Hydrol. Earth Syst. Sci. Discuss., 9, 2717–2762, 2012
www.hydrol-earth-syst-sci-discuss.net/9/2717/2012/
doi:10.5194/hessd-9-2717-2012

Hydrology and Earth System Sciences Discussions

Water-balance and groundwater-flow estimation for an arid environment: San Diego region, California

L. E. Flint[1], A. L. Flint[1], B. J. Stolp[2], and W. R. Danskin[3]

[1]US Geological Survey, Sacramento, CA, USA
[2]US Geological Survey, Salt Lake City, UT, USA
[3]US Geological Survey, San Diego, CA, USA

Received: 10 February 2012 – Accepted: 17 February 2012 – Published: 1 March 2012

Correspondence to: L. E. Flint (lflint@usgs.gov)

Published by Copernicus Publications on behalf of the European Geosciences Union.

Abstract

The coastal-plain aquifer that underlies the San Diego City metropolitan area in southern California is a groundwater resource. The understanding of the region-wide water balance and the recharge of water from the high elevation mountains to the east needs to be improved to quantify the subsurface inflows to the coastal plain in order to develop the groundwater as a long term resource. This study is intended to enhance the conceptual understanding of the water balance and related recharge processes in this arid environment by developing a regional model of the San Diego region and all watersheds adjacent or draining to the coastal plain, including the Tijuana River basin. This model was used to quantify the various components of the water balance, including semi-quantitative estimates of subsurface groundwater flow to the coastal plain. Other approaches relying on independent data were used to test or constrain the scoping estimates of recharge and runoff, including a reconnaissance-level groundwater model of the San Diego River basin, one of three main rivers draining to the coastal plain. Estimates of subsurface flow delivered to the coastal plain from the river basins ranged from 12.3 to 28.8 million $m^3\,yr^{-1}$ from the San Diego River basin for the calibration period (1982–2009) to 48.8 million $m^3\,yr^{-1}$ from all major river basins for the entire coastal plain for the long-term period 1940–2009. This range of scoping estimates represents the impact of climatic variability and realistically bounds the likely groundwater availability, while falling well within the variable estimates of regional recharge. However, the scarcity of physical and hydrologic data in this region hinders the exercise to narrow the range and reduce the uncertainty.

1 Introduction

Estimating the water balance and groundwater flow for an arid environment requires multiple approaches to provide consensus because of the general lack of water and sparse hydrologic data. Similarly, the spatial and temporal distribution of recharge and

HESSD
9, 2717–2762, 2012

discharge processes becomes important due to highly variable topography, elevation, geology, soils, and episodic weather patterns. Historically, recharge estimates have relied on monthly water-balance models that incorporate evapotranspiration (Alley, 1984), inverse modeling (Sanford et al., 2001), or lysimetry and tracer tests (Gee and Hillel, 1988). Water balance modeling to assess both recharge and runoff has been done at the site scale (Flint et al., 2001; Ragab et al., 1997) and integrated with various measurements addressing different spatial scales (Flint et al., 2002). Water-balance modeling at a regional scale has been done by Hevesi et al. (2003), Flint et al. (2004), Kavvas et al. (2006), and Flint and Flint (2007). Combining water-balance results with surface-water and groundwater discharge data, and formal routing of water through the system, provides a method for evaluating regional-scale groundwater flow. Although budget estimation is possible, sparse hydrologic data, spatial and temporal variability in precipitation and evapotranspiration, accumulation of snow and subsequent snowmelt, and the partitioning of surface-water runoff and groundwater recharge create varying degrees of uncertainty.

This work is part of several parallel efforts to further the conceptual understanding of water resources for the San Diego region and estimate the subsurface groundwater flow to the coastal plain. 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. This paper focuses on the spatial distribution of recharge and runoff on the landscape, independent data to support the distribution, and a reconnaissance-level quantification of groundwater recharge that also relies on independent data. The objective is to provide first pass estimates of a water balance and associated uncertainties, supported by multiple approaches that rely on independent data, to highlight data and information gaps and the paths necessary to refine and constrain estimates of groundwater flow available to the coastal plain.

A water balance for the San Diego region was quantified using the Basin Characterization Model (BCM) (Flint and Flint, 2007; Micheli et al., 2012). The groundwater flow regime implied by the water balance was evaluated for the San Diego River basin. The San Diego River basin is 1 of 6 arid basins that are referred to as the San Diego region of southern California (Fig. 1). Precipitation is the source of water inflow to the region. Most precipitation occurs in the eastern mountains and is evaporated, transpired, or sublimated back into the atmosphere. The difference between precipitation and evapotranspiration/sublimation becomes excess water that is partitioned into surface-water runoff and groundwater recharge. Drainage basins drain the mountains in the east and flow west across the coastal plain to the Pacific Ocean (Fig. 1). The Tijuana River basin drains a large basin to the southeast of the coastal plain. Recharge originates from direct infiltration of precipitation and indirectly from surface runoff that seeps into the ground along ephemeral stream channels. Baseflow moves from the subsurface along the lower-elevation reaches of the drainage basins to estuaries or directly into the Pacific Ocean. Groundwater moves toward and discharges to streams and/or the coastal plain, with eventual discharge to the Pacific Ocean. The coastal plain underlies the San Diego City metropolitan area (Fig. 1) and is considered a groundwater resource. To develop groundwater from the coastal-plain aquifer, subsurface inflows need to be quantified. Thirteen multi-level monitoring wells show the coastal-plain as having a complex stratigraphic structure, and heterogeneous water-quality/water-level distributions.

Groundwater flow is estimated as the difference of basin gains, BCM-derived surface-water runoff (BCM_{run}) plus recharge (estimated as water that percolates below the root zone – BCM_{rch}), and basin losses, gaged/estimated streamflow. These estimates are extrapolated for the entire coastal plain, and are supported with several approaches, including data collection and groundwater modeling. The estimated groundwater flow is evaluated by numerically simulating groundwater flow through the San Diego River basin using MODFLOW (Harbaugh, 2005), including data independent of the BCM, and constraining it with realistic aquifer geometry and hydraulic properties. Steady-state conditions, based on median annual conditions for 1982–2009, are simulated for the San Diego River basin.

Study area

The San Diego region includes 6 basins that drain to the ocean across the coastal plain; San Dieguito Creek, Los Pensaquitos Creek, San Diego River, Sweetwater River, Otay River, and Tijuana River, and two coastal plain drainages (Fig. 1). These drainages and associated tributaries make up a drainage area of approximately 8000 km^2 that ranges in elevation from sea level at the coast to 3700 m along the eastern boundary. The climate is arid in the coastal plain and transitions to semi-arid 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 rapidly with distance inland, and decreases slightly along the coast from north to south (Elwany et al., 1998). Hydrologic 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) ~1300 (700–1600) mm yr^{-1}), little or no summer rainfall, and highly seasonal streamflows. Precipitation generally exceeds PET in the winter months, which is when recharge occurs. The region has the highest variability of streamflow in the United States (Pryde, 1976). Stream discharge 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.

2 Methods

Two modeling approaches were used to estimate basin recharge to the coastal plain. Methods of model development and calibration are described for both the BCM water-balance modeling of the San Diego region, and the MODFLOW groundwater simulation

HESSD
9, 2717–2762, 2012

for the San Diego River basin. Streamflow reconstruction and estimation of groundwater flow are also described.

2.1 Water balance

To determine the spatially distributed hydrologic processes, and resulting water balance for the San Diego region, the Basin Characterization Model (BCM; Flint and Flint, 2007a) was applied to the region, including the Tijuana River basin. The BCM is a regional water-balance model that has been applied to the state of California at a fine scale of 270-m grid cells, and calibrated to streamflow at 138 basins to assess historical hydrologic processes and impacts of climate change on both water availability and ecosystems (Flint 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.

The BCM mechanistically models the pathways of a basin's precipitation into evapotranspiration, infiltration into soils, runoff, or percolation below the root zone to recharge groundwater. Groundwater recharge (recharge) is also tied to runoff and the relationship between the two is driven by bedrock permeability (Fig. 2).

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 was 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 was aggregated to monthly and cloudiness corrections were made using cloudiness data from National Renewable Energy Laboratory (NREL). Modeled PET for the southwest United States was then calibrated to measured PET from California Irrigation Management Information System (CIMIS) and Arizona Meterological Network (AZMET) stations, and is shown for the San Diego/Tijuana region in Fig. 3a. It is clear from the map that the highest PET is on high slopes (Fig. 3b) with southern facing

HESSD
9, 2717–2762, 2012

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. 4) 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 in general. The months with precipitation are indicated as red points (November–April), but the months with the most recharge during springtime snowmelt, April and May, 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., 2004; 800-m (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. 2). 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 downscaled or interpolated to the 270-m grid resolution for model application following Flint and Flint (2012). For the San Diego region, we combined the climate surfaces and monthly PET with maps of elevation, bedrock permeability estimated on the basis of geology (Jennings, 1977; Fig. 3c), and soil-water storage from the SSURGO soil databases (Natural Resource Conservation Service; http://soils.usda.gov/survey/geography/ssurgo/). Total soil-water storage is calculated as porosity multiplied by soil depth (Fig. 3d). 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) (Fig. 2). 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

HESSD
9, 2717–2762, 2012
Water-balance and groundwater-flow estimation for an arid environment
L. E. Flint et al.

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 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, else it will recharge at K until field capacity. Model calibration to partition excess water into recharge and runoff is done by comparing model results for runoff with measured streamflow 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 streamgages and to create streamflow recession and baseflow that can extend throughout the dry season. As described, BCM simulates recharge (BCM_{rch}) and runoff (BCM_{run}) for each 270-m grid cell for each month (i). To compare them to gaged mean monthly streamflow, all gridcells upstream of the streamgage are summed for each month to create time series for BCM_{run} and BCM_{rch}. To transform these results into a form that can be compared to the pattern and amount of gaged streamflow, the water balance is conceptualized as consisting of two units that are hydraulically connected through a shallow storage zone ($GW_{shallow(i)}$). The two units are the basin discharge ($Stream_{(i)}$), and regional aquifer ($GW_{deep(i)}$). A set of empirical flow-routing equations defines storage in successive time-steps (i) and performs partitioning (Fig. 5). $GW_{shallow(i)}$ is the computational method used to extend streamflow for time-steps when $BCM_{run(i)}$ and $BCM_{rch(i)}$ are zero (e.g., during seasonal and annual dry periods). For time-steps when $BCM_{run(i)}$ and $BCM_{rch(i)}$ are non-zero, the amounts are accumulated for the grid cells upstream of a streamgage. Initially the water in $GW_{shallow(i)}$ is evaluated as

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

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

$$GW_{stor(i)} = \left(GW_{shallow(i-1)}\right)^{\exp}. \tag{2}$$

The overland flow component is comprised of the direct runoff and baseflow. The direct runoff is calculated (Eq. 3) from $BCM_{run(i)}$ and the Runscaler (from Eq. 1), and the baseflow/recession component is partitioned from $GW_{shallow(i)}$ minus carryover to the next month ($GW_{stor(i)}$, see Eq. 2) using the parameter Rchscaler (<1).

$$\text{Runoff}_{(i)} = BCM_{run(i)} \cdot \text{Runscaler} + \text{Baseflow}_{(i)} \tag{3}$$

$$\text{Baseflow}_{(i)} = \left(GW_{shallow(i)} - GW_{stor(i)}\right) \cdot \text{Rchscaler}. \tag{4}$$

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

$$\text{Stream}_{(i)} = \text{Runoff}_{(i)} + \text{Baseflow}_{(i)}. \tag{5}$$

$\text{Stream}_{(i)}$ 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 ($GW_{stor(i)}$) to the next month,

$$GW_{deep(i)} = GW_{shallow(i)} - GW_{stor(i)} - \text{Baseflow}_{(i)}, \tag{6}$$

which is equivalent to (1-Rchscaler) + $\text{Baseflow}_{(i)}$. 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. 7).

$$\sum \text{BCM}_{\text{run}} + \sum \text{BCM}_{\text{rch}} - \sum \text{Discharge} - \sum \text{GW}_{\text{deep}} = 0. \quad (7)$$

The mass-balance, aggregated for all time steps, is checked (see Eq. 7). In practice, Runscaler is estimated to visually match measured streamflow peaks, and exp is adjusted to preserve the mass balance described in Eq. (7). The parameter *Rchscaler* is then used to match measured streamflow. 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, $\text{Stream}_{(i)}$, and the mass balance.

BCM_{rch} and 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. Figure 5 illustrates the conceptual model of the contribution of runoff and recharge to the groundwater system, the partitioning into Stream and GW_{deep}, and resulting streamflow discharge and groundwater recharge.

2.2 Streamflow

Gaged streamflow records exist for 15 locations within the San Diego region (Fig. 1, Table 1). Streamflow at all the locations are impaired (altered) to some degree by reservoirs, urban runoff, imported water, and diversions. For the San Diego River basin, impaired streamflows 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 streamflow 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 the possibility of error greater than 8 % of the reported flow (Water-Data Report,

HESSD
9, 2717–2762, 2012

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 (Water-Data Report, 2010). The gage data are used to calculate mean streamflow for the day and month. The mean best describes actual streamflow for a specific day or month. For this analysis, the median of the monthly mean was used to calculate annual and monthly flows. The median statistic was chosen to minimize the influence of extremely low and high streamflow specific to the 1982–2009 time period.

Cuyamaca, El Capitan, and San Vicente Reservoirs regulate streamflow at the Mast and FV gages (locations noted on Fig. 1). To reconstruct, the amounts of water entering and leaving El Capitan and San Vicente Reservoirs were examined (Cuyamaca regulation is aggregated with San Vicente). The amounts are recorded on a monthly basis by the San Diego City Water Department and account for water leaving the reservoirs via evaporation, seepage, and export; entering water includes import through aqueducts, precipitation on the reservoir surface, and runoff (J. Pasek, oral and written communications, City of San Diego, July 2011). Water exported from the reservoirs is primarily for municipal use and agriculture, negligible return flow is assumed. The monthly change in stored reservoir water is considered unimpaired streamflow at the location of the reservoir. Those amounts were added to the measured streamflow at both Mast and FV to estimate unregulated flow. Summation implies that the additional water does not affect the overall magnitude of surface-water/groundwater interaction between reservoirs and gages.

2.3 Groundwater flow

The components of groundwater flow in the San Diego River basin are (1) total inflowing water, (2) surface-water flowing out to the Pacific Ocean, and (3) the groundwater flow to the aquifer. Total water entering the basin is considered to be BCM_{run} plus BCM_{rch}; reconstructed streamflow at FV is assumed a fair measure of surface-water flow to the Pacific Ocean (Fig. 1). Streamflow at FV integrates upstream partitioning, loss/gain, and the carryover empirically described by Eqs. (1) to (6). Imposing these

HESSD
9, 2717–2762, 2012

assumptions, the difference between total water entering the basin and reconstructed streamflow at FV is considered groundwater flow to the aquifer, and a first-pass estimate of subsurface flow through the coastal plain bordering the San Diego River basin. The ratio of subsurface flow to BCM-derived total water entering the San Diego River Basin is used to estimate subsurface flow for the coastal areas adjacent to the Sweetwater and Otay River basins.

To further develop the conceptual model of groundwater recharge and flow in the San Diego region, a steady-state reconnaissance-level numerical flow simulation was constructed of the San Diego River basin using MODFLOW (Harbaugh, 2005). This further develops the conceptual model of groundwater recharge in the San Diego region by providing independent evidence to support the estimates of groundwater flow to the coastal plain. By using the BCM estimates of recharge and runoff as boundary conditions for the model, calibrating it with reasonable aquifer properties, and then matching it to measured water levels and streamflow, results offer additional bounds to the estimates and reduce uncertainty. An additional objective was to compare the physically-based routing scheme used in the MODFLOW model to the empirical routing used to calibrate the BCM, and lend further support for the BCM water-balance estimates.

The model domain was 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 by 500-m grid cells. Vertically, the domain consists of 3 layers. 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 the altitude datum that is used to establish layer 2 and 3 top/bottom altitudes. Contact altitudes come from the San Diego River System Conceptual Groundwater Management Plan (CH2MHILL, 2003).

The Drain (DRN) Package (Harbaugh, 2005, p. 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 fresh- and sea-water hydrostatic pressures at the midpoint altitude of each layer; DRN hydraulic conductance is set equal to the simulated coastal-plain sediments.

The San Diego River and two tributaries (San Vicente and Boulder Creeks) were simulated using the Streamflow Routing (SFR) Package (Niswonger and Prudic, 2003). Eight segments were used to represent the stream network (Fig. 7a, Table 2). Stream segment altitudes were determined from the 10-m digital elevation model of the basin. The RUNOFF term (Niswonger and Prudic, 2003, p. 24) for each of the segments was set equal to BCM_{run} for the portion of the drainage bisected by the segment. The Recharge (RCH) Package (Harbaugh, 2005, p. 8–37) was used to simulate groundwater recharge into the model domain across the uppermost layer. Recharge corresponds to the spatial distribution and amount of BCM_{rch} for 1982–2009. MODFLOW uses BCM_{rch} and BCM_{run} as boundary conditions and internally partitions them to determine aquifer recharge, independent of the empirical flow routing using Eqs. (1)–(6).

Horizontal and vertical hydraulic conductivities were lumped and put into zones 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 lumped 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. 3c and 7b–c). For zone-1 and zone-4, the ratio of horizontal to vertical hydraulic conductivity was set at 1.0. For zone-2 and zone-3 the ratio was fixed at 10.0. Observations and prior knowledge used to calibrate model parameters are (1) reconstructed streamflow estimates at Mast and FV, (2) water-level elevations in the Mission Valley and Santee/El Monte alluvial sub-basins, and (3) the ephemeral nature of streamflow along the upper reaches of the San Diego River, San Vicente Creek, and Boulder Creek. Groundwater withdrawal at wells, return flows from irrigation and lawn watering, and waste water treatment plant effluent were not simulated.

3 Results

Calibration results for the BCM and the MODFLOW model are presented. Climate and water-balance results are discussed for the BCM and simulation results are discussed for the MODFLOW model. Comparisons between models are made to substantiate the conceptual model of groundwater recharge and flow in the San Diego region and specifically for the San Diego River basin.

3.1 Climate

Average annual precipitation over 4 of the 6 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. 8, 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 calibration period, 1982–2009, indicated on Fig. 8, has the same average precipitation as the long-term average for 1940–2009, 389 mm yr^{-1}, but the variability about the mean is 11 to 12 % higher for the calibration period, which has more years with low precipitation, and more years with high precipitation.

HESSD
9, 2717–2762, 2012

3.2 Water balance

The spatial distribution of BCM_{run} and BCM_{rch} averaged for 1940–2009 is shown in Fig. 9. Very little 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 basins to the north, where high elevation mountains in the east receive somewhat more precipitation than those in the Tijuana basin (Fig. 9). 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. 3d). Thick soils hold moisture in the profile making it available for evapotranspiration and lost to recharge or runoff.

To illustrate the dominant contributions to the groundwater system, and the major physical features controlling the spatial distribution of recharge, total average annual recharge is shown for all river basins, geologic units, and regions with differing groundwater source elevations (Tables 3, 4, and 5). These tables do not include the variable contribution of runoff to groundwater recharge. Most of the volume of 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 $\mathrm{mm\,yr^{-1}}$, most of the recharge occurs in the San Diego River, Sweetwater River, San Dieguito River, and Otay River basins (Table 3). As a result, when 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 occurs in the San Dieguito River basin, this river drains to the ocean and does not directly intersect the coastal plain. The three river basins that contribute recharge to the coastal plain are the San Diego River, Sweetwater River, and Otay River basins, and have a long term (1940–2009) average volume of 91.4 million $\mathrm{m^3\,yr^{-1}}$, and a recent (2000–2009) average volume of 29.7 million $\mathrm{m^3\,yr^{-1}}$. Recharge is related to precipitation by a power function (Fig. 10), whereby recharge increases exponentially with increase in precipitation (Flint and Flint, 2007). The recent decade with approximately 65 % of long-term precipitation yielded about one-third the amount of recharge.

HESSD
9, 2717–2762, 2012
Water-balance and groundwater-flow estimation for an arid environment
L. E. Flint et al.

Within the San Diego basin, a large proportion, at least an order of magnitude more, of the recharge is located in the region defined by hard rock geology, and dominated by granites (Fig. 3c; Table 4). This suggests 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. To further confirm this observation, 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 to assess which elevations contributed the most to the recharge (Fig. 3c). The recharge was calculated for each of the three contributing river basins (Table 5). The most recharge occurs in the −50 ‰ δ Deuterium zone, which coincides with the high elevation, hard rock zone.

Bedrock permeability, which is the BCM parameter that controls partitioning of excess water between recharge and runoff, is provided in Table 4 for each geologic unit. Bedrock permeability was estimated iteratively by comparing BCM results to gaged streamflow at 15 locations (Table 1), using the empirical flow-routing equations outlined in the Methods section. Gaged streamflow 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 (Eq. 5) for the Mast gage is shown in Fig. 11 for the period of record. The r^2 calculated from the gaged streamflow at Mast and modeled basin discharge is 0.83, and the Nash-Sutcliffe efficiency statistic (Nash and Sutcliffe, 1986), calculated as the mean squared error divided by the variance for the period of record, is 0.86, indicating a good fit.

3.3 Streamflow in the San Diego River basin

The measured median monthly and annual streamflow for 1982–2009 at Mast and FV and the reconstructed streamflow at El Capitan and San Vicente Dams are in Table 6. Summing measured and reconstructed streamflow, the median annual streamflow at Mast is estimated at 20.5 million $m^3\,yr^{-1}$ and 28.8 million $m^3\,yr^{-1}$ at FV. The 1982–2009

HESSD
9, 2717–2762, 2012

time period, spans across the wettest and driest decades of available precipitation data for the San Diego region (Fig. 8).

3.4 Groundwater flow in the San Diego River basin

The mean annual BCM-derived water entering the San Diego River basin ($BCM_{rch} + BCM_{run}$) during 1982–2009 is 47.6 million $m^3\,yr^{-1}$, and is considered total water entering the basin (Table 7). Reconstructed streamflow at FV is estimated at 28.8 million $m^3\,yr^{-1}$ at FV, and assumed a fair measure of surface-water flow to the Pacific Ocean. Subtracting reconstructed streamflow at FV from BCM input, groundwater flow to the aquifer is 18.8 million $m^3\,yr^{-1}$ (Table 7). This is the first-pass estimate of subsurface flow through the coastal plain adjacent to the San Diego River basin. The groundwater flow estimate defines a 60/40 partitioning ratio between runoff and recharge. Sixty percent of the water entering the basin exits as streamflow; forty percent exits as subsurface or groundwater flow.

The mean annual BCM-derived water entering the San Diego River basin is used as boundary conditions to numerically simulate groundwater flow using MODFLOW. As stated previously, the primary purpose of the simulation was to further develop the conceptual model of groundwater recharge and provide independent evidence to support groundwater flow estimates for the coastal plain. The mean annual BCM_{rch} for the San Diego River Basin during 1982–2009 (33.1 million $m^3\,yr^{-1}$) is input to the model domain using the RCH boundary. The mean annual BCM_{run} during 1982–2009 is 14.5 million $m^3\,yr^{-1}$ and was subdivided, as per the BCM_{run} distribution, and applied to eight SFR 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}), streambed hydraulic conductivity (Fig. 7, K_{h3}), and portion of BCM_{rch} applied to the top model layer were adjusted. Simulation results were compared to reconstructed streamflow at Mast and FV, eight water-levels in the

Mission Valley and Santee/El Monte alluvial sub-basins, and the estimated groundwater flow to the aquifer.

An initial model calibration (model alternative 1) focused on minimizing the misfit between observed and simulated conditions without enforcing constraints on the subset of parameters that are allowed to vary. In general, K_{h1} controls water levels in the Santee/El Monte alluvial sub-basin, K_{h4} controls simulated streamflow and subsurface flow at Mast, and K_{h2} controls the partitioning of water to streamflow at FV and subsurface flow through the coastal plain to the Pacific Ocean. The parameter values resulting in the lowest sum of squared weighted residuals are listed in Table 8 under Model Alternative 1. For alternative 1, subsurface flow through the coastal plain is simulated at 18.0 million $m^3\,yr^{-1}$ (in comparison to 18.8 calculated by adding BCM_{rch} and BCM_{run} and subtracting reconstructed flows). Simulated water levels at the Santee/El Monte and Mission Valley alluvial sub-basins are an average of 3.7-m above observed water levels.

For comparison, model alternative 2 was calibrated within a constrained solution space defined by (1) $K_{h1} <= 1.0\,m\,day^{-1}$, (2) $K_{h4} < K_{h1}$, and (3) $K_{h2} > K_{h1}$. To meet the K_{h1} constraint BCM_{rch} was reduced by 30 % (Table 8). The 30 % reduction results in a similar reduction in subsurface flow through the coastal plain of 12.3 million $m^3\,yr^{-1}$ (5.7 million $m^3\,yr^{-1}$ less than alternative 1; Table 8). Simulated water levels at the Santee/El Monte and Mission Valley alluvial sub-basins are an average of 9.1-m above observed water levels. The water-level match can be improved by further reduction of BCM_{rch}, but that also results in less simulated streamflow, and increased misfit to reconstructed streamflow at FV. Model alternative 2 represents a compromise between reducing K_{h1}, not simulating excessively high water levels, and maintaining streamflow.

A secondary model objective was to compare a physically-based routing scheme to the empirical process used by the BCM to post-process the grid-based BCM_{rch} and BCM_{run} into streamflow, baseflow, and subsurface groundwater flow. Average annual basin discharge at FV (Stream), using Eq. (1) to (7), is 34.6 million $m^3\,yr^{-1}$, and is about 15 % higher model alternative 1 simulated streamflow at FV (Table 7). The annual

HESSD
9, 2717–2762, 2012

average GW_{deep} at FV is 13.0 million $m^3\,yr^{-1}$, which is 28 % lower than the simulated subsurface flow of 18.0 million $m^3\,yr^{-1}$. Comparisons with model alternative 2, where BCM_{rch} was scaled by 0.7, were not made because the empirical process assumes all BCM_{rch} enters the basin.

In order to provide an estimate of subsurface groundwater flow in the coastal plain for the entire San Diego region it is necessary to estimate contributions from the Sweetwater and Otay River basins. This is done on the basis of the partitioning ratio of 60/40 developed from the BCM-calculated unimpaired inflows (as $BCM_{rch} + BCM_{run}$) to the San Diego River basin and reconstructed streamflows. This assumes the physical and hydrologic processes that govern the relative proportion of recharge and runoff in the San Diego River basin are the same as in the Sweetwater and Otay River basins. Long-term average estimates for 1940–2009 estimates were used to estimate the volume of total input to the basins that result in subsurface groundwater flow (Table 9). The value of subsurface groundwater flow to the coastal plain from the San Diego River basin is 28.3 million $m^3\,yr^{-1}$ for the long term average, which is higher than the 18.8 million $m^3\,yr^{-1}$ calculated from the 1982–2009 model calibration period. This is because of the non-linear relation of recharge to precipitation, which is shown for the three river basins for 1940–2009 using the average BCM_{rch} for each basin (Fig. 10). The average precipitation for 1982–2009 is the same as 1940–2009, but the variability is higher in the last 28 years and there are more average precipitation years in the first half of the record. As recharge is limited by the ability of the soil to store water, and high precipitation results in more runoff rather than more recharge, if the variability of precipitation increases, there will be a reduction in the average recharge and an increase in the average runoff. For the 1940–2009 time period, the total estimate of subsurface groundwater flow delivered from these three river basins to the coastal plain is 46.6 million $m^3\,yr^{-1}$ (37 800 acre-feet yr^{-1}; Table 9). 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. 10.

HESSD
9, 2717–2762, 2012

4 Discussion

The average climatic conditions during the 1982–2009 calibration period were the same as the 1940–2009 long-term average, 389 mm yr^{-1}, with the largest variability occurring during the 1982–2009 period. The decadal variations ranged from the highest to the lowest of the 70 years within the last two decades. The seasonal trends in climate did not change significantly over the long-term, but precipitation declined approximately 0.35 mm yr^{-1}, maximum and minimum monthly air temperature increased 1.1 °C and 1.6 °C respectively, driving a slight increase in potential evapotranspiration of 7 mm yr^{-1} over the 70-yr period.

The BCM water-balance results are generally well constrained on the basis of precipitation and PET data, which are used to calculate excess water, and a mechanistic process-based algebraic calculation to partition excess water into recharge and runoff. Uncertainties are associated with the spatially distributed geology and soils data, and the calibration to streamflow data with unknown impairments. In a region with few unimpaired streamflow constraints, calibrated bedrock permeabilities are likely not unique. This creates error in the partitioning of excess water into recharge and runoff. In addition, almost all water-balance schemes assume there is no subsurface flow across topographic basin boundaries. Despite these limitations, the physically-based and spatially-distributed data approach used in the BCM is a reasonable estimate of the relative magnitude of recharge and runoff across the landscape.

To numerically simulate the amount of groundwater flow defined by the residual between BCM water-balance and reconstructed streamflow at FV, a hydraulic conductivity of 8.0 m day^{-1} was assigned to the crystalline rock in the eastern 2/3 of the San Diego River basin (Fig. 6b and c, zone-1). Using this value for hydraulic conductivity, simulated streamflow at Mast and water levels in the Santee/El Monte alluvial sub-basin are a reasonable match to reconstructed streamflow and observed water levels. However, a hydraulic conductivity of 8.0 m day^{-1} is typical of clean- to silty-sand and an order of magnitude above the typical range for igneous and metamorphic rock (Fitts,

HESSD
9, 2717–2762, 2012

2002, Table 3.1). The abnormally high zone-1hydraulic conductivity needed to match observed conditions indicates a potential misunderstanding of aquifer boundaries and recharge. Reduced recharge (model alternative 2) allows for a decrease in zone-1 conductivity (1.0 m day^{-1}), but still at the upper end of typically observed values for crystalline rock. Weathering and fracturing can enhance conductivity but likely not to the degree and spatial extent implied by flow model calibration. A reduction in recharge implies that (1) the treatment of topographic divides as no-flow boundaries may be an oversimplification, (2) some water-balance process represented in BCM is not properly parameterized, and/or (3) the generalized geology incorporated into the model (Fig. 8) does not adequately represent the "true" aquifer geometry. Specific uncertainties include the potential for groundwater flow beneath topographic divides, and a negative bias to the potential evapotranspiration estimates used in the BCM water balance.

As discussed, higher than typical hydraulic conductivities are needed to simulate subsurface flow in the coastal plain that is 40 % of the total BCM-estimated water entering the San Diego River basin. Model alternative 2, which utilizes lower hydraulic conductivity and recharge, poses the potential of less subsurface flow in the coastal plain. The range of sub-surface values derived from the residual between the BCM water-balance and reconstructed streamflow, and numerical calibration within the constrained solution space suggests that 18.8 million m^3 yr^{-1} might be an upper limit for subsurface flow during 1982–2009. The BCM-derived subsurface flow could be as high as 28.3 million m^3 yr^{-1}. Without subjectively weighting (favoring) streamflow data, BCM water balance, or groundwater-flow simulation, subsurface flow is assumed to range from 13.0 to 28.3 million m^3 yr^{-1}. Extrapolating the long term values to the entire coastal plain results in an estimated subsurface flow of 48.8 million m^3 yr^{-1}.

HESSD
9, 2717–2762, 2012

5 Summary and conclusions

The long term average recharge to the groundwater system calculated in the San Diego region was 1.5 % of the long term precipitation, which is within the range of arid and semi-arid regions throughout the world (0.1–5 %; Scanlon et al., 2006). Recharge is calculated as 5.8 mm yr^{-1} for 1940–2009 for the average precipitation rate of 389 mm yr^{-1} in the San Diego region. In studies across the southwest US (Scanlon et al., 2006), the average recharge rate was 11.2 mm yr^{-1}, ranging from 0–1612 mm yr^{-1}, which corresponded to an average precipitation of 301 mm yr^{-1}.

The geographical disparity between dominant recharge zones in this region estimated by the BCM, and coastal groundwater withdrawals requires additional data for hydrogeological resolution and subsurface transmissive properties across large areas. These uncertainties confound the large scale estimates of groundwater flow from the high elevation hard rock zones to the coastal plain and are needed to properly constrain simulations of westward flow.

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

- The Tijuana River basin is likely not a significant source of groundwater to the San Diego coastal plain.

- Water-balance and groundwater-flow estimates for other regions with similar climatic and geologic settings indicate partitioning ratios that range from 60/40 to 80/20. This suggests that the probable range of 12.3 to 28.3 million m^3 yr^{-1} for the San Diego River basin likely has a positive bias. The most probable subsurface flow should be considered less than the mean value of 20.4 million m^3 yr^{-1}. The long term value of 48.8 million m^3 yr^{-1} for the entire coastal plain, corresponding to 5.8 mm yr^{-1} recharge for the San Diego region is well within the regional estimates for arid and semi-arid regions, and corresponds to 1.5 % of precipitation, also within the regional estimates of 0.1 to 5 %.

HESSD
9, 2717–2762, 2012

– As a result of the non-linear relation of recharge to precipitation, changes in climate that either reduce the average precipitation, or result in more frequent years with less than 200 mm are likely to reduce recharge within the coastal plain of the San Diego region.

Several enhancements to the approaches described here could be applied to reduce the uncertainty of the estimates. Although it is unlikely that that errors in the calculations of potential evapotranspiration contribute significantly to the quantification of the water balance in this region, there are ways to improve estimates by using remote sensing and energy balance techniques to better quantifying evapotranspiration parameters used in the BCM water-balance modeling. Better defining the surface-water/groundwater interactions between Mast and FV gages will help constrain subsurface flow estimates in zone-4 of the groundwater flow model. Incorporation of available aquifer test information can help to quantify realistic ranges for the hydraulic conductivities of the crystalline and meta-volcanic rocks. Better representation of the coastal plain sediments in terms of (1) incorporating estimates of the sea-water/freshwater interface geometry, (2) integrating the 3-dimensional structure of the coastal plain sediments into the MODFLOW model design, and (3) designing and implementing a 7-day (4-log cycle) aquifer test with dedicated observation wells, will significantly improve the numerical simulation of river basin and regional groundwater flow. Both the BCM and the reconnaissance-level groundwater flow models should be used to assess the statistical significance of the various data-collection options in terms of better defining system parameters.

Acknowledgements. The authors would like to acknowledge the funding entities, the City of San Diego and Sweetwater Authority, for their support of this study. We would also like to thank Don Poole, US Geological Survey, for providing a timely and helpful review, and two anonymous reviewers. All reviews provided comments and suggestions to greatly improve the manuscript.

HESSD
9, 2717–2762, 2012

References

Alley, W. M.: On the treatment of evapotranspiration soil moisture accounting, and aquifer recharge in monthly water balance models, Water Resour. Res., 20, 1137–1149, 1984.

Anderson, E. A.: A point energy and mass balance model of a snow cover, Technical Report NWS 19, US National Oceanographic and Atmospheric Administration (NOAA), Silver Spring, MD., p.150, 1976.

CH2MHILL: San Diego River system conceptual groundwater management plan, Prepared for the City of San Diego, May 2003.

Daly, C., Gibson, W. P., Doggett, M., Smith, J., and Taylor, G.: Up-to-date monthly climate maps for the conterminous United States, Proc., 14th AMS Conf. on Applied Climatology, 84th AMS Annual Meeting Combined Preprints, 13–16 January 2004, Paper P5.1, Amer. Meteorological Soc., Seattle, WA, http://www.prism.oregonstate.edu/pub/prism/docs/appclim04-uptodate_monthly_climate_maps-daly.pdfCD-ROM (last access: March 2011), 2004.

Elwany, M. H. S., Flick, R. E., and Aijaz, S.: Opening and closure of a marginal southern California lagoon inlet, Estuaries, 21, 246–254, 1998.

Fitts, C. R.: Groundwater science, Academic Press, San Diego, 2002.

Flint, A. L. and Childs, S. W.: Calculation of solar radiation in mountainous terrain, J. Agr. Forest Meteorol., 40, 233–249, 1987.

Flint, A. L. and Childs, S. W.: Modification of the Priestley-Taylor equation to estimate evapotranspiration for soil water limited conditions, J. Agr. Forest Meteorol., 56, 247–260, 1991.

Flint, A. L. and Flint, L. E.: Application of the Basin Characterization Model to estimate in-place recharge and runoff potential in the Basin and Range carbonate-rock aquifer system, White Pine County, Nevada, and adjacent areas in Nevada and Utah, US Geol. Surv. Scientific Invest. Rep. 2007–5099, p.18, 2007.

Flint, A. L., Flint, L. E., Kwicklis, E. M., Bodvarsson, G. S., and Fabryka-Martin, J. T.: Hydrology of Yucca Mountain, Rev. Geophys., 39, 447–470, 2001.

Flint, A. L., Flint, L. E., Kwicklis, E. M., Fabryka-Martin, J. T., and Bodvarsson, G. S.: Estimating recharge at Yucca Mountain, Nevada, USA: Comparison of methods, Hydrogeol. J., 10, 180–204, 2002.

Flint, A. L., Flint, L. E., Hevesi, J. A., and Blainey, J. M.: Fundamental concepts of recharge in the Desert Southwest: a regional modeling perspective, in: Groundwater Recharge in a

Desert Environment: The Southwestern United States, edited by: Hogan, J. F., Phillips, F. M., and Scanlon, B. R., Water Science and Applications Series, vol. 9, Amer. Geophys. Union, Washington, D.C., 159–184, 2004.
Flint, L. E. and Flint, A. L.: Regional analysis of ground-water recharge, in: Ground-water recharge in the arid and semiarid southwestern United States, edited by: Stonestrom, D. A., Constantz, J., Ferré, T. P. A., and Leake, S. A., US Geol. Surv. Prof. Paper 1703, 29–59, 2007a.
Flint, L. E. and Flint, A. L.: Estimation of hourly stream temperatures in unmeasured tributaries to the lower Klamath River, California, J. Environ. Qual., 37, 57–68, 2007b.
Flint, L. E. and Flint A. L.: Downscaling future climate scenarios to fine scales for hydrologic and ecologic modeling and analysis, Ecol. Process., 1, 1–15, 2012.
Flint, L. E., Flint, A. L., Thorne, J. H., and Boynton, R.: Fine-scale hydrological modeling for regional applications: using watershed calibrations to assess model performance for landscape projections, in review, 2012.
Gee, G. W. and Hillel, D.: Groundwater recharge in arid regions: review and critique of estimation methods, Hydrol. Process., 2, 255–266, 1988.
Glockhoff, C.: Mapping the San Diego underground, San Diego Association of Geologists quarterly meeting, El Cajon, CA., 18 May 2011.
Harbaugh, A. W.: MODFLOW-2005: The US Geological Survey modular ground-water model – the Ground-Water Flow Process, US Geolog. Surv. Techniques and Methods 6-A16, variously p., 2005.
Hevesi, J. A., Flint, A. L., and Flint, L. E.: Simulation of net infiltration and potential recharge using a distributed-parameter watershed model of the Death Valley region, Nevada and California, US Geol. Surv. Water Resour. Invest. Report 03–4090, p.161, 2003.
Hillel, D.: Fundamentals of soil physics, Academic Press, New York, 1980.
IAEA: Isotope based assessment of groundwater renewal in water scarce regions, IAEA Tec Doc 1246, IAEA, Vienna, p.273, 2001.
Jennings, C. W.: Geologic map of California: California Division of Mines and Geology Geologic Data Map Number 2, scale 1:750,000, Sacramento, CA, 1977.
Kavvas, M. L., Yoon, J., Chen, Z. Q., Liang, L., Dogrul, E. C., Ohara, N., Aksoy, H., Anderson, M. L., Reuter, J., and Hackley, S.: Watershed environmental hydrology model: environmental module and its application to a California watershed, J. Hydrol. Eng.-ASCE, 11, 261–273, 2006.

Micheli, E., Flint, L. E., Flint, A. L., Weiss, S. B., and Kennedy, M.: Downscaling future climate projections to the watershed scale: A North San Francisco Bay Estuary case study, San Francisco Estuary and Watershed Science, accepted, 2012.
Milliman, J. D., Farnsworth, K. L., Jones, P. D., Xu, K. H., and Smith, L. C.: Climatic and anthropogenic factors affecting river discharge to the global ocean, 1951–2000, Global Planet. Change, 62, 187–194, 2008.
Nash, J. E. and Sutcliffe, J. V.: River flow forecasting through conceptual models, I, A discussion of principles, J. Hydrol., 10, 282–290, 1970.
Niswonger, R. G. and Prudic, D. E.: Documentation of the Streamflow-routing (SFR2) Package to include unsaturated flow beneath streams-A modification to SFR1: US Geological Survey Techniques and Methods 6-A13, p.50, 2003.
Pryde, P.: Chapter 8: Water supply for the county, edited by: Pryde, Pr. R., in: San Diego, an introduction to the region, Kendall/Hunt, Dubuque, IA, 103-120, 1976.
Ragab, R., Finch, J., and Harding, R.: Estimation of groundwater recharge to chalk and sandstone aquifers using simple soil models, J. Hydrol., 190, 19–41, 1997.
Sanford, W. E., Plummer, L. N., McAda, D. P., Bexfield, L. M., and Anderholm, S. K.: Estimation of hydrologic parameters for the groundwater model of the Middle Rio Grande Basin using carbon-14 and water-level data, in: US Geological Survey Middle Rio Grande Basin Study: Proceedings of the Fourth Annual Workshop, Albuquerque, NM, 15–16 February 2000, edited by: Cole, J. C., US Geol. Surv. Open-File Rep. 00-488, 4–6, 2001.
Scanlon, B. R., Keese, K. E., Flint, A. L., Flint, L. E., Gaye, C. B., Edmunds, W. M., and Simmers, I.: Global synthesis of groundwater recharge in semiarid and arid regions, Hydrol. Process., 20, 3335–3370, 2006.
US Geological Survey: Water-Data Report 2010, 11023000 San Diego River at Fashion Valley, at San Diego, CA, http://wdr.water.usgs.gov/wy2010/pdfs/11023000.2010.pdf (last access: June 2011), 2010.
Williams, A. E. and Rodoni, D. P.: Regional isotope effects and application to hydrologic investigations in southwestern California, Water Resour. Res., 33, 1721–1729, 1997.

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

Streamgage 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	11026000	46.5	1956–1978	granite, mixed granite
San Diego R at Mast Rd near Santee CA	11022480	150.7	1982–2008	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 bel Poway Creek near Poway	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	granite
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

HESSD
9, 2717–2762, 2012

Water-balance and groundwater-flow estimation for an arid environment

L. E. Flint et al.

Table 2. Streamflow estimated from Basin Characterization Model, BCM_{run}, distributed among stream segments 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

HESSD

9, 2717–2762, 2012

Water-balance and groundwater-flow estimation for an arid environment

L. E. Flint et al.

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	Average annual recharge			
	(km^2)	(million $m^3\,yr^{-1}$)		($mm\,yr^{-1}$)	
		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

HESSD
9, 2717–2762, 2012

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) (million m^3 yr^{-1})	Mean recharge (2000–2009) (million 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

HESSD
9, 2717–2762, 2012

HESSD
9, 2717–2762, 2012

Water-balance and groundwater-flow estimation for an arid environment

L. E. Flint et al.

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 ‰.

River basin	Average annual recharge (million $m^3\ yr^{-1}$)							
	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

Discussion Paper | Discussion Paper | Discussion Paper | Discussion Paper |

HESSD

9, 2717–2762, 2012

Water-balance and groundwater-flow estimation for an arid environment

L. E. Flint et al.

Table 7. The mean annual Basin Characterization Model-derived water entering the San Diego River basin, MODFLOW-simulated streamflow in the San Diego River at Fashion Valley and subsurface flow in the coastal plain adjacent to the San Diego River basin, and Basin Characterization Model empirically routed streamflow and deep groundwater flow, for 1982–2009.

Method	Estimated flow (million $m^3\,yr^{-1}$)
Groundwater flow model (alternative 1)	
Runoff + Recharge, (equivalent to BCM_{rch} + BCM_{run})	47.6
Streamflow at Fashion Valley	29.5
Subsurface flow in coastal plain	18.0
Basin Characterization Model	
Basin discharge (Stream)	34.6
GW_{deep}	13.0

HESSD
9, 2717–2762, 2012

Water-balance and groundwater-flow estimation for an arid environment

L. E. Flint et al.

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

	Model Alternative 1	Model Alternative 2
Horizontal hydraulic conductivity (m day^{-1})		
K_{h1}	8.0	1.0
K_{h2}	2.7	2.0
K_{h3}	10.0	20.0
K_{h4}	0.2	0.3
BCM_{rch} (million m^3 yr^{-1})	33.1	23.2
Subsurface groundwater flow (million m^3 yr^{-1})	18.0	12.3
Sum of squares weighted residual	293	431

HESSD
9, 2717–2762, 2012
Water-balance and groundwater-flow estimation for an arid environment
L. E. Flint et al.

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 % of total, San Diego region, California.

River basin	BCM_{rch}	BCM_{run}	Subsurface groundwater flow
	(million $m^3 yr^{-1}$)		
San Diego River	53.9	16.9	28.3
Sweetwater River	25.3	10.6	14.4
Otay River	12.2	3.0	6.1
Total	91.4	30.6	48.8

HESSD
9, 2717–2762, 2012

Water-balance and groundwater-flow estimation for an arid environment

L. E. Flint et al.

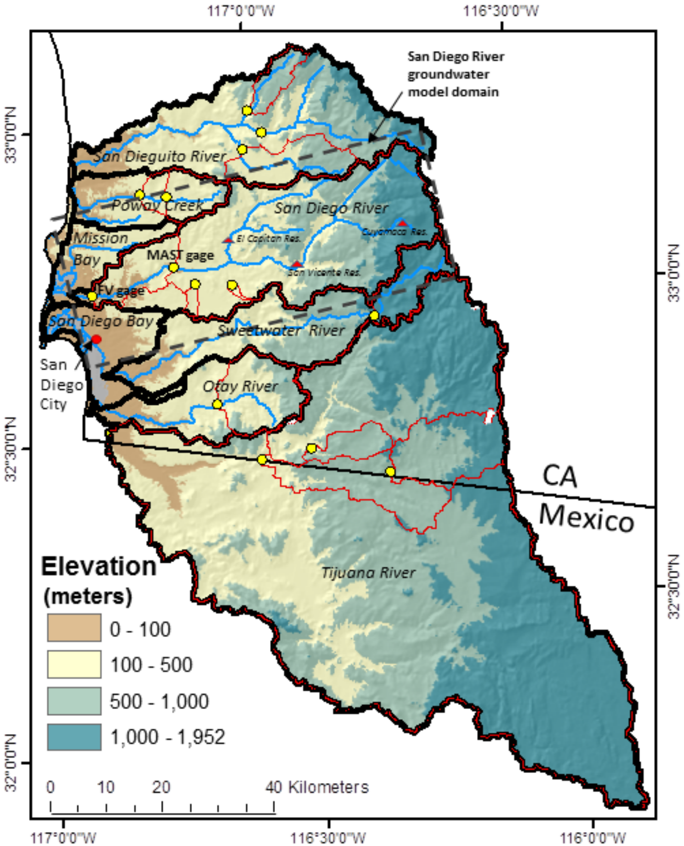

Fig. 1. Map of study area with major river basins outlined in black and calibration basins in red. Streamflow gages are noted as yellow points. San Diego River groundwater model, Mast Rd (MAST), and Fashion Valley (FV) gages, and reservoirs are indicated.

HESSD
9, 2717–2762, 2012
Water-balance and groundwater-flow estimation for an arid environment
L. E. Flint et al.

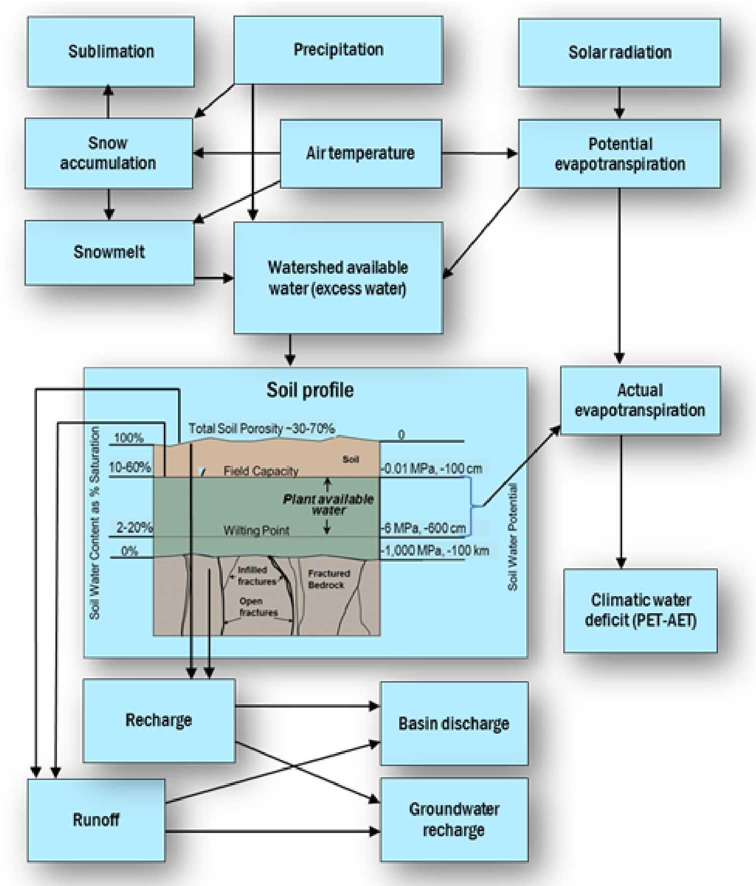

Fig. 2. Schematic illustrating the relation of the various components of the Basin Characterization Model.

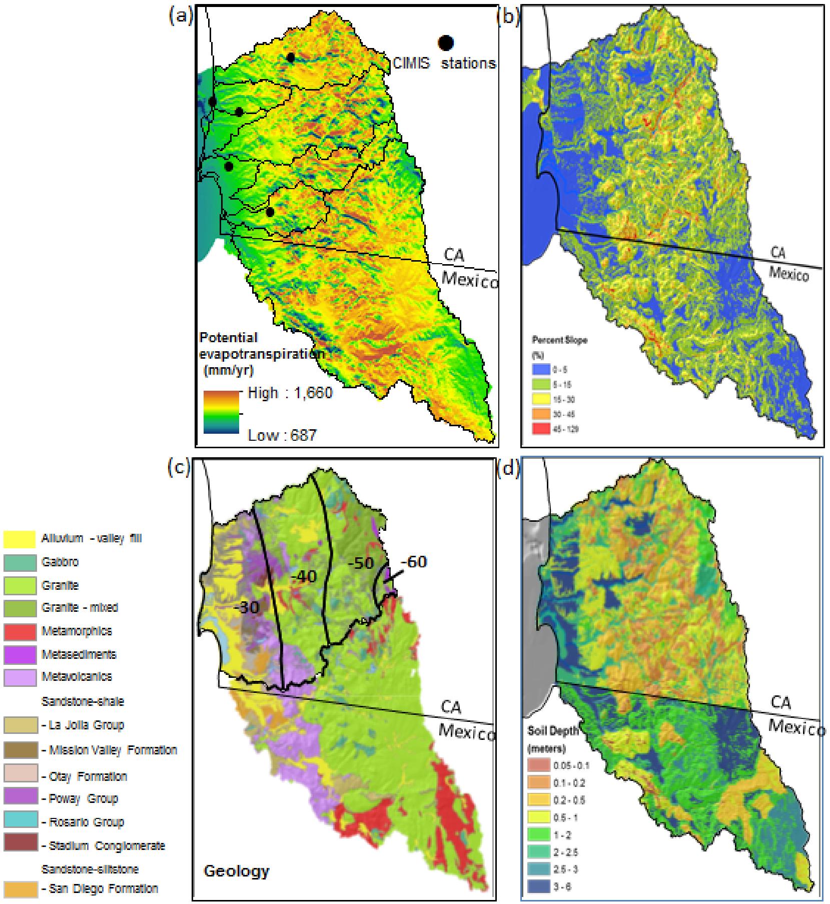

Fig. 3. Maps of input parameters for the Basin Characterization Model illustrating spatially distributed **(a)** average annual potential evapotranspiration, **(b)** slope, **(c)** geology and isotopic zones, indicated by values of δ Deuterium ‰, and **(d)** soil depth for the San Diego/Tijuana model.

HESSD
9, 2717–2762, 2012

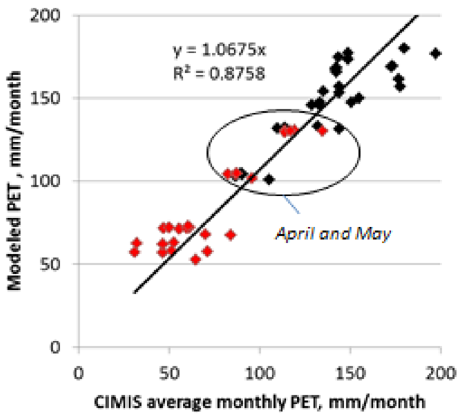

Fig. 4. Comparison of modeled potential evapotranspiration (PET) and PET measured at five stations from the California Irrigation Management Information System (CIMIS) in the San Diego region. Red points indicate November–April, black points indicate May–October.

HESSD
9, 2717–2762, 2012

Water-balance and groundwater-flow estimation for an arid environment

L. E. Flint et al.

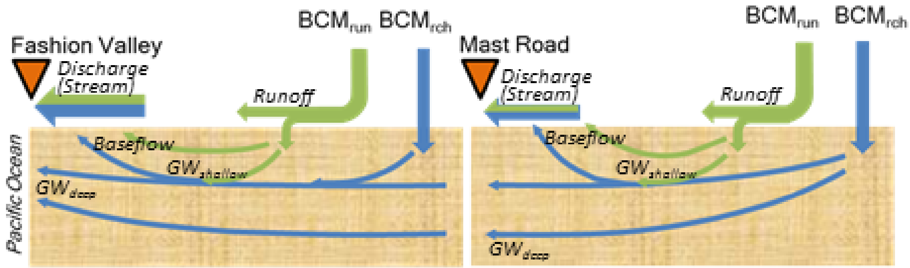

Fig. 5. Schematic illustrating the application of runoff and recharge from the Basin Characterization Model to the surface water and groundwater system.

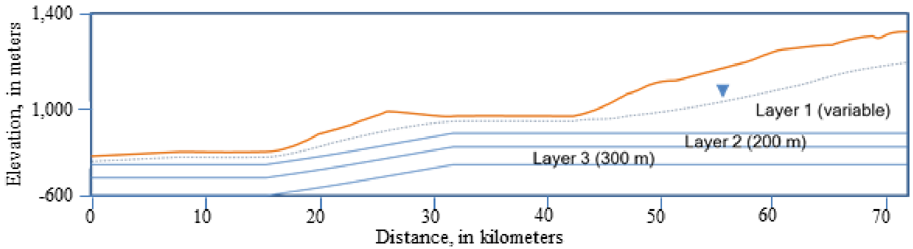

Fig. 6. Schematic cross-section of the San Diego River basin groundwater flow model domain showing generalized vertical model structure, land surface elevation, and depth to groundwater.

HESSD
9, 2717–2762, 2012
Water-balance and groundwater-flow estimation for an arid environment
L. E. Flint et al.

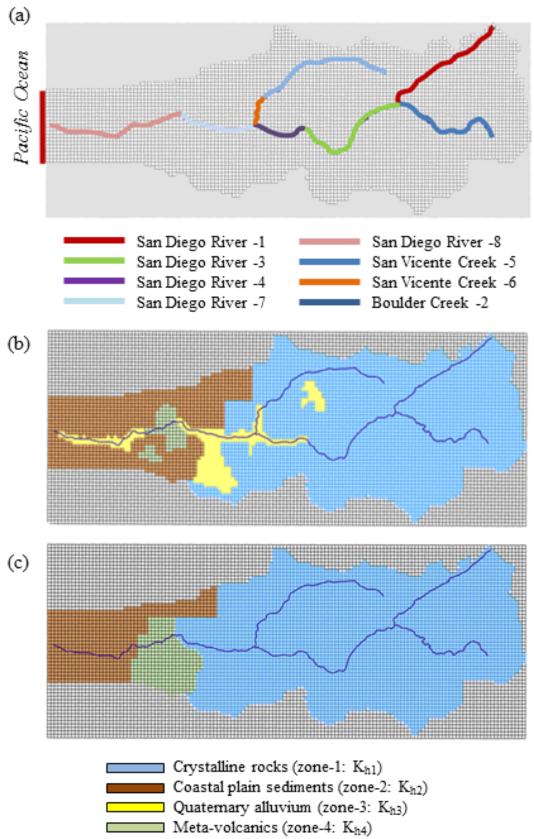

Fig. 7. Plan view of the San Diego River basin groundwater flow model domain illustrating the **(a)** grid, Streamflow Routing (SFR) Option boundary stream segments (indicated in legend, numbers correspond to Table 2), and Drain (DRN) Package boundary (red bar labeled as Pacific Ocean), and the geologic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

HESSD
9, 2717–2762, 2012

Water-balance and groundwater-flow estimation for an arid environment

L. E. Flint et al.

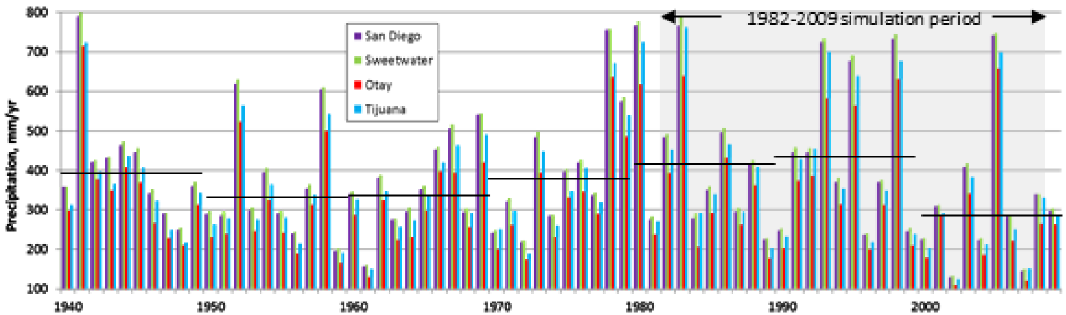

Fig. 8. Annual precipitation for the four major river basins in the San Diego/Tijuana study area with decadal mean precipitation indicated by the horizontal black lines. The gray shaded region indicates the groundwater model simulation period.

HESSD
9, 2717–2762, 2012

Water-balance and groundwater-flow estimation for an arid environment

L. E. Flint et al.

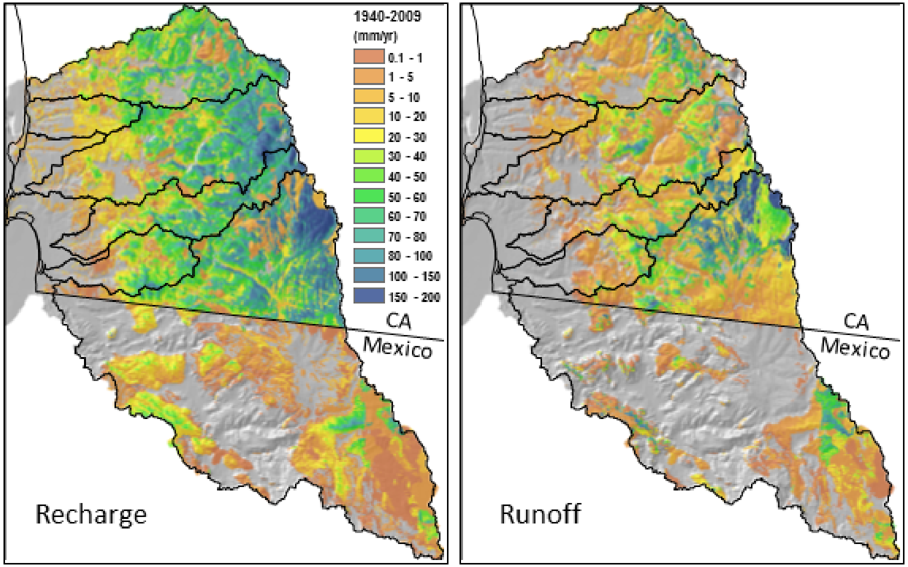

Fig. 9. Maps of average annual recharge and runoff for 1940–2009 calculated using the Basin Characterization Model for the San Diego/Tijuana study area.

HESSD
9, 2717–2762, 2012

Water-balance and groundwater-flow estimation for an arid environment

L. E. Flint et al.

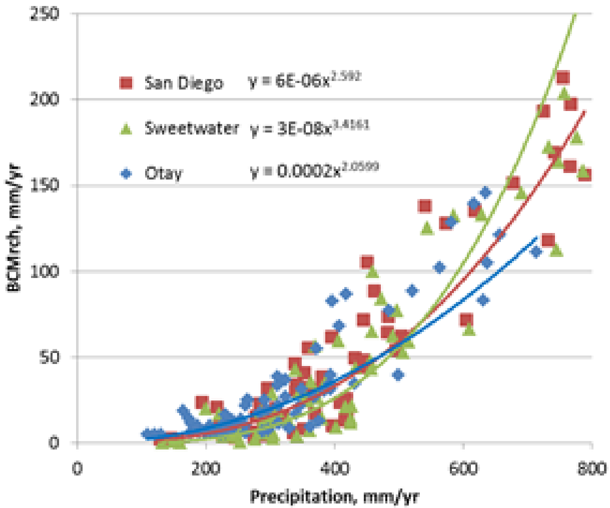

Fig. 10. Relation of BCM_{rch} to precipitation for 1940–2009 for three river basins.

HESSD
9, 2717–2762, 2012

Water-balance and groundwater-flow estimation for an arid environment

L. E. Flint et al.

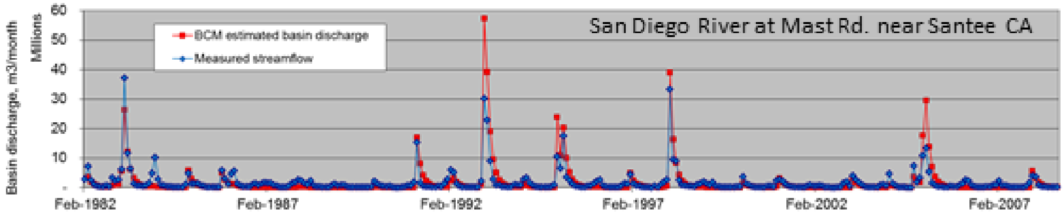

Fig. 11. Comparison of measured streamflow at the Mast Rd streamgage on the San Diego River with basin discharge estimated using the Basin Characterization Model.

HESSD
9, 2717–2762, 2012