Analysis of the Winter Low-Flow Balance of the Semi-Arid White River, Nebraska and South Dakota

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ABSTRACT

Low-flow studies improve our understanding of flow paths during critical base-flow periods and are needed to assess the effects of water consumption on stream flow, water quality, groundwater resources, and contaminant transport. The inflows to a river from its subbasins and corresponding alluvial aquifers in a semi-arid cold region are most readily quantified in winter. We investigated the low-flow water balance of eight subbasins of the White River at a monthly time scale over seven consecutive winters. Water going into or out of storage as ice or melt, obtained with a temperature index model, can be a dominant component of the water balance. The point estimate method is used to account for parameter uncertainty and variability, providing the mean, variance and limits of dependent variables such as water storage as ice and inflow from a subbasin. Negative water yield from subbasins of several thousand square kilometers occurred regularly through the period, indicating a significant flow from the river to the alluvial aquifers. We discuss the winter water balance by subbasin and between years. The results suggest a perched river or a coupled surface water–groundwater hydrologic system in particular subbasins, consistent with the field investigations of Rothrock (1942). The winter flow exchange between the surface and subsurface can be used to estimate the annual exchange for both conditions.

KEY WORDS: Perched rivers, Point estimate method, River ice growth

INTRODUCTION

Water resource development in semi-arid regions can lead to declining groundwater levels and streamflow in valleys with permeable soils and interconnected surface–subsurface flow systems, without a corresponding decrease in precipitation. Evapotranspiration losses, water withdrawals, and irrigation return flow affect both the stream and shallow groundwater levels. Soil and geologic characteristics determine the total near-surface aquifer storage, the groundwater recharge from precipitation, and the locations and rates of exchange between groundwater and surface water. For a given basin, the annual precipitation and its distribution in time and space determine the quantity of water in storage, and the air temperature regime affects the rate of water loss through evapotranspiration and the storage of water as snow and ice. Planners need tools for evaluating the effect of proposed changes in water usage in a basin on river flow, potential aquifer yields, water quality, contaminant migration and other issues (ASCE 1980). In regions with water shortages, an improved understanding of flow pathways and the effects of water consumption is especially important. During periods of low streamflow the surface water is of largely groundwater origin. However, groundwater recharge and discharge are difficult parameters to quantify. Sophocleous and Perkins (1993) developed a coupled stream–aquifer model with an annual time step and applied it to bound the hydrologic budget imbalance resulting from irrigation development in Kansas. Lacher et al. (1994) measured streamflow, evaporation rates, soil conductivities, pumping rates, and well hydrographs to estimate the rate of aquifer recharge from the Santa Cruz River in Arizona. Abdulrazzak and Sorman (1994) used a water balance approach to estimate flood water losses from ephemeral streams in arid regions, but large spatial and temporal parameter variability caused difficulties.

Low flows typically occur in the same season each year. The late summer and winter are low-flow periods in the northern United States and southern Canada (Rogers and Armbruster 1990, Melloh 1990, Wuebben et al. 1992). Kuusisto (1986) reported mean winter to summer low flow ratios that decrease significantly with distance north in Finland. Winter has been

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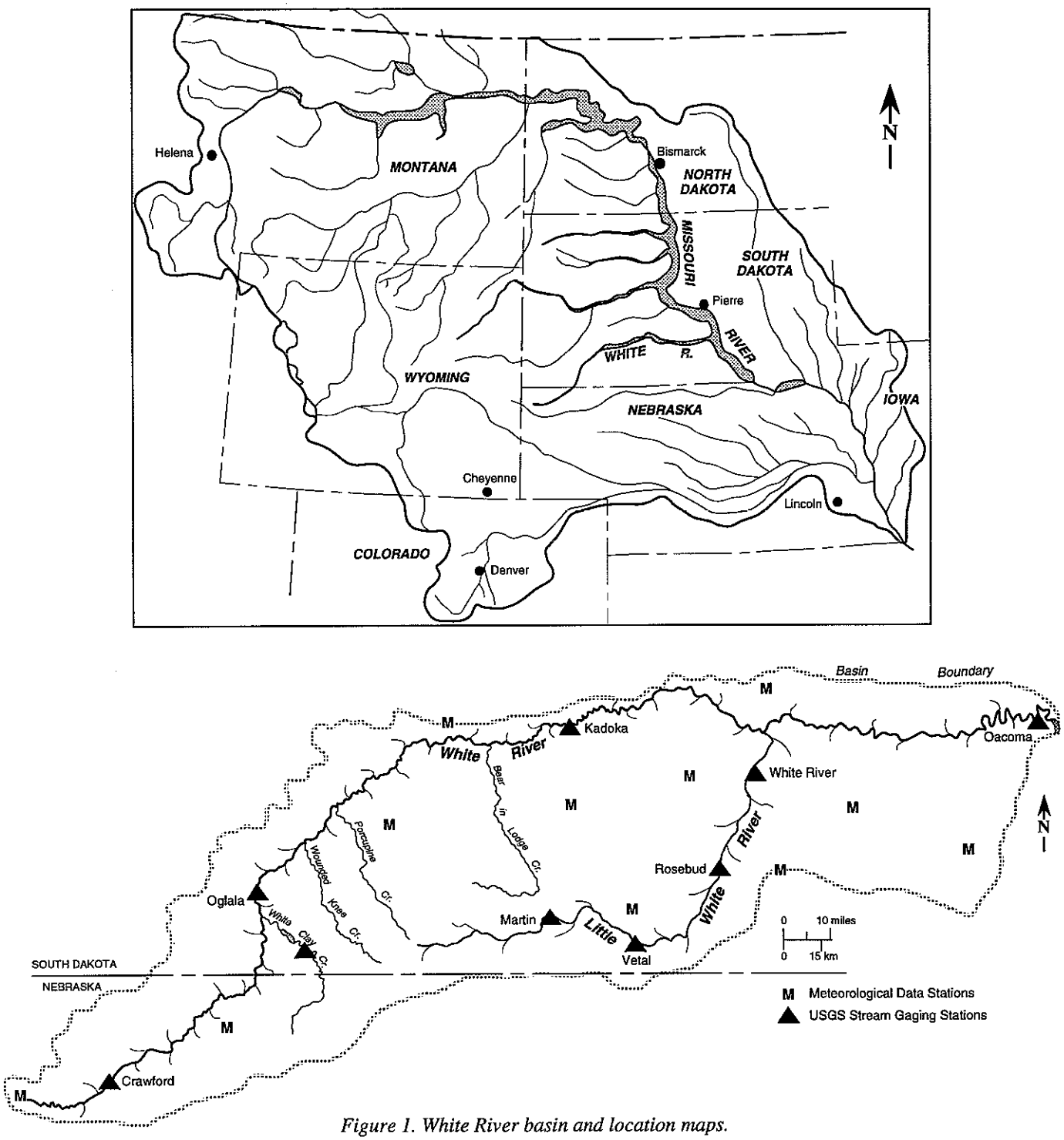

Figure 1. White River basin and location maps.

generally considered a hydrologically dormant period and has not been extensively studied. However, the exchange of water between a river and its near-surface aquifers is most readily quantified during the winter months. The winter low-flow water balance is simplified because there is negligible evapotranspiration, irrigation water withdrawals and diversions are halted, and precipitation occurs largely in the form of snow, minimizing the spatial and temporal variability of surface runoff. A complication is that the ice produced in the river can be a large component of the water balance for semi-arid basins in even moderately cold regions. The White River in Nebraska and South Dakota, an uncontrolled tributary of the Missouri River, has a basin of

26,400 km^2, but typical winter monthly average flows are less than 3 m^3/s. The simple characterization of basins by drainage area and precipitation is useful only in homogeneous basins. Subbasin yields at times of low flow can vary widely within a relatively small basin as a result of diverse water-bearing properties of underlying soils and rocks (Schneider 1965, Gerard 1981). Riggs (1972) suggested low-flow discharge measurements at several locations along a stream to define the base flows and the hydrologic homogeneity or heterogeneity of a basin. The White River basin is heterogeneous, but flows throughout are low and stable in winter.

In this paper we will investigate the winter water balance in eight subbasins of the White River. The water balance is written as a monthly average for river reaches bounded upstream and downstream by flow gages. The flow contributions from the corresponding subbasins and the water storage in the river due to ice production are computed for a series of seven winters, from November 1974 to February 1981. Water going into or out of storage as ice or melt is calculated from a temperature index model applied to the basin. The point estimate method (PEM) of Rosenblueth (1975) allows us to apply deterministic relations for ice growth or melt, water storage as ice, and subbasin water balance while still accounting for uncertain or variable parameters in the calculations and flow measurements. The PEM provides an expected value, variance and estimated limits of the probability distributions that characterize the dependent variable in each of these calculations. Our objective is to quantify the hydrologic balances in the White River basin, including the surface-subsurface flow exchanges.

GENERAL HYDROLOGY OF THE BASIN

The White River basin lies in an unglaciated part of the Missouri Plateau characterized by undulating uplands and wide floodplains along the larger streams. Location and basin maps for the White River are given in Figure 1, indicating the basin boundaries, the primary tributaries, and the USGS stream gaging stations and meteorological data stations used in this study. The streamflow gages on the main river and the Little White River are listed in Table 1, along with the drainage area of each nested subbasin, the gage datum, the approximate river location of each gage, the average channel slope, the annual and winter average discharges for the period of record, and the linear distance between gages. The winter average discharge was obtained from November through February monthly averages. The channels of the White and Little White Rivers are highly mobile within the floodplain, and river location is not published by the USGS. The approximate river locations in Table 1 were obtained from recent maps using a map wheel. The White River is generally more sinuous than the Little White River,

Table 1. Basin, flow and river parameters corresponding to USGS gages on the White and Little White Rivers.

Location	*Drainage area (km^2)*	*Avg discharge (m^3/s)*			*Gage datum or elev. (m)*	*Approx. length of river (km)*	*Approx. river slope*	*Linear distance (km)*	*River/linear distance ratio*
		Annual	*Winter*	*Ratio*					
White River									
Crawford	810	0.57	0.59	1.04	1115.5	804			
Cr-Og	4890	0.96	0.06	—		196	0.0013	78.9	2.48
Oglala	5700	1.53	0.53	0.35	869.8	608			
Og-Ka	7300	6.03	1.21	0.20		260	0.00086	118.5	2.19
Kadoka	13000	7.56	1.74	0.23	646.8	348			
Little White River									
Martin	800	0.54	0.38	0.71	928.1	196			
Ma-Ve	700	0.99	0.69	0.70		62.0.0013	33.2	1.87	
Vetal	1500	1.53	1.07	0.70	847.6	134			
Ve-Rb	1140	1.61	1.40	0.87		65.6	0.0022	37.6	1.74
Rosebud	2640	3.14	2.47	0.79	699.5	68.4			
Rb-WR	1430	0.48	0.17	0.35		45.1	0.0026	32.5	1.39
WhiteRiver	4070	3.62	2.64	0.73	583.0	23.3			
WR-Confl					541.0	0.0	0.0018		
White River									
Ka+WR	17070	11.2	4.37	0.39					
KaWR-Oa	9330	3.7	0.90	—					
Ka-Oa						342	0.00066	158.0	2.16
Oacoma	26400		14.9	3.47		0.23	419.8	6	

reflected in generally higher river/linear distance ratios. The Little White River below Martin has an average slope of 0.0020, more than twice that of the White River below Crawford at 0.00087.

Sando (1991) detailed the surface water diversions and groundwater withdrawals for irrigation over the last 50 years in the White River basin upstream of the South Dakota state line. Water use is zero for November through April but currently averages 50% of the average flow at the state line gage for the remainder of the year. All tributaries in the basin above the state line are ephemeral (Sando 1991), and more generally, streams in the White River basin are unreliable sources of water (Ellis and Adolphson 1971). Most streamflow occurs in response to precipitation and runoff during spring and early summer, with a large component of the March flow due to snowmelt. Rothrock (1942) reported extensive field investigations throughout the river valley over a 43-km reach from Kadoka downstream. The alluvium there is between 7.5 and 12.5 m deep, composed of mostly sands and gravels, overlying Pierre shale. Near-surface aquifers in the alluvial deposits are permeable and readily exchange water with the river. The groundwater level is below the river surface during dry seasons, rising to river level during wet seasons. When river flow ceases, deep pools remain wet, reflecting the water table level. Short reaches of the perennial Bear-in-the-Lodge, Porcupine, Wounded Knee and White Clay Creeks intercept the water table, allowing seepage from groundwater to supplement surface runoff. The larger tributaries supply groundwater with a different chemical signature directly to the alluvium of the White River. The wind-blown sand deposits in the Little White River basin above Rosebud are permeable, minimizing surface runoff and providing more consistent flows than found elsewhere in the basin (Ellis et al. 1971).

The subbasin above the Crawford gage has consistently stable flows throughout the year. Significant groundwater input to the river and minimal surface runoff cause this stability, interrupted only occasionally by large spring and summer events. The annual hydrograph of monthly average flows at Oglala has peaks in both March and June and low flows in the fall and winter. The average winter flow at Oglala is less than that at Crawford (Table 1), even though the basin is seven times larger. Rothrock (1942) reported that "in the upper part of the valley" a considerable flow will frequently disappear within 35 km due to groundwater recharge. The annual hydrograph of monthly average flows at Kadoka also has double peaks and the same general shape as that at Oglala, with flow decreasing through the fall and into midwinter. However, the increase in discharge between these gages is generally significant, due largely to flow contributions of the perennial creeks. The composite response of the Little White River is indicated by gage WR at 23 km above the White River confluence. Spring flows on the Little White are high and variable, while fall and winter flows are lower and more stable. Summer flows are occasionally high but generally consistent with the groundwater-inflow-dominated fall and winter conditions. The annual hydrographs of monthly average flow for all gages in the Little White subbasin have single peaks in either March or April. The monthly average White River flow at the Oacoma gage near the mouth can vary dramatically between seasons, especially spring and summer, and years. Winter flows are more consistent and extremely low by comparison, with 0.6 m^3/s at Crawford and 2.6 m^3/s from the Little White River providing almost the entire flow at Oacoma from subbasins representing only 18% of the total drainage area.

Figure 2 provides a breakdown of annual and winter average water yields of sequential subbasins of the White River basin, delineated by the primary streamflow gages. The highest annual and winter water yields in the entire basin occur in a subbasin of the Little White River between Martin and Rosebud. This subbasin has an annual water yield 2.5 times greater, and a winter yield 8.7 times greater, than those of the complete basin. The ratio of winter to annual average discharge is greater than 1 for the subbasin above Crawford and greater than 0.7 for all subbasins of the Little White River above the gage at Rosebud. However, for

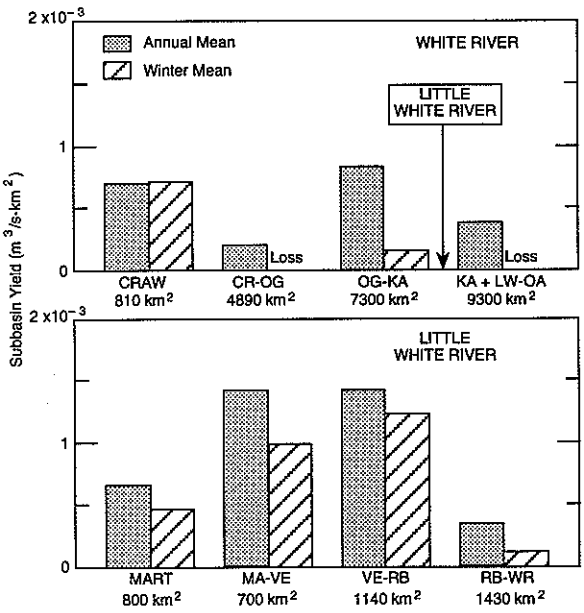

Figure 2. Annual and winter average water yields of sequential subbasins of the White River basin.

most of the mainstem White River and the Little White River below Rosebud, the winter flows are only 23-35% of the annual average. The subbasin between Crawford and Oglala has the lowest annual yield in the entire basin and a negative winter yield. Other subbasins with low yields both annually and in winter are the Little White below Rosebud and the White below Kadoka. Unlike the other mainstem subbasins below Crawford, the subbasin between Oglala and Kadoka has a high annual yield and a positive winter yield. The hydrogeologic maps of Ellis and Adolphson (1971) and Ellis et al. (1971) support the conclusion that the primary reason for the widely differing yields of the White River subbasins is differences in near-surface geology.

RIVER ICE GROWTH AND MELT

The extreme low flows in winter on the main stem of the White River greatly increase the importance of water storage as ice in the water balance. We will now develop a method to quantify the storage of water as ice and its release as melt. Temperature index models provide good estimates of the growth or melt of river ice, and we use the physically based equation given by Ashton (1989):

$$\frac{dh}{dt} = \frac{1}{\rho_i L}\frac{(T_m - T_a)}{\left(\frac{h}{k} + \frac{1}{H_{ia}}\right)} \qquad (1)$$

where h = ice thickness (m),
t = time (s),
k = is thermal conductivity of ice (W/m °C),
L = is latent heat of fusion (J/kg),
ρ_i = density of ice (kg/m^3),
H_{ia} = ice-air heat transfer coefficient (W/m^2 °C),
T_m and T_a = ice melting point and air temperatures (°C), respectively.

Integrating (1) we obtain the final or end-of-the-month ice thickness h_f as

$$h_f = \left[h_i^2 + \frac{2k}{H_{ia}}h_i + \left(\frac{k}{H_{ia}}\right)^2 - \frac{2k}{\rho_i L}T_a\Delta t\right]^{1/2} - \frac{k}{H_{ia}} \qquad (2)$$

where h_i is a given initial or start-of-the-month ice thickness, and Δt is the monthly time increment (s). In the case of ice melt, the ice thickness does not resist the flow of heat. The h/k term in the denominator of (1) is deleted, and integration leads to

$$h_f = h_i - \frac{H_{ia}}{\rho_i L}T_a\Delta t \; . \qquad (3)$$

With h_f known from either (2) or (3), we can obtain the monthly average flow Q_{ice} that has gone into or out of storage as ice or melt:

$$Q_{ice} = \frac{\Delta x}{\Delta t}\frac{(B_{in} + B_{out})}{2}(h_f - h_i) \qquad (4)$$

where Δx is the reach length (m), and B is channel width at the upstream (in) and downstream (out) ends of the reach. Equation (4) assumes that the average width of the river in a reach can be obtained by averaging the widths at each end.

WINTER SUBBASIN WATER BALANCE

The annual and winter water yields for the period of record in eight subbasins of the White River indicated a wide range of hydrologic conditions in the basin. In particular, major differences in subbasin yields were evident in the winter. We will now develop a detailed winter water balance on a monthly time scale that includes the changes in water storage in the river channel that accompany increases or decreases in flow and the formation or melt of river ice, and the subbasin flow exchange with the river. The effects of unsteadiness on the water balance are negligible during low-flow periods and will be neglected. The net inflow to the river from a subbasin Q_{sub} has tributary and groundwater components:

$$Q_{sub} = Q_{gw} + Q_{t1} + Q_{t2} + Q_{t3} + \ldots \qquad (5)$$

where Q_{gw} is groundwater inflow and Q_{t1}, Q_{t2} and Q_{t3} are tributary inflows.

The flow storage in the channel Q_{st} caused by significant changes in the monthly average flow can be computed for a river reach as

$$Q_{st} = \frac{\Delta x}{\Delta t}\left[\frac{B_{in}\Delta Y_{in} + B_{out}\Delta Y_{out}}{2}\right] \qquad (6)$$

where ΔY (m) is the channel depth change at the upstream (in) and downstream (out) ends of the reach, assuming that depth change can be adequately described by averaging the end values. The depth changes can be determined from the measured average discharge at each gage for the present and previous months and corresponding river stage data.

The winter water balance for an incremental subbasin delineated by a pair of stream gages is depicted

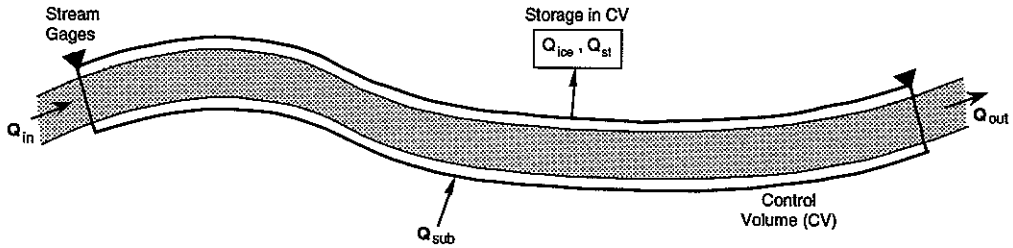

Figure 3. Schematic diagram of the control volume used to obtain the winter water balance for a river reach and its corresponding subbasin.

in Figure 3 and written as

$$Q_{in} + Q_{sub} - Q_{ice} - Q_{st} - Q_{out} = 0 \qquad (7)$$

where Q_{in} and Q_{out} are the flows measured at the upstream and downstream gages, respectively. In low-flow months the subbasin flow exchange may be almost exclusively with the groundwater. As tributary inflows are always nonnegative, $Q_{sub} < 0$ implies groundwater recharge from the river.

All of these computations are nested, with changes in ice thickness obtained for a given month and then used to find Q_{ice}. With Q_{st} and Q_{ice} known from (6) and (4), and Q_{in} and Q_{out} known from gage records, we can obtain Q_{sub} with (7). Finally, tributary inflows are used with Q_{sub} to obtain the groundwater discharge or recharge Q_{gw} from (5).

POINT ESTIMATE METHOD

Mean values can now be determined for Q_{ice}, Q_{st}, Q_{sub}, and Q_{gw}, but they would not account for the known variability of input parameters such as air temperature, heat transfer coefficient, and channel width. Also, many of the measured or estimated parameters contain uncertainty that contributes to the uncertainty of the corresponding dependent variable. We will use the Rosenblueth (1975) point estimate method (PEM) to account for and quantify the uncertainty in our deterministic winter water balance. The independent variables in each deterministic equation that contain uncertainty are considered random variables. The first two or three moments of each random variable and the correlation coefficients between variables are given as input, quantifying the variability or uncertainty. The PEM provides the mean, variance and limits of the dependent variable, uniquely specifying a Beta distribution (Harr 1977) that describes the uncertainty of a function of random variables. The estimated mean value is equivalent to a second-order Taylor series approximation, and the variance is a first-order estimate. The method is algebraic, replacing the distribution of each random variable by point estimates and not requiring the computation of derivatives. The PEM offers several advantages over a deterministic approach. A computed mean value has much greater importance when the variance is small, but variance is unknown in a deterministic model. The random variables contributing most of the uncertainty to the results can be readily identified, which can help to refine data collection. Also, the interpretation of PEM results is straightforward. For example, though a river reach may have positive inflow from its subbasin based on mean values, there may be a significant probability that the flow direction is the opposite.

We apply the PEM to the ice growth or melt in (2) or (3) by considering air temperature, the air-ice heat transfer coefficient and the initial ice thickness as the primary random variables. The mean of H_{ia} was taken as 20 W/m^2 °C with a standard deviation of 5 W/m^2 °C, representative of the data presented by Ashton (1989). Ice density and thermal conductivity are correlated random variables, but their variability is minor. In computing Q_{ice} with (4) the independent random variables are river distance between the gages, channel width at each gage, and correlated initial and final ice thicknesses. The estimated mean river distances between the gages are given in Table 1. Based on multiple trials, the measurement error in obtaining these distances from maps was about 2% of the distance, and in addition, the movement of the river within the floodplains could alter the distances from those shown on the maps by a few percent of the length. Therefore, we assume a coefficient of variation of 5% for reach length. The channel width of the ice/water interface is measured by the USGS each time a discharge rating is done at a gage. The channel cross section and discharge at which these measurements were made varied. We used all available measurements during ice-covered flow conditions to obtain the mean width and its variance at each gage. The mean width varied from 4 m on the Little White River at Martin to 23 m on the White River at Oacoma. The coefficient of variation of the river width varied between 0.11 and 0.46.

The Q_{st} computation in (6) has reach length and stream width, and correlated depth changes at the ends of the reach, as independent random variables. We assume that Q_{st} is generally negligible but evaluate it for

the pairs of months of largest flow increase and decrease in the period of record. Q_{sub} in (7) can be evaluated considering each component of the water balance as a random variable. The measurement error for discharge at the gages in winter is typically about 8%, as reported on the USGS discharge measurement notes. We will use this value as the coefficient of variation for the measured monthly average discharges at the gages. Q_{in} and Q_{out} correlations were computed for each pair of gages over the period of study and have coefficients that increase with distance downstream.

RESULTS

Seven consecutive winters from November 1974 to February 1981 provide a representative range of temperature and hydrologic conditions for analysis. Mean monthly air temperatures are available at the eleven meteorological stations in Figure 1 for the seven-winter study period. To account for the variability indicated by these temperatures, a basin mean temperature and standard deviation were obtained and are presented in Table 2. The coldest winter of the study period was 1978–79, and January 1979 was the coldest month. Five of the six other winters had less than half of the freezing degree-days of this winter. With these temperatures as input we obtained the ice growth or melt for each month, and the mean thickness and standard deviation at the end of the month are given in Table 2. December 1977 and January 1980 were the months of maximum ice growth, and February 1976 was the month of maximum melt in this period of record.

Figures 4 and 5 give results for January 1979, the coldest month of the study. The mean, standard deviation and corresponding Beta distribution for Q_{ice} and Q_{sub} of each mainstem White River reach are presented in Figure 4. The storage of water as ice is an important term in the water balance of the White River below Oglala but is of less significance farther upstream and on the Little White River, where the stream widths are small and the flows are relatively high. The Crawford-Oglala subbasin had a negative yield. The distributions for Q_{ice} and Q_{sub} have similar shapes in the two reaches below Oglala. There is a small probability that the Oglala-Kadoka subbasin had a negative yield and that the Kadoka-Oacoma subbasin had a positive yield. The mean and standard deviation data are repeated in Figure 5 from upstream (left) to downstream (right) together with corresponding flows at the gages. The river flows diminished from Crawford to Kadoka and then recovered somewhat at Oacoma, due to a significant inflow from the Little White River. The mean water storage as ice increased in successive reaches downstream, as did its variance. The variance of subbasin flow exchange also increased in the downstream direction.

Results for each mainstem subbasin are detailed in Figure 6 for 1978-79, the coldest winter of the study. The ice production was consistent over the winter except for a drop in February, with increasing mean and variance in the downstream direction. The mean subbasin flow exchanges were consistently negative for Crawford to Oglala and Kadoka to Oacoma, but positive between. The magnitudes of Q_{ice} and Q_{sub} are generally comparable to or larger than the river flows at the gages for this entire winter. Figure 7 gives corresponding results for 1979–80, another low-flow winter with only average temperatures. The flow storage due to ice growth was reduced somewhat compared to 1978-79, but subbasin yields were generally comparable. Water storage as ice must be considered in a winter water balance for semi-arid regions, but extreme cold may not significantly alter basin hydrologic response.

The Crawford to Oglala subbasin had only 3 of 28 months with a net inflow to the river: February 1976, the month of maximum ice melt in the period with relatively high flows indicating runoff throughout the basin, and two months in the 1979-80 winter. Excluding the four highest inflow months, the mean Q_{sub} for the period was –0.318 m^3/s with a standard deviation of 0.125 m^3/s. The only perennial stream in the reach is White Clay Creek. To obtain Q_{gw} the analysis was repeated

Table 2. Mean and standard deviation for monthly air temperature over the basin T_a (°C) and corresponding ice thickness h_f (m).

Year	*Variable*	*November*	*December*	*January*	*February*
1974–75	T_a	2.8, 1.2	–2.2, 1.1	–4.4, 0.9	–7.2, 0.9
	h_f	0.0, 0.0	0.19, 0.072	0.40, 0.056	0.61, 0.047
1975–76	T_a	1.1, 1.0	–1.7, 1.3	–5.0, 1.4	1.7, 1.2
	h_f	0.0, 0.0	0.15, 0.096	0.40, 0.074	0.12, 0.23
1976–77	T_a	–0.6, 0.6	–2.8, 0.8	–10.6, 1.3	1.1, 1.2
	h_f	0.06, 0.06	0.26, 0.054	0.63, 0.046	0.45, 0.20
1977–78	T_a	0.6, 2.1	–6.1, 1.2	–12.8, 1.4	–10.0, 1.4
	h_f	0.0, 0.0	0.38, 0.052	0.75, 0.046	0.93, 0.045
1978–79	T_a	–3.9, 1.0	–8.3, 1.4	–13.9, 0.9	–8.9, 1.8
	h_f	0.28, 0.051	0.58, 0.051	0.89, 0.041	1.0, 0.045
1979–80	T_a	0.0, 1.7	0.6, 1.3	–6.1, 1.1	–3.3, 1.1
	h_f	0.0, 0.0	0.0, 0.0	0.38, 0.048	0.49, 0.052
1980–81	T_a	3.9, 1.5	–1.7, 1.0	–0.6, 0.7	–1.7, 0.9
	h_f	0.0, 0.0	0.16, 0.073	0.20, 0.080	0.28, 0.076

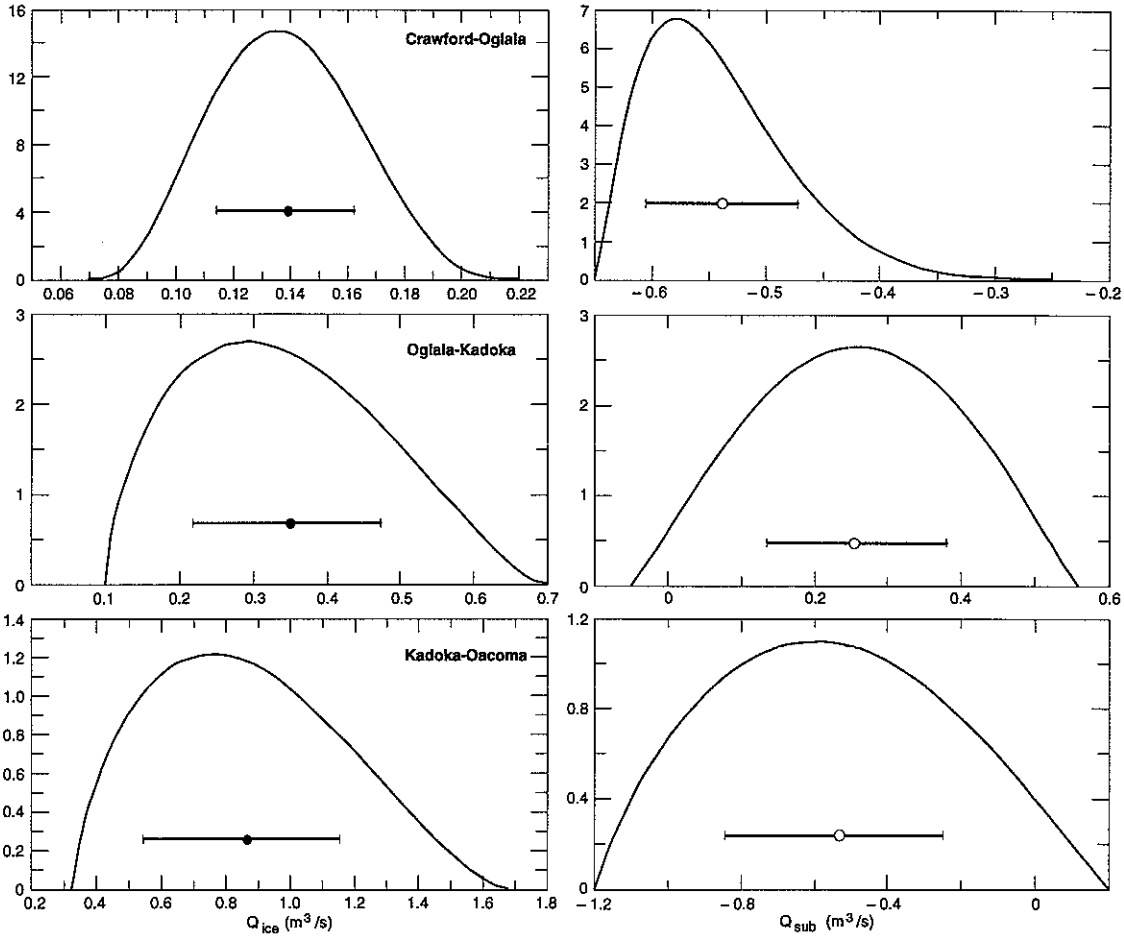

Figure 4. Beta distributions with means and standard deviations indicated for water storage as ice and subbasin inflow in January 1979, by reach between mainstem White River gages. Positive Q_{ice} is water loss due to ice growth, and positive Q_{sub} is flow to the river from the subbasin.

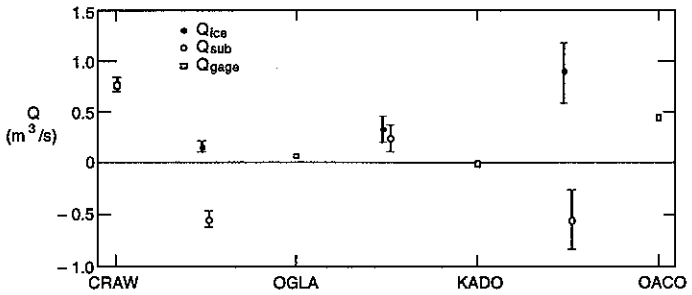

Figure 5. January 1979 monthly flow at mainstem White River gages, water storage as ice and subbasin inflows of the reaches between gages. Both the mean and standard deviation are given for each parameter.

with Crawford plus White Clay Creek providing Q_{in}. With this change the correlation between reach inflow and outflow increased from 0.409 to 0.721. The mean Q_{gw} obtained was –0.483 m^3/s, with a standard deviation of 0.091 m^3/s. Groundwater recharge occurs consistently at a steady rate during winter and represents a significant flow loss from the river.

The Oglala to Kadoka subbasin exhibited consistently positive yields. The three large inflow months were the melt in February 1976, and pre-freezeup combined with relatively high precipitation in November 1977 and 1979. The months of small local inflows were those with low flows at both gages and minor ice production. Excluding the four highest inflow months,

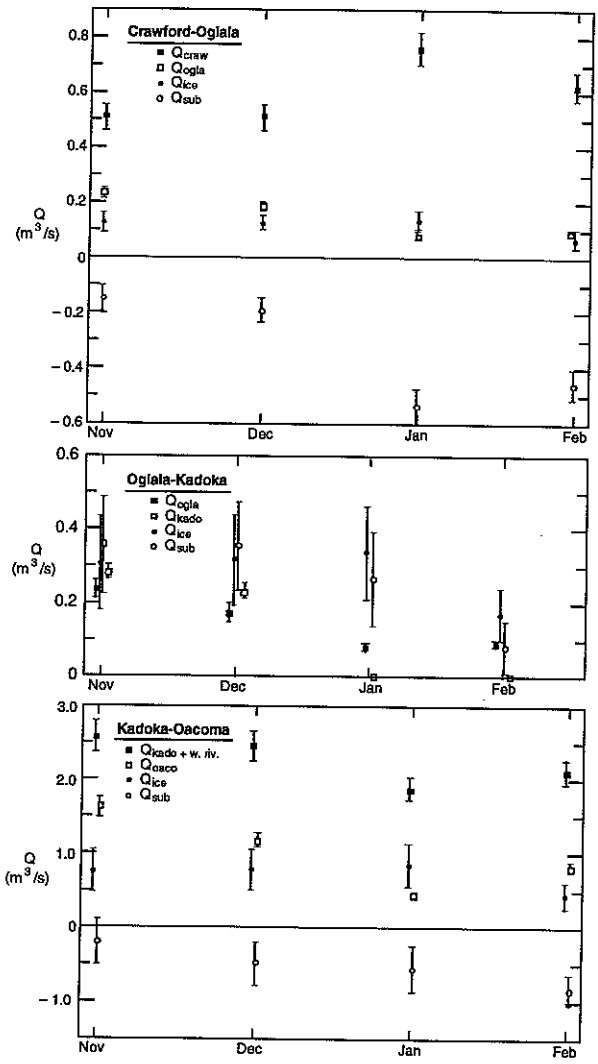

Figure 6. Winter 1978–79 monthly flows at mainstem White River gages, water storage as ice and subbasin inflow for the reaches between the gages.

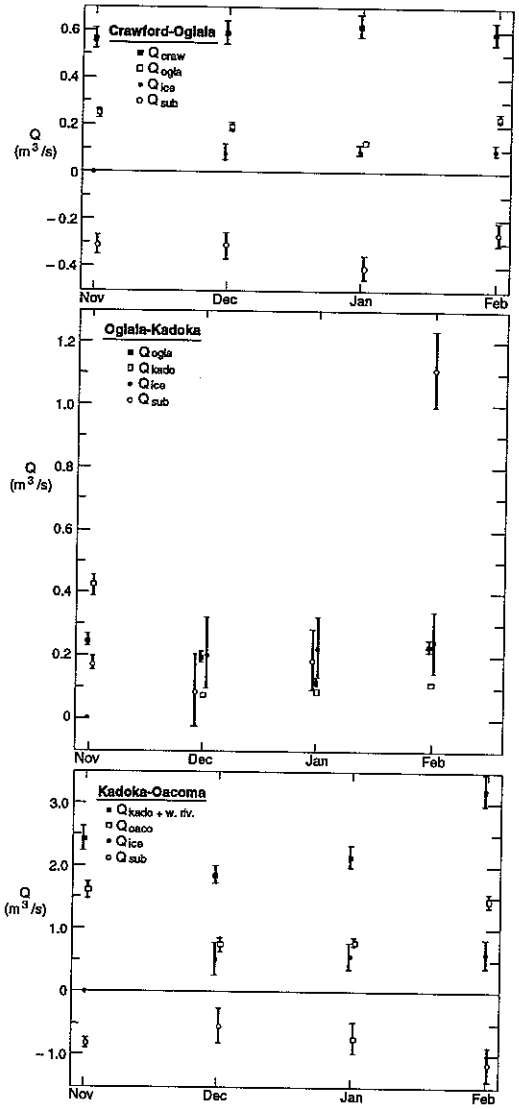

Figure 7. Winter 1974–75 monthly flows at mainstem White River gages, water storage as ice and subbasin inflow for the reaches between the gages.

the overall mean Q_{sub} was 0.39 m^3/s, with a standard deviation of 0.32 m^3/s and all positive monthly means. The river-groundwater exchange in this reach is clearly different from that in the adjacent subbasin upstream, but the quantity and direction are masked by inflow from three perennial creeks. Two of these creeks were gaged in 1992–93, having a combined winter flow 1.7 times that of White Clay Creek. We will use two times the White Clay Creek flow as a conservative estimate of tributary inflow to this reach. The overall mean Q_{gw} was obtained, again excluding the four high inflow months, as 0.11 m^3/s with a standard deviation of 0.31 m^3/s and 12 negative mean inflow months. The river-groundwater exchange in either direction near Kadoka is consistent with Rothrock (1942), but a more quantitative understanding will require additional tributary flow data.

The monthly subbasin inflows on the Little White River above Rosebud are always positive. The mean flows downstream at the gage below White River are usually positive, but there is a significant probability of negative flows in almost all months of the study period. These significantly reduced subbasin yields represent a transitional behavior between the high yields of the Little White basin upstream and the low yields of the White River subbasin immediately downstream. The Little White River inflow to the Kadoka-to-Oacoma reach is typically much greater in winter than the White River mainstem flow at Kadoka. Significant positive subbasin inflows to this downstream-most White River subbasin were computed for February 1976, November 1977 and November 1979, the same months as the adjacent subbasin upstream. However, the local monthly subbasin inflows are again

generally negative in the winter. Flow losses in 7 of 28 months exceeded 1.1 m^3/s, with a maximum loss of 2.5 m^3/s. Neither the coldest months nor those with maximum ice growth correspond to months of maximum flow loss from this reach.

We will now consider the assumption of negligible channel storage in the water balance. The problem with quantifying this term is that the stage change corresponding to a change in discharge may not be readily obtainable. For most of the seven study winters the change in average flow of consecutive months was small throughout the basin, and Q_{st} could be neglected without significant error. The largest flow increase of the study period on the mainstem White River occurred between January 1976, a cold, low-flow month, and February 1976, a month of significant melt, runoff and relatively high flows. The consecutive monthly flows at Oglala and Kadoka increased by factors of 8 and 29, respectively. We obtained Q_{st} using USGS gage rating data from these months. The mean subbasin inflow increased from 9.55 to 9.71 m^3/s when channel storage was considered, a change of less than 2 percent. Similarly, a flow decrease by a factor of 20 occurred at Kadoka between October and November 1980. Mean subbasin inflow decreased from 0.24 to 0.02 m^3/s with channel storage included in the water balance. The direction of Q_{gw} can be reversed by the channel storage term for decreasing flow conditions.

HYDROLOGIC IMPLICATIONS

The winter hydrologic balance is useful for quantifying the seasonal flow exchanges, but annual exchanges are needed for water resource evaluation. The balance between groundwater consumption and recharge determines the long-term availability of the resource. Conversely, increased flow losses from the river affect water quality, aquatic habitat and surface water availability. Figure 8 depicts the possibilities for a near-surface aquifer that is recharged by the river during winter. The first condition is the river perched above the water table throughout the year, resulting in a continuous loss of water from the river by unsaturated flow. For given alluvial bed conditions the flow loss would be proportional to the wetted perimeter and depth of the river, each generally increasing with river flow. The other possible condition is a coupled stream-aquifer system, with a fluctuating relationship between river and groundwater levels throughout the year. In the case depicted, a summer of groundwater withdrawals and insufficient recharge has caused the water table to fall to a minimum by the end of October. The end of the irrigation season, together with recharge from the river over the winter, causes a recovery of the levels. By April the water table is at river level, and it continues to rise through June. After that time groundwater discharge and withdrawals cause the levels to fall, reaching river level in July and continuing down with sustained withdrawals toward the fall minimum. Flow loss from the perched river is depicted in Figure 9,
together with the mean annual hydrograph of the White River at Oglala. With a perched river the winter flow loss extrapolated through the year would provide a lower-bound estimate of annual groundwater recharge from the river. The case of a coupled hydrologic system is depicted in Figure 10. "Net groundwater loss from irrigation" in this figure is the withdrawal restricted to the period with the water table at or below river level. An extrapolation of the winter losses would provide an upper bound for recharge from the river.

The net inflows of three mainstem subbasins are given in Table 3 for relatively dry months (January 1977 and 1979) and wet months (February 1976 and November 1977). Tributary inflows considered apart from the remainder of the subbasin were White Clay Creek (WCC), 2x White Clay Creek (2WCC) and the Little White River at WR. In a given month with no additional tributary inflows, the subbasin flow exchange in Table 3 is the groundwater exchange. With few exceptions the Crawford-Oglala subbasin had consistent net unit losses averaged over the channel area between 4.0 and 4.9×10^{-5} cm/s, a midrange seepage velocity for fine sands and silts (Bear 1972). The evidence supports the hypothesis of a predominantly perched river through this reach. Small, variable flow losses and yields from the other two subbasins with significant uncertainties suggest more complex coupled hydrologic systems in these reaches. Additional measurements of relative river-alluvial aquifer levels over the year, local groundwater withdrawals, and tributary inflows would reduce the uncertainty in the winter hydrologic balance and allow reliable estimates of the annual flow exchanges.

CONCLUSIONS

The semi-arid White River basin is heterogeneous, with highly variable annual and winter average subbasin yields caused by differences in soils and underlying rocks. However, winter is the season of minimum flows throughout the basin. The winter water balance is simplified because of the absence of quantities that are large in other seasons and have large uncertainties. We have developed a methodology for quantifying inflow from a subbasin, and the relatively small but important flow exchange between the river and its alluvi-

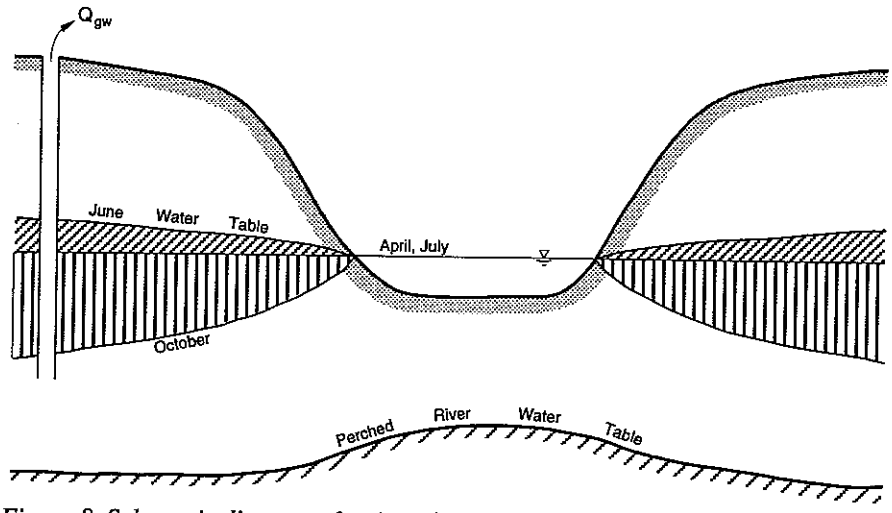

Figure 8. Schematic diagram of a river that recharges the groundwater during winter. Groundwater withdrawal is indicated. The river may be perched above the water table or directly connected to the water table in a coupled hydrologic system. Water tables in the diagonal shading discharge to the river, and those in the vertical shading or below are recharged by the river.

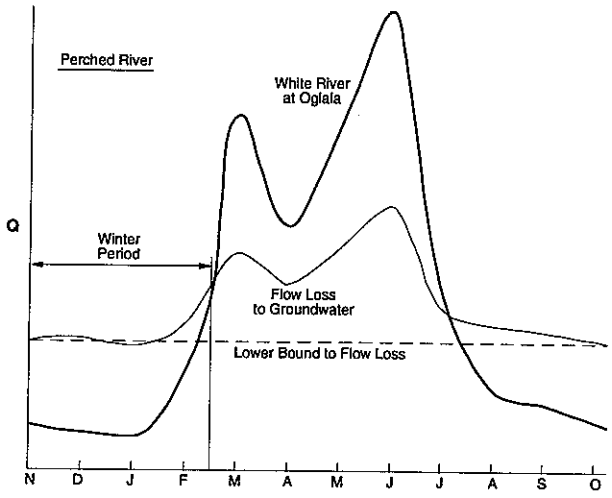

Figure 9. Hypothetical flow loss from the river to the groundwater for a perched river. The mean annual hydrograph for the White River at Oglala is used as a reference. The flow loss during winter extended over the year provides a lower bound for the annual flow loss.

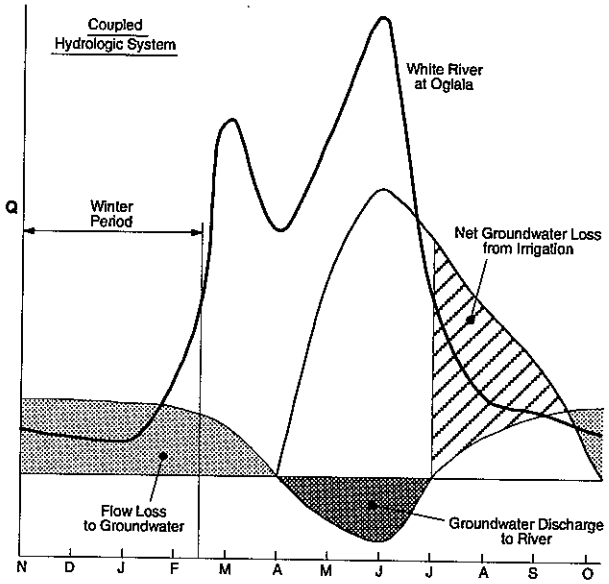

Figure 10. Hypothetical flow exchange between the river and the groundwater for a coupled hydrologic system. The mean annual hydrograph of the White River at Oglala is used as a reference. The lighter shading represents volumetric flow loss from the river, in balance with net volumetric groundwater loss from irrigation (diagonal shading), and the darker shading represents volumetric groundwater flow to the river.

al aquifer, by month through the winter. Important elements of the method are a winter water balance equation with a river ice growth-melt term, and a point estimate method that allows the application of deterministic models to problems with variable or uncertain parameters.

The variable severity of the winters in our seven-year study period did not significantly affect the water balance. Water storage as ice is generally a dominant component of the water balance on the mainstem White River below Oglala, where the channel becomes wide. The large Crawford to Oglala and Kadoka to Oacoma subbasins on the mainstem did not contribute flow to the river in most months of the study period. Even relatively mild winters did not produce inflows from these subbasins, unless a runoff event occurred.

Table 3. Mean, standard deviation for Q_{sub} and net subbasin yield or flow loss per unit area.

Subbasin	Q_{sub} (m^3/s)	Net yield $\times 10^{-4}$ (m^3/s–km^2)	Net unit loss $\times 10^{-7}$ (m^3/s–km^2)
February 1976			
Cr + WCC –G	0.49, 0.12	1.01, 0.25	
Og + 2WCC – Ka	9.05, 0.86	12.4, 1.18	
Ka + WR = OA	0.53, 0.70	0.57, 0.75	
January 1977			
Cr + WCC –G	–0.48, 0.06		–4.09, 0.50
Og + 2WCC – Ka	0.13, 0.15	0.18, 0.21	
Ka + WR = OA	–0.20, 0.34		–0.30, 0.50
November 1977			
Cr + WCC –G	–0.56, 0.05		–4.84, 0.45
Og + 2WCC – Ka	1.94, 0.18	2.65, 0.25	
Ka + WR = OA		1.81, 0.20	1.94, 0.21
January 1979			
Cr + WCC –G	–0.57, 0.06		–4.87, 0.55
Og + 2WCC – Ka	0.21, 0.13	0.28, 0.18	
Ka + WR = OA	–0.55, 0.31		–0.81, 0.45

In contrast, the Oglala to Kadoka subbasin, situated between the others, consistently contributed flow to the river. The flow to this reach from three perennial creeks is the probable cause of this anomalous behavior. Very consistent monthly flow losses from the river at a sand-silt seepage velocity provide evidence of a predominantly perched river between Crawford and Oglala. Small, variable flow yields and losses suggest coupled hydrologic systems downstream, with the alluvial water table near (Oglala-Kadoka) or below (Kadoka-Oacoma) the river level during winter. These hypothetical hydrologic systems, based on the results of this study, are consistent with the field investigations of Rothrock (1942).

The mean, variance and extremes obtained with the PEM for dependent variables, such as water storage as ice and local subbasin inflow, allow definitive conclusions to be developed or identify the independent variables responsible for uncertainty in the results. Computation of air temperatures by subbasin instead of over the complete basin, and additional river width data to characterize a reach, would reduce the uncertainty in the present water balance. Improved estimates of the exchange between the river and the alluvial groundwater in a subbasin can be obtained by gaging all perennial creeks. A well-defined water balance that quantifies the winter river exchange with the alluvial aquifer in semi-arid regions, together with measurements of the relative river–alluvial aquifer levels throughout the year, can provide reliable estimates of the annual flow exchange.

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