Natural Recharge of Groundwater Symposium

Symposium Proceedings

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Tempe, Arizona
June 2, 2000

Sponsored by
Arizona Hydrological Society, Arizona Department of Water Resources,
Salt River Project, U.S. Water Conservation Laboratory of USDA-ARS
and U.S. Geological Survey
Natural Recharge of Groundwater Symposium

Symposium Proceedings
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FOREWORD

From the Organizing Committee:

Presenters at the Symposium on Natural Recharge held at the Embassy Suites Hotel in Tempe Arizona on June 2, 2000, provided the following papers. The organizing committee wishes to thank the authors for their contribution to the success of this symposium. Our intent was to provide the reader and/or symposium participant with the latest research on this important topic. Natural recharge is a hydrologic process that has been difficult to measure in the field and is therefore one of the least understood components of basin-wide water balance. In the arid Southwest U.S., evaluation of water supply requires an understanding of natural recharge as well as other hydrologic influences. We hope that these papers add to this understanding and contribute in some small way to better groundwater management.

The organizing committee appreciates the support of our sponsors and the various individuals who have donated time to putting the symposium together. We would especially like to thank Suzanne Kirk of URS – Dames & Moore for orchestrating the details of the symposium and for her organizational skills. Without her, the symposium could not have happened.

Doug Bartlett
Organizing Committee Chairman
SPONSORS

The organizers of the Natural Recharge of Groundwater Symposium wish to express their gratitude to the following sponsors for their financial support of this symposium.

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FRIDAY - June 2, 2000

INTRODUCTION
8:15 a.m. - 8:35 a.m.
Frank Corkhill

SESSION 1
8:35 a.m. - 9:55 a.m.
Herman Bouwer
Bridget R. Scanlon
The Recharge of Groundwater
Relationship Between Geomorphic Settings and Unsaturated Flow in an Arid Setting

9:55 a.m. - 10:10 a.m.
Break

SESSION 2
10:10 a.m. - 12:10 a.m.
Jan M.H. Hendrickx
Dennis Williams
John Izbicki
Localized Natural Recharge in Arid Basins
Natural Recharge in the Cadiz Area, San Bernardino County, California
Chloride and Tritium Concentrations in a Thick Unsaturated Zone Underlying an Intermittent Stream in the Mojave Desert, Southern California, USA

12:10 a.m. - 1:30 p.m.
Lunch

SESSION 3
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Allen Flint
Frank Putman
Characterizing Natural Recharge using Geoscientific Information Systems (GSIS)
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SESSION 4
3:05 p.m. - 4:25 p.m.
Stan Leake
David C. Goodrich
M. Susan Moran
Southwest Groundwater Resources Project
San Pedro River Studies and Remote Sensing of Evapotranspiration

4:25 p.m. - 4:30 p.m.
Closing Remarks

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CHLORIDE AND TRITIUM CONCENTRATIONS IN A THICK UNSATURATED ZONE UNDERLYING AN INTERMITTENT STREAM IN THE MOJAVE DESERT, SOUTHERN CALIFORNIA, USA

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ABSTRACT

Previous studies indicated that small amounts of recharge may occur as infiltration of streamflow in intermittent washes, in the western part of the Mojave Desert, near Victorville, California. These washes typically flow only for a few days each year after large storms. To reach the water table, infiltrating water must pass through an unsaturated zone that is more than 130 meters (m) thick. Chloride concentrations in core material and cuttings from sites away from the wash were as high as 200 micrograms per gram within 12 m below land surface—the approximate depth of the root zone. On the basis of a chloride deposition rate of ~10 micrograms per square centimeter per year, measured as part of this study, it has been more than 10,000 years since water infiltrated to depths greater than 12 m at these sites. In contrast, chloride concentrations in unsaturated material underlying the wash were less than 10 micrograms per gram. Tritium was present in water extracted from core material under the wash to a depth of ~30 m. This corresponds to an average rate of movement of ~0.7 m per year, which suggests that at least 200 years is required for water to move through the thick unsaturated zone underlying Oro Grande Wash at the study site.

INTRODUCTION

The study area is in the western part of the Mojave Desert near Victorville, California, ~130 kilometers (km) east of Los Angeles (fig. 1). In recent years, population in the area has increased threefold, from ~90,000 in 1980 to more than 270,000 in 1993 (Victorville Chamber of Commerce, pers. comm., 1994). Water supply in the area is derived almost entirely from ground water, and pumping has increased with the population. In the past, most ground water was pumped from alluvial deposits along the Mojave River. These deposits are readily recharged by infiltration of stormflows in the Mojave River. Most recharge to the regional aquifer occurs near the front of the mountains to the south of the basin (Hardt, 1971), and the quantity of recharge (under present-day climatic conditions) is small relative to the quantity of water in storage and the quantity of water pumped from the aquifer. As a consequence, water levels in some wells completed in the regional aquifer declined more than 0.3 meters per year (m/yr) between 1985 and 1990 (Mendez and Christiansen, 1997).

On the basis of carbon-14 data, Izbicki et al. (1995) showed that most ground water in the regional aquifer was recharged between 5,000 and 20,000 years ago. However, they also identified areas where ground water in the regional aquifer was recharged between 500 and 2,400 years ago. The predevelopment water-level maps prepared by Hardt (1971) indicate that this water could not have been recharged by infiltration of stormflows in the Mojave River and, as a consequence, was interpreted to be part of a subregional groundwater-flow system within the regional aquifer. Because direct infiltration of precipitation is unlikely in desert areas, Izbicki et al. (1995) suggested that this flow system may be recharged as infiltration flow in intermittent streams. Water from these intermittent streams must infiltrate below the root zone (~10 to 12 m) and through a thick unsaturated zone (in some places more than 300 m) to reach the water table. The quantity of recharge from this source is unknown.

The purpose of this study was to determine if water from an intermittent stream can infiltrate below the root zone and ultimately move through a thick unsaturated zone to reach the water table. This paper discusses the results of work done at one site along Oro Grande Wash in the western part of the Mojave Desert. Results presented in this paper are preliminary and part of a larger study of infiltration from intermittent streams at a number of sites in the western part of the Mojave Desert. The study was funded cooperatively by the Mojave Water Agency and the U.S. Geological Survey.

HYDROLOGY

The study area overlies alluvial fan and basin-fill deposits that in places are more than 1,000 m thick. These deposits consist of unconsolidated to moderately consolidated interbedded gravel, sand, silt, and clay (California Department of Water Resources, 1967) and
Figure 1. Location of study area.
constitute the aquifer known locally as the regional aquifer. Under predevelopment conditions the regional aquifer drained to the alluvial deposits along the Mojave River. Low precipitation, low humidity, and high summer temperatures characterize the climate of the area. Precipitation in the area is generally less than 150 millimeters per year (mm/yr). Precipitation is greater in the mountains to the south of the study area and near Cajon Pass, a gap between the San Gabriel and San Bernardino Mountains. Moist air from the Pacific Ocean enters the Mojave Desert through Cajon Pass, and precipitation near the pass can exceed 1000 mm/yr (this study). Depth to water in the regional aquifer ranges from ~30 m on the bluffs overlooking the Mojave River to more than 300 m near Cajon Pass. Underlying the study site, the unsaturated zone is ~130 m thick.

In previous studies, most of the recharge to the regional aquifer was believed to occur near the fronts of the San Bernardino and San Gabriel Mountains, and ground water was believed to flow toward the Mojave River (Hardt, 1971). Recent work by Izbicki et al. (1995) suggested that some recharge may occur as infiltration of water from intermittent streams that cross the deposits composing the regional aquifer. Recharge from these streams may support a subregional groundwater flow system in which flow is also toward the Mojave River. It was unclear from previous work if the quantity of water infiltrated from intermittent streams is large enough to infiltrate below the root zone, through the thick unsaturated zone, and ultimately recharge the underlying aquifer.

Oro Grande Wash is an intermittent stream that flows only several days each year. On the basis of a relation between channel geometry and mean annual streamflow developed by Lines (1996), mean annual streamflow in Oro Grande Wash near the study site was estimated to be ~0.5 cubic hectometer (hm³). Most flow occurs during the winter months; however, observations in the field suggest occasional flow after summer thunderstorms. The channel of the wash is ~3 m wide and can be traced to near Cajon Pass, ~15 km upstream from the study site. The channel of Oro Grande Wash is incised 10 to 20 m into the surface of the regional alluvial fan deposits. Incision of the wash occurred after the opening of Cajon Pass ~500,000 years ago, and streamflow along the wash has followed nearly the same course since then. Because the head of the fan has been eroded by Cajon Creek (which drains to the Pacific Ocean), Oro Grande Wash does not drain the mountains to the south of the study area, and streamflow in the wash is a combination of runoff from precipitation that falls near the pass and streamflow from locally derived runoff farther downstream from the pass. In recent years, the amount of locally derived runoff may have increased as a result of urbanization in the Victorville area.

DATA COLLECTION

As part of this study three boreholes, LOGW-1, LOGW-2, and OGF (located in the wash, and ~45 m and 200 m west of the wash, respectively) were drilled to a depth of ~30 m. A fourth borehole, LOGW-3, was drilled 15 m west of the wash to a depth of ~15 m. LOGW-1 and OGF were drilled in 1994 and LOGW-2 and LOGW-3 were drilled in 1996. All boreholes were drilled using the ODEX air-hammer drilling method (Driscoll, 1986). To help ensure comparability of the data, all holes were drilled at the end of summer prior to the beginning of the rainy season. At all boreholes, drill cuttings were collected at 0.3-meter intervals and are believed to represent an average of the material in that interval. The drill cuttings were described visually in the field for grain size, sorting, color, mineralogy, and water content. A subsample was then sieved and leached with distilled water and the leachate was analyzed for specific conductance. Cores (10 cm in diameter and 60 cm long) were collected using a piston core barrel every 1.3 m at LOGW-1, LOGW-2, and OGF. Cores were not collected at LOGW-3. The core barrel was lined with four brass or aluminum core liners each ~0.175-m long. Each liner was immediately extracted from the core barrel, capped, labeled, wrapped in plastic, and sealed in individual aluminum pouches following the procedure of Hammermeister et al. (1986) to protect the cores from evaporation and contamination.

Analysis of Cores and Drill Cuttings

Subsamples from selected cores were analyzed for water content, water potential, chloride, and tritium (a radioactive isotope of hydrogen). Water content and water potential were measured on the same subsamples. Tritium was measured on water extracted from an adjacent subsample. Chloride was measured in core material and on drill cuttings.

Water content is the amount of water in a given amount (by volume or weight) of material. In this paper, water content is reported as volumetric water content (volume of water per volume of material). Water potential is a measure of how tightly the water in the material is held. Water potential is expressed as negative pressure; the more negative the water potential, the more tightly the water is held to the material. Water potential is a function of both the moisture content and the lithology (texture) of the material. In this paper, water potentials are reported as kilopascals (kPa) (1 kPa = 0.01 bar). The volumetric water content of core material was measured gravimetrically. The water potential of core material was measured using different methods depending on the expected water potential of the material. Moist samples were measured using either tensiometers or filter paper (Campbell and Gee, 1986); dryer samples were measured using a water-activity meter, commonly known as a chilled mirror hygrometer (Gee et al., 1992). Analyses for moisture content and water potential were done at the Desert Research Institute in Reno, Nevada.

Chloride concentrations were determined from analyses of leachates extracted from drill cuttings and core material collected from the four boreholes. Prior to extraction, the core material was sieved to obtain 50 grams (g) of material having a particle size less than 2 mm. The sieved sample was mixed with 50 milliliters (ml) of distilled water, shaken vigorously, allowed to stand for ~24 hours, and centrifuged to allow the remaining solids to settle. The supernate liquid extract was pressure-filtered through a 0.45-μm filter and analyzed for chloride using ion chromatography at the U.S. Geological Survey laboratory in San Diego, Calif. The chloride concentrations of the drill cuttings represent an average over the 0.3-meter collection interval, and the chloride concentration of the core material represents a point measurement at the depth at which the sample was collected. There was little difference in the chloride concentration of drill cuttings and core material collected at similar depths, and only the chloride concentrations of the drill cuttings are reported in this paper.
Tritium was measured in water extracted from core material by vacuum distillation at 80°C and analyzed by liquid scintillation with electrolytic enhancement (when sample volume was large enough). The detection limit is a function of the volume of water extracted from the core material. In general, more water was extracted from moist cores, resulting in lower detection limits. Analyses of tritium were done at the U.S. Geological Survey laboratory in Menlo Park, California.

Collection and Analyses of Bulk Precipitation

Bulk precipitation (wet fall plus dry fall) was collected at five sites between December 1994 and November 1997 (fig. 1). These data were used to estimate chloride deposition rates. Collectors were based on a design by Friedman et al. (1992) and consisted of a 75-mm diameter straight-sided Buchner funnel supported on a stake ~1 m above the ground. The funnel was connected to a 1-liter plastic bottle placed below the ground. The bottles contained a thin layer of mineral oil to prevent evaporation of the water. (Analysis of distilled water and mineral oil mixtures showed that the mineral oil did not contribute chloride to the water.) Bulk precipitation was collected semiannually, in April and November (at the end of the rainy season and at the end of summer, respectively), to ensure comparability with regionalized precipitation data collected during previous studies (Friedman et al., 1992). After collection, the volume of the sample was measured to determine precipitation quantity, and the water was analyzed for chloride by ion chromatography at the U.S. Geological Survey laboratory in San Diego, Calif.

WATER CONTENT AND WATER POTENTIAL IN CORE MATERIAL

The unsaturated zones underlying most of the Mojave Desert and other arid areas are very dry. Water-content and water-potential data from core material collected from LOGW-2 and OGF are typical of data from unsaturated zones underlying the Mojave Desert. Volumetric water contents at these sites ranged between 1 and 8 percent in the upper 10 to 12 m (fig. 2). Although water content increased with depth, the maximum volumetric water content did not exceed 13 percent, and water contents as low as 2 percent were present in some coarse-grained materials at depths below the root zone. Changes in water content may be related to changes in lithology: finer-grained deposits tend to have higher water contents than do coarser-grained deposits. In both LOGW-2 and OGF, the more negative water potentials were in the upper 10 to 12 m, with the most negative water potential, ~14,000 kPa, at OGF near the base of the root zone, ~10 m below land surface. Below the root zone, water potentials were less negative; however, in both LOGW-2 and OGF, water potentials were more negative than ~1,000 kPa.

In contrast to the very dry deposits sampled at LOGW-2 and OGF, the unsaturated zone at LOGW-1 (underlying Oro Grande Wash) is characterized by volumetric water contents as high as 27 percent and by water potentials between ~1.8 and ~50 kPa (fig. 2). Water contents appear to be lower and water potentials more negative within the upper 3 m. This may have resulted from evapotranspiration of shallow soil moisture during the summer months prior to sample collection or from the drainage of water through the comparatively coarse-grained recent alluvial fill underlying the channel of Oro Grande Wash. Water potentials were near zero (~1.8 kPa) between 6 and 12 m below land surface and became more negative below 12 m. However, water potentials as negative as those encountered away from the stream channel at LOGW-2 and OGF were not encountered in LOGW-1.

Water-content and water-potential data at LOGW-1, underlying Oro Grande Wash, differ from those at LOGW-2 and OGF. These differences indicate that water from intermittent flows in Oro Grande Wash infiltrates into the unsaturated zone and penetrates to depths below the root zone. Although water-content and water-potential data show that the unsaturated zone underlying Oro Grande Wash is wetter than the unsaturated zone underlying the surrounding desert, and that surface water from Oro Grande Wash infiltrates below the root zone (and presumably to the water table at greater depth), these data provide no information on the rate of movement of water at these sites.

CHLORIDE CONCENTRATIONS IN DRILL CUTTINGS

Because of its high solubility, chloride is readily dissolved and moves with infiltrating water. Consequently, chloride concentrations in sediments from unsaturated zones can provide an estimate of the long-term downward water flux in arid regions (Allison et al., 1985; Johnston, 1987; Scanlon, 1991; Phillips, 1994; Tyler et al., 1995; Prudic, 1994). If the chloride deposition rate is known, the length of time since water has passed through the unsaturated deposits can be calculated by dividing the total mass of chloride in the unsaturated zone by the chloride deposition rate. This calculation assumes that bulk precipitation (wet fall plus dry fall) is the only source of chloride to the sediments and that the deposition rate has remained constant through time. If large amounts of chloride are present, it may have been many thousands of years since water has moved through the unsaturated zone. As noted above, chloride data collected from the unsaturated zone at other sites in the Mojave Desert suggest that it has been several thousand years since water has infiltrated through the thick unsaturated zones underlying much of the area. For example, Prudic (1994) estimated that the age of water below the root zone (at ~10 m) at two sites in the Mojave Desert near Beatty, Nev., and Ward Valley, Calif., was between 16,000 and 30,000 years, respectively. This corresponds to a time when the climate was wetter and cooler. Conversely, if only a small amount of chloride is present, water may be moving through the unsaturated zone toward the water table under present-day conditions.

The average atmospheric chloride deposition rate measured at five sites between 1994 and 1997 (fig. 1) as part of this study is ~10 micrograms per centimeter squared per year (μg/cm²yr). Although precipitation varied widely at these sites—ranging from 8.7 mm/year to more than 1,000 mm/year near Cajon Pass, the chloride deposition rate was similar at all the sites measured. Friedman and Smith (1992) calculated that as many as 5 to 7 years of data collection are required to accurately estimate precipitation and chloride
Figure 2. Water content, water potential, chloride, and tritium profiles in the unsaturated zone underlying and near Oro Grande Wash. Boreholes LOGW-1, LOGW-3, LOGW-2, and OGF are located in the wash, and about 15, 45, and 200 meters west of the wash, respectively (see figure 1). Vertical position reflects relative elevation at borehole location.
deposition rates in the Mojave Desert. As a result, the chloride deposition rate presented in this paper should be considered preliminary, and the actual estimate of chloride deposition may change as more data are collected. However, the deposition rate measured as part of this study is similar to the chloride deposition rates used in similar calculations at other sites in the Mojave Desert (Fouty, 1989; Prudic, 1994; Phillips, 1994).

Chloride concentrations in drill cuttings ranged from less than 1 microgram per gram (μg/g) to more than 200 μg/g of soil (Fig. 2). At the three sites west of Oro Grande Wash (LOGW-3, LOGW-2, and OGF), chloride concentrations were highest in the upper 10 to 15 m and decreased with increasing depth. These chloride profiles are typical of desert soils where little or no water infiltrates below the root zone (Johnston, 1987; Allison et al., 1994) and have been measured at sites throughout the Mojave Desert in several studies (Prudic, 1994; Phillips, 1994; Fouty, 1989). In these settings, chloride from precipitation and dry deposition can accumulate over many thousands of years. Given the mass of chloride in the unsaturated zone and the deposition rate of 10 (μg/cm²)/yr measured as part of this study, chloride has been accumulating at these sites for more than 10,000 years.

In contrast, the distribution of chloride with depth at LOGW-1, located in the active channel of Oro Grande Wash, is greatly different. Chloride concentrations at this site were low in the upper 13 m below land surface, increased below 13 m, reached a maximum near 25 m, and subsequently decreased to low values. A chloride profile having this shape is not typical of desert soils, and similar profiles have not been previously reported. The high chloride concentrations at depth may be a regional feature, possibly an ancestral soil horizon. However, high chloride concentrations were not encountered at depth at LOGW-2. This suggests that they are not a regional feature and that the high chloride concentrations at depth at LOGW-1 are related to the infiltration of water from Oro Grande Wash. In addition to the differences in depth, the shape of the chloride profile at LOGW-1 is different from the chloride profile at LOGW-2, LOGW-3, and OGF. Chloride concentrations at LOGW-1 increase above background concentrations at ~13 m below land surface and increase to high concentrations near 25 m below land surface (Fig. 2). It is possible that the high chloride concentrations at LOGW-1 were initially located at a shallower depth and have moved with infiltrating water to the deeper depths. Increased streamflow in Oro Grande Wash as a result of urbanization could provide a mechanism for increased infiltration and subsequent mobilization of chloride.

TRITIUM CONCENTRATIONS IN WATER EXTRACTED FROM CORE MATERIAL

Tritium is a naturally occurring radioactive isotope of hydrogen, with a half-life of ~12.4 years. Tritium data are presented in this paper in tritium units (TU); each tritium unit equals 1 tritium atom in 10¹⁸ atoms of hydrogen. Prior to 1952, the tritium concentration of precipitation in coastal California was ~2 TU. Beginning in 1952, tritium was released to the atmosphere as a result of the atmospheric testing of nuclear weapons, and the tritium concentration of precipitation increased. Tritium in precipitation reached a maximum in about 1962 and decreased after the atmospheric testing of nuclear weapons ended. Because tritium is part of the water molecule, tritium concentrations are not affected significantly by reactions other than radioactive decay. As a result, tritium is an excellent tracer of the movement of water on timescales ranging from 10 to less than 100 years before present. Tritium has been used to determine the rate of movement of water through thick unsaturated zones underlying the Mojave Desert (Phillips, 1994; Striegl et al., 1996) and in other arid regions (Allison and Hughes, 1978; Allison et al., 1994).

Tritium concentrations in water extracted from core material ranged from 6.8 TU to less than the detection limit (Fig. 2). In general, more water was extracted from the comparatively moist cores obtained from LOGW-1 and detection limits were lower, ranging from 0.6 to 1.2 TU; less water was extracted from the very dry core material obtained from LOGW-2 and OGF and detection limits were slightly higher, ranging from 1.0 to 2.4 TU. For the purposes of this study, water having tritium concentrations greater than the detection limit is interpreted as water that infiltrated after 1952, and water having the highest tritium concentrations may be interpreted as water infiltrated near 1962.

Tritium was detected only near the surface at OGF. Although tritium was not detected near the surface at LOGW-2 and tritium was not analyzed for at LOGW-3 (cores suitable for analysis were not collected at this site), tritium is presumed to be present near the surface at these locations. These distributions with depth are typical for tritium in desert soils in areas where rainfall is scant, evapotranspiration is high, and precipitation does not infiltrate to great depths.

In contrast, at LOGW-1, directly underlying Oro Grande Wash, tritium was present to depths as great as 29 m. This represents water recharged after 1952, and corresponds to an average infiltration rate of about 0.7 m/yr. The highest tritium concentration, 6.8 TU, at about 16 m below land surface may represent water recharged during the early 1960s; about the time of the peak tritium concentrations in precipitation. This corresponds to an average infiltration rate of about 0.4 m/yr. On the basis of these data, it would take between 200 and 300 years for infiltrating water to reach the water table about 130 m below land surface. These infiltration rates are consistent with water-flux data estimated for this site by Nimmo (1995). However, layering and other discontinuities in the lithology may result in lateral spreading of the recharge water. As a result of this lateral spreading, it may take significantly longer than 200 to 300 years for the infiltration water to reach the water table.

Tritium was not detected in water extracted from core material collected between 22 and 25 m below land surface within the deep chloride pulse at LOGW-1—even through tritium was present at greater depths (Fig. 2). At least two explanations are possible for this distribution. First, water may have moved through the chloride pulse largely through cracks and macropores, and tritium extracted from cores collected within the chloride pulse was diluted to concentrations less than the detection limit by less-mobile, older (pre-1952) water. Second, water containing tritium may have infiltrated along the Oro Grande Wash upgradient of LOGW-1 and moved laterally around the high-chloride zone. The presence of tritium in water extracted from core material collected 31 m below land surface from LOGW-2 (about 45 m from the wash) is evidence for the lateral movement of water through the unsaturated zone and the spreading of infiltrating
water with depth. Because of the high chloride concentrations near the surface and because tritium was not detected in LOGW-2 at shallower depths, water containing tritium at LOGW-2 is believed to have infiltrated into the unsaturated zone underlying the wash and moved laterally to the sample location. Lateral spreading of infiltrating water is consistent with decreases in water-flux with depth estimated at this site by Nimmo et al. (1996, 1997).

DISCUSSION

Water movement through thick unsaturated zones in desert environments has been subject to increasing scientific study ever since it was proposed that these areas may be suitable for the disposal of radioactive and other hazardous wastes. However, these studies have not addressed water-supply or water-management issues associated with ground water recharge in arid areas. This study focuses on infiltration from intermittent streams where surface water is occasionally present.

Chloride data collected in this study show that water has not infiltrated below the root zone away from Oro Grande Wash for more than 10,000 years. However, the data do show that under present-day conditions, brief intermittent streamflow can infiltrate below the root zone and, presumably, to the water table even in very arid environments such as the Mojave Desert. On the basis of tritium data, this water may require at least 200 to 300 years to reach the water table at the study site. Because water recharged from the wash spreads laterally away from the wash with increasing depth, the flux of water decreases with increasing depth. As a result, under present-day conditions, it may take longer than 200 to 300 years for infiltrating water to reach the water table. This is consistent with the results of previous studies (Izbiicki et al., 1995) that identified a subregional groundwater-flow system containing ground water recharged between 500 and 2,400 years ago.

Although the quantity of recharge water from Oro Grande Wash is small and the length of time required for water to reach the water table is long, it may be possible to supplement natural recharge with imported water. This would increase the quantity of water infiltrated in the wash and decrease the time required for this water to reach the water table. Artificial recharge in washes where natural recharge occurs presents several advantages to artificial recharge at other locations. First, the unsaturated zone is not as dry as most areas underlying the desert. Less water will be lost to storage within the unsaturated zone underlying intermittent streams than in the very dry unsaturated zone away from these streams. Second, there may be reaches underlying some washes where high chloride layers are not present. Poor-quality water and dissolution of soil minerals within the unsaturated zone would not degrade the quality of water recharged in these areas.

Results presented in this paper are part of a larger study of infiltration from intermittent streams at a number of sites in the western part of the Mojave Desert. The purpose of the larger study is to regionalize data collected at this site to other reaches of Oro Grande Wash and to nearby washes such as Sheep Creek Wash, and to estimate the length of time required for artificially recharged water to reach the water table.

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Drywell Recharge and Contaminant Loading
Metropolitan Phoenix, Arizona

Frank Putman (editor)
Arizona Department of Water Resources

Abstract

The Arizona Department of Water Resource conducted a study of drywell recharge and contaminant loading in the urban area of metropolitan Phoenix, Arizona. The study estimated runoff volumes and contaminant loading to drywells, but did not deal with transport through the vadose zone. Activities include a survey of municipal drainage control ordinances, examination and analysis of drillers records and drywell registrations and field verification of data.

This information was combined with estimates of runoff for various land uses using the Rational Method and with available water quality data to derive estimates of the number of existing drywells for each land use type, the amount of runoff disposed of by drywells for each land use area, and the contaminant loading to drywells in each land use type.

About 20,000 to 25,000 drywells were estimated to be in the Phoenix Metropolitan area. Phoenix receives an average of seven inches of rain annually, and about 4500 acre-feet of runoff is diverted to drywells. Industrial area make up 5% of the urban area but contain 17% of the drywells. Commercial are make up 6% of the urban area and contain 41% of the drywells. Residential land uses make up 35% of the urban area and contain 35% of the wells. The remaining drywells are found mostly in public use areas such as schools, parks and golf courses. Existing data on runoff quality indicated that most pollutants received by drywells are total metals associated with sediment load. These metals are generally immobile in the vadose zone and only rarely were groundwater problems noted that were associated with recharge from drywells. Rising groundwater levels in the Phoenix area have a potential to increase the number of drywells in the direct contact with the groundwater and caution should be used in determining the separation distance between the bottom of the drywell and the water table.

An attempt to develop a transferable methodology for such assessments was not successful due primarily to difficulty in definitively linking use to land covers. A prioritization of land use/drywell combinations for future attention was presented.
Estimation of Infiltration and Recharge for Environmental Site Assessment

Health and Environmental Sciences Department

API PUBLICATION NUMBER 4643

PREPARED UNDER CONTRACT BY:

DANIEL B. STEPHENS & ASSOCIATES, INC.
ALBUQUERQUE, NEW MEXICO

JUNE 1996
<table>
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<tr>
<th>Geographic Region</th>
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<th>MAP 1</th>
<th>E 2</th>
<th>Vegetation</th>
<th>Topography</th>
<th>Soil/Aquifer Type</th>
<th>Estimated Annual Recharge (mm/yr [%P])</th>
<th>Estimation Technique</th>
<th>Reference</th>
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<td>810-1370 a</td>
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<td>33 [6]</td>
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¹ MAP refers to the Model Aquifer Properties system, which is used to estimate aquifer properties.
² E refers to the Parameter E, which is used in hydrologic models to estimate recharge rates.

Table A.1. Estimates of diffuse annual recharge by geographic region.
<table>
<thead>
<tr>
<th>Geographic Region</th>
<th>Site</th>
<th>MAP</th>
<th>E</th>
<th>Vegetation</th>
<th>Topography</th>
<th>Soil/Aquifer Type</th>
<th>Estimated Annual Recharge (mm/yr [%P])</th>
<th>Estimation Technique</th>
<th>Reference</th>
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<td>1900</td>
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<td>Basin floor</td>
<td>Unconsolidated sand and gravel</td>
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<td>Soil-water balance</td>
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<tr>
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<td>Amargosa River Basin</td>
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<td>Basin and mountain front</td>
<td>Alluvial basin</td>
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<td>Great Lakes Region</td>
<td>Hiram, Ohio</td>
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<td>610</td>
<td>Abundant</td>
<td>Flat-lying river valley</td>
<td>Sand and gravel</td>
<td>433 [42]</td>
<td>Soil-water balance</td>
<td>Lyford and Cohen, 1988</td>
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<tr>
<td>Great Lakes Region</td>
<td>Penn Yan, New York</td>
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<td>508</td>
<td>Abundant</td>
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<td>Sand and gravel</td>
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<td>1.5-10 [1]</td>
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<td>Sand and gravel</td>
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<td>661 [54]</td>
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*Site Hydrologic Conditions: MAP = Mean Annual Precipitation (mm), E = Evapotranspiration (mm), Vegetation = Type of Vegetation, Topography = Type of Topography, Soil/Aquifer Type = Type of Soil or Aquifer. Estimation Technique = Method used to estimate recharge. Reference = Source of data.
### Table A-1. Estimates of diffuse annual recharge by geographic region

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<th>Geographic Region</th>
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<th>E ²</th>
<th>Vegetation</th>
<th>Topography</th>
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<th>Estimated Annual Recharge (mm/yr [%P])</th>
<th>Estimation Technique</th>
<th>Reference</th>
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<td>1016-1651 ³</td>
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<td>High plains</td>
<td>Ogallala sandstone</td>
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<td>Glacial drift</td>
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<td>No report</td>
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<th>Topography</th>
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<th>Estimation Technique</th>
<th>Reference</th>
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<td>Muskingum Subbasin</td>
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<td>River valley</td>
<td>Alluvium, glacial outwash</td>
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<td>Streamflow measurements 4</td>
<td>Boyd, 1974</td>
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<td>MAP¹</td>
<td>E²</td>
<td>Vegetation</td>
<td>Topography</td>
<td>Soil/Aquifer Type</td>
<td>Estimated Annual Recharge (mm/yr [%F])</td>
<td>Estimation Technique</td>
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<td>Glacial sand and gravels</td>
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<td>Water level fluctuation</td>
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<td>Moraine and lake deposits</td>
<td>51-203 [5.9-23]</td>
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<td>72-236</td>
<td>Groundwater model</td>
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<th>MAP (E^2)</th>
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<th>Topography</th>
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<td>710-1140</td>
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<td>Rolling hills</td>
<td>Basalt with overlying unconsolidated sediments</td>
<td>Up to 190 [-38]</td>
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<td>Mountains and alluvial valleys</td>
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<td>Phillips et al., 1988</td>
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<td>Fine sand, sandy loam</td>
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<td>1780°</td>
<td>Sparse (four-winged saltbush, creosote)</td>
<td>Flat-lying floodplain terrace</td>
<td>Fine sand, sandy loam</td>
<td>2.6-3.0 [1.3-1.5]</td>
<td>Chlorine-36 peak</td>
<td>Phillips et al., 1988</td>
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<td>-1960°</td>
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<td>Flat alluvial plain</td>
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<td>230</td>
<td>2390°</td>
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<td>Las Cruces, New Mexico</td>
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<td>2390°</td>
<td>Sparse (grass and shrubs)</td>
<td>Terrace/pediment</td>
<td>Sandy loam</td>
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<td>Chloride mass balance</td>
<td>Phillips et al., 1988</td>
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<td>2390°</td>
<td>Sparse (grass and shrubs)</td>
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<td>Sandy loam</td>
<td>2.5 [1.1]</td>
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<td>Limestone</td>
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<td>Basin outflow</td>
<td>Fleisher &amp; Nye, 1933</td>
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Table A-1. Estimates of diffuse annual recharge by geographic region
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<table>
<thead>
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<th>Geographic Region</th>
<th>Site</th>
<th>MAP</th>
<th>$E^2$</th>
<th>Vegetation</th>
<th>Topography</th>
<th>Soil/Aquifer Type</th>
<th>Estimated Annual Recharge (mm/yr [%P])</th>
<th>Estimation Technique</th>
<th>Reference</th>
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<tr>
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<td>Roswell Basin</td>
<td>305-373</td>
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<td>Mountain slope to river</td>
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<td>Gross et al., 1979</td>
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<tr>
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<td>385</td>
<td>&gt;1000 a</td>
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<td>Sand</td>
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<td>Stone, 1986</td>
</tr>
<tr>
<td>Rio Grande Basin</td>
<td>San Juan Basin, northwest New Mexico</td>
<td>145</td>
<td>1420 a</td>
<td>Sparse to none</td>
<td>Alluvial terrace</td>
<td>Alluvium</td>
<td>2.29 [1.58]</td>
<td>Chloride mass balance</td>
<td>Stone, 1986</td>
</tr>
<tr>
<td>Rio Grande Basin</td>
<td>San Juan Basin, northwest New Mexico</td>
<td>145</td>
<td>1420 a</td>
<td>Sparse to none</td>
<td>Badlands</td>
<td>Eolian sand, unconsolidated sediments</td>
<td>0.25-0.89 [0.17-0.61]</td>
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</tr>
<tr>
<td>Rio Grande Basin</td>
<td>Sunland Park, New Mexico</td>
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<td>1270 a</td>
<td>Sparse (creosote bush)</td>
<td>Gently sloping escarpment</td>
<td>Sandy alluvium</td>
<td>0.19 [0.095]</td>
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<td>Stephens &amp; Coons, 1994</td>
</tr>
<tr>
<td>Rio Grande Basin</td>
<td>Sunland Park, New Mexico</td>
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<td>1270 a</td>
<td>Sparse (creosote bush)</td>
<td>Gently sloping escarpment</td>
<td>Sandy alluvium</td>
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<td>1270 a</td>
<td>Sparse (creosote bush)</td>
<td>Gently sloping escarpment</td>
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<td>788 a</td>
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<td>Mesa, valley floor</td>
<td>Alluvium and sedimentary bedrock</td>
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<tr>
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<td>Coastal plain and rolling highlands</td>
<td>Unconsolidated and consolidated sediments</td>
<td>150 [10.9]</td>
<td>Streamflow measurements</td>
<td>Cederstrom et al., 1979</td>
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<tr>
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<td>South Carolina</td>
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<td>No report</td>
<td>Abundant</td>
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<td>Clayey</td>
<td>&lt;25</td>
<td>No report</td>
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Table A-1. Estimates of diffuse annual recharge by geographic region
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<table>
<thead>
<tr>
<th>Geographic Region</th>
<th>Site</th>
<th>Site Hydrologic Conditions*</th>
<th>Estimated Annual Recharge (mm/yr [%PI])</th>
<th>Estimation Technique</th>
<th>Reference</th>
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<tr>
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<td>South Carolina</td>
<td>1450</td>
<td>1015 (^a)</td>
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<td>Sand hills of coastal plain</td>
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<td>No report</td>
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<td>Flat coastal plain</td>
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<td>Tennessee River Basin</td>
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<td>760 (^a)</td>
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<tr>
<td>Tennessee River Basin</td>
<td>Coastal Plain/Big Sandy River</td>
<td>1270</td>
<td>760 (^a)</td>
<td>Abundant</td>
<td>River valley and rolling plain</td>
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Table A-1. Estimates of diffuse annual recharge by geographic region
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<table>
<thead>
<tr>
<th>Geographic Region</th>
<th>Site</th>
<th>MAP ¹</th>
<th>E ²</th>
<th>Vegetation</th>
<th>Topography</th>
<th>Soil/Aquifer Type</th>
<th>Estimated Annual Recharge (mm/yr [%P])</th>
<th>Estimation Technique</th>
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<tr>
<td>Texas-Gulf Region</td>
<td>Regional alluvial aquifers</td>
<td>510-1120</td>
<td>510-1520 ³</td>
<td>Moderate</td>
<td>Floodplain</td>
<td>Sand, gravel, and clay</td>
<td>34 [4.2]</td>
<td>Soil-water balance</td>
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<td>Ogallala Aquifer of Texas High Plains</td>
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<td>High plain</td>
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<td>Baker &amp; Wall, 1976</td>
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<td>Ogallala Aquifer at Portales, New Mexico</td>
<td>-355</td>
<td>No report</td>
<td>Sparse</td>
<td>Depression in high plains with sand dunes</td>
<td>Sand and gravel</td>
<td>12.7 [3.6]</td>
<td>Basin outflow</td>
<td>Theis, 1937</td>
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<td>High plains</td>
<td>Sand and gravel</td>
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<td>Sparse</td>
<td>High plains</td>
<td>Sand and gravel</td>
<td>6.4-10.7 [1.8-3.0]</td>
<td>Water level fluctuations</td>
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<td>Region</td>
<td>Site</td>
<td>Geographic Region</td>
<td>Texas</td>
<td>Estimated Annual Runoff (10^3 acre-feet)</td>
<td>Sediment Yield</td>
<td>Vegetation Type</td>
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<td>Mapped?</td>
<td>Site Hydrologic Conditions</td>
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<tr>
<td>Mississippi</td>
<td>Floyd's 1976</td>
<td>Texas-Central North</td>
<td>1740.0</td>
<td>1.2</td>
<td>750</td>
<td>Woody Regime</td>
<td>Sandy Soil</td>
<td>No</td>
<td>Yes</td>
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<td>Floyd's 1975</td>
<td>Texas-Central North</td>
<td>1740.0</td>
<td>1.2</td>
<td>750</td>
<td>Woody Regime</td>
<td>Sandy Soil</td>
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<td>Yes</td>
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<tr>
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<td>Texas-Central North</td>
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<td>750</td>
<td>Woody Regime</td>
<td>Sandy Soil</td>
<td>No</td>
<td>Yes</td>
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<tr>
<td>Lower Mississippi</td>
<td>Floyd's 1975</td>
<td>Texas-Central North</td>
<td>1740.0</td>
<td>1.2</td>
<td>750</td>
<td>Woody Regime</td>
<td>Sandy Soil</td>
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Table A.1. Estimates of Diffuse Annual Runoff by Geographic Region.
<table>
<thead>
<tr>
<th>Geographic Region</th>
<th>Site</th>
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<th>Vegetation</th>
<th>Topography</th>
<th>Estimated Annual Recharge (mm)[R/R]</th>
<th>Estimation Technique</th>
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<td></td>
<td>Chippewa Block Subbasin</td>
<td>792</td>
<td>Abundant</td>
<td>Alluvial, glacial outwash</td>
<td>90 [12]</td>
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<td></td>
<td>Big Muddy Subbasin</td>
<td>1067</td>
<td>Abundant</td>
<td>Alluvial, glacial outwash</td>
<td>32 [3]</td>
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</tr>
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<td>Upper Mississippi River Basin</td>
<td>Mississippi Headwaters</td>
<td>680</td>
<td>Abundant</td>
<td>Alluvial, glacial outwash</td>
<td>90 [14]</td>
<td>Streamflow measurements</td>
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<tr>
<td></td>
<td>Meramec Subbasin</td>
<td>650</td>
<td>Abundant</td>
<td>Alluvial, glacial outwash</td>
<td>90 [14]</td>
<td>Streamflow measurements</td>
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<td>Upper Mississippi River Basin</td>
<td>Lake County, Illinois</td>
<td>70</td>
<td>No report</td>
<td>Glacial outwash, sand and gravel</td>
<td>76-310</td>
<td>Soil temperature</td>
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<td>Lake County, Illinois</td>
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<td>Glacial outwash, sand and gravel</td>
<td>76-310</td>
<td>Basin outflow</td>
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Table A-1: Estimates of diffuse annual recharge by geographic region
Table A-1. Estimates of diffuse annual recharge by geographic region
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<thead>
<tr>
<th>Geographic Region</th>
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<th>Topography</th>
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<th>Estimation Technique</th>
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<tr>
<td>Upper Mississippi River Basin</td>
<td>Lake County, Illinois</td>
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<td>No report</td>
<td>Moderate</td>
<td>River valley</td>
<td>Glacial drift (clay, sand and gravel); dolomite bedrock</td>
<td>80-107</td>
<td>Basin outflow (^a)</td>
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<td>Des Moines Subbasin</td>
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<td>686 (^a)</td>
<td>Abundant</td>
<td>River valley</td>
<td>Alluvium, glacial outwash</td>
<td>24 ([3])</td>
<td>Streamflow measurements (^4)</td>
<td>Bloyd, 1975</td>
</tr>
<tr>
<td>Upper Mississippi River Basin</td>
<td>Skunk Subbasin</td>
<td>813</td>
<td>686 (^a)</td>
<td>Abundant</td>
<td>River valley</td>
<td>Alluvium, glacial outwash</td>
<td>50 ([6])</td>
<td>Streamflow measurements (^4)</td>
<td>Bloyd, 1975</td>
</tr>
<tr>
<td>Upper Mississippi River Basin</td>
<td>Cannon-Zumbo-Root Subbasins</td>
<td>737</td>
<td>635 (^a)</td>
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<td>River valley</td>
<td>Alluvium, glacial outwash</td>
<td>74 ([10])</td>
<td>Streamflow measurements (^4)</td>
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<tr>
<td>Upper Mississippi River Basin</td>
<td>Iowa-Cedar Subbasins</td>
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<td>686 (^a)</td>
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<td>Alluvium, glacial outwash</td>
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<td>Streamflow measurements (^4)</td>
<td>Bloyd, 1975</td>
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<tr>
<td>Upper Mississippi River Basin</td>
<td>Salt Subbasin</td>
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<td>Bloyd, 1975</td>
</tr>
<tr>
<td>Upper Mississippi River Basin</td>
<td>Wapsipinicon Subbasin</td>
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<td>660 (^a)</td>
<td>Abundant</td>
<td>River valley</td>
<td>Alluvium, glacial outwash</td>
<td>80 ([10])</td>
<td>Streamflow measurements (^4)</td>
<td>Bloyd, 1975</td>
</tr>
<tr>
<td>Upper Mississippi River Basin</td>
<td>Fox-Wyaconda-Fabius Subbasins</td>
<td>889</td>
<td>737 (^a)</td>
<td>Abundant</td>
<td>River valley</td>
<td>Alluvium, glacial outwash</td>
<td>18 ([2])</td>
<td>Streamflow measurements (^4)</td>
<td>Bloyd, 1975</td>
</tr>
</tbody>
</table>

* Climate classifications according to Thornthwaite (1948):
  Arid: MAP/E <0.5
  Semiarid: 0.5 < MAP/E < 1.0
  Humid: MAP/E >1.0

\(^1\) Mean annual precipitation (mm/yr)
\(^2\) Evaporation (mm/yr)
\(^3\) Soil-water balance computations based on data provided for mean annual precipitation, evapotranspiration and runoff

\(^4\) Assumes a 60-percent flow parameter equals baseflow
\(^5\) Average recharge estimated from average baseflow
\(^6\) Range of recharge estimated from average baseflow
\(^7\) Recharge, approximated by steady-state yield, may be overestimated
\(^8\) Where basin outflow is limited to known pumping discharge

\(^a\) Potential evapotranspiration
\(^b\) Complimentary relationship areal evapotranspiration (CRAE) (Morton, 1993)
\(^c\) Lake evaporation
\(^d\) Report of annual recharge estimate as %P based on MAP reported by Zuroski (1978)
REFERENCES


R-9


LOCALIZED NATURAL RECHARGE

IN ARID BASINS

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Contribution to

“Natural Recharge of Groundwater Symposium”

June 2, 2000
Tempe, Arizona
LOCALIZED NATURAL RECHARGE IN ARID BASINS

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EXTENDED ABSTRACT

This abstract and my presentation are based on two earlier papers. The first one is Chapter 2 “Recharge from precipitation” by Hendrickx and Walker in “Recharge of Phreatic Aquifers in (Semi-) Arid Areas”, I. Simmers (Editor), 1997, Volume 19:19-114, International Association of Hydrogeologists, Balkema, Rotterdam, The Netherlands (balkema@balkema.nl). The second paper is “Uniform and preferential flow mechanisms in the vadose zone” by Hendrickx and Flury. This publication is a contribution to the “Conceptual Models of Flow and Transport in the Fractured Vadose Zone”, National Academy of Sciences, Irvine, California, March 18-19, 1999. It will be published this year by the U.S. National Committee for Rock Mechanics, National Academy of Sciences/National Research Council. For more details and figures I refer to these two publications.

In comparison to direct recharge, localized and indirect recharge are often considered at least as significant if not the most important sources of natural recharge in arid and semi-arid lands (Gee & Hillel 1988; Lerner et al. 1990; Stephens 1994; Wood & Sanford 1995). Localized recharge implies horizontal movement of surface and/or near-surface water and occurs in weathered bare hardrock or limestone terrain, topographical depressions, minor wadis or arroyos, and in mountain front systems. To account for localized recharge it is necessary to measure or estimate local runoff and runon volumes so that these can be included in the water balance. Such estimates and measurements are often complicated by subsurface components of runoff and runon flow (Anderson & Burt 1990) and by the fact that they frequently occur on a scale too detailed to map for engineering studies (Lerner et al. 1990).
Localized Recharge at Landscape Scale is often detectable by observing topographic features. A classical example occurs in the numerous depressions dotting the Great Plains of North America. These features can measure tens to thousands of meters across, are often occupied by wetlands or lakes and are referred to as "potholes", "sloughs", or "playas". Meyboom (1966) was one of the first investigators to quantify localized recharge during a one-year study of a till plain pothole with a watershed contributing area of 0.8 ha in south-central Saskatchewan (Canada). His pothole has a bottom diameter of 40 m and the height of its surrounding rim varies between 3 and 8 m. It was determined that the pothole or local depression, with 15% of the total surface area, contributed 70% of the recharge. Hayashi et al. (1998) present measurements about the fate this localized recharge. They found that most of the infiltrated water flows laterally through the unconfined aquifer in the shallow subsurface to the wet margin of the pond and further to the upland, where it is consumed by evapotranspiration without recharging the deep ground water. Other investigators have also reported studies that confirm the large contribution of localized recharge described by Meyboom (1966); examples are Freeze & Banner (1970), Miller et al. (1985), Winter (1986). Using chemical techniques, Wood & Sanford (1995) estimated that approximately half (4 to 5 mm/year) the annual recharge (9 to 10 mm/year) to the Ogallala Aquifer on the southern High Plains in the U.S.A. occurs through playa floors that cover only 6% of the area.

The effects of topographic, soil, and climatic conditions on the magnitude of depression focused recharge for specific sites is difficult to measure in the field. Therefore, Nieber et al. (1993) and Boers (1994) have developed mathematical models. Nieber et al. (1993) assumed for simplicity that their catchment was circular in form and contained a circular shaped depression with one single drainage outlet. Boers (1994) developed a similar procedure for the design of rainwater harvesting catchments in arid and semi-arid zones that can also be used for the assessment of localized recharge. His method is based on actual evapotranspiration predictions using a numerical soil water balance model, while the runoff component is predicted by a runoff model (Boers et al. 1986). He defines a micro-catchment that consists of a runoff area with a maximum flow distance of 100 m and an adjacent basin area (the depression) with a tree, bush, or row crop. The objective of rainwater harvesting is to induce runoff, collect and store the water
in the basin area and conserve it in the root zone for consumptive use by the vegetation. The components of the annual water balance are:

\[ D = P + R - E_i - E_w - E_{act} - T_{act} - \Delta W \]  

(1)

where \( D \) is deep percolation or recharge, \( P \) is precipitation, \( R \) is runoff calculated over the basin area where it is collected, \( E_i \) is evaporation of water intercepted by the vegetation, \( E_w \) is open water evaporation, \( E_{act} \) is the evaporation from bare soil, \( T_{act} \) is actual transpiration by the vegetation and \( \Delta W \) is the increase in soil water storage in the root zone. Table 1 presents the annual water balance components at Sadoré (Niger) for one Neem tree in a basin of 8 m² with a soil profile comprising 3 m fine sand above 2 m laterite gravel. If the basin receives no runon water, no recharge takes place during a dry, average, or even a wet year. In a dry year, runoff areas of 20 and 40 m² do not generate any recharge although they increase actual transpiration and, thus, the growth rate of the tree. In an average year a modest increase in runoff area from 20 to 40 m² produces a twenty-two fold increase in recharge from 5 to 113 mm, whereas in a wet year a similar increase in area ratio produces a four-fold increase in recharge from 38 to 185 mm. These results and those presented by Tosomeen (1991) demonstrate the great sensitivity of localized recharge to rather small changes in topography, soil type and climate.

Localized recharge can also take place as a result of discontinuities in subsurface layers. An example of this situation are pipes through calcic horizons in New Mexico. The La Mesa surface in southwestern New Mexico has developed on Rio Grande deposits mainly consisting of sands and gravels. The surface was abandoned by the Rio Grande River during the middle Pleistocene and since that time a calcic soil with an indurated calcic horizon has formed. A characteristic of such an indurated calcic horizon is the presence of pipes that develop through the horizon to the underlying sediments. Most of the upper soil profile has been eroded exposing the calcic horizon. However, in the last few hundred years the exposed calcic horizon has been covered with sand dunes with a loamy sand texture. The sand has taken the form of coppice dunes which develop underneath the shrub vegetation. Consequently, the pipes in the carbonate horizon are mostly buried underneath 0.5-1.5 m of loamy sand and are impossible to identify.
Table 1: Annual precipitation ($P$), predicted runoff ($R$), actual transpiration ($T_{act}$) and recharge ($D$) in mm for one Neem tree in a 8 m$^2$ basin at Sadoré, Niger, for precipitation only and for precipitation and runoff from 20 m$^2$ and 40 m$^2$ runoff areas in three different precipitation years (Boers, 1994).

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<tr>
<td>$P$</td>
<td>545</td>
<td>545</td>
</tr>
<tr>
<td>$R$</td>
<td>0</td>
<td>232</td>
</tr>
<tr>
<td>$T_{act}$</td>
<td>409</td>
<td>633</td>
</tr>
<tr>
<td>$D$</td>
<td>0</td>
<td>5</td>
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<tr>
<td><strong>Dry year</strong></td>
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<tr>
<td>$P$</td>
<td>258</td>
<td>258</td>
</tr>
<tr>
<td>$R$</td>
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from the surface. Rodriguez-Marin and Hendrickx observed in October 1998 on the La Mesa Surface pipe densities from one per 14 to 38 m² along two transects of 5.6 and 3.2 km in a 2.2 m deep trench dug for a gas pipeline. Since the pipes cover approximately 15 to 20 percent of the total surface, the ratio between catchment area (around the pipe) and through flow area (inside the pipe) varies from about 7 to 5 on the La Mesa surface. It is anticipated that such localized pipes serve as preferential flow conduits for water flow and may increase ground water recharge by an order of magnitude.

Harrison and Hendrickx have observed pipes in highway cuts near Hatch and Albuquerque and along escarpments with exposed calcic horizons near Socorro and El Paso. These field observations along a 400 km long stretch from El Paso to Albuquerque indicate that pipes through indurated calcic horizons are widespread in New Mexico and West Texas.

*Localized Recharge through Macropores* is not so easy to detect in the landscape. Although its impact on the total amount of ground water recharge may be relatively small, macropore flow can have a major impact on contaminant movement. For a good understanding of the different flow processes in the vadose zone, it is useful to recognize the two different flow mechanisms that drive potential groundwater recharge through the vadose zone: a capillary and a viscous one. Capillary flow takes place in pores with a diameter less than approximately 3 mm in which capillary forces, together with gravity, determine the flow process. A porous medium in which capillary forces dominate behaves like a sponge; i.e. no free drainage occurs even at high water contents and capillary rise causes water to move upwards against the pull of gravity. In macropores that are easiest defined as pores with a diameter or width larger than 3 mm (Germann, 1990), the effects of capillarity are no longer felt and the flow process is dominated by viscous forces and gravity. *Macropore* is a common name for a wide range of large pores such as cracks in clay soils, rock fractures, fissures in sediments, worm holes and old root channels. Saturated flow through macropores can be quantified using Poiseuille's equation.

The principal difference between capillary and viscous flow mechanisms from the view point of groundwater recharge is the difference in the velocity with which water moves from the soil surface to the water table. For example, consider a 10 m deep vadose zone of fine sand with a volumetric soil water content of 0.15 and unsaturated hydraulic conductivity of 0.15 cm/day.
arid regions it is common that the hydraulic gradient below the root zone is equal to unity so that the water flux becomes equal to the hydraulic conductivity. Since the water velocity is equal to the flux divided by volumetric water content, i.e. \(0.15/0.15=1\) cm/day, it will take 1000 days for a water particle to travel from the soil surface to the groundwater table. Because such capillary flow takes places in small pores under negative soil water pressures that prevent drainage into larger pores, all macropores are empty. However, after a high intensity precipitation producing some local runoff, macropores that start at the soil surface will quickly become conduits for the runoff water. Considering a "small" macropore with a diameter of only 1 mm that goes direct from the surface to the water table, one can calculate with the viscous flow model given by Poiseuille's law (e.g. Jury, et al. 1991) that water particle velocity in the macropore is approximately 30 cm/sec. Thus it takes only 33 seconds for the particle to travel from a ponded soil surface to the water table through the macropore. This means that the difference in travel time between a capillary flow mechanism through the soil matrix and the viscous one through macropores is approximately six orders of magnitude. It is therefore not surprising that the importance of flow through fractures and macropores was recognized as early as 1864 by Schumacher in his book "Die Physik des Bodens" (Germann, 1990) and was verified eighteen years later with measurements at Rothamsted (England) by Lawes, et al. (1882). However, this early work was overshadowed by theoretical and experimental work on capillary flow through the soil matrix until the late 1970s when environmental concerns prompted a renewed interest in macropore flow because it could cause immediate shallow aquifer contamination (e.g. Bouma & Raats, 1984; Evans & Nicholson, 1987).

The process of macropore flow is similar to localized recharge - albeit on a much smaller scale - since horizontal water movement is required. When the overall water input from precipitation or irrigation, \(q^*(t)\), exceeds infiltration capacity of the soil, \(i(t)\), a horizontal overland flow, \(o(t)\), is generated that causes a water flux into the macropores, \(q(0,t)\). This flux causes water content inside the macropore, \(w(z,t)\), to increase. A fraction of the water, \(r\), that occupies a macropore at a given depth will be absorbed by the soil matrix through the macropore walls. The remainder will percolate downwards into the macropore, \(q(z,t)\). The interplay of precipitation or irrigation rates with dynamics of the infiltration rate over time add to the
macropore flow mechanism complexity. When the infiltration rate, \( i(t) \), decreases with time and with increasing antecedent soil water content, the opportunity for overland flow, \( o(t) \), and macropore flow, \( q(0,t) \) increases. Different aspects of macropore flow are illustrated in the following case studies.

**Case 1:** German (1986) analysed drainage responses to storms observed during a seven year period in the Coshocton monolith lysimeters (Northeast Experimental Watershed, USDA-ARS, Coshocton, Ohio, U.S.A.). Rains of only 10 mm/day caused a drainage response at 2.4 m depth on the same day as precipitation when volumetric water content in the upper 1 m of the undisturbed profile exceeded a threshold value of 0.3 m\(^3\)/m\(^3\), whereas at soil water contents below this threshold value storms greater than 50 mm/day were found not to cause any drainage flow.

**Case 2:** Gunn (1983) investigated by which mechanisms flow is concentrated and transmitted to the underlying aquifer for a karst area in the Waitomo district of New Zealand. Although his study was conducted in an area with mean annual precipitation and potential evapotranspiration of 2370 mm and 775 mm respectively, it reveals a number of horizontal flow and fracture flow mechanisms that are relevant for recharge processes in arid zones after high intensity storms. He found that closed depressions (solution dolines, sinkholes, cockpits) act as funnels and collect near surface water through three concentrating mechanisms: (1) overland flow, defined as any water flowing along the ground surface; (2) throughflow, defined as any water flowing laterally within the soil; and (3) subcutaneous flow, defined as water flowing laterally through the upper, weathered layer of limestone. This is an example of localized recharge where the horizontal movement of water not only takes place at the soil surface, but also below it.

Three vertical flow mechanisms were recognized for water transmission through the vadose zone: (1) film flow (named shaft flow by Gunn), defined as water flowing underground as films on the walls of vertical shafts; (2) fracture or macropore flow (named vadose flow by Gunn), defined as vertically moving water which flows for a major part of its course in enlarged joints and fractures; (3) vadose seepage, defined as vertically moving water which percolates through small, tight joints and fissures or as intergranular flow. Gunn (1983) combined these six
flow mechanisms into a conceptual vadose zone flow model and qualitatively validated it by determining water travel times for each flow mechanism using measurements of water temperature and calcium and magnesium ion concentrations. This validation revealed the principal characteristics for each model component.

Overland flow is quantitatively insignificant in the Waitomo district, while subcutaneous flow appears to supply more water to vertical shafts than does throughflow. Travel times for subcutaneous and throughflow vary between 0 and 14 weeks with a mean of approximately 6 weeks. The concentration of near-surface water makes film flow more important than fracture flow, while vadose seepage contributes less than 5% to aquifer recharge. Two different categories of fracture flow were identified: flows through open joints and fissures, and flows through soil-filled fissures. The first responded directly to rainfall with a mean travel time of less than a week, whereas the second had a mean travel time of 8 weeks. The travel times for vadose seepage varied between 0 to 19 weeks, indicating that part of the water flows rapidly through small fractures as macropore flow while the remainder flows slowly through the porous material as capillary flow. Although Gunn (1983) cautions that in other climate-soil-regolith regimes the relative importance of these flow mechanisms may be different, his conceptual model and field observations offer much insight to the dynamics of flow processes through fractured rocks. Striking is the fact that capillary flow accounts for less than 5% of recharge volume.

Case 3: The simultaneous occurrence of capillary and macropore flow within the same soil mass without the presence of clearly defined macropores appears to be quite typical. Under these conditions macropore flow cannot be detected from visual observations in the field, but is inferred from analysis of solute profiles or drainage responses at the groundwater table as seen in the first case study. Johnston (1987a, b) used solute profiles in combination with groundwater table measurements to demonstrate macropore flow in a deep clayey regolith in southwest Western Australia. The regolith showed marked heterogeneity over horizontal and vertical distances of only a few metres which resulted in a complex water movement pattern. Thirteen vertical distributions of natural chloride were used to estimate recharge rates through the 16 m deep unsaturated vadose zone. Although over most of the 700 m² experimental area recharge
rates varied from 2.2 to 7.2 mm/year, a small portion of the site had rates between 50 to 100 mm/year. As a result of this preferred flow a groundwater mound was observed in piezometer 1356 below the localized recharge area within 12-14 hours of intense rainstorms and dissipated over a period of 2-4 days.

Case 4: Scanlon (1992) demonstrated the existence of preferred pathway flow through fissured sediments in the Chihuahuan desert of Texas (U.S.A.), with annual precipitation and approximate potential evapotranspiration of 280 mm and 3000 mm respectively. The term fissure refers to the alignment of discontinuous surface collapse structures, or gulleys: the underlying extensional feature is filled with sediment. A somewhat unexpected result from Scanlon's study is that water fluxes calculated from vertical chloride profiles in the fissured sediments ranged from 1 to 8 mm/year and were as much as 350 times higher than those calculated for adjacent ephemeral stream sediments. This difference could not be attributed to differences in soil texture as fissures and ephemeral streams have similar textures varying from muddy, sandy gravel to loamy sand and clay loam. However, open cavities and fractures in the fissured sediments were filled with loose sediments to give lower bulk densities and looser soil structure, so that fast macropore flows are promoted rather than slow capillary flows. Much of the precipitation occurring as high-intensity convective summer storms infiltrates to depths of only 0.1-0.3 m in the ephemeral streams beds and is readily lost by evapotranspiration, whereas deep infiltration in the fissured sediments prevents such losses and results in a much larger recharge rate.

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Arizona Hydrological Society, Arizona Department of Water Resources,

June 2, 2000 Tempe, Arizona

NATURAL RECHARGE IN THE CADIZ AREA,
SAN BERNARDINO COUNTY, CALIFORNIA

by

Dennis E. Williams¹

Project Overview

The Metropolitan Water District of Southern California (MWD), in conjunction with Cadiz, Inc., is evaluating the feasibility of operating a groundwater storage and transfer project near Cadiz, California in the eastern Mojave Desert. Specifically, the proposed Project area is located in the eastern Mojave Desert approximately 200 miles east of Los Angeles, 60 miles southwest of Needles, and 40 miles northeast of Twentynine Palms. The program area is within and around Fenner Gap, located between the Marble and Ship mountains east of Cadiz, California. (see Figure 1). The Project would entail transporting surplus water from the Colorado River Aqueduct (CRA) and artificially recharging it through a series of surface spreading basins located in the vicinity of Fenner Gap. A pipeline would transport water from the CRA during storage operations (i.e. "put" operations) as well as transport pumped groundwater from the Cadiz area back to the CRA during withdrawal operations (i.e. "take" operation). The Project has been given the title "Cadiz Groundwater Storage and Dry-Year Supply Program."

Geologic and Hydrologic Setting

Geology, Geohydrology

The Fenner Gap area is located within a topographically closed drainage system that includes three main drainage basins or watersheds: Bristol, Cadiz, and Fenner. These watersheds are considered

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one drainage system because all surface and groundwater within them drains to a central lowland area (i.e. Bristol and Cadiz dry lakes). The Bristol, Cadiz, and Fenner watershed system is separated from the surrounding watersheds by topographic divides (generally mountain ranges). Surface and groundwater on the other sides of these divides (i.e. outside of the three watershed area) flows toward other areas.

The total area of the Bristol, Cadiz and Fenner watershed system is approximately 2,710 square miles. The Bristol watershed is 1,170 square miles, the Cadiz watershed is 540 square miles, and the Fenner watershed is 1,000 square miles. Groundwater flow within the Bristol watershed flows toward, and terminates in, Bristol Dry Lake. Groundwater flow within the Cadiz watershed flows toward, and terminates in, Cadiz Dry Lake. Groundwater flow within the Fenner watershed flows through Fenner Gap. Some of the groundwater flowing through Fenner Gap migrates toward Bristol Dry Lake and some flows toward Cadiz Dry Lake.

In the Fenner Gap area, the principle geologic deposits that store and transmit groundwater (i.e. aquifers) can be divided into three units: an upper alluvial aquifer, a lower alluvial aquifer, and a bedrock aquifer. In general, the three units are in hydraulic continuity with each other and the separation is primarily due to stratigraphic differences. The upper aquifer consists mainly of Quaternary alluvial sediments ranging in thickness from 200 to 800 feet. The upper aquifer is very permeable in places and can yield 3,000 gallons per minute or more to wells with less than 20 feet of drawdown. The lower alluvial aquifer consists of older sediments that yield water freely to wells but are generally less permeable than the upper aquifer sediments. The maximum thickness of the lower aquifer is unknown but may reach over 6,000 feet in the vicinity of Bristol Dry Lake. Based on findings during recent drilling work in the Fenner Gap area, carbonate bedrock of Paleozoic age, located beneath the lower alluvial aquifer, contains groundwater and is considered a third aquifer unit. Groundwater movement and storage in the carbonate bedrock aquifer primarily occurs in secondary porosity (i.e., cracks, faults and cavities that have developed in the rocks over time). However, the extent, potential yield, and storage capacity of this aquifer have not been tested.

**Climate**

The eastern Mojave Desert is characterized as an arid desert climate with low annual precipitation, low humidity, and relatively high temperatures. Winters are mild and summers are hot with a relatively large range in daily temperature. Temperature and precipitation vary greatly with altitude, with lower temperatures and higher precipitation amounts in the higher elevations. Due to the desert climate of the Bristol, Cadiz, and Fenner watershed area (with high daytime temperatures and low amounts of precipitation), surface drainages do not flow water throughout the year (Freiwald, 1984). The primary natural sources of replenishment to the groundwater in the area are rain and melting snow runoff in the higher elevations of the hills and mountains surrounding the

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basins and infiltration in the washes during storm events (Izbicki et al., 1997). Surface runoff that does not infiltrate into the ground or evaporate, eventually reaches either Bristol or Cadiz dry lakes where it ponds temporarily before evaporating. Most of the precipitation in the eastern Mojave Desert accumulates during the winter months of November through March (Thompson, 1929). Early summer and late fall are typically periods of little rainfall. However, the frequency and intensity of rainfall from year to year is unpredictable.

Winter rainfall is more or less steady, falling in events with a duration of several hours to a day or more. In the higher mountains, snow frequently accumulates during these months. The amount of precipitation in the Bristol, Cadiz, and Fenner watersheds vary with differences in altitude. Average annual precipitation ranges from approximately 3 inches on the Cadiz and Bristol dry lakes (elevations of 545 to 595 feet amsl) to over 12 inches in the Providence and New York mountains (elevations over 7,000 feet amsl). However, most of the study area receives, on average, 4 to 6 inches of rain annually (CDWR, 1975; Freiwald, 1984; Bedinger et al., eds., 1989).

Methods Used to Estimate Natural Recharge

Estimates of the amount of groundwater recharge replenishing the Bristol, Cadiz, and Fenner watershed areas are still being refined and are based on a number of different methods. These methods range from simplistic estimates involving area and precipitation calculations to complex relationships between daily precipitation, evapotranspiration, soil moisture considerations and surface and subsurface runoff. The combined result of these methods is being used to estimate the range of natural recharge and the project area safe yield. In the subsequent sections, emphasis is placed on recharge in the Fenner basin as the project area safe yield is more dependent on this recharge estimate than other areas.

In general, the methods may be grouped into two categories:

1 - Lumped Parameter Methods

2 - Distributed Parameter Methods

Lumped Parameter Methods

Lumped Parameter Methods include basin-wide estimates resulting from calculations involving gross areas, precipitation or evaporation factors. Other basin-wide methods include empirical data and curves relating precipitation and recharge.
Estimates Based on a Percentage of Precipitation or Evapotranspiration

In a closed hydrologic system one method used to estimate recharge is simply to assume that groundwater recharge is generally in the range of 3% to 5% (and as high as 8%) of total average annual precipitation within a watershed. This can be done either on a gross area x precipitation x percentage basis or by weighting precipitation, elevation and area.

Another simplistic calculation applicable to closed hydrologic systems such as the Cadiz, Fenner and Bristol basins is to estimate evapotranspiration since evapotranspiration is approximately equal to recharge. Relatively high temperatures, low humidity, and frequent strong winds cause a high rate of evaporation in the Bristol/Cadiz/Fenner watersheds (CDWR, 1979). Surface water evaporation (as measured using an evaporation pan) is the greatest in the valleys and the lowest in the higher mountain weather stations. Measured evaporation rates are available for the Amboy and Iron Mountain weather stations. The CDWR reports an average annual evaporation of 158 inches at the Amboy station, and an average of 119 inches at the Iron Mountain station (Table 1; US Ecology, 1989).

Evaporation rates from Bristol and Cadiz dry lakes were estimated based on fresh-water evaporation pan data corrected for both large surfaces as well as salinity of the local groundwater beneath the dry lakes. The depth to groundwater was then determined and empirical relationships between potential evapotranspiration, depth to groundwater and evapotranspiration rate applied to estimate the range of potential evapotranspiration from the two dry lakes.

Maxey-Eakin Method

The Maxey-Eakin method was developed to estimate the recharge for the basins in eastern Nevada. Maxey and Eakin established a relationship between precipitation zones and recharge based on 21 valleys in Nevada. Maxey and Eakin constructed the system using six zones of assumed equal rainfall.

It is a common mistake for the Maxey-Eakin method to be applied outside of Nevada, and that it should not be utilized to estimate recharge outside of this area unless that watershed has been independently calibrated. One method which may be used to calibrate the Maxey-Eakin method in a closed hydrologic system is to estimate evapotranspiration with the assumption that evapotranspiration is approximately equal to recharge. When data validation can be provided for the Maxey-Eakin model, the result is empirical. An empirical model is derived from practical experience rather than basic theory. Therefore, a validated Maxey-Eakin model in one groundwater basin does not necessarily translate to a different one.
Some investigators have characterized the Maxey-Eakin model as "a simple linear model". However, data actually form a power function with an equation of the form (Davisson, 2000):

\[ y = 8.0 \times 10^{-10} x^{4.1}, \quad r^2 = 0.998. \]

Where:

- \( y \) = annual recharge (mm)
- \( x \) = annual precipitation (mm)

This equation is empirical and does not resemble more simple expressions typically describing theoretical processes in nature.

It is possible that a power function can describe the relationship between precipitation rates and recharge rates for many groundwater basins in desert regions. However, as commonly observed when using empirical equations, the exponent and correlation coefficient of the power function probably vary among different basins. These variations would likely record differences in percolation rates (i.e. different lithology), rates of precipitation, and type of precipitation (i.e. snow versus rain).

Modification of the Maxey-Eakin method was made which incorporates the underlying physical processes governing the variation in precipitation as a function of elevation. Mean annual precipitation-elevation data were plotted for 70 sites throughout southeastern California and the southern half of Nevada. Data were obtained from the Desert Research Institute's Western U.S. Climate Historical Summaries (http://www.wrcc.sage.dri.edu/climsun.html), which represent long-term records for each station, most of which are normalized from 1961-1998.

An exponential regression through these data yields a correlation coefficient of \( r^2 = 0.43 \). Such a poor correlation coefficient is consistent with previous observation, for example Hevesi et al. (1992), who suggested effects of rain-shadows, latitude-elevation changes, and the types and rates of high elevation precipitation may contribute to the scatter. The less data used in the regression typically results in better correlation coefficients, but may result in increased error due to data limitations.

In order to better understand what causes the data scatter, precipitation-elevation data were correlated to geographic regions. It should be noted that the distribution of precipitation in southeastern California is controlled by two first-order processes:

1) The Sierra Nevada create an obvious rain shadow effect for southeastern California and Nevada, particularly during winter storm events originating from the Pacific Ocean. The severity of this effect varies with proximity to the Sierra Nevada Mts.
2) The ratio of winter to summer storm events decreases from a west to east direction (see Figure 2).

The significance of either of these processes in any given area varies as a function of longitudinal position. Therefore, the precipitation-elevation data was separated into longitudinal groupings comprising those > 116 deg W, those between 116 deg W and 115 deg W, and those < 115 deg W. The 116 deg W demarcation is an important low elevation trend extending south from the Salton Sea trough and north toward Hot Creek Valley in Central Nevada. Prominent high elevation areas such as the Fenner watershed and the Spring Mountains lie east of 116 deg W, while the San Bernardino Mts., the Sierra Nevada, and the White Mts. lie to the west. These latter three mountain ranges create most of the rain shadow effects of western Nevada and the western Mojave Desert during winter storms. However, high elevation regions east of 116 deg W receive the bulk of the precipitation during summer storm events originating from the Gulf of California.

In summary, analysis shows that Maxey-Eakin calibrations are geographically dependent, a result expected for an empirical model. Using the modified Maxey-Eakin recharge and precipitation-elevation curves, a range in annual recharge in the Fenner watershed was estimated which is consistent with natural recharge predicted by both the watershed model and the calibrated groundwater flow model as well as other simplistic estimates.

Chloride Mass Balance

The chloride concentration of mountain front recharge decreases with increasing recharge amounts. Subsequent mixing of this recharge with low elevation recharge sources in the washes skews the groundwater chloride concentration toward the low elevation concentration. This mixed chloride concentration will suggest higher evaporation rates and will result in a lower estimated recharge rate using a chloride mass balance approach.

The chloride concentration for precipitation used in the chloride mass balance was based on data collected in the Mojave Desert areas (Feth, 1967, USGS). Rosen (1989) also used Feth's data to calculate chloride budgets for Bristol Dry Lake. The area used for the chloride mass balance was the entire Fenner Watershed, because the recharge from precipitation in the Fenner Watershed either infiltrates directly in the mountain areas or as runoff infiltrating in the sandy bottom washes.

The range of recharge to the Fenner Watershed based on the chloride mass balance falls within the range of recharge as determined by other estimates.
Distributed Parameter Methods

Watershed model

A watershed model is a detailed water budget which takes into account all of the quantifiable variables that affect the water balance of the watershed. These variables include daily precipitation, infiltration, runoff, vegetation interception, evapotranspiration, soil moisture, and percolation. Use of a watershed model to predict catchment behavior is not new, although manipulation of the large databases required to develop the models was not practical until the creation of high speed computers. The first model developed is a widely used digital (computer-based) model of catchment behavior, designed by researchers at Stanford University in the 1960s (Crawford et al., 1966). This was soon followed by the development of a mathematical model by the United States Geological Survey (USGS) (Dawdy and O'Donnell, 1965). The Stanford Watershed Model, now called HPSF (Hydrological Simulation Program - Fortran), has since been updated to include water quality and sediment transport packages, and is widely used by both the United States Environmental Protection Agency (USEPA) and the USGS.

A watershed model provides a detailed accounting of the water balance of a basin by combining numerous input parameters (i.e., precipitation, runoff, evapotranspiration, soil moisture, and percolation) which affect the water balance of a particular basin. The USGS and Stanford watershed models begin by calculating the amount of water infiltrating into the groundwater basin. That amount is subtracted from the total volume of water available to the watershed, and the models then determine the surface runoff and other components of the water balance as fractions of the remaining water. On the other hand, the Fenner watershed model first determines the runoff component of the water balance. Once established, the runoff is subtracted from the total amount of water, and all other components (i.e., infiltration, vegetation interception, evapotranspiration, soil moisture, and percolation) are calculated based on the amount of water remaining.

GEOSCIENCE constructed the Fenner watershed model using guidelines set out in the San Bernardino County Hydrology Manual, which was itself based on guidelines developed by the United States Department of Agriculture in the Soil and Conservation Service National Engineering Handbook (USDA, 1969).

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2 The runoff component is calculated based on data from the San Bernardino County Hydrology Manual (Williamson and Schmid, 1986a).
This approach is widely accepted throughout Southern California, and has been used in several similar watershed assessments (e.g., Riverside County Flood Control and Water Conservation District, 1978; Williamson and Schmid, 1986b).

The Fenner watershed model was created in order to estimate recoverable water on the basis of different characteristics (parameters) of the Fenner watershed. The first step in the development of the Fenner watershed model was to define the boundaries of the watershed itself. In order to do this, subareas of the watershed were grouped by soil type (type A, B, C, or D as defined on the maps of the San Bernardino Hydrology Manual) and vegetation density (determined from field reconnaissance and examination of aerial photographs). In the case of the Fenner watershed, the soil type D group was further subdivided in the Providence Mountains to account for higher precipitation in areas of higher elevation. A total of five subareas were identified in the Fenner watershed.

Once the subareas had been identified, a long-term isohyetal map was constructed based on precipitation records from weather stations at Twentynine Palms, Amboy, Needles, Mitchell Caverns, Mountain Pass, Kelso, and Yucca Grove. For areas of the basin where data were not available, rain intensity isohyetal maps provided in the San Bernardino County Hydrology Manual were used instead. The average precipitation for each subarea was calculated from the isohyetal map as a weighted average, and long term daily precipitation records were created for each subarea based on precipitation records from both the Amboy and Mitchell Caverns weather stations.

The Fenner watershed model was developed based on several parameters which define the hydrogeology of the Fenner watershed. The next step in developing the model was to assign values to the model parameters of each subarea. Values for each parameter were determined as follows:

- **Curve Number** - Based on soil type, vegetation type and density, and antecedent moisture conditions (as specified in the San Bernardino County Hydrology Manual, p. C-6, Fig. C-3);

- **Soil Thickness, Field Capacity, and Apparent Specific Gravity** - Determined from field observation and comparison with published data from similar areas;

- **Daily Evapotranspiration Rate** - Calculated using Thornthwaite’s formula (Hantush, 1959);

- **Vegetation Density** - Determined from field observation and examination of aerial photographs;

- **Vegetation Interception** - Determined from field observation; and
Initial Moisture Content of the Soil - Assumed.

The watershed model determined a range of recoverable water estimates which fall within or overlap other estimates of recharge for the Fenner Basin.

Recharge Estimates Based on Flownet analyses and a Groundwater Flow Model

Alternative methods used to estimate the flow through the Fenner Gap include a groundwater flownet analysis and a multiple layered groundwater flow model (MODFLOW). Flownet methods require the assumption of isotropic and homogenous aquifer materials within various flownet cells and estimate underflow based on Darcian flow calculations.

The Cadiz groundwater flow model covers approximately 945 square miles with a three layer variable-grid network consisting of 120 nodes in the north-to-south direction (I-direction), 120 nodes in the west-to-east direction (J-direction), for a total of 43,200 nodes. The smallest node represents an area 500 ft (north-south) by 500 ft (east-west). Nodes near the edges of the model are of variable size, ranging up to 5,000 ft by 5,000 ft. The Layer 1 and Layer 2 model area is bounded by a head-dependent or open (active) boundary with the exception of the Clipper, Ship, Calumet, Bristol, and Marble mountains. These mountainous areas represent impermeable boundaries and as such were assigned as no flow or inactive cells in the model. Layer 3 consists of the Paleozoic dolomite rocks and is only active from Fenner Gap area to approximately 5 miles northeast of the Fenner Gap.

The groundwater flow model, by design, can more accurately treat the heterogeneous nature of the aquifer system. The heterogeneity of the system is well documented in the pumping test results, well logs and geophysical logs of the area. However, in the course of developing a model, it is generally accepted practice to rely on Darcian or flownet estimates to establish a “first look” and recognize the limitations of these estimates due to the assumptions that are associated with them (e.g. isotropic and homogenous aquifer materials).

In summary, independent calculations using both flow net analyses and a distributed parameter groundwater flow model resulted in a range of recharge estimates for the Fenner basin consistent with other independent recharge calculations.

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Figure 1 - Watersheds of the Eastern Mojave
Relationship between geomorphic settings and unsaturated flow in an arid setting

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Abstract. Because geomorphology can readily be mapped, our ability to characterize unsaturated flow over large areas would be greatly enhanced if relationships between geomorphic settings and unsaturated flow could be identified. The purpose of this study was to evaluate relationships between geomorphic settings and spatial and temporal variability in unsaturated flow at a field site in the Chihuahuan Desert of Texas. This study differs from most previous studies in the variety of geomorphic settings studied, including drainage areas (Blanca Draw and Grayton Lake playa) and interdrainage areas (basin-fill deposits, eolian sheets, alluvial fans, and a fissure), density of data (~50 sampled boreholes 3–31 m deep), and variety of techniques (physical, chemical, and electromagnetic) used to quantify unsaturated flow. Spatial variability in unsaturated flow parameters is related to geomorphic settings. The various geomorphic settings form distinct groups on a plot of chloride versus water potential. Interdrainage areas have low water potentials and high chloride concentrations, indicating low water fluxes. Mean water fluxes estimated from chloride data ranged from 0.02 to 0.05 mm yr⁻¹. In contrast, localized topographic depressions (fissure, gully, and borrow pit) have high water potentials and low chloride concentrations which indicate high mean water flux (~100 mm yr⁻¹). These topographic depressions occupy <1% of the basin area. Drainage areas have low water potentials, which indicate low water fluxes and low to moderate chloride concentrations, which indicate higher water fluxes in the past (≤40 mm yr⁻¹). Short-term variability in response to precipitation events was only monitored in topographic depressions. The age of the pore water in the interdrainage areas spanned paleoclimatic variations (≤136,000 years at 25 m depth); however, most interdrainage profiles show negligible response to past climate fluctuations. Some profiles in eolian sheets showed increased water flux in response to Pleistocene climate change. The findings from this study indicate that geomorphology can provide valuable information on unsaturated flow and underscore the importance of localized topographic depressions for focusing unsaturated flow.

1. Introduction

Variations in geomorphology may include differences in topography, soil texture, and vegetation. Previous studies have shown that surface water collects in topographic depressions and results in focused flow through the unsaturated zone. Water fluxes are much higher beneath sinkholes (≥60 mm yr⁻¹) than in surrounding vegetated topographic settings (0.06–1.7 mm yr⁻¹) in South Australia [Allison et al., 1985]. Ephemeral lakes or playa lakes in the Southern High Plains of Texas and New Mexico also focus recharge [Stone, 1990; Wood and Sanford, 1995; Scanlon and Goldsmith, 1997]. Fissured sediments in the Chihuahuan Desert of Texas concentrate surface runoff, and water fluxes beneath these fissures are up to 3 orders of magnitude higher than in surrounding areas [Scanlon, 1992; Scanlon et al., 1997b]. Texture of surficial sediments also plays an important role in controlling unsaturated flow. Fine-grained surficial sediments provide a storage capacity large enough to retain infiltrated water near the surface, where it is available for evapotranspiration. Cook et al. [1992] noted an apparent negative correlation between clay content in the upper 2 m and the recharge rate. High water fluxes (23 mm yr⁻¹) have been inferred on the basis of tritium data in coarse sand dune sediments in Saudi Arabia [Dincer et al., 1974]. Vegetation also affects unsaturated flow in desert systems. Lysimeter studies in Hanford, Washington, and Las Cruces, New Mexico, show deep drainage ranging from 10% to >50% of the annual precipitation in bare, sandy soils [Ge et al., 1994]. The presence of plants in other areas of the Hanford site reduced deep drainage by up to 2 orders of magnitude [Pych, 1995; Murphy et al., 1996]. Studies in Cyprus show that recharge rates were highest in areas of sparse vegetation and were lowest in areas of bush vegetation [Edmunds et al., 1988].
The aforementioned studies indicate that basic variables among geomorphic settings, such as topography, sediment texture, and vegetation, greatly affect unsaturated flow. Identifying relationships between geomorphic settings and unsaturated flow would increase our ability to characterize unsaturated flow over large areas because geomorphology can be readily mapped. Allison et al. [1994] found that recharge in some areas was quite variable locally on a 100 m scale and noted that use of geomorphic techniques for estimating recharge over large areas is still a "seminal problem" in recharge evaluation.

What techniques can be used to evaluate unsaturated flow within and among various geomorphic settings? There has been much debate about the reliability of various techniques as indicators of unsaturated flow. Traditional techniques can generally be subdivided into direct physical methods such as lysimeters, indirect physical methods such as water balance and Darcy's law approaches, and environmental tracers such as chloride and tritium. Although reviews of techniques in unsaturated zone hydrology suggest that physical methods are less accurate than environmental tracers in estimating water flux [Gee et al., 1994], the two approaches, physical methods and environmental tracers, provide information on unsaturated flow over different timescales. Physical measurements allow evaluation of dynamic processes under current climatic conditions, whereas environmental tracers generally provide information on net water fluxes over longer time periods (up to thousands of years for chloride). A comprehensive understanding of unsaturated flow requires both physical and environmental tracer data.

Noninvasive techniques, such as electromagnetic (EM) induction and ground-penetrating radar, are becoming increasingly popular for evaluating unsaturated flow because they can be used to evaluate unsaturated flow rapidly over large areas and because they provide an evaluation of conditions between point measurements from boreholes. In an Australian study the correlation coefficient between apparent electrical conductivity and recharge estimated according to unsaturated zone chloride data was 0.7 [Cook et al., 1992]. These data suggest that although EM induction cannot estimate recharge directly, it may be useful in reconnaissance and interpolation between borehole measurements.

The objectives of this study were to evaluate relationships between geomorphic settings and the spatial and temporal variability in unsaturated flow at a field site in an arid region and to determine fundamental controls on unsaturated flow. This work differs from previous field studies in arid settings in the variety of geomorphic settings evaluated, size of the study area examined (60 km²), the density of data (~50 boreholes 3–31 m deep in different geomorphic settings), and the variety of techniques (physical, chemical, and isotopic).

2. Site Description

The study area (~60 km²; 31°17'N, 105°16'W), ~120 km southeast of El Paso, lies in northwest Eagle Flat basin, in the Chihuahuan Desert of Texas (Figure 1). Northwest Eagle Flat basin, a closed topographic depression ~500 km² in area, is a sediment-filled basin within the Basin and Range physiographic province [Gile et al., 1981]. The unsaturated zone is extremely thick, as indicated by the depth of the regional potentiometric surface (198–230 m).

The regional climate is subtropical arid [Larkin and Bomar, 1983]. Long-term meteorological data were obtained at Sierra Blanca, on the west edge of the study area. Mean annual precipitation is 320 mm for a 25 year record. Precipitation during sample collection and monitoring in this study ranged from 179 mm in 1994 to 353 mm in 1993 (Figure 2). Most precipitation falls in local, intense, short-duration convective storms during the summer, when temperature and potential evaporation are highest. Minor winter frontal storms are of longer duration.

The geomorphic evolution of the landscape was described by Scanlon et al. [1999]. The study area has been subdivided into interdrainage and drainage areas. The interdrainage area consists of fine-grained basin-fill deposits and eolian sheets surrounded by a narrow rim of alluvial fans at the margin of the basin. An earth fissure is also found in the interdrainage area. The drainage area includes Blanca Draw and Grayton Lake.

The floor of Eagle Flat basin consists mostly of muds overlain by the Arispe Surface, which has well-developed soils. The basin-fill deposits are overbank deposits from the braided streams and from the toes of the alluvial fans. These deposits are stable, vegetated landforms that do not exhibit channels or erosional or depositional features resulting from fluvial or alluvial activity. Three calcic soil horizons are found at depths of 0 to 1, 3, and 6 m, which suggest extremely stable conditions [Jackson et al., 1993]. The vegetation consists of scattered short bunch grasses and small Tobosa (Hilaria mutica) grass flats. Shrubs are much rarer, and the overall vegetation is less dense than that of the adjacent eolian sheets.

The eolian sheets are characterized by irregular microtopography consisting of 0.1–0.2 m hummocks and swales superimposed onto larger-scale eolian bedforms. Black gramma (Bouteloua eriopoda) and other grasses are dense and common, along with soaptree yucca (Yucca elata) and mesquite (Prosopis glandulosa).

An earth fissure that was found ~0.3 km west of Blanca Draw in an interdrainage setting was described by Jackson et al. [1993]. The surface expression of the fissure is ~640 m long. On aerial photographs as far back as 1957 it can be distinguished by a linear vegetation feature (mesquite bushes). The fissure consists of an alignment of horizontal pipes or collapse pits up to ~1 m in diameter underlain by a tension fracture that is filled with sediment. The width of the fracture is ~0.2 m at 2–6 m depth. The maximum vertical extent of the fracture is unknown because we did not dig trenches deep enough to see the base of the fracture. In addition to natural geomorphic settings, some areas have been subjected to anthropogenic influence. Borrow pits next to the highway were excavated in the 1960s for road construction (Figure 1).

Blanca Draw is the axial drainage system for Eagle Flat basin and is flanked by moderate slopes formed when the drainage incised into the surrounding basin-fill deposits (Figure 1). The slopes, ~250 m in width, have an average gradient of 0.013. Locally, the slopes are dynamically eroding. The floor of Blanca Draw is fairly stable and has no active channel with mobile sediment. A discontinuous gully lies in one section of Blanca Draw. Although Blanca Draw is generally dry, water may pond for long periods in the gully after heavy rains. The dominant vegetation in the area of Blanca Draw that was studied is scattered dense mesquite trees (~2 m²) interspersed with grasses and other shrubs (Figure 3). Thickly vegetated grass flats containing dense patches of Tobosa grass form isolated patches both in the slope area next to Blanca Draw and near Grayton Lake, where the wash floor is wide and the gradient low.
Blanca Draw drains into Grayton Lake (20 km²), an ephemeral playa that was flooded between May 1992 and October 1993 (Figure 1). When not flooded, it is sparsely vegetated with herbs. The floor of the playa consists of clay containing mud cracks resulting from shrinking and swelling of the sediment.

3. Methods

3.1. Theory

Three basic approaches were used to evaluate unsaturated flow in the study area: electromagnetic induction, physical measurements, and chemical measurements. Electromagnetic induction was used to obtain information on large-scale spatial variability in unsaturated zone characteristics and to interpolate and extrapolate data from point estimates provided by boreholes. Physical data provided information on current flow processes, whereas chemical tracers provided information on net water fluxes on timescales from years to thousands of years.

Because borehole data provide only point estimates of hydraulic and hydrochemical parameters, it is important to evaluate variability between boreholes and between geomorphic settings using noninvasive techniques such as EM induction. The apparent electrical conductivity ($EC_a$) of the subsurface is directly proportional to the conductivity of the pore water, to the water content, and to the solid phase conductance [Rhoades et al., 1976, 1989]. The solid-phase conductance is determined primarily by the clay content and the cation exchange capacity of these clays. To evaluate factors affecting...
EC_a measured on the land surface with the EM31 meter, average values of water, chloride, and clay content (Table 1) were calculated by weighting the depth distribution of these parameters (represented by \( x \) in (1)) according to the sensitivity of the EM31 meter in the vertical dipole mode (subscript \( v \), restricted to 6 m depth) [McNeill, 1980]:

\[
x_v(z) = \frac{4z}{(4z^2 + 1)^{\frac{3}{2}}}
\]  

(1)

Although the physical approach may not provide accurate estimates of water flux, it provides invaluable information on current unsaturated flow processes. Variations in water content cannot be used to assess the direction of water movement because water content is discontinuous across sediment types. Monitoring water content is generally useful in evaluating movement of water pulses in areas of moderate to high water flux. In contrast to water content, energy potential is continuous across the interfaces between different sediment types under steady flow conditions and is typically used to infer flow direction. Evaluation of the flow direction requires information on the head gradient. Thermocouple psychrometers are required to measure the low water potentials (sum of matric and osmotic potentials), generally \(< -1 \) MPa, characteristic of arid settings. In addition to using the gradient, we can also use the position of the measured matric potential relative to a static equilibrium matric potential to assess the flow direction under steady state conditions [Bear, 1972. Figure 9.4.16. p. 506].

Chloride concentrations in pore water have been used to evaluate water fluxes in semiarid systems over timescales up to thousands of years [Allison et al., 1985]. Chloride from rain and dry fallout concentrates in the root zone as a result of evapotranspiration because the volatility of chloride is negligible and chloride uptake by plants is minimal [Gardner, 1967]. Volumetric water flux below the root zone (\( q_v, \) L T\(^{-1}\)) can be estimated on the basis of the degree of enrichment of chloride in pore water relative to its concentration in precipitation according to the chloride mass balance equation

\[
q_v = J_{Cl} c_{Cl,uz} = P c_{Cl,P} c_{Cl,uz}
\]  

(2)

where \( J_{Cl} \) is the chloride mass flux or chloride deposition flux at the surface (M L\(^{-2}\) T\(^{-1}\)), \( c_{Cl,uz} \) is the pore water chloride concentration (M L\(^{-3}\)), \( P \) is precipitation (assumed to include dry fallout, L T\(^{-1}\)), and \( c_{Cl,P} \) is the chloride concentration in precipitation and dry fallout (M L\(^{-3}\)). This equation ignores hydrodynamic dispersion and assumes that chloride moves by piston flow. The age or residence time \( t \) represented by chloride at depth \( z \) can be evaluated by dividing the cumulative total mass of chloride from the surface to that depth by the annual chloride deposition flux

\[
t = \int_0^z \theta c_{Cl,uz} dz / J_{Cl} = \int_0^z \rho_b M_{Cl,uz} dz / J_{Cl}
\]  

(3)

where \( \theta \) is the volumetric water content (L\(^3\) L\(^{-3}\)), \( \rho_b \) is the dry bulk density (M L\(^{-3}\)), and \( M \) is the chloride concentration.

Figure 3. Aerial photograph of Blanca Draw and surrounding interdrainage area. Plots labeled 1 and 2 represent 15 m\(^2\) areas where the number of mesquite trees was counted and ranged from 26 trees in 1 to 19 trees in 2. An electromagnetic (EM) transect is located from EF 94 in Blanca Draw gully to EF 28 and EF 111 in the interdrainage basin-fill deposits.
(M Cl M⁻¹ soil). The main assumptions of the chloride mass balance method are (1) one-dimensional, vertically downward piston flow and (2) constant chloride deposition flux for the period being considered. If other sources, such as leaching from the mineral phase or runon, contribute chloride to the system, such sources should be included. Runon represents surface water input from an area. The simple chloride mass balance equation was modified by Wood and Sanford [1995] to include the effects of runon into playas:

\[ q_a = P_{CIP} - \frac{C_{runon}}{C_{Cl_{sat}}} + R_0 \left( \frac{A_{runon}}{A_{f}} \right) \]  

(4)

where \( C_{runon} \) (M L⁻¹) is the chloride concentration in runon, \( R_0 \) is runon (L), \( A_s \) (L²) is the area of the basin, and \( A_f \) is the area of the playa floor.

Chloride mass deposition flux can be estimated from (1) chloride concentrations in precipitation and dry fallout times the mean annual precipitation or (2) the estimated atmospheric fallout of \(^{36}\)Cl at the latitude of the site (23 atoms m⁻² s⁻¹ [Bentley et al., 1986]) divided by the measured prebomb \(^{36}\)Cl/Cl ratio in the unsaturated zone (samples from 3.1 to 10.7 m in borehole EF 60 (490 \times 10⁻¹⁵ [Scanlon et al., 1998]). Method 1 resulted in a precipitation weighted average chloride concentration of 0.14 g m⁻³ (approximate monthly measurements of chloride for 2 years), whereas method 2 resulted in a chloride deposition flux of 87 mg m⁻² yr⁻¹, which is equivalent to a chloride concentration in precipitation and dry fallout of 0.27 g m⁻³ calculated on the basis of a long-term mean annual precipitation of 320 mm. Method 2 is considered more valid for this study because it provides an estimate of long-term chloride mass deposition flux [Scanlon et al., 1998]. If the variations in \(^{36}\)Cl fallout with precipitation are taken into account as described by Scanlon et al. [1997], the resultant chloride deposition flux is 102 mg m⁻² yr⁻¹. This value implies an average chloride concentration in precipitation of 0.32 g m⁻³. The range of deposition fluxes (87-102 mg m⁻² yr⁻¹) gives some indication of the uncertainties in this parameter.

Tritium (half-life 1243 years) is produced naturally by cosmic ray neutrons in the upper atmosphere, and tritium concentrations in precipitation range from 5 to 10 tritium units (TU). Tritium concentrations increased from 10 to \( \geq \)2000 TU during atmospheric nuclear testing [International Atomic Energy Agency (IAEA), 1983] that was initiated in 1952 and peaked in 1963-1964. The subsurface distribution of bomb-pulse tritium can be used to estimate water fluxes and to evaluate preferential flow.

3.2. Field and Laboratory Methods

Detailed descriptions of methods are given by Scanlon [1994] and Scanlon et al. [1998]. The EM31 ground conductivity meter (Geonics Limited, Mississauga, Ontario, Canada) was used to measure \( E_{DC} \) of the subsurface [McNeill, 1992]. The intercoil spacing in the EM31 is 3.7 m. The exploration depth is \( \approx 3 \) m when the instrument is operated in the horizontal dipole mode (both coils vertical on the ground) and \( \approx 6 \) m when the instrument is operated in the vertical dipole mode (both coils horizontal on the ground). Electromagnetic induction transects were conducted perpendicular to Blanca Draw (EF 94 to EF 111; Figure 3), Grayton Lake, and the fissure (Figure 1).

Boreholes were drilled with a hollow-stem auger without any drilling fluid, and samples were collected with a split spoon. Particle-size analyses were conducted on sediment samples from 37 boreholes using sieving and hydrometer analyses [Gee and Bauder, 1986] (Table A1'). Sediment samples were collected from 30 boreholes for laboratory determination of gravimetric water content by oven drying the samples for 24 hours [Gardner, 1986] and for chloride content (Figure 1 and Table A1). Many samples were collected from the same boreholes as those that had been sampled for texture. To determine chloride content, double-deionized water was added to the dried sediment sample in a 3:1 ratio. Both ion chromatography (detection limit 0.1 g m⁻³) and potentiometric titration (detection limit 2 g m⁻³) were used to analyze chloride concentrations in the supernatant. Chloride concentrations are expressed as g Cl⁻ m⁻³ pure water (equivalent to milligrams Cl⁻ per liter of water). Water content was monitored by a neutron moisture probe (model 503 DR, CPN Corporation, Martinez, California) in nine neutron probe access tubes (Figure 1). Monitoring depths ranged from 0.3 m to maximum depths of 2.2-8.6 m in different access tubes.

Sediment samples were collected from 47 boreholes to a maximum depth of 31 m for water-potential measurements in the laboratory (Figure 1 and Table A1). Water potential was measured with a water activity meter (model CX-2) and a thermostate psychrometer sample changer (model SC-10A), both manufactured by Decagon Devices, Inc., Pullman, Washington. The water activity (\( A_w \)) was converted to water potential (\( \psi_w \)) using the Kelvin equation [Gee et al., 1992]. Because water potentials measured with the two instruments were similar [Scanlon et al., 1999], the results section presents only water potentials measured with the Decagon SC10A thermocouple psychrometer. Field psychrometers consisted of screened-caged, thermocouple psychrometers (model 74, PST 66, J.R.D. Merrill Specialty Equipment, Logan, Utah) and were installed at different times beginning in April 1993. Water potentials and temperatures were logged daily at depths of 0.3-19.3 m at 0900 LT. Good agreement was found between duplicate psychrometers installed at 0.3, 0.4, 0.5, 7.6, and 19.3 m depth. At all other depths, one or both of the duplicate psychrometers stopped functioning.

Table 1. Mean Water Content, Chloride Content, and Clay Content Calculated From Sediment Samples in the Upper 6 m From Boreholes According to Equation (8) and Compared With \( E_{DC} \) Measured at the Surface

<table>
<thead>
<tr>
<th>Borehole</th>
<th>Geomorphic Setting</th>
<th>Water Content Mean, g g⁻¹</th>
<th>Chloride Mean, mg Cl⁻ kg⁻¹ sediment</th>
<th>Clay Content Mean, %</th>
<th>( E_{DC} ) (VD) Mean, mS m⁻¹</th>
</tr>
</thead>
<tbody>
<tr>
<td>EF 94</td>
<td>gully</td>
<td>0.17</td>
<td>14</td>
<td>50</td>
<td>135</td>
</tr>
<tr>
<td>EF 110</td>
<td>Blanca Draw</td>
<td>0.13</td>
<td>5</td>
<td>45</td>
<td>30</td>
</tr>
<tr>
<td>EF 93</td>
<td>slope</td>
<td>0.09</td>
<td>602</td>
<td>45</td>
<td>110</td>
</tr>
<tr>
<td>EF 28</td>
<td>interdrainage</td>
<td>0.07</td>
<td>184</td>
<td>52</td>
<td>50</td>
</tr>
<tr>
<td>EF 111</td>
<td>interdrainage</td>
<td>0.12</td>
<td>380</td>
<td>45</td>
<td>50</td>
</tr>
<tr>
<td>GL 2</td>
<td>playa</td>
<td>0.15</td>
<td>59</td>
<td>61</td>
<td>66</td>
</tr>
<tr>
<td>GL 4</td>
<td>solon sheet</td>
<td>0.05</td>
<td>89</td>
<td>21</td>
<td>30</td>
</tr>
<tr>
<td>EF 35</td>
<td>fissure</td>
<td>0.12</td>
<td>10</td>
<td>24</td>
<td>75</td>
</tr>
<tr>
<td>EF 36</td>
<td>10 m from fissure</td>
<td>0.08</td>
<td>511</td>
<td>39</td>
<td>61</td>
</tr>
</tbody>
</table>

\( E_{DC} \), apparent electrical conductivity; VD, vertical dipole.

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A Table A1 is available on diskette or via anonymous FTP from kosmos.agu.org, directory APEND (Username = anonymous, Password = guest). Diskette may be ordered by mail from AGU, 2000 Florida Avenue, NW, Washington, DC 20008 or by phone at 800-966-2481; $15.00. Payment must accompany order.
Samples were collected for tritium analysis from boreholes EF 79 and EF 117 in the interdrainage eolian sheet, EF 92 beneath the fissure, EF 96 10 m from the fissure, and GL 2 in Grayton Lake. Water was extracted from core samples in the laboratory by toluene azeotropic distillation and purified using paraffin wax [Ingraham and Shadel, 1992]. The samples were enriched and analyzed using liquid scintillation methods at the University of Arizona Tritium Laboratory or using gas proportional counting at the University of Miami Tritium Laboratory.

4. Results and Discussion

4.1. Electromagnetic Induction

Electromagnetic induction was used to evaluate large-scale spatial variability in unsaturated zone characteristics among geomorphic settings. The surface EM transect from the gully in Blanca Draw (EF 94) to the interdrainage basin-fill deposits (EF 28 and EF 111) showed a wide range of EC (Figures 1, 3, and 4a). Because clay content is fairly uniform in profiles along the transect (45%-52%), differences in surface conductance associated with variations in clay content cannot be used to explain the variations in EC (Table 1). High values of EC were mapped in the gully, which correspond to high water contents (EF 94; mean 0.17 g g⁻¹) attributed to frequent ponding of water in the gully. The large decrease in EC from the gully (EF 94) to the surrounding draw (EF 110) (vertical dipole (VD): ~30 mS m⁻¹) is consistent with a 24% reduction in mean water content and with the threelfold decrease in chloride content. Values of EC increased markedly from the draw, where mesquite is found, to the slope (VD: 80-110 mS m⁻¹).
where grasses occur. This increase in EC_a correlates with a 120-fold increase in chloride content from the draw (EF 110) to the slope (EF 93). The EC_a remained high for ~100 m in the slope area and decreased gradually to lower values typical of the interdrainage basin-fill deposits (VD: 23–50 mS m⁻¹), which is consistent with the twofold or threefold decrease in chloride. Similar trends in EC_a were measured in four other transects at right angles to the draw, which showed low EC_a in the draw and sharp increases in EC_a values in the adjacent slope areas [Scanlon et al., 1999].

The floor of Grayton Lake playa was characterized by fairly uniform values of EC_a, which is consistent with the similarity in water and chloride contents in three profiles beneath the playa (Figures 5a, 5c, and 5g and Table 1). Higher values of EC_a in the playa relative to the adjacent eolian sheet area are attributed to the higher clay and water content beneath the playa. Chloride profiles are similar in both settings (Figures 5f and 5g).

In the vicinity of the fissure, EC_a was approximately 2 times higher than in the surface adjacent to the fissure (Figure 6a). The EM transect was conducted adjacent to profiles EF 35 and EF 36. Chloride is flushed out beneath the fissure, resulting in lower pore water conductivity. Low chloride content in the pore water is characteristic of fissured sediments [Scanlon et al., 1997b]. Although the 50-fold lower chloride content and 38% lower clay content beneath the fissure (EF 35) should result in lower EC_a, these two effects are offset by the 50% higher water content beneath the fissure than 10 m from the fissure (EF 36). The insensitivity of the EM response to the higher chloride content adjacent to the fissure suggests that the water content in this zone is below the critical level and that the EM31 is primarily measuring surface conductance.

When water content is below the critical level, electromagnetic induction is insensitive to variations in chloride content in the pore water, as seen in the data adjacent to the fissure. Above the critical water content, EC_a varies with all three parameters, water content, chloride content, and clay content. In most cases, although high water flux is associated with high water content and low chloride content which have opposite effects on EC_a, the data from the gully and the fissure indicate that water-content variations dominate the EM variations. Variation in chloride content is a secondary control and is seen in many of the profiles in the drainage-interdrainage transect (Figure 4).

4.2. Spatial Variability in Sediment Texture and Water Content

Sediments throughout the basin are generally fine grained. Mean clay contents in the drainage and interdrainage areas, excluding the eolian sheets, range from 38% to 56% (Tables A1 and 2). The finest sediments are found in Grayton Lake playa (mean clay content 56%). Clay minerals in the playa include smectite, illite, and kaolinite with moderate shrink/swell potential [Scanlon et al., 1998]. The only geomorphic setting with coarse-textured sediments is the young eolian sheet (mean sand content 56%). Local zones of coarse-grained sediments with up to 50% gravel are found beneath the eolian sheets and reflect paleochannels (E. C. Collins, personal communication, 1996). Sand contents in the older eolian sheet (mean 41%) are slightly lower than those in the young eolian sheet.

Spatial variability in water content is controlled primarily by differences in texture of the sediments. Mean water content is negatively correlated with percentage of sand ($R = -0.67$ and $n = 37$; Figure 7a) and is positively correlated with percentage of clay ($R = 0.69$ and $n = 37$; Figure 7b). All correlations are statistically significant at $\alpha = 0.05$. These data indicate that water content can be used to infer the texture of these sediments. The lowest water contents (mean 0.07 g g⁻¹; Tables A1 and 2), which correspond to the coarsest sediments, were found in the young eolian sheet, and the highest water contents (mean 0.15 g g⁻¹), which correspond to the highest clay contents, were found in Grayton Lake. Higher mean water contents in topographic depressions (fissure, gully, and borrow pit) relative to those in other geomorphic settings with similar textures are attributed to higher water fluxes in these depressions (Table 2 and Figure 7).

4.3. Spatial Variability in Water Potential and Chloride

Mean water potentials were plotted against mean chloride concentrations to evaluate variations in unsaturated flow in the different geomorphic settings (Figure 8). The various geomorphic settings form distinct groups on this plot. Significance test...
conducted on mean differences, given unknown and unequal variance, showed that the geomorphic settings are significantly different [Scanlon et al., 1998]. The water potential/chloride plot can basically be divided into four unequal sections: high water potential, low chloride (localized topographic depressions) (Figure 8a); high water potential, moderate to high chloride (Figure 8b); low water potential, low chloride (Blanca Draw) (Figure 8c); and low water potential, moderate to high chloride (Grayton Lake, interdrainage eolian sheet, and interdrainage other) (Figure 8d). Note that none of the profiles plot in Figure 8b because high water potentials imply high current water fluxes, which would flush out chloride. The high water potentials and low chloride concentrations in the localized topographic depressions indicate high water fluxes. The low

Table 2. Summary of Texture, Water Content, Water Potential, Chloride, Tritium, and Water Fluxes for Different Geomorphic Settings

<table>
<thead>
<tr>
<th>Geomorphic Setting</th>
<th>Sand Mean, %</th>
<th>Silt Mean, %</th>
<th>Clay Mean, %</th>
<th>Water Content Mean, g cm⁻³</th>
<th>Water Potential Mean, MPa</th>
<th>Chloride Mean, g m⁻³</th>
<th>Tritium Range, TU</th>
<th>Water Flux Mean, mm yr⁻¹</th>
<th>Water Flux Range, mm yr⁻¹</th>
</tr>
</thead>
<tbody>
<tr>
<td>Interdrainage (eolian sheet)</td>
<td></td>
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<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
</tr>
<tr>
<td>eolian sheet (old)</td>
<td>41</td>
<td>24</td>
<td>34</td>
<td>0.09</td>
<td>-5.0</td>
<td>3264</td>
<td>0.1-11.4</td>
<td>0.03 (2)</td>
<td>0.02-0.05 (2)</td>
</tr>
<tr>
<td>eolian sheet (young)</td>
<td>56</td>
<td>17</td>
<td>25</td>
<td>0.07</td>
<td>-4.9</td>
<td>1789</td>
<td>0.1-36</td>
<td>0.06 (2)</td>
<td>0.01-0.36 (2)</td>
</tr>
<tr>
<td>Interdrainage (other)</td>
<td></td>
<td></td>
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<tr>
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<td>23</td>
<td>20</td>
<td>38</td>
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<td>-7.6</td>
<td>5757</td>
<td>0.02 (1)</td>
<td>0.01-0.02 (1)</td>
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<td>basin-fill deposits</td>
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<td>29</td>
<td>48</td>
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<td>3187</td>
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<td>0.01-0.05 (1)</td>
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<tr>
<td>slope (Blanca Draw)</td>
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<td>26</td>
<td>46</td>
<td>0.11</td>
<td>-5.8</td>
<td>6230</td>
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<td>0.01-0.10 (1)</td>
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<td>-6.8</td>
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<td>6.2-15.2</td>
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<td>22</td>
<td>56</td>
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<td>-5.1</td>
<td>544</td>
<td>3.2-16.8</td>
<td>0.23 (1 er)</td>
<td>0.07-0.98 (1 er)</td>
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<td>3.8 (1 er)</td>
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<td>Localized topographic depressions</td>
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<td>349</td>
<td>1.5 (3 er)</td>
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<td>-0.3</td>
<td>10</td>
<td>13.4 (er)</td>
<td>2.1-32.9 (er)</td>
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Water fluxes are estimated from chloride data.

* Calculated mean using samples (1), below 1 m depth; (2), below 2 m depth; (3), upper 5 m; (4), upper 7.5 m (EF 120); and (5) upper 5.0 to 9.1 m; (er) excludes runon.
water potentials and high chloride concentrations characteristic of the interdrainage areas indicate low water fluxes. Data from the eolian sheet plot in the center and indicate higher water fluxes than those of other interdrainage settings. An apparent inconsistency between low water potentials in Blanca Draw, which indicate low water fluxes, and low chloride concentrations, which indicate high water fluxes, can be resolved by considering the different timescales represented by water potential and chloride data. Water potential data represent current water fluxes, which are low, whereas chloride data represent long-term net water fluxes and indicate that water fluxes were higher at some time in the past. Because water ponds in Grayton Lake and was ponded during the study for ~1 year, chloride concentrations were expected to be negligible; however, chloride concentrations were higher than expected. The higher mean chloride concentration in Grayton Lake (mean 544 g m$^{-3}$) relative to that in Blanca Draw (87 g m$^{-3}$) may be attributed to the higher clay content in Grayton Lake, which decreases water fluxes (Tables A1 and 2).

Vertical water potential profiles can be used to evaluate the direction of water movement. Typical laboratory-measured water potential profiles outside the localized zones of surface ponding had low values near the surface that increased with depth (Figures 4g–4i, 5d, and 5e). The upward decrease in water potentials indicates an upward driving force for water flow. In addition to estimation of the flow direction from the upward gradient in water potential, water potentials generally plotted to the left of the equilibrium line in most profiles, which also indicates upward flow under steady flow conditions. Mean water potentials in Blanca Draw (~3.6 MPa; Figure 4g and Tables A1 and 2) were higher than those in the profiles in the adjacent slopes (~6.0 MPa; Figure 4h) and water potentials plotted to the right of the equilibrium line at depths ≥6–15 m (Figure 4g), which suggests drainage at depth below the draw. Water potentials in Grayton Lake were uniformly low (mean ~3.8 to ~5.7 MPa) throughout the profiles (Figure 5e and Table A1).

High water potentials in the localized topographic depressions, such as the fissure (Figure 6c), the gully (Figure 4f), and the borrow pit, generally plotted to the right of the equilibrium line, which indicates drainage under steady flow conditions. High water potentials beneath the fissure were generally re-
Figure 9. Mean water fluxes calculated from chloride data for the various geomorphic settings. Data from depths >1 m were used for all settings except the interdrainage eolian sheets, where data from depths >2 m were used. Water fluxes in the drainage areas and in the topographic depressions were increased by an order of magnitude to account for chloride in runon.

restricted to the upper 6–10 m of the profiles. Below this zone, water potentials decreased sharply to values similar to those in profiles 10 m away from the fissure.

Vertical chloride profiles were also examined. In the interdrainage areas outside the eolian sheets, peak-chloride concentrations (4534–17,821 g m⁻³; Table A1) were found at 0.5–2.6 m depth, and chloride concentrations decreased gradually below the peak (Figures 41 and 4m). The highest chloride concentrations were found in the slope area adjacent to Blanca Draw (EF 93; Figure 41) and adjacent to the fissure (EF 36; Figure 6d). Surface water runon in these areas may have added chloride to these profiles in addition to that provided by precipitation. Chloride was flushed out (concentrations ~250 g m⁻³) in the upper 0.5–2 m depth in the eolian sheet because the sediments are coarse-grained. Peak chloride concentrations ranged from 1716 to 7831 g m⁻³ in the eolian sheet (Table A1). All three chloride profiles in Grayton Lake had bulge shapes and similar concentrations, with peak values from 1084 to 1315 g m⁻³ (Figure 5g). Vertical chloride profiles beneath Blanca Draw (mean 87 g m⁻³), including the gully (mean 349 g m⁻³), were generally uniformly low (Figures 4j and 4k).

The localized topographic depressions generally had uniformly low vertical chloride profiles. Low chloride concentrations were restricted to the upper 6–10 m beneath the fissure, whereas concentrations in the profiles 10 m away from the fissure were high (Figure 6d and Table A1). This variation in chloride concentration is similar to the variations in water potential and indicates higher water fluxes restricted to the upper 6–10 m zone. Chloride was leached throughout the profile in the borrow pit (mean 10 g m⁻³; EF 15; Tables A1 and 2).

4.4. Water Flux Estimates From Meteoric Chloride

Calculated water fluxes from chloride data (equation (2); CI deposition flux 87 g m⁻³ yr⁻¹) were uniformly low in the eolian sheets below 2 m depth (mean 0.05 mm yr⁻¹) and in the other interdrainage areas below 1 m depth (mean 0.02 mm yr⁻¹) (Tables A1 and 2 and Figure 9). Surface leaching of chloride to a depth of ~2 m corresponded to higher water fluxes in near-surface coarse eolian sediments. Water fluxes in the slope region adjacent to Blanca Draw and to the fissure may be underestimated because additional chloride provided by runon was not accounted for in the analysis. The higher water fluxes in the eolian sheets relative to the other interdrainage settings is also corroborated by the shorter residence times of pore water in the eolian sheet (mean ~30,000 years at 10 m depth) relative to the other interdrainage settings (mean ~83,000 years at 10 m depth).

It is difficult to estimate water fluxes from chloride profiles in the drainage areas because runon or chloride concentrations in runon were not quantified. Minimum mean water fluxes below the upper meter based on (2) that neglect runon ranged from 0.1 to 1.0 mm yr⁻¹ (mean 0.2 mm yr⁻¹) in Grayton Lake and from 0.2 to 20 mm yr⁻¹ (mean 4 mm yr⁻¹) in Blanca Draw (Table 2). The effect of runon on water flux was evaluated by using (4) with the following parameters: runon (estimated to be ~10% of the precipitation (32 mm yr⁻¹)), area of basin (northwest Eagle Flat) = 500 km², area of playa floor (Grayton Lake) = 20 km², and chloride concentration in runon (estimated to be ~5 times that in precipitation (1.4 g m⁻³)). The resultant water fluxes were about an order of magnitude greater than those based on (2) that neglected runon (Figure 9).

Water fluxes were also estimated for the topographic depressions. The EF 120 chloride profile beneath the fissure that was fairly completely flushed was used to estimate water flux (2). The mean chloride concentration for EF 120 in the upper 7.5 m was 21 g m⁻³, and the mean water flux was 8 mm yr⁻¹ (Table 2). Inclusion of runon effects should increase this estimate by about an order of magnitude to 80 mm yr⁻¹. Water flux can also be estimated from the depth of the chloride front by assuming, on the basis of aerial photographs, that the fissure has been active for at least 50 years. The approximate location of the chloride front was 9 m (EF 35); therefore the resultant water velocity was ~180 mm yr⁻¹. The resultant water flux was 36 mm yr⁻¹ (mean β: 0.2 m² m⁻³). Because water flow is episodic in desert systems, most of the flux may have taken place after a sequence of rainfall events. Minimum estimates of water flux for the borrow pit ranged from 2 to 33 mm yr⁻¹ (mean 13 mm yr⁻¹) (Tables A1 and 2). Low chloride concentrations in the upper 5 m beneath the gully in Blanca Draw indicate a mean water flux of 1.5 mm yr⁻¹. It is difficult to estimate the amount of runon into these topographic depressions; however, we estimate at least an order of magnitude increase in flux as a result of runon.

4.5. Spatial Variability in Tritium

High tritium concentrations (3.1–11.4 TU) in an interdrainage profile (EF 79) in the upper 5 m probably reflect the tail of the bomb pulse (Figure 10). Tritium levels below this zone were low in EF 79 and EF 117 (range 0.14 ± 0.24–1.71 ± 0.44 TU ±2 standard deviations). In some cases the tritium levels were ~2 standard deviations, indicating no tritium. The tritium level in a procedural blank (0.98 ± 0.5 TU; dead, tritiated water added to an oven-dried sample) was similar to the tritium levels found in EF 79 and EF 117 profiles at depth and indicates no tritium in these settings [Scanlon et al., 1999]. The absence of tritium is consistent with the high chloride concentrations and low water fluxes estimated for these sediments.

High tritium levels were found beneath and adjacent to the fissure and beneath the playa (Figure 10). Tritium penetrated much deeper beneath the fissure (~20 m) than expected from the depth of chloride leaching (~10 m; Figure 6d). The most
plausible explanation for the high tritium concentrations is preferential flow [Scanlon et al., 1997b].

4.6. Temporal Variability in Unsaturated Flow

The previous section focused on mean water fluxes; however, water fluxes may change over time. Temporal variability in unsaturated flow was evaluated over different timescales: short-term variability based on water content and water potential monitoring during the time of the study, longer-term variability based on comparison of water potential and chloride data, and long-term variability based on chloride data.

Water content monitoring with a neutron probe in the various geomorphic settings showed that penetration of water was restricted to the upper 0.6 m depth except in areas subject to ponding, such as the fissure (Figure 11c) and the gully in Blanca Draw (Figure 11a). Water penetrated to 1.2 m depth beneath the fissure after 131 mm of rain fell in July 1993 (38 mm in 1 day); however, the infiltrated water was removed from the subsurface by evapotranspiration in ~3 months. These data show how effectively the creosote bushes along the fissure remove infiltrated water. The zone of infiltration beneath the fissure was localized, as shown by the lack of variations in water content in a neutron probe access tube 10 m from the fissure (Figure 11d). High rainfall in September 1995 (116 mm) resulted in water penetration to 0.8 m in the gully (EF 100NP; Figure 11a). Ponding probably occurred after this wet period in the gully because 70 mm of rain fell in 1 day. Water content did not vary with time in Blanca Draw outside the gully (Figure 11b).

Water ponded ephemerally in Blanca Draw and Grayton Lake. Blanca Draw was flooded in May 1992, and water levels were up to 0.4 m above ground surface in some areas; however, the flooding lasted <1 week, and samples from boreholes drilled in October 1992 had low water contents (mean 0.09 g g⁻¹; EF 41 and 85; Figure 4e) similar to those in profiles in adjacent geomorphic settings (mean 0.11 g g⁻¹; EF 43; slope area; Figure 4d). Grayton Lake was also flooded in May 1992 and remained flooded until approximately October 1993. Mean water content in sediment samples from GL 2 (0.15 g g⁻¹) drilled in September 1994 was similar to mean water contents in GL 5 (0.16 g g⁻¹) and GL 6 (0.14 g g⁻¹), which were drilled in April 1996 (Figure 5c and Table A1). These data indicate little infiltration of water after flooding.

In situ water potentials were monitored in the interdrainage eolian sheet area by thermocouple psychrometers (Figure 12). Water potentials at 0.3–0.5 m depth were quite variable. The greatest increase in water potentials in the shallow zone oc-
curred after 70 mm of rain fell on September 15, 1995, and June 27, 1996. Water potentials at 0.3 m depth increased only to \(-0.5\) MPa 3 days after these rainfall events, indicating that the sediments did not become saturated at this depth. Sequential increases in water potential at 0.3, 0.4, and 0.5 m depth over time suggest uniform downward movement of water in this zone. Water potentials at 1.7 and 4.9 m depth showed small-scale seasonal fluctuations attributed to seasonal temperature fluctuations [Scanlon and Milly, 1994], whereas water potentials at 7.6 m depth were fairly uniform over time. Water potentials at depths \(>10\) m converged to \(-3\) MPa. Vertical profiles of water potential on May 28, 1996, showed that water potentials increased with depth and suggest an upward driving force for water movement, which is consistent with laboratory-measured data (Figure 13). Field-measured water potentials were slightly higher than laboratory-measured water potentials in a nearby borehole (EF 66), which is attributed to slight drying of sediment during sample collection and analysis.

Comparison of soil physics and environmental tracer data gives some indication of temporal variability in unsaturated flow. Low water potentials in Blanca Draw indicate current low water fluxes, whereas low chloride concentrations indicate higher water fluxes at some time in the past (Figures 4g and 4k). The time period represented by the current low water potentials is difficult to estimate. It depends partly on how efficiently mesquite trees can remove water from the subsurface. Vertical chloride profiles can also be used to estimate temporal variability in water flux. Beneath the fissure, water fluxes changed from low (\(-0.02\) mm yr\(^{-1}\); chloride \(-5000\) g m\(^{-3}\) at depth) to high (\(-80\) mm yr\(^{-1}\) including runoff; chloride 3-30 g m\(^{-3}\) near the surface; mean 13 g m\(^{-3}\)) (EF 120).

The long period represented by chloride profiles in the interdrainage settings spans paleoclimatic variations (Figures 14a, 14c, and 14e). Response to paleoclimate varied among these settings and was examined using plots of cumulative chloride versus cumulative water content (Figures 14b, 14d, and 14f). Profiles in the interdrainage basin-fill deposits and old eolian sheet did not respond to postulated increased precipitation during the Pleistocene, as shown by the uniform slope in the cumulative chloride versus cumulative water plot (Figure 14b). In contrast, there is a marked change in slope of the cumulative chloride versus cumulative water plots for some profiles in the young eolian sheet (EF 91, 101, and GL 4) that corresponds to a chloride mass balance age of 8900 (EF 101) to 16,400 years (GL 4) (Figure 14f). Uncertainty in these ages is at least 30%, based on uncertainties in the chloride deposition.
flux. Higher water fluxes before that time correspond approximately to the Pleistocene period. Water fluxes increased by up to a factor of 5 in GL 4 (0.15 mm yr\(^{-1}\) prior to 16,400 years to 0.03 mm yr\(^{-1}\) after 16,400 years). Some profiles in the young eolian sheet may not have extended deep enough to reflect the Pleistocene climate change (EF 95; Figure 14c). Other shallow profiles in the young eolian sheet, however, showed no response to the Pleistocene climate change (EF 70, 75, and 76; Figures 14c and 14d). It is difficult to determine what controls the system's response to increased precipitation in the Pleistocene. The sediment texture in the shallow zone in profiles EF 91, EF 101, and GL 4 is similar to those in other profiles in the young eolian sheet that did not show any response to Pleistocene climate change, suggesting that some factor other than texture is responsible. It is possible that microtopographic variations may have been important.

4.7. Relationship Between Geomorphic Setting and Unsaturated Flow

Results of this study show that geomorphology and unsaturated flow are highly related. Variability in physical and chemical parameters within geomorphic settings was generally much less than variability between geomorphic settings. The large number of boreholes within each geomorphic setting provided detailed information on spatial variability that was supported by data from noninvasive geophysical techniques between boreholes. Physical and chemical results were consistent.

Interdrainage settings, including alluvial fans, basin-fill deposits, and eolian sheets, are generally characterized by low water fluxes, as indicated by low water contents, low water potentials, and high chloride concentrations. Mean water fluxes estimated from the chloride data ranged from 0.05 mm yr\(^{-1}\) in the eolian sheets below the upper 2 m zone to 0.02 mm yr\(^{-1}\) in the other interdrainage areas below the upper 1 m zone (Figure 9 and Table 2). Some chloride profiles in the young eolian sheet indicate leaching to a 2 m depth associated with coarse-textured sands.

The drainage system includes Blanca Draw and Grayton Lake playa. Current water fluxes in Blanca Draw are low outside the gully, as shown by low water contents and low water potentials. Low chloride concentrations in these profiles indicate that chloride either never accumulated or was flushed out in the past. The mean water flux estimated from chloride data is 4 mm yr\(^{-1}\) (excluding runon; Tables A1 and 2). Including the effects of runon would increase the water fluxes by about an order of magnitude (Figure 9). Grayton Lake playa is characterized by very uniform profiles of low water content and low water potentials, indicating low water fluxes during the...
Figure 15. Relationship between mean water potentials and mean chloride concentrations calculated from profiles drilled in drainage and interdrainage areas in the Hueco Bolson [Scanlon, 1994], interdrainage alluvial fans at Beatty, Nevada, and Ward Valley, California [Prudic, 1994], and at the Nevada Test Site [Tyler et al., 1996]. Confidence intervals for mean values for different geomorphic settings in the Eagle Flat basin are included for comparison.

Geomorphology may have also influenced the response of the unsaturated zone to higher precipitation in the Pleistocene. This variation in response is suggested by some deep profiles (≤25 m depth) in selected areas of the young eolian sheet, which had low chloride concentrations at depth (EF 91, 101, 102, and GL 4) that are attributed to higher water fluxes during the Pleistocene. Most other profiles in interdrainage settings showed no variations in chloride concentration in response to Pleistocene climate.

4.8. Comparison With Data From Geomorphic Settings in Other Basins

Unsaturated zone hydrologic studies have been conducted in arid systems throughout the world. In many cases, however, the studies have been restricted to a particular geomorphic setting, and few studies have examined the relationship between different geomorphic settings and unsaturated flow. Detailed studies of unsaturated flow were previously conducted in drainage and interdrainage settings in the Hueco Bolson, immediately northwest of the Eagle Flat basin [Scanlon, 1992, 1994] and in alluvial fan settings in Ward Valley (California), Amargosa Valley (Beatty, Nevada) [Prudic, 1994], and the Nevada Test Site (Nevada) [Tyler et al., 1996]. Data from these settings are shown on a plot of water potential versus chloride (Figure 15). Data from the interdrainage settings are similar to those in the Eagle Flat basin and suggest low water fluxes under current climatic conditions. The plotted points represent data from the upper 10 m at the Beatty site and from the upper 30 m at the Nevada Test Site. Chloride concentrations at the Beatty site decreased to 50 g m⁻³ at depths ≥10 m, indicating an increase in water flux from 0.01 (1-10 m zone) to 2 mm yr⁻¹ (≥10 m depth). The higher water fluxes at depth were attributed to the Amargosa River being more active during the Pleistocene [Prudic, 1994]. Low chloride concentrations at depths of 30–60 m (mean 18 g m⁻³) at one of the Nevada Test Site profiles were attributed to its location at the confluence of alluvial fans, which affected the system's response to higher precipitation during the Pleistocene. Data from the Hueco Bolson drainage have about an order of magnitude higher mean chloride concentration (mean 880 g m⁻³) than those in the Blanca Draw (mean 87 g m⁻³). The higher chloride concentration and resultant lower water flux in the Hueco Bolson drainage are attributed to the lower topographic expression (maximum relief ~0.6 m) relative to Blanca Draw (maximum relief ~3 m) in the Eagle Flat basin. These data suggest that drainage size is important and that small drainages, such as those studied in the Hueco Bolson, may not be effective in creating high water fluxes through the unsaturated zone.

Although the data are limited, many basins show the effect of higher recharge associated with drainage areas, either currently or during Pleistocene times. Lower water fluxes are generally found in interdrainage settings. The similarities between profiles in coarse-grained alluvial fan sediments in Ward Valley, Amargosa Valley, and the Nevada Test Site and those in fine-grained sediments in Eagle Flat basin suggest that sediment texture is not the primary control on water flux in the unsaturated zone.

4.9. Comparison of Different Methods of Evaluating Unsaturated Flow

Results of this study show that multiple independent lines of evidence are required to obtain a comprehensive understanding of unsaturated flow processes. For example, use of chloride...
data alone in the drainage area (Blanca Draw) would suggest high water fluxes, whereas water-content and water-potential data indicate that current water fluxes are low.

Electromagnetic induction provides valuable information on interborehole variability. The transects are inexpensive and can be conducted quickly. Apparent electrical conductivity varies with water content, chloride, and texture. In arid settings, however, water content is commonly below the critical level, and electromagnetic induction simply maps the surface conductance or textural variability. EM induction is a good reconnaissance tool; however, borehole data are required to determine accurately what is controlling variations in EC, mapped with the EM instruments.

Data on physical parameters provide valuable information on current, unsaturated flow processes. Spatial variability of water content was of limited use in this study because of the high correlation between water content and sediment texture (Figure 7). Water-content monitoring was very useful in delineating zones of high water flux. Zones of high water flux were accurately delineated by water-potential data, such as beneath the fissure, the gulley, and the borrow pit. Water-potential measurements also provide information on the direction of water movement. The water-potential data suggest current generally upward water movement in interdrainage and drainage areas and downward water movement in localized topographic depressions. The significance of low water potential measurements with regard to unsaturated water movement depends on how long it takes to develop such profiles. In areas of dense vegetation such as mesquite with deep roots, water may readily be removed from the subsurface and low water potentials may develop quickly.

Information on long-term net water fluxes is provided by environmental tracers. Low chloride concentrations indicate high water fluxes beneath the fissure, the borrow pit, and the draw. The timing of these high water fluxes cannot be determined from chloride data alone because once flushed out of the sediment, chloride takes a long time to accumulate, and low chloride concentrations could be a relic of some past period of high water fluxes. The relic status of low chloride concentrations in profiles beneath the draw setting is suggested by the discrepancy between the chloride data, which indicate high water fluxes, and the low water-content and water-potential data, which suggest current, low water fluxes.

4.10. Controls on Unsaturated Flow

Previous studies suggest that topography, sediment texture, and vegetation can affect unsaturated flow. It is difficult to determine which of these parameters controls unsaturated flow because flow in the natural system reflects a delicate balance between the various parameters. The most important control on high current water fluxes in the study area is topographic depressions that pond water episemically, such as the fissure, the gulley, and the borrow pit. In some cases, topographic control is not as obvious. Higher water fluxes during the Pleistocene in some areas of the young eolian sheet are attributed to microtopographic variations. Texture is also important. For example, the effects of water ponding in Gray Lake on unsaturated flow are damped by swelled clays that minimize water flux. The vertical extent of chloride leaching beneath the fissure is attributed to natural capillary barriers that retain water.

Vegetative cover in the various geomorphic settings is distinctive. Plants are opportunistic and concentrate in areas where water is available. Mesquite, found along the fissures and in the draw, reflects zones of high water flux. Active roots of mesquite plants have been found to depths of at least 6 m in fractures beneath fissures in the Hueco Bolson [Scanlon, 1992]. The distribution of these plants therefore is a good indicator of zones of high water movement. The plants nevertheless do not appear to control the areas of high water flux but simply concentrate in those areas.

4.11. Conceptual Flow Model

Results of physical and environmental tracer studies can be used to develop a conceptual model of unsaturated flow in the Eagle Flat basin. Most precipitation in the study area falls as local, intense, short duration convective summer storms. After such intense rainfall events, because water runs off and collects in topographic depressions such as the fissure, the gulley, and the borrow pit, these topographic depressions spatially focus water. The depth of penetration of most water beneath the fissure is restricted to the upper 6–10 m and is controlled partly by layering of sediments and natural capillary barriers. Preferential flow occurs to greater depths, as indicated by tritium data. Surface water runoff in the areas adjacent to the fissure and the slope region adjacent to the draw results in chloride addition to profiles adjacent to the fissure and in the slopes. The low permeability of the sediments results in shallow penetration of this runoff water and subsequent evapotranspiration, which concentrates chloride near the surface. Ponding of water also occurs in the draw and in the playa. Initial ponding in the playa results in preferential flow of water in desiccation cracks, as shown by deep penetration of tritium. Long-term ponding in the playa is attributed to expansion of the clays, which greatly reduces permeability of the clays after they have been wetted. Evaporation of the ponded water and slow infiltration results. Interdrainage areas outside localized topographic depressions are generally characterized by low water fluxes. Most of these areas show negligible response to past climate change such as during the Pleistocene.

5. Conclusions

The relationship between geomorphic settings and unsaturated flow is clearly delineated in the Eagle Flat study area. The various geomorphic settings form distinct groups on a plot of water potential versus chloride. Interdrainage areas have low water potentials and high chloride concentrations, which indicate low water fluxes. Mean water fluxes range from 0.02 to 0.05 mm yr⁻¹. In contrast to interdrainage areas, localized topographic depressions (fissure, gulley, and borrow pit) have high water potentials and low chloride concentrations, which indicate higher water fluxes. The mean water flux in topographic depressions was estimated from chloride data to be ~10 mm yr⁻¹. Inclusion of runoff would increase this estimate by about an order of magnitude to 100 mm yr⁻¹. The depth of penetration of the high water fluxes was restricted to the upper 10 m zone beneath the fissure. The topographic depressions occupy <1% of the basin area. Drainage areas have low water potentials, which indicate low current water fluxes, and low to moderate chloride concentrations, which indicate higher water fluxes in the past. Minimum estimates of water fluxes estimated from chloride data range from 0.2 to 20 mm yr⁻¹ in Blanca Draw. Water fluxes beneath Gray Lake ranged from 0.1 to 1.0 mm yr⁻¹. Inclusion of runoff would increase these estimates of water flux by about an order of magnitude. Tritium
down to a 17 m depth beneath the playa is attributed to preferential flow along pathways that result from desiccation of the shrink-swell clays.

Geomorphic settings are also related to the degree of temporal variability in unsaturated flow. Pore water residence times in interdrainage areas were up to 136,000 years at 25 m depth, which spans paleoclimatic variations. Most interdrainage profiles showed negligible response to past climate fluctuations, with the exception of some profiles in eolian sheets, which showed increased water flux in response to Pleistocene climate change. Short-term variability in response to current precipitation events was restricted to topographic depressions. These findings indicate that geomorphology can provide valuable information on unsaturated flow, and they underscore the importance of localized topographic depressions for focusing unsaturated flow.

This study emphasizes the importance of using multiple techniques to investigate unsaturated flow because each technique has limitations. Electromagnetic induction allowed evaluation of interborehole variability. Spatial variability in water content was controlled primarily by textural variability and was not a very good indicator of water movement. Water-content monitoring clearly delineated the zones of high water flux. Water potential and chloride were good indicators of areas of low and high water flux. The variety of techniques used in this study provided multiple independent lines of data to explain flow processes comprehensively.

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San Pedro River Studies and Remote Sensing of Evapotranspiration

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ABSTRACT

A variety of studies have been carried out in the San Pedro Basin as part of the Semi-Arid Land-Surface-Atmosphere Program ("SALSA"). SALSA is a multi-agency, multi-national research effort that seeks to evaluate the consequences of natural and human-induced environmental change in semi-arid regions. The ultimate goal of SALSA is to advance scientific understanding of the semi-arid portion of the hydrosphere-biosphere interface in order to provide reliable information for environmental decision making. The SALSA Primary Science Objective is: To understand, model, and predict the consequences of natural and human-induced change on the basin-wide water balance and ecological complexity of semi-arid basins at event, seasonal, interannual, and decadal time scales. SALSA approaches this goal and objective through a program of long-term, integrated observations, process research, modeling, assessment, and information management that is sustained by cooperation among scientists and information users. In this presentation general program background information and the critical nature of semi-arid regions is presented. A brief description of the Upper San Pedro River Basin, the initial location for focused SALSA research follows. Several overarching research objectives under which much of the interdisciplinary research was conducted will be. Principle methods, primary research sites and data collection used by numerous investigations during 1997-1999 are then presented. Particular emphasis will be place on the use of in-situ and remote sensing methods to estimate evapotranspiration. The results from SALSA research will be described in more detail in a upcoming special issue of the Journal of Agricultural and Forest Meteorology (in press).
THE RECHARGE OF GROUNDWATER

Herman Bouwer

Introduction

The two major components of the world's water are salt water in oceans (97.2%) and fresh water in glaciers and ice caps (2.1%). This leaves only 0.7% as liquid fresh water. Of this, 0.62% occurs as groundwater (0.31% within a depth of 800 m and 0.31% between 800 and 4000 m depth) and 0.0091% as surface water in streams and lakes. Atmospheric water accounts for only about 0.001%, water in the biomass for 0.004%, and water in the mostly unsaturated zone above the groundwater (also called vadose zone) for 0.005% of the world's water (Feth, 1973; Bouwer, 1978). Thus, there is about 67 times more groundwater than fresh surface water in this world, and groundwater comprises more than 98% of the world's liquid fresh water resources. This indicates the global importance of groundwater as a water resource and the need to manage it wisely to avoid depletion and contamination. It also shows that underground formations are capable of storing vast amounts of water through natural and artificial recharge.

Natural recharge

The source of practically all groundwater is atmospheric precipitation that has infiltrated into the soil and moved down through the vadose zone to underlying aquifers. Basically, this natural recharge rate thus is the amount of precipitation minus surface runoff and minus evaporation from the soil and transpiration by vegetation. The latter two often are combined into one parameter called evapotranspiration or ET. Additional recharge occurs from losing streams or lakes. On a long-term basis and for relatively permeable soil without much surface runoff, this natural recharge was about 30% of precipitation for agricultural land and forests of deciduous trees (20% for pine trees) in the humid, temperate climate of The Netherlands where the mean precipitation is about 77 cm per year (Querner, 2000). These percentages increase with increasing precipitation, like the 70 cm annual recharge or 53% of the 132 cm annual precipitation achieved south of Auckland, New Zealand (Rosen et al., 1999). In East Java, recharge rates were estimated at 30 to 50% of precipitation in an area with 175 cm average annual rainfall (Nielsen and Widjaya, 1989). On Bali, a more conservative recharge percentage of 25% was used for a 175 cm average annual rainfall zone. Mediterranean-type climates may produce recharge rates of 10% to 20% of precipitation, while hot, dry climates may produce natural recharge rates of less than 1% of average annual precipitation (Bouwer, 1978 and 1989, and references therein; Tyler et al., 1996). Since precipitation in dry climates may be considerably less than 10 cm per year, natural recharge rates at 1% of precipitation would be less than 1 mm/yr. These small recharge rates are very sensitive to precipitation characteristics and, hence, also to climate change. The water reaching the aquifer via natural recharge normally is of good quality, except in areas with intensive agriculture where it can be contaminated with agricultural

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chemicals (nitrates and pesticide residues) and also salts if rainfall is insufficient and crops must be irrigated (Bouwer, 1978 and 1990). This nonpoint source pollution of groundwater is becoming a major threat to groundwater quality as increasing populations need more food, which can only be realized by increasing the agricultural land area and using more fertilizer and other chemicals, and irrigation to increase yields (Bouwer et al, 1999).

In regions with significant rainfall, groundwater levels may be higher than water levels in streams or lakes, so that groundwater moves into these surface waters to maintain base flow in streams and water levels in lakes. These streams are then perennial, gaining streams. In drier regions, groundwater levels may be may be below streams, which then are losing streams that contribute to the natural recharge of underlying aquifers. Losing streams may not flow year-round, but intermittently (ephemeral streams) in response to rainfall and surface runoff. Some streams or rivers are gaining at higher elevations with more rain and snow where they originate, and become losing as they flow to lower elevations where precipitation is less and groundwater is deeper.

In natural systems, groundwater recharge is in balance with discharges from the aquifer. These include discharge into gaining streams or lakes and uptake of groundwater by roots of trees and other vegetation. In dry climates, the groundwater may eventually reach low areas where it can form oases or salt flats as water evaporates from water-logged areas. Pumping groundwater from wells interferes with the natural balance between recharge and discharge of groundwater, which can be a factor in safe yield determinations. Thus, knowledge about natural recharge rates is important because it gives an idea of the safe yield of aquifers, which is the rate at which groundwater can be pumped or otherwise collected from an aquifer without depleting the groundwater or without producing other undesirable effects of falling groundwater levels. The latter include reduction of groundwater flow into gaining streams and resulting reduction in streamflow, loss of trees and other vegetation that depend on groundwater, decay of wooden piles that begin to rot when they protrude above falling groundwater levels and cause foundation problems, land subsidence, sea water intrusion, higher pumping costs, and water quality declines of well water if the deeper groundwater is of lower quality than the upper groundwater (Bouwer, 1978). Regions with relatively dry climates are the most vulnerable to groundwater depletion because groundwater often is the main or only water resource and natural recharge rates are very low because they are small percentages of small annual precipitations and very vulnerable to climate change. Many regions in the world experience declining groundwater levels and it is frightening to think what will happen when the wells go dry and other water resources are not available.

Groundwater resources are sustainable only when groundwater pumping does not exceed safe yield levels. Exceptions can be made for short term uses of groundwater in excess of safe yield to provide water in periods of below average rainfall or to help start a new settlement that with time will build an economy and tax base that can afford to develop other, more expensive water resources like artificial recharge, water reuse, desalting of brackish or ocean water, etc. Development of groundwater resources often is quite inexpensive and produces “cheap” water. As a result, the cost of the extracted water tends to be much less than its value which, unfortunately, fosters depletion of groundwater. However, sound economics dictates that the value of water also be considered. Water in the aquifer itself (in-situ groundwater) also has value, since it protects against subsidence, sea
water intrusion and other adverse effects of declining groundwater levels. Leaving water in the aquifer also provides a reserve of water for periods of droughts and for the future (U.S. Nat. Acad. of Sciences, 1997).

Incidental recharge

Another type of groundwater recharge is incidental recharge, which is not planned but occurs as a side effect to human activities. Incidental recharge takes place below septic tank leach fields, cess pits, unlined lagoons, constructed wetlands, surface impoundments, etc. In irrigated areas, more irrigation water must be applied than needed for evapotranspiration so that salts and other chemicals brought into the root zone with the irrigation water can be leached out of the root zone with drainage or deep percolation water to maintain a salt balance in the root zone that avoids salinization of the soil (Tanji, 1990). However, this deep percolation water has high concentrations of salts and possibly agricultural chemical residues (Bouwer, 1990) and it will continue to move downward to underlying groundwater. Also, as cities grow and need more water, more and more sewage effluent will be used for irrigation (Bouwer, 1994 and 2000). The underlying groundwater then will not only be contaminated with salts and the normal agricultural chemicals, but also with "sewage" chemicals like disinfection byproducts, pharmaceuticals like endocrine disruptors, and a whole spectrum of other synthetic organic chemicals (Bouwer et al., 1999), and possibly also viral and bacterial pathogens.

Where groundwater levels in irrigated areas become too high, agricultural drains must be installed to keep water tables well below root zones to prevent the soil from becoming too salty. The salty discharge from these drains must be managed to avoid undesirable environmental effects (Madramootoo et al., 1997). Where the groundwater is deeper, the deep percolation water will join the aquifer and cause deterioration of ground-water quality. In the absence of adequate drainage or groundwater pumping, the deep percolation water also will cause groundwater levels rise. This could damage sewers and other underground pipelines, flood basements, rise into landfills or cemeteries, damage vegetation, and produce other undesirable effects, including the eventual water logging and salinization of surface soils.

Incidental recharge also occurs when urbanization of areas in dry climates creates impermeable surfaces like roads and roof tops etc. that produce more surface runoff and concentrate rain water and infiltration in the remaining soil areas. Small rains that normally would not produce significant recharge can then produce significant infiltration and recharge of groundwater. Where cities discharge storm water into detention and infiltration basins or dry wells or dry streambeds, additional recharge is obtained.

Induced recharge

Induced recharge is achieved by increasing the pumping of groundwater from aquifers near streams to create losing streams or to lower groundwater levels near already losing streams so as to "pull" more water from the stream into the aquifer. This principle is used in bank filtration where wells are located near streams to pump stream water that has been filtered through the aquifer. Bank filtration is used where groundwater is preferred over surface water and the availability of
groundwater is limited, or where surface water is of impaired quality and needs to be filtered through the aquifer before it goes to the water treatment plant into the water supply system. Bank filtration can also protect against accidental pollution of surface water by relying on attenuation of the pollutants in the aquifer and/or by shutting off the wells until the pollution plume has passed.

Inducing more recharge from streams by lowering adjacent groundwater levels through pumping from wells is effective only where groundwater levels are not too deep and the seepage outflow in the aquifer is mainly in horizontal direction and controlled by the slope of the water table away from the stream. If the water table is already deep, the seepage flow is mainly downward and controlled by gravity, so that further lowering of the water table does not increase the seepage rate and, hence, does not enhance recharge of the aquifer (Bouwer, 1999). If the wetted perimeter of the stream is covered by fine-particle deposits and products from biological activity to form a clogging layer, this layer becomes the bottleneck for the seepage flow system and controls seepage rates. The underlying soil will then be unsaturated (Bouwer, 1982) and there is no direct hydraulic connection between the stream and the aquifer as long as the top of the capillary fringe above the water table is below the original stream bottom. For most systems, this will be the case if the groundwater table is at least about 1 m below the stream bottom. Thus, once the groundwater level is more than 1 m below the stream bottom, further lowering of groundwater levels will not increase seepage rates under these conditions (Bouwer, 1999).

Where groundwater levels are relatively high, for example, at depths of less than 1 m, the net recharge of groundwater can also be increased by lowering groundwater levels via underground drains or pumped wells to where they are below the root zone of the vegetation. This will prevent plant roots from drawing on groundwater to meet potential evapotranspiration rates during periods of insufficient rainfall. This reduces the evapotranspiration component of the water balance and, hence, increases the net recharge of groundwater. Studies in the Netherlands (Quermer, 2000) have shown that recharge rates for deeper groundwater levels (average depth more than 1.2 m) were considerably more than for shallower groundwater levels (average highest levels at depths of less than 0.4 m and average lowest levels at depths of 0.8 to 1.2 m). The respective recharge rates were 34 vs. 24 cm/yr for pasture, 34 vs. 22 cm/yr for cultivated land, 22 vs. 14 cm/yr for pine forest, and 33 vs. 26 cm/yr for deciduous trees, all for an average precipitation of 77 cm/yr. The main reason that deciduous trees gave more recharge than pine trees probably was greater interception of precipitation on the evergreen pine trees than on the deciduous trees, so that more water evaporated directly from the pine needles and never reached the ground below the pine trees.

Lowering groundwater levels near streams and in floodplains has also been suggested as a way to reduce uptake of groundwater by deep rooted riparian vegetation, like cottonwoods, willows, mesquites, salt cedars, and other "phreatophytes" that have deep roots that can directly tap the aquifer and allow the trees to grow and survive in dry climates (Bouwer, 1975). While the active rooting depth of phreatophytes typically is on the order of 10 m, depths of as much as 30 m have been found during construction of the Suez Canal (Renner, 1915; Bouwer 1978). Lysimeter studies with young salt cedars in the Gila River west of Phoenix, Arizona, showed that evapotranspiration decreased from 2.4 m/yr to about 1 m/yr by lowering the groundwater level from a depth of 1 m to 2.5 m (van Hylckama, 1975; Bouwer, 1978). Thus, lowering groundwater levels in flood plains not
only increases recharge by increasing seepage losses from the stream, but also by reducing evapotranspiration by phreatophytes.

Enhanced recharge

Enhanced recharge is additional recharge obtained by non-invasive, non-engineering processes such as cloud seeding to increase precipitation. This is already done to increase streamflow, but it can also be expected to increase infiltration in the watershed and recharge of underlying aquifers. Another technique is to decrease discharge of groundwater via evapotranspiration by managing the vegetation. For example, vegetation can be completely removed to create bare soil. This will be most effective for increasing the recharge where soils are sandy and flat. Finer soils and sloping surfaces can produce more surface runoff and soil erosion. Exposure of finer soils to rain can create surface crusts (Sumner and Stewart, 1992), which reduce infiltration and recharge of underlying groundwater. Usually, however, vegetation management consists of replacing deep-rooted vegetation like trees by shallow rooted vegetation like grasses or grain crops. This reduces evapotranspiration and, hence, increases infiltration and groundwater recharge, causing water tables to rise (George et al., 1999). Also, existing vegetation could be replaced by plants that intercept less precipitation, mainly by their leaves, so that less water evaporates directly back to the atmosphere and more reaches the ground. This could be achieved, for example, by replacing evergreen trees by deciduous trees, as discussed under “Induced Recharge.”

Artificial recharge

Artificial recharge is the use of engineered systems that alter the earth’s surface to create spreading, impounding, or other surface facilities to which water can be applied for infiltration and subsequent movement to underlying groundwater. Such surface infiltration systems can be constructed on land areas (off-channel systems) or in streambeds (in-channel systems). Where sufficiently permeable soils for surface infiltration systems are not available, artificial recharge can also be achieved with subsurface infiltration facilities like trenches or large-diameter wells in the vadose zone, or with deeper wells that inject water directly into the aquifer (Bouwer, 1999).

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