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LOSS RATES FROM LAKE POWELL AND THEIR IMPACT ON MANAGEMENT OF THE COLORADO RIVER[1]

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ABSTRACT: As demand for water in the southwestern United States increases and climate change potentially decreases the natural flows in the Colorado River system, there will be increased need to optimize the water supply. Lake Powell is a large reservoir with potentially high loss rates to bank storage and evaporation. Bank storage is estimated as a residual in the reservoir water balance. Estimates of local inflow contribute uncertainty to estimates of bank storage. Regression analyses of local inflow with gaged tributaries have improved the estimate of local inflow. Using a stochastic estimate of local inflow based on the standard error of the regression estimator and of gross evaporation based on observed variability at Lake Mead, a reservoir water balance was used to estimate that more than 14.8 billion cubic meters (Gm^3) has been stored in the banks, with a 90% probability that the value is actually between 11.8 and 18.5 Gm^3. Groundwater models developed by others, observed groundwater levels, and simple transmissivity calculations confirm these bank storage estimates. Assuming a constant bank storage fraction for simulations of the future may cause managers to underestimate the actual losses from the reservoir. Updated management regimes which account more accurately for bank storage and evaporation could save water that will otherwise be lost to the banks or evaporation.

(KEY TERMS: Lake Powell; reservoir bank storage; surface water/groundwater interactions; water conservation; water supply; reservoir operations simulations.)

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INTRODUCTION

As demand for water in the southwestern United States (U.S.) increases and climate change potentially decreases the inflow to the Colorado River system (Christensen and Lettenmaier, 2007; Barnett and Pierce, 2008; Barsugli *et al.*, 2009; Miller and Piechota, 2011), the need to optimize the water supply will increase. The Colorado River has 73.4 billion cubic meters (Gm^3) of available storage in its 10 largest reservoirs (USBR, 2011), which is approximately four times the river's average annual flow. The live storage in those reservoirs on October 1, 2011, was 47.7 Gm^3 (USBR, 2011), after the wettest runoff year in at least 15 years. The available storage space, 25.7 Gm^3, is about one and a half years of long-term average inflow, estimated to be 18.5 Gm^3 per year (Gm^3/yr) (USBR, 2007). Lake Powell, the second largest reservoir on the river, has a maximum storage equal to 32.1 Gm^3. As of August 2010 there was 22.3 Gm^3 stored in the banks of the reservoir according to an ongoing water balance (http://www.usbr.gov/lc/region/g4000/NaturalFlow/documentation.html, accessed August 23, 2010).

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The U.S. Bureau of Reclamation (USBR) manages the Colorado River and reservoirs according to a complex system of laws, treaties, and agreements known as the Law of the River (http://www.usbr.gov/lc/region/g1000/lawofrvr.html, accessed September 6, 2012). The USBR simulates the Colorado River system using the Colorado River Simulation System (CRSS) (USBR, 1986), which has been incorporated in RIVERWARE software (Zagona *et al.*, 2001). Assumptions used in CRSS, including reservoir loss rates, affect the results of the simulations (USBR, 2007) and therefore potentially the decision-making processes. The assumptions include loss rates from Lake Powell.

Lake Powell loses water from the river system in two ways — to evaporation from the free water surface and bank seepage. Bank seepage is flow to the banks that does not return to the river system whereas bank storage may return to the reservoir or river system. CRSS does not simulate seepage but rather assumes a change in bank storage equal to 8% of the monthly change in reservoir storage (Jerla, 2005), which averages 0.53 Gm^3/month. The simulation assumption does not account for the USBR's published values of bank storage. No studies have been completed that demonstrate that water which seeps into the banks actually returns to the reservoir to justify treating all of it as bank storage. The fact that the amount of water stored in the banks approximates a year's worth of river flow suggests that unaccounted for, the bank storage is a large potential error in the simulations and a source of inaccuracy in the management of the river system that is the source of water supply for about 30 million people in the southwestern U.S.

The objective of this article was to improve the estimate of monthly and cumulative bank storage and seepage on Lake Powell by improving the estimate of local inflow, which is ungaged and estimated as a fraction of the difference in gages on the mainstem river as described below. Variability in the estimates of bank storage is estimated using the stochastic properties of local inflow and evaporation. The article also discusses the differences in actual and simulated bank storage and makes recommendations for utilizing this new information to make more accurate predictions for the management of the water resources in Lake Powell.

HYDROLOGY OF LAKE POWELL AND THE UPPER COLORADO RIVER BASIN

Lake Powell lies near the downstream end of the upper basin of the Colorado River system which is divided politically into upper and lower basins at Lees Ferry, Arizona (Figure 1). The total area of the upper basin is 293,200 square kilometers (km^2), which is roughly split between the Rocky Mountains headwaters and the Colorado Plateau (Fenneman, 1931). The Colorado River at Lees Ferry gaging station 9380000 (Figure 1 and Table 1), which lies about 25 kilometers (km) downstream from Glen Canyon Dam, measures the flow from the upper to the lower basins. The flow at this gage has varied with time (Figure 2) with construction of Lake Powell causing the largest change to the flow from the upper basin.

The Colorado Plateau consists of nearly horizontal sedimentary strata deeply incised by major stream systems and interrupted by north-south trending monoclines, structural domes, and basins along with widely scattered extrusive and intrusive igneous features (Blanchard, 1986). The Navajo Sandstone forms the walls of Glen Canyon, and contains and

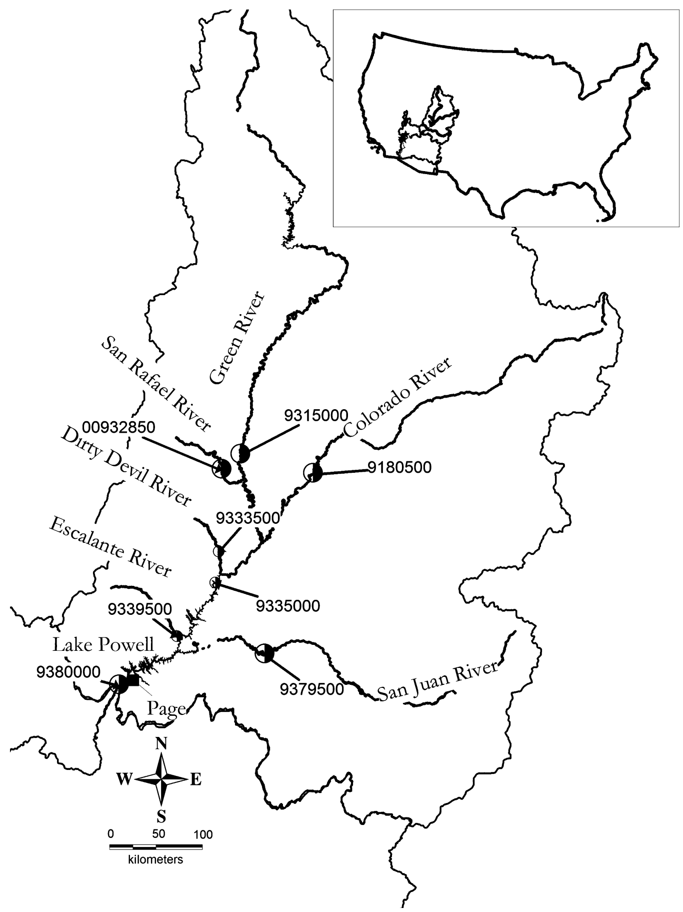

FIGURE 1. Upper Colorado River Basin and Select Gaging Stations. Gaging station 9380000 — Colorado River at Lees Ferry, Arizona; 9335000 — Colorado River at Hite, Utah; 9315000 — Green River at Green River, Utah; 9180500 — Colorado River near Cisco, Utah; 9379500 — San Juan River, near Bluff, Utah; 9339500 — Dirty Devil River above Poison Spring Wash, Utah; 9339500 — Escalante River at mouth near Escalante, Utah; 9328500 — San Rafael River near Green River, Utah. See Table 1 for drainage area and periods of record.

TABLE 1. U.S. Geological Survey Gaging Stations in the Upper Colorado River Basin, Used for This Analysis. See Figure 1 for the location of these stations within the basin. Data from U.S. Geological Survey National Water Information System for Utah and Arizona (http://waterdata.usgs.gov/ut/nwis/sw) and Arizona (http://waterdata.usgs.gov/az/nwis/sw/).

Gaging Station Name	Gaging Station Number	Drainage Area (km^2)	Period of Record[1]	Avg. Annual Flow (Gm^3/yr)	Month Count
Colorado River near Cisco, Utah	9180500	62,400	10/13-present	6.43	1,092
San Juan River near Bluff, Utah	9379500	59,600	10/14-present	2.00	1,019
Green River at Green River, Utah	9315000	116,000	10/1894-present	5.42	1,315
Sum of river inflow gages		238,000		13.8	
Colorado River near Lees Ferry, Arizona	9380000	289,600	-	18.5[2] 14.6[3] 12.1[4]	1,227
Other Gages					
Colorado River at Hite, Utah	9335000	198,000	8/47-9/58	12.0	134
Escalante River near Escalante, Utah	9339500	829	10/42-9/55; 12/71-present	0.00957	610
Dirty Devil River above Poison Creek near Hanksville, Utah	9333500	10,800	10/43-9/98; 6/2001-present	0.0885	640
San Rafael River near Green River, Utah	9328500	4,260	10/09-9/18; 10/45-present	0.118	876

[1]The period of record may include periods without measurement, however, the intent is to use only gages with a mostly complete record.
[2]Lees Ferry gage natural flows as estimated by USBR for 1906 through 2007 (Prairie and Callejo, 2005).
[3]Actual Lees Ferry gage flows for 10/21-9/63.
[4]Actual Lees Ferry gage flows 10/63-9/2009.

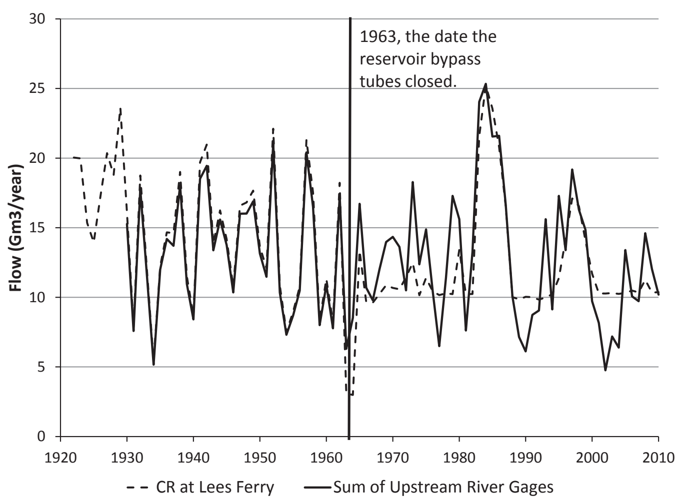

FIGURE 2. Historic Annual Flows on the Colorado River, by Water Year, Below Lake Powell (CR at Lees Ferry) and Main River Inflow to Lake Powell. See Table 1 for gaging station numbers and data citation.

transmits any bank storage and seepage. Thomas (1986) found that the Navajo Sandstone dips northward from the reservoir with a conductivity of about 0.4 meters per day (m/day). Jacoby *et al.* (1977) estimated that 10.4 Gm^3 of water had been stored in the banks between 1964 and 1976. Thomas (1986) developed a groundwater model which essentially verified Jacoby *et al.*'s estimate of bank storage and indicated that much of the water in the banks would not return to the reservoir.

Natural river flows are the flows that would have occurred without upstream human-induced consumptive use (Prairie and Callejo, 2005). Natural flows at Lees Ferry from 1906 to 2007 averaged 18.5 Gm^3/yr and for 1963 through 2007 averaged 17.8 Gm^3/yr. Historic flows, the actual gaging station flow measured at the Lees Ferry gage, averaged 14.8 Gm^3/yr for 1927 through 1962 and 11.9 Gm^3/yr for 1963 through 2007 (Figure 2). The difference between natural and historic flows is the USBR's estimate of consumptive use within the upper basin. The difference between the two periods includes the effect of Lake Powell and any changes in the watershed condition.

Mainstem river inflow to Lake Powell is estimated as the sum of flow at three upstream gages and outflow is the measured flow at the Lees Ferry gage downstream from the reservoir (Table 1 and Figure 1). Errors in the flow rates are primarily due to measurement error and normally distributed around zero (Haan, 1977). Local inflow is ungaged inflow that enters the river or reservoir from the 51,400 km^2 that lies between the upper and lower gaging stations. Local inflow from this semiarid area includes groundwater discharge, local runoff, and several perennial rivers. Groundwater inflow would be relatively constant to the river or to the reservoir basin although development of the reservoir would change the hydraulic gradient so that the discharge to the reservoir would be lessened or even reversed (Blanchard, 1986; Thomas, 1986). Local runoff and stream inflow would continue unchanged due to the reservoir except for a portion of the area covered by the reservoir.

The USBR estimated local inflow for reservoir management as a fraction of the difference in average monthly pre-dam flows measured at the three upstream and one downstream gaging stations (Figure 3), similar to Jacoby *et al.* (1977) (Rick Clayton, USBR Salt Lake

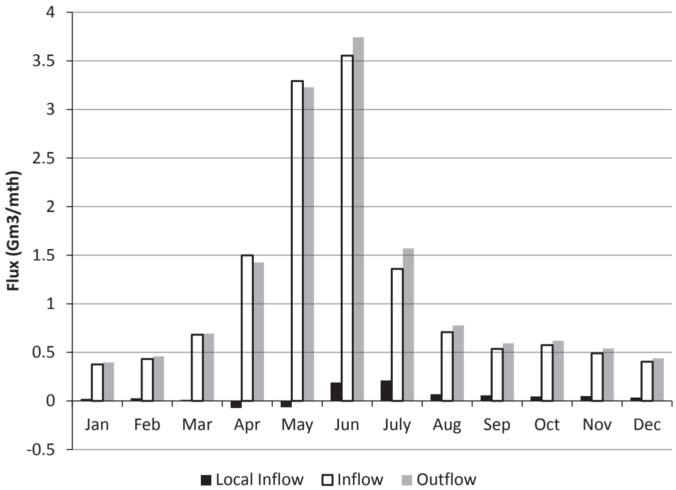

FIGURE 3. Local Inflow, Inflow, and Outflow Determined as the Difference between Inflow and Flow to the River Reach between the Upstream Gages Used as Inflow to Lake Powell and the Downstream Gage Used as Outflow from the River Reach for the Period of Record 1927 Through 1962 (Jacoby *et al.*, 1977).

City, November 29, 2010, personal communication). Both methods estimate that 0.063 Gm^3/month is the average for the 1927 through 1962 period, just prior to the closure of the bypass tubes of Glen Canyon Dam in 1963, but the fractions differ by month (Figure 3). Local inflow for the period before the reservoir began filling is a gain during all months, except for April and May (Figure 3) when high flows are recharging the banks. The estimates do not vary among wet or dry years or account for actual watershed conditions, but do account for development in the basin in that the estimates have been adjusted to represent natural flows.

METHOD OF ANALYSIS

Reservoir Water Balance

Reservoir water balance may be described with Equations (1-3).

$$\text{Inflow-Outflow} = \text{Change in Storage} + \text{Residual} \quad (1)$$

$$\text{Inflow} = Q_i + Q_l + P \quad (2)$$

$$\text{Outflow} = E + Q_o \quad (3)$$

Q_i is major river inflow, Q_l is local inflow, and P is precipitation on the reservoir surface. E is evaporation from the reservoir surface and Q_o is river outflow. Change in reservoir storage may be estimated with stage/volume relationships (USBR, 1985, 2007). If all of the water balance factors are estimated independently, bank storage is a residual that also accounts for errors in the estimation or measurement of any of the factors. Accuracy in the bank storage estimates depends on the accuracy of the estimation of the terms in Equations (2) and (3).

The local inflow estimate may be the largest source of uncertainty in reservoir bank storage estimates. Local inflow may depend less on the high elevation runoff from the Rocky Mountains than on local factors which are not represented by the three upstream gages. Assuming that relations between local gaging stations and the calculated local inflow are the same after 1963 as before, statistical analyses including correlation and multiple linear regression were used to revise the estimate of local inflow used as input to the reservoir water balance analysis. The multiple linear regression included indicator variables to account for differences among months and the Durbin-Watson test of residuals used to test whether significance depends on autocorrelation (Neter *et al.*, 1985).

The residual of the reservoir water balance, calculated using historic flows with Equations (1-3), is flow to the bank. Monte Carlo simulations with local inflow and net evaporation estimated as stochastic variables were used to account for uncertainty. The estimated variables were assumed to be normally distributed, using the polar method for estimating $N(0,1)$ random variants (Law and Kelton, 1991). For local inflow, the actual standard error from the regression was used for adjusting to actual inflow estimates. The standard deviation for the monthly evaporation rate equaled 10%, based on data presented in Westenburg *et al.* (2006). Simulations continued until the moving average of all simulations fluctuated within 2% of the deterministic value of bank storage, determined from the local inflow regression and average evaporation. The upper and lower 90% confidence bands were determined as the simulation, which yielded the final bank storage within 5% of the lowest and highest simulated values. Multiple linear regression of bank storage with reservoir storage characteristics using the deterministic bank storage values was used to consider the controls on the rate that bank storage accumulates or seepage is lost.

RESULTS

Local Inflow Estimation

The correlation coefficient of historic local monthly inflow with the sum of the historic flow at the three upstream gages, for the 1927 through 1962 period, is 0.230 and with the San Juan, Colorado, and Green River gages is 0.24, 0.22, and 0.23, respectively. This shows that the flow from above these mainstream gages explains only a small amount of variation in local inflow.

Table 2 provides monthly and annual statistics for historic local inflow and other rivers that enter the river within the area considered for local inflow (Figure 1). Monthly local inflows from 1927 to 1962 have positive skew and the mean (0.063 Gm^3/month) exceeds the median by 1.6 times. Local inflows have high variability as indicated by the standard deviation being 2.2 times the mean and numerous negative values due to the river recharging the banks. Accounting for local factors is essential for understanding the variability of local inflow.

TABLE 2. Average Monthly, Average Annual Flow, and Monthly Standard Deviation at Gaging Stations on Three Lake Powell Tributaries (Escalante River near Escalante, Dirty Devil River above Poison Creek, and San Rafael River near Green River) for Their Period of Record to 2010 and Local Ungaged Inflow to the Lake Powell Reach from 1927 to 1962. Statistics for local tributaries based on complete period of record (Table 1). Count is the number of months included in calculating the statistics, and the effective count reflects the reduced information content in the mean based on lag 1 autocorrelation.

Month	Local Ungaged Inflow (1927-1962)	Escalante	Dirty Devil	San Rafael
	Average Flow (Mm^3/month)			
January	21.53	0.62	7.53	3.15
February	26.89	0.69	9.27	4.59
March	10.10	0.93	10.43	7.33
April	–68.80	0.95	7.72	6.99
May	–39.86	1.73	6.23	20.71
June	190.21	1.40	4.79	37.76
July	211.27	0.47	3.88	10.66
August	99.30	0.64	6.69	6.22
September	57.32	0.49	6.21	5.33
October	49.18	0.60	9.52	6.89
November	52.03	0.51	9.18	4.65
December	33.20	0.53	7.12	3.32
Annual (Mm^3/yr)	642.38	9.56	88.59	117.59
	Standard Deviation			
January	21.43	0.45	2.84	2.31
February	27.04	0.45	2.60	2.71
March	51.91	0.61	3.64	7.91
April	75.64	0.95	5.71	9.35
May	148.11	2.19	5.94	26.90
June	151.55	2.65	7.44	44.84
July	107.93	0.53	4.76	14.61
August	94.49	0.58	8.33	6.07
September	76.49	0.61	9.07	5.01
October	67.17	0.56	16.24	9.24
November	48.15	0.39	10.14	3.91
December	23.74	0.36	2.14	1.78
	Autocorrelation of Monthly Flows			
Lag 1	0.27	0.54	0.27	0.48
Lag 2	–0.04	0.24	0.08	0.15
Lag 3	–0.18	0.15	0.07	0.05
Lag 12	0.47	0.04	0.14	0.44
Count	427	610	640	876
Effective count	245	183	370	309

Three gaged perennial tributary rivers enter the Colorado River within the local inflow reach — the Escalante, Dirty Devil, and San Rafael (Table 1, Figure 1). Their total average annual flow is about 0.216 Gm^3/yr for years with coinciding periods of record meaning they account for more than a third of the average local inflow, leaving almost two-thirds unaccounted for (Figure 4 and Table 2). Measured flows from these rivers and the computed local inflow have decreased over the period of record (Figure 4), with decreases for local inflow, San Rafael River, Dirty Devil River, and the Escalante River equaling 1.19, 0.905, 0.00937, and 0.00395 million cubic meters per year (Mm^3/yr), respectively. The cause of the decrease could be development, of which there has been little, or long-term flow changes, as in the Colorado River watershed (Woodhouse *et al.*, 2006; Brekke *et al.*, 2007; Meko *et al.*, 2007).

Computed local inflow can be negative whereas measured tributary flow is positive. The highest local inflow months correspond with the highest tributary flows on the Escalante and San Rafael Rivers, but not the Dirty Devil (Table 2). The Escalante and San Rafael Rivers have much higher elevation headwaters with snowmelt controlled flow. The Dirty Devil River flows more in response to the rainfall/runoff regime,

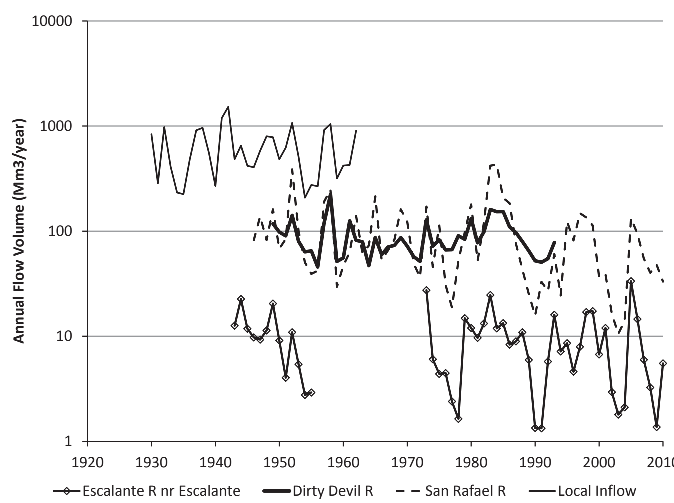

FIGURE 4. Water Year Flow at Gaging Stations on Three Lake Powell Tributaries (Escalante River near Escalante, Dirty Devil River above Poison Creek, and San Rafael River near Green River) for Their Period of Record Through 2010, and Local Inflow to the Lake Powell Reach from 1927 to 1962. See Tables 1 and 2 for gaging station numbers, data citations, and flow statistics and Figure 1 for location.

so high flows can occur during any month, but least frequently during summer due to small areal storm coverage. The months May through August were most variable, as shown by the standard deviation of the monthly flows, with the exception of October on the Dirty Devil River (Table 2).

The autocorrelation of monthly local inflow reflects how antecedent conditions can influence runoff, but the differences of autocorrelation among local gaged sites reflects how small-scale events likely influence runoff in the semiarid region contributing to local inflow (Table 2). The 12-month lag is tantamount to year-to-year autocorrelation and is high for overall local inflows and the San Rafael River gage simply because these represent larger drainage areas (Table 1).

Simple linear regression analyses between the monthly local inflow and the gaged tributary flows for their common periods of record, yielded low coefficients; the highest coefficient of determination (R^2) equals 0.16 for local inflow with San Rafael River flow and the other two are less than 0.01. Simple linear regression of local inflow with the sum of the three tributary gages for the period of record that overlaps all three tributary gages explained 27% of the variance. Multiple regression of local inflow with the three gages for the corresponding period from October 1948 through September 1955 during which all gages have records resulted in the following regression (adjusted R^2 [adj R^2] = 0.25)

$$Q_l = 8.1 + (-0.00437)Q_{\mathrm{Esc}} + 0.000803Q_{\mathrm{DD}} + 0.00306Q_{\mathrm{SR}} \tag{4}$$

Q_l is local inflow, Q_{Esc}, Q_{DD}, and Q_{SR} are monthly flows (Mm^3/month) at Escalante, Dirty Devil, and San Rafael Rivers, respectively (Table 2). Only the coefficient for Q_{SR} was significant, with $p = 5.7 \times 10^{-6}$, which is reasonable because its drainage area represents more of the local inflow area. Overall, the correlation and regression results reflect that many factors control local inflow and that the gages measure flow from only a portion of the total tributary area.

Indicator variable regression (Neter *et al.*, 1985) with Q_{SR}, with monthly indicators equaling 1 for observations during that month and 0 otherwise, helps to control for differences in the average flow from the tributary gage and the local inflow among months. This regression (data not shown) showed there is a substantial difference in flows among months, but it implicitly assumes the slope of the flow relation is constant among months. The slope and intercept were 0.00293 and 19.5 Mm^3/month, just slightly different than Equation (4), but the indicator variables for April, May, June, July, and August showed the flows differed from the monthly average by −140, −94.6, 71.3, 156, and 52.7 Mm^3/month (p = 1.21×10^{-5}, 0.003, 0.032, 1.63×10^{-6}, and 0.081, respectively). The adj R^2 increased to 0.715.

To test whether the flow relation might differ by month, indicators for the four months which had significant coefficients ($p < 0.05$) were retained, and used to create an interaction term wherein the indicator, 0 or 1, is multiplied by the flow for each month. The interaction term will account for how the relationship varies among months. The resulting equation is

$$\begin{aligned} Q_l = {} & 16.2 + 0.00502Q_{\mathrm{SR}} - 83.5\mathrm{APR} - 123.\mathrm{MAY} \\ & + 100.\mathrm{JUN} + 105.\mathrm{JUL} - 0.010Q_{\mathrm{SR}}\mathrm{APR} \\ & - 0.00068Q_{\mathrm{SR}}\mathrm{MAY} - 0.00281Q_{\mathrm{SR}}\mathrm{JUN} \\ & + .005485Q_{\mathrm{SR}}\mathrm{JUL} \end{aligned} \tag{5}$$

The adj R^2 is 0.78, indicating that Equation (5) explains a substantial amount of variance. The regression coefficients for the monthly indicator variables reflect the difference in average local inflow by month. The regression coefficient of the month and San Rafael River flow interaction terms is the amount that the slope of the flow relationship for that month differs from the overall regression slope. The differing relations by month reflect the snowmelt and beginning of the summer monsoon season. The Durbin-Watson statistic is 2.37 which indicates the residuals are not autocorrelated and the relationship may be used for further analysis without adjustment (Neter *et al.*, 1985).

With some exceptions, the local inflow predicted with this regression developed herein tends to be less than the USBR estimate of local inflow (Figure 5).

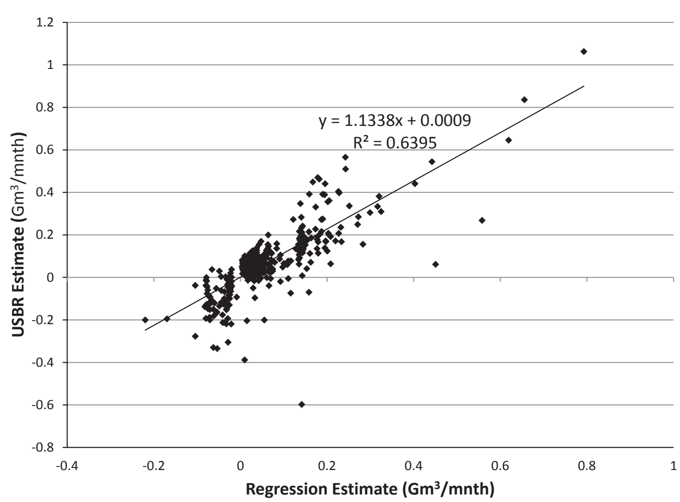

FIGURE 5. Relation of the U.S. Bureau of Reclamation Estimated Local Inflow and the Local Inflow Based on Equation (5).

The slope is 1.13 which indicates the USBR estimate is about 13% higher. Several high outliers occur for situations that regression prediction is higher than the USBR estimate. In May 1984, high predicted local inflow causes a large residual because the negative coefficient and high river flow caused the USBR to estimate -0.616 Gm^3/month (Figure 5).

Precipitation and Evaporation

Annual precipitation at Page, Arizona (Figure 1), equals 16.4 cm/yr (Western Regional Climate Center, http://www.wrcc.dri.edu/cgi-bin/cliMAIN.pl?azpage, accessed November 16, 2010) (Figure 6). The monthly precipitation volume, equaling the average precipitation over the monthly reservoir areas, was used directly in the water balance analysis.

The USBR estimates monthly evaporation based on a net value of 121 cm/yr distributed as shown in Figure 6 (Clayton, 2004; R. Clayton, "Reclamation Evaporation Methodology for Lake Powell," unpublished report, U.S. Bureau of Reclamation, Upper Colorado Region, Salt Lake City, Utah, 2008). Net evaporation equals gross evaporation minus pre-reservoir evapotranspiration (ET). Gross evaporation is 176 cm/yr and pre-reservoir ET loss within the inundated reservoir area (at full pool 664 km^2) is 0.280 Gm^3/yr (Jacoby *et al.*, 1977). Pre-reservoir ET loss included evaporation loss from the hillside area (0.0784 Gm^3/yr), transpiration from the riparian vegetation, and evaporation from the river surface. Jacoby *et al.* (1977) assumed that an average water surface area of 506 km^2 for a two-thirds full reservoir is representative of the long-term operations of the reservoir. At this area, the total gross evaporation loss is 0.893 Gm^3/yr. Subtracting the 0.280 Gm^3/yr pre-reservoir losses from the gross evaporation yields a net evaporation loss of 0.617 Gm^3/yr, which for a surface area of 506 km^2 is 1.22 meters per year (m/yr), the rate used for CRSS modeling (USBR, 2007).

Reservoir Water Balance

As of September 2009, the estimated deterministic cumulative bank storage is 15.0 Gm^3 (Figure 7). Considering stochasticity, the cumulative bank storage ranges from 11.9 to 18.8 Gm^3 for the fifth through ninety-fifth percentile estimate (Figure 7). The highest and lowest bank storage estimates of 125 traces were 20.8 and 10.3 Gm^3, respectively.

The deterministic cumulative bank storage first exceeded 14.8 Gm^3 in June 1983. Since that time it has fluctuated between 17.3 Gm^3 and slightly less than 14.8 Gm^3, the lowest value of which was 14.2 Gm^3 in April 2008, which also coincided with the lowest reservoir volume since its initial filling. From the peak in June 1989, cumulative bank storage decreased slowly until February 1995, losing about 1.0 Gm^3. Reservoir levels then began to increase and the cumulative bank storage crested again in December 1997 and remained mostly constant for several years while the reservoir remained at higher levels. Cumulative bank storage then began to decrease to its nadir in April 2008, losing about 2.47 Gm^3 in 10 years during a long-term drought.

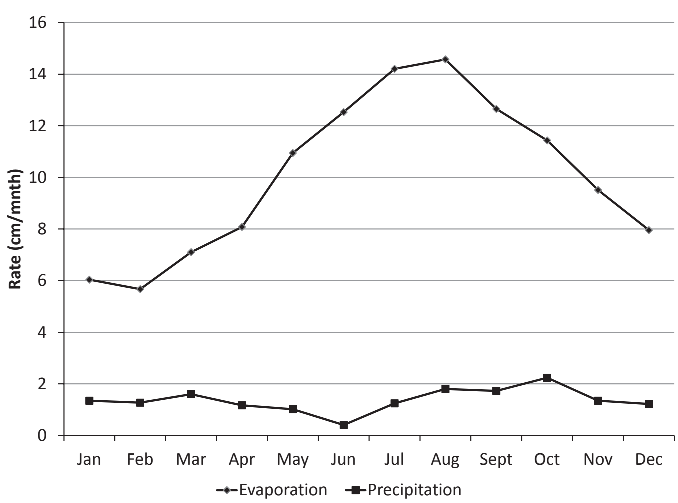

FIGURE 6. Evaporation and Precipitation Rates for Lake Powell as Simulated in the Water Balance. Evaporation as simulated by USBR (2007) and precipitation for 1957 through 2005 (Western Regional Climate Center, http://www.wrcc.dri.edu/cgi-bin/cliMAIN.pl?azpage, accessed November 30, 2010).

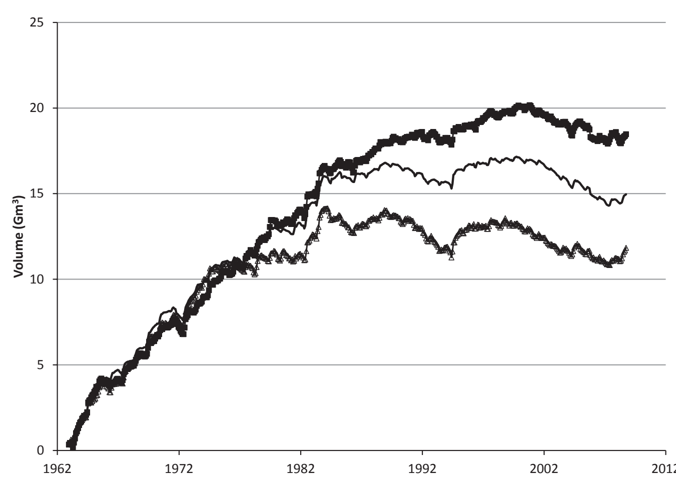

FIGURE 7. Deterministic and 5th and 95th Percentile Cumulative Bank Storage Curves.

The fluctuations since 1984 suggest that cumulative bank storage decreases with the reservoir level suggesting that water returns to the reservoir, but at a much lower rate than it had entered the banks during filling. The following regression of bank and reservoir storage over the entire time period shows that monthly bank storage is less when cumulative bank storage is higher and higher when the reservoir storage is high.

$$\begin{aligned}\Delta V_b = {} & 0.0448 - 0.0107V\mathrm{b} + 0.0410\Delta V + 0.00457V \\ & + 0.0301\mathrm{Fall} - 0.0346\mathrm{Spring} + 0.112\mathrm{Summer}\end{aligned} \quad (6)$$

V_b is bank storage, V is reservoir storage, Δ means "change in," and Spring, Fall, and Summer are indicator variables specifying the season. Adjusted R^2 is 0.49 and all coefficients are highly significant ($p < 0.0016$). The errors are not autocorrelated as the Durbin-Watson statistic is 1.73 (Neter *et al.*, 1985). Bank storage is larger both during months with large changes in reservoir storage and/or a near full reservoir, which reflects the hydraulic gradient for flow into the banks. Based on the sign and coefficients for the seasonal indicator variables, the largest amount of water entering bank storage occurs in the summer and water generally leaves the banks during the spring. Overall, the estimated bank storage values are quite variable which reflects their calculation as a water balance residual that has all of the uncertainty in the reservoir water balance relationship.

DISCUSSION

The deterministic bank storage estimate made herein is about two-thirds of the 2010 USBR estimate. None of the Monte Carlo simulations predicted bank storage as high as the USBR estimate. This estimate is lower because estimated local inflow in the Lake Powell reach is less (Figure 5). The USBR estimate depends on upriver flow which is not as predictive of local inflow as is gaged local tributary flow, which has decreased with time, leading to a decrease in estimated total local inflow. The USBR also uses natural flow estimates; therefore, the bank storage includes error inherent in that estimate as part of its bank storage estimate. Natural flows are higher than historical flows, therefore, bank storage estimates using natural flow are higher, but include water that has been lost to consumptive use.

The USBR changed its long-term bank storage estimate effective January 1, 2012 (Katrina Grantz, USBR Salt Lake City, February 1, 2013, personal communication). In its 24-month operations report issued 12/9/2011, the USBR reported that bank storage at the end of November 2011 equaled 23.5 Gm^3 but the next month's report, issued 1/12/2012, reported that bank storage equaled 6.6 Gm^3 (Operation Plan for Colorado River System Reservoirs (24-Month Study), http://www.usbr.gov/lc/region/g4000/24mo/, accessed August 8, 2012). The methodology has not been documented in a formal study (Katrina Grantz, USBR Salt Lake City, February 1, 2013, personal communication).

The deterministic bank storage accumulated since 1964 is almost equivalent to a year's worth of inflow to the reservoir. There are three apparent questions to consider about bank storage or seepage in Lake Powell: (1) where does the water go?; (2) will it return to the reservoir as the reservoir lowers?; and (3) how much more water will be stored in the banks?

Where Does the Water Go?

The reservoir water seeps into the Navajo Sandstone and other sedimentary rocks around the reservoir. When the reservoir is full, the total inundated river length is about 299 km with about 233 km along the Colorado River and 66 km along the San Juan River portion of the reservoir, for a total river bank of about 599 km. The unsaturated thickness from the original river level to an approximate average depth is about 152 m at the dam (Thomas, 1986). The pre-reservoir groundwater level sloped up from the river to a point where the groundwater level approximates the full reservoir elevation, about 8 km from the river at the dam (Blanchard, 1986). The volume of an unsaturated wedge in the sandstone surrounding the reservoir is about 180 Gm^3 accounting for the riverbed slope. If the sandstone porosity varies between 0.01 and 0.1 (Anderson and Woessner, 1992), the available bank storage is about 1.80-18.0 Gm^3 below the pre-reservoir groundwater level.

Groundwater flowing toward the river likely no longer reaches it due to backwater caused by the rising reservoir because of backwater into the sediments above the reservoir level (Blanchard, 1986; Thomas, 1986), as seen in rising groundwater levels near the reservoir (Figures 8 and 9). Thomas' (1986) groundwater model simulations found that a long-term equilibrium would result in about 400 years with half of the total bank storage having occurred by 1983. He projected that 36 and 57% of the second half of the equilibrium storage would be reached within 50 and 100 years, respectively, although his estimates do not account for fluctuation in the reservoir level. This suggests that another 14.8 Gm^3 will accumulate in the banks over the next 400 years. Because

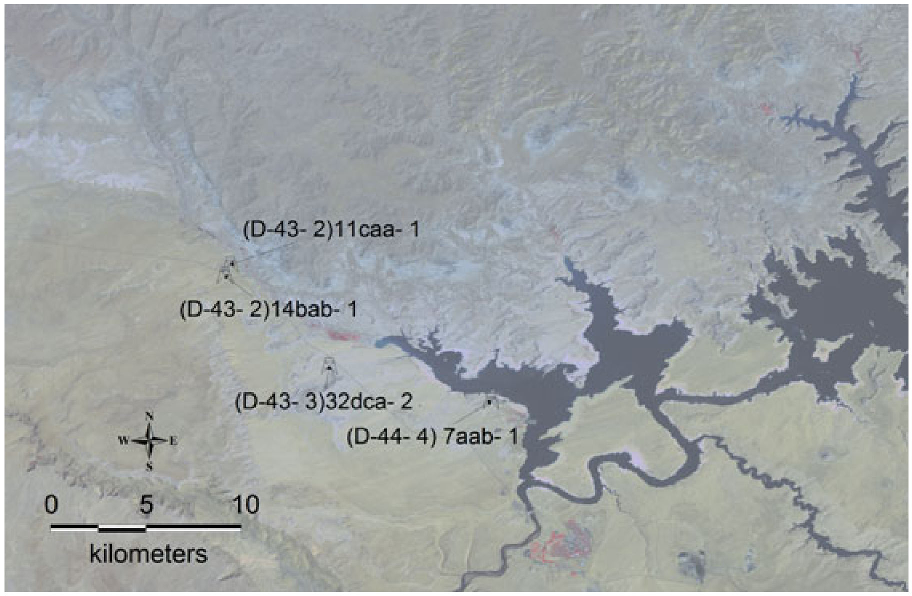

FIGURE 8. Location of Four Monitoring Wells Near Lake Powell. All wells are northwest of the dam within 10 km of the reservoir. Site map upper right is the Upper Colorado River region, see Figure 1.

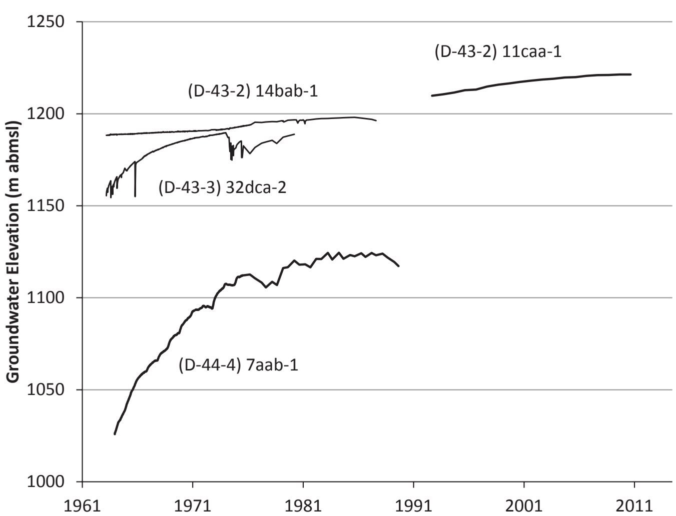

FIGURE 9. Groundwater Elevation at Wells Monitored Near Lake Powell (see Figure 8). The full lake level is 1,128 m above mean sea level.

the groundwater levels near the reservoir have already risen to the level of equilibrium storage, most of the future bank storage will be from groundwater inflow.

Will the Water Return to the Reservoir?

Water returns to the reservoir as the reservoir level lowers because the gradient reverses near the reservoir, but the regression relations indicate the rate of return is much lower than for water flowing into the banks. Because the sandstone dips downward to the north, water in the banks to the north may have barriers to overcome to return to the reservoir or river system and some may flow past a point where geology prevents its return.

Thomas (1986) indicated that the reservoir had not yet affected regional flow patterns in the sandstone, but that "levels in wells within 1 mile of the lake shoreline indicate that the direction of groundwater movement near the lake reverses following the seasonal fluctuations of the lake level" (Thomas, 1986, p. 16). His groundwater model showed that within 20 years, groundwater levels will have increased 100 m near the downstream end of the reservoir and more than 7 m, up to 40 km from the Colorado River. However, the potentiometric surface would still slope toward the river, as found by Blanchard (1986). Neither simulations nor observations suggest a groundwater divide has or will form to prevent water from returning to the reservoir; a divide could form downgradient of the dam causing water to return to the river far below the dam (Thomas, 1986).

How Much More Water Can Be Stored in the Bank of Lake Powell?

The answer depends on the combination of seepage and how groundwater inflow to the Colorado River and inundated tributaries now backs up due to the reservoir level. If the bank storage was half full in 1983, then full storage is about 29.0 Gm^3 and there is room for an additional 15 Gm^3 over the next 400 years. Bank storage accumulates above the full pool elevation because groundwater flowing toward the reservoir mounds up against the inflow to sandstone units from the reservoir as shown by groundwater levels near the downstream end of the reservoir (Figure 9).

Effect of Simulating Reservoir Management Using a Constant Bank Storage Fraction

As noted, the USBR CRSS simulations of the operation of Lake Powell assume that bank storage is 8% of the change in reservoir storage; this value has not changed even though the USBR estimated cumulative bank storage has changed, as noted above. This effectively means that the USBR assumes the storage which accumulates in the reservoir is 8% greater than the measured water volume increase. It also assumes that water returns to the reservoir at the same rate as the reservoir volume lowers and ignores the long-term storage.

The actual fraction that the monthly change in bank storage is of the monthly change in reservoir storage varies substantially but the median negative and positive bank storage fraction has been −0.12 and 0.15 for the entire time period (Figure 10). Since 1983, the median fraction has been 0.12, both positive and negative, indicating more water than simulated both enters and leaves the reservoir bank storage.

Bank storage returns to the reservoir slowly as the reservoir volume decreases — much slower than the water flowed to the banks while it was filling — because the reservoir levels generally decrease more slowly than they increase. Since 1983, bank storage decreased at small rates for two periods; the first was quickly made up as the reservoir filled in 1998 (Figure 7). Statistics and modeling indicate that the bank storage will continue to increase if the reservoir returns to high levels. If the long-term trend is toward low reservoir levels some of the bank storage may return, though some may be downstream from the dam. If this were the case, CRSS may actually underestimate the long-term bank storage return.

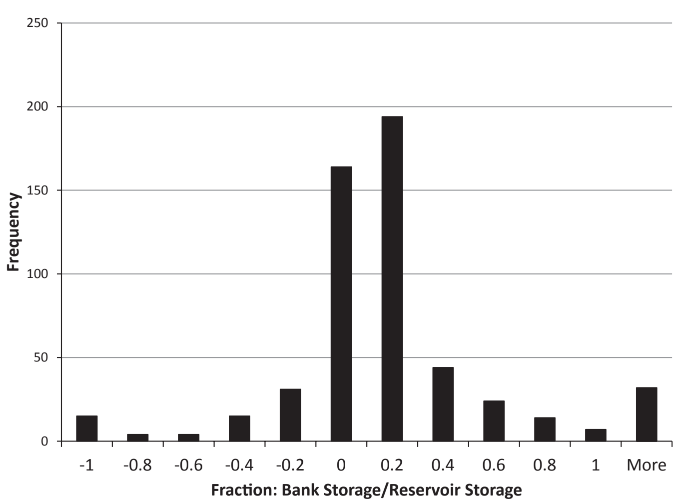

FIGURE 10. Histogram of the Fraction That Change in Bank Storage by Month Is of Change in Reservoir Storage by Month. The figure does not include higher or lower fractions which primarily occurred during periods with very small reservoir storage changes.

Effect of Simulating Net Evaporation

Assuming evaporation based on the difference between gross evaporation and the salvaged losses for a particular reservoir storage amount biases the evaporation estimate. When the reservoir volume is low, the surface area is low and the water considered salvaged by a full reservoir will actually still be lost in the river channel and hillside areas not inundated. It is more accurate to estimate gross evaporation and account for precipitation onto the water surface and proportional salvaged ET. Salvaged ET is the rate determined for a full reservoir adjusted by the proportion of the reservoir area actually inundated.

The effect USBR's assumptions would have on CRSS results depends on the average reservoir level. During low reservoir stage, the assumption would underestimate the actual evaporation loss because use of a net value assumes water is not lost that actually is lost. When the reservoir is near full, the method may estimate a net value that is too high because the full pool may actually salvage more pre-reservoir losses than assumed. In the future if the reservoir level remains low, the USBR assumption may allow too little evaporation loss and allow the USBR to overpredict reservoir storage.

Local Inflow between Lake Powell and Lees Ferry

The only way to estimate inflow or outflow between Glen Canyon Dam and the gage at Lees Ferry is to determine the difference between the gage and the dam releases. U.S. Geological Survey (USGS) gage number 09379910, Colorado River below Glen Canyon Dam, has operated intermittently just 1,370 m downstream from the dam, from October 1989 to March 1993 and from March 2000 to September 2002. The flow difference between 1989 and 1993 averaged −0.397 cubic meters per second (cms) and between 2000 and 2002 averaged 3.43 cms.

Using flow measurements through the power plant since 2000, the USBR has estimated seepage between the dam and Lees Ferry gage to be about 187.5 Mm^3/yr, which is about 5.95 cms (Rick Clayton, USBR Salt Lake City, November 29, 2010, personal communication). Seepage to the river below the dam from the Navajo Sandstone below the dam began as early as 1983, after 20 years of reservoir filling, and continues to this day (David Wegner, former Program Manager, Glen Canyon Environmental Studies, August 6, 2011, personal communication).

Water that enters the river between the dam and Lees Ferry is essentially returning bank storage that is counted as part of the discharge from the system. The fact there is seepage into the river below the dam indicates the estimated bank storage is a net value, with some of the water entering the reservoir returning to the system. Seepage below the dam is accounted for at the Lees Ferry gage, therefore, it should not affect the water balance or the bank storage estimate unless seeps discharge below Lees Ferry, more than 25.6 km below the dam.

CONCLUSION

Lake Powell has lost or stored more than 14.8 Gm^3 of water in its banks since the bypass tubes were closed in 1963 according to the water balance presented above. Groundwater models completed by others had predicted that amount and indicated that about that much more will flow into the banks within 400 years. The accumulating bank storage is not an error in the estimate of local inflow.

Water loss to the banks has lessened but has not yet reached steady state. Equilibrium may require as much as 400 years, but much of the loss may actually be due to groundwater not reaching the reservoir rather than to water flowing into the banks from the reservoir. Bank storage returns to the reservoir when the reservoir is low but accumulates much faster when the reservoir fills. Keeping the reservoir lower or even empty may salvage water lost to the banks, reduce evaporation, and possibly recover up to 14.8 Gm^3 of water already stored there. Keeping the reservoir near empty, however, could affect other values generated in the river, but these are beyond the scope of this article.

Simulating the reservoir operations with a constant bank storage estimate that returns all of the water to the reservoir fails to account for some system losses. Using the higher bank storage fraction determined herein and also accounting for cumulative bank storage separately would provide superior estimates and improve the reservoir management.

Bank storage relationships are not constant with time and factors not considered in the simple water balance calculation affect the amount of water lost to and returning from the banks. A network of monitoring wells and piezometers around the reservoir would improve the understanding and management of bank storage. Because of differing transmissivity values, there may be levels at which more water discharges into the banks. A detailed groundwater model of bank storage linked to the river simulation models could help to integrate bank storage into surface water management. Detailed monitoring and modeling could allow the USBR to optimize its reservoir storage with bank storage rather than assuming that all water lost will return.

Local inflow and evaporation estimates have a large uncertainty. Installing gaging stations on as many local tributaries as feasible and updating the evaporation data so that year-to-year and seasonal variability is better known would also improve the understanding of reservoir water losses. Both of these water budget components could change with time due to a changing climate, a factor which increases the importance of improving the data. Including the updated estimates in CRSS would change the management of bank storage and evaporation which currently leads to an estimate of more water availability, especially as the reservoir levels trend downward. The evaporation rate accounts for too much salvage of pre-reservoir water loss; long-term downward trends in the reservoir level could result in more predicted bank storage returning to the reservoir than is realistic.

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