Water Balance of Lake Powell
An Assessment of Groundwater Seepage and Evaporation

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Executive Summary

Glen Canyon is a deep, majestic canyon with numerous small seeps and springs flowing into the Colorado and San Juan Rivers. It is located in southern Utah and northern Arizona Flow from the seeps into the river were small. Evaporation from the vegetation and river surface was a small loss to the flow of the rivers through canyon. Lake Powell, the reservoir formed by Glen Canyon Dam, loses water to both evaporation and bank seepage. Many argue that the increase represented by this loss over the natural loss is excessive and represents an important reason to consider draining the reservoir.

This study had five objectives. The first is to provide background on how the Bureau of Reclamation estimates losses and how those losses are included in the "Law of the River". The second and primary objective is an estimate of the net total water loss and its component parts of evaporation and bank seepage due to Lake Powell. The third objective is to determine the range of evaporation estimates and estimate the impact that those estimates have on the operations of the river. The fourth objective is an independent estimate of seepage loss with a groundwater model to verify the water balance estimate, to extend the prediction into the future and to assess how much seepage returns to the reservoir. The fifth objective is to determine whether losses are increasing, decreasing or static.

The Bureau of Reclamation uses monthly evaporation coefficients based on a mass transfer analysis. It is a reasonable method that yields approximately 69 inches per year of gross evaporation loss, but they reduce the estimates by pre-dam evaporation which is not appropriate. Their final estimates are less than 48 inches and they have underestimated evaporation since 1963 by more than 5,000,000 af. For monthly operations, they should consider the gross reservoir evaporation.

Based on the water balance analysis, cumulative seepage peaked around 1983 at 11,000,000 af after which it dropped to less than 10,000,000 af in 1992. Between 1992 and 1997, cumulative seepage fluctuated between 10,000,000 and 10,500,000 af as the reservoir rose and fell. A groundwater model confirmed the mechanics and pattern of seepage loss, but underestimated the quantity by about 50%. The primary reasons for the difference in seepage quantity are errors in the measurements and estimates used in the water balance analysis, the sensitivity of total seepage in the groundwater model to the storage coefficients, and potential inaccuracies in the assumed gradient of the groundwater surface near the reservoir.

Running the groundwater model for 1500 years into the future suggests that equilibrium will occur in about 1400 years and about 21,600,000 af will be lost. Over the total period, half of the seepage is lost in the first 37 years because the available storage becomes filled and the gradient has from the reservoir to the bank has significantly decreased. The equilibrium depends on the assumed reservoir level (3680'). The actual equilibrium will be dynamic with water movement into and out of the banks. The loss rate drops significantly and the time to equilibrium is very long because water moves into the sandstone canyon walls very slowly. The natural groundwater flow is toward the river from both south and north, therefore there is no continuous flow away from the reservoir to the south or north.

Evaporation losses up to 1997 are about double those to seepage. Total evaporation from the reservoir until 1997 has been about 23,500,000 af. Evaporation without the reservoir was probably about 102,000 af/year on the 18,000 acres of riparian area within the canyons of the two rivers. Since dam closure, a high estimate is that 3,500,000 af would have been lost without the reservoir. The net evaporation loss due to the reservoir is about 20,000,000 af since 1963. The evaporation loss is about 2.1 percent of a full reservoir (27,000,000 af).

Total losses from Lake Powell due to the reservoir have been about 30,000,000 af. This is about 2 1/4 years of average annual flow at the Lee's Ferry gage or 6.3 percent of the annual flow. It is also about 11 percent more than the volume of the reservoir when full. Considered as an average over 34 years, the annual loss is about 3.3 percent of a full reservoir. The evaporation loss will continue at a rate of about 570,000 af/year after considering predam losses. The seepage loss will rapidly slow down and become much less than has been observed for the first 34 years.

In conclusion, seepage and evaporation represent a substantial loss of water from the Colorado River system. As the demands on the river's flows exceed the annual flow prior to reservoir losses, the loss as a percentage of inflow will likely change. Evaporation will continue into the future at current rates but seepage will decrease as equilibrium is approached. Only a detailed operations analysis will resolve the question of whether the increased flow in the river that would result from draining Lake Powell would be more important than the decreased certainty caused by the lost reservoir storage. In addition to the Bureau of Reclamation changing the way it accounts for evaporation in its operations models, the completion of a detailed operations analysis into the future is the primary recommendation from this report. Only such an analysis will allow an assessment of the tradeoffs between water deliveries and drought uncertainties thought to be mitigated by Lake Powell.

Introduction

Glen Canyon is a deep, majestic canyon with numerous small seeps and springs flowing into the Colorado and San Juan Rivers. It is located in southern Utah and northern Arizona (Figure 1). Flow from the seeps into the river were small. Towering cottonwood groves and exotic salt cedar stands dominated the riparian vegetation. Evaporation from the vegetation and river surface was a small loss to the flow of the rivers through canyon.

Lake Powell, the reservoir formed by Glen Canyon Dam, loses water to both evaporation and bank seepage. Many argue that the increase represented by this loss over the natural loss is excessive and represents an important reason to consider draining the reservoir. Loss estimates vary and are subject to some controversy (Dawdy, 1991). The U.S. Bureau of Reclamation (USBR) does not consider bank storage in its operations of the Colorado River. This study quantifies total losses from Lake Powell and segregates that loss between evaporation and bank storage.

The Bureau of Reclamation and most Colorado River basin water purveyors consider Lake Powell to be an essential part of the Colorado River water resources system. It stores water for drought years. It is important to document water lost to the system caused by Lake Powell in light of the reservoir's role in storage and reducing uncertainty in river deliveries to the lower basin. Managers in the upper Colorado basin believe that the reservoir is essential to meet its 75 million acre-ft, 10-year obligation to the lower basin.

Objectives

There are five objectives of this study. The first objective is to provide background how the Bureau estimates losses and how those losses are included in the "Law of the River". The second and primary objective is an estimate of the net total water loss and its component parts of evaporation and bank seepage due to Lake Powell. The third objective is to determine the range of evaporation estimates and the impact that those estimates have on the operations of the river. The fourth objective is an independent estimate of seepage loss to verify the water balance estimate and to extend the prediction into the future and to assess how much seepage returns to the reservoir. The fifth objective is to determine whether losses are increasing, decreasing or static.

Background

The USBR operates the system of dams and reservoirs on the Colorado River on behalf of the Secretary of the Interior according to the Law of the River. The Law of the River is a collection of statutes and court cases that divide the waters of the river. Ingram et al (1991), a book published by the National Academy of Sciences, provide an excellent discussion of the application of the Law with respect to Lake Powell. Only a basic outline of the Law will be provided here along with an assessment of how reservoir losses can impact the operations of the river.

The Colorado River Compact of 1922 involved the seven basin states including Colorado, Wyoming, Utah, New Mexico and a small portion of Arizona in the Upper Basin and Nevada, the remainder of Arizona and California in the Lower Basin. The Upper and Lower Basin dividing line on the river was at Lee's Ferry at the mouth of Glen Canyon. Based on highly erroneous estimates (Dawdy, 1991), the negotiators assumed that the river flow equaled 16,000,000 af/year at Lee's Ferry. In an effort to provide "equity between the upper and lower basins" (Ingram et al, 1991, page 13), the flows were divided to provide for 7,500,000 af/y for each basin. Surpluses went to the lower basin. The compact's text requires the Upper Basin to provide at least 75,000,000 af to the Lower Basin over any ten-year period[1]. Because the text requires a delivery to the lower basin rather than specifying a consumptive use for the Upper Basin, all losses from Lake Powell should be charged to the Upper Basin. For this reason, any saved losses will benefit the Upper Basin first and to the extent that savings lead to a surplus will benefit the Lower Basin. Incorrect loss estimates will lead to an improper apportionment of losses among Upper Basin states.

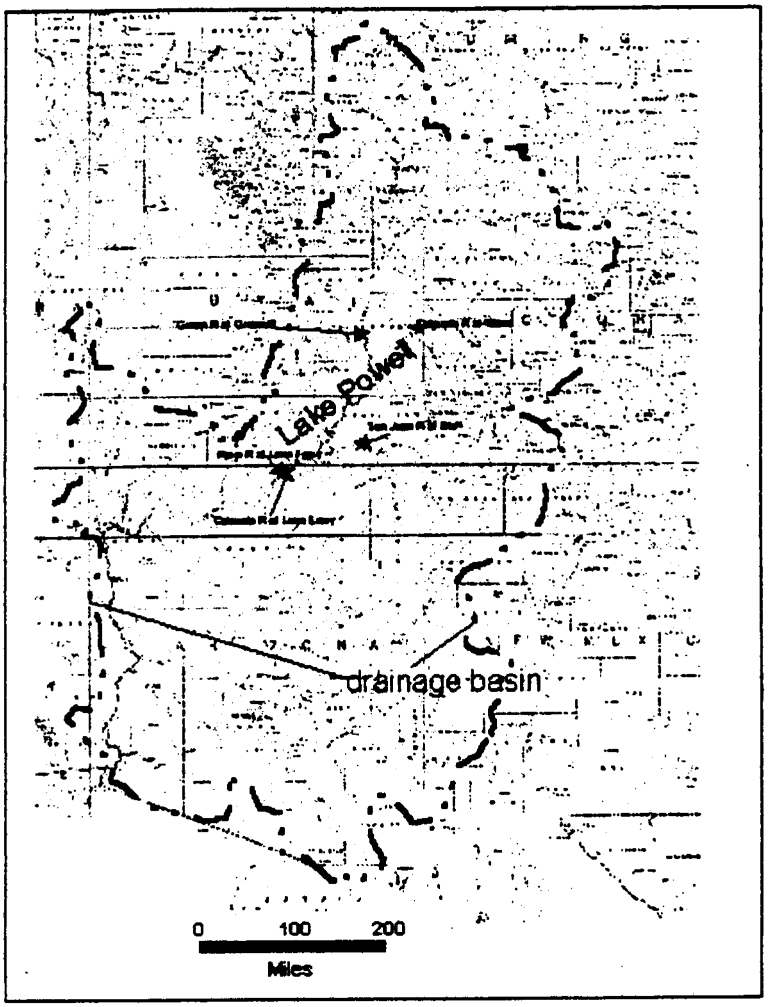

Figure 1: Location of Lake Powell, the Colorado River drainage basin, and gaging stations used in this study.

Many assume that Lake Powell reservoir storage is essential for drought security. If losses from the reservoir approach 10 to 15 percent of the required Lee's Ferry delivery, as will be demonstrated below, it is important to assess the losses in conjunction with drought flows.

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[1]Colorado River Compact, 1922, Article III(d): "The States of the Upper Division will not cause the flow of the river at Lee (sic) Ferry to be depleted below an aggregate of 75,000,000 acre-feet for any period of ten consecutive years reckoned in continuing progressive series beginning with the first day of October next succeeding the ratification of the compact."

Water Balance Method of Analysis

Total Water Loss

Total water loss results from a simple calculation of reservoir inflow minus outflow minus storage change. Inflow, Qi, is the sum of stream flow into and precipitation onto the reservoir, outflow, Q_o, is the flow through the dam, and storage change, Δs, is the change in volume associated with changing reservoir levels. If those three values are estimated, the error term is the unmeasured portion of the balance which included evaporation and bank storage changes. Total loss is the error term in the water balance equation.

$$\Delta S = Q_i - Q_o - e \tag{1}$$

The storage change and the inflow and outflow are measured by the Bureau of Reclamation and the US Geological Survey, respectively. e is the loss and includes both evaporation and groundwater seepage. Breaking it down into its component parts:

$$\Delta S = Q_i - Q_o - E - GW \tag{2}$$

In this equation, ΔS is positive when inflow exceeds the sum of outflow, evaporation E, and seepage GW. Both E and GW are positive values to represent losses to the reservoir. Qi consists of mainstem river flows on the Green River, Colorado River and San Juan River (see Appendix 1 for a listing of the stations), local inflow and precipitation on the reservoir surface.

Local Inflow

The USBR estimates the local inflow based on monthly proportions developed from the flow differences between the summed upstream gages and the Lee's Ferry gage. For this water balance analysis, the author examined these proportions and developed more accurate monthly estimates. The regression analysis presented in Appendix 1 is a 23% improvement over the proportional estimates used by the USBR. The regression estimates also do not bias the estimates as the proportions do when considering the 1929 to 1962 flow data. The USBR proportions add a negative bias of about 4900 af/mnth or about 59000 af/year. In a water balance analysis, the error term is usually the seepage term and would be underestimated by about 4900 af/mnth using the USBR proportions. The regression relations showed significant scatter and different correlations between months that corresponds to the varying seasonal effects in the watersheds controlling local inflow and the watershed controlling snowmelt runoff to the mainstream gages.

The regression equations occasionally predict negative inflows to Lake Powell. This is expected because of the long distance between the upstream gages and Lee's Ferry. During dry periods, the river losses flow resulting in a negative inflow. With Lake Powell, the total river length is cut by about a third, but there is still a significant loss in the river before the inflow reaches the gages. During dry periods, the reservoir is dropping and predicted evaporation accounts only for losses from the water surface. Additional loss occurs through evaporation from wetted shoreline and exfiltration of reservoir water infiltrating near the shore. The negative inflow also helps to account for this loss.

Precipitation

Precipitation on the water surface was based on average monthly precipitation (USBR, 1986). USBR based their estimate on seven nearby stations, Page, Wahweap, Bullfrog, Canyonlands, Hite, Mexican Hat, and Natural Bridges[2] having data from 1963 to 1980. USBR adjusted downward the averages based on a linear regression of precipitation and elevation because the reservoir elevation is less than at the stations. The data from these stations is incomplete with many missing monthly data points, therefore the USBR's estimates were accepted for use in this study. These values are:

Month	Jan	Feb	Mar	April	May	June	July	Aug	Sept	Oct	Nov	Dec
Inches	0.6	0.6	0.8	0.8	0.8	0.8	1.0	1.0	1.0	0.8	0.8	0.8
Feet	.05	.05	.067	.067	.067	.067	.083	.083	.083	.067	.067	.067

The average total annual precipitation is 6.33 inches. Surface area ranges from 28,000 to 162,000 acres, therefore an error of one inch for a month has a maximum error of 13,500 af/mnth. This error would be a doubling of the average for any month. The total error compared to predicted seepage and evaporation losses discussed below is very small. Additional analysis to create a monthly time series of actual precipitation is not warranted because improved estimates would not increase the precision of the water balance. Errors inherent in other estimates used in the calculations exceed the full magnitude of precipitation values.

According to Dawdy (1991) in a book published by the National Academy of Sciences, precipitation on the lake surface is not included in the USBR's daily modeling of the lake. However, the USBR does improperly subtract a portion of the precipitation that would have fallen on the terraces and drained into the river prior to construction of the reservoir (USBR, 1986). See the evaporation discussion below. The use of monthly estimates herein represents an improved estimate.

Outflow

The USBR estimates releases Lake Powell using hydropower rating curves (Dawdy, 1991). Dawdy expressed reservations with the outflow estimates:

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[2]The USBR provides no station number for these gages.

> Discharge is computed incorrectly based on turbine ratings. The discharge at the Lee's Ferry gaging station of the USGS should be used. If more accuracy is required, then a study should be undertaken to improve the accuracy at that station. If turbine ratings are used for day-to-day operations, then each turbine should be calibrated based on the USGS gage. The gravel bars immediately below Glen Canyon Dam should affect the different turbine ratings differently. A study should be undertaken to determine whether, in fact, some turbines are more efficient and what can be done to improve the performance of the less effective ones. (Dawdy, 1991, page 46)

Consideration of the accuracy of the Bureau's estimates or the calibration of the turbines is beyond the scope of this report. This study uses flows measured at the Lee's Ferry gage.

Evaporation Estimates

Because neither evaporation nor seepage can be measured, alternative estimates are required. Most reservoir balance models assume all losses are to evaporation and that any seepage is bank storage which will return to the reservoir. A purpose of this report is to assess the differential losses between evaporation and seepage. This section estimates evaporation losses from the reservoir and considers the USBR methodology.

Dawdy (1991) expressed substantial concerns with the methods used by the USBR for modeling evaporation from the reservoir:

> Evaporation loss from Lake Powell was assumed by La Rue to be about five acre-feet per acre per year, which amounts to about 750,000 acre-feet per year for the approximately 150,000 acres of surface area of the lake. Lake Mead evaporation was found to be about 7 feet per year. A similar evaporation at Glen Canyon would produce about 1 million acre-feet of evaporation per year. The U.S. Bureau of Reclamation (USBR) appears to use slightly under four feet per year, with a constant distribution in the year, irrespective of climate variation. USBR computes evaporation as a function of stage, however, so that their computed evaporation varies from 560,000 acre-feet for 1989 to 633,000 acre-feet for 1983. The U.S. Weather Bureau (USWB) estimates 80 inches per year for a Class A pan and a coefficient of .68 to convert to lake evaporation for an average loss of 4.5 feet, or about 15% higher than the USBR figure, which would give 650,000-730,000 acre-feet of loss per year. The USWB says 74% of the evaporation should be in May through October, USBR shows only 63%. Therefore, if lake evaporation is underestimated, it is in the spring runoff and summer months, when the reservoir will be highest in stage. (Dawdy, page 45, citations omitted)

The Bureau of Reclamation currently estimates evaporation based on a report prepared by the Upper Colorado Regional office in 1986 (USBR, 1986). The report used data collected during the 1970s to estimate evaporation based on vapor pressure deficits and wind speed and evaporation pans. (An U.S. Weather Bureau Class A pan (Shuttleworth, 1992), assumed to have been used here, is circular, 3.97 feet in diameter and 0.84 feet deep, and is made of galvanized iron. It is mounted on a wooden frame platform more than 6 inches above the ground. The pan is filled with water and the change is depth is the pan evaporation. To obtain actual ground surface or reservoir evaporation, the pan evaporation is multiplied by a predetermine pan coefficient.) Their estimates of total evaporation for both methods were about 69 inches which is very reasonable.

Standard vapor pressure estimates of evaporation are based on the following equation:

$$E=N\times u\times(e_o-e_a) \tag{3}$$

In this equation, u is wind speed in miles per hour, N is a mass transfer coefficient, e_o is the saturation vapor pressure in millibars corresponding to the water surface temperature and e_a is the vapor pressure[3] of the air in millibars at a meters above the water surface. All values are daily averages. This quasi-empirical formula considers two meteorological variables that affect evaporation the most. The first is wind speed; the faster the wind blows, the more water evaporates. The second is humidity. Here it is expressed as a gradient of vapor from the water surface to a point above the surface, usually 2 meters. It considers water temperature in that the saturation vapor pressure is the water vapor pressure that would occur if the air was saturated and the temperature equaled that of the water surface. If the air is warm and humid, it is possible for the gradient to be towards the water surface as the air vapor pressure exceed the saturation vapor pressure at the lake surface. This is what causes fog, a very rare event at Lake Powell, to form.

At Lake Mead, the USBR had determined the mass transfer coefficient for estimating reservoir evaporation is 2.65e-3. Using several years of data collected from rafts at three points in the reservoir, they determined the following empirical equation.

$$E=3.27e-3\times u\times(e_o-0.005(H_{max}+H_{min})e_{as}) \tag{4}$$

where H is relative humidity and e_{as} is saturation vapor pressure in millibars at average daily air temperature. The USBR used these data and measurements to calculate the Lake Powell evaporation rate for each month from 1965 to May, 1979. This yielded the following average monthly evaporation rates:

Month	Jan.	Feb.	March	April	May	June	July	Aug.	Sept.	Oct.	Nov.	Dec.
Inches	2.90	2.33	3.42	4.56	6.94	7.61	8.42	9.24	8.31	6.23	4.94	4.35

[3]Each gas that makes up the atmosphere, including water vapor, has causes a certain amount of the total atmospheric pressure. Vapor pressure is the portion of the total atmospheric pressure that consists of water vapor. Saturation vapor pressure is the maximum amount of vapor that the atmosphere can hold at a given temperature. Relative humidity is the percentage that the actual vapor pressure is of the saturation vapor pressure.

Feet	.242	.194	.285	.380	.578	.634	.702	.770	.692	.519	.412	.362

The total is 69.25 inches[4]. These monthly averages were used for calculated gross monthly evaporation for the water balance analysis.

While these data match pan data very well when the pan was on a float in the reservoir, Sellers and Hill (1973) report that the ratio of pan evaporation on land to that on a reservoir is substantial and variable at Lake Mead. From October through January, pan evaporation on the lake exceeds that on the land by as much as 44%, presumably because the lake surface maintains higher temperatures and because there is a greater wind fetch. During the summer, the effect is opposite with the lake pan evaporating as little as 67% of a land pan during June. Sellers and Hill (1973) report the following land pan evaporation for Page:

Month	Jan.	Feb.	March	April	May	June	July	Aug.	Sept.	Oct.	Nov.	Dec.
inches	m	2.81	5.91	8.24	11.19	12.77	13.34	11.35	8.78	5.19	2.41	m

Even without December and January, the pan evaporation total of 82 inches or 6.8 feet far exceeds the pan rates on the reservoir. This may be the source of Dawdy's USWB pan estimate quoted above, however his pan coefficient actually reduced the reservoir evaporation estimate more than the gross estimate from the vapor pressure deficit equation. Land-based evaporation pans provide poor estimates of reservoir evaporation.

There is annual variability in the data. Annual evaporation ranged from 58.32 inches in 1967 to 76.84 inches in 1977 (USBR, 1986). Monthly values range to about +- 30% from the mean. In the long run, the average evaporation loss is accurate and will yield an accurate seepage loss estimate from the reservoir. However, assuming constant monthly evaporation rates may cause short term errors.

For daily operations the USBR reduces the evaporation rate by subtracting evaporation that occurred during the pre-dam years. According to the CRSS methodology:

> The area of the reservoir is determined using an area-elevation table. This area is multiplied by an interpolated effective evaporation coefficient to get effective lake surface evaporation. Then the evaporation of the river (pre-dam), the evaporation of the streamside (pre-dam) and the evaporation of the terrace (pre-dam) are calculated. These three values are then subtracted off of the effective lake surface evaporation to get the total net evaporation for Lake Powell. (Rick Clayton, USBR, Salt Lake City, personal communication, March, 1999)

The USBR uses 68.32 inches of gross evaporation calculated above but then subtracts evaporation that would have occurred had the reservoir not been constructed. It is appropriate to

[4]The USBR (1986) reported the sum as 68.32 inches.

subtract accurate predam evapotranspiration estimates from the gross reservoir evaporation for estimating the losses due to the reservoir. In the opinion of this author, doing so for daily or monthly operations is inappropriate. The difference between gross evaporation and that used by the USBR is very large. The total gross evaporation from 1963 through 1997 is about 23,500,000 af while the USBR estimate[5] is 15,500,000 af. These totals yield rates of 5.69 and 3.95 feet per year, respectively. Between 1979 and 1997 when the reservoir reached almost full conditions, the USBR average evaporation has been 590,000 af/year while the gross evaporation estimates are 850,000 af/year. The annual difference based on rates when the reservoir has a surface area of 150,000 acres is 261,000 af.

As discussed, total evaporation from the reservoir until 1997 has been about 23,500,000 af. However, subtracting off the amount that would have evaporated from the river and streamside is difficult. The total length of rivers inundated by the reservoir on average is about 250 miles. The average width of river and riparian area was about 600 feet[6]. For 250 river miles, about 18,000 acres of riparian and water surface area are now under the reservoir. For reasons described above, the river probably did not evaporate at the same rate as the reservoir level. Being very conservative, we will use the 5.69 feet/year estimate as determined for gross evaporation from the reservoir. Evaporation without the reservoir averaged about 102,000 af/year. Since dam closure, about 3,500,000 af would have been lost from the river areas without the reservoir. The net loss due to the reservoir is about 20,000,000 af since 1963. This is 4,500,000 af more than predicted by the USBR.

The USBR estimates of the amount of river, streamside and terrace evaporation may be incorrect. For river evaporation, they use the gross rate calculated for the reservoir surface for the preexisting river surface. River evaporation may not be the same as reservoir surface evaporation because the river is protected from wind and sun by the canyon walls. The river surface temperature is probably less than the reservoir surface temperature. Both terms in the equations above, vapor pressure deficit and wind speed would be much less than estimated for the reservoir surface.

The USBR estimated streamside evapotranspiration (ET) using the Blaney-Criddle method. The USBR uses the simplest form of Blaney-Criddle:

$$u=fk \tag{5}$$

[5]USBR estimates were provide by Rick Clayton and Randy Peterson of the USBR Salt Lake City office in March, 1999.

[6]Any estimate of width is subject to wide variation, both seasonally and dependent on the canyon width. A 20,000 cfs at 5 feet per second and 10 feet deep will be 400 feet wide. Riparian areas probably average about 100 feet per side.

u is evapotranspiration rate in inches per month, k is an empirical coefficient based on the type of vegetation and f is tp/100 where t is mean monthly air temperature in degrees Fahrenheit and p is the monthly percent of annual daylight hours. They chose 1.1 and 0.9 for the streamside and terrace vegetation, respectively. The report does not provide final calculated values for evaporation rates for streamside or terrace evaporation.

There are at least three problems with the method. First, it was derived for irrigated agriculture. Plant roots are receiving consistent water percolating from just a few inches above. The plants are also evenly spaced. Natural vegetation draws its water from groundwater and is unevenly spaced. Even the USBR acknowledges as much before they proceed to use it. "Consumptive use coefficients have been determined mostly for irrigated crops so there has been little investigation of native vegetation" (USBR, 1986, page 18). Second, it is unlikely that the terrace vegetation, which is upland desert, uses 0.9/1.1 or 82% as much water as streamside vegetation which included willow, cottonwoods and tamarisk in the predam canyon. It is very likely that the USBR subtracts too much evaporation from the gross evaporation to estimate the impact of Lake Powell. This results in the USBR underestimating the water losses caused by Lake Powell by up to 250,000 af/y. Third, the USBR may subtract this evaporation for operations because they assume that it would be a part of the consumptive use in the ungaged inflow estimates (see regression analysis below). Because this evaporation occurred well above the river and did not result from any of the tributary inflow, it is unlikely that it was accounted for in any of the gages.

During 1996, using the USBR estimates, almost 68% of the annual evaporation occurred from May through October. In the citation above, Dawdy states that the USBR uses a uniform evaporation through the year and then states that the USBR assumes that 63% of evaporation occurs from May to October. One of his assumptions must be wrong. A uniform distribution would have 50% of the evaporation occurring from May through October. Either estimate contrasts with values provide to this author for this study by the USBR. For this study, the gross evaporation estimate of 68% estimated above with the mass transfer method was used.

For the water balance analysis, monthly evaporation estimates are the product of a gross evaporation rate and the reservoir surface area. The Bureau reports water level as an elevation. Ferrari (1988) reported new water level, volume and surface area relations based on new capacity measurements. Data were reported at 20-foot intervals. For interpolation, best fit lines were determined with linear regression for the estimation of both surface area and reservoir volume.

A third order regression between area and elevation provided the best predictive equation (R^2=0.999).

$$Area = -28598459.8 + 0.00088201E^3 - 8.41660945E^2 + 26839.44E \quad (6)$$

This equation was used with the time series of monthly water levels to determine a time series of monthly surface area values. Multiplying this by the gross reservoir evaporation provided a time series of evaporation values for use in the water balance.

Reservoir Volume

The change in reservoir volume is an important part of the water balance calculation. It is determined by subtracting the previous month's volume from the current month's volume. The data in Ferrari (1986) was also used to determine a volume:area relationship as follows (R^2=0.999):

$$Volume = -78499209384 + 0.22962807E^3 - 2233.369E^2 + 7248497E \quad (7)$$

Using the water level time series, this equation was used to determine a time series of month-ending volumes.

Calculation of Seepage

Equation 2 can be rearranged to calculate groundwater seepage.

$$GW = Q_i + Q_L + P - Q_o - E - \Delta S \quad (8)$$

Here, GW is seepage (negative values represent returns to the reservoir), Qi is mainstem inflow, Ql is local inflow, P is precipitation, Qo is outflow measured at Lee's Ferry, E is gross evaporation, and ΔS is change in storage. Calculations proceed on a monthly basis beginning in December, 1963.

Results

Seepage began quickly with most values less than 150,000 af/mnth, but with some more than 200,000 af/mnth and one month of 470,000 af/mnth (Figure 2 and Appendix 2). After the cumulative

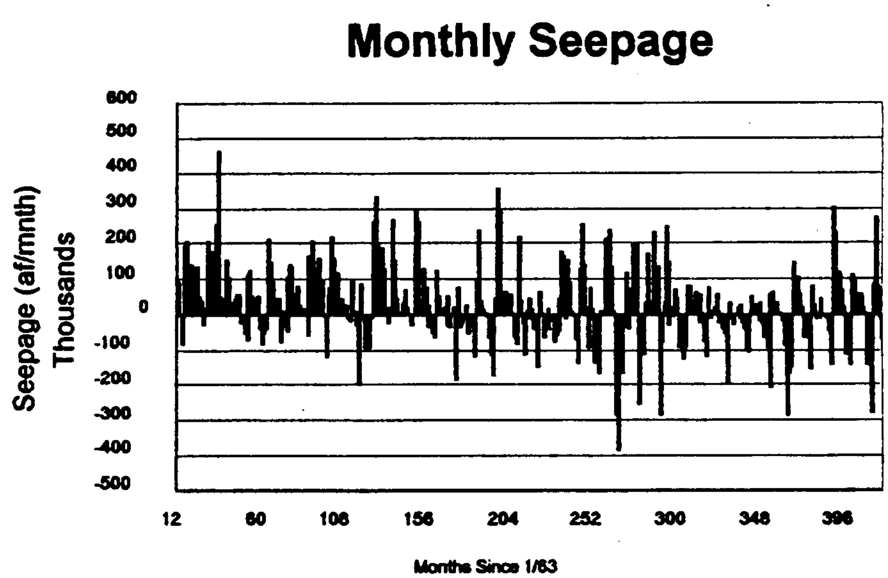

Figure 2: Monthly seepage from Lake Powell by month since 1963.

seepage reached a quasi-equilibrium ranging from 10,000,000 to 11,000,000 af after 1979 (Figure 3, month 193), monthly seepage began to fluctuate around 0.0 (Figure 2). For several years from month 300 to 348 (1988 through 1991), the cumulative storage dropped in response to the dropping reservoir levels. The reservoir dropped to about 3620 around 1992 from its peaks greater than 3700 in 1986. The reservoir had previously been at 3620 in 1978. Between 1965 and 1978, about 8,000,000 af of water had seeped. It peaked at over 11,000,000 af in the early 1980s. By the time the reservoir level dropped back to 3620 in 1992, about 1,200,000 af had returned to the reservoir. This is less than the 3,000,000 af that had seeped as the reservoir rose from 3620 to 3700. This indicates that substantial amounts of the water may be lost. As the reservoir recovered from 1992 to 1997, about 500,000 af of additional seepage occurred.

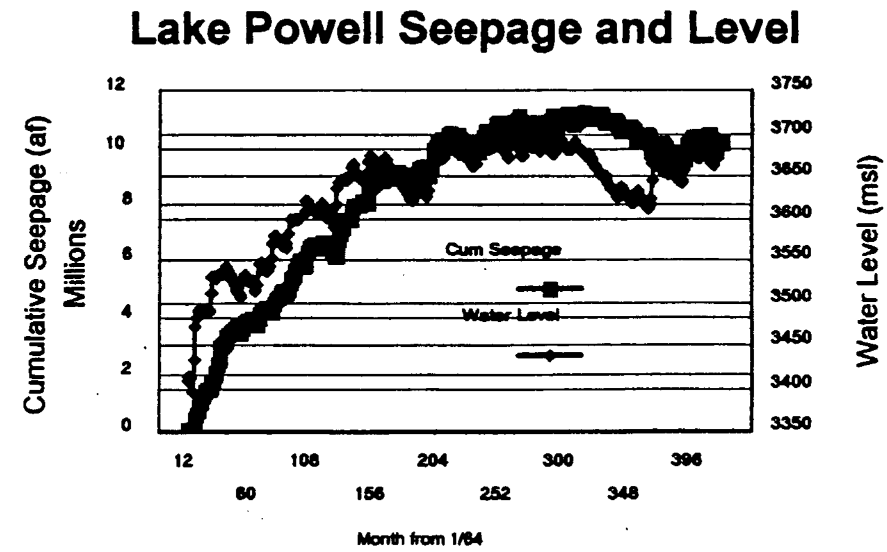

Figure 3: Cumulative seepage and water level.

When reservoirs fill, they cause an increase in head over the groundwater levels in the adjacent aquifers by as much as the total depth in the reservoir. In Lake Powell, the increase ranges to 564 feet (3136 to 3700 feet). As the level rises, flow into previously dry rock begins and depends on the hydraulic conductivity and the presence of fractures. Different layers probably have different seepage rates because of differing aquifer properties. Unfortunately for the analysis of loss into different layers, the reservoir rose to over 3500 within three years and the monthly rises were quite variable. Seepage was rapid and consistent until the levels peaked temporarily at 3550 and the seepage slowed and even was negative (water returned to the reservoir) for a few months. There is no way to analyze any relation between seepage and head change at this level.

In most cases, negative and positive values (flow to and from the reservoir) occurred in clusters. Water level drops cause negative seepage, or return flow, and water level increases cause seepage. However, there were usually lags of variable duration before the seepage changed signs after the reservoir level switched directions. In other words, the direction of movement across the reservoir perimeter required a period of time to reverse once the reservoir level changes changed. From 1992 to 1997, abrupt water level rises of 10 to 20 feet followed and coincided with months of return flow (Figure 4). The rapid increase in water level started after 6 to 8 month periods of seepage which continued even as levels dropped several feet per month. Rapid increases in reservoir level extend the water level above that in the surrounding rock. Several months of reservoir drops are required for the level in the bank to equal the reservoir level.

Because seepage for several months is controlled by one or two months of rapid rise, there is no monthly relationship. The scatter shown in Figure 5 has a correlation coefficient of only 0.09. Many months with negative seepage correspond with positive level changes and vice versa. The lag time between reservoir level changes and equalization with aquifer heads renders statistical analysis of seepage and reservoir levels impossible to determine whether loss rates are greater at different levels is impossible. Even if statistics were possible, the changing perimeter of the reservoir would render this analysis more difficult.

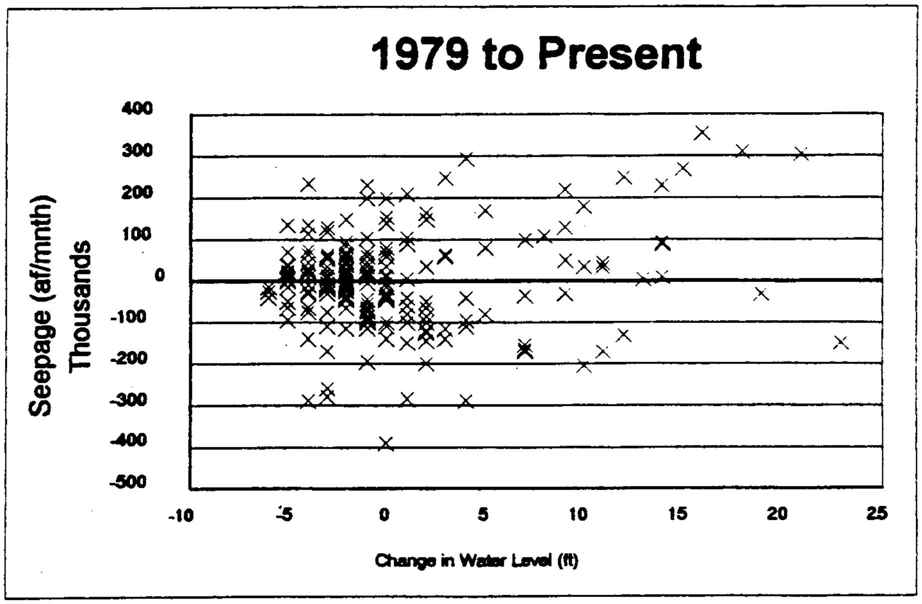

Figure 5: Scatter plot of monthly seepage and change in water level.

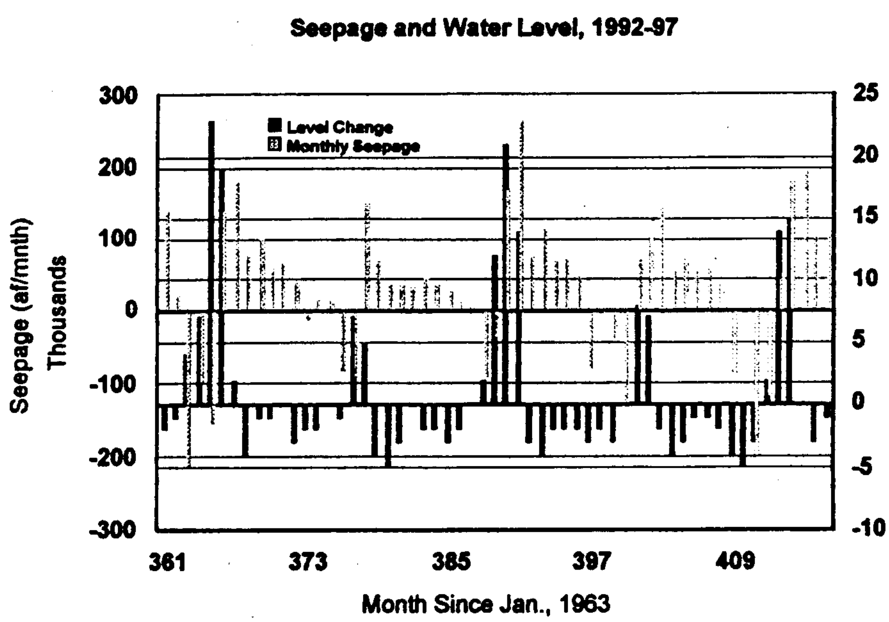

Figure 4: Monthly seepage and level change between 1992 and 1997.

Total Losses

Annual seepage losses varied depending on the rate of rise and the time period considered. Cumulative seepage peaked around 1983 at 11,000,000 af after which it dropped to less than 10,000,000 af in 1992. Between 1992 and 1997, cumulative seepage fluctuated between 10,000,000 and 10,500,000 af. The water balance analysis does not suggest that cumulative seepage is increasing. During months with rapid rises, several hundred thousand acre-feet of water seep, but then the return flow over several months reduces cumulative seepage back to the level existing before the rapid rise. Aquifer levels rise as seepage from the reservoir continues. This causes seepage decreases as the gradient from the reservoir level to the aquifer water level decreases. Because the aquifers are extensive, seepage could wet aquifers for many miles both north and south.

Based on the water balance analysis, the total loss of water due to Lake Powell between 1963 and 1997 was 30,000,000 af, or more than 850,000 af/year. This is about 6.3 percent of the average annual flow at Lee's Ferry. Total evaporation from the reservoir has been about 23,500,000 af and the net loss due to the reservoir is about 20,000,000 af. The seepage loss was about 10,000,000 af. Evaporation and seepage loss, respectively, are about two-thirds and one-third of the total loss caused by Lake Powell.

As upper basin development decreases the flow into the reservoir, the percent of inflow lost will increase if the reservoir remains at 1997 levels. Of course, with decreased inflow and the outflow set by decree, the reservoir level will decrease. As this occurs, the surface area will decrease faster than the relative volume. The area subject to evaporation will decrease and the total loss will be less. The decreased reservoir level may allow some seepage to return to the reservoir.

Groundwater Model of Seepage from Lake Powell

The only way to estimate seepage independent from the water balance analysis is with a well calibrated groundwater model. Available geologic information is limited, but sufficient for the writing of a basic regionwide groundwater model. The purpose of this model is to simulate seepage amounts and the change in head in the aquifers around the reservoir. The goal is to determine whether the magnitude estimated by the water balance is appropriate and to estimate the mechanism and location of the seepage, and to predict future seepage and time to equilibrium.

Hydrogeology

Many books have written about the general geology of the Grand Canyon, Glen Canyon and the Four Corners region. Almost nothing has been written about the hydrogeology, or the movement of groundwater within the stratified layers in the study area. Two reports prepared by the U.S. Geological Survey (Blanchard, 1986 and Thomas, 1986) provide the most up-to-date information about the region and are used as the basis for the analysis herein.

Geologic Layers: The following table describes the stratigraphic units found within the study area and used within the model. The description follows Blanchard (1986) and the hydrologic characteristics follow Thomas (1986).

Table 1: Stratigraphic Units in the Study Area

#	Age	Stratigraphic Layer	Description	Hydrologic Properties
7	Jurassic	Entrada Sdstn	Massive medium- to fine-grained crossbedded sandstone.	Yields water to springs east of Henry Mtns.
6	Jurassic	Carmel Form.	Thin beds of limy siltstone, fine-grained Friable sandstone, limestone, and gypsum, all of marginal marine origin	Yields less than 1 gpm to springs. Generally a confining bed.
5	Jurassic and Triassic	Navajo Sdstn	Massive medium-to fined-grained sandstone, exhibiting large-scale aeolian cross-bedding.	Yields several 100 gpm to wells and springs thru-out study area.
4	Triassic	Kayenta Form.	Fluvial sandstone, siltstone, shale, and minor shale-pellet conglomerate and freshwater limestone.	Yields small amt water; much less permeable than Navajo sdstn
3	Triassic	Moenave Form.	Composed of two members: Springdale sandstone member is medium-grained, micaceous sandstone and minor siltstone; and underlying Dinosaur Canyon Sandstone member is coarse- to fine-grained parallel-bedded sandstone and siltstone.	Not known to yield water
2	Triassic	Wingate Sdstn	Fine-grained, thickly crossbedded, calcareous aeolian sandstone.	Yields less than 5 gpm to wells near Navajo Mtn.
1	Triassic	Chinle Form.	Varicolored beds of fluvial and lacustrine origin, generally sandy at top; limey, muddy, bentonitic in the middle; and sand and conglomeratic near base.	Less than 1 gpm in sandy spots

Units 2 through 5 are the Glen Canyon group. The Navajo sandstone is the best aquifer material in the area. Most springs occur at the base of the formations. This indicates that near the discharge points the aquifers all tend to be unconfined. The underlying layers act as a barrier (an aquiclude) causing water to move horizontally to springs or seeps. That the layer above the spring is dry indicates that the layer near the discharge points are unconfined. Thomas (1986) indicates that several miles to the north, closer to recharge zones, these layers are under pressure as the heads within a layer exceed the top of the layer.

Based on laboratory analysis of core samples, Blanchard (1986) estimated hydraulic conductivity of the Navajo and Wingate sandstone to be 3.5 and 1.0 ft/day, respectively. Fracture zones have a much higher conductivity. Localized hydraulic conductivity may be several orders of magnitude higher than the bulk aquifer material. Fractures exist along the

Waterpocket monocline and on the steeply dipping structures near the Henry and Navajo Mountains. In these areas, igneous intrusions have bent the sandstone layers causing fractures. Isolated fractures are too small to consider in a regional groundwater model because modeling uses cells with size many orders of magnitude greater than the fracture.

Recharge/Discharge: Between 1929 and 1962, the local discharge into the Colorado River between Cisco, Bluff, Green River and Lee's Ferry was about 500,000 af/year. Most of this discharge is surface runoff that never reached the groundwater and is not relevant to the groundwater model. Based on spot measurements, Blanchard (1986) estimated base flow to be about 4000 af/year in tributaries. Groundwater inflow usually supports baseflow, but it is not known how much of this groundwater inflow would be from the regional flow system and therefore connected to this model. The difference is that floodplain flows recharge shallow near surface aquifers that are not connected to regional groundwater (Workman and Serrano, 1999).

Blanchard (1986) reported many small springs and seeps emanating from the Glen Canyon Group prior to reservoir formation, but estimated total inflow to be less than 10,000 af/year. These are small when compared to the river flow. There is no measurement of the flow from seeps directly into the river. In a steady state system, some discharge directly to the river from the aquifers probably occurs. But it is unlikely that more than 10% of the 500,000 af/year of local inflow can be linked to groundwater discharge to the river. For this reason, one of the calibration targets was keeping river seepage below 50,000 af/year.

Recharge to the study area occurs in mountains north and south of the model domain and in mountains within the domain. Blanchard (1986) estimated there was very little recharge into the principal aquifers, the Navajo and Wingate Sandstone. This was due in part because most recharge occurred above 8000 feet and most outcroppings were below this. The only recharge reaching the aquifers would be through fractures in the overlying aquifer. He estimated that recharge in the Henry Mountains was only 3000 acre-feet per year for all aquifers. Similar recharge amounts and processes occur on Navajo Mountain. Blanchard also suggests that small quantities of recharge occur in the region between the rivers due to precipitation directly on the layers.

Groundwater levels near the town of Escalante are high and likely due to recharge on Boulder Mountain north of the town (Thomas, 1986). There is likely a substantial unmeasured inflow from the north that substantially exceeds recharge in the mountains within the model domain. The recharge from Boulder Mountain will enter the domain through the boundaries. Similar arguments but smaller flow amounts are likely from the south side of the model domain, from the Navajo Reservation area, as well.

Water Levels: Glen Canyon is the low point in a regional groundwater system (Figure 6). Prior to reservoir filling, groundwater flow was toward the river from all directions for

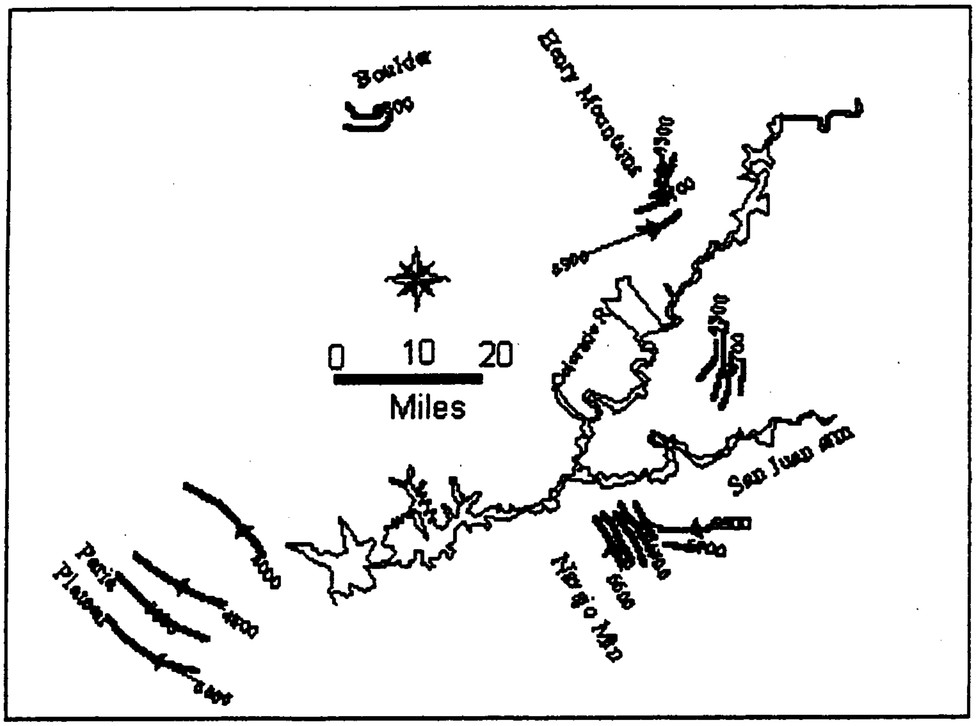

Figure 6: Groundwater contours in the vicinity of Lake Powell showing that the flow direction is toward Glen Canyon and the lake from all directions.

which groundwater contours have been estimated. In the north, recharge on Boulder Mountain and the Henry Mountains cause the level to be 2500 feet above the river. From the west, recharge on the Arizona Strip cause a significant gradient dropping to the east with levels as high as 5500 feet. Because these contours are in the Navajo sandstone, they represent artesian pressure under rivers such as the Paria (Thomas, 1986, Figure 6). To the south recharge on Navajo Mountain raise groundwater levels to at least 5500 feet (Blanchard, 1986, figure 10). Unfortunately, Blanchard's map does not extend further to the south and east into Arizona. However, the San Juan River gains flow indicating that flow toward the river occurs and that a groundwater divide exists south of the river. Blanchard (1986, Figure 10) also indicates that water levels between the rivers are as high as 4900 feet. Glen Canyon clearly was a gaining stream and the regional water table constrains seepage from the reservoir so that there is no long-term loss in any direction. The obvious exception is seepage from the reservoir around the dam to the river below, but this does not represent a significant loss to the river system.

Groundwater Model

Simulating the seepage from Lake Powell with a groundwater model satisfies at least two purposes. First, a model can show that the pattern of seepage observed in the water balance is possible with reasonable estimates of regional hydraulic parameters. Second, the model will provide an estimate as to the future magnitude of seepage from the reservoir and when a quasi-equilibrium between bank storage and return flow will be reached.

For this purpose, a coarse groundwater model using the US Geological Survey three-dimensional finite difference code MODFLOW (McDonald and Harbough, 1988) was used to simulate seepage from the reservoir. Groundwater Vistas, version 2.29, was the graphic user interface used to prepare input files and process the output.

The task of groundwater modeling was accomplished with the following steps:

1. A conceptual model of the domain was prepared. Using existing hydrogeologic and geologic information, the grid structure, boundary conditions, and model aquifer layers were determined.

2. A steady state calibration was completed. Calibration is the process of setting hydraulic parameters, primarily hydraulic conductivity, so that various target values are adequately simulated. The targets used herein were head levels at various point in the domain as determined from Blanchard (1986) and Thomas (1986) and from the level of the river. Assumed seepage limits to the river were also used. A steady state calibration uses pre-stress conditions; the largest stress in this model is the filling of Lake Powell.

3. A transient calibration and verification was performed. The goal of this step is to be certain that the model simulates changing water levels. Hydraulic conductivity was held constant as based on the steady state calibration. Storage coefficients are adjusted so that the aquifer responds as close as possible to the observed. The problem with this model is that transient observed levels only exist near the dam, primarily in Wahweap Bay.

4. The model was then used to predict future seepage by running it for 1500 years to determine the point that equilibrium occurs.

Grid Structure: The model domain extends about 106 and 73 miles, respectively, in the east-west and north-south directions (Figure 7). The model was large enough to allow regional flow to come to equilibrium with the reservoir while allowing for seepage from the reservoir to back up toward the boundaries. The grid was rotated 42 degrees to the east so that rows roughly parallel the rivers. Grid cell sizes varied from 5280 feet square for the aquifer at large to 560 feet square near the dam. The finer discretization was necessary to model the steep gradient expected for seepage around the dam.

Because the model simulates seepage from a reservoir that does not affect or is not affected by water movement in layers well above the Glen Canyon groups primarily intersecting the reservoir, two model layers spanning elevation 2000 to 4000 feet were used. Aquifers above the Glen Canyon group is ignored except for the recharge assumed to occur through it. The top layer is 850 feet thick with the top elevation being 4000 feet while the bottom of the layer is at 3150.

Boundaries: The domain was surrounded by variable flux boundaries using the General Head Boundary (GHB) module in MODFLOW (Figure 7). Lake Powell was also simulated with GHBs. The river during steady state conditions prior to the reservoir was simulated with the variable flux RIVER boundary package. Variable flux boundaries limit flow from or to the boundary by calculating a conductance for each cell according to the following equation:

$$C=\frac{K\times A}{L} \tag{9}$$

C is conductance, K is hydraulic conductivity of the porous material between the boundary and the water source, A is cross-sectional area perpendicular to the flow path, and L is the distance along the path of flow. Boundaries are assumed to be perpendicular to the flow path. The units are $length^2/time$, or ft^2/day. Conductance is multiplied by the head drop across the boundary to determine flow.

Boundary head elevations were set based on the best knowledge available which is primarily Figure 10 in Blanchard (1986) and Figure 6 in Thomas (1986). Figure 6 summarizes these levels. GHBs require a hydraulic conductivity, aquifer thickness, distance to the head elevation, and width of the cell to determine the conductance. Hydraulic conductivity was set equal to 1.0 ft/day, a magnitude that will simulate an extension of the existing aquifer material. In the top layer, aquifer thickness depended on the location of the phreatic surface and in the bottom layer thickness was 1150 feet. Distance to the head was 20,000 feet to allow the head at the boundary to fluctuate reasonably if conditions warranted. The width of the cell in the boundary was the width of the cell in the grid.

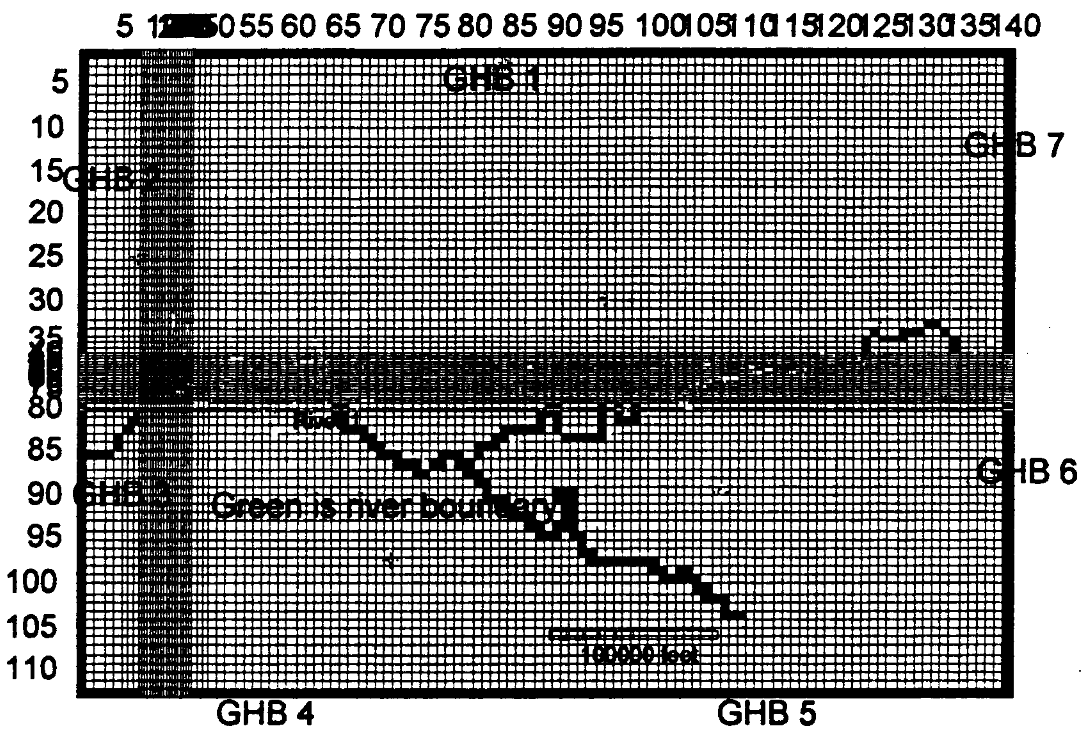

Figure 7: Grid and boundary conditions for the Glen Canyon regional model. GHB is general head boundary which were used to model and control the flow into and from the model domain. The green cells are the river boundaries for pre-dam conditions.

The steady state model used for calibration had variable flux boundaries for the Colorado and San Juan Rivers. This was the RIVER package in MODFLOW (Figure 7). Other rivers were ignored because they were primarily above the top layer and groundwater flow interaction with them was expected to be slight because of the fluvial soils constrained by the boundaries of canyons carved in sandstone. Required input for river boundaries includes water surface and bottom elevations, bottom layer thickness, bottom layer hydraulic conductivity, river width, and length within each cell to determine conductance for flow across the bottom layer. The river elevations were measured from maps with the depth assumed to be 10 feet. The Colorado and San Juan Rivers were assumed to be 300 and 200 feet wide, respectively. Bottom layer thickness was 1 foot with a 4 ft/day hydraulic conductivity.

The rising reservoir level used was modeled with transient GHBs. Anderson and Woessner (1992) recommended using a constant head boundary to provide an unlimited source of water for lakes and reservoirs, but MODFLOW does not allow this type of boundary to vary with time. Therefore, the GHB boundary used here had high hydraulic conductivity to minimize the resistance. Storage coefficients within the reservoir area were set to 0.001 so that little water would go into or come out of storage within the reservoir volume and confuse the analysis of the results. The distance to the head was just 10 feet to rapidly transmit the changing reservoir levels to the aquifer. Other parameters in the transient model are as determined for the steady state model.

The other boundary used in each model was a constant flux boundary at the Henry Mountains, Navajo Mountain and the area between the rivers. This was minor recharge equaling about 5500 af total for the model domain. Note that this exceeds slightly the amount suggested by Blanchard (1986). Recharge was increased slightly during the steady state calibration to raise the water levels in the mountains without decreasing the hydraulic conductivity.

There was no evapotranspiration used in either the steady state or the transient model because the ground surface was above the top layer in most cases and because vegetation is sparse on the sandstone in the area.

Hydraulic Properties: There have been no published pump tests performed on the aquifer materials within this domain. Hydraulic conductivity has been measured in the laboratory to be less than 4 ft/day for all layers. Laboratory measurements are extremely unreliable because they represent a point value rather than a continuous value over an element or cell of aquifer material. The Moenave formation conductivity is probably less than 0.01 ft/day because of its imbricated structure and lack of water yield. Thomas (1986) estimated specific yield to range from 0.05 to 0.15 and storativity to vary from 0.0001 to 0.001. The specific yield values are low when compared to alluvial aquifers which range from 0.15 to 0.30 but the storativity values are similar to that for confined alluvial aquifers. The specific yield values are in the low end of the 0.02 to 0.41 range suggested by Anderson and Woessner (1992). Final values were estimated during the calibration and will be discussed below.

Hydraulic conductivity zones were determined based on the geologic layers thought to exist given model elevation (Figure 8). Blanchard (1986) provided maps of surface outcrops and thickness of all layers. Based on land surface contours and interpolation of the maps, it was possible to estimate the formation existing at each location. Hydraulic conductivity was set during a steady state calibration phase so that heads equaled expected values. Only the parameter values were adjusted, not the zone area itself. It is assumed that prior to the filling of the reservoir, groundwater flow was at equilibrium. During a transient phase, the storage coefficients were adjusted so that simulated head levels approximated that in five observation wells. This phase also represents a verification of the seepage magnitude simulated with that determined from the water balance model. Reservoir level changes were the only stresses applied during the transient phase.

Steady State Calibration

The goal of a calibration is to set model parameters, in this case hydraulic parameters of the aquifers and the boundaries, so that various targets are matched. Groundwater levels around the region and a maximum seepage to the rivers were the primary targets. A secondary target was the requirement to not to increase flow from the boundaries significantly more than expected in the regional flow.

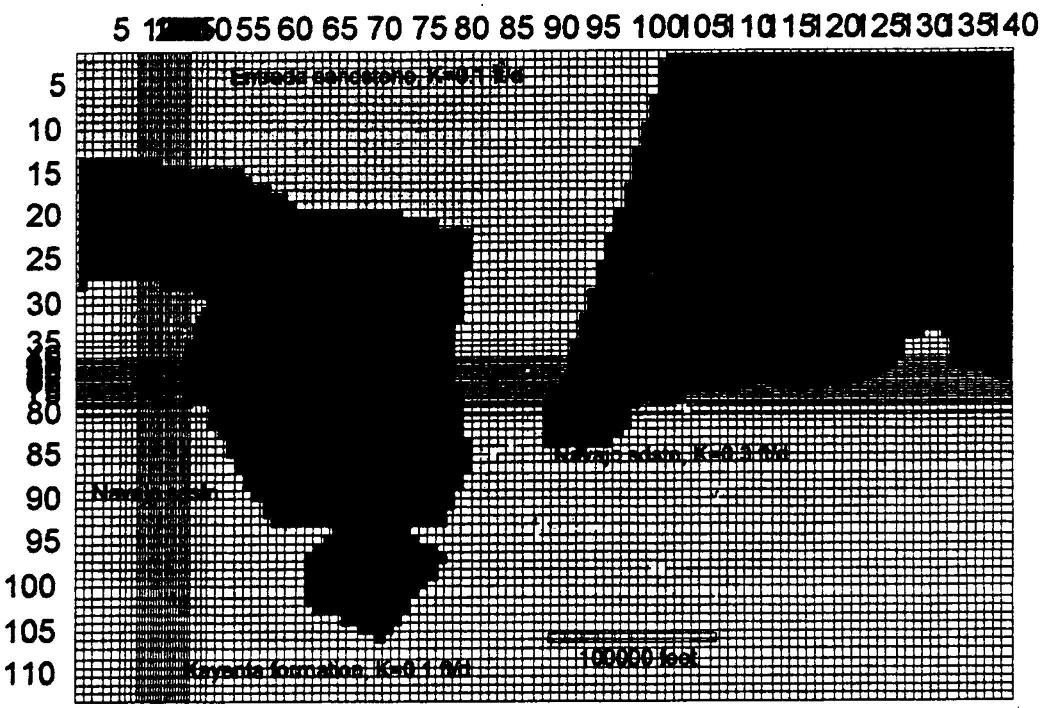

Figure 8: Calibrated hydraulic conductivities for the Glen Canyon regional model.

Steady state groundwater flow conditions exist when few stresses exist on the system. Prior to aquifer development, such as a well field, the only changes in flow that occur are seasonal. Viewed at a time scale of decades, seasonal changes are ignored and the system is in steady state. For this analysis of regional groundwater flow to the Colorado River, the primary stress is the filling of Lake Powell. These changes far exceed any stresses caused by the pumping from the few wells existing prior to 1963. For steady state calibration, we assumed that conditions existing in 1963 were in steady state. We also assumed that reservoir seepage has not yet reached locations more than ten miles from the reservoir so that current observations approximate steady state conditions. This assumption holds because well levels reported by Blanchard (1986) do not fluctuate with changing reservoir levels.

Using water levels obtained from both Blanchard (1986) and Thomas (1986) along with the water level in the Colorado River, several targets were established where the water level during a steady state analysis would be simulated. (This assumes seepage directly to the rivers; the amount and water level closely approximates the seeps and springs found before reservoir filling.) Also used for calibration was the water balance of the region and especially the flow to the rivers. (It was essential to not allow inflows from GHBs to exceed reasonable values because this water had to have come from somewhere.) As Blanchard (1986) indicates, little recharge occurs on the plateaus. Most recharge comes from higher elevations in all directions from the domain. Thomas (1986) indicates flow into the region from the west, but that it flows south right at the boundary corresponding with GHBs on the west of this model.

By altering hydraulic conductivity values in the predetermined zones and recharge in all three zones, steady state groundwater levels approximated the targets within 100 feet (see the crosses in Figure 7). Because groundwater levels can drop as much as 100 feet in just a mile, the size of one cell, matching targets closer than this is not necessary. Also, the relief of groundwater levels in the model range from 3200 to 6300 feet amsl, therefore the residuals are less than 5 percent of the total change in head across the model domain.

The final hydraulic conductivity values (Figure 8) are an order of magnitude less than suggested above by both primary references (Blanchard, 1986 and Thomas, 1986). This was necessary to limit the flow to the river and to avoid the head values being several hundred feet above the river. In other words, the higher initial conductivity values resulted in the head at the river being several hundred feet above the river. This resulted in a high gradient across the bottom of the river and a flow that was several times greater than assumed to reach the river (see the discussion about seepage to the river above). Decreasing river conductivity in the boundary could have reduced the flow, but would have increased the head above the river level even more and resulted in unrealistic levels. For these reasons the hydraulic conductivity of the aquifers was decreased.

Total flow from the GHBs into the model equaled 38,095 af/year, recharge equaled 5607 af/year and flow from the model to the river equaled 43,757 af/year. Excluding the springs, seeps, and tributaries discussed above, the river seepage is about 28,000 af/year. The error is only 0.12 percent. The following table summarizes the values for each GHB. Negative flow represents flow from the model and positive flow represents flow into the model. Negative river flows means that was lost from the model to the river.

Table 2: Summary of Flows from Each Boundary

Layer	Boundary Type	Number	Flow (af/year)
1	**GHB**	1	12,699
1		2	3,142
1		3	34
1		4	4,000
1		5	1479
1		6	-420
1		7	688
1		25	693
1	**River**	1	-11,435
1		2	-17,061
1		3	-6724
1		4	-8537
2	**GHB**	11	14,644
2		12	-3646
2		13	-615
2		14	3373
2		15	1944
2		16	1124
2		17	1196
2		26	-2240
1	**Recharge**	5607	

The majority of inflow comes from the GHBs on the north side of the domain (GHB 1 and 11). Lesser amounts come from the south and east. Small amounts of water flow out to the west in the northern end of the domain. Flow to the Colorado River is maximum upstream from the confluence with the San Juan River (river 2).

Transient Analysis

The purpose of the transient analysis was to simulate responses in the aquifer as determined by well levels observed near the dam from 1990 to 1999 and flow into storage around the reservoir as determined in the water balance above. The only well observations are from 1990 to 1999[7]. This analysis was done to verify the hydraulic parameters determined in the steady state calibration and to determine the appropriate storage coefficients.

Starting in 1963 when the reservoir began to fill, 66 stress periods were used to simulate the seepage. A stress period is a time period during which a stress is applied to the system. In this model, the stress is changing reservoir levels. The length of each period varied according to

[7]These were obtained from the US Bureau of Reclamation in Page, AZ in March, 1999.

the magnitude of the stress and the time period between well observation. The yearly periods begin on January 1 with the observed reservoir elevation on that date holding for the year long period. Each stress period is divided into 20 time steps using an expansion coefficient of 1.2. The first period length is determined with the following equation:

$$\Delta(1) = T \times \frac{(1-M)}{(1-M^n)} \tag{10}$$

M is the expansion coefficient, n is the number of steps and T is the period length. Following time steps are the expansion coefficient (1.2) times the preceding time step. For a 365 day period, the first step is 1.96 days and the twentieth step is about 62 days.

Table 3 shows the stress periods, period length, and reservoir levels. The shortened periods 21 through 24 reflect the rapid level changes caused by high water in the early 1980s. The shortened periods beginning with number 31 reflects the additional detail required to simulate the well level changes.

Table 3: Table of Stress Periods and Reservoir Levels

Period	Length	Res.	Period	Length (d)	Res. Elev.	Period	Length (d)	Res. Elev.	Period	Length (d)	Res. Elev.
1	365	3410	17	365	3673	33	122	3637	49	122	3654
2	365	3492	18	365	3681	34	122	3629	50	93	3649
3	365	3535	19	365	3667	35	93	3627	51	75	3645
4	365	3521	20	365	3685	36	93	3638	52	92	3680
5	365	3527	21	183	3707	37	93	3628	53	93	3687
6	365	3539	22	182	3685	38	122	3622	54	93	3680
7	365	3572	23	183	3702	39	93	3629	55	93	3672
8	365	3600	24	182	3685	40	122	3622	56	75	3688
9	365	3610	25	365	3687	41	93	3628	57	93	3679
10	365	3606	26	365	3684	42	93	3633	58	93	3673
11	365	3649	27	365	3685	43	122	3622	59	93	3663
12	365	3648	28	365	3680	44	75	3615	60	92	3694
13	365	3668	29	365	3658	45	93	3619	61	93	3690
14	365	3655	30	90	3633	46	93	3668	62	92	3682
15	365	3630	31	93	3651	47	93	3662	63	93	3673
16	365	3633	32	93	3650	48	92	3659	64	92	3696
									65	93	3688
									66	93	3687

Specific yield was adjusted downward to 0.06 from the starting value of 0.15; storativity was dropped to 0.001 for all aquifer materials in the domain. This was required to force the

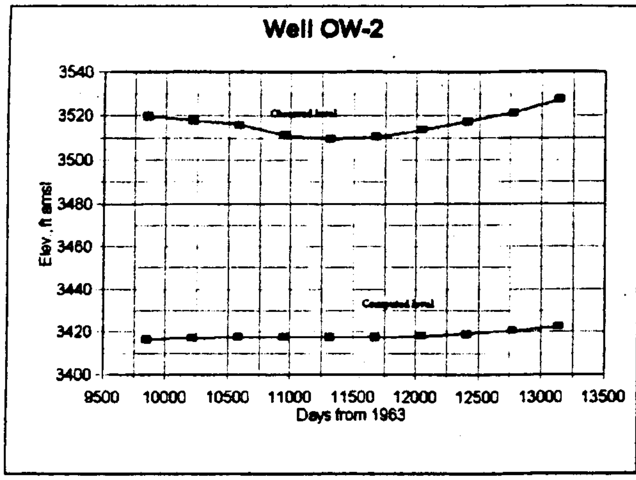

Figure 9a: Computed and observed levels, well OW-2.

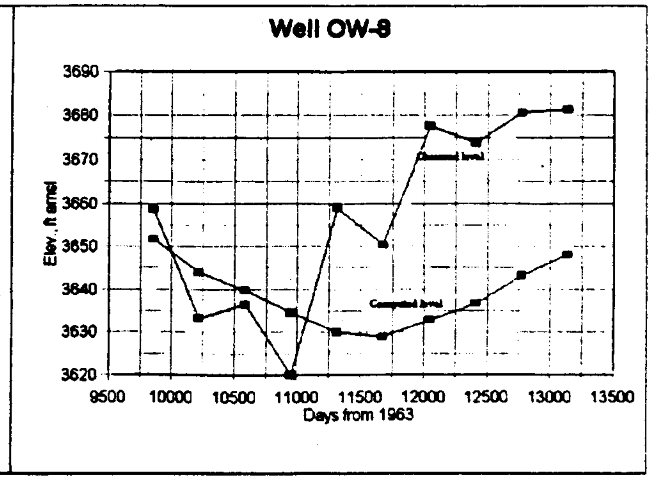

Figure 9b: Computed and observed levels, well OW-8.

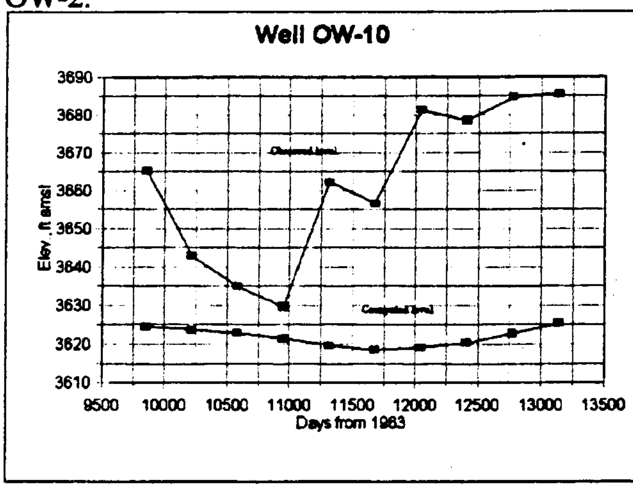

Figure 9c: Computed and observed levels, well OW-10.

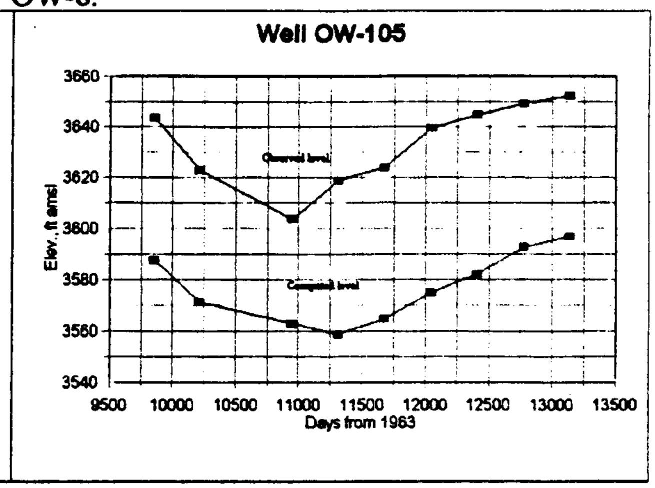

Figure 9d: Computed and observed levels, well OW-105.

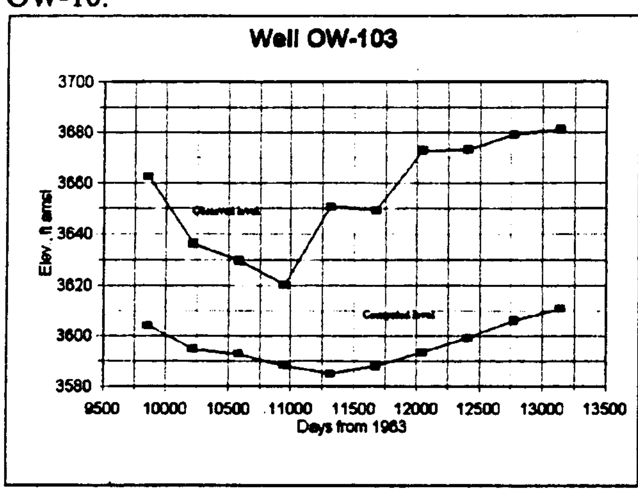

Figure 9e: Computed and observed levels, well OW-103.

mounding due to reservoir rises to move further away from the reservoir as indicated in the observation wells. The final model came within 50 feet of the observation wells in 1990, the start of the observation period. After 28 years of simulation with no observations to calibrate to,

this is excellent. More importantly, after 1990, most of the well levels parallel the observation wells (Figure 9) suggesting that water moves into and out of the banks in a realistic manner. The exception is well OW-2 (Figure 9a).

Cumulative seepage paralleled the estimate determined in the water balance analysis, but results in about half as much total seepage (Figure 10). The most probable reason for the difference is errors in both analyses. Flow gaging stations may be incorrect by up to 20%. The surface evaporation estimates probably have a 30% range on them. The local inflow calculations vary by up to 22% just from the regression (see Appendix 1). The sensitivity of flow to the reservoir due to the hydraulic conductivity of the surrounding material also approaches 20%. Actual seepage losses are probably close to the water balance estimate of 10,000,000 af because of the unknown aquifer layers and fractures that decrease the reliability of this groundwater model.

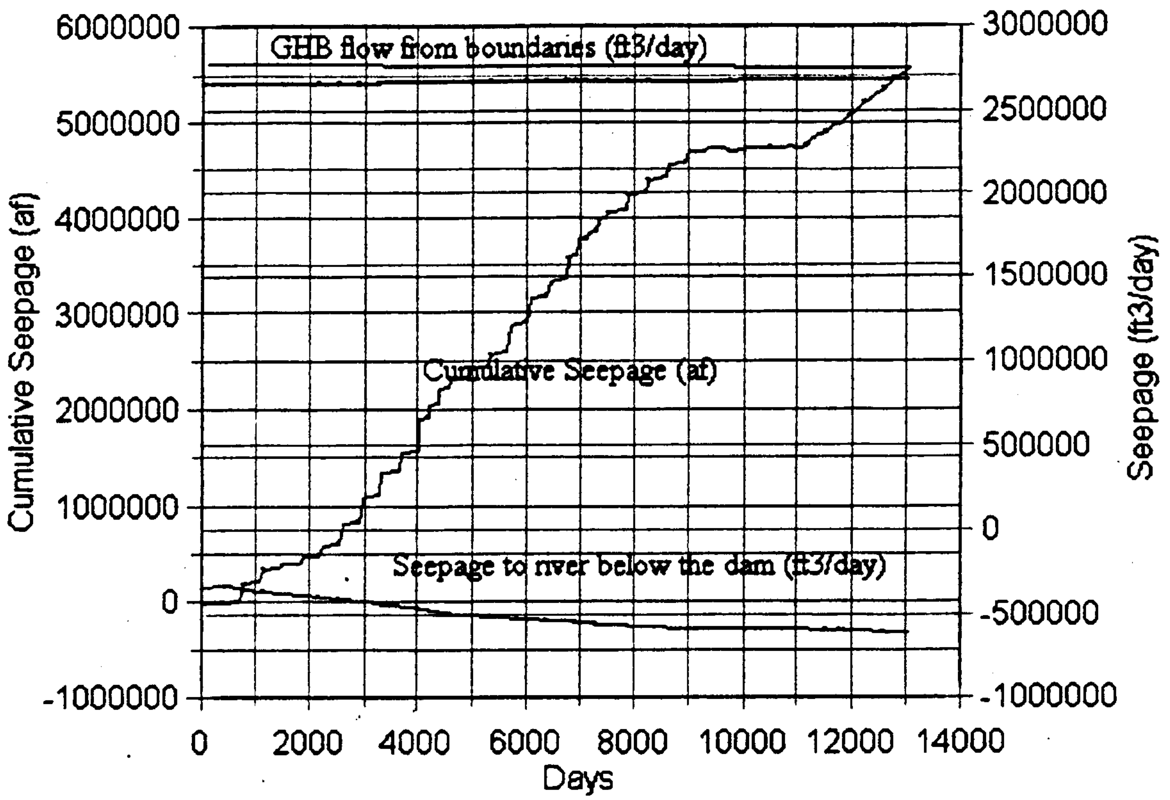

Figure 10: Seepage from Lake Powell during the transient simulation, 1963-99.

Seepage to the river below the dam also increased about 300,000 ft^3/day, or about 2500 af/year. This is expected due to the rising reservoir levels and reflects the very steep gradient near the dam from the reservoir to the river. This seepage reflects seeps and springs that have been found below the dam. It also represents a small source of error in the difference between seepage volumes in the model and water balance discussed above. The water balance would not account for any seepage reaching the river downstream from the gage at Lee's Ferry.

Flow from the GHBs changes only slightly with time through the transient runs. The slight change reflects a slightly higher flow downstream through the western GHB. The lack of change from other boundaries indicates that the stress has not reached the boundaries. The total drawdown from 1963 to 1999 (Figure 11) reflects this lack of impact. Approximately two thirds of the model domain are only slightly affected by the stress.

Time to Equilibrium

To test the time to equilibrium, the model was run with a 67th stress period added to the 66 periods used from 1963 to 1997. It was 1500 years, or 547,500 days, long. According to equation 10, the first time step of this period was about 15 days and the last was about 100,000 days. Because there is no information about future levels or cycles, the reservoir level was set equal to the long-term average level of 3680 feet similar to Thomas (1986).

Cumulative seepage after 1500 years is about 10,800,000 af (Figure 12). That the storage rate becomes negative after 1400 years suggests that the reservoir will lose water to the banks for many years. This contrasts with Thomas (1986) who suggested a much shorter time to

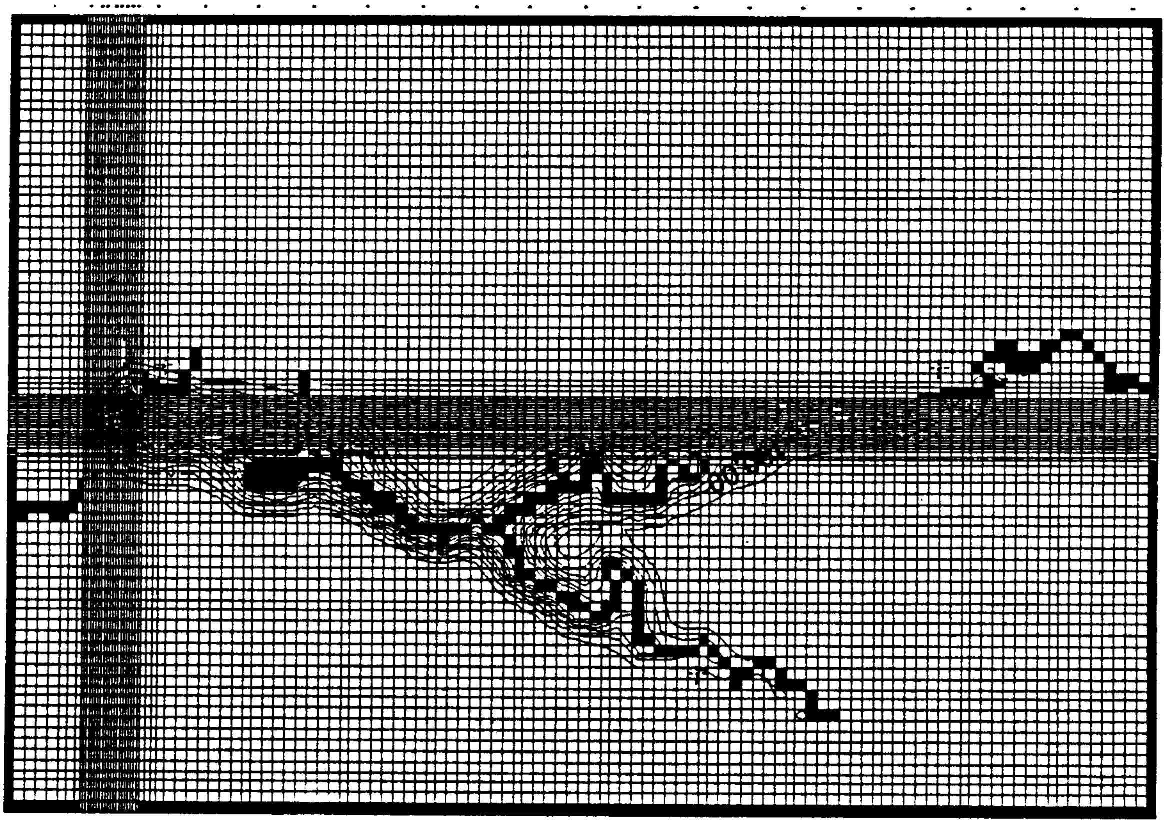

Figure 11: Predicted mounding in 1998 due to filling Lake in 50 foot contours. There is no drawdown in the area away from the reservoir. The mound decreases toward the upstream end of the reservoir.

equilibrium. The rate of loss will be much less however as reflected by the fact that the total cumulative seepage after 1500 years is only twice that of the first 34 years.

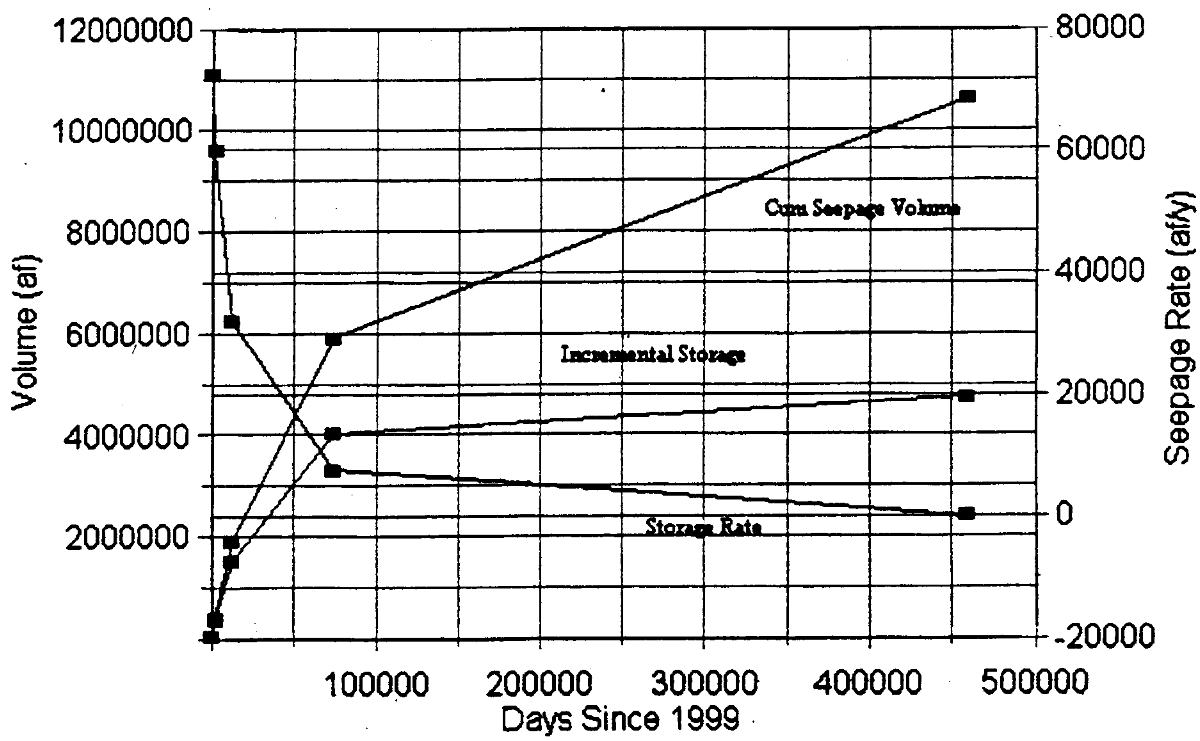

Figure 12: Long-term storage relations for the 1500 years beyond 1999.

Differences Between This Model and Thomas (1986): Thomas (1986) simulated groundwater flow from the north side of the reservoir from near Wahweap Bay to about 20 miles downstream of the dam to the Paria Plateau. He calibrated in steady state conditions by changing hydraulic parameters and recharge from the Paria Plateau. During transient calibration, he calibrated by altering seepage from Lake Powell. A difference in model structure is that we used a GHB on the Paria Plateau to simulate recharge. This model only drew about 1000 af from this boundary which is less than the 5000 to 15000 af/y range used by Thomas[8]. His estimates were based on no data and seem very high when compared with the 3000 af/y estimate by Blanchard (1986) for the Henry Mountains.

[8]Thomas acknowledged that his model did not match the expected levels in the northwest boundary near the location of the high recharge on page 37.

Thomas' calibration values for hydraulic conductivity are an order of magnitude higher than found here. At least two differences between the models explain the difference in estimate. First, Thomas forced large amounts of recharge into his models on the west and north sides. Higher hydraulic conductivity was required to allow this water to reach the river without substantially increasing water levels. Second, during transient simulation, Thomas settled on a specific yield of 0.02, an extremely low value for this material. The low value would increase the rate that water levels in the aquifer rise in response to the reservoir. To counter that rate, Thomas used higher conductivity values so that water would move away from the reservoir faster and the rate of rise would be decreased. Also, he tested a range of specific yields that included that used herein and found that the whole range would potentially work. His results were not very sensitive while the flow to the river herein is very sensitive to hydraulic conductivity.

Thomas' predicted that long-term equilibrium would occur in 400 years with 57% having occurred after 100 years, that total bank storage would be double the 1983 bank storage and that between 1963 and 1983, bank storage increased by 25,000 af/mile of reservoir. According to the digitizing of the lake boundary completed for this study, the perimeter is 564 miles. This indicates that 14,100,000 af has seeped into the bank.

This study found a longer time to equilibrium than Thomas for two reasons. First, the boundaries were much further from the reservoir which allowed for a "backwater" in the regional flow to the reservoir to extend beyond the limitations imposed by Thomas' limited domain. Backwater allowed the levels to continue rising, albeit slowly, for many centuries. Second, the Lake Powell stresses were applied to the regional flow simulated in the steady state model. The regional flow allowed the bank storage to move further from the reservoir in both upstream and downstream directions.

Thomas' estimate of total seepage was significantly higher than that estimated herein either by the water balance or the groundwater model. There are three faults with Thomas' estimate of seepage that indicate that the model presented herein is a more realistic estimate. First, Wahweap Bay is at the downstream end of the reservoir where the depth is maximum. There is much more depth over which the reservoir interacts with the banks at this point than along most of the reservoir. It is a poor assumption to argue that the depth of the reservoir averages one-half the depth at Wahweap because of the long, flat length of Glen Canyon. However, as the river bottom (reservoir bottom) raises toward the upstream end of the reservoir, the area available for seepage decreases substantially. Because of the bays and many complexities of the canyon, a detailed analysis of area for seepage is beyond the scope of this study. However, one-third would be a reasonable reduction in Thomas' seepage estimate. This reduces the estimate 9,400,000 af for the time period.

Second, the total seepage estimate should be reduced because the perimeter estimate includes shoreline on peninsulas that would quickly fill with water and not continue to pass seepage at the same rate as predicted by a groundwater model. A key example of this is the land between the San Juan and Colorado River arms of the reservoir. Estimating the applicable reduction due to ineffective shoreline is difficult. However, it is clear that Thomas' total seepage estimate should be reduced to less than 9,400,000.

Third, Thomas' method of modeling reservoir seepage was to force it into the banks. He did not consider the amount returning to the river below Glen Canyon Dam. The model herein accounted for that return flow and did not count it.

Because of these faults, Thomas' estimate should be reduced to very close to the total amount predicted by the model presented herein. The two separate models when properly interpreted yield compatible results.

Discussion

Based on the water balance analysis, cumulative seepage peaked around 1983 at 11,000,000 af after which it dropped to less than 10,000,000 af in 1992. Between 1992 and 1997, cumulative seepage fluctuated between 10,000,000 and 10,500,000 af as the reservoir rose and fell. A groundwater model confirmed the mechanics and pattern of seepage loss, but underestimated the quantity by about 50%. The primary reasons for the difference in seepage quantity are errors in the measurements and estimates used in the water balance analysis, the sensitivity of total seepage in the groundwater model to the storage coefficients, and potential inaccuracies in the assumed gradient of the groundwater surface near the reservoir.

Potential errors in the water balance are up to 20% on each of the gaging stations and to 30% of the regression estimated inflow (but the magnitude of these areas will be small compared to the overall seepage from the model). Errors in evaporation estimates are possibly substantial but not possible to estimate. Evaporation estimate errors stem from substantial differences in wind velocity and direction in all of the bays and coves of the reservoir.

The shape of the water table near Lake Powell was estimated from Figure 10 in Blanchard (1986) and Figure 6 in Thomas (1986). Both sources based their groundwater levels on little well level information but rather depended on observed springs and river water levels. The mounding shown in Figure 10 (this report) represents about 5,500,000 af of bank storage. (Because of the complete lack of wells around most of the reservoir, there is no way to verify the shape of the mound.) The mound is up to 600 feet high near the reservoir. If the extent of the mound were increased, the volume also would increase. The primary control on the extent of the mound, given the calibrated hydraulic parameters, is the location of natural, regional water table. Errors in the assumptions of this water table which was calibrated to in the steady state model and which formed the initial conditions in the transient model would also limit the total seepage from the reservoir.

For predictions into the future, it is reasonable to use the groundwater model but double the amount of loss to reflect the water . This is reasonable because the regional water table limits the extent of the mound. The storage coefficients limit the volume of water stored within the mound. Because the best measured values in the water balance indicate seepage is about twice that found in the model, it is reasonable to double the seepage values from the model for discussion. (Note, it is not reasonable to change the storage coefficients in the model because, as discussed under calibration, they were adjusted to provide the best fit to the observed transient well levels.)

Running the model for 1500 years into the future suggests that equilibrium will occur in about 1400 years and about 21,600,000 af will be lost. Over the total period, half of the seepage is lost in the first 37 years because the available storage becomes filled and the gradient has from the reservoir to the bank has significantly decreased (consider the mounds around the reservoir in Figure 10). The equilibrium depends on the assumed reservoir level (3680'). The actual equilibrium will be dynamic with water movement into and out of the banks. The loss rate drops significantly and the time to equilibrium is very long because water moves into the sandstone canyon walls very slowly. The natural groundwater flow is toward the river from both south and north (Blanchard, 1986), therefore there is no continuous flow away from the reservoir to the south or north.

Evaporation losses up to 1997 are about double those to seepage. Total evaporation from the reservoir until 1997 has been about 23,500,000 af. Evaporation without the reservoir was probably about 102,000 af/year on the 18,000 acres of riparian area within the canyons of the two rivers. Since dam closure, a high estimate is that 3,500,000 af would have been lost without the reservoir. The net evaporation loss due to the reservoir is about 20,000,000 af since 1963. The evaporation loss is about 2.1 percent of a full reservoir (27,000,000 af).

Total losses from Lake Powell due to the reservoir have been about 30,000,000 af. This is about 2 1/4 years of average annual flow at the Lee's Ferry gage or 6.3 percent of the annual flow. It is also about 11 percent more than the volume of the reservoir when full. Considered as an average over 34 years, the annual loss is about 3.3 percent of a full reservoir. The evaporation loss will continue at a rate of about 570,000 af/year after considering predam losses. The seepage loss will rapidly slow down and become much less than has been observed for the first 34 years.

Conclusion

Seepage and evaporation represent a substantial loss of water from the Colorado River system. Currently the loss exceeds 6 percent of the average annual inflow, but will reduce to about 5 percent as seepage decreases. As the demands on the river's flows exceed the annual flow prior to reservoir losses, the loss as a percentage of inflow will likely change. Only a detailed operations analysis will resolve the question of whether the increased flow in the river that would result from draining Lake Powell would be more important than the decreased certainty caused by the lost in-channel storage.

Evaporation will continue into the future at current rates. According to the groundwater model, the fact the water balance model predicted seepage peaked in 1983, and Thomas (1986), seepage losses will be reduced substantially in the future.

Recommendations

The results and conclusions of this report may be used to analyze the reliability of Lake Powell within the current Colorado River system. Lake Powell loses about 6 percent of the flow into it to evaporation and seepage. However, the report considers Lake Powell by itself and not as a part of a river system full of water projects including dams and diversions. As the upper basin develops, consumptive use will increase and the inflow to Lake Powell will decrease. Decreased inflows will cause reservoir levels to decrease. As this occurs, the evaporative area and the gradient driving seepage will decrease. Without performing detailed operation analyses, it is not possible estimate the decreased evaporation. Also, the total seepage and the time to equilibrium will also likely be decreased if the long-term average reservoir level is less than modeled here.

An important recommendation that the Bureau of Reclamation should change now is its current operating assumptions to fix the errors identified above. Most importantly, they should use bulk evaporation estimates rather than estimate the evaporation from the banks and subtract it from the reservoir evaporation. This evaporation loss prior to the reservoir would likely not have been consumptive use in the canyon and is not reflected in the calculated inflows for the Glen Canyon reach.

The primary recommendation for future studies, probably as a part of a full environmental impact statement, is to consider evaporation accurately and seepage in the operations modeling of future water availability. Only with this type of analysis will downstream users understand the true tradeoffs associated with Lake Powell. True evaporation estimates might decrease the certainty that users perceive in the delivery of water and increase the time period that demands are not met.

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Appendix 1

Calculation of Local Inflow into Lake Powell

Introduction

Inflow to Lake Powell is not directly measured at the upstream end of the reservoir and in the many tributaries that flow into the lake. There are three gages on the Green, Colorado and San Juan Rivers, respectively, that represent much of the flow from the mainstem rivers into the reservoir. However, these gages are many miles upstream from the reservoir and intervening river losses and tributary flows render the sum of these flows imprecise as an inflow to Lake Powell. The purpose of this appendix is to consider the method the Bureau of Reclamation uses to estimate the intervening local inflow and to provide a regression based improvement to that estimate for use in the water balance study of Lake Powell.

US Bureau of Reclamation Method

The three major rivers are gaged at various distances upstream from Lake Powell. The mainstem is also gaged at Lee's Ferry, just below Glen Canyon Dam. These gages include the Colorado River at Cisco, the Green River at Green River, San Juan River at Bluff, UT, and Colorado River at Lee's Ferry (Table 1). The gages will be referred to as the Cisco, Green River, Bluff, and Lee's Ferry gages, respectively, in this appendix.

Table 1: Gages Used in This Analysis

Name	Gage ID No.	Drainage Area (m2)	Datum (Ft. Msl)
CR at Lee's Ferry	9380000	111800	3106
CR near Cisco, UT	9180500	24100	4090
San Juan at Bluff, UT	9379500	23000	4048
Green R at Green River, UT	9315600	44850	4040

Substantial inflow and river loss occurs between the upstream gages and Lake Powell. For the purpose of estimating inflows to Lake Powell, the Bureau of Reclamation estimates local inflow as a coefficient, or proportion of the total flow measured at the Green River, Cisco and Bluff gages. This estimate was based on pre-dam, 1927 to 1962 flow data. The USBR's Reservoir Allocation Procedure Report[9] states:

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[9]As provided by Richard Clayton, USBR Salt Lake City Office, by email, March 29, 1999.

Inflow into Glen Canyon is calculated by adding up the San Juan River at the Bluff Gage, the Green River at the Green River Gage, and the Colorado River at the Cisco Gage and then adding or subtracting a fraction of this water using monthly coefficients. These monthly coefficients represent gains and losses between these gages and the Lee's Ferry gage. These were determined by the pre dam (1927-1962) relationship between the sum of the three gages and the Lee's Ferry gage. These coefficients are as follows:

Jan.	Feb	March	April	May	June	July	Aug	Sept	Oct	Nov	Dec
.0572	.0642	.0170	-.0485	-.0197	.0535	.1553	.0985	.1081	.0806	.1039	.0894

Note that historically, losses occurred in April and May.

Using data from 1929 through 1962, this author was able to collaborate these values. However, there is a great deal of scatter around the data (Figures 1-12). For example, during January when the USBR uses a coefficient equal to 0.0572, in at least five years the observed coefficient would have exceeded 0.10 and in several months it was less than 0.0 (Figure 1). In July and August, there were months in which the coefficient exceeded 0.3 while the estimate is

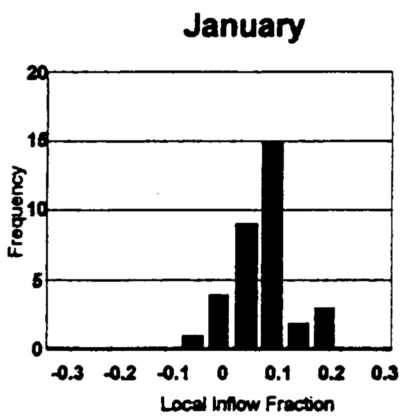

Figure 1: Local inflow fraction in January.

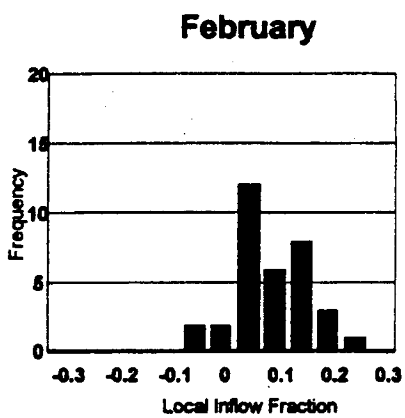

Figure 2: Local inflow fraction in February.

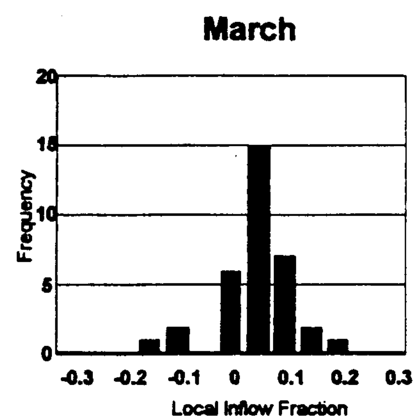

Figure 3: Local inflow fraction in March.

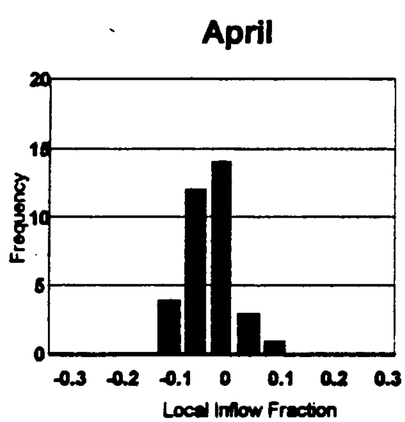

Figure 4: Local inflow fraction in April.

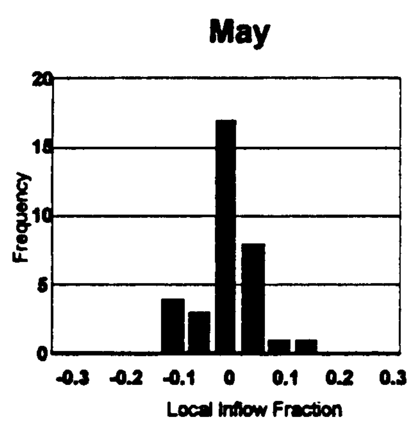

Figure 5: Local inflow fraction in May.

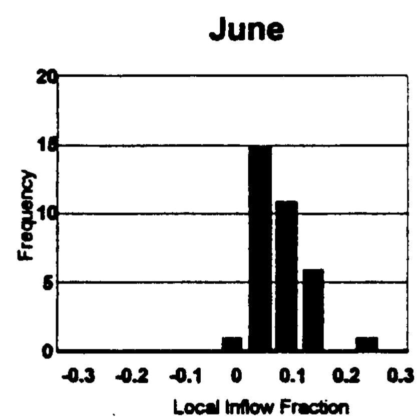

Figure 6: Local inflow fraction in June.

less than half of that (Figures 7 and 8). The average inflow in July and August exceeds 1,000,000 and equals 540,000 af/month, respectively. Based on average flows, the error during the high coefficient months

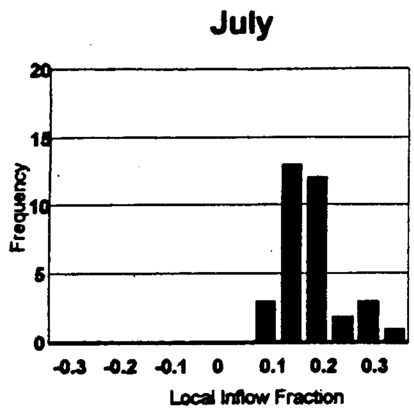

Figure 7: Local inflow fraction in July.

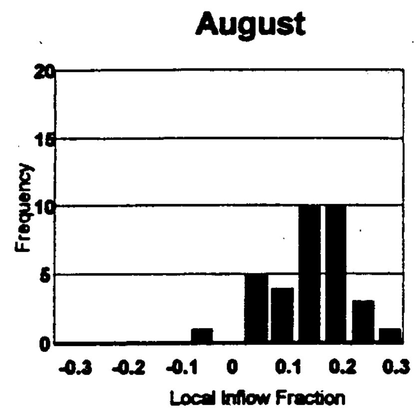

Figure 8: Local inflow fraction in August.

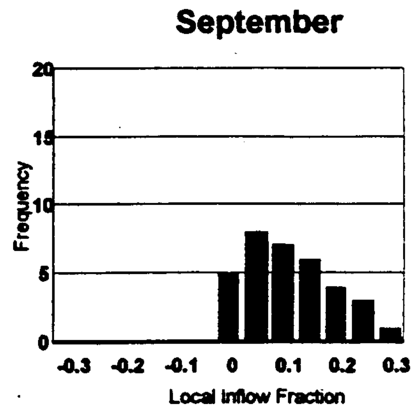

Figure 9: Local inflow fraction in September.

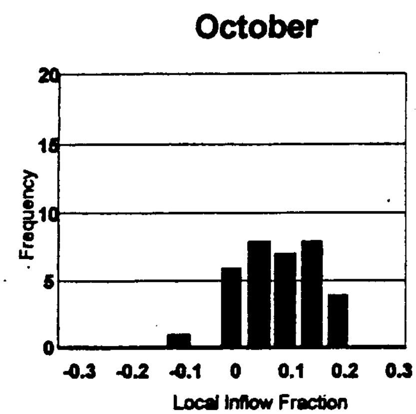

Figure 10: Local inflow fraction in October.

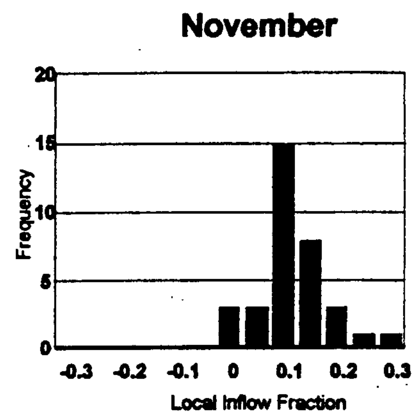

Figure 11: Local inflow fraction in November.

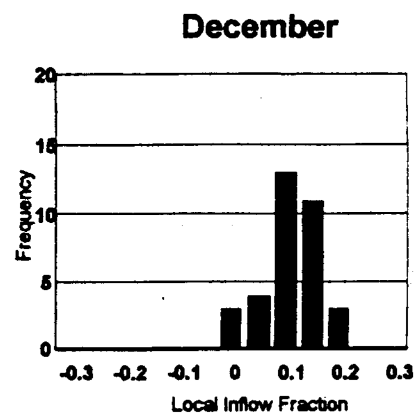

Figure 12: Local inflow fraction in December.

could be more than 150,000 af. During high flow months the error could be as high as half a million acre-feet.

The coefficients reflect tendencies in the Colorado River basin. From late fall through early spring, flows and losses are low and depend on baseflow. During mid-spring, April and May, before there is substantial runoff from the mountains, the riparian vegetation begins to transpire and losses exceed inflows. During the summer, monsoon storms cause localized high inflow rates. During all months, there is to high variability caused by seasonal changes.

Linear Regression Estimates of Local Inflows

The variability discussed above introduces many errors into a water balance analysis intended to determine seepage losses from the reservoir. Therefore, for this water balance analysis a detailed linear regression between the local inflows, determined as the difference between the Lee's Ferry and the sum of the three upstream gages and the three upstream mainstem gages individually was performed on the data from 1929 through 1962 (Haan, 1977).

Two models were tested. Both first and second order relations were considered as follows:

$$Q_l = B_0 + B_1 Q_{BL} + B_2 Q_{GR} + B_3 Q_{CI} \tag{1}$$

The βs are coefficients, Q_l is local inflow, and Q_{GR}, Q_{CI} and Q_{BL} are flow at Green River, Cisco, and Bluff, respectively. Because of the various differences among runoff relations by month, the regression analyses were completed by month. The best model between equations 1 and 2

$$Q_l = B_0 + B_1 Q_{BL} + B_2 Q_{GR} + B_3 Q_{CI} + B_4 Q_{BL}^2 + B_5 Q_{GR}^2 + B_6 Q_{CI}^2 \tag{2}$$

was chosen based on the adjusted coefficient of determination. If the increase in R^2 for adding the three additional terms in equation 2 was less than 0.04, equation 1 was chosen.

Table 1 presents the results of the analysis.

Table 1: Regression Model Results
Local Inflow as a Function of Upstream River Flow

Month	Model	B_0	B_1	B_2	B_3	B_4	B_5	B_6	R^2
Jan	2	-75452	-1.1399	0.84419	1.06122	.0000151	-.000002	-.000005	0.3633
Feb	2	42,844	-.45617	-.23825	.25186	.0000033	.0000007	-.000001	0.2828
March	2	-104102	.98475	.07589	.84021	-.000004	-3.8E-7	-.000002	0.2191
April	2	-13129	.42325	.38628	-.68705	-3.7e-7	-2.6e-7	.0000004	0.3053
May	2	-276412	-.42286	-.069569	.54563	.0000005	.0000001	-2.3e-7	0.3920
June	1	55478	.28758	.095018	-.07867				0.2701
July	2	8029	.06261	.15645	.23210	.0000001	-2.6e-9	-1.0e-7	0.7698
August	2	-33949	-0.08229	.165087	.46339	.0000012	-1.2e-8	-6.2e-7	0.9015
Sept.	2	-24682	.77988	1.38055	-.90743	-.000004	-.000004	.0000028	0.6103
October	1	-38320	.20209	-.08958	.32196				0.7717
November	1	-31591	.85783	.22505	-.02988				0.7894
December	1	1685	-.03206	-.53155	-.00905				0.1806

There are several trends in the relations. The magnitude of the coefficients indicates the relative importance of events in the various drainages in explaining local inflow. Most obvious is the small magnitude of some of the coefficients for the Bluff gage. Considering that this is also the gage with the least flow, it is apparent that events which control the flow at Bluff are more independent from the local inflow than the gages on the Green and Colorado Rivers. The best correlations occur in summer and early fall when monsoon rainfall likely controls local inflow. The regressions suggest that similar meteorological conditions control the flows at the mainstem gaging stations. This relationship probably only holds for monthly data while the variability of daily storms would preclude prediction at shorter time scales.

The poorest correlations occur in June and December when local inflow is lowest. During June the flow at the three gages is controlled by snowmelt runoff which is not occurring in the local watersheds. Snowmelt runoff does not explain either local inflow or evapotranspiration rates. During December, runoff producing storms in the local watersheds that produce local inflow would either not coincide with storms in the mountains or would coincide with snow in the mountains that causes little mainstem river flow.

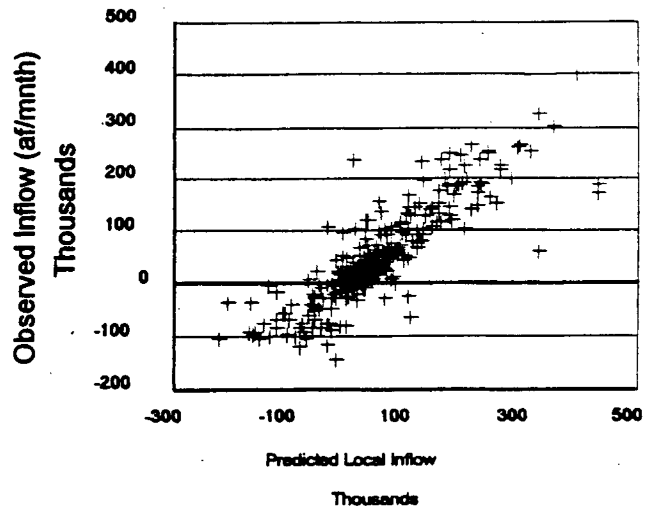

Figure 13: Variation of observed inflow with predicted inflow using the equations in Table 2.

The change from model 1 to model 2 suggests that runoff production from different parts of the Colorado River watershed varies seasonally. Model 2 generally reflects more local inflow production while the magnitude of the coefficients distinguishes among the influence of storms in different watersheds. For example, the Bluff gage is the primary influence on local inflow in January, February, July, August and September. Late winter storms strongly influence the flow in the San Juan River and local inflow. Summer monsoon events have similar effects.

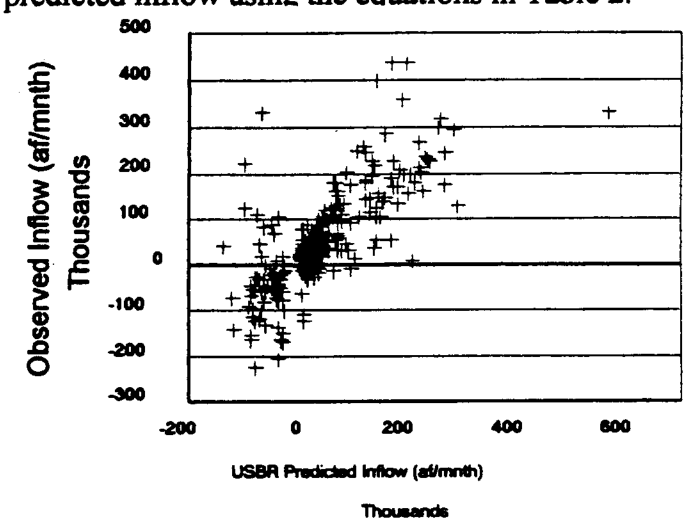

Figure 14: Variation of observed inflow with the inflow predicted by the USBR.

The regression equations in Table 2 yield a closer fit than the USBR estimates for the period 1929 through 1962 (Figures 13 and 14). The predicted mean residual for the regression equations is 0 while for the USBR estimates is -4916 af/mnth. The mean should equal 0.0 so that the estimate does not bias the prediction. The difference indicates the USBR underestimates the local inflow. The standard deviations are 45665 and 59372 af/mnth for the regression and USBR estimates, respectively. In a water balance analysis, the error term is usually the seepage term and would be underestimated by about 4900 af/mnth by using the USBR proportions. The lower standard deviation indicates the absolute magnitude of residuals is closer to 0.0. Residuals of a perfect fit are all 0 while residuals of any imperfect estimate will scatter around 0.0. However, the individual residuals may have a large magnitude. The lower standard deviation of the regression equations suggests that the regression provides estimates that are about 23% better than the USBR proportions.

Conclusion

The analysis in this appendix provides an improved estimate of local inflow into Lake Powell. The regression analysis presented in equations 1 and 2 and Table 2 is about a 23% improvement over the proportional estimates used by the USBR. The regression estimates also do not bias the estimates as the proportions do when considering the 1929 to 1962 flow data. The proportions appeared to add a negative bias of about 4900 af/mnth or about 59000 af/year.

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Appendix 2

Water Balance Calculations for Lake Powell
1963-97

This appendix presents the results of the water balance analysis beginning in January, 1963 and ending in October, 1997. The column headings are described in the following definitions.

San Juan: flow on the San Juan River near Bluff
Green River: flow on the Green River at Green River
Cisco: flow on the Colorado River at Cisco
USGS Inflow: sum of the three mainstem inflows
Lee's Ferry: flow on the Colorado River at Lee's Ferry
res evap: evaporation from Lake Powell based on the average monthly evaporation rate and the surface area of the reservoir based on month ending volume
res precip: precipitation on Lake Powell based on the average monthly precipitation and the surface area of the reservoir based on month ending volume
area: reservoir surface area (acres)
volume: reservoir volume in af
volume change: change in reservoir volume during the month
local inflow: inflow below the three mainstem gages based on regression equations developed in Appendix 1
seepage: calculated seepage from the reservoir
cum seepage: sum of seepage since the beginning of the analysis
water level: elevation of Lake Powell
total inflow: sum of USGS inflow and local inflow

month	year	month	San Juan	Green R	Cisco	USGS inflow	Lee's Ferry	res evap	res precip	area	volume	volume change	local inflow	seepage	cum seepage	water' level	total inflow
January	1963	1	25146	71056	163152	259354	168637	m	m	m	m	m	3192	m	m	m	262546
February	1963	2	39036	119275	192931	351242	368201	m	m	m	m	m	12713	m	m	m	363955
March	1963	3	39711	99277	219008	357996	187466	m	m	m	m	m	6354	m	m	m	364350
April	1963	4	64279	153886	244154	462318	60390	m	m	m	m	m	-79166	m	m	m	383152
May	1963	5	95080	398059	515731	1008869	62014	m	m	m	m	m	-99220	m	m	m	909650
June	1963	6	46621	309177	331353	687151	140184	m	m	m	m	m	72194	m	m	m	759345
July	1963	7	14488	50625	114345	179457	89476	m	m	m	m	m	42097	m	m	m	221554
August	1963	8	48405	71832	167409	287646	61697	m	m	m	m	m	36879	m	m	m	324525
September	1963	9	69413	94686	182299	346397	59776	m	m	m	m	m	34117	m	m	m	380514
October	1963	10	40550	47389	133967	221907	61380	m	m	m	m	m	8761	m	m	m	230668
November	1963	11	46997	73805	178457	299259	59420	m	m	m	m	m	20002	m	m	m	319261
December	1963	12	48264	84075	138085	270424	62608	10163		28074	2775976	0	27532		0	3410	297957
January	1964	1	43782	108484	131729	283995	70785	7047	1456	29120	2915103	139127	26271	94763	94763	3415	310266
February	1964	2	29789	113494	120721	264003	230155	5657	1458	29162	2920791	5688	26057	50018	144782	3415	290061
March	1964	3	28445	128106	128264	284815	387625	8072	1898	28322	2808833	-111957	-5823	-2850	141932	3411	278992
April	1964	4	29676	189565	214038	433279	769230	9805	1729	25804	2480076	-328758	-66513	-81782	60150	3398	366767
May	1964	5	102287	632610	859162	1594058	318305	19444	2254	33640	3536127	1056051	-8115	194397	254547	3435	1585943
June	1964	6	120740	723571	778516	1622828	59481	27575	2914	43493	4980347	1444220	97704	192170	446717	3472	1720532
July	1964	7	113335	343114	275557	732006	59974	33106	3914	47160	5543680	563333	126215	205722	652439	3484	858221
August	1964	8	131482	196099	240570	568151	174062	37804	4075	49096	5845987	302306	83238	141292	793730	3490	651389
September	1964	9	55507	139293	152302	347102	156236	34289	4113	49550	5917329	71342	50787	140135	933865	3492	397889
October	1964	10	36464	195485	163231	395180	268033	25753	3325	49621	5928522	11193	4091	97617	1031482	3492	399271
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January	1965	1	121275	299732	161469	582476	557251	11969	2473	49460	5903107	-26942	95464	138135	1408235	3491	677941
February	1965	2	119651	302267	140303	562221	513691	9627	2481	49621	5928522	25415	38517	54487	1462722	3492	600738
March	1965	3	85170	360776	153985	599930	555212	14139	3324	49612	5926995	-1527	5217	40647	1503369	3492	605147
April	1965	4	164716	517216	561350	1243282	1220036	18733	3303	49299	5877779	-49216	-88148	-31117	1472251	3491	1155134
May	1965	5	287555	817245	1269378	2374178	2279574	28664	3323	49592	5923941	46162	-8457	14643	1486894	3492	2365721
June	1965	6	418433	1205028	1651320	3274781	2319372	35536	3755	56051	6956168	1032227	160393	51794	1538689	3511	3435174
July	1965	7	294683	544995	1114542	1954220	725868	44608	5274	63544	8189674	1233506	254118	209630	1748319	3531	2208338
August	1965	8	217364	227759	445876	891000	869220	49059	5288	63713	8217829	28155	124595	74450	1822769	3531	1015595
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January	1966	1	197663	180992	199247	577903	450351	15790	3262	65246	8474553	38859	390800	466965	2981503	3535	968702

February	1966	2	128918	165409	168518	462845	482308	12590	3245	64899	8416316	-58237	24857	54286	3035788	3534	487702
March	1966	3	198871	392693	277655	869220	620631	18769	4412	65856	8577035	160720	-26979	46534	3082322	3537	842241
April	1966	4	252054	389090	437659	1078803	823878	25371	4473	66766	8730460	153424	-46544	34058	3116381	3539	1032259
May	1966	5	266231	564538	695633	1526402	975942	39923	4628	69071	9120634	390174	-84286	40704	3157085	3545	1442116
June	1966	6	126661	324779	428116	879556	752677	43881	4637	69213	9144783	24149	89081	152567	3309652	3545	968637
July	1966	7	53474	146144	185011	384629	656350	47165	5576	67187	8801384	-343399	73994	104083	3413735	3540	458623
August	1966	8	43643	146223	119632	309498	680665	49868	5375	64763	8393639	-407745	35178	27263	3440998	3534	344675
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January	1967	1	58250	195822	145867	399938	613246	14136	2921	58414	7341403	-247291	38540	61309	3664455	3517	438479
February	1967	2	64091	168696	135452	368238	533432	11174	2880	57596	7207649	-133754	17818	-21915	3642539	3515	386057
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May	1967	5	77533	502979	460746	1041258	877061	32039	3714	55431	6855788	100945	-105129	-70201	3512019	3509	936129
June	1967	6	88963	1132164	712285	1933413	697019	39852	4211	62857	8075202	1219414	132600	113940	3625959	3529	2066013
July	1967	7	39236	506662	326462	872360	639995	45055	5327	64181	8296131	220929	154311	126019	3751978	3533	1026671
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January	1968	1	35555	248292	204930	488777	632095	14831	3064	61284	7813904	-126953	-12855	-40986	3950518	3525	475922
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March	1969	3	143055	353767	249361	746183	707157	19576	4602	68687	9055530	54921	6702	-24167	4280066	3544	752885
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May	1969	5	270270	1093356	985644	2349270	761805	46833	5429	81027	11186308	1481592	41736	106204	4337450	3571	2391006
June	1969	6	237422	682922	729808	1650152	873774	53946	5701	85088	11902144	715836	131228	143525	4480974	3580	1781380
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March	1970	3	115771	193248	276170	585189	485298	23063	5422	80923	11168130	93844	17297	5702	4925880	3571	602486
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June	1970	6	138184	1017126	1337490	2492800	798633	60142	6356	94862	13649766	1557796	86636	169220	5200298	3599	2579436
July	1970	7	73842	419542	536758	1030143	767210	67247	7951	95794	13818085	168319	174025	209342	5409640	3601	1204168
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March	1972	3	119275	321988	278487	719750	376834	29152	6853	102289	14999086	328661	961	-7082	6622661	3613	720711
April	1972	4	64493	323096	201287	588876	780971	38411	6773	101082	14778713	-220373	-12223	-15584	6607077	3611	576653
May	1972	5	80954	634234	452153	1167341	900306	58858	6823	101830	14915146	136433	-97977	-19410	6587666	3612	1069364
June	1972	6	117911	832214	757489	1707613	861181	67315	7114	106175	15711661	796515	108867	98584	6686250	3620	1816481
July	1972	7	17054	246055	191189	454297	913057	72872	8616	103807	15276865	-434795	88165	-56	6686195	3615	542463
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March	1973	3	177071	341411	239897	758380	1093198	26883	6320	94327	13553308	-245464	5472	-104445	6301512	3598	763852
April	1973	4	260073	302485	387288	949846	1674684	34364	6059	90432	12853516	-699792	-43987	-97338	6204173	3591	905859
May	1973	5	484783	1165606	1553904	3204293	646708	60358	6997	104426	15390481	2536965	18874	-13867	6190307	3617	3223167

June	1973	6	462746	1067418	1554498	3084662	749331	73665	7785	116191	17567419	2176938	167678	260191	6450497	3636	3252340
July	1973	7	397604	519829	797247	1714680	655004	84945	10043	121004	18468124	900705	251815	335885	6786383	3644	1966495
August	1973	8	221404	302762	330026	854192	565607	93761	10107	121768	18611615	143491	140141	201581	6987963	3645	994333
September	1973	9	194416	232135	220097	646648	423502	84604	10148	122260	18704109	92494	16112	72307	7060271	3646	662760
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January	1974	1	132779	230591	310999	674368	844490	29753	6147	122947	18833188	-212489	-11701	7060	7511299	3647	662667
February	1974	2	92110	150044	293575	535729	298683	24027	6192	123849	19003043	169854	-27357	22000	7533298	3648	508371
March	1974	3	103257	299594	362716	765567	387803	35787	8413	125570	19327561	324518	-51925	-26053	7507245	3651	713642
April	1974	4	65118	355885	360677	781680	493673	48015	8466	126355	19475813	148252	-80466	19739	7526985	3652	701214
May	1974	5	105556	1176912	1014552	2297020	802039	76880	8912	133011	20737949	1262136	103688	268565	7795550	3662	2400708
June	1974	6	68952	890525	745609	1705085	912107	86529	9144	136482	21399748	661800	101263	155056	7950606	3667	1806348
July	1974	7	39186	268231	312880	620296	1224234	93527	11058	133229	20779444	-620304	115199	49097	7999703	3663	735496
August	1974	8	24910	192040	160420	377370	1210572	99062	10678	128652	19910356	-869088	54421	1923	8001626	3656	431791
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March	1975	3	111547	248965	240016	600528	506722	34859	8195	122311	18713738	91329	24065	-122	8140274	3646	624593
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June	1975	6	396218	1250766	1241064	2888048	985545	87213	9217	137560	21605864	1723183	190627	291950	8398628	3669	3078675
July	1975	7	360756	896980	805088	2062823	1218690	99494	11764	141730	22404774	798910	304667	262161	8660789	3675	2367491
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July	1996	7	48900	287654	477041	813596	985842	105468	12470	150239	24045195	-181917	143967	60639	10338873	3686	957563
August	1996	8	29181	144837	192654	366672	909077	113245	12207	147072	23433087	-612109	54642	23308	10362180	3682	421314
September	1996	9	51054	152163	253064	456281	821027	100306	12031	144951	23024240	-408847	77417	33243	10395423	3679	533698
October	1996	10	93203	175131	284962	553295	665141	74745	9649	144018	22844633	-179607	56573	59238	10454661	3678	609868
November	1996	11	58212	225581	290525	574319	631937	59184	9625	143650	22773901	-70732	60431	23985	10478646	3677	634750
December	1996	12	47130	216117	267439	530686	905058	51280	9491	141658	22391084	-382818	35673	2329	10480975	3675	566358
January	1997	1	50983	225443	277180	553606	1070586	33663	6955	139102	21900843	-490240	-87985	-141432	10339543	3671	465621
February	1997	2	45075	256608	259439	561122	1236312	26326	6785	135699	21250371	-650472	-6094	-50353	10289190	3666	555028
March	1997	3	156372	459617	469379	1085369	1529154	38087	8954	133638	20857409	-392962	-199764	-279720	10009470	3663	885605
April	1997	4	158657	561132	726403	1446192	1299870	51152	9019	134611	21042758	185349	-117703	-198863	9810607	3665	1328489
May	1997	5	344896	1325412	1655082	3325390	1287594	83593	9690	144625	22961473	1918715	44162	89340	9899947	3679	3369552
June	1997	6	474368	1464408	1920204	3858980	1482822	98881	10450	155964	25155835	2194363	179970	273334	10173282	3694	4038950
July	1997	7	144500	430769	641698	1216967	1319868	109176	12908	155521	25069706	-86130	193770	80732	10254014	3694	1410738
August	1997	8	153014	250094	462013	865121	1258587	118236	12745	153554	24687642	-382063	103753	-13141	10240873	3691	968874
September	1997	9	163677	368458	429502	961637	1134164	105951	12708	153108	24601190	-86453	109179	-70139	10170734	3690	1070815