

EXPLANATION

WELL FROM WHICH WATER SAMPLE WAS COLLECTED—First entry, 14, is measured water temperature at wellhead in degrees Celsius. Second entry, 66M, is calculated Na-K-Ca geotemperature of last water-rock equilibration in degrees Celsius (M, Na-K-Ca geotemperature corrected for magnesium concentration; R, percentage of magnesium is greater than 50). Third entry, 69, is calculated quartz-silica and, (44), chalcodony-silica geotemperatures in degrees Celsius. Fourth entry, 0.05, is boron concentration in milligrams per liter. Letter, F, next to symbol indicates that the chemical analysis for this site is shown in table

SPRING FROM WHICH WATER SAMPLE WAS COLLECTED—First entry, 39, is measured water temperature at spring orifice in degrees Celsius. Second entry, 43M, is calculated Na-K-Ca geotemperature of last water-rock equilibration in degrees Celsius (M, Na-K-Ca geotemperature corrected for magnesium concentration). Third entry, 114, is calculated quartz-silica and, (89), chalcodony-silica geotemperatures in degrees Celsius. Fourth entry, 9.10, is boron concentration in milligrams per liter. Letter, I, next to symbol indicates that the chemical analysis for this site is shown in table

FAULT—Dashed where approximately located. U, upthrown side; D, downthrown side

APPROXIMATE AREA OF HIGH SILICA GEOTEMPERATURES—Quartz-silica geotemperatures are more than 100°C, and chalcodony-silica geotemperatures are more than 70°C

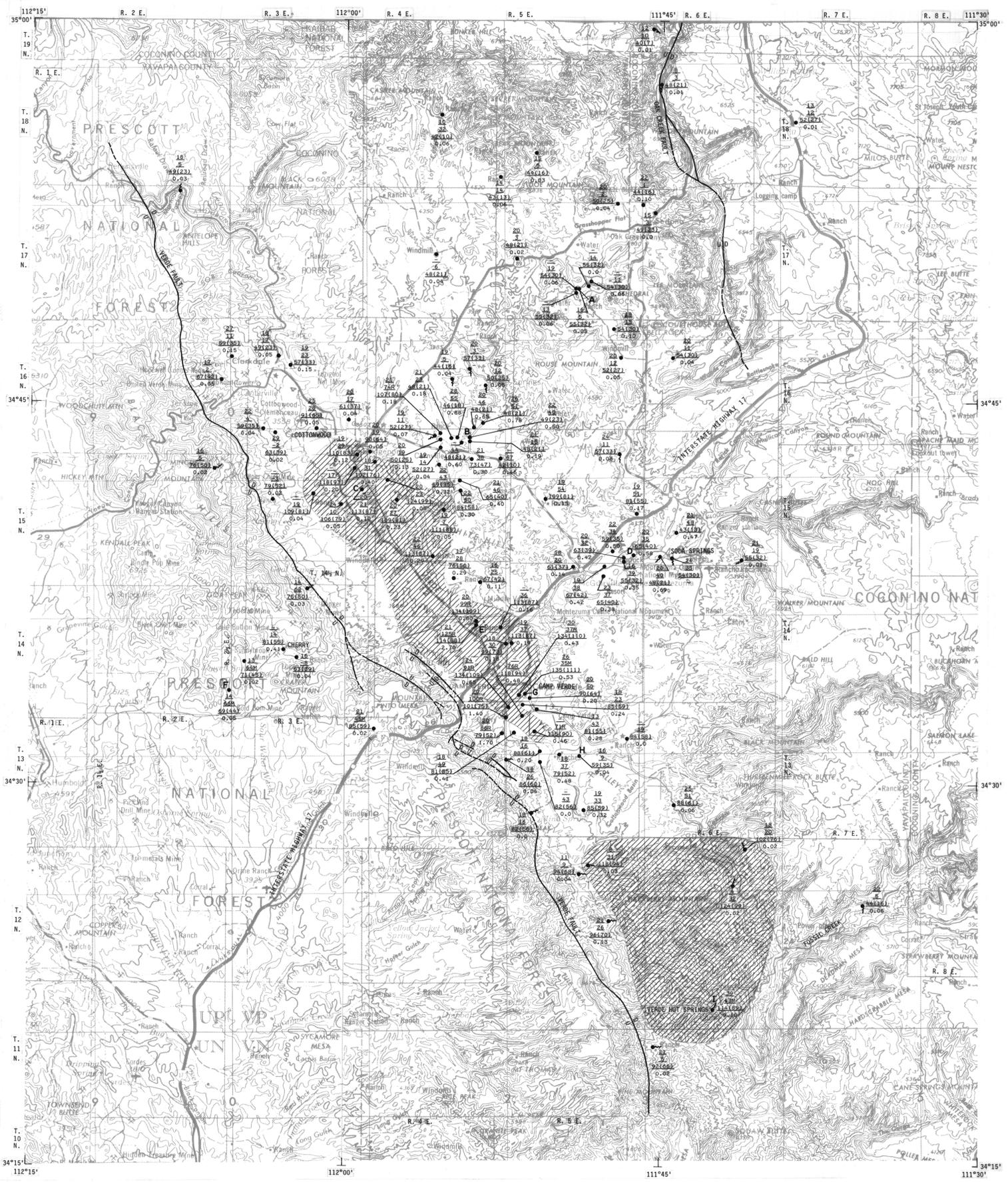
CONVERSION FACTORS

For readers who may prefer to use metric units rather than inch-pound units, the conversion factors for the terms used in this report are listed below:

Multiply inch-pound unit	By	To obtain metric unit
mile (mi)	1.609	Kilometer (km)
square mile (mi ²)	2.590	square kilometer (km ²)

INDEX MAP SHOWING AREA OF REPORT (SHADED)

**CONTOUR INTERVAL 200 FEET
WITH SUPPLEMENTARY CONTOURS AT 100-FOOT INTERVALS
NATIONAL GEODETIC VERTICAL DATUM OF 1929**



Verde Valley—an area of complex geology where the crust of the North American plate changes from unstable to stable (Lucchitta, 1974, p. 351)—is in central Arizona in the transition zone between the Basin and Range and the Colorado Plateau tectonic provinces of Swenberg and others (1977, fig. 12). Volcanism and tectonism have been progressing northeastward along the transition zone for about 20 million years, and volcanoes were active in the San Francisco volcanic field 40 mi north of Verde Valley as late as 900 years ago (Eastwood, 1974, fig. 2; Smiley, 1958, p. 190). The San Francisco volcanic field is an area of recent faulting, where 27 earthquakes having magnitudes of 2.2 to 5.1 on the Richter scale have occurred since 1936 (Shoemaker and others, 1974, fig. 9). The geologic record thus is replete with implications of geothermal activity, and hydrothermal activity is evinced by measured ground-water temperatures that are 10° to 20°C higher than the average ground-water temperature of 18°C. Subsurface temperatures sometimes can be estimated from the chemical composition of water that has been heated at depth and subsequently discharged by wells or springs. The study in Verde Valley was undertaken by the U.S. Geological Survey to delimit the potential geothermal resources and to estimate water temperatures at depth using geochemical methods.

More than 100 water samples were collected from wells and springs using the sampling procedures described by Brown and Truesdell (1970). The samples were analyzed for common chemical constituents; temperature, pH, and specific conductance were determined at the site. Three 250-ml (milliliter) samples were collected in polyethylene bottles at each site. One sample was filtered through a 45-micrometer membrane filter, acidified with reagent-grade nitric acid to a pH of 3.0 or less, and used for cation determination. The second sample was filtered through a 45-micrometer filter, untreated, and used for anion determination. The third sample was unfiltered and untreated, chilled to 4°C, and used for nutrient determination.

In Verde Valley temperatures of the potential geothermal resources were calculated using two geochemical methods or geothermometers. One method is an empirical geothermometer that uses the molal relations of dissolved sodium, potassium, and calcium ions to calculate the temperature of the last water-rock equilibration (Fournier and Truesdell, 1973). The equation used for the estimate is

$$T = \frac{1647}{2.24 + P(2)} - 273.15,$$

where

$$T = \text{temperature in degrees Celsius, and}$$

$$P(2) = \log \left(\frac{\text{Na} + \text{K}}{\text{Ca}} \right) + \log \left(\frac{\text{Ca}}{\text{Mg}} \right)$$

in which

Na = molal concentration of sodium,
 K = molal concentration of potassium,
 Ca = molal concentration of calcium, and
 Mg = molal concentration of magnesium.

If magnesium is present in significant quantities, the Na-K-Ca geotemperature must be corrected (Fournier and Potter, 1978); the steps for applying the correction are:

1. Calculate the Na-K-Ca geotemperature as described above, and do not apply a magnesium correction if the calculated temperature is less than 70°C.
2. Calculate the percentage of magnesium compared to the concentration of magnesium, calcium, and potassium as follows:

$$R = \frac{\text{Mg}}{\text{Mg} + \text{Ca} + \text{K}} \times 100,$$

where

R = percentage of magnesium,
 Mg = equivalent concentration of magnesium,
 Ca = equivalent concentration of calcium, and
 K = equivalent concentration of potassium.

3. If R is greater than 50, then assume that the water comes from a cool underground environment that has a temperature about equal to the measured temperature, regardless of the high calculated Na-K-Ca geotemperature.
4. If the calculated Na-K-Ca geotemperature is more than 70°C and R is less than 50, then calculate the temperature correction as follows:

$$\Delta T_{\text{Mg}} = 10.66 - 4.7415R + 325.867 (\log R)^2 - 103210 (\log R)^3 / (2+273)^2 - 19683000 (\log R)^4 / (2+273)^4 + 16053000 (\log R)^5 / (2+273)^5$$

where

ΔT_{Mg} = temperature correction in degrees Celsius,
 R = percentage of magnesium as calculated in step 2, and
 T = Na-K-Ca geotemperature in degrees Celsius.

The second geochemical method is based on the solubility of silica minerals along two-phase curves for water plus steam (Fournier, 1973, fig. 2). In general, quartz seems to control the dissolved-silica concentration in aquifers that contain ground water having temperatures greater than 150°C and possibly greater than 110°C (Fournier, 1973, p. 124; Truesdell, 1976, p. 59). When ground-water temperatures are less than 100°C and possibly less than 180°C, chalcodony may control the dissolved-silica concentration (Fournier, 1973, p. 133; Truesdell, 1976, p. 59). Because of the significant amount of overlap in the temperature ranges, both quartz-silica and chalcodony-silica geotemperatures were considered in this analysis. The curves used in the analysis assume that the geothermal water cools entirely by conduction of heat into the surrounding environment because there is no indication that boiling ground water or steam is escaping at the surface in Verde Valley.

The equations that describe the curves are:

$$T_{\text{QS}} = -9.40711 + 426.1057(SiO_2/100) - 652.4019(SiO_2/100)^2 + 612.8814(SiO_2/100)^3 - 315.15123(SiO_2/100)^4 + 82.15363(SiO_2/100)^5 - 8.48271(SiO_2/100)^6,$$

where

T_{QS} = the quartz-silica geotemperature in degrees Celsius, and
 SiO_2 = silica concentration in milligrams per liter, and

$$T_{\text{CS}} = \frac{1032}{4.69 - \log SiO_2} - 273,$$

where

T_{CS} = the chalcodony-silica geotemperature in degrees Celsius, and
 SiO_2 = silica concentration in milligrams per liter.

The quartz-silica equation is effective to about 280 mg/L SiO_2 , and the chalcodony-silica equation is effective to about 500 mg/L SiO_2 .

The accuracy of the geothermometers is dependent on certain basic assumptions that must be valid to insure the reliability of the calculated temperatures. The assumptions are (1) temperature-dependent reactions occur at depth, (2) all constituents involved in a temperature-dependent reaction are sufficiently abundant, (3) water-rock equilibration occurs at reservoir temperature, (4) negligible change occurs as the water flows from the heat reservoir to the surface, and (5) the hot water from deep in the system does not mix with cooler shallow ground water (Fournier and others, 1974, p. 259). The first three assumptions generally hold true for the Na-K-Ca and silica geothermometers (Renner and others, 1975, p. 51). The fourth assumption probably is not correct for natural systems, for example, in Verde Valley, the geothermal water may be taking place, especially with respect to calcium and magnesium. In the area between Cottonwood and Camp Verde chemical analyses indicate that the water is a calcium magnesium bicarbonate type, which is consistent with the composition—carbonate and evaporite deposits that are rich in calcium and magnesium—of the main aquifer (Madell, 1972, p. 43 and 72). Dissolved calcium in excess of the amount expected at the source

water-rock interface produces erroneously low temperatures when calculated by the Na-K-Ca geothermometer (Fournier and Truesdell, 1973, p. 1273). As water moves into the upper part of the aquifer, the concentration of dissolved magnesium by reequilibration increases to extremely high magnesium temperature corrections to the Na-K-Ca geotemperature, which again produces erroneously low Na-K-Ca geotemperatures. Therefore, the use of the Na-K-Ca geothermometer in Verde Valley is tenuous at best. The fifth assumption frequently is invalid; however, Swenberg and others (1977, p. 7) indicated that it is sometimes possible to recognize a geothermal water even though it may have migrated into a cooler shallow aquifer. The geothermal water will cool faster than it will lose its geothermal chemical identity because the speed at which reequilibration occurs becomes slower at lower temperatures, especially during silica reequilibration, which becomes extremely slow in the low-temperature range—less than 100°C (Fournier, 1973, p. 123).

In Verde Valley the ground water should have average quartz-silica geotemperatures between 60° and 75°C (Swenberg and others, 1977, fig. 12). The average silica concentration of 109 background samples was 32 mg/L, which gives a quartz-silica geotemperature of 65°C and a chalcodony-silica geotemperature of 35°C. The Camp Verde and Verde Hot Springs areas, which include about 120 mi², contain ground water that has anomalously high quartz-silica and chalcodony-silica geotemperatures. The geothermal water will cool faster than it will lose its geothermal chemical identity because the speed at which reequilibration with amorphous silica and evaporation by loss of steam from boiling water are the major factors in the concentration of silica. Because there is no indication of boiling water in Verde Valley and because the main aquifer consists mostly of carbonate and evaporite deposits, the silica anomaly may indicate a silica equilibrium with a source rock other than the main aquifer. Therefore, the geotemperatures calculated using the silica geothermometers are more credible than the geotemperatures calculated using the Na-K-Ca geothermometer.

The Verde Hot Springs area is the focal point of the intersection of three major lineaments, as shown by Haman and others (1978). The north- to south-trending lineament is an extension of the Oak Creek fault. Verde Hot Springs has the highest measured temperature, 39°C, in Verde Valley and a quartz-silica geotemperature of 114°C. Sheep Bridge Spring, which is 20 mi south along the extension and outside the map area, has a measured temperature of 36°C and a quartz-silica geotemperature of 89°C. Soda Springs, which is 20 mi north of Verde Hot Springs, was designated as a hot spring by Haman and others (1978) and has a measured temperature of more than 30°C. Castle Hot Springs, which is outside the map area, is along the lineament that extends from Soda Creek southwest to the springs and also is designated as hot by Haman and others (1978). The northwest- to southeast-trending lineament coincides with the Verde fault, and the anomalous geotemperature, indicated by the silica geothermometers closely parallel the lineament.

Fluoride and boron also are used as qualitative indicators of geothermal activity (Swenberg, 1975, p. 180). In the Camp Verde and Verde Hot Springs areas, the average fluoride and boron concentrations are 1.18 and 0.83 mg/L, respectively, which are significantly above the average concentrations of 0.28 mg/L fluoride and 0.17 mg/L boron in the rest of Verde Valley. The occurrence of high fluoride and boron concentrations in the areas of high silica geotemperatures may be the result of equilibration with the large evaporite deposits.

Renner and others (1975, p. 7) divided hot-water convection systems into three temperature ranges: (1) more than 150°C (the systems may be considered for generation of electricity), (2) 90° to 150°C (the systems are attractive for space and process heating and cooling), and (3) less than 90°C (the systems generally are used only for heat in local favorable circumstances in the United States). The estimated silica geotemperatures in the Camp Verde and Verde Hot Springs areas range from 74° to 135°C; however, these geotemperatures are low if precipitation of silica or dilution of geothermal water by shallow ground water is taking place. If the silica geotemperatures accurately depict geothermal conditions at depth, the ground water in the Camp Verde and Verde Hot Springs areas is in the low- to medium-temperature range. The geochemical evidence to substantiate a geothermal resource in Verde Valley is not conclusive, and further work, including test drilling, should be done to verify the silica geotemperatures given in the map.

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MAP SHOWING POTENTIAL GEOTHERMAL-RESOURCE AREAS, AS INDICATED BY THE CHEMICAL CHARACTER OF GROUND WATER, IN VERDE VALLEY, YAVAPAI COUNTY, ARIZONA