GROUNDWATER RESOURCES OF NORTHERN CALIFORNIA: AN OVERVIEW

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INTRODUCTION

California has substantial water resources; however, as more and more people compete for these resources, coordinated utilization and planning on a regional scale is becoming increasingly critical. Groundwater is a particularly important component in the water supply/demand equation, and continues to be an attractive source of water for individual farmers, agro-businesses, rural homeowners, land planners, and water purveyors. We geologists compete with the dowsers and experienced drilling outfits in the tasks of exploration and design of suitable wells for groundwater extraction; however, we take on almost the full burden of solving the problems generated by groundwater development, such as overdraft and progressive depletion of storage, declining water tables, deterioration of quality in freshwater aquifers due to seawater encroachment, and land subsidence due to the compaction of the underlying water-bearing sediments. In addition, we overlap considerably with civil and geotechnical engineers, land planners, and business managers insofar as groundwater basin management and protection are concerned.

This paper is intended to be a point of introduction to the hydrogeology of the major groundwater basins of Northern California (Figure 1), organized loosely by geologic province. The summaries presented are necessarily selective, as the database of California hydrogeology is vast. Hopefully, this overview will familiarize the new generation of hydrogeologists and engineering geologists with some of the basic, "classic" references that shaped much of our current understanding of California hydrogeology. In keeping with the theme of the volume, I have tried to include pertinent information on practical approaches to exploration and groundwater basin management. The interested reader will also want to refer to the excellent summaries prepared by Thomas and Phoenix (1983), and Planert and Williams (1995).

In preparing this summary I have drawn heavily on the work of the U.S. Geological Survey (USGS), the California Department of Water Resources (DWR) and the California Department of Conservation's Division of Mines and Geology (DMG), which, in my opinion, are three of the foremost geologic and hydrogeologic agencies in the world. Those of us that work in California are lucky to be able to stand on the shoulders of giants!

THE CENTRAL VALLEY

The Central Valley is the largest groundwater basin in the state, not only in terms of total storage capacity, but also in terms of its high utilization rate. A conservative estimate for the year 1995 puts the annual extraction rate at 9,000,000 acre-feet (~11 km^3), largely to support California's foremost industry—agriculture (DWR, 1998).

The basin is recharged by direct precipitation and infiltration along the beds of the San Joaquin and Sacramento river systems (which in turn receive most of their discharge from rainfall and snowmelt in the Sierra Nevada). Major withdrawals are through evapotranspiration, subflow into the Sacramento delta, and pumping. With regard to management of the groundwater resource, Bertoldi et al. (1991) have noticed that development of

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the resource has greatly modified the total flow through the basin, increasing from an estimated 2 million acre-feet in the pre-development period (early 1900s) to nearly 12 million acre-feet in the late 1980s. In the mid 1980s. the total volume of fresh water storage in the upper 1,000 feet of the aquifer system was estimated at about 800 million acre-feet, with an estimated annual net-depletion rate of 800,000 acre-feet.

Stratigraphy and structure. The stratigraphic and structural setting of the fresh groundwater basin has been conveniently summarized by Page (1986). The valley is a synclinal trough that has a surface area of about 20,000 square miles (Figure 1). It is bound to the west by the Coast Ranges, and to the east by the Sierra Nevada, and is often divided into four areas: (1) A northern basin drained by the Sacramento River and its tributaries. (2) The delta region, where the Sacramento, American, and San Joaquin rivers join to empty into Suisun Bay (and from there into the bays of San Pablo and San Francisco). (3) A meridional basin drained by the San Joaquin River. (4) The southernmost Tulare basin, which under natural conditions had internal drainage into the now vanished Tulare Lake. A good percentage of the fluvial inflow into the Tulare basin is diverted toward agricultural irrigation, and the balance empties into the Central Valley aqueduct network.

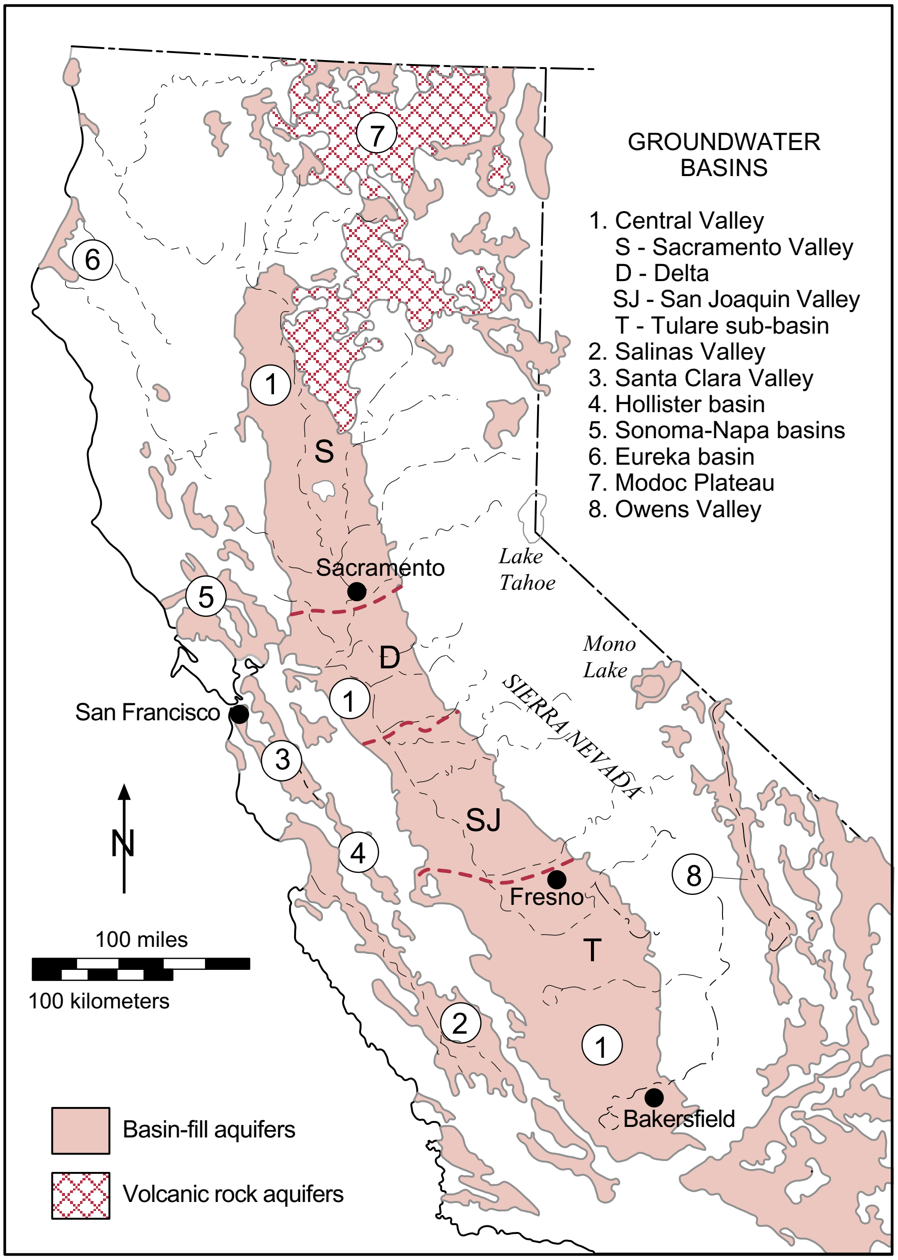

Figure 1. Main groundwater basins in Northern California (modified from Planert and Williams, 1995).

The trough is filled with a thick sequence of late Cretaceous to Holocene sediments, which at the axis of the syncline vary in thickness from 50,000 feet in the north to 30,000 feet in the south. The sedimentary section decreases in thickness toward the margins of the valley and, eventually, pinches out against the metamorphic foothills of the Sierra Nevada or against the fault-bound Franciscan basement of the Coast Ranges. Fresh groundwater aquifers are most often found in post-Eocene units, and the highest yields are obtained mainly from aquifers hosted by Miocene to Holocene units such as the Miocene Mehrten Formation; the Pliocene/Pleistocene Tuscan, Tehama, Laguna, Kern River, and Tulare formations; and the Pleistocene/Holocene Victor, Turlock Lake, Riverbank, and Modesto formations. The Appendix contains summary descriptions and key bibliographic references to these formations.

Hydrogeology. When describing the aquifers of the Central Valley, it has been traditional to regard the Sacramento Valley basin as having a single unconfined aquifer, and the San Joaquin Valley basin as having an upper unconfined aquifer, an intervening aquitard (the Corcoran Clay), and a lower confined aquifer. This simplified conception is adequate for general description purposes, but Williamson et al. (1989) have convincingly argued that the continental deposits of the Central Valley form, in fact, a single heterogeneous aquifer system, in which lateral and vertical differences in hydraulic conductivity lead to local variations in the degree of aquifer confinement. Consequently, the exploration hydrogeologist should not be surprised to find only trivial head differences across the Corcoran Clay in west Fresno County, but a couple hundred feet difference across some of the minor clay lenses in Kings County. At depth, the freshwater aquifer boundary is "defined" by salinity contents higher than 2,000 milligrams per liter (obviously an arbitrary boundary, but a convenient one for defining the groundwater resource; Page, 1986, Plate 3). The base of the fresh water aquifer lies at an average depth of 3,000 feet in the southern San Joaquin Valley, 1,000 feet in the northern San Joaquin Valley, 200 to 2,000 feet in the Delta area, and 1,500 to 3,500 feet in the Sacramento Valley.

Williamson et al. (1989) reconstructed the likely configuration of the uppermost equipotential surface of the aquifer at the turn of the century, based

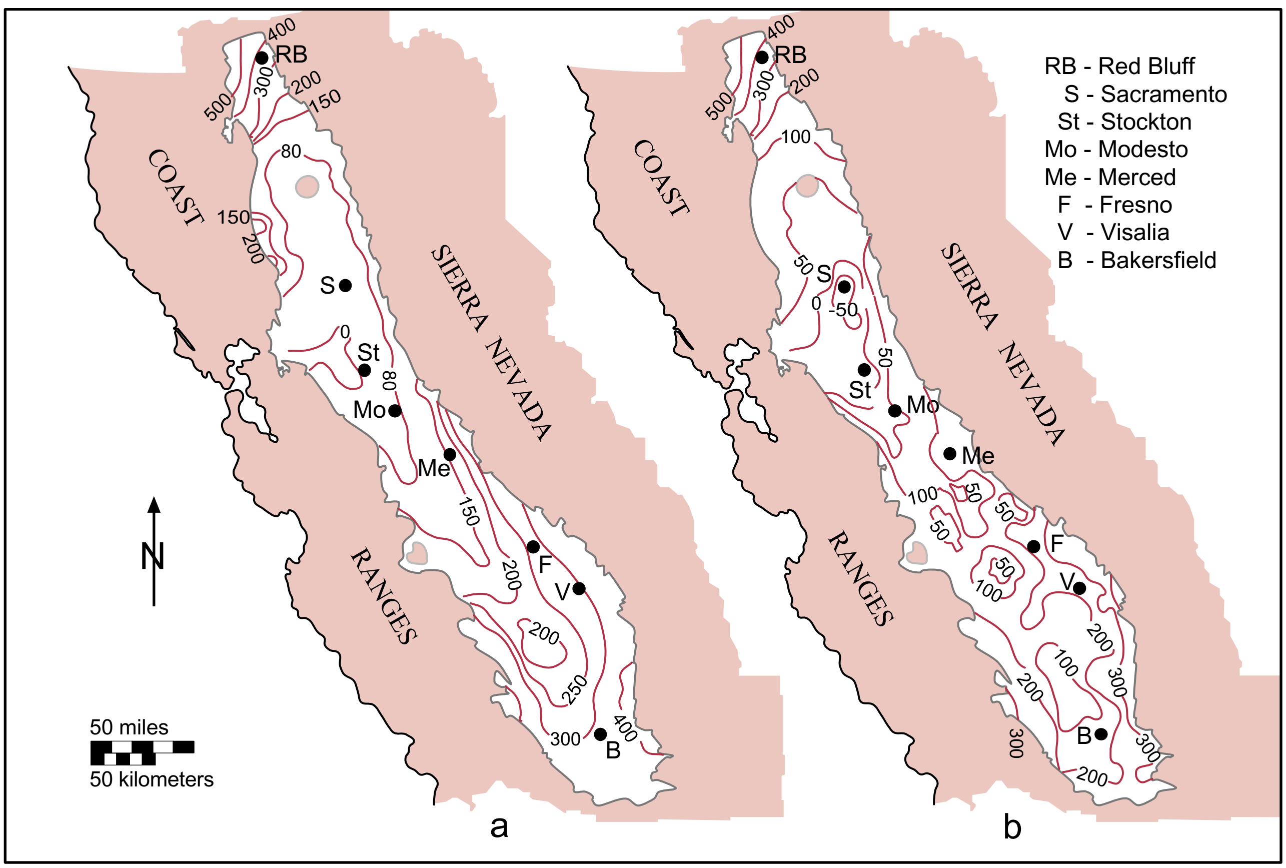

Figure 2. Elevation of the water table throughout the Central Valley. (a) Estimated configuration of the water table at the turn of the century (modified from Williamson et al., 1989). (b) Configuration of the water table in 1997. The contours for the San Joaquin Valley are based on data from DWR (1997a); the contours on the Sacramento Valley are loosely controlled from the data of DWR (1993, 1994, and 1997b).

on the data and observations of Hall (1889), Bryan (1915), and Mendenhall et al. (1916) (Figure 2a). At that time, groundwater generally moved from recharge areas in the higher ground at the edges of the basin toward its topographically lower axis, and from there to its discharge point at the Sacramento delta (or to the secondary discharge area of Tulare Lake). Prior to the development of large-scale agriculture and irrigation pumping, evapotranspiration along the many *tules* (marshes) that covered the floor of the valley was a major mechanism of aquifer discharge (*tule*, from the Nahuatl *tulli*, was the word used by the native inhabitants to describe the bulrushes growing in the saturated soils of the Central Valley). Otherwise groundwater was discharged to the streams as base flow, and eventually lost to the delta. Quite simply, the groundwater basin was "full to capacity".

Heavy pumpage from wells during the last 60 years has changed considerably the geometry of the equipotential surface. As shown in Figure 2b, water table levels have dropped considerably between Sacramento and Stockton, and deep, regional depression cones have formed north of Fresno and Bakersfield. Because of the lowering of the water table most of the *tule* marshes have disappeared, evapotranspiration losses of recharge water have become practically insignificant, and the rivers recharge the basin throughout most of their lengths. These changes have certainly modified the environment of the Central Valley; however, from the standpoint of utilization of the resource, groundwater extraction has certainly reduced the loss of precipitation and snowmelt water into the delta (at the same time that it has furthered the development of California's prime industry—agriculture).

Water quality. Bertoldi et al. (1991) summarized relevant issues regarding the quality of the groundwater resources. In general, the high level of recharge from Sierra streams contributes to lower total dissolved solids levels in the eastern portion of the basin, whereas groundwater in the western half of the basin has consistently higher salinities. Concentrations of dissolved solids are generally lower in the northern half of the Central Valley than in the southern part, perhaps due to the fact that marine sediments with saline connate waters form a larger proportion of the stratigraphic column in the San Joaquin Valley. (The last marine sediments in the Sacramento Valley are Eocene in age, whereas the southern half of the basin was the locus of localized marine deposition as recently as the Pliocene.) The basin is also subject to seawater intrusion in the Sacramento delta area.

Engineering geology. Shallow water table levels are of concern to civil works, particularly in the Sacramento delta area. For example, some portions of the delta are susceptible to liquefaction under seismic loading, largely due to the low consolidation of the sediments and shallow depths to the zone of saturation. Drainage of water-logged soils for agriculture has also led to oxidation of peat and subsidence. Levees have been built to protect the sinking ground from tidal flooding, but this in turn has created a potential hazard for catastrophic flooding in the event of levee failure.

Particularly high rates of groundwater extraction in the period between the early 1940s and the mid 1970s triggered extensive subsidence throughout the southern and central portions of the Central Valley. Poland et al. (1975) did an extensive study of the subsidence problem and concluded that nearly one-half of the area of the San Joaquin Valley (approximately 5,200 square miles) had been affected by subsidence, with as little as 1 foot of settlement in the less affected areas and as much as 30 feet in the Los Baños-Kettleman Hills area, 12 feet in the Tulare-Wasco area, and 10 feet in the Maricopa-Arvin area. Poland et al. (1975) were able to correlate the magnitude of subsidence with the local pumping rates and with the presence of thick lenses of montmorillonitic clays in the local stratigraphic column. Pumping-related subsidence has created agricultural drainage problems, has compromised roads and railroad tracks, and has triggered large expenditures for maintenance of the California Aqueduct (Poland et al.,1975). A recent review of the problem of subsidence in the San Joaquin Valley can be found in the paper by Swanson (1998).

A larger volume of surface water imports during the late 1970s and decreased rates of extraction during the last 20 years have contributed to a virtual cessation of subsidence in the Central Valley.

THE COAST RANGES

The narrow and elongated ranges and valleys of this province have a predominant north-northwesterly trend. Reed (1933) was the first to emphasize the presence of two types of "basements" within this province. Between the San Andreas fault and the Central Valley, the oldest exposed rocks belong to the Franciscan Series (Lawson, 1895), a subduction zone assemblage with a predominance of graywackes, submarine lavas, and serpentinites that have experienced low-temperature, high-pressure metamorphism. Between the San Andreas and Nacimiento faults (the Salinian Block of Compton, 1966), in contrast, the basement is formed by gneisses, schists, quartzites, marbles, and granulites that have been intruded by plutons of quartz diorite, granodiorite, adamellite, and granite. Finally, west of the Nacimiento fault and all the way to the coast, the basement is once again formed by Franciscan rocks. Clearly, the San Andreas and Nacimiento faults are major zones of structural discontinuity that have juxtaposed significantly different terranes.

A thick blanket of Upper Cretaceous and Cenozoic clastic rocks covers the basement rocks and bears record of intermittent but persistent crustal deformation. Folds, thrust faults, steep reverse faults, and strike-slip faults developed as a consequence of Cenozoic deformation, some of which continues to date (Page, 1966). Some of these deformed units host small-volume aquifers that can adequately supply small rural communities, but the truly significant aquifers of this province are found in Pliocene to Recent alluvial fill of active of active structural basins such as the Salinas Valley (a faulted synclinal structure between the Nacimiento and San Andreas faults) and the Santa Clara Valley (between the San Andreas and Hayward faults). Smaller groundwater basins such as the Sonoma and Napa valleys north of San Francisco Bay, or the Eureka basin in Humboldt County, have been formed by fluvial aggradation processes.

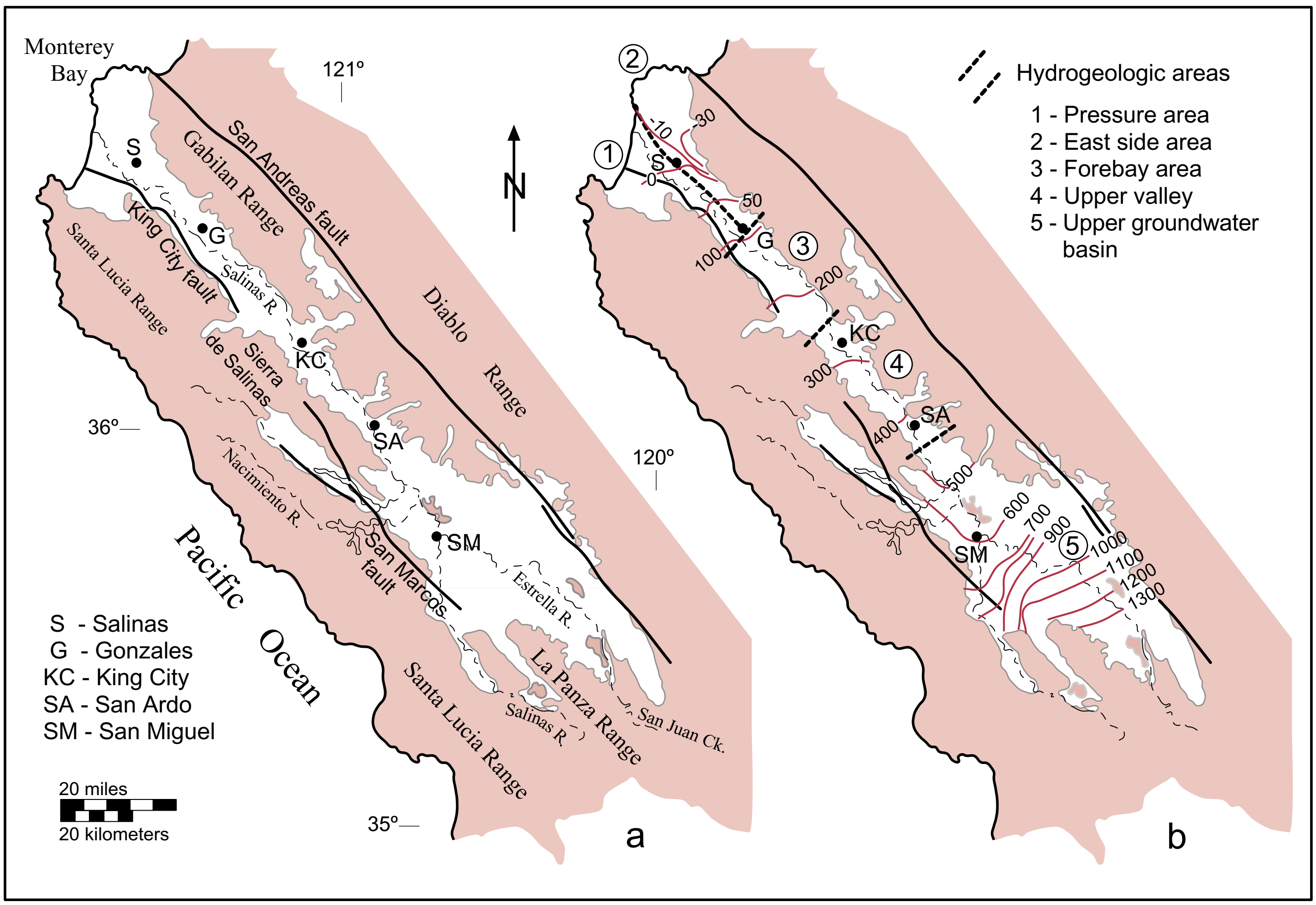

Figure 3. (a) General map of the Salinas Valley (modified from Planert and Williams, 1995, and from Durbin et al., 1978). (b) Hydrogeologic subdivisions of the Salinas groundwater basin, and general configuration of the water table in the early 1980's (modified from Planert and Williams, 1995, and from Showalter et al., 1983).

The Salinas Valley

Stratigraphy and structure. The Salinas Valley (Figures 1 and 3) is an elongated, intermontane fluvial valley formed by the Salinas River and its tributaries. The Salinas Valley groundwater basin has been divided into two sub-basins by Planert and Williams (1995). The first, called the upper basin, extends from the headwaters of the Salinas River and its tributaries (e.g., the Estrella River) to San Ardo. The groundwater resources of this basin have not been widely developed, so its hydrogeology is not well understood, but the unconsolidated deposits are reported to be as much as 1,750 feet in the alluvial deposits of the Estrella River. The second, lower basin, extends from San Ardo to Monterey Bay. The alluvial prism of the lower basin is 70 miles long, about 3 miles wide at San Ardo, and 10 miles wide at Monterey Bay. The lower basin has an average thickness of 1,000 feet of saturated sediments, but locally the alluvial prism is as much as 2,000 feet thick. It supports a thriving agricultural industry, and its groundwater resources are being actively utilized.

On the southwest, the Salinas Valley is separated from the Pacific Ocean by the Santa Lucia Range, and from the Carmel River by the Sierra de Salinas. On the northeast, it is separated from the San Joaquin Valley by the Diablo Range, and from the Hollister basin by the Gabilan Range. The history of structural deformation of these ranges, and of the intervening structural trough, is complex (e.g., Durham, 1974), but it appears that the basement under the valley has either been folded downward, down-dropped by faults, or both. The trough was invaded by the sea during the late Cretaceous, and was the site of intermittent marine deposition until the Pliocene. The marine deposits span the spectrum from conglomerates through mudstones, including the cherts and diatomites of the Miocene Monterey Formation. The ocean retreated during the Plio-Pleistocene, and the older marine deposits were widely blanketed by the sandstones and conglomerates (and subordinated mudstone, fresh-water limestone, and lignite) of the Paso Robles Formation. Durham (1974) classified sediments younger than the Paso Robles as alluvium, and differentiated it into: (a) old alluvium associated with old land surfaces in the hills, (b) old alluvium in valleys and lowland areas, (c) modern alluvium in stream beds, (d) debris-flow material, and (e) dune sand (described as the Aromas Sand by Dibblee (1973) and Dibblee et al. (1979)).

Hydrogeology. Durbin et al. (1978) followed local usage and divided the lower basin into four areas: the Pressure Area, the East Side Area, the Forebay Area, and the Upper Valley Area (Figure 3b). Groundwater moves from one area to another, so they are not sub-basins in a hydrogeologic sense; rather, they are a pragmatic way of recognizing variations in the degree of confinement of the aquifer, or regional variations in the specific capacity of production wells.

The Pressure Area extends from about 6 miles offshore beneath Monterey Bay to Gonzales. In this area of estuarine deposition massive clay units underlie much of the area between Monterey Bay and Salinas, and divide the unconsolidated deposits into an upper aquifer (the so-called 180-foot aquifer) a lower aquifer (the 400-foot aquifer), and a deep aquifer (the 900-foot aquifer). As the name of the area implies, within its footprint the aquifers are confined. The East Side Area encompasses the area east of the line that joins Gonzales and Salinas, up to the base of the Gabilan Range. Groundwater under semi-confined conditions is found in sand and gravel lenses that are interbedded with thick deposits of fine-grained sediments. Finally, groundwater in the Forebay and Upper Valley areas is mostly unconfined.

Specific capacity values (i.e., yield of the well divided by the drawdown) are smallest in the northern end of the basin, and tend to increase to the south. Average values of specific capacity are 25 gal/min/ft for the East Side Area and 60 gal/min/ft for the Pressure Area. In contrast, the average specific capacity of wells in the Forebay Area is 100 gal/min/ft, and 150 gal/min/ft in the Upper-Valley Area. Durbin et al. (1978) caution that the specific capacities of individual wells within each area are quite variable.

Recharge to the lower basin is largely by infiltration along the channel of the Salinas River (~30% of total recharge) and its tributaries (~20%). The second major source of recharge is irrigation return water (~40%). The remaining recharge is contributed by direct recharge from precipitation over the valley floor, subsurface inflow, and seawater intrusion. Outflow from the basin is dominated by pumping (~95%) and evapotranspiration by riparian vegetation (~5%). DWR (1995) estimated basin inflow at 532,000 acre-feet per year, and basin outflow at 550,000 acre-feet per year. The Salinas Valley Water Project is currently being implemented by the

Monterey County Water Resources Agency to mitigate groundwater overdraft and seawater intrusion. The project includes mitigation measures such as construction or retrofit of recharge dams, protection of recharge areas, and injection of recycled water into the impacted aquifers.

Not much information has been published on the hydrogeology of the upper basin, along the headwaters of the Salinas, Estrella, and San Juan Rivers. Showalter et al. (1983) state that groundwater in deep deposits is confined but that shallow groundwater is largely unconfined.

The general direction of groundwater flow is down the valley, from the headwaters of the Salinas and Estrella rivers, to San Ardo, to Monterey Bay (Figure 3b). Between San Ardo and Monterey Bay the average hydraulic gradient is 0.001 ft/ft, closely following the gradient of the Salinas River. Locally, however, pumping depression cones have imparted a distinctive cross-valley gradient to the potentiometric surface. Such is the case at the latitude of Gonzales, where water levels on the northeast side of the valley are about 30 feet lower than on the southwest side, or at the latitude of Salinas, where the difference is as high as 60 feet. Water levels in much of the Pressure and East Side areas are below sea level during a large part of the year. As a result, at Monterey Bay the direction of groundwater movement is inland and seawater intrusion is occurring (DWR, 1975).

Water quality. According to Planert and Williams (1995), groundwater in the Salinas Valley basin is generally acceptable for most uses, with dissolved solid concentrations ranging between 200 and 700 mg/liter. Exceptions are the Bitterwater area in the upper basin (high boron and arsenic), San Lorenzo Creek (high sulfate due to dissolution of gypsum beds), and the area between Soledad and Salinas (organic pollutants and high nitrate concentrations as a result of industrial and agricultural activity).

As reported by et al. (1978), seawater intrusion has considerably degraded water quality in the Pressure Area. Seawater intrusion was first noted in the 1930s and led to abandonment of several wells screened in the 180-foot aquifer. This degradation led to development of the 400-foot aquifer, but by the late 1960s this aquifer was also being degraded. By 1970 seawater intrusion had extended about 4 miles inland in the 180-foot aquifer, and about 2 miles in the 400-foot aquifer. The problem continues to affect the area, and as of 1995 seawater intrusion had extended up to 6 miles inland.

Engineering geology. The Salinas Valley has not been strongly affected by subsidence due to withdrawal of groundwater or by the problems commonly associated to shallow water tables. Nevertheless, this area promises to be a focus of engineering geology activity during the first few years of the new millennium. The works associated with the Salinas Valley Water Project will no doubt be the ground where many young engineering geologists will learn their craft. The planning horizon for the project is the year 2030, and it includes:

1. Spillway modifications at Nacimiento Dam to accommodate the probable maximum flood event, thus allowing full use of the capacity of the dam.
2. Operation of the Nacimiento and San Antonio reservoirs during the spring and summer months, first to increase recharge through the Salinas River bed, and ultimately for downstream diversion.
3. Storage and use of recycled water from the Monterey County water recycling projects.
4. Diversion of the Salinas River for direct delivery to agricultural users or to be stored in a newly-constructed reservoir for urban use.
5. Construction of treatment and water-conveyance facilities.

San Francisco Bay—Santa Clara Valley groundwater basin

Stratigraphy and structure. San Francisco Bay and the Santa Clara Valley occupy a linear, northwest-trending intermountain structural depression in the Central Coast Ranges (Figures 1 and 4). The depression is bound by the Mesozoic marine formations and Franciscan assemblage of the Santa Cruz Mountains and the San Francisco Peninsula on the west, and the Franciscan graywackes and serpentinite bodies of the Diablo Range on the east. The structural depression itself formed in response to movement along the San Andreas fault across the Santa Cruz Mountains and the San Francisco Peninsula, the Hayward fault along the eastern edge of the trough, and the Calaveras fault in the Diablo Range.

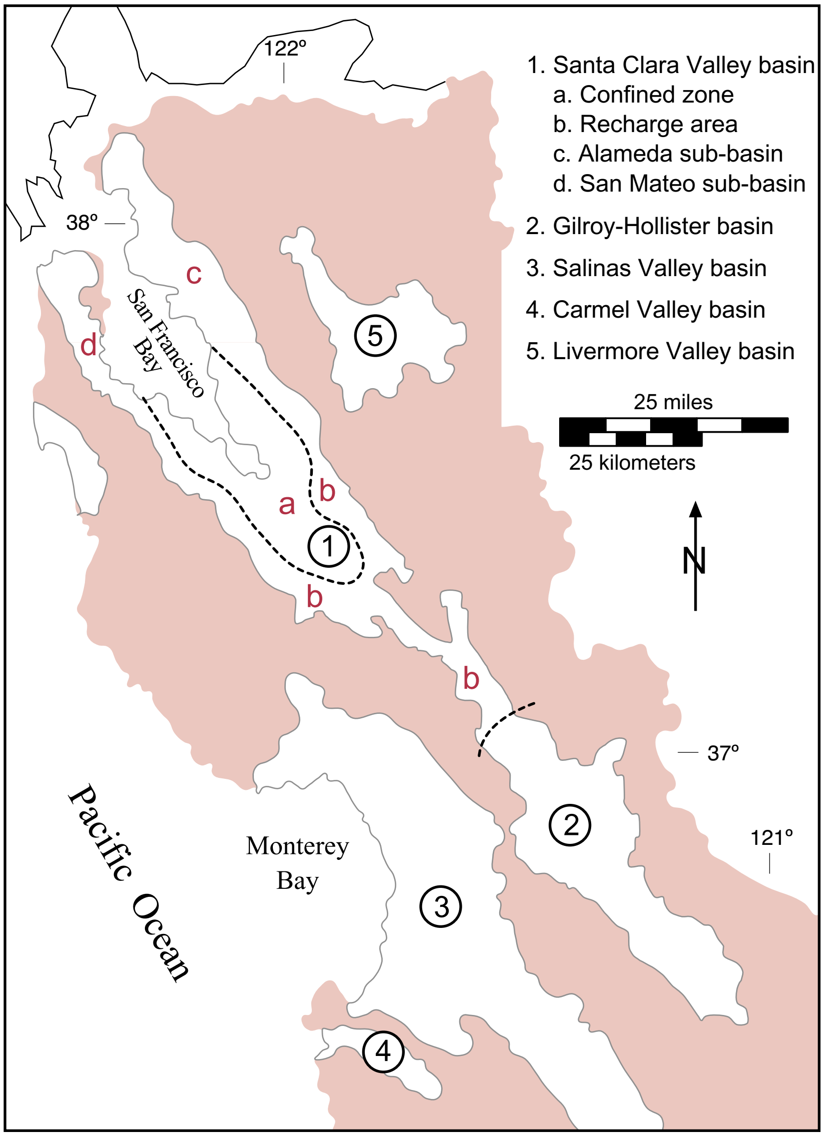

Figure 4. General map of the Bay Area, with hydrogeologic subdivisions (modified from Planert and Williams, 1995, Iwamura, 1995, and DWR, 1975).

Santa Clara Valley proper ends to the north against the San Francisco Bay, which occupies the northern end of the structural depression, but the alluvial fill “forks around” the Bay to merge with the plains and baylands of San Mateo on the west, and of Alameda on the east. The groundwater basin can thus be divided into five sub-basins (DWR, 1980): the Alameda Bay Plain and Niles Cone sub-basins east of San Francisco Bay, the San Mateo sub-basin west of San Francisco Bay, and the Santa Clara Valley and Coyote Valley sub-basins south of San Francisco Bay.

Of these five, the Santa Clara Valley sub-basin is in many regards representative of the whole structural trough. Stratigraphically, its alluvial fill is divided into older, lightly consolidated Plio-Pleistocene alluvium of the Santa Clara Formation (Bailey and Everhart, 1964; Dibblee, 1966; Cummings, 1972), and younger, unconsolidated Pleistocene-Holocene alluvium. The Santa Clara Formation does not yield large volumes of water to wells along the margins of the structural trough, but it is water-yielding toward the center of the basin, where it becomes indistinguishable from young alluvium in terms of lithology and consolidation (or lack thereof).

The younger alluvium was deposited as a series of coalescing alluvial fan deposits off from the surrounding mountain ranges. The sediments of the upper fan areas are coarse-grained, and form thick accumulations of permeable gravels and sands. The finer-grained, distal fan deposits interdigitate with shallow marine and tidal deposits of San Francisco Bay. Hence, the sedimentary deposits near the axis and mouth of the valley show distinctive stratification and have marked variations in vertical hydraulic conductivity. Many of the sand and gravel bodies found in the axis of the basin represent buried stream channels with limited lateral continuity. These sinuous channels apparently carved their courses across the interdigitating marine and distal fan deposits.

Hydrogeology. The hydrogeology of the Santa Clara Valley has been discussed recently by Iwamura (1995), who is the primary source of the following discussion. Other key references to the hydrogeology of the area are Clark (1924), and DWR (1967, 1975b, 1981).

For practical purposes, the aquifers of the Santa Clara Valley can be grouped into three hydrogeologic units: (1) a recharge area (called “forebay” by Iwamura, 1995), (2) an upper aquifer zone, and (3) a lower aquifer zone that hosts the primary drink-

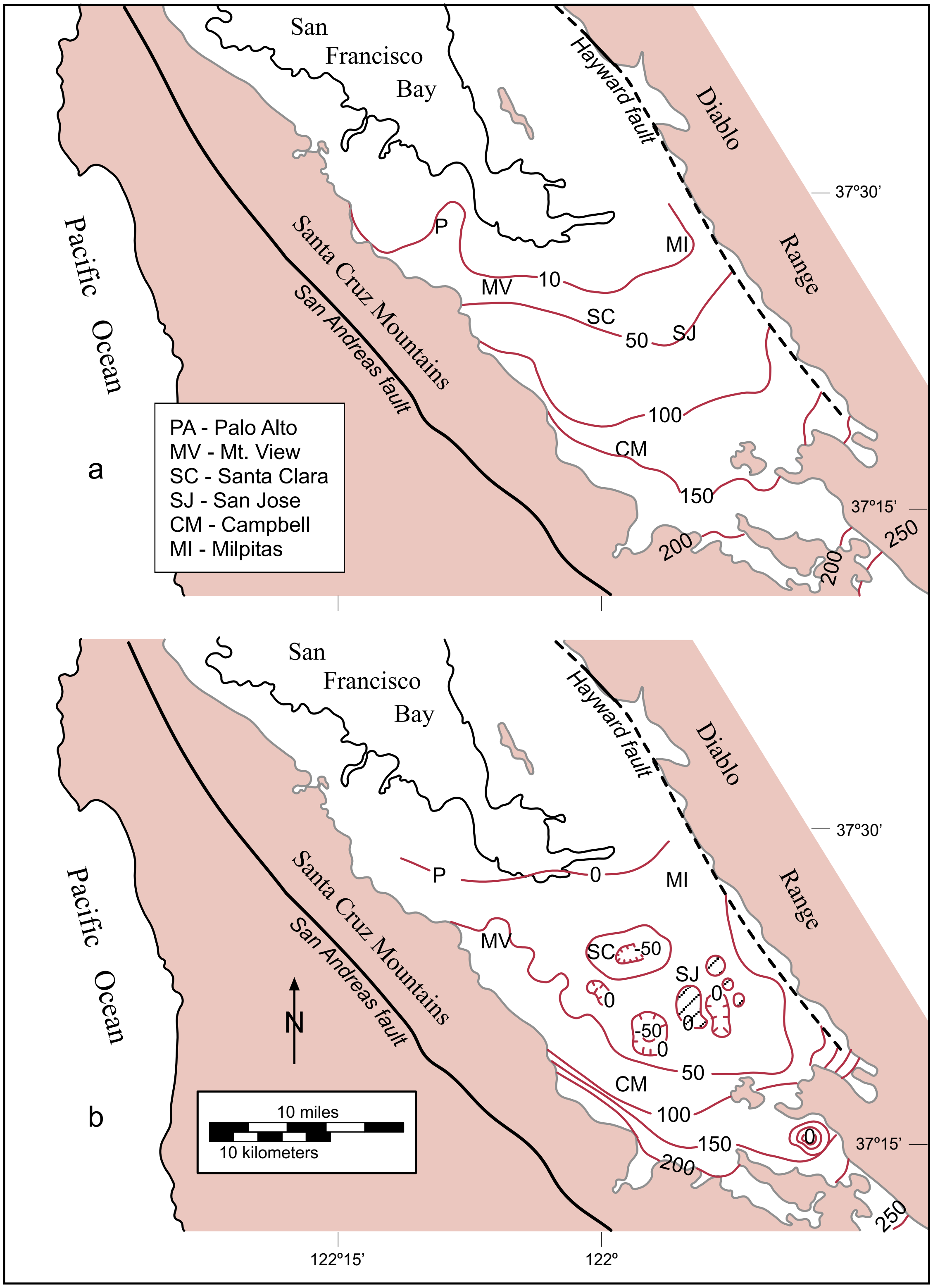

Figure 5. (a) Configuration of the water table throughout the Santa Clara Valley in the early 1900's (modified from Poland and Ireland, 1968 and from Planert and Williams, 1995). (b) Configuration of the water table throughout the Santa Clara Valley in September 1997 (modified from Moll, 1998).

ing water aquifers. The recharge area of the basin comprises the upper alluvial fan areas (including the foothills of Palo Alto/San Mateo on the west and of Alameda on the east). The sediments are highly permeable, though intervening leaky aquitards of small lateral extent are common.

Within the central portion of the basin the distal facies of the sloping alluvial fans become distinctively divided into discrete aquifers within a predominantly clayey section. The “upper aquifer zone” is the term used to group aquifers that occur within 150 feet from the surface. In contrast, the “lower aquifer zone” is the collective name given to aquifers that are deeper than 150 feet. The distinction is admittedly arbitrary, but it is also convenient because all the aquifers in the lower aquifer zone are confined. In the upper aquifer zone, groundwater is either unconfined or confined by leaky aquitards.

The two aquifer zones are separated by an extensive, thick, compacted aquitard that is essentially impermeable. The aquitard pinches out toward the medial portion of the alluvial fan apron, which enables recharge of both aquifer zones through lateral flow from the common recharge area in the upper reaches of the alluvial fans.

In the early 1900s, before significant development of the groundwater resource, groundwater flowed in a simple pattern, from the elevated recharge areas along the flanks of the basin toward San Francisco Bay (Figure 5a). Recent maps of the potentiometric surface (Moll, 1998) show a significant lowering of the water level and the hydraulic gradient in the urban area around San Jose (compare, for example, the change in the 50 foot elevation contour between Figures 5a and 5b). Wide depression cones have formed around the major pumping centers in the interior of the basin (e.g., around San Jose and Santa Clara in Figure 5b), and local recharge mounds have formed in response to local reinjection (e.g., hatched areas around San Jose in Figure 5b).

As previously mentioned, recharge to the basin occurs largely by infiltration from streams in the upper alluvial fan areas. However, return irrigation, infiltration of areal precipitation, and artificial recharge through the ponds operated by the Santa Clara Valley Water District are significant contributors to the inflow balance of the basin. Outflow is mainly by pumping withdrawal.

Many similarities exist between the hydrogeology of the Santa Clara Valley sub-basin and the Alameda Bay Plain, Niles Cone, and San Mateo sub-basins, particularly with regard to the upper alluvial fan areas. For example, in the vicinity of Niles and Hayward one can recognize the shallow Newark aquifer, and the deeper Centerville aquifer. Saline water has advanced 2-5 miles into the shallow Newark aquifer on a broad front, so this aquifer is not widely utilized. The deeper Centerville aquifer is a viable resource, although faulty and abandoned wells appear to have allowed downward leakage of salty water from the Newark aquifer at some locations (DWR, 1960a; 1968).

Before development started in the early 1900s, the Santa Clara groundwater basin was essentially full to capacity, and surface streams emptied their “rejected recharge” into San Francisco Bay (Iwamura, 1995). Thus, the first wells drilled found groundwater at very shallow depths or, if drilled into the lower aquifer zone, were naturally flowing artesian wells. As groundwater production increased, however, the water table declined in the upper aquifer zone, and the artesian pressure decreased. The decline of the water table also led to a reversal in the gradient of the upper aquifer zone in the San Francisco bayfront area, which in turn led to saltwater intrusion into the upper aquifer (Tolman and Poland, 1940). The water levels and pressures started to rise in the mid-1930s after construction of several artificial recharge reservoirs. Water levels dropped again between 1944 and 1965 in response to pumping overdraft, and the accompanying pressure reduction in the lower aquifer zone triggered localized seawater intrusion, albeit not as extensive as in the upper aquifer zone. Overdraft ceased in 1965, when pumping decreased in response to water imports from the State Water Project and Hetch Hetchy Reservoir.

Currently, most groundwater is pumped from either the confined lower aquifer zone or from the unconfined gravels of the recharge area. Aquifers of the upper zone are little used, in part because the agricultural industry has declined significantly in the valley, and in part because of the high salinity triggered by seawater intrusion. Furthermore, local contamination plumes of organic solvents and fuels have affected the upper aquifer zone.

Water quality. The aquifers of the Santa Clara Valley basin have been partially affected by seawater intrusion, rise of deep-seated connate waters with high contents of dissolved solids, nitrate and pesticide accumulation, gasoline and solvent leaks, and bacterial pollution due to poor disposal practices of sewage and refuse (Iwamura, 1980, 1995). The Santa Clara Valley Water District and the Regional Water Quality Control Board have given high priority to the investigation of these problems, and their remediation or containment will no doubt occupy the attention of a few generations of engineering geologists.

With respect to migration of contamination between the upper and lower aquifer zones, the intervening aquitard appears to be an effective barrier against natural migration. Unfortunately many wells are screened in both aquifers, so contaminant migration through the boreholes themselves, or through their gravel packs, is probably the main mechanism for dispersal of contaminants into the lower aquifer zone.

Engineering geology. The major claim to engineering geology fame of the Santa Clara Valley is the extensive subsidence that was induced by groundwater withdrawals between 1920 and 1969. Subsidence triggered a host of remedial actions, including levee construction along the bayfront and tributary stream banks to prevent inland encroachment of the waters of San Francisco Bay (Tolman and Poland, 1940). Construction of water conservation reservoirs in the mountainous watershed enhanced recharge of the aquifers, which led to a partial recovery of groundwater levels and pressures between 1935 and 1944, and to greatly diminished subsidence. The problem arose again in the mid 1940s. when increased pumping triggered anew the onset of subsidence, particularly in the area between southeast San Jose and downtown Mountain View. Poland and Ireland (1968) estimated up to 8 feet of subsidence for downtown San Jose for the 1945-1968 period, and a total aggregate subsidence of 13 feet for the 1935-1969 period (Poland, 1969). Recent reviews of the problem of subsidence in California can be found in the volume edited by Borchers (1998).

The import of water from the State Water Project and Hetch-Hetchy Reservoir led to decreased groundwater pumping and virtual cessation of land subsidence. Since water imports started, pressure in the lower aquifer zone has slowly increased, to the extent that some of the wells within the interior of the basin have recovered their artesian character.

Gilroy-Hollister groundwater basin

Stratigraphy and structure. In terms of stratigraphy and structure, the Gilroy-Hollister Valley is the southernmost extension of the Santa Clara and Coyote Creek Valleys (Figure 4). Topographically, the boundary between the two drainage basins is the apex of the Morgan Hill alluvial fan, which forms a drainage divide between the watershed of Coyote Creek (draining to San Francisco Bay), and that of the Pajaro River (draining to the Pacific Ocean). This drainage divide generally corresponds with the groundwater divide that forms the northwestern hydrogeologic boundary of the Gilroy-Hollister groundwater basin. The basin itself is hosted by the alluvial deposits of the headwater tributaries to the Pajaro River: the Llagas, Uvas, Pacheco, Tequisquita, and Santa Ana Creeks, and the San Benito River. The alluvial basin is about 20 miles long in the northwest direction and 6 miles wide, and is bound by the Franciscan of the Diablo Range to the northeast and faulted Tertiary sedimentary units to the southwest, within the zone of structural deformation of the San Andreas and Calaveras fault systems (Figure 6).

The thickness of the alluvial fill of the basin ranges between 500 feet at the Morgan Hill alluvial fan to more than 1,000 feet in the center and south of the basin, and 2,000 feet in the San Benito area. These thickness estimates include the underlying, slightly consolidated sands and gravels of the Santa Clara Formation and the San Benito Gravel. According to Kilburn (1972), the older sedimentary deposits have been subjected to several episodes of faulting and folding. The Holocene alluvium does not appear to be folded, but it is broken by currently active faults. The structural setting of the basin is complex: Two major faults, the San Andreas and Calaveras faults, dominate the tectonic fabric of the region, but a host of smaller faults cut the basin and act as groundwater barriers (e.g., the Park Hill West, Asuyama, Santa Ana Valley, Tres Pinos, and Bolado Park faults). Separating the Hollister and San Juan valleys are the faulted and folded sandstones of the Purisima Formation in the Lomerias Muertas and Flint Hills. For convenience, this zone of structural deformation is here referred to as the “Sargent anticline” (Figure 6).

Hydrogeology. From a hydrogeologic standpoint, the Gilroy-Hollister groundwater basin has been divided into four sub-basins (Kilburn, 1972; Iwamura, 1989): the Llagas sub-basin (a political

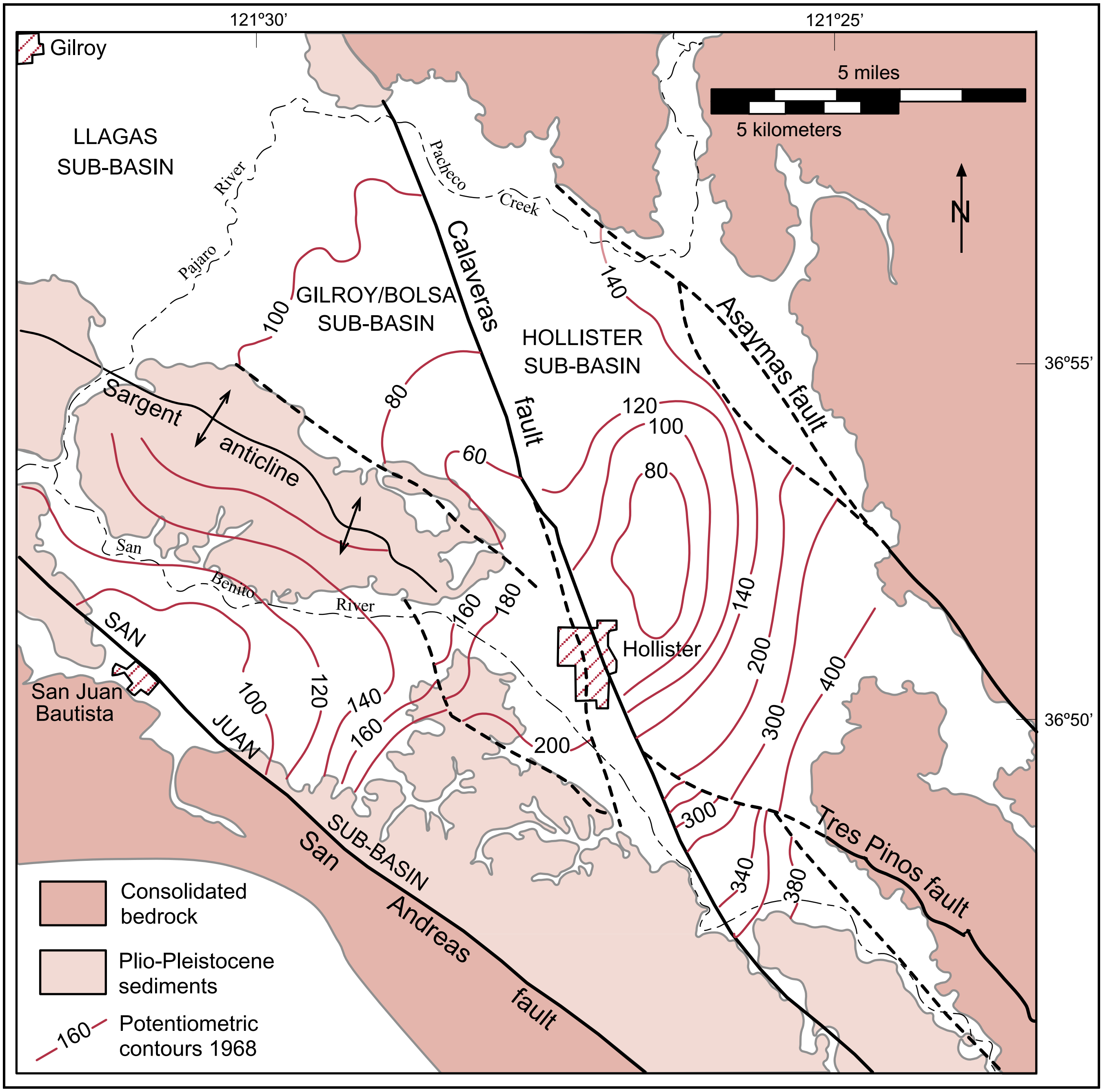

Figure 6. Map of the Gilroy-Hollister basin, with hydrogeologic subdivisions, and general configuration of the water table throughout the Gilroy-Hollister basin in 1968 (modified from Kilburn, 1972).

division used to refer to that part of the Gilroy-Hollister basin that is within Santa Clara County), the Gilroy-Bolsa sub-basin (which forms the natural extension to the southeast, into San Benito County, of the Llagas sub-basin), the Hollister sub-basin, and the San Juan sub-basin (Figure 6). Most of the sub-basin boundaries are defined by bedrock outcrops, the trace of the San Andreas or Calaveras faults, and the crest of the Sargent anticline (with the exception, of course, of the political boundary of Santa Clara and San Benito counties).

The combined Llagas-Gilroy-Bolsa sub-basin has a similar hydrostratigraphy to that found in the Santa Clara Valley groundwater basin (Iwamura, 1989): (1) A recharge area defined by the upper reaches of the alluvial fans, and (2) a flat interior portion where upper and a lower aquifer zones are separated by an intervening aquitard. The recharge area has the typical stratigraphy of the proximal facies of an alluvial fan, with coarse gravel units interbedded with minor fine-grained lenses. The upper aquifer zone consists of several aquifers interbedded with thin aquitards. The uppermost aquifer is unconfined, whereas the others are confined or partially confined. The top of the intervening aquitard is found at depths that vary between 20 and 100 feet, and its thickness ranges between 40 and 100 feet. Numerous individual confined aquifers occur in the lower aquifer zone. Wells tapping into the lower zone south of Old Gilroy used to be flowing artesian wells, but agricultural pumping has decreased the aquifer pressures, and only very few wells retain their artesian character.

At its southeastern terminus, the Llagas-Gilroy-Bolsa sub-basin is bound by the "V" formed by the Calaveras fault and the axis of the Sargent anticline. The Pajaro River entrenched itself into the anticline as the latter formed during the Pleistocene-Holocene, forming a narrow gap that provides for surface drainage of the basin. However, the alluvial fill in the gap is not deep enough to allow for significant underflow discharge out of the basin. Prior to development of the groundwater resource in the early 1900s) groundwater discharge was chiefly by upward movement into the bed of the Pajaro River and evapo-transpiration from the marshy land between the Pajaro and Tequisquita Rivers (Clark, 1924). The sub-basin has now been extensively developed with irrigated agriculture, and pumping has created a broad depression cone at the southeastern end of the basin, a couple of miles west of the Hollister municipal airport (Figure 6).

The Hollister sub-basin is bound by the Calaveras fault to the west, and the front-range faults of the Diablo Range (Asuyamas and Santa Ana Valley faults) to the east. The boundary faults are relatively impermeable, with piezometric levels varying by as much as 100 feet across the faults. According to Kilburn (1972), prior to development groundwater moved generally to the northwest from recharge areas in the southern and eastern sides of the basin. Discharge was through artesian flow into the streams and marshes in the northern half of the basin. Probably little groundwater flowed across the Calaveras fault into the adjacent San Juan and Gilroy-Bolsa sub-basins. With the onset of irrigation pumping the piezometric surface changed, however, and the pattern of groundwater flow is now toward a broad depression centered 3 miles northeast of downtown Hollister and toward a secondary depression cone in the structural sliver formed between the Asuyamas and Santa Ana Valley faults.

The San Juan sub-basin is bound by the Sargent anticline to the north, the Calaveras fault to the east, the Bird Creek Hills to the south (formed by outcrops of folded sandstones of the Purisima Formation), and the San Andreas fault to the west. Based on a compilation of water-level records, Kilburn (1972) distinguished two aquifers in the San Juan and Hollister sub-basins: (1) a semi-confined aquifer that extends to a depth of as much as 300 feet below ground surface, and (2) an underlying confined aquifer of undetermined thickness. The basin appears to receive most of its recharge from streambed infiltration along the upper reach of the San Benito River, with perhaps some underflow from the Hollister sub-basin across the Calaveras fault. Before basin development, the San Benito River appears to have been a losing stream east of Hollister, and a gaining perennial stream west of Hollister. Groundwater flow at the time was to the northwest, toward the confluence of the Pajaro and San Benito Rivers. This configuration has changed considerably now that the basin is being actively pumped. For one thing, the San Benito River is now a losing stream throughout most of its extent and for most of the year, so the basin is now recharged whenever the river flows (in contrast, before development the groundwater basin was "full", and any additional recharge was "rejected" in the form of subflow to the lower reaches of the river). The general groundwater flow direction continues to be to the northwest but, instead of reaching the Pajaro River, it now gravitates toward a broad depression cone that has developed just east of San Juan Bautista.

Livermore groundwater basin

Livermore Valley is an intermontane valley nestled in the heart of the Diablo Range. The Livermore groundwater basin is hosted by the alluvial fill of the valley, and is elongated in an east-west direction (Figure 7). To the northwest it has a narrow extension that follows the trend of the Calaveras fault (the Dublin-San Ramon sub-basin). To the southeast it merges with Sunol Valley, but the latter has been traditionally considered a separate groundwater basin. The Livermore groundwater basin encompasses a surface area of approximately 100 square miles (~65 mi^2 underlain by Quaternary alluvium and 35 mi^2 underlain by Livermore Formation).

Major cities within the basin are Livermore on the east end, Pleasanton on the west end, and San Ramon Village and Dublin in the narrow northwestern extension.

Livermore Valley is drained by Arroyo de la Laguna and its tributaries (starting from the north and proceeding clockwise South San Ramon Creek, Alamo Creek, Tassajara Creek, Arroyo Las Positas, Arroyo Mocho, and Arroyo Valle). At the mouth of Livermore Valley, Arroyo de la Laguna joins Alameda Creek, which flows southerly through Sunol Valley (where it is impounded by Sunol Dam), cuts across the Diablo Range through Niles Canyon, and eventually reaches San Francisco Bay.

Stratigraphy. The Jurassic through Miocene formations that bound Livermore Valley are too indurated to host significant groundwater resources. The Pliocene Orinda Formation that crops out just north of Livermore Valley, and the Plio-Pleistocene Livermore Formation that crops out in the hills to the south are not indurated, and contain significant amounts of gravel and sand. Their yields are marginal and erratic, however, because of the relatively high proportion of fines. The Orinda Formation reaches a thickness of 9,000 feet north of the valley, whereas the Livermore can be up to 4,000 feet in thickness to the south. Locally, these units may play an important role in recharging the Quaternary alluvial aquifers.

The following summary of the Plio-Pleistocene geologic history of the area is largely based on Barlock (1988) and Andersen (1995) (see also Crittenden, 1951; Hall, 1958; DWR 1966, 1974; Dibblee, 1980a, 1980b): The Livermore intermontane basin between the Calaveras fault to the west and the Greenville fault to the east has been filled with continental detritus since the Late Miocene, in response to spasmodic Coast Range uplift. The lower portion of the Livermore Formation was deposited between 5 and 2.5 million years ago, by sand-dominated braided streams as a result of uplift in the Altamont Hills. 2.5 million years ago the sediments of the upper Livermore Formation record the uplift of the central Diablo Range and the development of an alluvial fan complex that reversed the direction of sediment transport. The braided streams and debris flows of this fan spreaded Franciscan detritus northward across Livermore valley. These deposits have been warped and tilted by Pleistocene deformation, so the Livermore Formation now dips 10 to 20° degrees to the north in the hills south of the valley.

The main aquifers of the Livermore basin are in Upper Pleistocene and Recent fluvial gravels and sands, which are collectively referred to as Quaternary alluvium. DWR (1966) “mapped” the changes in the paleogeography of the streams by contouring the proportion of sand and gravel in the 0 to 100, 100 to 200 and 200 to 300-foot depth intervals (the alluvium is as much as 500 feet thick along the axis of the valley). These contour maps, and a careful analysis of the stratigraphy of selected wells, have revealed a fascinating sedimentologic history: After northward tilting of the Livermore Formation in the mid-Pleistocene, streams draining the newly created highlands flowed north into the Livermore depression, crossed it from east to west while accumulating alluvium, and eventually flowed out through San Ramon Valley to empty into the ancestral Sacramento River into the area now occupied by Suisun Bay. The southern streams “cleaned up” gravels eroded from the Livermore Formation and spread them over nearly the entire floor of the valley. Gradually, great sheets of clean gravel accumulated as the streams worked their way back and forth over the valley floor.

The outlet to the northwest through San Ramon Valley seems to have been blocked from time to time, probably in response to rupture of the Calaveras fault and associated landsliding. At the times the outlet was blocked, the carrying capacity of the streams was reduced, and swamps and lakes formed in the western portion of the valley, so fairly continuous bodies of silt and clay accumulated on top of the previously deposited gravel layers. At least four thick layers separated by extensive gravel beds are known to be present in the western portion of the valley. The uppermost of these clay layers accumulated in recent time—a small remnant of the lake in which it accumulated persisted until the early 1900s in the area northwest of Pleasanton—and formed the 60-foot upper aquitard that extends to the ground surface.

The present stream outlet to the south through Sunol Valley was apparently established at the time when the uppermost gravel—called the upper aquifer—was being deposited. The upper gravel is considerably thicker in the southwest corner of the valley, parallel to the stream course of Arroyo de la Laguna, than it is in the San Ramon Valley. Accumulation of the upper aquifer gravel came to an

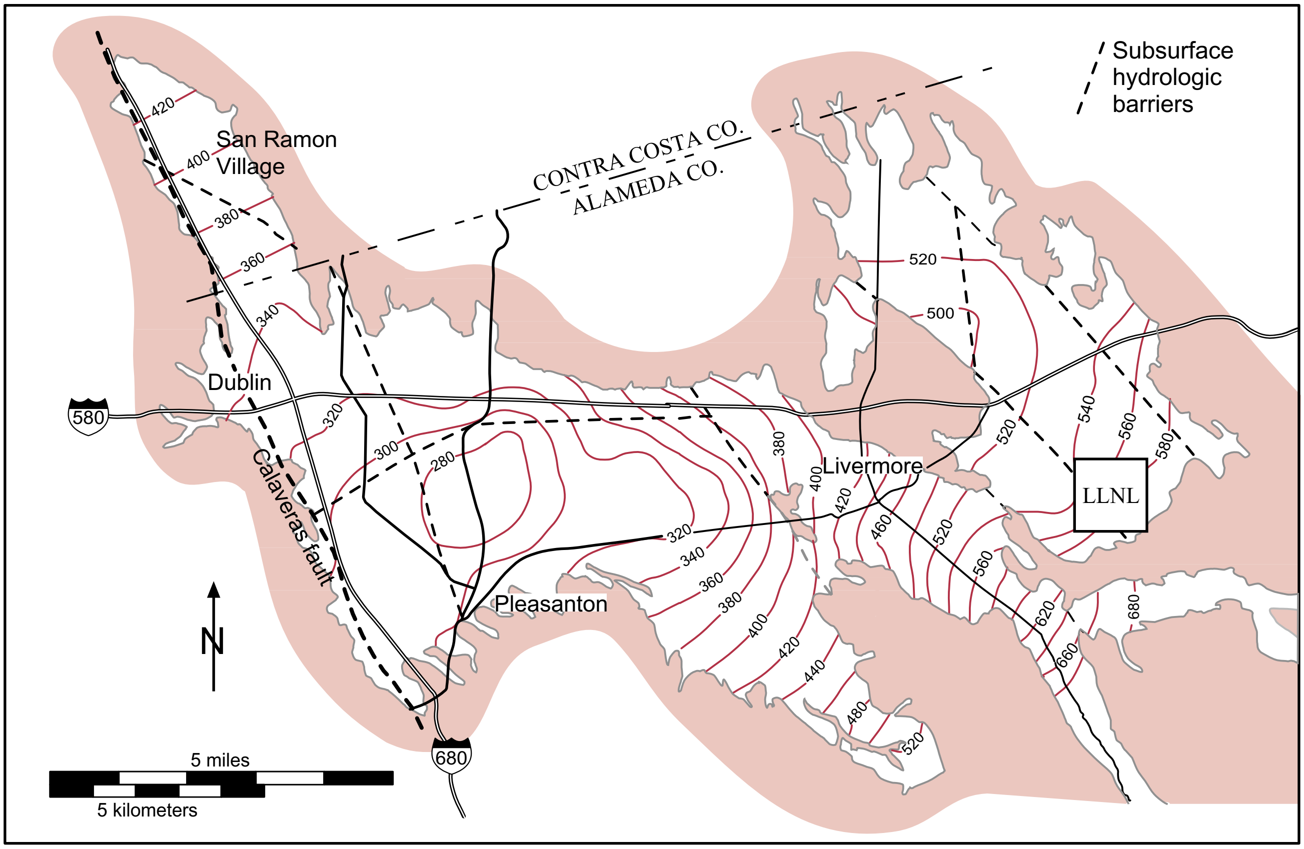

Figure 7. Configuration of the water table throughout Livermore Valley in October 1998 (modified from Gates, 1998). Thin unlabeled lines are major local roads for reference. LLNL marks the site of the Lawrence Livermore National Laboratory.

end suddenly when the lake referred to in the last paragraph formed.

Hydrogeology. As shown in Figure 7, the potentiometric surface of the upper aquifer(s) indicates that water moves from the periphery of the basin toward a depression cone located north of Pleasanton. Pumping in the Pleasanton area began as early as 1898, and large quantities of water have been produced ever since, both for local consumption and for exporting to the City of San Francisco.

Four distinct gravel aquifers can be recognized in the western and central portions of the basin, to an average depth of 400 feet. The highly permeable gravels are separated by four distinct clay aquitards. Originally, the upper aquifer was confined by the upper aquitard, and wells drilled into it had artesian flow. The pressure has declined considerably after nearly a century of pumping, however, so the upper aquifer is now a water-table aquifer. The deeper aquifers remain confined and are the ones from which most of the water is now pumped. The correlation of hydrostratigraphic units becomes less distinct to the east, but even as far as the Lawrence Livermore National Laboratory (LLNL on the far right of Figure 7) detailed stratigraphic work has revealed an alternance of hydrostratigraphic units of different permeability (LLNL, 1995).

DWR (1966) divided the basin into several subbasins, based on the presence of subsurface hydrologic barriers. These barriers are shown as dotted lines in Figure 7. They obviously do not have a marked effect in the potentiometric contours of the upper aquifer(s), but in the original work DWR (1966) reported differences of 10 to 50 feet in the levels of deep-screen wells on opposite sides of some of these barriers.

Typical yields from wells in the gravel aquifers range between 50 and 2,500 gallons per minute. The low values may reflect wells screened in the Livermore Formation, which is significantly less permeable than the Quaternary alluvium.

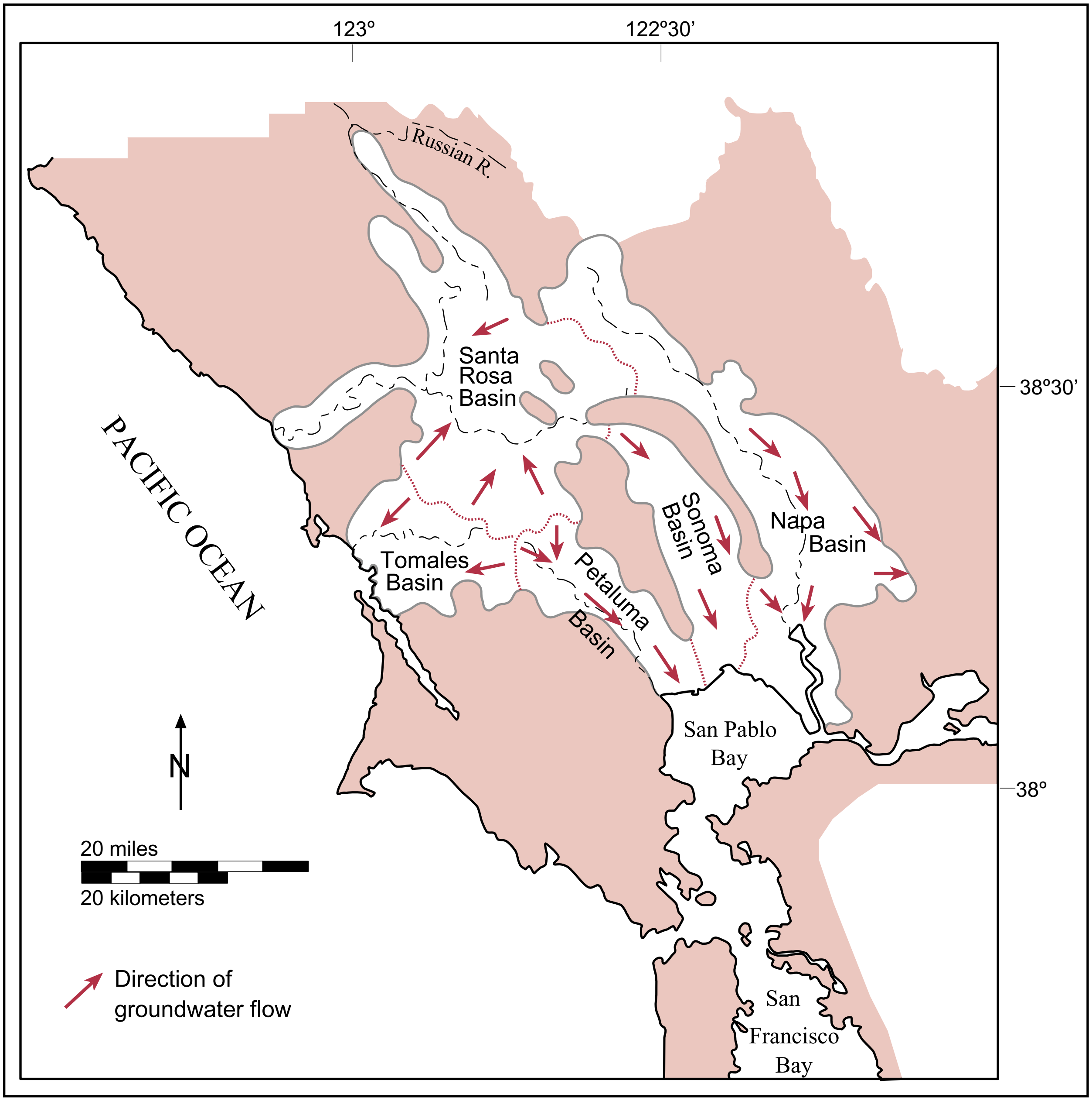

Figure 8. General map of the Petaluma, Sonoma and Napa Valleys (modified from Planert and Williams, 1995).

Water quality. Groundwater in the Livermore Valley is generally of good quality, although there are some areas in which relatively high contents of dissolved solids limit its domestic use (DWR, 1974). Pollution from point sources is a relatively minor concern in the west portion of the basin, where the upper aquiclude provides a significant amount of protection. On the east side of the basin, however, industrial pollution is of concern. The Lawrence Livermore National Laboratory has proved to be the source of several contaminant plumes of volatile organic compounds, hydrocarbons, chromium, and tritium, but active remediation efforts are ongoing and offsite plumes seem to have stabilized (LLNL, 1995).

The Petaluma, Sonoma, and Napa Valleys

The pleasant climate and fertile soils of the valleys north of San Francisco and San Pablo bays have made them a target for specialty agriculture (e.g., wine grapes) and for upscale rural residential development (Figure 8). This development has been aided by the availability of water from the Russian River for irrigation and urban consumption purposes. However, shallow wells are the basic source of water for rural residences, so thousands of shallow wells have been drilled in substrates that range from fractured Franciscan graywackes and Tertiary volcanic rocks to alluvial fill sediments. As in so many other California basins, the alluvial sediments have proved to be the most consistent target for groundwater exploration and development, even though the thickness of alluvial fill in these valleys is considerably less than in other valleys of the Coast Ranges (Cardwell, 1958; DWR, 1982a, 1982b). For example, DWR (1982c) reports an average thickness of only 250 feet for the alluvium of the Sonoma Valley.

The morphology of the Petaluma, Sonoma and Napa valleys appears to be less controlled by faulting than by volcanism, fluctuations in sea level, and normal fluvial processes. To be sure, the rocks that form the flanks of these valleys are cut by some prominent faults, but none of them seems to have promoted the accumulation of thick basin fills. Instead, volcanic activity during the Pliocene led to the formation of rhyolitic domes and andesitic/basaltic cinder cones, the emplacement of lava flows, and the accumulation of air-fall tuffs, ignimbrites, and volcaniclastic sediments (collectively known by the names of Sonoma Volcanics or Clear Lake Volcanics). These volcanic landforms controlled to some extent the development of drainage basins that became entrenched by fluvial erosion during periods of lowered sea level during the Pleistocene. The deepening of the valleys seems to have favored the accumulation of alluvial fan deposits along their flanks, now reflected in the Pleistocene gravels and sands of the Pleistocene Glen Ellen and Huichica Formations (Kunkel and Upson, 1960) (the sediments in these units are generally consolidated, so yields to wells can be quite low). Sea level rose during the Holocene, with the resulting encroachment of bay mud deposits into the valleys. In Sonoma Valley, for example, bay mud can be found as far inland as Schollville, 3 miles inland from the present shoreline of San Pablo Bay. The rise in sea level also triggered alluvial accumulation in the valleys, as the rivers adjusted their base profiles, eventually resulting in the fertile valley floors where wine grapes thrive today.

Because of the small volume of alluvial deposits, the storage capacity of the valleys is modest. For example, DWR (1982c) estimated the usable groundwater in storage in Sonoma Valley at about 472,000 acre-feet. Since the potential recharge is high, however, the basins have been able to support thousands of rural users without major depletion.

Maintenance of water quality should be a major consideration to major users, such as water companies, on three accounts. First, as in most coastal basins, these valleys are susceptible to salt water intrusion. Seawater has intruded into the pumped aquifers of the Petaluma, Sonoma and Napa Valleys, not by subsurface inflow from the bay, but by infiltration of surface water in tidal channels (Thomas and Phoenix, 1983). This problem is compounded by the encroachment of bay muds several miles inland from the present shoreline. These mud deposits contain entrapped saline water that "taints" the water chemistry of most wells near the shoreline. The second source of concern is thermal water associated with the volcanic centers of the Sonoma and Clear Lake Volcanics. The heat appears to be a remnant of the Pliocene pulse of volcanism, and meteoric groundwater that comes in close proximity to some of the Pliocene volcanic centers becomes hot enough to dissolve some of the ions contained in the surrounding rocks. The result is thermal water with high contents of boron, sodium, and total dissolved solids, which could conceivably impact the low salinity water characteristic of the valleys (and of many wells screened in fractured volcanic rocks). Third, the valleys are subject to intense agricultural exploitation, and this industry often requires the use of fertilizers and pesticides. No extensive contamination by these compounds has been recognized to date, but users of the basin are aware that they pose a latent threat.

THE BASIN AND RANGE PROVINCE

Owens Valley

Owens Valley is a narrow, north-trending graben bound by the Sierra Nevada to the west and the White and Inyo Mountains to the east. The valley extends for about 200 miles from the Nevada border just north of Mono Lake south to Haiwee Reservoir. A drainage divide south of the reservoir separates

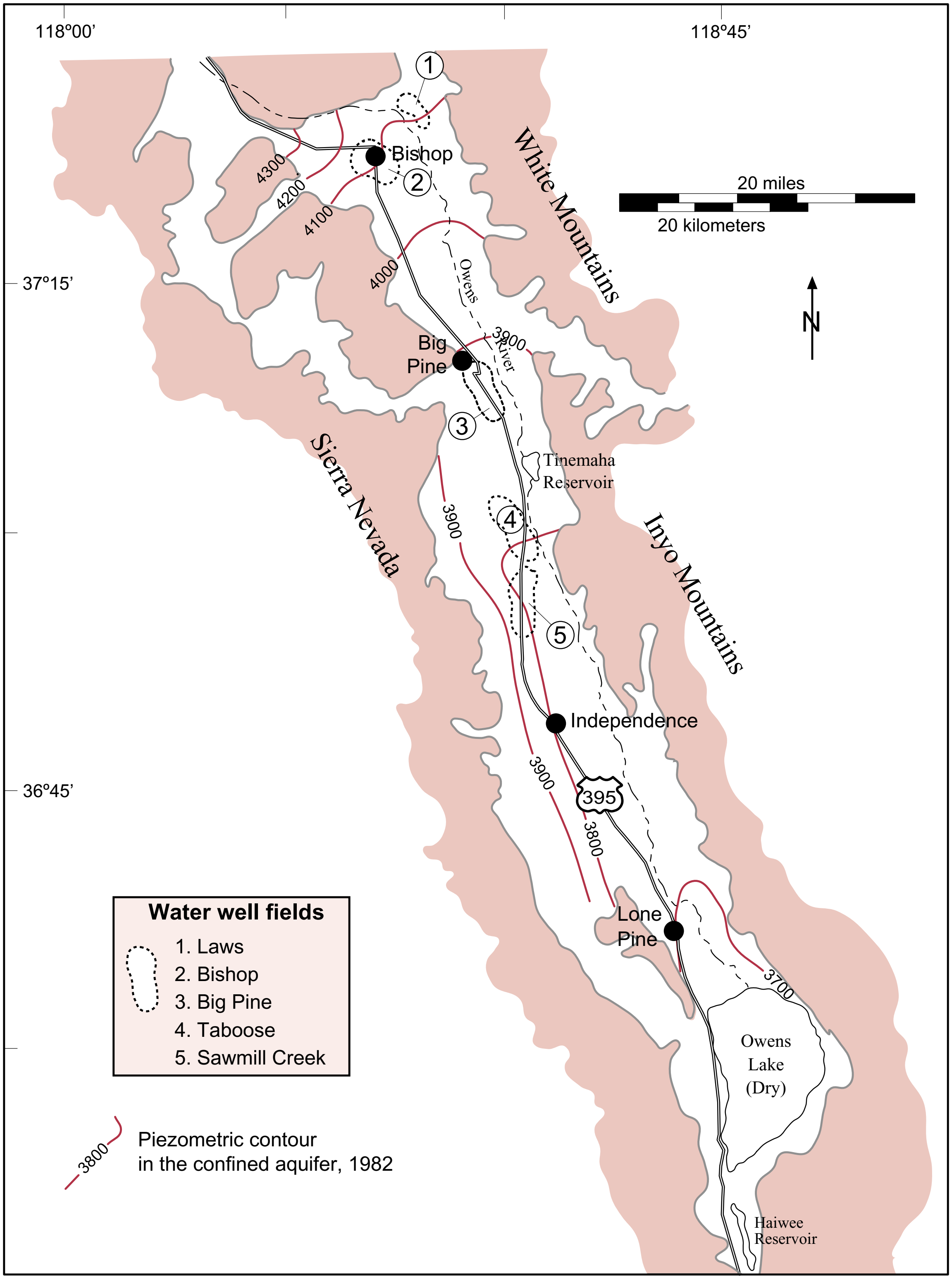

Figure 9. Map of the Owens Valley, with general configuration of the water table throughout the Owens Valley in 1985 (modified from Rogers et al., 1987).

the Owens Valley closed-drainage basin from the China Lake drainage basin to the south. Located in the rain shadow of the Sierra Nevada, the valley has an arid climate with an annual average of less than 6 inches of precipitation. However, because of the streamflow from the Sierra Nevada, the valley plays a crucial role in supplying water to the City of Los Angeles, 230 miles to the south.

The assessment of the water resources of the basin has been a matter of concern ever since the early 1900s, when the City of Los Angeles started acquiring much of the property and water rights along the axis of the valley. Key studies include those of Lee (1906, 1912), DWR (1960b), Griepentrog and Groenveld (1981), LADWP (1972, 1979 among many others), and Rogers et al. (1987). The following discussion is largely based on this last reference.

In terms of its hydrologic budget, inflow to the basin comes largely from partial infiltration of streamflow from the Sierra Nevada, with minor contributions from areal precipitation. Outflow is dominated by pumping (to boost the water supply to the City of Los Angeles) and evapotranspiration (discharge to surface springs was significant prior to development, but pumping has reduced spring discharge to an insignificant level).

The initial export of water from Owens Valley started with completion of the first aqueduct in 1913. Average export volume was about 300,000 acre-feet per year, out of which only a modest 10,000 acre-feet was derived from wells tapping the groundwater resource. A second aqueduct was completed in 1970, and the average annual export increased to about 500,000 acre-feet, with nearly 100,000 acre-feet currently produced by groundwater pumping. In 1987 the City of Los Angeles operated 92 deep wells distributed in nine well fields (but the total number of wells and test holes is in excess of 475). Nearly 50% of the yield was derived from two fields, Big Pine-Crater Mountain and Taboose-Aberdeen, in which production was largely from fractured volcanic rocks that are interbedded with the alluvial fill of the valley (Figure 9). Some of these wells yield as much as 9,000 gpm (in contrast, wells screened in sedimentary units have characteristic yields between 1,000 and 5,000 gpm).

The Owens Valley graben is filled with alluvial-fan gravel and sand deposits that interdigitate with lacustrine clay layers, air-fall tuffs, ignimbrites, and lava flows. The valley fill deposits range in thickness between 1,000 and 8,000 feet along the axis of the valley and, in the case of the alluvial-fan units, thicken considerably toward the bounding mountain ranges. Prominent among the volcanic units is the Bishop Tuff, which includes rhyolitic air-fall and ignimbrite units erupted from the Long Valley silicic volcanic center (in the northern third of the Owens Valley) 700,000 years ago (Bailey et al., 1976). Smaller in volume, but more significant in terms of groundwater production, are the basaltic scoria and lava flows of the Big Pine cinder cone field, south of the town of Bishop. Interbedded fine-grained lake-bed deposits, like the ones exposed in the dry bed of Owens Lake, serve as confining units for the aquifers developed in the coarse-grained alluvial deposits and the fractured volcanic units.

Groundwater moves from the flanks of the surrounding mountains toward the center of the valley and then southward toward Owens Lake (Figure 9). The lake is now dry, but apparently still functions as the ultimate “sink” for groundwater flow due to intense evaporation. Pumping creates local depression cones, but none of these cones seems to have affected the general pre-development flow pattern. As to vertical movement, the presence of interbedded lacustrine clays often leads to confined conditions in the deep aquifers, so most deep wells and springs have some artesian pressure.

OTHER AQUIFERS

My intent in this overview paper was not to make an exhaustive inventory of the groundwater resources of California, partly because of my own limited stamina and knowledge, and partly because much work remains to be done to characterize other aquifers. However, the following key references may be useful as a starting point for practitioners interested in the hydrogeology of Shasta Valley (DWR, 1964), Eureka Valley (Evenson, 1959), the basin-fill aquifers of northernmost California (Wood, 1960; U.S. Department of the Interior, 1980), the volcanic-rock aquifers of the Cascades and the Modoc Plateau (Planert and Williams, 1995), and Death Valley (Hunt and Robinson, 1966).

In addition to the high-yield alluvial and volcanic-rock aquifers referred to above, much remains to be learned about the fractured-rock “aquifers” of the Sierra Nevada and the Klamath Mountains. These “aquifers” are comparatively small in terms of total storage, and are often hosted by fractured igneous and metamorphic rocks. Fortunately surface water is comparatively abundant in these regions, and agricultural and urban demands are low, so groundwater extraction is modest and often limited to rural homesteads. Nevertheless, the exploration and development of fractured-rock groundwater resources remain two of the most challenging professional tasks for California hydrogeologists.

EXPLORATION METHODS

Alluvial aquifers. Finding groundwater in basins such as the Central Valley or the Santa Clara Valley basins is not the problem. Just dig and you will eventually find it. The real challenge faced by the exploration hydrogeologist is to site and design high yield wells. For example, in the northern San Joaquin Valley a typical domestic well would be less than 200 feet deep, would be screened in the lower 40 feet (8-inch diameter), and would have a safe yield of about 100 gpm. In contrast, an agricultural well equipped with an electric pump (and operated rationally during the low-tariff times of 10 pm to 10 am) would need to have a safe yield of about 2,000 gpm to service an orchard area of up to 300 acres. A farmer's dream would be a well that is no more than 500 feet deep, and is screened in the lower 200 feet (16-inch diameter).

How do we go about finding this dream well? My prime exploration strategies are careful surveys of neighbors' wells and thorough interviews with local drilling companies. Borehole geophysical surveys could be very useful for characterization of local aquifer conditions, but most existing wells have steel casing (which eliminates most types of resistivity logs) and active-source radioactive methods would be unadvisable because of the hazard that a loss of the active-source tool could represent to the aquifers. Natural gamma logs are very helpful and allow good discrimination of clay and sand units.

Vertical resistivity soundings remain the prime exploration tool in areas where neighboring wells are scarce, but the method cannot discriminate between low-yield and high-yield horizons. Ultimately, at least one pilot hole has to be drilled and carefully logged to characterize local aquifer conditions.

Fractured-rock aquifers. My prime exploration strategies when dealing with fractured-rock aquifers are lineament studies, surface fracture surveys (for both orientation and spacing of the fractures), surface geophysics, and drilling. The best aquifer management tool, in turn, is borehole geophysics.

The detailed analysis of surface fractures is a task that would try the patience of Job, but it is of crucial importance for pinpointing and characterizing zones of intense fracturing. Data must be collected, in a systematic way, about the orientation of each fracture, the spacing between fractures, and characteristics such as openness, mineral fill, or annealing. The interpretation of structural orientations requires the use of stereographic projections to represent and analyze the three-dimensional data in two dimensions. Fractures and other discontinuities (e.g., dikes or veins) are plotted in the pole format in order to detect the presence of preferred orientations, thus defining discontinuity sets, and to determine mean values for the orientations of these sets. This process can be facilitated by contouring to accentuate and distinguish the repetitive features from the random features. I believe that careful analysis of structural data eventually leads to the recognition of fracture directions that make "good geologic sense", in that they can be reconciled with the tectonic stress regime of the region. It is these regional fracture sets that I normally look for in an exploration program.

Spacing between discontinuities, and patterns of spatial distribution, can be characterized through the use of standard statistical techniques (e.g., Swan and Sandilands, 1995). For example, spacing between fractures can be easily represented, and visualized, by simple frequency histograms. Modal spacings of less than 1 foot are often indicative of a zone of intense fracturing.

Surface geophysics can be of some assistance in locating fracture clusters with high hydraulic conductivity, but it can hardly be considered a sure-fire method. The most promising approach is the so-called VLF method (short for very low frequency electromagnetic surveying). The VLF method relies on the fact that the U.S. and other coastal nations operate long wavelength, or very low frequency, radio stations for communication with submarines. The electromagnetic emissions from the VLF antennas propagate as air, water, and groundwaves, with magnetic and electric field components. Far away from the transmitter, the VLF field can be regarded as a uniform electromagnetic field that is oriented parallel to the surface of the ground and perpendicular to the bearing of the transmitter. In the ground, the primary (source) field propagates vertically away from the transmitter. Upon encountering an electrical conductor (e.g., a fluid-filled fracture), the propagation of the source field causes the flow of secondary electrical currents, which in turn generate a secondary magnetic field that adds its strength to the total magnetic field. In the presence of a lateral change in conductivity, the secondary field is shifted in phase relative to the primary field. To the extent that the observed VLF anomalies are caused by relatively vertical and narrow structures elongated parallel to the bearing to the transmitter, the interpretation of the data is straightforward: the surface trace of the structure is inferred to be where there is a positive anomaly in the total magnetic field and where the in-phase response changes sign (the crossover point). Interpretation is complicated, however, by non-geologic conditions (e.g., power lines or grounded metal fences), topography, or departures from the ideal assumption that the conductor is narrow, steeply dipping, and parallel to the bearing to the transmitter.

Ultimately, the exploration program has to be put to the test by drilling. We have identified a lineament that coincides with a cluster of fractures, the fractures appear to have formed in response to a large scale tectonic stress regime, and surface geophysics suggests that a vertical, narrow conductive zone can be found at depth. Hence, we advise the property owner that it is time to retain a driller, and our client reasonably asks "How deep should we drill?" Personally, I feel inclined to abandon a hole that is more than a few hundred feet in favor of a new location. Chasing fractures can be an extremely costly proposition, as witnessed by many "dusters" drilled to depths of two or three thousand feet. Page et al. (1984) reached a similar conclusion after analyzing more than 200 well records from Nevada County in the Sierra Nevada. They found that most producing wells had depths of less than 180 feet, with yields commonly in the range between 5 and 60 gpm. In contrast, wells that had been advanced more than 215 feet in search of a producing fracture had yields that were often less than 5 gpm. In a separate study, Davis and Turk (1964) compiled the yields of 239 wells in crystalline rocks of the Sierra Nevada, and found that the median flow for wells with a depth of less than 100 feet was 10 gpm or more, but less than that in wells that had been advanced more than 200 feet. I stress the fact that these are empirical observations, to which no doubt many exceptions can be found. After all, fluid-filled fractures can remain open under lithostatic loads of tens of thousands of feet. If only we had the budget to keep looking for them!

Great advances have been made in recent years in the analysis of fractured-rock aquifers by borehole geophysics. For example, a pilot study at the Raymond site, in the foothills of the Sierra Nevada (Cohen et al., 1996), used impeller flowmeter logging, thermal-pulse flowmeter logging, hydrophysical logging, and straddle-packer injection profiling to determine the location of fluid-bearing fractures and their respective transmissivities. Cohen et al. (1996) concluded that the hydro-physical logging measurements were the most precise and enabled confident assessment of the relative magnitudes of the hydraulic conductivity of individual fractures. Hydrophysical logging is based on a variant of the borehole dilution method (Drost et al., 1968; Freeze and Cherry, 1979). The horizontal average linear velocity of groundwater moving through a fracture, or group of fractures, is estimated through the introduction of a "tracer" in an isolated interval of the well and periodic measurements of the concentration as the tracer is diluted by groundwater flowing through that interval. The rate of concentration decay is related to the average velocity of groundwater moving through the formation and across the borehole. In hydrophysical logging, deionized water is used as the tracer and fluid electric conductivity is measured repeatedly with a fluid conductivity probe to keep track of changes in the "concentration" of the tracer. Interpretation of the data is accomplished using the methods presented in Anderson et al. (1993), Tsang et al. (1990), Pedler et al. (1988), and Drost et al. (1968).

EPILOGUE

Preparing this summary started as a short project and ended being a never-ending story. Nevertheless, I enjoyed myself and greatly increased my admiration for the tremendous work performed by the "old guard" of California geologists. I hope our generation is remembered in the same light, and that our combined works will be a source of inspiration and encouragement to young engineering geologists and hydrogeologists. I can see that they will face unique challenges, such as the management of an enormous volume of accumulated data and the extensive use of a vital but finite resource. I feel we need to help them in what will be a difficult task by maintaining the highest standards in our colleges (yes, Physics and Field Geology are still essential in the Geology curriculum); by providing internship and professional training opportunities; by becoming involved, as a profession, in the management of this resource; and by applying our ebullient scientific imagination to the solution of meaningful hydrogeologic problems. Here's to a wonderful profession!

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APPENDIX

IMPORTANT STRATIGRAPHIC UNITS OF THE CENTRAL VALLEY

Sacramento Valley

Tuscan Formation. This formation crops out from Red Bluff to Oroville (Harwood et al., 1981), and can be recognized in the subsurface to a distance of about 5 miles west of the Sacramento River (Olmstead and Davis, 1961; DWR, 1978). The Tuscan Formation was accumulated as alluvial fan deposits off the Cascade Range, and thus thins from east to west from 1,600 feet at the Cascade Range to about 300 in the subsurface under the Sacramento Valley (Lydon, 1969; DWR, 1978). In the subsurface it consists largely of black volcanic sands and gravels, with interbedded layers of tuffaceous clay and tuff breccia. The Tuscan Formation yields large quantities of fresh water to wells (900 to 3,000 gallons per minute; DWR, 1978), often from aquifers confined by tuffaceous beds.

Tehama Formation. This formation crops out in the western flank of the Sacramento Valley basin, From Red Bluff to Vacaville. The Tehama Formation extends easterly from the west marging of the basin toward the trough of the valley, where it interfingers with the Tuscan and Laguna Formations. In areas where it is exposed it has an average thickness of 1,800 feet and consists of poorly-sorted, thick-bedded sandy silt and clay that yield poorly to wells. Gravel and sand interbeds are usually thin (< 50 feet) and discontinuous, but may host high-yield localized aquifers (200 to 2,500 gallons per minute) (Olmstead and Davis, 1961; DWR, 1978). Because of its extension and thickness, the Tehama Formation is the principal water-bearing formation in the western half of the Sacramento Valley (but aquifer development challenges the hydrogeologist because of facies changes and large proportions of low-yield silt and clay interbeds).

Laguna and Fair Oaks Formations. The Laguna Formation is well exposed in the southeastern part of the Sacramento Valley, where it forms many of the low, rolling foothills southeast of Sacramento and south of the American River (formally, the equivalent sediments north of the American River are grouped under the Fair Oaks Formation, but the sedimentology and hydrogeology of the two formations are very similar). DWR (1978) has described the Laguna and Fair Oaks Formations as a westward-thickening wedge that was deposited by streams draining the Sierra Nevada. In outcrop they are up to 180 feet thick and rest conformably over the Mehrten Formation. To the west they dip toward the valley trough, where they interdigitates with the Tehama and Tuscan Formations.

The Laguna Formation includes beds of silt, clay, and sand with lenticular bodies of gravel. Some of the sands are clean and well sorted, whereas some of the gravels are silty and poorly sorted. Where fine-grained, the Laguna Formation yields poorly to wells, but in areas where well-sorted arkosic sands predominate the yields can be quite high (~1,750 gallons per minute).

Victor Formation. The Victor Formation overlies the Laguna and Fair Oaks Formations, and forms most of the valley floor east of the Sacramento River. According to DWR (1978) it was deposited on a plain of aggradation, now partly dissected, so it is composed of a heterogeneous assemblage of fluvial sediments deposited by streams that drained the Sierra Nevada. These streams left sand and gravel that grade laterally and vertically into silt and clay, which results on laterally-discontinuous, thin aquifers that are difficult to correlate. The Victor Formation is the most important water-bearing formation for domestic and shallow irrigation wells on the eastern half of the Sacramento Valley basin, even though yield to wells is generally modest (< 1,000 gallons per minute).

Northern San Joaquin Valley

Mehrten Formation. This formation crops out discontinuously along the eastern flank of the valley, between the Bear and Chowchilla rivers, and dips gently to the southwest beneath the valley. The formation was originally defined by Piper et al. (1939) to refer to a 190-foot section of clay, silt, and lithic sandstone and breccia (with characteristic black andesite detrital grains) in the Mokelumne area, laid down by streams carrying andesitic debris from the Sierra Nevada (see also Marchand and

Allwardt, 1981). The detailed makeup of the stratigraphic section changes markedly from one location to another, but the comparative abundance of andesitic detrital grains remains a diagnostic characteristic. In the Sacramento Valley the formation is as much as 200 feet thick where exposed, and in the subsurface it ranges in thickness from 400 to 500 feet. The unit apparently thickens to the south: in the northeastern part of the San Joaquin Valley Davis and Hall (1959) report an aggregate outcrop thickness of more than 700 feet, and a subsurface thickness of nearly 1,200 feet. The "black sands" of the Mehrten Formation generally yield large quantities of water to wells (3,000 to 4,000 gallons per minute being common), which makes them a preferred exploration target in the eastern half of the Central Valley (Davis and Hall, 1959).

Turlock Lake, Riverbank, and Modesto Formations. These formations are easily differentiated from the underlying Mehrten Formation because the silts, sands, and gravels that form them contain large proportions of quartz and feldspar (i.e., they are arkosic or quartz-feldspathic sediments), which give them a light color. In contrast, the underlying Mehrten is characterized by dark-colored andesitic clasts. The change in sediment type, from andesitic to granitic, was brought forth by the enormous supply of sediments "released" by the Pleistocene glaciation of the High Sierra. Technically, then, most of the sediments found in the Turlock Lake, Riverbank, and Modesto Formations are glacial outwash sediments.

Mapping of the poorly exposed sediments has traditionally been based on degree of soil development. Because the Turlock Lake sediments have been exposed for a longer period of time than the Riverbank or Modesto sediments, their soil profiles have developed thicker, more compact B horizons (local farmers refer to these compact soil horizons as "hardpan"). According to Davis and Hall (1959), the Turlock Lake Formation forms dissected, rolling hills with up to 60 feet of local relief. The Riverbank Formation, in contrast, forms low, slightly dissected hills with 10 to 20 feet of local relief to nearly flat land. Finally, the Modesto Formation has a nearly flat topographic relief with a gentle westward slope. The distinction between the three units is not practical in the subsurface, where the total combined thickness of these Pleistocene units can be as much as 1,000 feet. Typical yields for agricultural wells vary between 2,000 and 3,500 gallons per minute.

Southern San Joaquin Valley

Kern River Formation. Miocene marine and nearshore deposits at the southern end of the Central Valley (e.g., the Jewett and Olcese Formations) are overlain by Plio-Pleistocene sediments transported by streams draining westward from the Sierra Nevada. These sediments are grouped as the Kern River Formation, which is dominated by stacked channel-fill sands that includes lenses and layers of cobbly, locally bouldery gravels with sand/clay matrix, interbedded with medium- to very coarse-grained sands, clayey sands, and sandy clays. Some of the sandy clays contain minor layers of very fine pebbles (Hackel et al. 1965); Nicholson, 1980; Bartow and Pittman, 1983; Miller, 1986; Olson et al., 1986; Graham et al., 1988; Kuespert, 1990; Kuespert and Sanford, 1990; and Bartow, 1991). Borehole data and surface exposures indicate that the Kern River Formation ranges in thickness from 50 to 750 feet. According to Miller and Graham (1995), the depositional setting of the Kern River Formation evolved from a coarse-grained marine delta that fed deep-water, coarse clastic deposits during the Late Miocene, to a system of glaciogenic fluvial deposits and lacustrine deltas during the Plio-Pleistocene. The presence of Hemphillian (late Miocene-early Pliocene) vertebrate fossil assemblages found near the base of the formation (Savage et al., 1954; Bartow, 1991) constrain the lower age for the Kern River Formation at about 8.2 Ma. The upper age constraint of the Kern River Formation is based on a volcanic ash with a radiometric age of 6.13 Ma (Miller et al., 1998).

The Kern River Formation dips gently (<10º) toward the San Joaquin Valley, and overlies the easternmost end of the Bakersfield Arch (a broad, southwest plunging anticline). These uplifted and tilted strata are incised by the present day Kern River and are exposed in outcrops along either side of the river valley. Several faults cut through these sediments. Most are normal faults that strike either north-northwest or west-northwest (Nicholson, 1980; Bartow, 1991).

The Kern River reservoir sands have produced more than 1 billion barrels of oil from a 4 billion barrel resource, which, of course, makes them useless as a fresh-water aquifer at or near the oil fields. Away from the oil fields, however, the formation provides valuable groundwater storage and supports high yields to wells.

Tulare Formation. The Pliocene-Pleistocene-Recent(?) Tulare Formation is a widespread nonmarine unit that crops out along the western margin of the San Joaquin basin (Watts, 1894; Anderson, 1905). It consists of diverse sedimentary facies ranging from alluvial fan (Lennon, 1976; Lettis, 1988) to fluviodeltaic (Miller et al., 1990) and lacustrine units (Woodring et al., 1940; Lennon, 1976). Sediments range from conglomerates to silts and clays.

The Tulare Formation contains at least three major units in the southern San Joaquin Valley basin, which are separated by clay units deposited during widespread lacustrine flooding of the basin. The base of the Tulare is the *Amnicola* sand (named for the common occurrence of the freshwater gastropod) that ranges in age from 3.4 Ma to 2.2 Ma. Overlying the *Amnicola* sand are the Paloma Clay (Pacific Geotechnical, 1991), and an unnamed sand of varying thickness and extent. The latter is in turn overlain by the Corcoran Clay (Frink and Kues, 1954; Croft, 1972; Lettis, 1982, 1988), which is also referred to as the "E-clay" in the hydrogeologic literature (Page, 1986). Finally, extending from the Corcoran Clay to the present surface are Upper Pleistocene and Holocene alluvial sands.

The maximum reported thickness of the Tulare Formation along the western basin margin is 3,500 ft (Loomis, 1990; Woodring et al., 1940), with a similar thickness of equivalent section (3,800 ft) occurring in the southernmost part of the basin (Schwartz, 1990). Beneath the valley the thickness ranges from about 200 feet at North Belridge to about 5,000 feet beneath the Kettleman Plains (Wood and Davis, 1959). The Corcoran member of the Tulare Formation is recognized along the western San Joaquin Valley basin as far north as the San Luis Reservoir.

Three Pliocene-Pleistocene fluvial depositional systems provided sediment to the San Joaquin Valley basin prior to the deposition of the Corcoran Clay. The largest volume of sediment was contributed from the basin's southern margin as a combination of fluvial deposits from the Mojave province and the Sierra Nevada (e.g., Kern River). The second system represents sediment transport from the Sacramento Valley, as documented by paleocurrent measurements in the conglomerates of northernmost Kettleman Hills. Transport direction reversed itself after deposition of the lacustrine Corcoran Clay. The third and smallest depositional system originated in the Salinas basin, west of the present San Andreas fault, where paleocurrent directions in braided gravel deposits of the Pliocene-Pleistocene Paso Robles Formation indicate eastward transport of sediment and connection with the San Joaquin Valley basin across the present trace of the San Andreas fault (Galehouse, 1967). Deposition after the Corcoran Clay formed the alluvial fan and lacustrine systems that persist to the present day.

Along the western side of the valley, south of Tulare Lake, the Tulare Formation contains mostly saline water, whereas north of Tulare Lake it contains mostly fresh water (Hotchkiss and Balding, 1971; Miller et al., 1971). Because of the large proportion of clay in the deposits, yield to wells is generally low (< 500 gallons per minute), but Hotchkiss and Balding (1971) have reported average yields of more than 1,000 gallons per minute for wells in the Tracy-Dos Palos area.