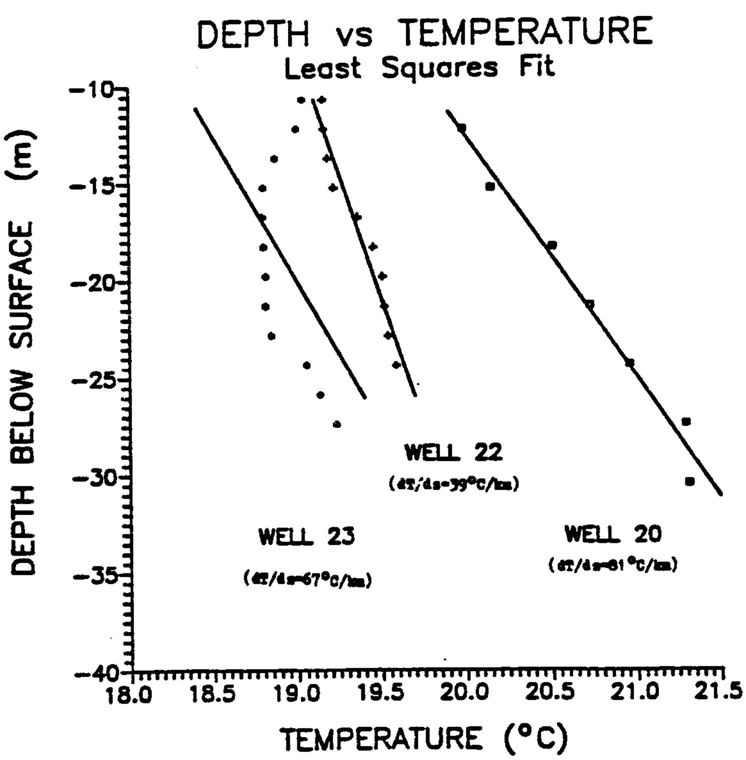

DEPTH vs TEMPERATURE
DEPTH BELOW SURFACE (m)
-10
-15
-20
-25
-30
-35
-40
-45
WELL 14
WELL 15
WELL 13
17.5
18.0
18.5
19.0
19.5
20.0
20.5
21.0
TEMPERATURE (°C)

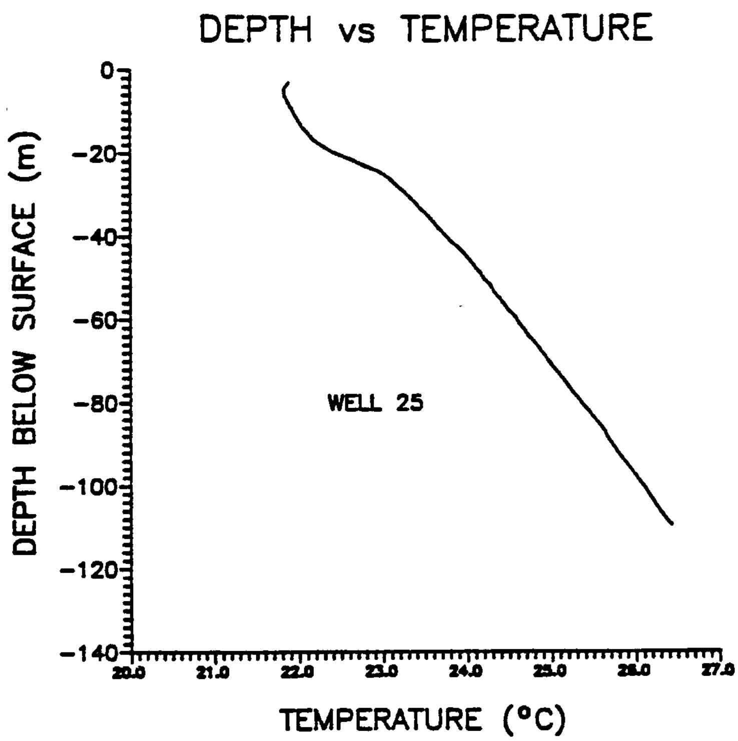

DEPTH vs TEMPERATURE
Least Squares Fit
DEPTH BELOW SURFACE (m)
-10
-15
-20
-25
-30
-35
-40
-45
WELL 14
(dT/dz=54°C/km)
WELL 15
(dT/dz=9°C/km)
WELL 13
(dT/dz=19°C/km)
17.5
18.0
18.5
19.0
19.5
20.0
20.5
21.0
TEMPERATURE (°C)

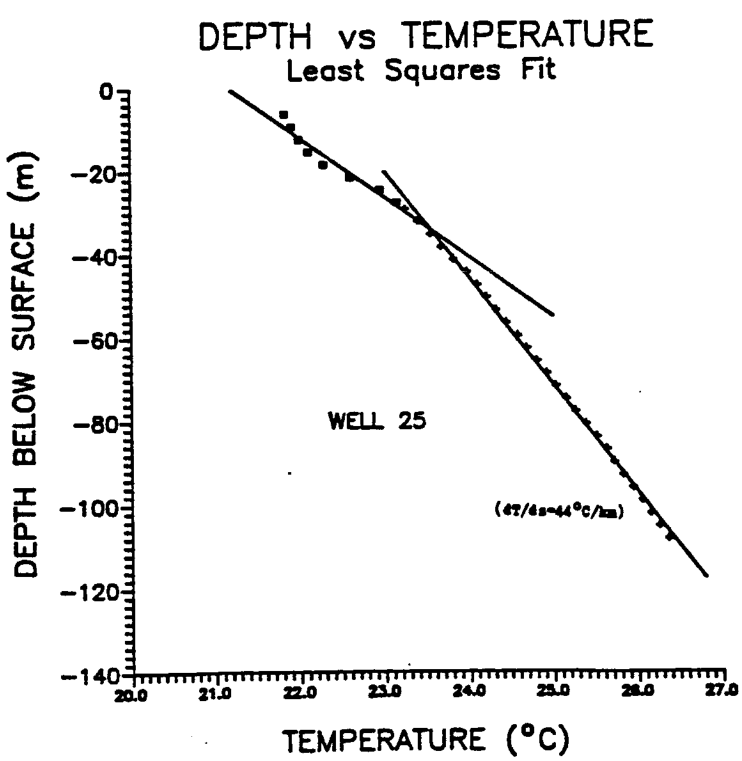

DEPTH vs TEMPERATURE
DEPTH BELOW SURFACE (m)
-10
-15
-20
-25
-30
-35
-40
-45
WELL 18
WELL 17
WELL 19
14.0
15.0
16.0
17.0
18.0
19.0
20.0
21.0
TEMPERATURE (°C)


DEPTH vs TEMPERATURE
Least Squares Fit
DEPTH BELOW SURFACE (m)
-10
-15
-20
-25
-30
-35
-40
-45
WELL 18
(dT/dz=-32°C/km)
WELL 17
(dT/dz=14°C/km)
WELL 19
(dT/dz=167°C/km)
14.0
15.0
16.0
17.0
18.0
19.0
20.0
21.0
TEMPERATURE (°C)


DEPTH vs TEMPERATURE
DEPTH BELOW SURFACE (m)
-10
-15
-20
-25
-30
-35
-40
18.0
18.5
19.0
19.5
20.0
20.5
21.0
21.5
TEMPERATURE (°C)
WELL 22
WELL 23
WELL 20

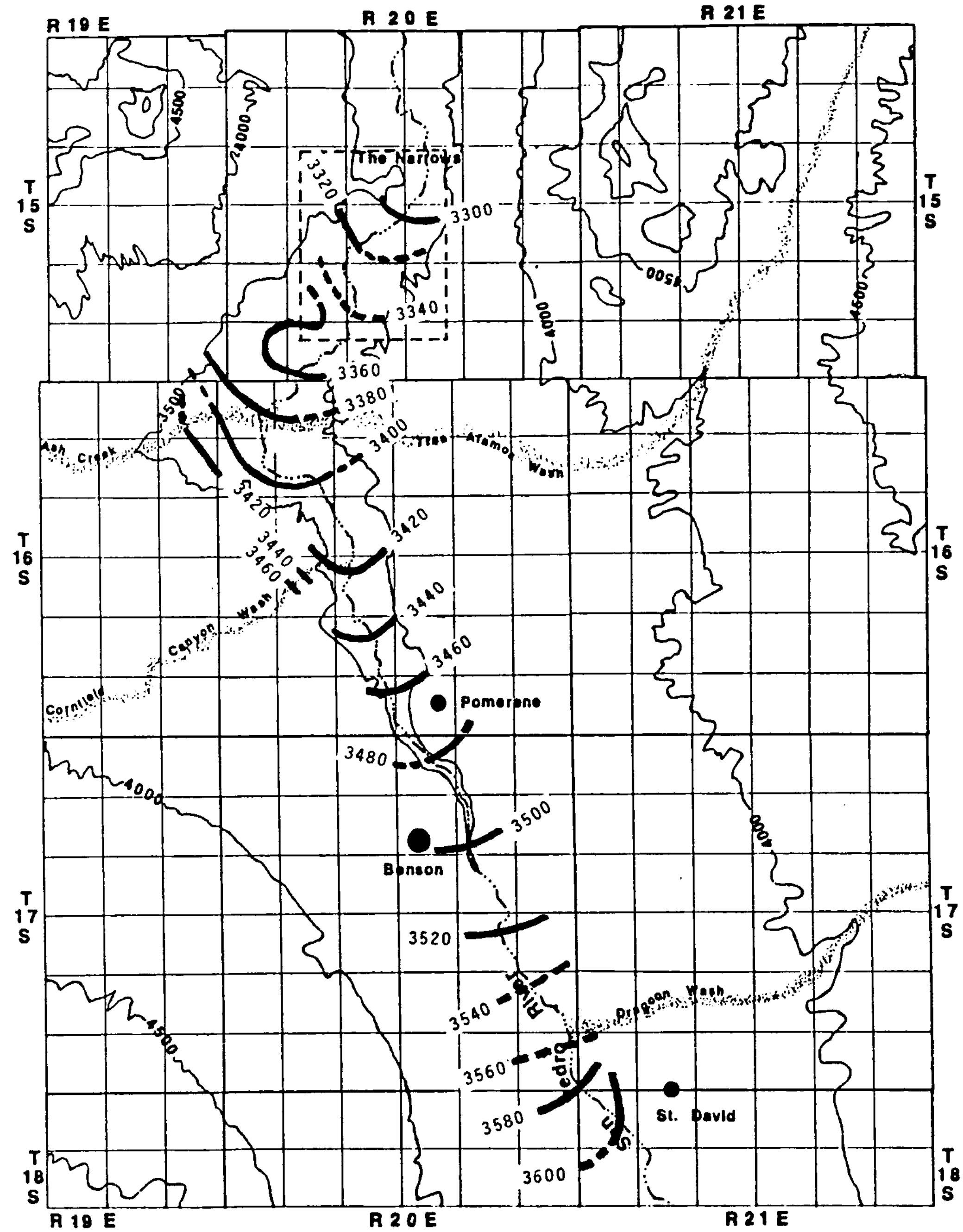

DEPTH vs TEMPERATURE
Least Squares Fit
DEPTH BELOW SURFACE (m)
-10
-15
-20
-25
-30
-35
-40
18.0
18.5
19.0
19.5
20.0
20.5
21.0
21.5
TEMPERATURE (°C)
WELL 22
(dT/dz=39°C/km)
WELL 23
(dT/dz=67°C/km)
WELL 20
(dT/dz=81°C/km)


DEPTH vs TEMPERATURE
DEPTH BELOW SURFACE (m)
0
-20
-40
-60
-80
-100
-120
-140
WELL 25
20.0
21.0
22.0
23.0
24.0
25.0
26.0
27.0
TEMPERATURE (°C)

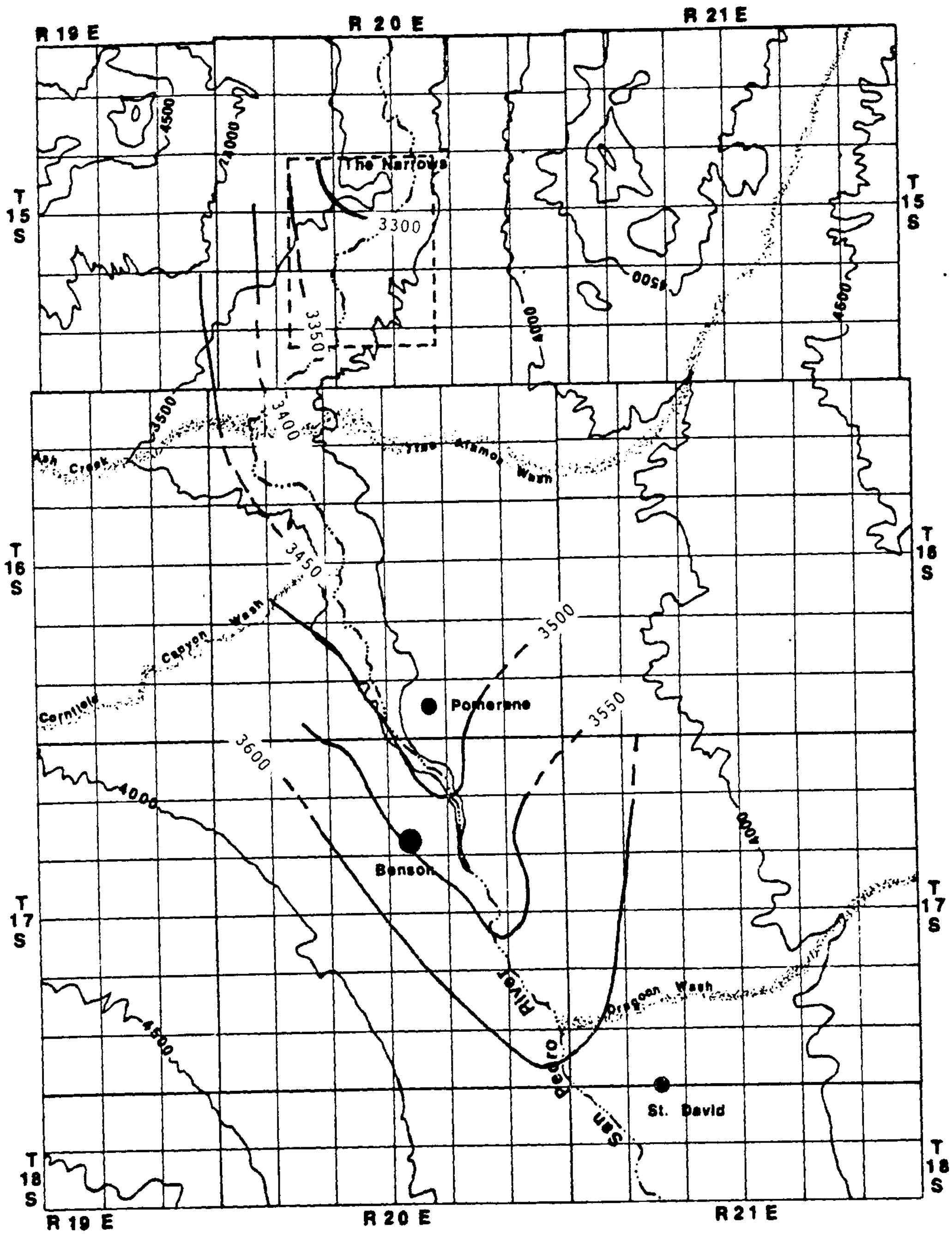

DEPTH vs TEMPERATURE
Least Squares Fit
DEPTH BELOW SURFACE (m)
0
-20
-40
-60
-80
-100
-120
-140
WELL 25
20.0
21.0
22.0
23.0
24.0
25.0
26.0
27.0
TEMPERATURE (°C)

APPENDIX C

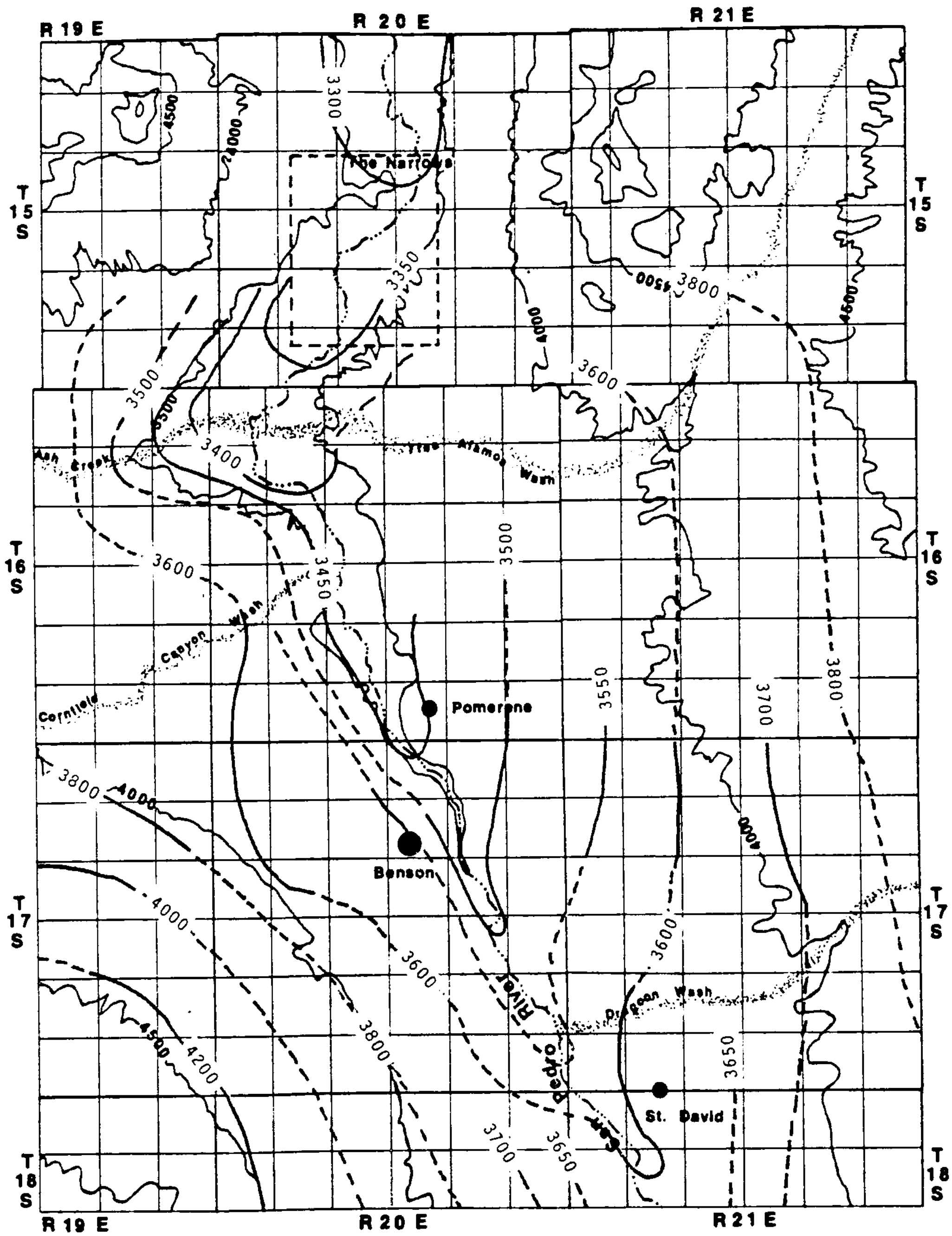
CONTOUR MAPS OF THE WATER TABLE AND PIEZOMETRIC SURFACE



Upper aquifer water table contours in June 1985, as mapped by the author. Contours are in feet above mean sea level.



Lower aquifer piezometric surface contours in June 1985, as mapped by the author. Contours are in feet above mean sea level.



Ground-water contours in 1968, with no differentiation between upper and lower aquifers. Contours are in feet above mean sea level. (After Roeske and Werrell, 1973.)

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