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A basin-scale approach for assessing water resources in a semiarid environment: San Diego region, California and Mexico

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Abstract. Many basins throughout the world have sparse hydrologic and geologic data, but have increasing demands for water and a commensurate need for integrated understanding of surface and groundwater resources. This paper demonstrates a methodology for using a distributed parameter water-balance model, gaged surface-water flow, and a reconnaissance-level groundwater flow model to develop a first-order water balance. Flow amounts are rounded to the nearest 5 million cubic meters per year.

The San Diego River basin is 1 of 5 major drainage basins that drain to the San Diego coastal plain, the source of public water supply for the San Diego area. The distributed parameter water-balance model (Basin Characterization Model) was run at a monthly timestep for 1940–2009 to determine a median annual total water inflow of 120 million cubic meters per year for the San Diego region. The model was also run specifically for the San Diego River basin for 1982–2009 to provide constraints to model calibration and to evaluate the proportion of inflow that becomes groundwater discharge, resulting in a median annual total water inflow of 50 million cubic meters per year. On the basis of flow records for the San Diego River at Fashion Valley (US Geological Survey gaging station 11023000), when corrected for upper basin reservoir storage and imported water, the total is 30 million cubic meters per year. The difference between these two flow quantities defines the annual groundwater outflow from the San Diego River basin at 20 million cubic meters per year. These three flow components constitute a first-order water budget estimate for the San Diego River basin. The ratio of surface-water outflow and groundwater outflow to total water inflow are 0.6 and 0.4, respectively. Using total water inflow determined using the Basin Characterization Model for the entire San Diego region and the 0.4 partitioning factor, groundwater outflow from the San Diego region, through the coastal plain aquifer to the Pacific Ocean, is calculated to be approximately 50 million cubic meters per year.

The area-scale assessment of water resources highlights several hydrologic features of the San Diego region. Groundwater recharge is episodic; the Basin Characterization Model output shows that 90 percent of simulated recharge occurred during 3 percent of the 1982–2009 period. The groundwater aquifer may also be quite permeable. A reconnaissance-level groundwater flow model for the San Diego River basin was used to check the water budget estimates, and the basic interaction of the surface-water and groundwater system, and the flow values, were found to be reasonable. Horizontal hydraulic conductivity values of the volcanic and metavolcanic bedrock in San Diego region range from 1 to 10 m per day. Overall, results establish an initial hydrologic assessment formulated on the basis of sparse hydrologic data. The described flow variability, extrapolation, and unique characteristics represent a realistic view of current (2012) hydrologic understanding for the San Diego region.

1 Introduction

Current hydrologic understanding of the San Diego region consists of generalized summaries, site-specific evaluations, and project-design engineering studies (Ellis and Lee, 1919; Izbicki, 1985; Bondy and Huntley, 2000; CH2MHILL, 2003). Characterization of area-scale recharge/runoff,

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groundwater movement, groundwater/surface-water interactions, discharge, and aquifer geometry do not exist. Because of limited local surface and groundwater resources and the widespread availability of imported water, there has historically been little need to identify these characteristics at the area scale. However, increasing water demands are creating a commensurate need for integrated understanding of local water resources. This paper presents a first-order water budget for the San Diego region and describes the methodology for deriving water budgets developed from sparse hydrologic data. The term "first-order" implies (1) only the largest inflows and outflows are considered, (2) annual flow values are a hybrid statistic that combines average and median values, and (3) all flow values are rounded to the nearest 5 million cubic meters per year (million $m^3\,yr^{-1}$).

Because of stream gage location limitations, a water budget was specifically formulated for the San Diego River basin and extrapolated for the entire San Diego region. The water budget was framed in terms of (1) total water inflow, (2) surface-water outflow as measured by US Geological Survey stream gaging station 11023000 (San Diego River at Fashion Valley, Fig. 1), and (3) groundwater outflow. Total water inflow is estimated using the Basin Characterization Model (BCM; Flint and Flint, 2007a, 2012b); surface-water outflow is modified from gaged information; and groundwater outflow is calculated as the difference between total water inflow and surface-water outflow. The hydraulic implications of the derived outflow were checked using a reconnaissance-level steady-state numerical simulation of groundwater flow (MODFLOW; Harbaugh, 2005).

The BCM is a distributed parameter water-balance model that uses mechanistic, process-based algebraic equations to perform water-balance calculations. The calculations are performed at a monthly time step and independently at an evenly distributed 270 square meter (m^2) grid cell spacing. The equations utilize (1) topography, soil properties, and geology datasets, which are essentially static with time, and (2) precipitation and temperature datasets, which are spatially interpolated from weather station information and vary monthly. Water balance is formulated in terms of precipitation inflow and evaporated/transpired/sublimated outflow. Excess water is partitioned into recharge (BCM_{rch}) and runoff (BCM_{run}) for each grid cell. Partitioning is used for BCM calibration and the MODFLOW simulation. Partitioned values are not used for water budget calculations. Instead, grid cell values are summed to quantify total water inflow for individual river basins, and tributary sub-basins, within the San Diego area. Additional details of the BCM are presented in Appendix A.

Historically, models that incorporate evapotranspiration (Alley, 1984), inverse modeling (Sanford et al., 2001), or lysimetry and tracer tests (Gee and Hillel, 1988) have been used to assess water inflow. Water balance estimates and segregation into groundwater recharge and surface-water runoff has been done at the site scale (Flint et al., 2001; Ragab et al., 1997) and integrated with various measurements addressing

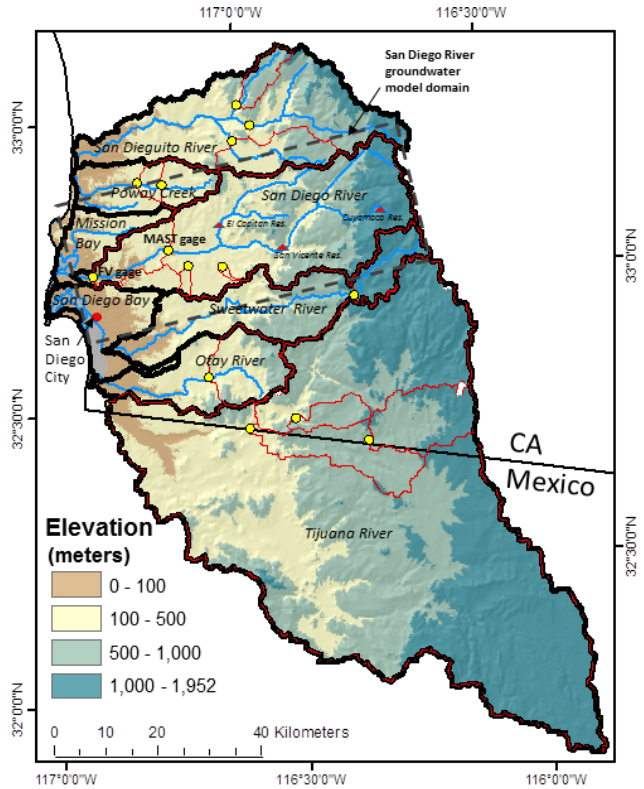

Fig. 1. Map of study area with major river basins outlined in black and calibration basins in red. Streamflow gages are noted as yellow points. San Diego River groundwater model domain, Mast Road (MAST) and Fashion Valley (FV) gages, and reservoirs in the San Diego River basin are indicated.

different spatial scales (Flint et al., 2002). Water-balance modeling has been done at a regional scale by Hevesi et al. (2003), Flint et al. (2004), Chen et al. (2004), and Flint and Flint (2007a, 2012b). A complete discussion of the use of the water balance to quantify hydrologic conditions in arid and semiarid regions is in Appendix B, and describes the episodic nature of recharge in locations where the precipitation occurs during months when the potential evapotranspiration is low, and there is little to no precipitation in months when the potential evapotranspiration is high. The BCM incorporates the historical knowledge by using monthly historical transient time series as climate input; the version used in this analysis has been updated and refined from earlier published versions, and includes refinements in the soils data, historical climate, and the potential evapotranspiration (PET) calculations. Also, an empirical flow-routing scheme is employed that calculates stream channel processes to estimate streamflow, baseflow, and losses to groundwater.

Many basins throughout the world have sparse hydrologic and geologic data, but have increasing demands for water and a commensurate need for integrated understanding of surface and groundwater resources. Better understanding of these resources is a stepwise process requiring multiple and parallel approaches. In addition to the information presented in this paper, the US Geological Survey San Diego Hydrogeology project (http://ca.water.usgs.gov/sandiego) includes drilling and construction of thirteen multi-level monitoring wells. Data collected from these monitoring wells are the basis for other concurrent investigations of the San Diego region hydrology. Those efforts include collection of borehole data and three-dimensional geologic mapping, geochemical signatures of the water, aquifer testing, development of a long-term hydrologic-data network, and seawater/freshwater dynamics. The fundamental goal of the work presented in this paper, and the San Diego Hydrogeology project as a whole, is to provide reliable hydrologic interpretations that can be used to make informed water utilization and management decisions.

1.1 Study area

The San Diego region includes 5 major basins that drain to the ocean across the coastal plain, which is generally defined as alluvial fill on the plain west of the mountains. The basins are San Dieguito River, San Diego River, Sweetwater River, Otay River, and Tijuana River (Fig. 1). These basins and associated tributaries make up a drainage area of approximately 8000 square kilometers (km^2) that ranges in elevation from sea level at the coast to 3700 m along the eastern boundary. The region has the highest variability of surface-water flow in the United States (Pryde, 1976). Surface-water flow is strongly correlated with the Pacific Decadal Oscillation (PDO) (Milliman et al., 2008). In addition to water resources, there are important ecological systems in the San Diego area. There are more endangered and threatened species in San Diego County than in any area in the nation. The coastal sage scrub ecosystem found in the county is one of the most endangered environments in the entire world (Cleland, 2009).

1.2 Climate

The San Diego region climate is classified as arid in the coastal plain and transitions to semiarid in the mountains to the east. Rainfall is closely associated with storms that approach from north, northwest, west, or southwest. Rainfall amounts vary from one local geographic area to another during each storm. Rainfall increases with distance inland as elevations increase, with orographic effects resulting in the highest rainfall at the highest elevations. The precipitation also decreases slightly along the coast from north to south (Elwany et al., 1998). Climatic conditions in the San Diego region are generally characterized by low rainfall (average annual precipitation of about 390 $mm\,yr^{-1}$), high evaporation rates (average annual potential evapotranspiration (PET) $\sim$ 1300 (700–1600) millimeters per year; ($mm\,yr^{-1}$)), and little or no summer rainfall.

Average annual precipitation over 4 of the 5 river basins within the San Diego region (San Diego River, Sweetwater River, Otay River, and Tijuana River basins) for the period 1940–2009 is shown in Fig. 2, with decadal averages indicated. Average precipitation ranges from about 150 to 750 $mm\,yr^{-1}$ and mean decadal values ranging from about 295 to 430 $mm\,yr^{-1}$, with 1990–1999 the wettest, and the last decade, 2000–2009, being the driest. For all years the San Diego River and Sweetwater River basins receive about 10 % more precipitation than the Otay River basin, which is at a lower elevation. There are several very wet years, such as 1983 and 1993, along with very dry years, when the Tijuana River basin receives nearly the precipitation of the San Diego River and Sweetwater River basins, but typically it receives about 5 % less.

The average precipitation during 1982–2009 (Fig. 2), the period used for BCM calibration and MODFLOW simulation, is the same as for 1940–2009, 389 $mm\,yr^{-1}$, but the variability about the mean is about 12 % higher. The more recent period has more years with low precipitation, and more years with high precipitation. The greatest decadal variations in precipitation from the last 70 yr of record occurred during 1989–2009. The seasonal trends in climate did not change significantly over the long term, but precipitation declined approximately 0.35 $mm\,yr^{-1}$, and maximum and minimum monthly air temperature increased 1.1 °C and 1.6 °C, respectively. The combined effect is a 7-$mm\,yr^{-1}$ increase in PET during 1940–2009.

2 Methods

A first-order water budget for the San Diego region was determined on the basis of total water inflow, surface-water outflow, and the difference between the two. The difference is considered groundwater outflow. Water-balance calculations were determined specifically for the San Diego River basin. The San Diego River basin was singled out because a stream gaging station is located near the terminal end (the Pacific Ocean coastline) of the river basin. The gaged flow is considered a reasonable representation of surface-water outflow from the basin. Flow at the gage was corrected to account for upper basin reservoir storage and imported water.

The ratio of groundwater outflow to total water inflow determined for the San Diego River basin was used to extrapolate groundwater outflow from the entire San Diego region. The ratio of groundwater outflow to total water inflow was compared to the streamflow components calculated from the BCM. Hydraulic rational and internal consistencies of the water balance were examined using an uncalibrated numerical simulation of groundwater flow.

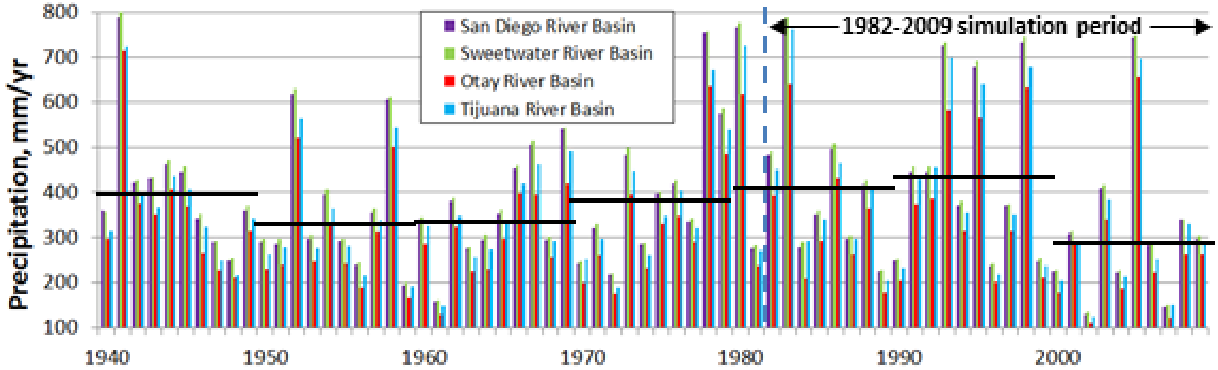

Fig. 2. Annual precipitation for 1940–2010 for the four major river basins in the San Diego/Tijuana study area with decadal mean precipitation indicated by the horizontal black lines. The gray shaded region indicates the groundwater model simulation period.

2.1 Total water inflow and water balance

The spatially distributed hydrologic processes and resulting total water inflow into the San Diego region were determined using the BCM (Flint and Flint, 2007a; Thorne et al., 2012; and Flint et al., 2011). To initiate the BCM, the San Diego region was gridded with a cell size of 270 m^2 and run monthly for 1982–2009. Model components that remained constant over the time period are soil properties (depth, water content at field capacity and wilting point, and porosity from SSURGO soil databases; NRCS, 2006; Fig. 3a) and topography (10-m digital elevation model; slope shown in Fig. 3b). Precipitation, air temperature (Parameter–Elevation Regressions on Independent Slopes Model, PRISM; Daly et al., 2008; 800-m transient dataset), solar radiation, and PET (Flint and Childs, 1987) vary monthly. Monthly values of PET (Fig. 3c) are accumulated from hourly calculations using the Priestley–Taylor equation (Flint and Childs, 1991). Actual evapotranspiration (AET) is calculated from changes in soil water storage.

Following the schematic illustrated in Fig. 4, once sublimation and AET are accounted for, excess water is partitioned into BCM_{run} and BCM_{rch} for each 270-m^2 grid cell. Total water inflow is the sum of BCM_{run} and BCM_{rch} and reflects natural hydrologic conditions. Diversions, reservoir storage or releases, urban runoff, groundwater pumping, or other impairments are not accounted for. The partitioning is controlled by shallow-depth bedrock permeability (K); the permeability values are initially estimated on the basis of geology (Jennings, 1977; Fig. 3d). To check and adjust BCM computations, results are compared to gaged surface water. Summing the grid cells that represent the drainage basin above a gaging location creates a monthly time series that can be compared to surface-water flow data. The time series are transformed using an empirical flow-routing scheme that conceptualizes surface-water discharge ($\mathrm{Stream}_{(i)}$), and regional groundwater flow ($\mathrm{GW}_{\mathrm{deep}(i)}$) in terms of the BCM_{run} and BCM_{rch} (Fig. 5). Empirical routing parameters and K are iteratively adjusted to achieve a "reasonable" match between average monthly surface-water flow and the BCM computed monthly $\mathrm{Stream}_{(i)}$ time series. Additional details of BCM datasets, computations, and empirical flow-routing are presented in Appendix A.

2.2 Surface-water flow

Stream gaging station records exist for 15 locations within the San Diego area (Fig. 1, Table 1). Surface-water flows at all the locations are impaired (altered) to some degree by reservoirs, urban runoff, imported water, waste water treatment plant effluent, and diversions. For the San Diego River basin, impaired surface-water flows for the San Diego River at Mast Road near Santee, California (hereafter referred to as Mast), and at Fashion Valley at San Diego, California (hereafter referred to as FV), were reconstructed to best reflect pre-development, unimpaired conditions. A continuous record of surface-water flow for 1982–2009 is available for both Mast and FV. The Mast gage is located 16.1 km upstream from FV, and the record has a possibility of error greater than 8 % of the reported flow (USGS, 2010). The FV gage is located 4.2 km upstream from the Pacific Ocean, and the record is considered accurate to within ±8 % of the reported flow (USGS, 2010). The estimated annual gaged flow estimate for 1982–2009 is summarized in 3 steps: (1) 10 277 average daily flows are used to calculate average flows for the 336 months of record, (2) the average monthly flows are summarized into 12 median monthly flows, and (3) the average of the median monthly flows are summarized into an annual flow value. The median statistic was used in step 3 to minimize the influence of extremely low and high flows specific to the 1982–2009 time period.

Cuyamaca, El Capitan, and San Vicente Reservoirs regulate surface-water flow at the Mast and FV gages (Fig. 1). To reconstruct surface-water flow to unimpaired conditions, the amounts of water entering and leaving El Capitan and San Vicente Reservoirs were examined (Cuyamaca regulation is aggregated with San Vicente). Reservoir conditions are recorded on a monthly basis by the City of San Diego

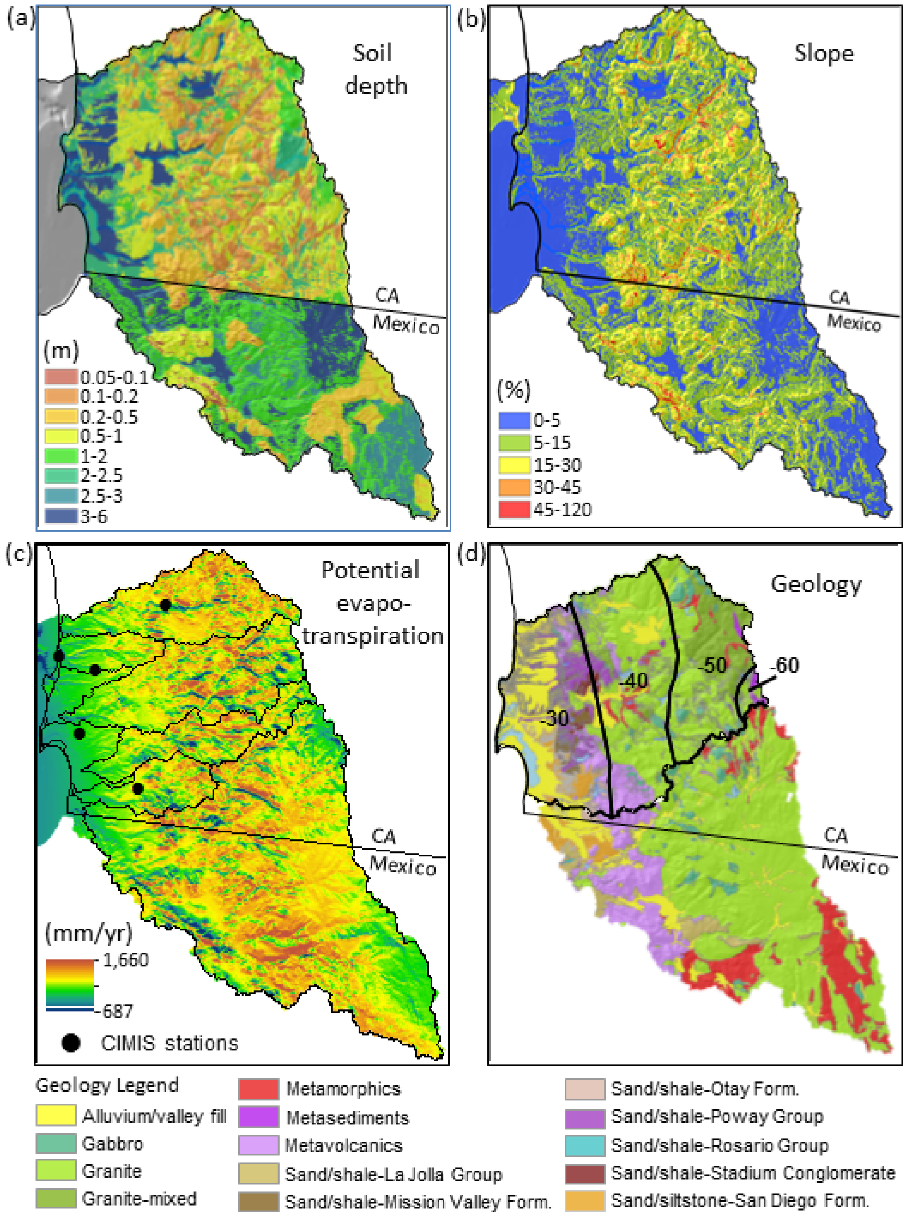

Fig. 3. Input maps for the Basin Characterization Model in the San Diego region study area illustrating **(a)** soil depth, **(b)** slope, **(c)** average annual potential evapotranspiration, and **(d)** geology and isotopic zones, as δ deuterium ‰.

Public Utilities Department, and records account for water leaving the reservoirs via evaporation, seepage, and export; entering water includes import through aqueducts, precipitation on the reservoir surface, and surface runoff from areas upstream of the reservoirs (J. Pasek, personal and written communication, City of San Diego, July 2011). A formal assessment of error associated with reservoir accounting has not been done, but an error of ±10 % is deemed reasonable (J. Pasek, personal communication, City of San Diego, July 2012). The qualitative error estimate is based on the number of outflow and inflow components considered in the calculations and the general consistency and thoroughness of the data. Also, efficient management of the reservoirs would be difficult if errors were greater than 10 % (i.e., other and better observations would have been implemented if errors were consistently greater than 10 %).

Using month to month accounting, increases in reservoir storage that exceed imported water were considered to be unimpaired surface-water flow at the dam location. It was assumed that 100 % of water exported from the reservoir is consumed, none returns to the stream below the reservoir. The median of calculated increases in reservoir storage were added to the measured surface-water flow at both Mast and FV. Flow alterations due to urban runoff, waste

Table 1. Stream gages used in the development of the Basin Characterization Model and San Diego River groundwater model.

Stream gage name	Station ID	Upstream area (km^2)	Period of record	Dominant geology
Guejito Ck. near San Pasqual, CA	11027000	159.7	1947–2007	granite
Santa Maria Ck. near Ramona, CA	11028500	88.9	1976–2007	granite
Santa Ysabel Ck. near San Pasqual, CA	11026000	46.5	1956–1978	granite, mixed granite
San Diego R. at Mast Road near Santee, CA	11022480	150.7	1912–2008	granite, mixed granite
San Diego R. at Fashion Valley at San Diego, CA	11023000	74.6	1982–2008	sandstone–shale
Los Penasquitos Ck. near Poway, CA	11023340	45.5	1969–1992	sandstone–shale
Los Penasquitos Ck. below Poway Creek near Poway, CA	11023330	45.0	1969–1993	sandstone–shale
Sweetwater R. near Descanso, CA	11015000	26.1	1956–2007	granite
Jamul Ck. near Jamul, CA	11014000	56.8	1949–1998	metavolcanics
Forester Ck. at El Cajon, CA	11022350	12.3	1983–1993	alluvium
Los Coches Ck. near Lakeside, CA	11022200	4.2	1983–2007	alluvium
Potrero Ck. Trib near Barrett Jct, CA	11011900	66.0	1966–1968	alluvium
Campo Ck. near Campo, CA	11012500	217.8	1939–2000	granite
Tijuana R. near Dulzura, CA	11013000	215.6	1939–1989	granite
Tijuana R. near Nestor, CA	11013500	3128.7	1939–1982	granite

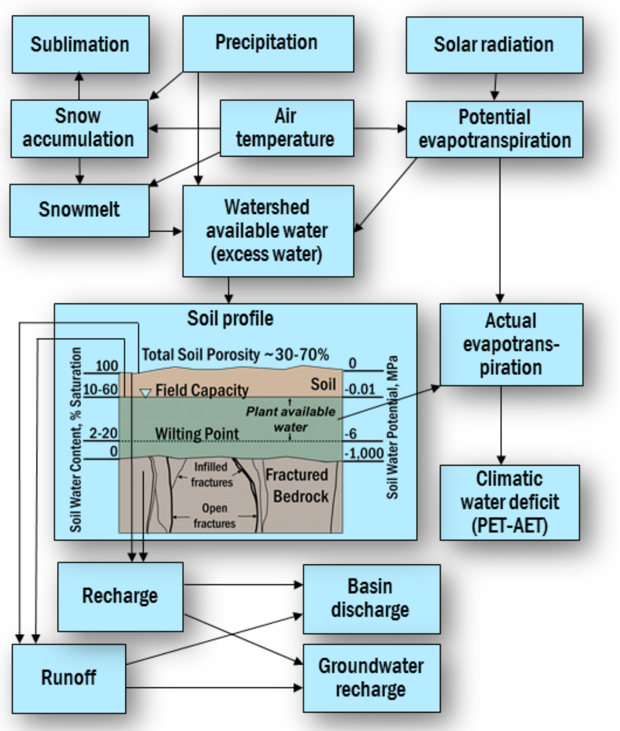

Fig. 4. Schematic illustrating the relation among the various components of the Basin Characterization Model.

water treatment plant effluent, and diversions are integrated into the surface-water flow record; any induced changes in surface-water/groundwater interaction are not considered for this reconstruction of total flow at the gages.

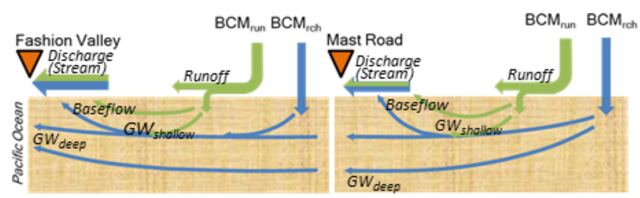

Fig. 5. Schematic illustrating the application of runoff and recharge from the Basin Characterization Model to the surface-water and groundwater system in the San Diego River basin.

2.3 Groundwater flow

The hydraulic rational and internal consistencies of the water balance derived from the difference between BCM-derived total water inflow and gaged/reconstructed surface-water outflow was assessed using a steady-state MODFLOW simulation of groundwater flow in the San Diego River basin. Groundwater withdrawal at wells, return flows from irrigation and lawn watering, and waste water treatment plant effluent were not simulated.

The model domain is delineated by no-flow boundaries that correspond to topographic divides for the eastern 2/3 of the basin and the bottom of the lowest model layer (Figs. 6 and 7). The domain was extended beyond the topographic divides for the western 1/3 of the basin to create a larger lateral interface with the coastal plain and Pacific Ocean. The model domain is horizontally discretized into 500-m^2 grid cells. Vertically, the domain consists of 3 layers that extend across the entire model domain. The altitude of the bottom of layer 1 (Fig. 6) is the Quaternary–Tertiary (or older) contact underneath the Mission Valley and Santee/El-Monte alluvial sub-basins. The lowest absolute altitude of the contact, and a linear interpolation between the two areas, defines

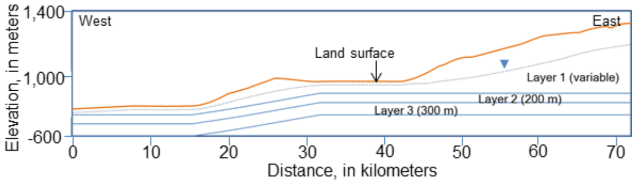

Fig. 6. Schematic cross-section of the San Diego River basin groundwater flow model domain showing generalized vertical model structure, land-surface elevation, and depth to groundwater.

Table 2. Runoff estimated from Basin Characterization Model, BCM_{run}, accumulated for each stream segment, and applied using Streamflow Routing Package.

Model stream segment	1	2	3	4	5	6	7	8
Runoff (million $m^3\,yr^{-1}$)	3.2	2.1	0.5	0.8	4.5	3.1	0.2	0.2

the altitude datum that is used to establish layer 2 and 3 top/bottom altitudes. Contact altitudes are based on information presented in the San Diego River System Conceptual Groundwater Management Plan (CH2MHILL, 2003).

Horizontal and vertical hydraulic conductivities were zoned in accordance with the surficial geologic map of the San Diego region and a three-dimensional geologic framework rendition of the coastal plain (Glockhoff, 2011). The regional geology is generalized into a crystalline rock zone that includes granite, gabbro, and unclassified crystalline rocks (zone 1), coastal plain sediments (zone 2), Quaternary alluvium (zone 3) and metavolcanics (zone 4) (Figs. 3d and 7b, c). For zone 1 and zone 4, the ratio of horizontal to vertical hydraulic conductivity was fixed at 1.0. For zone 2 and zone 3 the ratio was fixed at 10.0. The San Diego River and two tributaries (San Vicente and Boulder Creeks) were simulated using the Streamflow Routing (SFR-2) Package (Niswonger and Prudic, 2003). The stream network is represented with eight segments (Fig. 7a, Table 2). Stream segment altitudes were determined from the 10-m digital elevation model of the basin. Stream depth and width are fixed at 1.0 and 10.0 m for all segments. Depth and width estimates are rough estimates made from visual observations. Streambed conductance is set at 1.0 m per day (m d^{-1}). The RUNOFF term (Niswonger and Prudic, 2003, p. 24) for each of the eight segments was set equal to BCM_{run} for the portion of the drainage bisected by the segment (Table 2).

The Drain (DRN) Package (Harbaugh, 2005, pp. 8–43) is used to simulate the hydraulic connection with the Pacific Ocean (Fig. 6). The DRN altitudes assigned to layers 1, 2, and 3 are 0.252 m, 3.052 m, and 7.252 m, respectively, and correspond to the difference between freshwater and seawater hydrostatic pressures at the midpoint altitude of each layer; DRN hydraulic conductance is set equal to the simulated coastal-plain sediments. The Recharge (RCH) Package (Harbaugh, 2005, pp. 8–37) was used to simulate areal

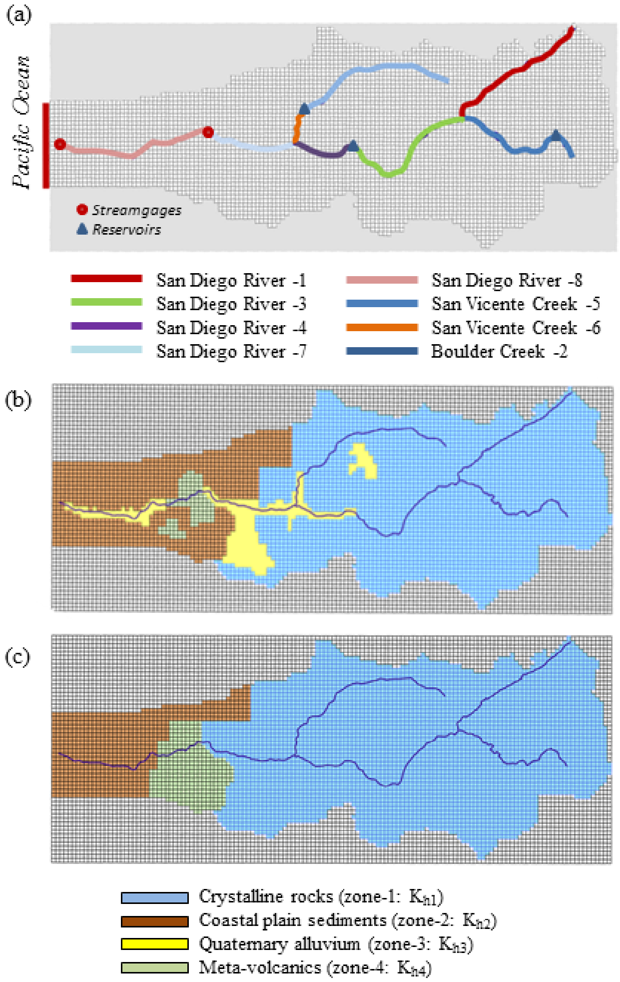

Fig. 7. Plan view of the San Diego River basin groundwater flow model domain illustrating the **(a)** grid, Streamflow Routing (SFR) Package boundary stream segments (indicated in legend, numbers correspond to Table 2), and Drain (DRN) Package boundary (red bar labeled as Pacific Ocean), and the geologic zones used in the **(b)** model layer 1, and **(c)** model layers 2 and 3.

groundwater recharge into the model domain across the uppermost layer and is set equal to BCM_{rch}. Both runoff and recharge corresponds to the spatial distribution and amount determined from the 1982–2009 BCM simulation.

Horizontal hydraulic conductivities and streambed conductance were adjusted so that the model (1) simulates the pattern of ephemeral surface-water flow along the upper reaches of the San Diego River, San Vicente Creek, and Boulder Creek, (2) reproduces the magnitude of reconstructed surface-water flow at Mast and FV, and (3) simulates a basin-scale hydraulic gradient that does not dramatically exceed or intercept land-surface topography. Flow model parameters were not formally adjusted to match specific flow and water-level observations.

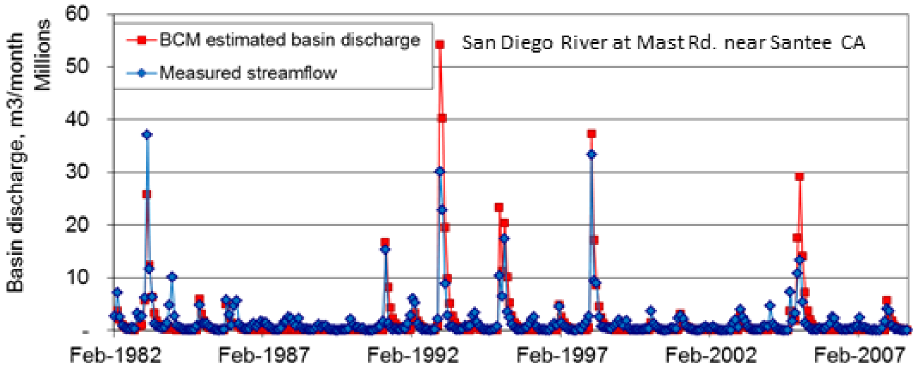

Fig. 8. Comparison of measured streamflow at Mast Road stream gage on the San Diego River with basin discharge estimated using the Basin Characterization Model.

3 Results

The differences in total water inflow as derived by the BCM, and surface-water outflow as described for the San Diego River at Fashion Valley, corrected for upper basin reservoir storage and imported water, are presented. The quantity is an estimate of groundwater flow through the coastal-plain aquifer adjacent to the San Diego River basin. A reconnaissance-level groundwater flow model for the San Diego River basin defines aquifer characteristics required by the groundwater estimate. The surface-water outflow and the groundwater outflow define a partitioning of the BCM-derived total water inflow.

3.1 Total water inflow and water balance

The average annual total water inflow to the San Diego area determined by the BCM during 1982–2009 was 50 million $m^3\,yr^{-1}$. Partitioned, BCM_{rch} is 20 million $m^3\,yr^{-1}$ and BCM_{run} is 30 million $m^3\,yr^{-1}$. Final shallow-depth bedrock permeabilities, the BCM parameter that controls partitioning, are listed in Table 3 for each geologic unit. Bedrock permeabilities were estimated iteratively by comparing BCM results to gaged surface-water flow at 15 locations (Table 1), using the empirical flow-routing equations described in Appendix A. Gaged surface-water flow with the least impairments (i.e., those upstream of reservoirs, without major diversions or urban runoff) were given more weight during calibration. Basin discharge, derived from the routing (Appendix A, Eq. A5) for the Mast gage, is shown in Fig. 8 for the period of record. The r^2 calculated from the gaged surface-water flow at Mast and modeled basin discharge is 0.83, and the Nash–Sutcliffe efficiency statistic (Nash and Sutcliffe, 1970), calculated as 1−(mean squared error/variance) for the period of record, is 0.86, indicating a good fit. The slight overestimation of peak flows by the BCM in comparison to measured flows is likely due to the retention of storm flows by the two reservoirs in the basin.

The spatial distribution of BCM_{run} and BCM_{rch} averaged for 1982–2009 is shown in Fig. 9. Very little estimated recharge or runoff occurs in an average year directly on the coastal plain, and much less occurs in the Tijuana basin than in the San Diego basin to the north, where high elevation mountains in the east receive somewhat more precipitation than those in the Tijuana basin (Fig. 9). Most of the BCM_{rch} (at least an order of magnitude more) is simulated for the eastern mountains, where the consolidated rock types are mainly granite and metavolcanic (Fig. 3d). The dominant factor, however, controlling the lower recharge and runoff in the Tijuana basin is the high elevation areas with low slopes (Fig. 3b) and thicker soils (Fig. 3a). Thick soils hold moisture in the profile making it available for evapotranspiration and loss to recharge or runoff.

To illustrate the conceptualization of recharge and runoff represented by the BCM, dominant contributions to the groundwater system, and the major physical features controlling the spatial distribution of recharge, total average annual BCM_{rch} (1982–2009) for individual river basins, geologic units, and regions with differing groundwater source elevations are included in Tables 3, 4, and 5. Although these tables do not include the variable contribution of runoff to groundwater recharge, they provide support to the concept that recharge occurs in the eastern higher elevation mountains and water flows to the western coastal plain. Most of the volume of simulated recharge is produced in the Tijuana River, San Diego River, and San Dieguito River basins (Table 3). However, when disregarding area and calculating as a rate in $mm\,yr^{-1}$, most of the simulated recharge occurs in the San Diego River, Sweetwater River, San Dieguito River, and Otay River basins (Table 3). As a result, when the calculated recharge is spread over such a large area, it is considered that the Tijuana River basin provides little recharge to the coastal plain and can be considered negligible. Although a large percentage of recharge is simulated to occur in the San Dieguito River basin, this river drains to the ocean and does not directly intersect the coastal plain aquifer. The three river basins that contribute recharge to the coastal plain are the

Table 3. Average annual recharge calculated using the Basin Characterization Model for all river basins in the San Diego/Tijuana study area for 1940–2009.

River basin	Area (km^2)	Average annual recharge (million $m^3\ yr^{-1}$)		Average annual recharge ($mm\ yr^{-1}$)	
		1940–2009	2000–2009	1940–2009	2000–2009
San Dieguito River	894	33.7	8.5	37.6	9.6
Poway Creek	244	4.1	1.5	16.9	6.1
Mission Bay	160	1.7	1.2	10.6	7.4
San Diego Bay	237	0.4	0.0	1.5	0.1
San Diego River	1121	53.9	17.5	48.1	15.6
Sweetwater River	564	25.3	7.4	45.0	13.2
Otay River	368	12.2	4.7	33.1	12.9
Tijuana River	4376	92.8	25.7	21.2	5.9

Table 4. Average annual recharge calculated using the Basin Characterization Model for geologic units in the San Diego region.

Geologic unit	Bedrock permeability ($mm\ day^{-1}$)	Area (km^2)	Mean recharge (1940–2009) (millions $m^3\ yr^{-1}$)	Mean recharge (2000–2009) (millions $m^3\ yr^{-1}$)
Alluvium	500.0	508	2.37	1.72
Gabbro	0.1	120	0.37	0.26
Granite	5.0	1437	49.70	33.30
Granite-mixed	10.0	387	31.52	19.74
Metamorphics – gneiss/schist	0.1	81	0.20	0.14
Metasediments	5.0	34	3.27	2.02
Metavolcanics	15.0	289	6.61	3.83
Sandstone La Jolla Group	5.0	165	0.81	0.64
Sandstone Otay Formation	50.0	34	0.21	0.07
Sandstone Poway Group	2.0	261	3.96	2.06
Sandstone Rosario Group	2.0	11	0.04	0.04
Sandstone San Diego Formation	5.0	131	0.55	0.12
Sandstone Mission Valley Formation	40.0	38	0.34	0.15
Sandstone Stadium Conglomerate	100.0	44	1.11	0.59

San Diego River, Sweetwater River, and Otay River basins, and have a long-term (1940–2009) average recharge volume of 91.4 million $m^3\ yr^{-1}$, and a recent (2000–2009) average volume of 29.7 million $m^3\ yr^{-1}$.

Within the San Diego River basin, a large proportion, at least an order of magnitude more, of the modeled recharge is located in the region defined by hard rock geology and dominated by granites (Fig. 3c; Table 4). This implies that the largest volume of recharge within the river basins is occurring east of the band of metasediments and metavolcanics that divide the coastal plain from the higher elevation hard rocks. In an effort to collect evidence supporting this preliminary conceptualization of the regional hydrology, groundwater data was collected from wells at a range of elevations throughout the region to determine the chemical characteristics of the locally recharged groundwater (as δ deuterium ‰; Williams and Rodoni, 1997). Those results were then compared to groundwater samples collected from basin aquifers on the coastal plain to assess which elevations may have contributed the most to the recharge (Fig. 3c). The recharge to the coastal plain was calculated for each of the three contributing river basins (Table 5). Although the data does not discriminate between river basin sources, it does indicate that the most recharge occurs in the −50 ‰ δ deuterium zone, which coincides with the high elevation, hard rock zone.

In addition, BCM output indicates that 90 % of simulated BCM_{rch} occurred during 3 % of the 1982–2009 period. A compilation of papers by IAEA (2001) based on field studies that estimate recharge at 44 benchmark sites showed that rainfall below 200 mm usually results in negligible recharge, similar to the model results shown in Fig. 9. An analysis of maps of recharge over a series of years clearly showed that very seldom does any recharge occur directly on the coastal plain, and only in years with very high precipitation. Additional details of episodic recharge in semiarid and arid environments are given in Appendix B.

Table 5. Average annual recharge calculated using the Basin Characterization Model for three river basins in the San Diego region for areas defined on the basis of measurements of δ deuterium, in per mil.

River basin	Average annual recharge (millions $m^3\,yr^{-1}$)							
	1940–2009				2000–2009			
	−30	−40	−50	−60	−30	−40	−50	−60
San Diego River	5.0	14.4	33.0	1.5	4.3	10.3	21.6	1.0
Sweetwater River	1.4	7.1	11.4	5.2	0.9	5.2	7.7	3.3
Otay River	3.4	8.7	n/a	n/a	2.8	6.0	n/a	n/a

Table 6. Gaged and reconstructed streamflow in the San Diego River basin, San Diego region, California, for 1982–2009.

Location	Median annual streamflow	Reconstructed median annual streamflow
	(million $m^3\,yr^{-1}$)	
San Diego River at El Capital Dam	n/a	7.6
San Vicente Creek at San Vicente Dam	n/a	2.5
San Diego River at Mast Road	10.4	20.5
San Diego River at Fashion Valley	18.7	28.8

3.2 Surface-water flow in the San Diego River basin

The measured median monthly and annual surface-water flow for 1982–2009 at Mast and FV and the reconstructed surface-water flow at El Capitan and San Vicente Dams are in Table 6. Summing measured and reconstructed surface-water flow, the annual surface-water flow was estimated to be 20 ± 3 million $m^3\,yr^{-1}$ at Mast and 30 ± 4 million $m^3\,yr^{-1}$ at FV.

3.3 Groundwater flow in the San Diego River basin

The total average annual BCM-derived water inflow to the San Diego River basin during 1982–2009 was calculated to be 50 million $m^3\,yr^{-1}$. Reconstructed surface-water flow at FV was estimated to be 30 million $m^3\,yr^{-1}$ at FV, and considered total surface-water outflow to the Pacific Ocean. This assumes no significant gain/loss of surface water along the 4.2-km stream reach between FV and the coast. Subtracting surface-water flow at FV from total water inflow, groundwater flow through the coastal-plain aquifer adjacent to the San Diego River basin is estimated to be 20 million $m^3\,yr^{-1}$ (Table 7). These flow values equate to a 0.4 partitioning factor. Forty-percent of the water inflow to the San Diego River basin ultimately exits the basin as groundwater flow to the Pacific Ocean; 60 % exits as surface-water flow. Groundwater and surface-water routing for the San Diego River basin were numerically simulated using MODFLOW and the SFR-2 boundary package. The average annual BCM_{rch} for the San Diego River basin during 1982–2009 (35 million $m^3\,yr^{-1}$) was input to the model domain as areal recharge using the RCH boundary. The average annual BCM_{run} during 1982–2009 (15 million $m^3\,yr^{-1}$) was subdivided, as per the BCM_{run} distribution, and applied to eight SFR-2 stream segments (Table 2). Model parameters that control horizontal hydraulic conductivity of the crystalline rock (Fig. 7, K_{h1}), metavolcanic (Fig. 7, K_{h4}), coastal plain conductivity (Fig. 7, K_{h2}), and streambed hydraulic conductivity (Fig. 7, K_{h3}) were adjusted. Values of parameters are listed in Table 8 along with the estimated and simulated surface-water and groundwater flows for which the simulated partitioning factor is 0.38.

Table 7. Groundwater flow to the coastal plain of the San Diego River for the calibration period, 1982–2009, calculated using two approaches, a mass balance approach with average annual, spatially distributed BCM recharge and runoff accumulated for the basin from which reconstructed streamflow is subtracted, and deep groundwater flow calculated from the partitioning of BCM recharge and runoff into streamflow components.

Method	Estimated flow (million $m^3\,yr^{-1}$)
Mass balance	
Runoff + Recharge, (equivalent to $BCM_{rch} + BCM_{run}$)	47.6
Reconstructed streamflow at Fashion Valley	28.8
Calculated subsurface flow in coastal plain	18.8
Basin Characterization Model	
Basin discharge at Fashion Valley (Stream)	34.6
Calculated subsurface flow in coastal plain (GW_{deep})	13.0

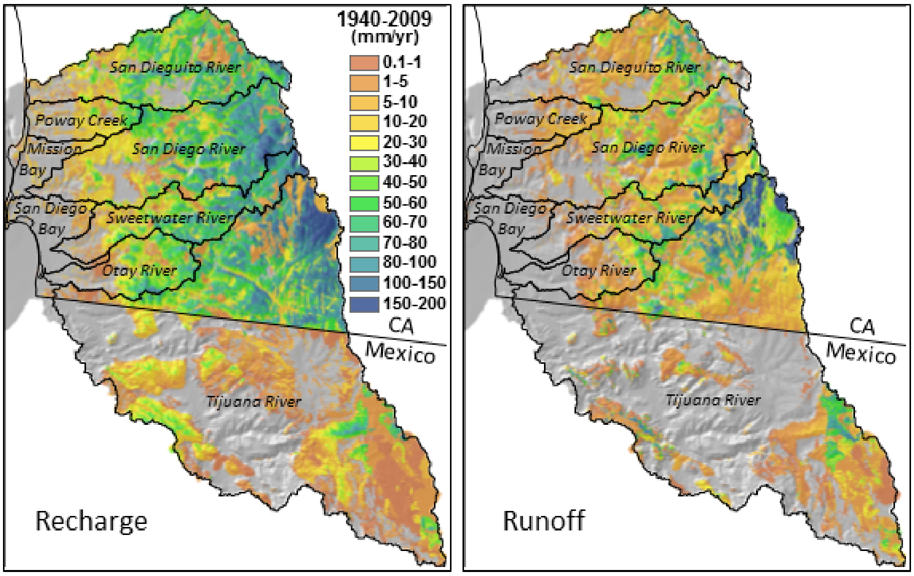

Fig. 9. Maps of average annual recharge and runoff for 1940–2009 calculated using the Basin Characterization Model for the San Diego/Tijuana study area. Gray indicates no recharge or runoff, and the scale applies to both figures.

Table 8. Groundwater-flow model parameter values and simulation results, San Diego region, California.

Horizontal hydraulic conductivity (m day^{-1})	Parameters
K_{h1}: crystalline rocks, zone 1	8.0
K_{h2}: coastal plain sediments, zone 2	2.7
K_{h3}: quaternary alluvium, zone 3	10.0
K_{h4}: metavolcanics, zone 4	0.2
BCM_{rch} (million m^3 yr^{-1})	33.1
Subsurface groundwater flow (million m^3 yr^{-1})	18.0
Sum of squares weighted residual	293

4 Discussion

General evidence from multi-completion monitoring wells and water-supply wells shows that groundwater in the coastal-plain aquifer is a mixture of freshwater and seawater. The fact that the coastal plain is not fully inundated by seawater necessitates a degree of freshwater inflow; as groundwater originating from the eastern mountains or from losing streams. Utilizing the available data, incorporating clearly defined physical processes, and accounting for the spatial and temporal variations, the BCM simulates a scientifically and intuitively reasonable estimate of precipitation that becomes total water inflow to the terrestrial hydrologic cycle. The quantity, timing, and pattern of surface-water flow are a comprehensive integration of the terrestrial hydrologic processes. Measurements of surface-water and reservoir inflow/outflow make it possible to quantify surface-water outflow for selected river basins of the San Diego area. Exploiting the difference between total water inflow and surface-water outflow, the amount of groundwater flow through the coastal plain aquifer was estimated.

The BCM-derived total water inflow and reconstructed surface-water outflow for the San Diego River basin (for 1982–2009) quantified ratios of surface-water outflow and groundwater outflow to total water inflow as 0.6 and 0.4, respectively. Using total water inflow determined from the BCM applied to the entire San Diego region (120 million m^3 yr^{-1} for 1940–2009), and the 0.4 partitioning factor, groundwater outflow from the San Diego area and through the coastal plain aquifer to the Pacific Ocean was estimated to be 50 million m^3 yr^{-1} (Table 9). The possible range of groundwater outflow cannot be objectively quantified.

Within the BCM, the level of spatial and temporal detail built into precipitation and PET, and their physically-based interactions, are designed to make full use of commonly available area-scale datasets, deterministic calculations, and calibrations to measured data. Also recognizing the episodic nature of recharge (occurring only when precipitation far exceeds PET (see Appendix B)), the total water inflow is considered to be generally well-constrained. Using the monthly median to describe an annual inflow for 1982–2009 does not inappropriately weight months when differences between precipitation and PET are extremely small. These factors are tempered by the inherent difficulty of estimating heterogeneous physical responses at all spatial and temporal scales. Uncertainties associated with shallow-depth bedrock permeability and soil characteristics (depth and storage capacity) are for the most part irrelevant to the water budget analysis. These uncertainties create error in the partitioning of total water inflow, but the partitioned BCM_{rch} and

Table 9. Recharge and runoff derived from the Basin Characterization Model for 1940–2009 for the San Diego, Sweetwater, and Otay River basins and proportion of subsurface groundwater flow to the coastal plain, calculated as 40 percent of total $BCM_{rch} + BCM_{run}$ for each basin, San Diego region, California.

River basin	BCM_{rch}	BCM_{run}	Subsurface groundwater flow	Subsurface groundwater flow
	(million $m^3\,yr^{-1}$)			Acre-feet yr^{-1}
San Diego River	53.9	16.9	28.3	22 940
Sweetwater River	25.3	10.6	14.4	11 670
Otay River	12.2	3.0	6.1	4950
Total	91.4	30.6	48.8	39 560

BCM_{run} amounts are not used independently in the calculation of groundwater outflow; the sum (total water inflow) is used in to calculate groundwater outflow.

The reconstructed surface-water outflow has a cumulative error associated with the stream gaging record (±8 %) and the reservoir accounting (±10 %). Assuming that the stream and reservoir estimate errors are normally distributed and not related, the cumulative error is estimated at ±13 %. More elusive is the error associated with the assumption that "*100 % of water exported from the reservoir is consumed, none returns to the stream below the reservoir*". If some exported water returns to the stream, that portion of the "reconstructed" flow is integrated into the surface-water flow record, which results in double counting. The 100 % assumption insinuates that the reconstructed flow estimate is a maximum. Apart from the reservoir effects, the estimated surface-water flow captures (or integrates) the effects of urban runoff, waste water treatment plant effluent, and diversions.

Using the difference between total water inflow and surface-water outflow, groundwater flow through the coastal-plain aquifer adjacent to the San Diego River basin is estimated to be 20 million $m^3\,yr^{-1}$. Results of the un-calibrated numerical simulation of groundwater flow suggest that the 20 million $m^3\,yr^{-1}$ is near the upper plausible limit. To simulate that amount of groundwater flow, a horizontal hydraulic conductivity of 8.0 m day^{-1} was assigned to the bedrock in the eastern 2/3 of the San Diego River basin (Fig. 7b and c, K_{h1}). A hydraulic conductivity of 8.0 m day^{-1} is more typical of clean- to silty-sand and at least an order of magnitude above the typical range for igneous and metamorphic rock (Fitts, 2002, Table 3.1). It is possible that the extensional tectonic regime in the San Diego region has enhanced conductivity. Data are not available to quantify hydraulic characteristics of the bedrock. The K_{h1} value was adjusted on the basis of matching the pattern of gain/loss in the San Diego River upstream of El Capitan and San Vicente Reservoirs.

Using the 0.4 partitioning factor developed for the San Diego River basin and applied to the BCM-derived total water inflow for the San Diego region (120 million $m^3\,yr^{-1}$), groundwater outflow from the entire San Diego region is estimated at 50 million $m^3\,yr^{-1}$. Extrapolation implies that the physical and hydrologic processes that govern the relative proportion of recharge and runoff in the San Diego River basin are the same for the entire San Diego region. Extrapolation makes sense given that the topography and geologic structure in the San Diego River basin is similar to that of the entire San Diego region, and the BCM is developed for and calibrated to gage data throughout the entire region.

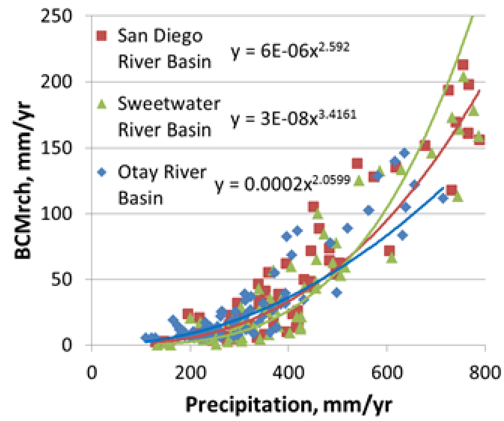

Fig. 10. Relation of BCM_{rch} to precipitation for 1940–2009 for three river basins.

Groundwater flow derived in this analysis is based on data/observations peculiar to the 1982–2009 time period and illustrates a fundamental complication associated with any water budget analysis. All hydrologic systems operate in two distinctly different time frames. Precipitation and runoff occur in minutes, hours, and possibly days. Groundwater flow occurs in years, centuries, and thousands of years. For the methodology used in this analysis, the disparity in time frames is exacerbated; surface-water outflow (which for the most part responds to short time-frame input) is used to quantify groundwater (reflecting long time-scale inputs). Even if it were possible to consider 100-yr meteoric and stream-gaging data, that still represents only a small interval of the time period imbedded in the regional groundwater flow system. Stream baseflow, regional spring discharge, and water levels are a more direct and accurate measurement of groundwater conditions. These data are not available or were not considered in this analysis.

BCM simulations indicate that the large variability in precipitation during 1982–2009 favors increased surface-water flows and decreased groundwater recharge due to the episodic nature of recharge in arid environments (see discussion in Appendix B). Figure 10 indicates that recharge increases exponentially with increases in precipitation in the river basins in the San Diego region, and Fig. 3, although having more years with high precipitation than the longer time period, also had many more low precipitation years, particularly during the last decade. This suggests that 1982–2009 surface-water flow may have a positive bias relative to the longer time-scale groundwater system. Use of the median statistic has removed some of the positive bias from the flow record, but the overall effect is under-predicting groundwater outflow. This somewhat tempers the previously stated "*upper plausible limit*" concerns as it applies numerically simulated groundwater flow.

5 Summary and conclusions

A first-order estimate of the average annual groundwater flow through the San Diego region coastal plain and out to the Pacific Ocean is approximately 50 million $m^3\,yr^{-1}$. The amount of groundwater flow is determined from the difference between total water inflow derived using the BCM distributed parameter precipitation–recharge–runoff model and gaged surface-water flow for the San Diego River at Fashion Valley. The BCM incorporates the physical system (soils and geology), and the climate variables of precipitation, air temperature and potential evapotranspiration defined on a monthly basis, in order to capture the temporal variability of the processes leading to total water inflow. Although interpolations, extrapolations, and parameter estimates introduce uncertainty, the episodic nature of recharge and runoff in semiarid environments insures a robust estimate of water inflow. The integrated nature of terrestrial hydrologic processes represented by the surface-water flow records also makes for a robust estimate of surface-water outflow. Using the difference between total water inflow and surface-water outflow, particularly for areas where hydrologic data are sparse, results in a reasonable first-order water budget.

Several conclusions can be made on the basis of the multiple approaches taken to substantiate preliminary estimates of groundwater flow to the coastal plain.

- Multiple lines of evidence (two models and geochemistry) support the conceptual model represented by the BCM that recharge primarily occurs in the eastern mountains of the region and that excess water is partitioned into recharge and runoff that eventually leave the upland basins via rivers and subsurface pathways.

- The Tijuana River basin is likely not a significant source of groundwater to the San Diego coastal plain on the basis of the distributed recharge calculated for that large basin.

- The groundwater flow estimate for the entire coastal plain corresponds to 5.8-mm yr^{-1} recharge for the San Diego region and is well within the regional estimates for arid and semiarid regions, and corresponds to 1.5 percent of precipitation, also within the regional estimates of 0.1 to 5 percent (Scanlon et al., 2006).

- Better defining the surface-water/groundwater interactions along the San Diego, Sweetwater, and Otay Rivers would help constrain regional groundwater flow estimates.

- Incorporation of available aquifer test information would help to quantify realistic ranges for the hydraulic conductivities of the crystalline and metavolcanic bedrock in the eastern portions of the San Diego region, and help to constrain potential groundwater flows.

- Better representation of the coastal plain sediments by (1) incorporating estimates of the seawater/freshwater interface geometry, (2) completely integrating the 3-dimensional structure of the coastal plain sediments into the MODFLOW model, and (3) designing and implementing aquifer testing would significantly improve the estimates of groundwater flow.

- Both the water-balance (BCM) and groundwater flow (MODFLOW) models should be used to assess the statistical significance of the various data-collection options in terms of better defining system parameters.

Appendix A

Basin Characterization Model

The Basin Characterization Model is a distributed parameter water-balance model that uses spatially distributed climate and physical properties, along with mechanistic, process-based algebraic equations to perform water-balance calculations. The calculations allocate precipitation into evapotranspiration, infiltration into soils, runoff, or percolation below the root zone to recharge groundwater. The relationship between runoff and recharge is driven by permeability of shallow-depth bedrock. Calculations are performed at a monthly time step and independently at evenly distributed 270-m^2 grid cell spacing. The BCM has been applied to the state of California and calibrated to streamflow at 138 basins to assess historical hydrologic processes and impacts of climate change on both water availability and ecosystems (Thorne et al., 2012). Because of the grid-based, simplified nature of the model, with no internal streamflow routing, long time series for very large areas can be simulated easily. More application and description of the model structure, input and output files, and model operation can be found in Thorne et al. (2012) and Flint et al. (2011). The BCM used in the San Diego area application has been updated and refined from earlier published versions, including refinements in the soils data, the historical climate, and the PET calibration.

The BCM relies on an hourly energy-balance calculation that is based on solar radiation, air temperature, and the Priestley–Taylor equation (Flint and Childs, 1991) to calculate potential evapotranspiration (PET; Flint and Childs, 1987). Clear sky PET is calculated using a solar radiation model that incorporates seasonal atmospheric transmissivity parameters and site parameters of slope, aspect, and topographic shading (to define the percentage of sky seen for every grid cell) (Flint and Flint, 2007b). Hourly PET is aggregated to a monthly rate and cloudiness corrections are made using cloudiness data from National Renewable Energy Laboratory (NREL). Modeled PET for the southwest United States was then calibrated to the measured PET rates from California Irrigation Management Information System (CIMIS) and Arizona Meterological Network (AZMET) stations, and is shown for the San Diego region in Fig. 3c. It is clear from the map that the highest PET is on high slopes with southern facing aspects. The modeled PET was compared to the CIMIS stations in San Diego County to estimate the local error associated with the regional calibration. Five stations are located in relatively low elevation agricultural areas around the region and have periods of record ranging from 1999–2010 to 2002–2010. A comparison of mean monthly PET for the five stations for the period of record for each station (Fig. A1) yielded a standard error of the regression of 13 mm month^{-1}, distributed variably throughout the year. When forced through zero, the regression equation has a slope of 1.067, indicating a slight overestimation of the simulated evapotranspiration in general. The months with precipitation are indicated as red points (November–April), but the months with the most recharge (during March and April snowmelt), have the least variability around the mean.

Using PET and gridded precipitation, maximum, and minimum air temperature (Parameter–Elevation Regressions on Independent Slopes Model, PRISM; Daly et al., 2008; 800-m transient dataset) and the approach of the National Weather Service Snow-17 model (Anderson, 1976), snow is accumulated, sublimated, and melted to produce available water (Fig. 4). Snow cover estimates for California were compared to Moderate Resolution Imaging Spectroradiometer (MODIS) snow cover maps (Flint and Flint, 2007a) and snow courses and sensors throughout the Sierra Nevada.

All input data is spatially downscaled or interpolated to the 270-m^2 grid resolution for model application following Flint and Flint (2012a). This downscaling approach was shown to not introduce additional uncertainty but indeed improved the estimate of the climate parameter by incorporating the

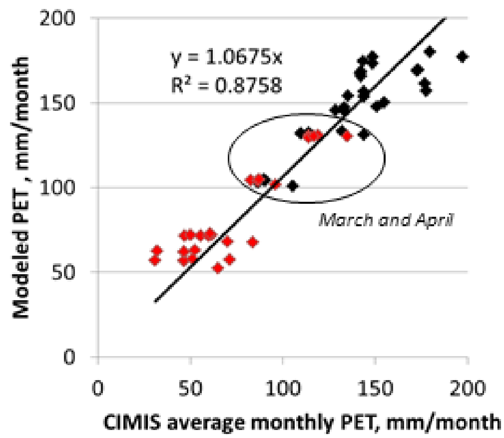

Fig. A1. Comparison of modeled potential evapotranspiration (PET) and PET measured at five stations from the California Irrigation Management Information System (CIMIS) in the San Diego region. Red points indicate November–April, black points indicate May–October.

deterministic influence (such as lapse rates or rain shadows) of location and elevation on climate. For the San Diego region, the climate surfaces and monthly PET were combined with maps of elevation, bedrock permeability estimated on the basis of geology (Jennings, 1977; Fig. 3d) and iteratively modified in the model calibration process, and soil-water storage from the SSURGO soil databases (NRCS, 2006). Total soil-water storage is calculated as porosity multiplied by soil depth (Fig. 3a), and plant available water (Fig. 4) is field capacity minus wilting point. Field capacity (soil water volume at −0.03 MPa) is the soil water volume below which drainage is negligible, and wilting point (soil water volume at −1.5 MPa) is the soil water volume below which actual evapotranspiration does not occur (Hillel, 1980). SSURGO data was not available for the Tijuana Basin; therefore available coarse soil property maps (Mexican National Institute of Statistic and Geography) were used to estimate porosity, field capacity, and wilting point. Soil depth was calculated by relying on the coarse maps and incorporating other information that is defined more finely, such as topographic description (Mexican National Institute of Statistic and Geography) and slope calculated from the 270-m digital elevation model (DEM).

Once available monthly water is calculated, water may exceed total soil storage and become runoff or it may be less than total soil storage but greater than field capacity and become recharge. Anything less than field capacity will be lost to actual evapotranspiration at the rate of PET for that month until it reaches wilting point. When soil water is less than total soil storage and greater than field capacity, soil water greater than field capacity equals potential recharge. If potential recharge is greater than bedrock permeability (K), then recharge = K and potential recharge that exceeds K becomes runoff, or else it will recharge at K until it reaches field capacity. Model calibration to partition excess water into recharge and runoff is done by comparing model results for runoff with measured surface-water flow and iteratively changing K until a reasonable match is achieved. This was done for 15 subbasins with varying amounts of impairment (Fig. 1, Table 1). The subbasins with the least impairments, those upstream of reservoirs, without major diversions or urban runoff, were considered for the calibrations.

Finally, basin discharge is calculated to more accurately reflect stream channel losses and gains between stream gages and to create surface-water flow recession and baseflow that can extend throughout the dry season. As described, BCM simulates recharge ($\mathrm{BCM_{rch}}$) and runoff ($\mathrm{BCM_{run}}$) for each 270-m^2 grid cell for each month (i). To compare them to gaged mean monthly surface-water flow, all grid cells upstream of the stream gage are summed for each month to create time series for $\mathrm{BCM_{run}}$ and $\mathrm{BCM_{rch}}$. To transform these results into a form that can be compared to the pattern and amount of gaged surface-water flow, the water balance is conceptualized as consisting of two units that are hydraulically connected through a shallow storage zone ($\mathrm{GW_{shallow}}_{(i)}$). The two units are the basin discharge ($\mathrm{Stream}_{(i)}$), and regional aquifer ($\mathrm{GW_{deep}}_{(i)}$). A set of empirical flow-routing equations defines storage in successive time steps (i) and performs partitioning (Fig. 5). $\mathrm{GW_{shallow}}_{(i)}$ is the computational method used to extend streamflow for time steps when $\mathrm{BCM_{run}}_{(i)}$ and $\mathrm{BCM_{rch}}_{(i)}$ are zero (e.g., during seasonal and annual dry periods). For time steps when $\mathrm{BCM_{run}}_{(i)}$ and $\mathrm{BCM_{rch}}_{(i)}$ are non-zero, the amounts are accumulated for the grid cells upstream of a stream gage. Initially the water in $\mathrm{GW_{shallow}}_{(i)}$ is evaluated as

$$\mathrm{GW_{shallow}}_{(i)} = \text{(1-Runscaler)} \cdot \mathrm{BCM_{run}}_{(i)} + \mathrm{BCM_{rch}}_{(i)} + \mathrm{GW_{stor}}_{(i-1)} \,. \tag{A1}$$

Runscaler is a coefficient (< 1) that is used to match peak flows, and (1-Runscaler) is the direct loss of peak flows to $\mathrm{GW_{shallow}}$. Carryover of groundwater storage from the previous time step ($\mathrm{GW_{stor}}_{(i-1)}$), is set by the parameter exp (< 1).

$$\mathrm{GW_{stor}}_{(i)} = (\mathrm{GW_{shallow}}_{(i-1)})^{\exp} \tag{A2}$$

The overland flow component is comprised of the direct runoff and baseflow. The direct runoff is calculated (Eq. A3) from $\mathrm{BCM_{run}}_{(i)}$ and the Runscaler (from Eq. A1), and the baseflow/recession component is partitioned from $\mathrm{GW_{shallow}}_{(i)}$ minus carryover to the next month ($\mathrm{GW_{stor}}_{(i)}$, see Eq. A2) using the parameter Rchscaler (< 1).

$$\mathrm{Runoff}_{(i)} = \mathrm{BCM_{run}}_{(i)} \cdot \text{Runscaler} + \mathrm{Baseflow}_{(i)} \tag{A3}$$

$$\mathrm{Baseflow}_{(i)} = (\mathrm{GW_{shallow}}_{(i)} - \mathrm{GW_{stor}}_{(i)}) \cdot \text{Rchscaler} \tag{A4}$$

To maintain mass balance, the carryover ($\mathrm{GW_{stor}}_{(i)}$) is subtracted from the $\mathrm{Baseflow}_{(i)}$. The sum of $\mathrm{Runoff}_{(i)}$ and $\mathrm{Baseflow}_{(i)}$ is the storage water partitioned to $\mathrm{Stream}_{(i)}$.

$$\mathrm{Stream}_{(i)} = \mathrm{Runoff}_{(i)} + \mathrm{Baseflow}_{(i)} \tag{A5}$$

$\mathrm{Stream}_{(i)}$ is the post-processed portion of the BCM water balance that is compared to the pattern and amount of gaged streamflow. The amount partitioned to the regional aquifer is the residual water in the shallow storage zone, minus carryover ($\mathrm{GW_{stor}}_{(i)}$) to the next month,

$$\mathrm{GW_{deep}}_{(i)} = \mathrm{GW_{shallow}}_{(i)} - \mathrm{GW_{stor}}_{(i)} - \mathrm{Baseflow}_{(i)}, \tag{A6}$$

which is equivalent to (1-Rchscaler) + $\mathrm{Baseflow}_{(i)}$. Together these equations represent the conceptual routing scheme illustrated in Fig. 5. It is not based on extensive system properties, nor is it a formal mass balance; however, it is an aggregate mass-balance check for all time steps in the water-balance period (Eq. A7).

$$\Sigma\mathrm{BCM_{run}} + \Sigma\mathrm{BCM_{rch}} - \Sigma\text{Discharge} - \Sigma\mathrm{GW_{deep}} = 0 \tag{A7}$$

The mass balance, aggregated for all time steps, is checked (see Eq. A7). In practice, Runscaler is estimated to visually match measured streamflow peaks, and exp is adjusted to preserve the mass balance described in Eq. (A7). The parameter Rchscaler is then used to match measured streamflow. Subsurface bulk bedrock permeability, which is initially assigned on the basis of geology, is also iteratively adjusted to improve the match between gaged streamflow and the basin discharge, $\mathrm{Stream}_{(i)}$, and the mass balance.

$\mathrm{BCM_{rch}}$ and $\mathrm{BCM_{run}}$ reflect natural hydrologic conditions and do not account for diversions, reservoir storage or releases, urban runoff, groundwater pumping, or other impairments, and therefore will not exactly match measured streamflow in impaired basins.

Appendix B

Episodic recharge in semiarid and arid environments

The conceptualization of recharge in the arid and semiarid southwest is complicated. The definition of climate regimes called arid was developed by United Nations Educational, Scientific, and Cultural Organization (UNESCO, 1979) on the basis of the ratio of mean annual precipitation to potential evapotranspiration. The San Diego region is classified as semiarid (Flint and Flint, 2007a), which means average annual precipitation is between 20 and 50 percent of potential evapotranspiration, suggesting little potential for recharge. However, recharge in a semiarid basin does not occur based on average annual conditions. In certain areas of a basin (in particular, the higher elevations), precipitation in some months can exceed potential evapotranspiration and soil storage, and net infiltration (defined as infiltration that reaches depths below which it can be removed by evapotranspiration processes) and/or runoff may occur, depending on the rate of rainfall or snowmelt, soil properties (including permeability, thickness, field capacity, and porosity), and bedrock permeability (Flint et al., 2001). For many basins, snow accumulated for several months provides enough moisture to exceed the soil storage capacity and exceed potential evapotranspiration for the month or months during which snowmelt occurs (Flint and Flint, 2007a). This leads to sporadic and sometimes spatially limited occurrences of net infiltration but can represent the majority of recharge in a basin. Net infiltration is the precursor to groundwater recharge that can occur months to decades after the net infiltration event and is dependent on the properties and thickness of the unsaturated zone.

On a global scale, Scanlon et al. (2006) determined that recharge in semiarid and arid regions throughout the world responds to climate variability. Average recharge rates estimated over large areas (40–374 000 km^2) range from 0.2 to 35 $mm\,yr^{-1}$, representing 0.1–5 % of long-term average annual precipitation. Extreme local variability in recharge, with rates up to $\sim 720\,m\,yr^{-1}$, results from focused recharge beneath ephemeral streams and lakes and preferential flow mostly in fractured systems. Interannual climate variability related to El Niño Southern Oscillation (ENSO) results in up to three times higher recharge in regions within the southwest United States during periods of frequent El Niños (1977–1998) relative to periods dominated by La Niñas (1941–1957).

The use of water balance approaches to estimate recharge in arid and semiarid environments has been disputed in the literature over the last two decades partially in response to Gee and Hillel (1988), who reported that the volumes of recharge in arid environments were too small to measure or estimate using anything other than approaches that integrated recharge over long time periods, such as lysimetry or chloride mass balance methods. Since then, major advances have been made in the understanding of how recharge occurs in arid and semiarid environments, as described above, and have been discussed and scrutinized by numerous authors (Lerner et al., 1998; Hendrickx and Walker, 1997; Zhang and Walker, 1998; Kinzelbach et al., 2002; Scanlon et al., 2002; Flint et al., 2002).

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