River Salinity Variations in Response to Discharge: Examples from Western United States During the Early 1900s

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ABSTRACT: Major controls on river salinity (total dissolved solids) in the western United States are climate, geology, and human activity. Climate, in general, influences soil-river salinity via salt-balance variations. When climate becomes wetter, river discharge increases and soil-river salinity decreases; when climate becomes drier river discharge decreases and soil-river salinity increases. This study characterizes the river salinity response to discharge using statistical-dynamical methods. An exploratory analysis of river salinity, using early 1900s water quality surveys in the western United States, shows much river salinity variability is in response to storm and annual discharge. Presumably this is because river discharge is largely supported by surface flow.

Introduction

Climate is one of the most important causes of variations in river (or stream) chemistry in the western United States. Other important factors include geology and human activity. How is climate connected to river salinity? Perhaps the most direct link is through river-basin salt-balance variations (*eg,* Ghassemi, Jackeman, and Nix 1995). Assume, initially, that the long-term mean soil and river salinity in natural systems is in a dynamic balance mostly controlled by river basin long-term mean precipitation. Salt stored in soils is generally increased by atmospheric deposition, evapo-concentration, and salt production via biotic and abiotic soil-forming processes including rock and mineral weathering and is generally decreased by flushing with fresh runoff or precipitation. If the climate becomes dryer on time scales ranging from months (see below) to millennia, the slow buildup of soil salt during the extended dry periods may exceed the intermittent flushing of salts during the less frequent wet periods. The soil-river system becomes more saline. Alternatively, if the climate becomes wetter, the soil-river system becomes less saline. In this simple conceptual model, the soil-river basin salt balances are controlled by the rate of flushing and, thus, by decreases or increases in precipitation (and all of the other complicating factors are assumed to be secondary). Climate controls salt dynamics largely through the rates of salt removal. Our ability to recognize climatic driven salinity variations may ultimately be controlled by the intersection of time scales at which flushing transits the system, the time scale of climatic forcing, and finally, the time scales of observational time series. A more complex model than

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our simple conceptual model would include feedback. For instance, with increasing precipitation/runoff and decreasing soil salinity, the rates of salt supply from weathering may also increase but less so than the rates of removal. In this paper, the utility of statistical-dynamical models is explored and an example of the role of water storage and time scales in salinity response is presented.

A long-term goal of our research is to better understand natural soil-river salinization processes and thereby to gain a broader perspective for interpreting artificial salinization processes. The north-south climate gradients in the west provide a natural laboratory with generally low salinities in the humid northwest and high salinities across the arid southwest. Presumably, with sufficient time, if this north-south climate gradient were reversed, the salinity regimes would also qualitatively reverse (increasing salinity in the northwest and decreasing salinity in the southwest). However, the responses of the soil-river salinities to changes in precipitation would not be instantaneous.

Statistical-dynamical methods have been used to study stream chemistry in response to variations in discharge (Whitehead *et al* 1986) as well as suspended sediment concentrations (Lemke 1991). One of our interests is to gain insight into the response of soils to regional differences in precipitation by analyzing the salinity-discharge responses of a wide range of river basins. A long-term goal is to understand the nature of the correlation between stream and soil salinity. We emphasize that the results given here are preliminary not only because of the simplified conceptual model used, but also because they are generally limited to analysis of annual cycles; analysis of longer and/or additional time series ultimately will be essential for building confidence in the results.

In the present analysis, our depiction of river chemistry is restricted to consideration of the concentration of total dissolved solids or salinity. As a further simplification, we selected the earliest, and therefore the least human influenced, statewide water quality surveys by the U.S. Geological Survey for California, Oregon, and Washington from about 1906 to 1912. Also, effects of mineralogical differences across this western region were not considered in this study.

Data and Methods

Early this century, water quality surveys were made in numerous river basins (Figure 1) in California (Van Winkle and Eaton 1910), Oregon (Van Winkle 1916b) and Washington (Van Winkle 1916a). In general, 100mL samples were collected each day, and 10 days of samples were combined as a composite sample for analysis of total dissolved solids and major ion composition (Dole 1909). Similarly, discharge was reported as 10-day mean values by combining daily observations.

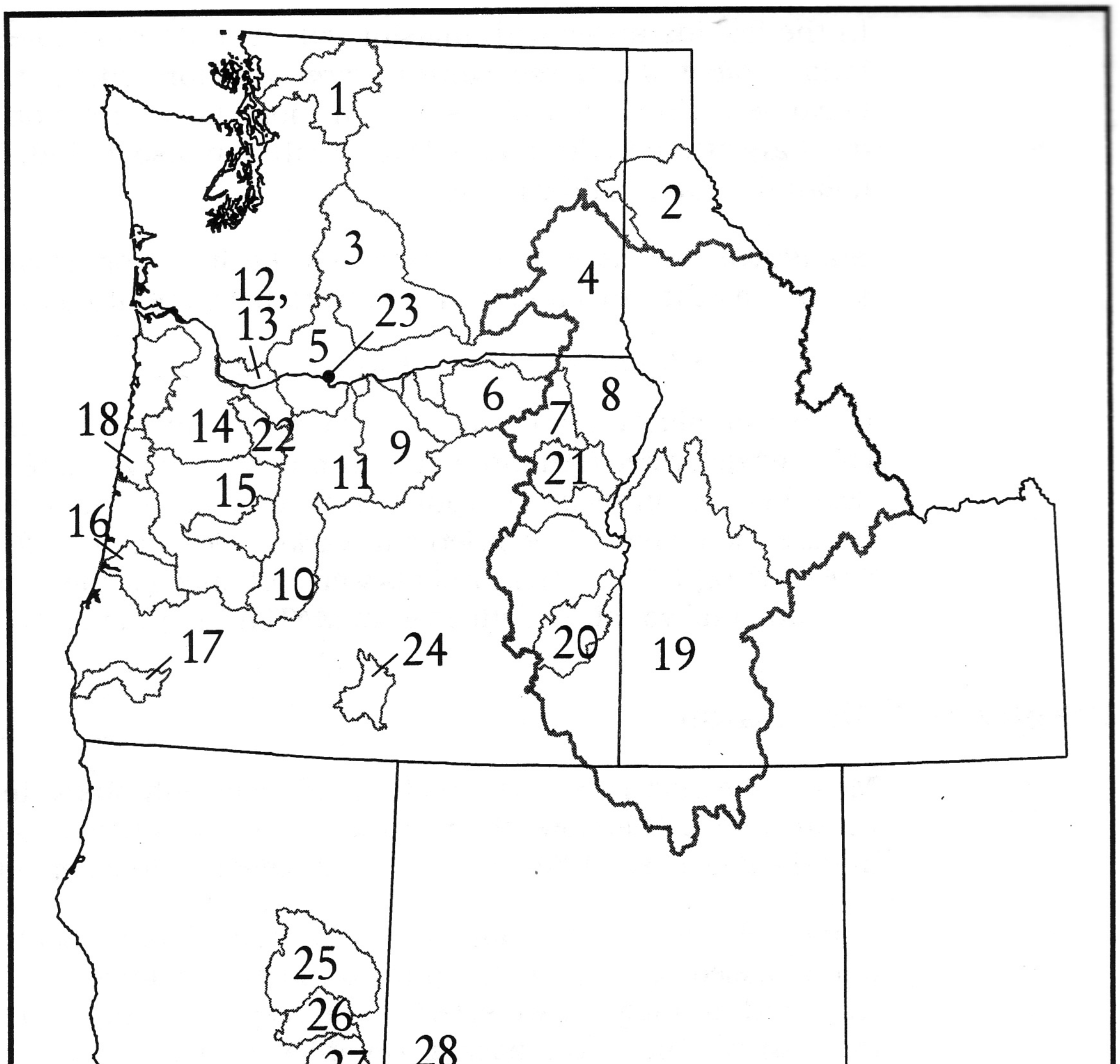

Figure 1 Study Region and Basin Locations
See Table 1 for names of river basins.

In the few instances with missing values, data were filled in by interpolation. Most discharge values were transformed as the log (base e) because salinity typically shows a more linear relationship to the log discharge than to discharge. Data treatment also included removal of the mean discharge and salinity.

Statistical models with discharge as the input or driving variable and salinity as the output or response variable were fitted according to:

$$y(t) = b_0u(t) + b_1u(t\text{-}1) + b_2u(t\text{-}2) + \ldots + e(t) \tag{1}$$

where y is salinity; u is discharge; t is present time; t-1 is a one-time internal (10-day) delay between input and output; t-2 is a two-time interval (20-day) delay between input and output with succeeding (t-x) at 10-day intervals; the b's are the response coefficients at various delays; and $e(t)$ is uncorrelated noise (Ljung 1987, 1991). Best fits were obtained for these models using the instrumental variable routines from MATLAB (Ljung 1991).

Results and Discussion

Most of the model results were identified using only one water year of 10-day mean values. Therefore, the parameter estimates (Table 1) are preliminary and the goodness-of-fits have not been tested on independent series.

The results are encouraging, with a mean correlation between observation and modeled values of 70%, and informative. To illustrate the usefulness of this kind of analysis, we selected the results of only two river basins for discussion. One of the basins, Umatilla River at Umatilla, shows a close discharge-salinity relationship; the other, Deschutes River at Moody, is an example of a weak relationship between discharge and salinity. Results for the Umatilla River are shown first, followed by those for the Deschutes River.

Umatilla River

Flow in the Umatilla River at Umatilla (Table 1) is characterized by low base flow and a high response to winter and spring precipitation and snowmelt (Figure 2). This response presumably indicates that water storage within the basin (and flow delays associated with such storage) is relatively small.

Under these presumed limited-storage conditions, the constant three coefficient model(b_0, b_1, b_2) captures about 90% of the variance (Table 1). However, a bias in the residuals (observed minus simulated values) indicates the model slightly overestimates concentrations (by about 10%) during annual rising discharge and underestimates (by about 10%) during falling discharge (Figure 3). This annual bias is probably due, in part, to saline irrigation water discharged into the river during summer (Van Winkle 1914a).

Table 1
DISCHARGE-SALINITY RESPONSE MODEL PARAMETERS AND STATISTICS

River and Location	Year	Observed Mean Discharge m^3s^{-1}	Observed Mean Salinity mgL^{-1}	Model Parameters b_0	b_1	b_2	b_3	Standard Deviation of Residual mgL^{-1}	Correlation Coefficient between Observed and Simulated	Range Between Base and Peak Discharge (% of Mean Discharge)
Washington, Oregon, and Adjacent States										
1 Skagit, Sedro Wolley	1910-11	449	48	-5.3	-28	0	0	11	0.12	145
3 Yakima, Prosser	1910-11	148	128	-20	-17	-7.6	0	24	0.90	330
4 Snake, Burbank	1910-11	1,870	131	-16	-3	-10	0	16	0.88	300
6 Umatilla, Yoakum	1911-12	29	100	-6.3	-2.7	-2.0	0	17	0.72	380
Umatilla, Umatilla	1911-12	25	188	-33	-10	-7.3	0	25	0.95	380
7 Grande Roude, Elgin	1911-12	54	98	-5.5	0	0	0	9.6	0.74	360
8 Wallowa, Joseph	1911-12	4.6	64	-0.2	0	0	0	3.0	0.50	310
9 John Day, McDonald	1911-12	165	142	-9.6	-4.9	-2.6	0	17	0.82	1,800
John Day, Dayville	1911-12	14	166	-25.7	0	0	0	20	0.87	360
10 Deschutes, Moody	1911-12	198	92	0	+7.8	0	0	6.5	0.39	140
Deschutes, Bend	1911-12	37	66	0	0	0	+7.0	4.9	0.24	59
11 Crooked, Prineville	1911-12	14	223	-2.7	-1.1	-0.6	0	37	0.89	350
12 Sandy, Brightwood	1911-12	24	49	-1.4	-1.9	-0.8	-1.8	3.3	0.95	220
13 Bull Run, Bull Run	1911-12	22	30	-3.2	-1.7	-0.5	-1.7	4.3	0.78	460
14 Willamette, Salem	1911-12	679	51	-3.1	-0.8	0	0	6.5	0.59	490
15 McKenzie, Springfield	1911-12	127	48	-7.2	-1.8	-0.5	-1.2	2.8	0.93	300
16 Umpqua, Elkton	1911-12	160	68	0	-2.5	0	0	5.1	0.48	380
17 Rogue, Tolo	1911-12	105	71	0	-5.1	2.3	0	5.2	0.66	350
18 Siletz, Siletz	1911-12	46	42	-0.04	0	0	0	5.1	0.41	600
19 Snake, Weiser	1911-12	621	220	-17	-13	-11.5	-16.7	21	0.84	300
20 Owyhee, Owyhee	1911-12	44	230	-6.7	-12.7	-7.9	-8.1	35	0.89	510
21 Powder, North Powder	1911-12	10	193	0	-7.9	0	0	36	0.50	450
22 Clackamas, Cazadero	1911-12	75	50	-8.1	-3.9	-1.6	0	3.7	0.91	280
23 Columbia, CascadeLocks	1910-11	6,700	91	0	0	-4.6	-6.5	9.5	0.62	220
Columbia, Cascade Locks	1911-12	5,610	98	-3.6	-16.8	0	0	8.1	0.87	290
24 Chewancan, Paisley	1911-12	4	85	-5.5	0	0	0	7.7	0.52	530
California										
25 Feather, Oroville	1906	275	99	-9.1	-3.6	6.6	0	11	0.83	360
26 Yuba, Smartville	1906	140	84	-3.3	-11	0	0	11	0.86	360
27 American, Fair Oaks	1906	298	80	-0.2	-4.4	-5.1	0	8.3	0.87	1,200
28 Mokelumne, Clements	1906	56	76	-7.4	-3.2	-4.1	0	15	0.76	370
29 Stanislaus, Knights Ferry	1906	302	83	-9.1	-15	0	0	27	0.48	210
30 Tuolumne	1906	136	73	0	0	-8.8	-5.3	16	0.80	350
31 Merced	1906	130	66	0	0	-12	-4.0	16	0.60	220
32 Owens, Charles Butte	1906	8.7	339	-45	-0.1	0	0	30	0.52	200
33 Kern, Bakersfield	1906	89	128	-15	-20	0	0	30	0.78	520
34 Arroyo Seco, Soledad	1906	12	281	-23	-9.5	-10	0	44	0.90	960
35 San Luis Rey, Pala	1906	143	320	-13	0	0	0	34	0.61	1,900

As expected for the simple conceptual model presented in the Introduction, total dissolved solids decrease with increasing discharge and, indeed, the response shows an approximately first-order exponential decrease in salinity with increase in discharge (Figure 4). An example of hysteresis in a salinity-discharge cycle is where the salinity response differs depending on whether discharge is increasing or decreasing (Figure 5). Much of this hysteresis was captured by the statistical-dynamical model (as indicated by similar departures of observed and simulated salinities from a smooth exponential decrease in concentration with increase in discharge). We believe an important mechanism for hysteresis is that the early seasonal flushing flows (rising discharge limb) carry more concentrated salts than the later flows (falling discharge limb).

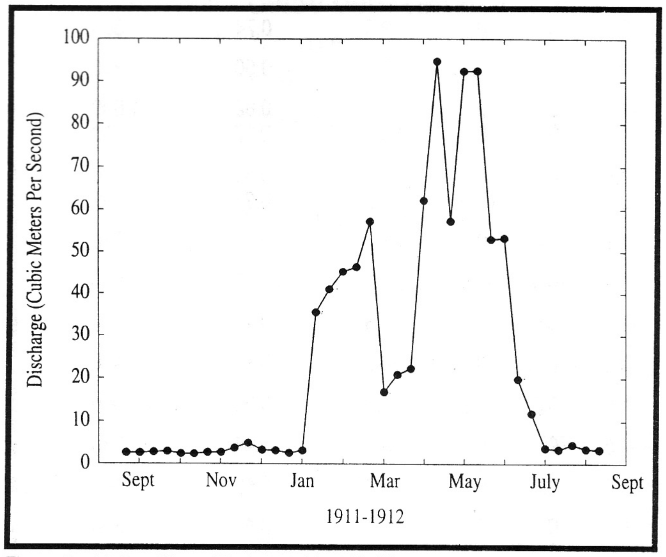

Figure 2 10-Day Mean Discharge Umatilla River, Umatilla, Oregon, 1911-1912

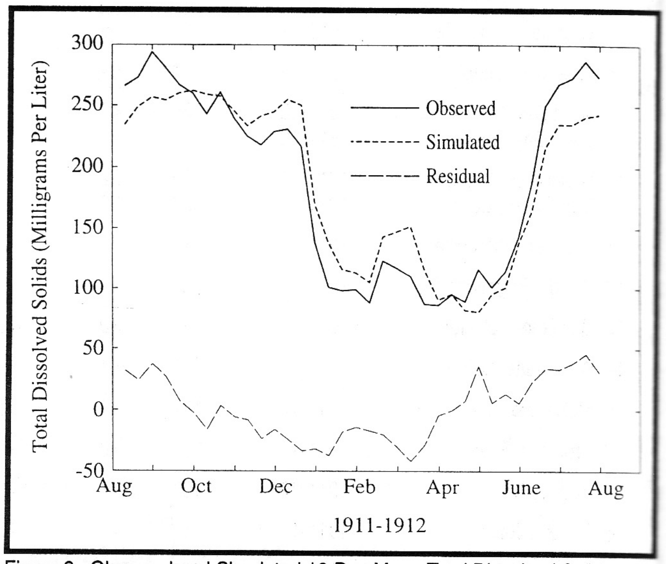

Figure 3 Observed and Simulated 10-Day Mean Total Dissolved Solids and the Residuals (Observed minus Simulated), Umatilla River at Umatilla, 1911-1912

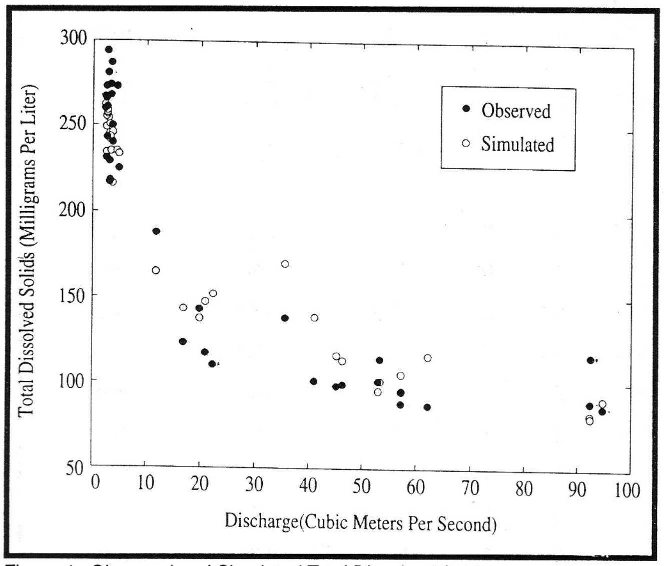

Figure 4 Observed and Simulated Total Dissolved Soids versus Discharge, Umatilla River at Umatilla, 1911-1912

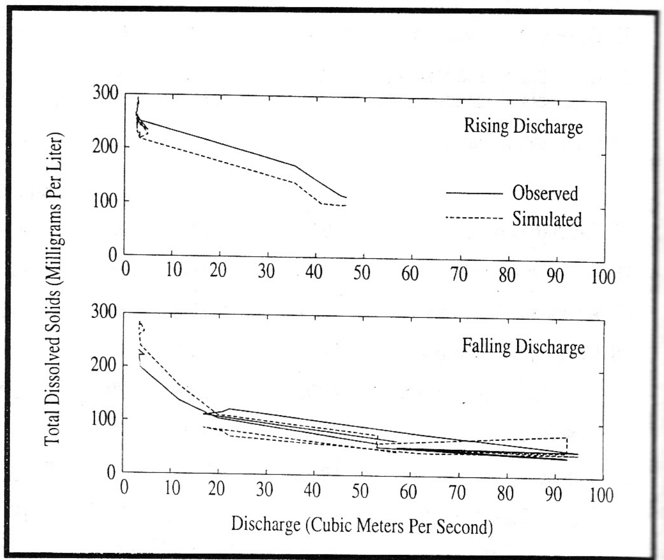

Figure 5 Upper Panel: Total Dissolved Solids in Relation to Rising Annual Discharge.
Lower Panel: Total Dissolved Solids in Relation to Falling Discharge, Umatilla River at Umatilla, 1911-1912

Deschutes River

In the Deschutes River system (Table 1), water storage in the soils and basalts underlying the basin dampen and delay the pulse of moisture input associated with winter precipitation (Shelton 1985). Thus, in contrast to the Umatilla River at Umatilla, the Deschutes River at Moody shows a high base flow relative to the winter-spring peak flows (Figure 6). For example, in the Umatilla River (1911-1912), base flow is about 20 times less than peak flows, whereas in the Deschutes River, base flow was about half the peak flows.

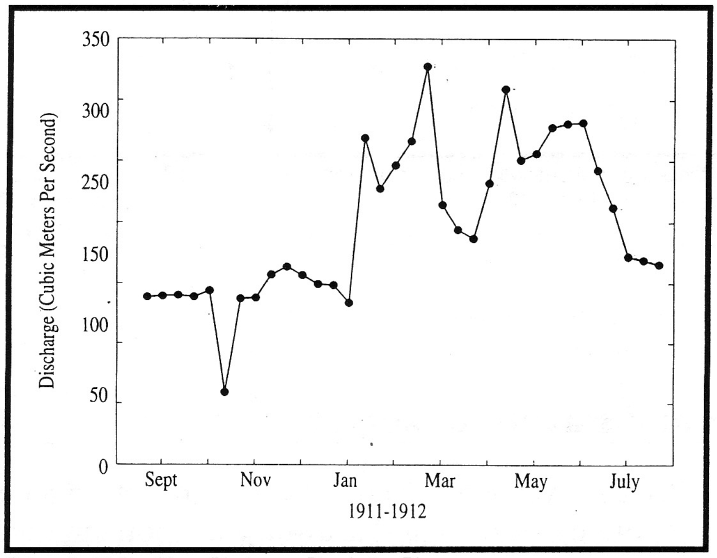

Figure 6 10-Day Mean Discharge, Deschutes River, Moody, Oregon, 1911-1912

The sensitivity of total dissolved solids concentrations to variations in discharge is low in the Deschutes River; in fact, total dissolved solids concentrations tend to increase with increasing discharge (Figure 7). We conclude that the poor correlation (correlation coefficient of 0.39) of observed and simulated total dissolved solids (Figure 8) is probably a result of dampening of salinity variations that “parallels” the dampening of discharge variations in this basin. Slow transit of subsurface storage within this basin, causes base flow and, indeed, the lion’s share of streamflow to undergo considerable damping, mixing and homogenization. Consequently, the discharge time scale of variation in this basin is probably at least annual and the salinity time scale is longer still (from the looks of Figure 8). Neither the model nor the 350-day time series is adequate to fit the slow variations in discharge and salinity that characterize this basin.

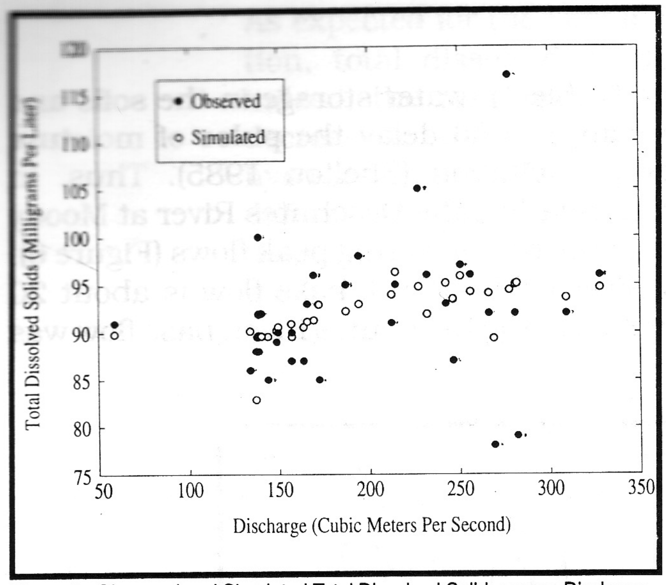

Figure 7 Observed and Simulated Total Dissolved Solids versus Discharge, Deschutes River at Moody, 1911-1912

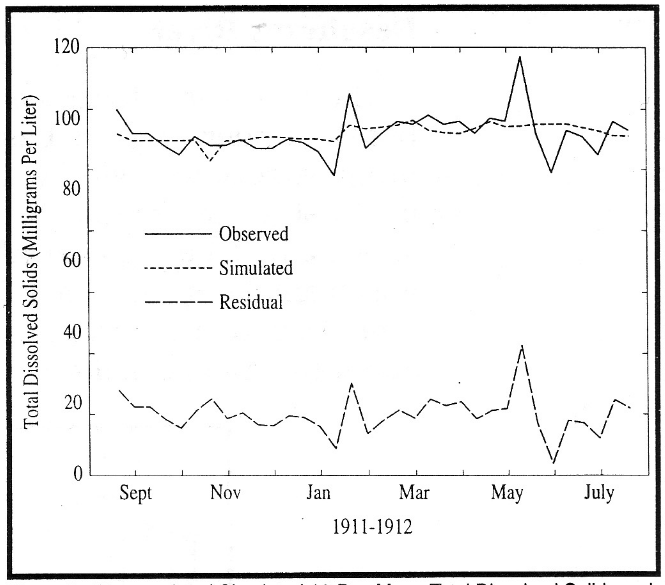

Figure 8 Observed and Simulated 10-Day Mean Total Dissolved Solids and the Residuals (Observed minus Simulated), Deschutes River at Moody, 1911-1912

Conclusion and Future Research

In the western United States, much river-salinity variability is in response to storm and annual variations in discharge, presumably because discharge is largely supported by surface flow. Basins with large storage and high base flow, such as the Deschutes River, are exceptions to this generalization. Also, however, not all basins with low base flow show a strong response of salinity to discharge.

Although considerable work needs to be done, statistical-dynamical modeling methods appear to offer an efficient way to characterize the complex variations in riverine chemistry, including river basin salt-balance phenomena. The preliminary success of the models is encouraging, but we need to compare the results with observations from other years. Further, the analysis of model parameters should be normalized to runoff (discharge per unit area). Also, a more complete evaluation of model parameters should include analysis of effects of the length of record and observational errors associated with inadequate sample collection, storage, and analysis as well as climatic (*eg*, snowpack) and human (*eg*, reservoirs) influences. Finally, discharge-response characteristics probably also vary for the different major ions. The Deschutes River example suggests that multiple-input models, with surface and subsurface discharge, may be needed.

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