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Geochemical Imaging of Flow Near an Artificial Recharge Facility, Orange County, California

by Jordan F. Clark¹, G. Bryant Hudson², M. Lee Davisson³, Greg Woodside⁴, and Roy Herndon⁴

Abstract

Critical for the management of artificial recharge operations is detailed knowledge of ground water dynamics near spreading areas. Geochemical tracer techniques including stable isotopes of water, tritium/helium-3 (T^3He) dating, and deliberate gas tracer experiments are ideally suited for these investigations. These tracers were used to evaluate flow near an artificial recharge site in northern Orange County, California, where $\sim 2.5 \times 10^8$ m³ (200,000 acre-feet) of water are recharged annually. T^3He ages show that most of the relatively shallow ground water within 3 km of the recharge facilities have apparent ages < 2 years; further downgradient apparent ages increase, reaching > 20 years at ~ 6 km. Gas tracer experiments using sulfur hexafluoride and xenon isotopes were conducted from the Santa Ana River and two spreading basins. These tracers were followed in the ground water for more than two years, allowing subsurface flow patterns and flow times to be quantified. Results demonstrate that mean horizontal ground water velocities range from < 1 to > 4 km/year. The leading edges of the tracer patch moved at velocities about twice as fast as the center of mass. Leading edge velocities are important when considering the potential transport of microbes and other "time sensitive" contaminants and cannot be determined easily with other methods. T^3He apparent ages and tracer travel times agreed within the analytical uncertainty at 16 of 19 narrow screened monitoring wells. By combining these techniques, ground water flow was imaged with time scales on the order of weeks to decades.

Introduction

Artificially enhancing recharge rates of surface waters into ground water aquifers at specially designed facilities is an attractive option for increasing the long-term storage capacity of potable water in arid areas such as the western United States. After residing in the subsurface for a period of time, which can be as short as a few days, the recharged water is extracted from these aquifers via production and irrigation wells. Artificial recharge has been conducted from many different types of facilities including spreading ponds, natural river channels, and injection wells (Bouwer

2002). It has also been used to combat sea water intrusion in some coastal areas.

Artificial recharge presents a few challenges, however. First, there needs to be an adequate supply of water for the recharge operation. This may involve importing water into the basin, diverting nearby rivers, capturing storm runoff, or recycling waste water. Second, there is a potential to degrade the quality of the native ground water because contaminants from industry, agriculture, or municipal sources may enter the surface water supply that will be used in the recharge operation. The introduction of disinfection byproducts, infective microorganisms, and organic compounds with unknown health risks into ground water supplies is a significant concern, especially when a component of the water supply is recycled waste water (National Research Council 1998).

Projects that propose to recharge recycled water into potable-supply aquifers in California may be evaluated in terms of residence time of recycled water in the aquifer, the amount of dilution with native ground water, and the documentation of in situ water quality changes within the aquifer. Development of field methodologies to evaluate these criteria will assist with permitting these projects as

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well as validating numerical models of flow near recharge operations. Geochemical tracers are ideally suited for such studies (Muir and Coplen 1981; Ma and Spalding 1996; Gamlin et al. 2001).

Tracers that have been commonly used to investigate flow characteristics of relatively young ground water include chlorofluorocarbons (CFCs) and tritium/helium-3 ($T/{}^3\text{He}$) dating techniques (Schlosser et al. 1989; Busenberg and Plummer 1992; Cook and Solomon 1997). These methods use preexisting signatures to determine ground water ages (up to 50 years) that typically have uncertainties of ± 2 years. Tracer ages are defined as the length of time that the ground water has been isolated from the soil atmosphere; hence, these geochemical clocks typically start at the time of recharge. They have been used to estimate recharge rates (Solomon et al. 1993), validate numerical models of flow (Reilly et al. 1994; Szabo et al. 1996; Tompson et al. 1999), determine rates of water-rock interactions (Rademacher et al. 2001), and provide chronologies to assess temporal variability in water quality parameters (Böhlke and Denver 1995). Because of their relatively large analytical errors, CFCs and $T/{}^3\text{He}$ dating are best suited for studying ground water flow over periods of decades that typically have path lengths of kilometers. These tracers are unable to resolve ground water ages between 0 and 2 years, and can give inappropriate travel times if mixing between different-aged waters occurs as is typical in long-screened production wells.

A second geochemical approach, such as deliberate tracer experiments, is needed to evaluate flow and transport during a period of 0 to 2 years. With this approach, trace substances are added to ground water in a controlled manner. By monitoring tracer concentration at wells, initial and mean arrival times can be determined. The initial arrival time represents the fastest flowpaths between the well and where the tracer was introduced. This information cannot be determined with a high degree of certainty using either geochemical dating techniques or numerical flow models. The mean arrival time (i.e., the arrival of the center of mass) is representative of the bulk flow or linear velocity.

Here, we summarize results from a study initiated in 1995 that used geochemical tracer techniques to examine ground water flow and transport near the Orange County Water District (OCWD) recharge operation. The study consisted of two phases. First, the $T/{}^3\text{He}$ dating technique was employed to determine long-term transport and identify areas of rapid flow near the recharge facilities. Second, deliberate tracer experiments were conducted to evaluate short-term transport. The latter included gas tracer experiments using sulfur hexafluoride (SF_6) and xenon isotopes (${}^{124}\text{Xe}$, ${}^{129}\text{Xe}$, ${}^{136}\text{Xe}$) and experiments of "opportunity" using the stable isotopes of water. All data and experimental details can be found in Davisson et al. (2004).

Study Location

A 9 km section of the Santa Ana River (SAR) and a series of spreading basins, including Warner Basin, Anaheim Lake, and Kraemer Basin, located near Anaheim, California, are used by OCWD as the principal recharge area for the Orange County ground water basin (Figure 1). Sur-

face water has been artificially recharged at these facilities for more than 40 years (Mills 2002). Currently, OCWD recharges $\sim 2.5 \times 10^8 \text{ m}^3$ (200,000 acre-feet) per year at these locations. This constitutes $\sim 60\%$ of the average annual ground water production within the basin (Orange County Water District 1998). The off-river recharge facilities are filled by either diverting SAR water, or importing Colorado River or California State Water Project water that has been transported hundreds of kilometers in aqueduct systems. The SAR water is a mixture of seasonal upper watershed runoff, tertiary-treated waste water, and rising ground water (Mills 2002).

The artificially recharged surface water flows into an alluvial aquifer system consisting of sand, gravel, silt, and clay layers. The spatial extent of any given layer is poorly known (Orange County Water District 1998; Tompson et al. 1999). The ground water basin covers an area of $\sim 900 \text{ km}^2$ and has a maximum thickness exceeding 600 m. The recharge facilities lie near the northeastern edge of the basin, in the "Forebay" or intake area, where the vadose zone is the thickest and impervious clay/silt deposits are generally thin and discontinuous. The total aquifer thickness in the vicinity of the recharge facilities increases from 30 m near the SAR at the Imperial Dam to 500 m near Kraemer Basin. The vadose zone thickness ranges from $< 5 \text{ m}$ adjacent to the SAR and recharge basins (during recharge) to $\sim 40 \text{ m}$. In this area, the ground water levels are the highest and the piezometric surface indicates that ground water near the spreading areas flows predominately to the west/southwest (Figure 1).

Principle of Tracer Techniques

Tritium/ ${}^3\text{He}$ Ground Water Dating

A large quantity of tritium, a radioactive isotope of hydrogen that decays to ${}^3\text{He}$, was released into the atmosphere in the late 1950s and early 1960s as a result of above-ground nuclear bomb tests. Since that time, this tritium has

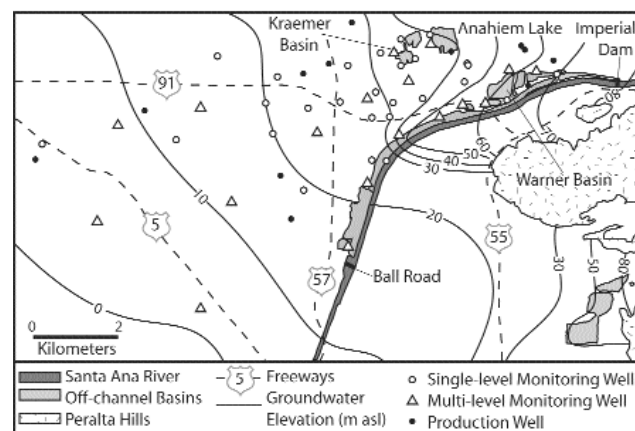


Figure 1. Map of the OCWD artificial recharge areas, wells sampled during this study, and ground water elevations (m asl). The ground water elevations were determined using only well data. The percolation zone within the SAR is between the Imperial Dam and Ball Road. The Peralta Hills are composed of nonwater producing formations and lie outside of the Orange County Water Basin.

entered the hydrologic cycle, significantly raising the tritium content of young ground water above natural levels. The T/³He method of dating is based on the radioactive decay law and the ground water age is determined with

$$t = \frac{t_{1/2}}{\ln(2)} \ln\left(1 + \frac{[{}^3\text{He}]_{\text{tri}}}{[T]}\right) \quad (1)$$

where t is time (apparent ground water age), $t_{1/2}$ is the half-life of tritium (12.43 years), $[T]$ is the tritium content, and $[{}^3\text{He}]_{\text{tri}}$ is the concentration of the ³He derived from the decay of tritium. While $[T]$ is measured directly, $[{}^3\text{He}]_{\text{tri}}$ is not; it is calculated from a helium mass balance that considers other helium sources (Schlosser et al. 1989; Cook and Solomon 1997).

Stable Isotopes of Water

Waters imported into ground water basins from great distances often can be distinguished from local water by their stable isotope compositions (i.e., ¹⁸O/¹⁶O and D/H ratios). This difference reflects the well known geographical variation in the isotope composition of precipitation. Typically, precipitation with lower ¹⁸O/¹⁶O and D/H ratios (i.e., isotopically light) falls inland, at cooler temperatures, and at higher elevations. Hence, imported water is often isotopically light and the migration of artificially recharged imported water can be easily traced in the subsurface (Muir and Coplen 1981; Ma and Spalding 1996). Williams (1997) demonstrated that the difference between local and imported water in Orange County is sufficient to use these tracers.

Deliberate Gas Tracers

Tracing the movement of artificially recharged water in the subsurface over a period longer than a year with deliberate tracers presents a number of challenges. First, a large volume of water (> 10⁶ m³) is typically needed to be tagged. Second, dispersion within the aquifer can significantly reduce the tracer's initial concentration. Third, to ensure that buoyancy does not influence the vertical flow, the density of the tagged water cannot be significantly different than the untagged water (Istok and Humphrey 1995). Hence, the tracer needs to be relatively inexpensive, have a very low background concentration, and behave conservatively in the aquifer. Gas tracers such as sulfur hexafluoride (SF₆) and noble gas isotopes (e.g., ⁴He, ²²Ne, ¹²⁴Xe, ¹²⁹Xe, and ¹³⁶Xe) are ideally suited for this application (Gupta et al. 1994; Wilson and Mackay 1996; Dillon et al. 1999; Gamlin et al. 2001; Clark, 2003). Because of gas transfer at the air-water interface in surface water, the use of these tracers presents a major challenge in that, after introduction, their concentrations in the recharge water must be monitored to determine the amount of tracer lost due to gas exchange. This is especially true if the artificial recharge is taking place in a shallow river or spreading pond.

Methods

Tritium/³He Ground Water Age Dating

Samples were collected between 1995 and 1997 from 57 locations near the OCWD recharge operation using both monitoring and public supply wells. At 15 locations, multi-

level monitoring wells were sampled. All tritium and ³He samples were collected in 500 mL glass bottles and 10 mL copper tubes, respectively. The copper tubes were sealed with steel pinch-off clamps after they were thoroughly flushed.

At Lawrence Livermore National Laboratory (LLNL), the copper tubes were attached to a high vacuum inlet system that led to a VG5400 noble-gas mass spectrometer. The ³He/⁴He isotope ratio and the concentrations of ⁴He, neon, and argon were determined using the method described by Radmacher et al. (2001). The mass spectrometer was calibrated with equilibrated water samples and known quantities of air. The uncertainties of ³He/⁴He isotope ratio and the ⁴He, neon, and argon concentration measurements were ± 0.5% and ±1%, respectively.

Tritium, which is reported in tritium units (1 TU equals a tritium to hydrogen ratio of 10⁻¹⁸), was measured by a ³He accumulation method (Clarke et al. 1976; Surano et al. 1992). The samples were vacuum degassed and shelved for ~60 days to allow for the growth of ³He from tritium decay. The uncertainty in the tritium measurement was ± 0.5 TU.

SF₆ Analysis

Samples of surface water for SF₆ analysis were collected in 50 mL glass syringes and analyzed within four hours on a gas chromatograph (GC) equipped with an electron capture detector using the headspace method described by Wanninkhof et al. (1987). SF₆ was separated from other gases with a molecular sieve 5A column held at room temperature. The detector response was calibrated about every 30 minutes with gas standards (~148.1 ppt and ~1947 ppt) certified by Scott-Marrin Inc. (Riverside, California, [909] 653-6780). The precision and detection limit were ± 3% and 0.04 pmol l⁻¹, respectively.

Following the tracer injection, OCWD staff collected ground water samples every two to 12 weeks from up to nine production wells and 31 monitoring wells, of which 11 were screened at multiple discrete depths. Initially, SF₆ samples were collected in copper tubes (~ 10 mL) that were sealed with steel pinch-off clamps and sent to the University of California, Santa Barbara (UCSB), where they were analyzed using a method developed by Gamlin et al. (2001). The detection limit and precision of the copper tube method were 0.2 pmol l⁻¹ and ± 15%, respectively. After two years, SF₆ samples were collected in a preweighed 15 or 20 mL Vacutainer™. These containers were partially filled (5 to 12 mL of water) at the wellhead and shipped to UCSB for analysis. (Tests have shown that samples can be stored for more than six months in a Vacutainer.) In the laboratory, these containers were weighed (to determine the sample size) and carefully filled with ultrahigh purity nitrogen gas (so that the final pressure was equal to ~1 atm). After one minute of shaking to equilibrate the nitrogen gas with the water sample, the headspace gas was analyzed on the GC. With this method, the detection limit was 0.03 pmol l⁻¹.

Stable Isotope Analysis

Stable isotope samples were collected in 60 mL glass bottles and sent to LLNL for analysis. The ¹⁸O/¹⁶O ratios were determined using the CO₂ equilibration method

(Epstein and Mayeda 1953) on an isotope ratio mass spectrometer. Analyses are reported in the standard δ notation (Craig 1961):

$$\delta^{18}\text{O} = \left(\frac{R}{R_{SMOW}} - 1 \right) \times 1000 \quad (2)$$

where R and R_{SMOW} are the $^{18}\text{O}/^{16}\text{O}$ isotope ratio of the sample and of the SMOW (standard mean ocean water) standard, respectively. The reproducibility was better than ± 0.1 ‰ for the $\delta^{18}\text{O}$ analyses.

Xenon Isotope Analysis

The xenon isotope samples were collected in 10 mL copper tubes sealed with hardened steel pinch-off clamps. At LLNL, the tubes were attached to a sample manifold using flare compression fittings that lead to a VG-5400 mass spectrometer. Each water sample was released into a 250 cm³ evacuated stainless steel flask by removing the bottom pinch-off clamp and heating the copper tube. The receiving volume was heated with a hot air gun to boil the sample for ~1 minute to completely degas the sample. The water was then frozen with liquid nitrogen (77 K) leaving the noncondensable gases in the headspace volume. This gas was expanded and cleaned using a series of cold-traps and titanium-alloy getters. Xenon was then collected on activated charcoal cooled to 77 K. After warming, the xenon was expanded into the mass spectrometer for isotope analysis.

Results and Discussion

$T^3\text{He}$ apparent ages of ground water in the Forebay area ranged from < 2 years near the artificial recharge facilities to > 40 years downgradient (Figure 2). Most of the ground water in the top 300 m of the aquifer was younger than 30 years and hence recharged after the initiation of the artificial recharge program. Downgradient from Anaheim Lake and Kraemer Basin, apparent ages increased with depth and to the west following the hydraulic gradient (Figure 2). Average linear velocities, implied by distances between apparent age contours, decrease away from the spreading centers. This is anticipated since the ground water basin increases in width and depth to the west. The $T^3\text{He}$ ages define an area of very rapid flow; apparent ages of < 2 years can be found at distances up to 3 km southwest of Kraemer Basin. Depth-specific age data plotted in cross section indicates that very young ground water (< 2 years) is penetrating to depths > 200 m below the northern spreading basins (Figure 2b).

$T^3\text{He}$ ages near the SAR indicate a much more complicated flow system than downgradient from Anaheim Lake and Kraemer Basin. Relatively young water (< 2 years) was found at the up- and downstream ends for the recharge section; however, ground water with apparent ages between 10 and 20 years was found near the river at relatively shallow depths between this young water.

$T^3\text{He}$ samples were collected from 11 wells more than once. Replicate ages agreed very well (within the analytical uncertainty), suggesting a relatively stable ground water flow system exists in the Forebay area.

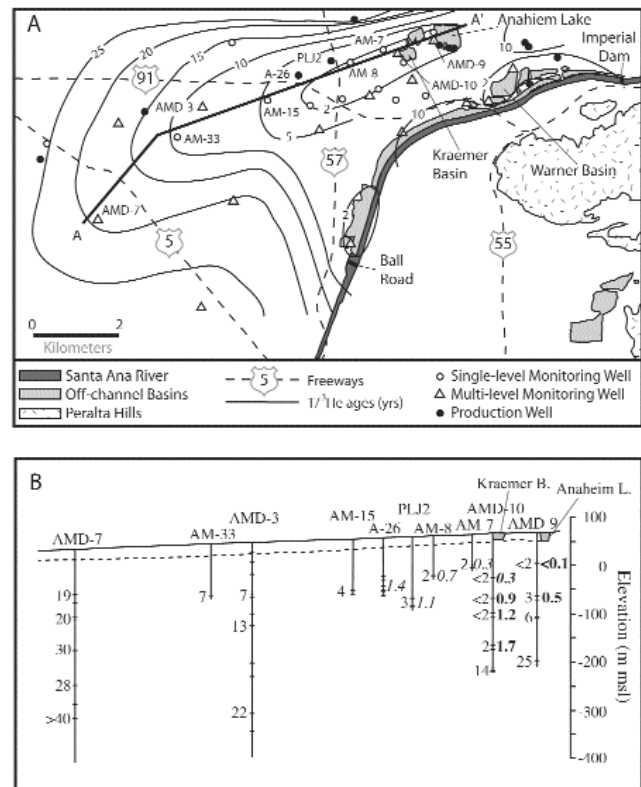


Figure 2. $T^3\text{He}$ apparent age data shown in (a) map view for the depth interval 0 to -80 m asl and (b) cross section down-gradient of Kraemer Basin and Anaheim Lake along A-A'. Next to each well screen on (b) are the $T^3\text{He}$ ages (to the left) and xenon isotope tracer arrival times (to the right). Xenon isotope arrival times are in bold and italic, respectively, for the Anaheim Lake and Kraemer Basin experiments. Tracer from both spreading areas arrived at AMD-10 #1.

SAR Tracer Experiment

99.8% pure SF_6 gas was released at three points along the length of the SAR for a period of 15 days in 1998 (July 20, 11:50 A.M. to August 4, 2:10 P.M.) by bubbling through diffusing stones (Gamlin et al. 2001). Concentration measurements in the SAR immediately downstream of the injectors indicate that only ~4% of the 7.8 kg released during the injection dissolved. The remainder was transferred directly to the atmosphere by the rising bubbles.

At the time of the experiment, the discharge rate in the SAR at the Imperial Dam (elevation = 80.5 m asl), the upstream end of the recharge section, averaged 2.8 m³/sec and the entire flow percolated into the subsurface before Ball Road (Figure 1). Thus, $\sim 4 \times 10^6$ m³ (3000 acre-feet) of surface water were tagged in the SAR recharge area with SF_6 . Mean concentrations in the SAR ranged between ~20 and ~300 pmol l⁻¹. The highest concentrations were found immediately below the gas injectors. Farther downstream, the concentration decreased as a result of gas loss across the air-water interface (Gamlin et al. 2001).

After percolating into the subsurface, the tracer was transported by the ground water flow system and observed at many wells. The duration the tracer was observed at each well varied considerably. Near the SAR, it was detected for as short as a tenth of a year (six weeks) while farther away it was observed for more than one year (Figure 3). The movement of the tracer within the ground water system was

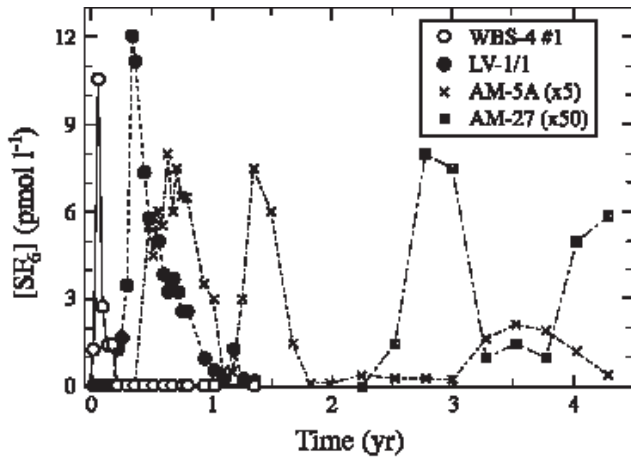


Figure 3. SF_6 breakthrough curves at selected wells determined during 1998 SAR experiment. See Figure 4 for well locations.

easily mapped (Figure 4). Initially, ground water flow was perpendicular to the river and two separate plumes formed up- and downstream of Warner Basin. At the time of the experiment, Warner Basin was recharging the aquifer with SF_6 -free water at a rate of $\sim 0.8 \text{ m}^3/\text{sec}$. Thus, Warner Basin created a water mound that diverted the flow of tagged ground water originating from the SAR. After about one year, these two plumes merged and began to flow to the west following the regional hydraulic gradient.

The effects of the water mound below Warner Basin were most dramatically seen at two wells, OCWD-FH1 and WBS-2A. At OCWD-FH1, which lies between the SAR and the basin (Figure 4), tracer arrived much later than at nearby wells. At WBS-2A, the tracer was never detected at the shallowest zone (59 m asl) while it was detected at the

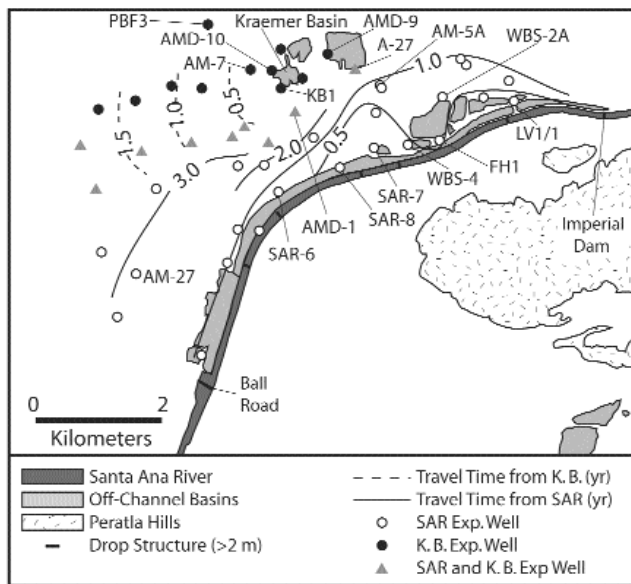


Figure 4. Map of the field area with travel times of the tracer center of mass (in years) from the SAR (solid contour lines) and Kraemer Basin (dashed contour lines) during the 1998 experiments. Wells sampled during each experiment are indicated on the map. The entire flow of the SAR had percolated into the ground by Ball Road during the tracer injection period.

deeper zones (47 and 33 m asl), suggesting that ground water from the SAR was flowing beneath water from Warner Basin.

At many wells, tracer was observed for more than a half-year showing the effects of dispersion and the complex flow field. The latter was best illustrated at a few wells such as AM-5A that showed multiple peaks in their breakthrough curves (Figure 3). Each peak probably represents a different flowpath from the river.

The tracer plume almost always arrived first at the shallower zones of the multilevel monitoring wells. A notable exception to this trend was observed at WBS-4 where three levels were monitored (zone 1, 53 m asl; zone 2, 36 m asl, zone 3, 9 m asl). At this well, the tracer arrived at zone 1 after 0.1 year, at zone 3 after 0.4 year, and was not detected at zone 2 until the third year. The relative late arrival at zone 2 is consistent with the T^3He data which indicates that zone 2 water (19 years) is substantially older than zone 3 (< 2 years).

The tracer data can be used to determine two important timescales of ground water flow. First, the initial arrival times of the tracer plumes represent the quickest flowpath from the SAR to the wells and are important when considering the transport of contaminants. The leading edge of the plume was traveling horizontally at rates between 0.1 and 4.3 km/year depending on location and depth of well (Figure 5a). Second, the arrival times of the center of mass were determined from the ratio of first and 0 moments following the procedure developed for river tracer experiments

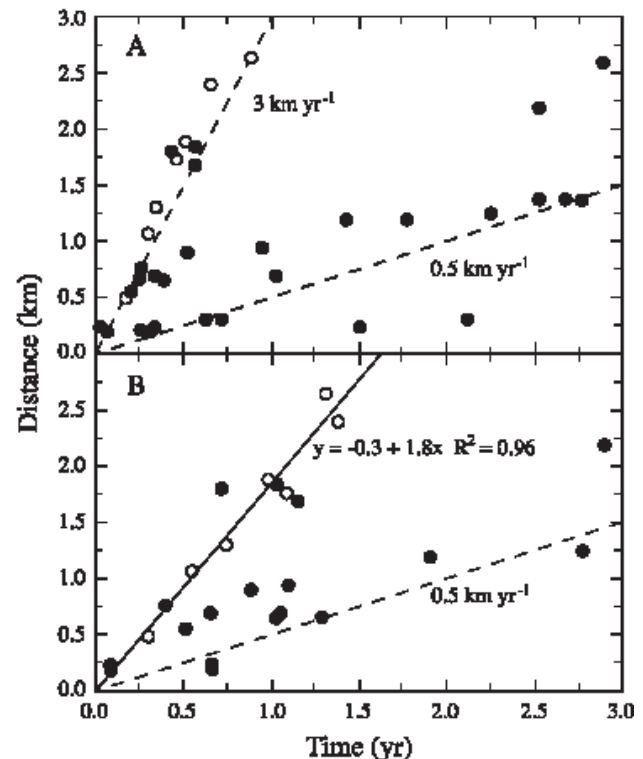


Figure 5. Tracer arrival times of (a) the first detection and (b) center of mass plotted as a function of distance along flowpaths that were inferred from the migration of the tracer plume and water table elevations (after Gamlin et al. 2001). Only tracer data from the SAR (filled circles) and Kraemer Basin (open circles) experiments are plotted. On (b), the best fit line through the Kraemer Basin data is plotted.

(Rutherford 1994). The center of mass moves at a rate equivalent to the mean linear ground water velocity. Thus, velocities calculated with the center of mass arrival times are comparable to flows calculated by Darcy's law. Near the SAR, linear velocities varied between ~0.3 and 2.9 km/year (Figure 5b). These variations in flow velocities are indicative of the complex hydrogeology of this basin.

Five multilevel and two single-level monitoring wells are located within 250 m of the SAR. The leading edge of the plume percolated into the top 50 m of the aquifer within 0.4 year (Figure 6), indicating the vertical velocity near the top of the aquifer averaged ~0.13 km/year, an order of magnitude less than the observed horizontal velocities. The plume arrived at the deepest monitoring wells (~90 m below the SAR) after about two years, demonstrating the vertical velocity was decreasing with depth.

The vertical migration was not uniform along the river. A zone with reduced vertical flow was identified with the SF₆ data midway through the recharge area near SAR-8, SAR-7, and WBS-4. This region is also apparent on the map of T³He ages (Figure 2). While the SF₆ plume showed very rapid flow to the shallowest zones (SAR-8 #1 and WBS-4 #1; > 53 m asl), the flow to deeper zones was slower than elsewhere along the river (Figure 6.). While tracer did arrive at WBS-4 #2 and #3, it did not arrive at SAR-8 #2 and #3 or at SAR-7 #1 and #2. This may be explained by the slight upward flow potential between the deeper and shallower zones at these wells.

Northern Basin Tracer Experiments

The northern recharge basins lie ~1.5 km from the SAR and recharge annually approximately the same amount of water as the river. These basins can be filled with either SAR or imported water. Over a period of ~6 weeks during the fall of 1996 and 1998, OCWD recharged, respectively, 7.4 × 10⁶ and 12 × 10⁶ m³ (6000 and 9500 acre-feet) of imported Colorado River water in Anaheim Lake and Kraemer Basin (1998 only). Prior to and after the introduction of imported water, these basins were recharging SAR water.

Because of its distant mountain origin, the δ¹⁸O composition of the imported water (~ -11.5 ‰) was ~4 ‰

lighter than both local ground water and SAR water (-7 ‰ to -8 ‰). The plume of recharged imported water down-gradient of Kraemer Basin and Anaheim Lake could be detected using δ¹⁸O to dilutions of ~10 to 1. During these experiments, rapid vertical transport was observed at wells adjacent to the spreading basins (A-27, A-28, AMD-9/1 near Anaheim Lake; and KB-1, AMD-10/1 near Kraemer Basin). As documented by the ~4 ‰ change in δ¹⁸O (Figure 7), the recently recharged imported water completely replaced earlier recharged local water within a month at some of these wells (A-27, AMD-9/1, KB-1). The arrival of imported water became difficult to determine with δ¹⁸O at wells > 1.5 km from the spreading ponds because of the natural variability of the δ¹⁸O composition of local ground water and dispersion.

About one week after filling the previously dry basins, isotopes of xenon gas (¹²⁴Xe in Anaheim Lake in 1996; ¹²⁹Xe in Anaheim Lake in 1998, ¹³⁶Xe in Kraemer Basin in 1998) were injected into the basin water. After introduction, the concentrations of the xenon isotopes were ~1000 times greater than the atmospheric solubility equilibrium values. Xenon isotope concentrations decreased exponentially with a timescale of ~1 week in the spreading basins; hence, the basins were enriched with xenon isotope tracers for ~3 weeks, half of the six-week percolation time of isotopically light imported water.

During the 1998 experiment, the ¹³⁶Xe tracer originating at Kraemer Basin could be detected at wells > 1.5 km from the spreading ponds and over a period of more than two years. The ¹³⁶Xe plume had a pie-wedge shape and was restricted to a region of high hydraulic conductivity material defined by the T³He ages (Figures 2 and 4). The arrival times of the xenon isotope tracers were approximately the same as δ¹⁸O, demonstrating that xenon was neither retarded within the aquifer nor lost in the vadose zone beneath the ponds (Figure 7.). The computed ground water linear velocity was 1.8 ± 0.2 km/year (Figure 5b). The rapid transport southwest of Kraemer Basin determined by the xenon tracer experiment is consistent with the T³He apparent ages which indicate that the ground water is < 2 years old. The half-peak width of the breakthrough curves, which

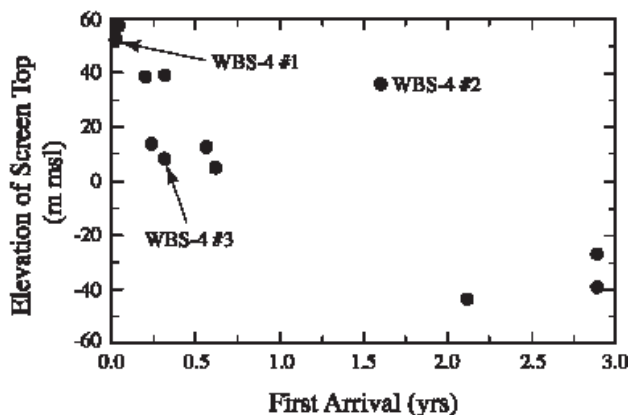


Figure 6. Vertical travel times to wells located within 250 m of the SAR determined during the SF₆ tracer experiment.

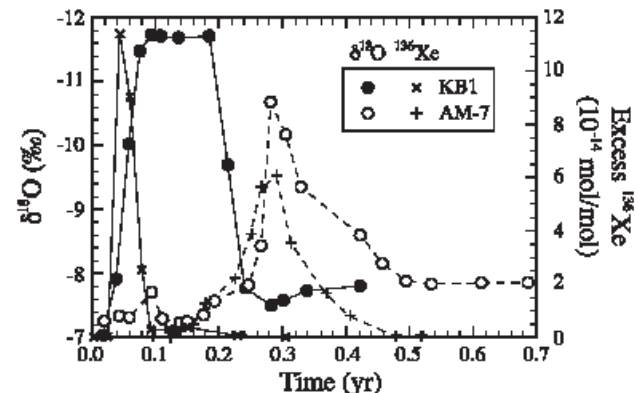


Figure 7. δ¹⁸O and ¹³⁶Xe breakthrough curves determined during the 1998 Kraemer Basin experiment. The sample dates have been adjusted so that 0 time corresponds to the start of each experiment. The δ¹⁸O axis has been reversed.

were calculated from the mean velocity and the duration tracer observed at each well, increased linearly with distance. The width was approximately two-thirds of the distance traveled (Figure 8).

The velocity field downgradient from Kraemer Basin appeared to be more uniform than near the SAR (Figures 4 and 5). This difference was partially caused by the geometry of the tracer source. In the case of Kraemer Basin, it was approximately a point source, whereas the SAR was a 9 km long line source.

Ground water recharged at Anaheim Lake was detected in very few wells away from the lake. The tracer plume was detected in the lower zones of AMD-10, which lie to the west of Kraemer Basin, indicating that the Anaheim Lake water was traveling beneath this basin (Figure 2). The plume water was also detected farther downgradient at two wells (AMD-11 and PBF3) outside of the plume emanating from Kraemer Basin. Based on these results, the water recharged at Kraemer Basin appears to deflect the westerly flow of water recharged at Anaheim Lake in a manner similar to the effect of Warner Basin's recharge mound on SAR-recharged water noted previously.

Summary

The geochemical techniques employed here successfully imaged the flow near the OCWD recharge operations over a range of temporal and spatial scales. T^3He measurements were able to identify large-scale features and broadly define ground water ages, while the deliberate tracer experiments were able to define flow patterns in detail over short periods. Most importantly, the deliberate tracer techniques developed for OCWD were able to determine flowpaths over a period of at least two years due to the sufficiently large volumes of recharge water tagged and low detection limits of the tracers used. Using the arrival time information, both mean (linear) and maximum ground water velocities were calculated from the experimental data.

The deliberate tracer data depict a rapid flow system in the upper 100 to 200 m of the aquifer near the recharge facilities. Horizontal velocities were on the order of 1 km/year. In the study area, the hydraulic gradient ranged

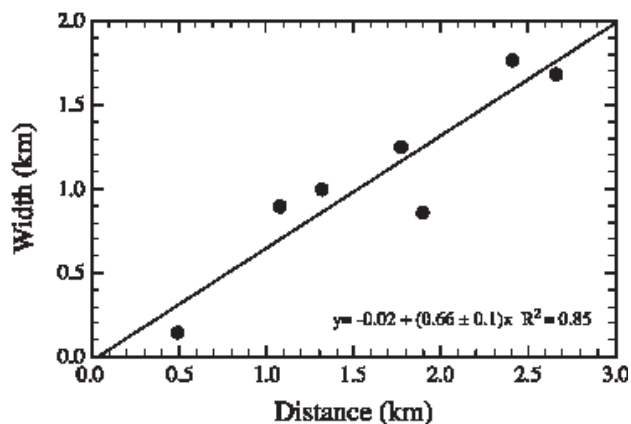


Figure 8. Half-peak width of the breakthrough curves at wells downgradient of Kraemer Basin.

between 0.010 and 0.005; hence, the tracer plumes were flowing through material with a hydraulic conductivity on the order of 10^2 m/day. Whether the plume was traveling through the entire aquifer or only the most conductive layers is difficult to determine directly with the tracer data; however, given the very heterogeneous hydrostratigraphy of the aquifer (OCWD 1998; Tompson et al. 1999) and the very high velocities observed, it is likely that the tracer plumes were moving primarily through the most conductive layers. This is especially true near the river where adjacent wells often displayed very different breakthrough curves that could include very long tails and multiple peaks.

Both T^3He ages and gas tracer travel times were determined at 19 narrow screened monitoring wells. With only three exceptions, the mean travel times and the apparent ages agreed within the analytical uncertainties of the methods. The agreement was also very good at two of the three production wells sampled for both tracers. The poor agreement was found at wells near the SAR where the flow and hydrogeology is more complicated.

The very fast travel times, good agreement with T^3He ages, and the similar $\delta^{18}O$ breakthrough curve indicate that the gas tracers behaved conservatively in the aquifer and were not significantly retarded during infiltration from the spreading facilities. Laboratory experiments have demonstrated gas tracers are retarded when traveling through partially saturated zones (Gupta et al. 1994). The absence of retardation indicates the tracers were infiltrating primarily through saturated material.

The tracer plume originating from the SAR was found to the south of the plume originating from Kraemer Basin. Tracer from the different experiments was never detected at the same well. It is likely that a very narrow mixing zone exists between these recharge plumes that could not be defined because of the limited number of existing monitoring wells. Nevertheless, the experimental results indicate Kraemer Basin is replenishing a different part of the aquifer than the SAR. In a similar fashion, recharge plumes originating from Anaheim Lake and the SAR were found to flow around or beneath downgradient plumes associated with recharge from Kraemer and Warner basins, respectively. There are no known geological features that could cause these plume separations; hence, they are features of the flow system.

In addition to providing direct measurements of travel times to wells and mapping the flow field, the tracer data can be used to validate future and existing models of flow (Thomson et al. 1999) and determine relatively large-scale hydraulic conductivity for the aquifer material. Numerical models are needed to assess how changes in recharge and pumping rates may affect travel times in the future. The tracer techniques also provide critical data for evaluating water quality variation by establishing hydraulic connections between the recharge locations and wells.

The timescales of flow determined by deliberate tracer experiments are needed to assess the potential transport of time-sensitive contaminants. These timescales are very difficult to determine with other means, especially if the flow is occurring preferentially in zones of high conductivity as observed in our experiments. Results from Orange County

indicate that potential recharge regulations for recycled water based solely on limiting how close production wells are to recharge areas would be inadequate for ensuring recharge water residence times of at least six months in the subsurface prior to extraction for potable supply. Close to the SAR and the northern spreading ponds, recharge water arrival times at downstream wells were a function of horizontal distance, inferred aquifer properties, and depth. Regulations need to consider all of these parameters.

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