

Interactive comment on “Water-balance and groundwater-flow estimation for an arid environment: San Diego region, California” by L. E. Flint et al.

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Abstract

Flint et al. estimated recharge from precipitation for the San Diego region using a water-balance model. A fundamental problem with the application of the model to the San Diego region is that in a semiarid climate annual evapotranspiration nearly equals the precipitation. Uncertainty in the evapotranspiration and precipitation is on

C1144

the same order of magnitude as the difference between those quantities. The uncertainty in the model recharge estimate can be assessed by considering the essential inputs to the model, the sensitivity of the resulting water-yield estimates to uncertainty in those inputs, and the uncertainty in the inputs. Such an exercise indicates that moderate uncertainty in the model inputs leads to large uncertainty in the estimate of the watershed-scale recharge. The coefficient of variation of the recharge estimate is about 100 percent, which means that the uncertainty is of the same order of magnitude as the recharge estimate itself.

1 Introduction

Flint et al. (2012) describe the application of the Basin Characterization Model (BCM) (Flint and Flint, 2007a; Flint and Flint, 2007b) to the San Diego region, California. The modeled area includes eight watersheds that individually discharge to the ocean north of the Mexican border. The BCM area covers about 8,000 km², including the San Diego River watershed, which has an area of about 1,100 km² (Figure 1). The climate within the San Diego region is semiarid. The average annual precipitation ranges from about 250 mm/yr near the coast, to 400 mm/yr within the middle watershed areas, and to 900 mm/yr within the upper watershed areas. The geographic distribution of precipitation is mostly a result of orographic effects. The average annual potential evapotranspiration ranges from about 700 mm/yr near the coast, to 1,700 mm/yr within the middle watershed areas, and to 800 mm/yr within the upper watershed areas. While mostly granitic rocks crop out within the middle and upper watershed areas, the granitic and other crystalline rocks are overlain by sedimentary deposits within the coastal areas. Flint et al. (2012) have postulated that a deep groundwater system within the crystalline rocks supports groundwater flow from upper watershed areas to the coast, where groundwater discharges to the ocean. The remainder of the water yield discharges to the ocean as streamflow.

The BCM is a water-balance model based on monthly accounting of inflow, outflow, and storage change within the root zone of the vegetation cover. The method is ap-

C1145

plied to an incremental area of a watershed, and the watershed water balance is the integration of the incremental areas over the watershed area. The BCM does not consider the possible redistributional effects of runoff. Correspondingly, the model treats each incremental area in isolation from other areas. For each incremental area, the water budget is expressed by the relation

$$r = p - et - ro - \Delta s \text{ (Eq. 1)}$$

where r is the recharge from an incremental area during a particular month, p is the precipitation during the month, et is the evapotranspiration, ro is the runoff, and Δs is the soil-moisture storage change. For a sufficiently long period (at least multi-decadal), the storage term vanishes, and the water yield (recharge plus runoff) is the precipitation that is not consumed by evapotranspiration.

Equation 1 represents the local recharge per unit area. The total recharge for a watershed is the integration of Equation 1 over the watershed area and averaged over time, or

$$R = \sum r_{ia_i} \text{ (Eq. 2)}$$

or

$$R = P - ET - RO \text{ (Eq. 3)}$$

where R is the average annual recharge for the watershed, r_i is the average annual recharge within the incremental, a_i is an incremental area, P is the average annual precipitation volume for the watershed, ET is the evapotranspiration volume, and RO is the runoff volume. Correspondingly, the summation is over the incremental areas comprising the watershed. For the application to the San Diego region, the incremental areas within the BCM area are defined by a rectilinear grid with 270-meter cells.

Flint et al. (2012) tested the BCM estimates of recharge and runoff using a stream-aquifer model of the San Diego River watershed. The model was constructed using MODFLOW. The recharge and runoff were superimposed on the model domain, and

C1146

the streamflow and groundwater underflow at the coast were simulated. The model was constructed using one of the streamflow-routing packages within MODFLOW to simulate stream-aquifer interactions along principal stream channels. The modeled streamflow was compared with measured streamflow, and the geographic and vertical distributions of hydraulic conductivity assigned to the groundwater model were adjusted based on that comparison. Given the BCM estimates of point recharge and runoff, the groundwater model was fitted to the measured streamflow.

Flint et al. (2012) estimated for the San Diego River watershed that the recharge is about 54×10^6 m³/yr (or 48 mm/yr) and runoff is about 19×10^6 m³/yr (or 17 mm/yr). The average annual precipitation within the watershed is about 530×10^6 m³/yr or 470 mm/yr. Correspondingly, the watershed yield equals 13 percent of the precipitation, and the recharge equals about 10 percent of the precipitation. Flint et al. (2012) report that the BCM relation between point recharge and precipitation can be approximated for the San Diego River watershed by a power function such that the recharge as a percentage of precipitation is 7 percent for precipitation of 300 mm/yr, 12 percent for precipitation of 500 mm/yr, and 26 percent for precipitation of 700 mm/yr. Flint et al. (2012) discuss the uncertainty in the recharge estimate, but they do not offer a quantification.

The stated purpose of the BCM application is to facilitate groundwater management within the San Diego region, but the uncertainty in the BCM recharge estimate is so large that the purpose is not achieved. Our comments on this application of the BCM to the San Diego region address the overall uncertainty in the estimate of recharge. The fundamental issue with the recharge estimate is that it is derived from the subtraction of two nearly equal uncertain quantities where the uncertainty is of the same order as the difference. Furthermore, the available data do not facilitate reducing the uncertainty through a model calibration. Finally, comments address mostly the recharge estimate for the San Diego River watershed within the BCM model area because Flint et al. (2012) provide the most complete information on that subarea.

C1147

2 Overall Uncertainty in Recharge Estimates

The recharge within the BCM area is a small percentage of the precipitation. Based on tables within Flint et al. (2012), the average annual runoff volume equals about 3 percent of the precipitation volume within the San Diego River watershed. The recharge equals about 10 percent of the precipitation. Correspondingly, the average annual evapotranspiration equals about 87 percent of the precipitation. The model area evapotranspiration is only slightly smaller than the precipitation, which leads to an exaggerated uncertainty in the water yield (Gee and Hillel, 1988). That uncertainty can be derived from the relation (Benjamin and Cornell, 1970)

$$\text{Var}[R] = \text{Var}[P] + \text{Var}[ET] + \text{Var}[RO] \quad (\text{Eq. 4})$$

This relational form assumes no correlation among errors in the independent variables, which probably is a reasonable representation of actual conditions. Equation 4 expresses the recharge uncertainty at the watershed scale, and that uncertainty can be derived by first assessing the recharge uncertainty at a point and then upscaling the uncertainty at a point to the watershed scale.

The uncertainty in point recharge was assessed by making simulations using a soil-water model to quantify the sensitivities of the simulated recharge to the model inputs. For that purpose, the soil-water module was extracted from the FEMFLOW3D groundwater-modeling program (Durbin and Bond, 1998). The module structure has similarities to the BCM structure. As in the BCM, the module is based on the water budget for a soil column, and it contains a relation for constraining evapotranspiration when the available soil moisture is limiting. The module dependencies are represented by

$$r = f(x_1, x_2, x_3, x_4, x_5, x_6) \quad (\text{Eq. 5})$$

where x_1 is the local infiltration of precipitation, x_2 is the local potential evapotranspiration, x_3 is the vegetation rooting depth, x_4 is a parameter related to the depen-

C1148

dence of evapotranspiration on soil moisture (Durbin and Bond, 1998), x_5 is the soil available water capacity, and x_6 is the vegetation cover. Given the functional relation expressed in Equation 5, the variance of the uncertainty in the recharge estimate is given by (Benjamin and Cornell, 1970)

$$\text{Var}[r] = \sum_{i=1}^6 \left(\frac{\partial r}{\partial x_i} \right)^2 \text{Var}[x_i] \quad (\text{Eq. 6})$$

where the uncertainties in independent variables are assumed to be uncorrelated. The partial differentials are approximated with finite differences derived from the soil-water module, where the partial differentials represent the sensitivities of the simulated recharge to the respective module inputs.

Equation 6 was applied to average annual infiltrations of 300, 500, and 700 mm. Monthly recharge was simulated for a 12-year period. Precipitation and potential evapotranspiration were derived from a California Irrigation Management Information System (CIMIS) (California Department of Water Resources, 2011) station within the San Diego region. The simulation results are summarized in Table 1 with respect to the coefficient of variation for point recharge. The coefficients are 1100 percent for average precipitation of 300 mm/yr, 360 percent for precipitation of 500 mm/yr, and 160 percent for precipitation of 700 mm/yr. The uncertainty in the point recharge estimates is based on the underlying uncertainty in the inputs to the simulations as listed in Table 2.

2.1 Uncertainty in Point Precipitation

The uncertainty in the point precipitation is based on comparisons between the PRISM maps (Daly et al., 2004) of monthly, annual, and average annual precipitation for the San Diego region. The precipitation input to the BCM was derived from monthly PRISM maps. The PRISM maps are a gridded representation of monthly precipitation based upon a regression of station precipitation data against variables describing orographic effects. Based on calendar year 2001, the comparison to monthly and annual precipitation with station data indicates monthly coefficients of variation that range from 20 to 300 percent for individual months. The coefficient of variation for the annual pre-

C1149

cipitation is about 20 percent. That is consistent with a comparison of PRISM maps of average annual precipitation in Nevada with station data (Jeton et al., 2005). The coefficient of variation for that comparison was about 15 percent, depending on the station set considered, but it tended to be larger with higher elevation.

2.2 Uncertainty in Point Evapotranspiration

Flint et al. (2012) used the Priestly-Taylor equation (Flint and Childs, 1991) for calculating hourly potential evapotranspiration, which was aggregated into monthly potential evapotranspiration. The approach requires measurements or estimates of net radiation, soil heat flux, air temperature, and atmospheric vapor density. The BCM was calibrated to monthly potential evapotranspiration derived from CIMIS climatic measurements at stations located within the northern coastal part of the model area (Figure 2). Nearly all the CIMIS stations are located within 4 km of the ocean, while the upper watershed boundary is about 80 km from the ocean. The average annual precipitation is about 300 mm within the region containing the stations, while the precipitation near the upper watershed boundary is as much as 900 mm/yr. Flint et al. (2012) indicate that most of the watershed recharge is generated within areas of higher elevation and precipitation. However, the BCM evapotranspiration functions were calibrated to stations within areas of lower precipitation and elevation, which are climatically distinct from areas of higher precipitation. Based on data compiled for Remote Automatic Weather Stations (RAWS) (Western Regional Climatic Center, 2011) for the San Diego region (Figure 2), the average annual relative humidity decreases from about 70 percent near the coast to about 45 percent at the upper watershed boundary. The coastal areas tend to be windier than mountain areas, but winds display high geographic variability.

The RAWS dataset includes information on potential evapotranspiration computed using the Kimberly-Penman equation (Wright, 1982). Those estimates of potential evapotranspiration characterize the general geographic distribution of potential evapotranspiration within the BCM area. Firstly, as shown on Figure 3, the RAWS data indicate a linear relation between average annual potential evapotranspiration and elevation over

C1150

the range of elevation within the BCM area. Secondly, the geographic application of that relation to the BCM area yields the map of average annual potential evapotranspiration shown on Figure 4. That map is significantly different than the corresponding map produced by Flint et al. (2012, Figure 3a). Their map indicates that average annual potential evapotranspiration is about 700 mm/yr within coastal areas, about 1,700 mm/yr within middle watershed areas, and about 800 mm/yr within upper watershed areas. In contrast, Figures 3 and 4 indicate that the average annual potential evapotranspiration is about 900 mm/yr within coastal areas, about 1,050 mm/yr within middle watershed areas, and about 1,150 mm/yr within upper watershed areas. While Flint et al. (2012) conclude that the highest potential evapotranspiration occurs within the middle watershed areas, the RAWS data indicates that the highest potential evapotranspiration occurs within the upper watershed areas. These differences suggest that the potential evapotranspiration maps generated by Flint et al. (2012) may contain considerable uncertainty.

2.3 Uncertainty in Point Land-Surface Characterizations

The land-surface characterizations include the vegetation cover density, vegetation rooting depth, and soil available water capacity. Flint et al. (2012) presumably used as a BCM input the U. S Geological Survey (2011) vegetation mapping, or similar mapping, to assess cover type, density, and rooting depth, which was the case for a previous BCM application (Flint et al., 2007a). For the San Diego River watershed, the U. S. Geological Survey mapping delineates general vegetation classes, such as live oak woodland and savanna, chaparral, coastal scrub, and coastal grass land. The mapping system is such that particular classifications include broad variations in vegetation composition, such as the oak woodland-savanna classification, which ranges from closed-canopy woodlands to mostly grasslands. The diversity within that particular classification represents different cover densities, rooting depths, and water use, which leaves the characterization at a point within the area delineated for the classification very uncertain. The same uncertainty exists within other vegetation classifications.

C1151

Flint et al. (2012) used as a BCM input a generalized soil map produced by the National Resources Conservation Service (2006). The map was created by generalizing more detailed soil survey maps. Where more detailed soil survey maps were not available, data on geology, topography, vegetation, and climate were assembled. The soils mapping was used to compile soil depth, field capacity, wilting point, porosity, and other parameters for BCM inputs. As for the vegetation map, the soil-map classifications include soils with different characteristics, which leaves the point characterization very uncertain.

2.4 Uncertainty in Watershed-Scale Recharge

The uncertainty in point process can be translated into watershed-scale processes using the variance-reduction method developed by Vanmarcke (2010). The watershed uncertainty is smaller than the point uncertainty because the point uncertainty is smoothed in the summation from the point recharge to the watershed recharge. The magnitude of the reduction depends on the correlation structure for the uncertainty in the recharge estimates as described by the relations

$$\text{Var}[R] = \gamma(A) \text{Var}[r] \quad (\text{Eq. 7})$$

where

$$\gamma[A] = \alpha/A \quad (\text{Eq. 8})$$

and

$$\alpha = \int_{-\infty}^{\infty} A(a) da \quad (\text{Eq. 9})$$

where γ is the variance-reduction factor, γ is the correlation function, and α is the characteristic area. These relations indicate that the variance reduction is smaller for higher spatial correlations.

The correlation distance for the point-recharge uncertainty is probably large. A large correlation distance most likely characterizes the PRISM precipitation maps because

C1152

of the geographical sparsity of station data, especially for the early years included in the BCM simulations. A large correlation distance most likely also characterizes the geographic distribution of potential evapotranspiration because of bias suggested by the comparison between the BCM simulations and RAWs data. Finally, large correlation distances apply also to the land-surface characterizations. The BCM incorporates generalized rooting depth, available water capacity, and vegetation-cover density. For a particular soil or vegetation class, the same parameter value is assigned throughout the BCM area. Concomitantly, errors occurring in the parameterization of a soil or vegetation class will be highly correlated across the BCM area. Assuming a linear correlation function for the uncertainty in the point recharge and a correlation distance of 15 km, the variance reduction factor is 20 percent, which means that the variance for the watershed-scale recharge is 20 percent of the variance for the point recharge.

The watershed-scale uncertainty is the composite of the point recharge uncertainty for different precipitation zones, which is characterized by a decrease in the coefficient of variation with an increase in precipitation. The result is a coefficient of variation for the San Diego River watershed recharge of about 100 percent. The recharge estimated by Flint et al. (2012) is 54×10^6 with an uncertainty of $\pm 54 \times 10^6$ m³/yr.

The uncertainty in recharge estimates produced by the BCM is described by Masbruch et al. (2011) for an application to a regional groundwater system within the Great Basin, Nevada and Utah. The BCM was applied much as it was for the San Diego region. However, for the Great Basin groundwater system, estimates were available regarding discharges from the groundwater system, which would be the equivalent to knowing the underflow to the ocean prior to applying the BCM to the San Diego region. For the Great Basin groundwater system, local adjustments were made to the BCM recharge estimates to match better the recharge implied by the discharge estimates. While the BCM recharge estimates were reduced by a specified factor in some subareas, the estimates were increased in other subareas. The adjustment factors ranged from 0.20 to 2.25, which indicates considerable disparity between the BCM recharge estimates

C1153

and the prior discharge estimates.

2.5 Uncertainty in Bedrock Hydraulic Conductivity

Flint et al. (2012) developed a groundwater model for the San Diego River watershed to partition the water yield between streamflow and underflow at the coast. Stream-aquifer interactions occur such that both streamflow and underflow comprise some mixture of point runoff and recharge. Using recharge and runoff generated by the BCM, the groundwater model was used to simulate streamflow at a streamgaging site on the San Diego River near the coast. Simulations were made assuming different hydraulic conductivity to characterize the deep groundwater system. The simulations respectively used conductivities of 8 and 1 m/d, but the higher conductivity produced a better fit of the groundwater model to the measured streamflow. The transmissivities corresponding to these conductivities respectively are about 4,000 and 500 m²/d. From the simulation results, Flint et al. (2012) conclude that the underflow at the coast may equal about 40 percent of the point recharge within the San Diego River watershed, or about 22x10⁶ m³/yr.

The underflow at the coast depends on the hydraulic characterization of the deep groundwater system. Different combinations of BCM recharge and groundwater-model hydraulic conductivity can fit the streamgaging measurements with correspondingly different quantities of BCM recharge and partition between streamflow and underflow. Correspondingly, the uncertainty in the recharge and partitioning is tied ultimately to the uncertainty in the hydraulic characteristics of the deep groundwater system. That uncertainty unfortunately, is large. Furthermore, the hydraulic conductivities used in the groundwater model probably are much larger than actually exists.

The most extensive information on the hydraulic conductivity of the deep aquifer system is a collection of specific-capacity tests reported by well drillers to the California Department of Water Resources. About 150 wells are located within middle and upper watershed areas where the crystalline rocks comprising the deep aquifer system crop

C1154

out. The geometric mean of the hydraulic conductivity derived from tests on these well is about 2x10⁻² m/d, which is about two orders of magnitude less than the conductivity used in the groundwater model. Kaehler and Hsieh (1994) evaluated the hydraulic conductivity of fractured rock within a subarea of the BCM area, and they derived a conductivity of about 10⁻³ m/d.

The specific-capacity data suggest a decay of conductivity with depth, which is similar to the findings of Page et al. (1984), Borchers (1996), and Boutt et al. (2010). For the San Diego region specific-capacity data, the depth decay is such that the aquifer transmissivity is 10 m²/d, which is two and three orders of magnitude smaller than the aquifer transmissivity used in the two separate groundwater model formulations. If the transmissivity derived from the specific-capacity tests were to be used in the groundwater model, the BCM recharge would need to be reduced substantially in order to fit the groundwater model to the streamgaging data, and the groundwater model would have simulated at least an order of magnitude less underflow at the coast.

3 Lack of Documentation

Flint et al. (2012) do not provide a citation that adequately describes formulation of the BCM simulator. Elements of the formulation appear in Hevesi et al. (2003), Flint and Flint (2007b), and U. S. Geological Survey (2008). However, Flint et al. (2012) do not identify where a comprehensive description of the BCM simulator can be found. The BCM appears to have evolved from the U. S. Geological Survey simulator INFIL (U. S. Geological Survey, 2008), based on comparisons among the INFIL documentation and various BCM narratives that appear in subsequent papers and reports (Flint et al., 2001a, Flint et al., 2001b, Flint et al., 2002, Flint et al., 2004, Flint and Flint, 2007a, and Flint and Flint, 2007b). The first specific reference to the BCM is in Flint et al. (2007b), but that report contains only a diagram of the BCM structure. Subsequent papers provide little additional information. Consequently, little information is available to judge the adequacy of the BCM structure.

C1155

4 Conclusions

A fundamental problem with the application of the BCM to the San Diego region is that in a semiarid climate annual evapotranspiration nearly equals the precipitation. Uncertainty in the evapotranspiration and precipitation is on the same order of magnitude as the difference between those quantities. The result is an exaggerated uncertainty in the recharge estimate. A second problem with the application is that the model requires calibration, because direct measures of model parameters are unavailable or incomplete. Correspondingly, the model development was based on generalized information of highly uncertain specificity. Were the model to be calibrated, the calibration target should be the water yield from the BCM area or a subarea.

The water yield of the BCM area ultimately discharges to the ocean as either streamflows or underflow. While streamgaging data allow a reasonable estimate of the streamflow discharge to the ocean from the San Diego River watershed, the available data facilitate only an order of magnitude estimate of underflow, which means that the water yield from the BCM area is essentially unknown, and no basis exists for a model calibration. Nevertheless, the uncertainty in the BCM recharge estimate can be assessed by considering the essential inputs to the BCM, the sensitivity of the resulting water-yield estimates to uncertainty in those inputs, and the uncertainty in the inputs. Such an exercise indicates that moderate uncertainty in the BCM inputs leads to large uncertainty in the estimate of the watershed-scale recharge. The coefficient of variation of the recharge estimate is about 100 percent, which means that the uncertainty is of the same order of magnitude as the recharge estimate itself. This is the expected result of applying a soil water-budget with an arid or semiarid environment (Gee and Hillel, 1988).

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C1156

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C1157

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C1158

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C1159

C1160

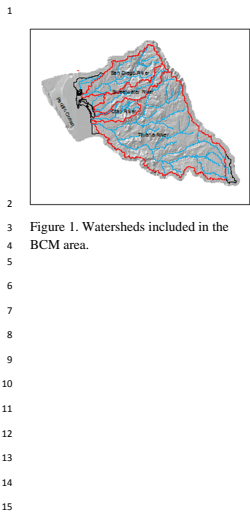


Fig. 1. Figure 1. Watersheds included in the

C1161

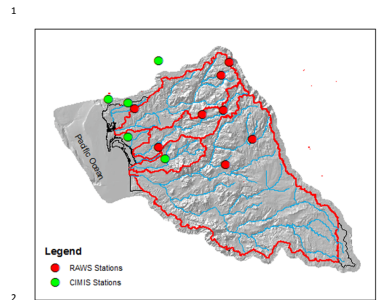


Figure 2. Locations of RAWIS and CIMIS climatic stations.

Fig. 2. Figure 2. Locations of RAWIS and CIMIS climatic stations.

C1162

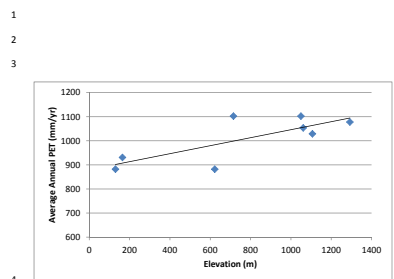


Figure 3. Relation between potential evapotranspiration and elevation based on RAWIS data.

Fig. 3. Figure 3. Relation between potential evapotranspiration and elevation based on RAWIS data.

C1163

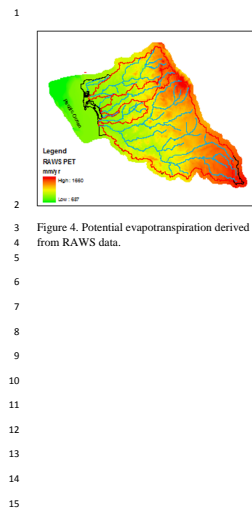


Fig. 4. Figure 4. Potential evapotranspiration derived

C1164

Table 1. Uncertainty in point recharge estimates derived from FEMFLOW3D soil-water module.

Precipitation (mm/yr)	Recharge (mm/yr)	Standard Deviation (mm/yr)	Coefficient of Variation (percent)
300	22	240	1100
500	85	300	360
700	210	340	160

Fig. 5. Table 1. Uncertainty in point recharge estimates derived from FEMFLOW3D soil-water module.

C1165

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Table 2. Uncertainty assigned to inputs to the FEMFLOW3D soil-water module.

Input	Coefficient of Variation (percent)
Precipitation	15
Potential evapotranspiration	20
Rooting depth	50
Soil-moisture parameter	20
Soil available water capacity	20
Vegetation coverage	30

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Fig. 6. Table 2. Uncertainty assigned to inputs to the FEMFLOW3D soil-water module.

C1166