

On the relationships among temporal patterns of evapo-transpiration, stream flow and riparian water levels in headwater catchments during baseflow

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Abstract Diel fluctuations in stream flow during baseflow discharge and coincident fluctuations in riparian water levels have been observed in many streams and are typically attributed to water losses from evapo-transpiration (ET). Riparian water level fluctuation may be due to 1) changes in stream stage, 2) changes in supply of subsurface hillslope water to the riparian zone, and/or 3) ET from riparian vegetation. We analyze data collected over the summer of 2004 from a previously-established network of wells located across a corridor of a headwater stream (~100 ha contributing area) in the central Cascade Mountains of Oregon, USA. Our findings indicate that groundwater-stream interactions link drawdown of riparian aquifers with streams in very complex spatial and temporal patterns even at the local scale (m to 10s of m). This is coincident with the effects of basin-wide limitation of hillslope water yield to riparian zones (and streams) that cause local changes in stream stage.

Keywords: groundwater-surface water interactions, riparian aquifers, diel fluctuations, stream flow

1. Introduction

A simple conceptual model of stream flow generation in headwater basins indicates that groundwater moves from hillslope and/or deeper aquifers to riparian zones and then from riparian zones to streams. Once in the stream, however, channel water is exchanged bi-directionally with riparian groundwater, flowing through meander bends or following preferential flow paths with higher hydraulic conductivity (e.g., abandoned stream channels). These interactions are not well understood when the hillslope-riparian-stream system is responding to temporal hydrologic changes, such as precipitation events, snowmelt, or even daily evapo-transpiration (ET). Part of the reason is that the spatial and temporal dynamics of the hydrologic processes are not completely understood, especially the effects of spatially heterogeneous patterns of hydraulic conductivity which are known to influence patterns of subsurface flow through the riparian zone (Cardenas et al., 2004).

Both riparian groundwater and stream flow respond temporally and spatially to local and basin-scale hydrological processes. At the basin scale, ET can provide a substantial loss of water, depending on climate and vegetation. The spatial pattern is relatively uniform (assuming relatively uniform vegetation coverage throughout the basin and no soil water limitation), and the temporal pattern cycles closely (particularly at the daily timescale) with vapour pressure deficit. The communication of these spatially-distributed losses of water down the catchment (i.e., from hillslopes to stream outlet) is complex, due to the fact that there are many temporal patterns being added or combined over a complex pattern of flowpaths (Wondzell et al., 2007). At the scale of 10s of meters within a riparian zone (i.e., plot scale), local ET is likely to generate some drawdown of local water tables (Meyboom, 1965). However, this demand is no different than the same ET demands occurring basin-wide, and it occurs in the spatial context of hydrologic responses to the basin-wide ET. That is, the down-catchment communication of these spatially distributed ET demands are ‘accumulating’ in riparian zones from all possible directions, and the local ET withdrawals are super-imposed upon these patterns. The temporal response, therefore, is reflected in the pattern of water table rise and fall within the riparian zone.

Given the complex nature of both the temporal transmission of ET signals within a headwater basin and the spatial pattern of hydraulic conductivity throughout hillslopes, riparian zones, and streambeds, riparian water table fluctuations and flow patterns are difficult to predict. Several studies have assessed riparian water level fluctuations (Bren 1997; Troxell 1936; Butler et al., 2007). However, the observed patterns of interactions between streams and riparian groundwater systems appear different in low gradient streams versus high gradient mountainous headwater streams. Here we analyze time series of riparian water table dynamics from 13 riparian wells (hillslope toe, mid-floodplain, stream edge, and sub-stream) in a mountainous headwater basin (101 ha). We compare the timing and amplitude of fluctuation in both the water table and stream stage in an attempt to identify critical controlling factors. That is, are diel riparian water table fluctuations caused by diel fluctuations in stream discharge; or do riparian water table fluctuations result from changes in the lateral inputs of hillslope water to the riparian zone; or are water table fluctuations only the result of local-scale ET demands from riparian vegetation?

2. Site Description and Methods

2.1. Site Description

We conducted this study in the lower portion of Watershed 1 (WS1) in the H. J. Andrews Experimental Forest in western Oregon (44° 10' N, 122° 15' W). WS1 is a small, steep-mountain stream draining a 101-ha catchment. The valley floor in the study reach of WS1 averages nearly 14 m wide and the longitudinal gradient averages 13%. The entire WS1 watershed was clear-cut over 4 years, from 1962 to 1966, and the wood was removed using an aerial cabling system so that no roads were constructed in the watershed. The watershed was burned in 1967 to remove understory vegetation and then aerially seeded with Douglas fir (*Pseudotsuga menziesii*). Regeneration was poor, so the watershed was hand-planted with 2-yr old Douglas fir seedlings in 1969. Today, Douglas fir dominates steep hillslopes and red alder (*Alnus rubra*) dominates the riparian corridor. Six well transects were established across the riparian corridor in the late 1990s (Fig. 1). The WS1 stream gauge is located approximately 150 m downstream of this well network.

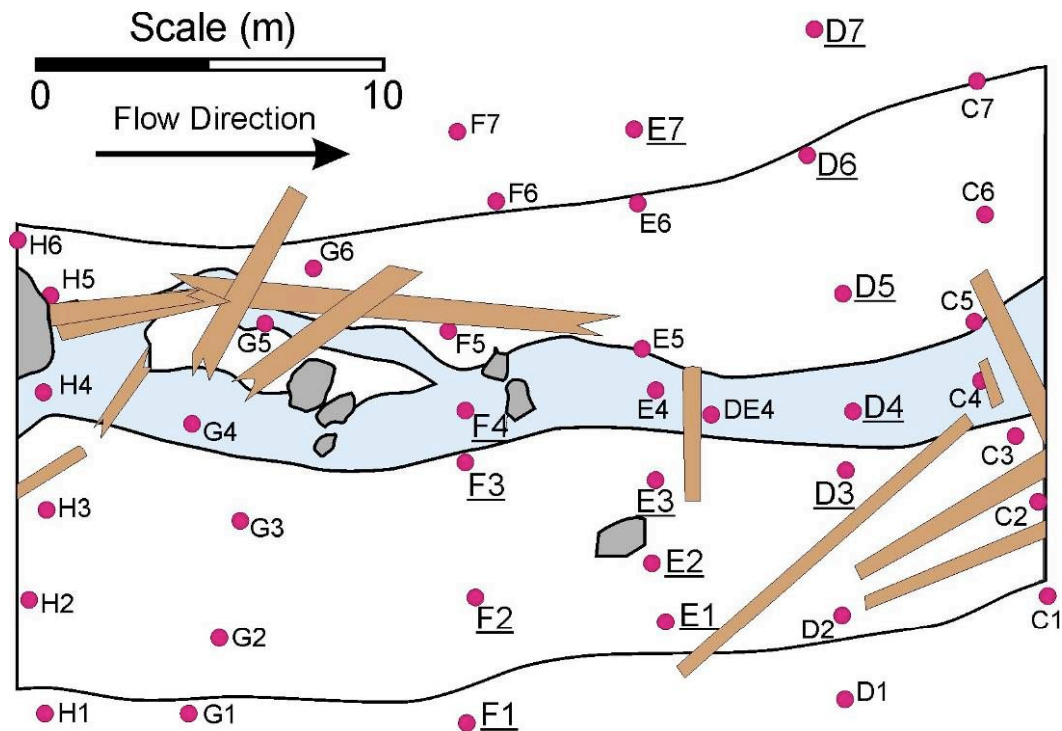


Fig. 1. Map of the WS1 riparian corridor well transects. Wells instrumented with capacitance rods are indicated in underlined bold text. Large boulders (grey polygons) and large logs (brown) are shown. Study site outline (solid black line) shows the limit of the valley floor. Wells outside of the study site

are located on the lower toe slope of adjoining hillslopes.

2.2. Methods

2.2.1. Water Levels

In late May 2004, 13 wells were instrumented with capacitance rods (Trutrack, Inc., New Zealand) that record water level (+/- 1mm) (Fig. 1). Water level data was collected every 10 min. for all wells except D3 which was collected hourly. The water level recorder in well E3 did not correctly acquire data until 3 July 2004, and the record from F1 was truncated on 22 July 2004 due to instrumentation error. Water level rods were retrieved 11 August 2004. We also analyzed stream stage and discharge data from the WS1 gauge (available at <http://www.fsl.orst.edu/lter/>).

We analyzed the water level and stream stage data for the period of 1 July to 10 August 2004, the period of summer-time baseflow recession. We do not have surveys for water levels, so water level data in each well is relative to an arbitrary datum and we cannot compare true elevations of water level among wells. Thus, we have focused our analysis on the magnitudes of the diel trends among wells and those of the stream. We computed anomalies (a_i) of diel fluctuations in water levels as:

$$h_{avg} - h_i = a_i \quad (1)$$

where h_{avg} is the average water level elevation computed daily (from midnight to midnight, mm) and h_i is the instantaneous water table elevation (mm). Analysis of the daily pattern of these anomalies allows for comparison of each well, relative to its location in the stream-riparian-hillslope system. We also computed daily water table elevation ranges (r) as

$$h_{max} - h_{min} = r \quad (2)$$

where, h_{max} and h_{min} are the maximum and minimum water table elevations for the day, respectively (computed from midnight to midnight, mm).

2.2.2. Stream Tracer Experiments

We analyzed the results of two stream tracer experiments conducted at low baseflow (1.04 L/s) in August 1997 and at high baseflow (4.47 L/s) in June 1998. Both tracer experiments were conducted with dissolved NaCl injected continuously for 102 hr at low baseflow and 69 hr at high baseflow. Electrical conductivity (EC) was measured in all wells and in the stream at each well transect to determine the percentage of stream water arriving at each well, based on an increase in EC above background (Gooseff and McGlynn, 2005). We used the EC measurements made at 69 hr from the end of the high-baseflow experiment to determine the percentage of stream water in each well. Measurements were made at 55 hr and 79 hr during the low-baseflow experiment, so we interpolated between the measurement times to estimate the percentage stream water in each well during low baseflow at 69 hours.

3. Results

Baseflow recession in WS1 in 2004 commenced in mid-June, reaching annual minimum stream flow sometime in late-August or early September (Fig. 2). The period was mostly rain-free. There were a few small storms that had little effect on stream discharge and water table heights. A larger storm occurred in late August (Fig. 2), during a period for which we did not analyze capacitance rods records. Discharge fluctuated daily, in response to basin-scale ET (Wondzell et al., 2007), but over the period of baseflow recession, the range in the daily fluctuations in discharge decreased dramatically (Fig. 2). During this time, we also observed daily fluctuations of the riparian water table (Fig. 3).

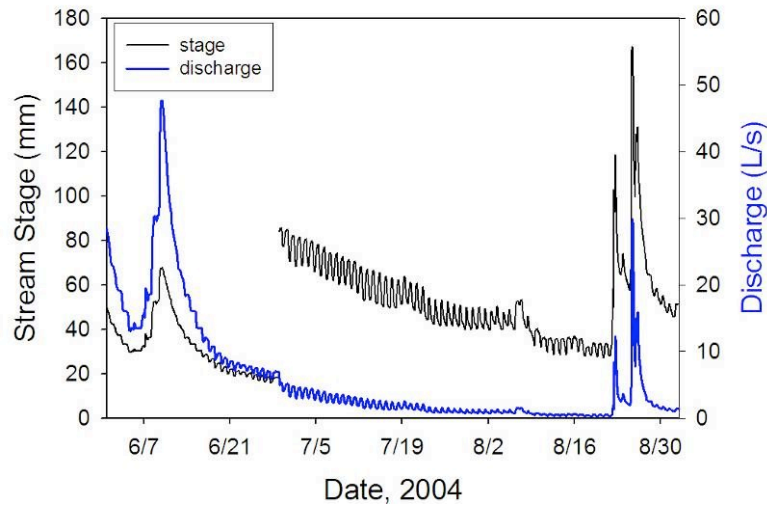


Fig. 2. Stage and hydrograph observed at the WS1 stream gauge. Discontinuity in late June stage record results from installation of a v-notch weir plate over the mouth of the trapezoidal flume to better gauge low flows.

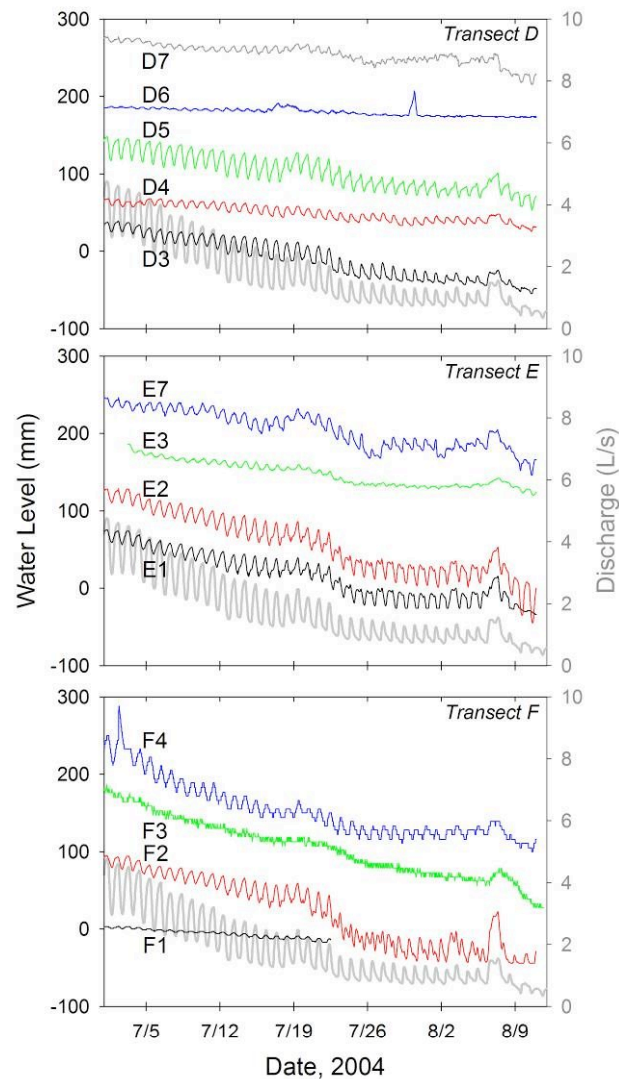


Fig. 3. Water table elevations from each well transect (arbitrary datum, all elevation records offset to provide separation for clarity) and concurrent discharge record (light grey, right-hand y-axis) in WS1 during seasonal baseflow recession.

All wells showed a long-term drawdown throughout the summer. Water table draw down, however, was relatively small, averaging approximately 10 cm. The water table drawdown over the summer was also spatially heterogeneous, ranging from a minimum of 3 cm (Well D6) to a maximum of approximately 16 cm (Well F4) (Fig. 3). Examination of early baseflow conditions (5-10 July) and later baseflow conditions (28 July - 2 August) show temporal shifts of diel patterns of some water table fluctuations, compared to stream stage variations (Fig. 4). In addition, the range of daily water table elevations, r , is spatially and temporally variable at these 13 sites (Fig. 5).

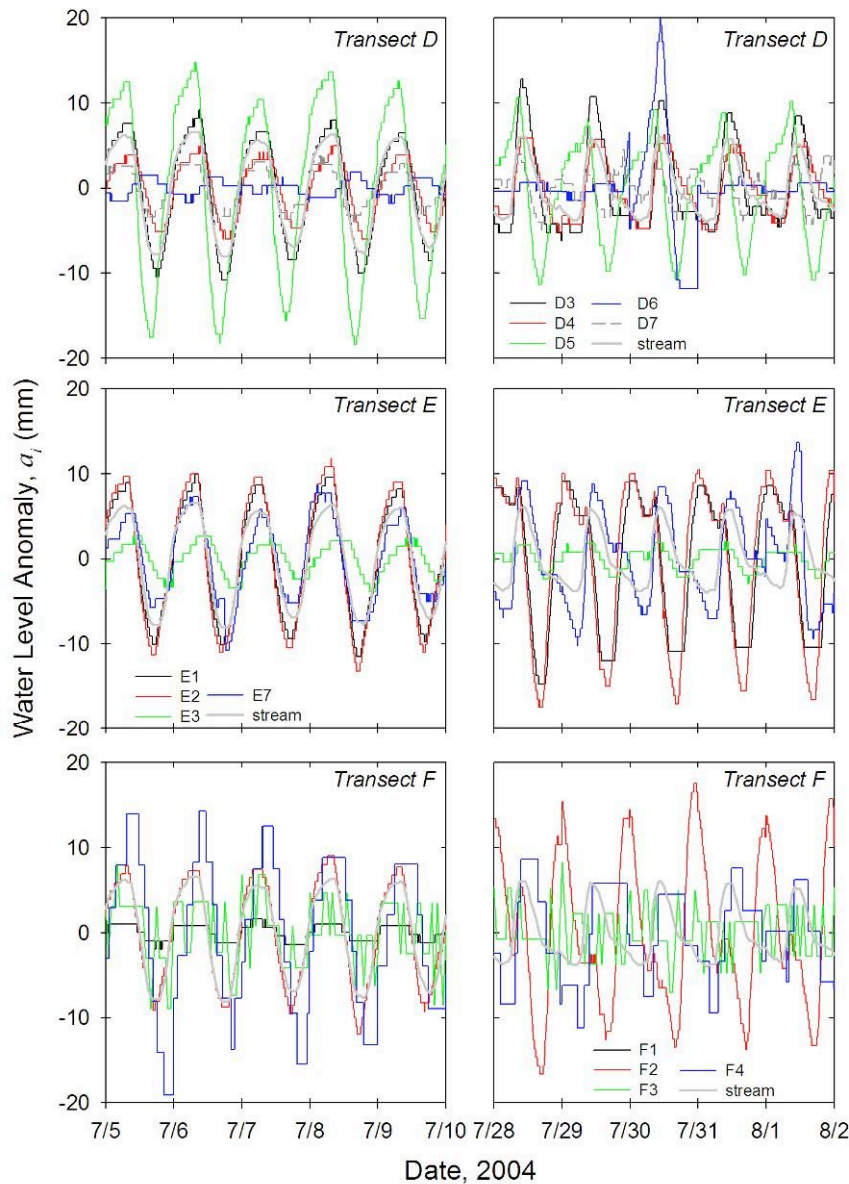


Fig. 4. Water table elevation and stream stage daily anomalies (arbitrary datum, all elevation records offset to provide separation for clarity) in WS1 during seasonal baseflow recession.

In transect D, daily water table fluctuation is approximately in phase for D3, D4, D5 and D7, while D6 is obviously out of phase (Fig. 4). During the later baseflow period, D3, D4, and the stream remain in phase, while peak water table heights occur earlier at D5 but much later at D7. The observed fluctuations in well D6 are quite noisy and irregular. Throughout baseflow recession, r is generally largest in the two stream-bank wells, D5 and D3, intermediate in the stream and the piezometers located in the center of the stream channel, D4, and smallest at the floodplain margin and toe-slope wells D6 and D7, respectively (Fig. 5).

In transect E, daily fluctuations of a in wells E1, E2, E7 and the stream are coincident in the early baseflow period (Fig. 4), whereas E3 lags the wells by several hours. In the latter baseflow period, a at well E7 and the stream remain tightly in-phase, whereas wells E1 and E2 become almost perfectly out of phase with stream stage. At this transect, the toe-slope well (E7), valley-margin well (E1), and the mid-floodplain well (E2) have the largest r , and the streambank well (E3) has the smallest r (Fig. 5). This spatial pattern is quite different than that observed at transect D, however the instrumented wells are located on opposite sides of the valley floor, suggesting that significant differences exist among locations on the valley floor and that these differences are not simply a function of the distance to the stream channel.

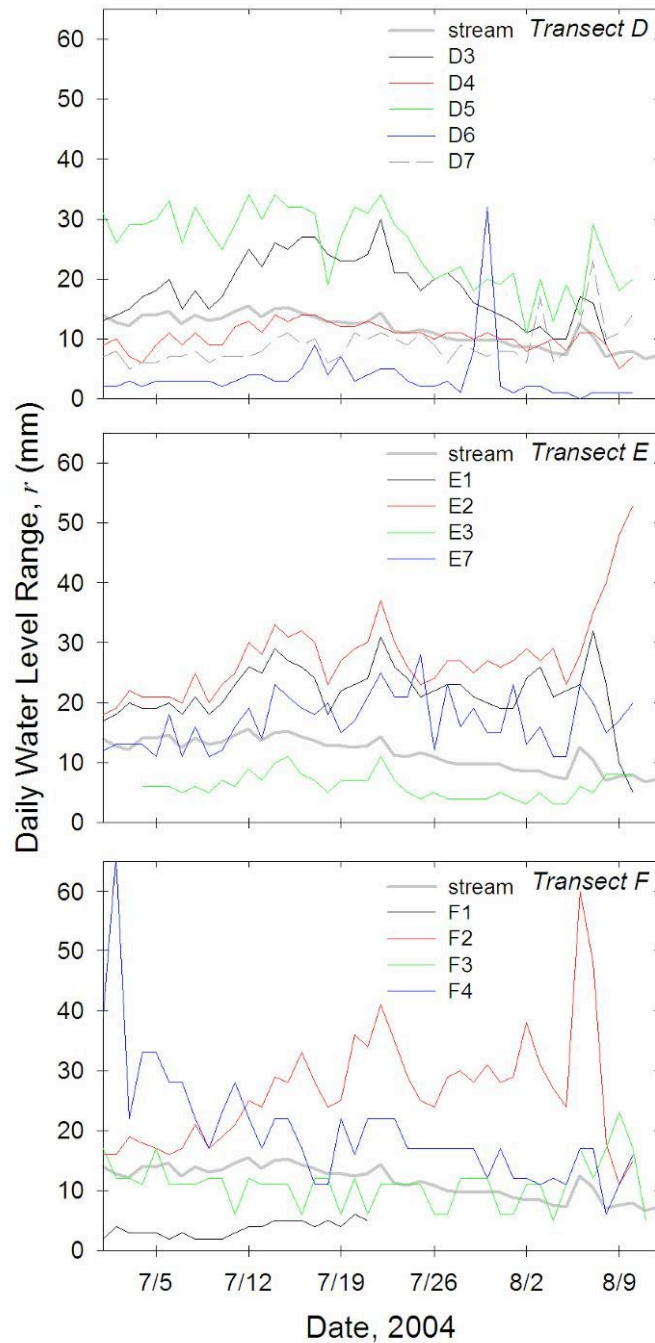


Fig. 5. Daily water table elevation and stream stage ranges in WS1 during seasonal baseflow recession. Stream stage range is repeated in all three panels for reference.

In transect F during the early baseflow period, daily fluctuations in wells are all in phase with stream stage, at least within a few hours (Fig. 4). In the later baseflow period, F2 a values increase in amplitude though the pattern is out of phase with the stream and the streambed piezometer, F4. Also, well F2 appears to develop a secondary peak in a , similar to that visible in well E2. The cycle of a at F3 becomes very noisy, lacking a discernable temporal pattern. Also, the temporal trend in r (Fig. 5) for well F2 is very similar to well E2.

The amount of tracer-labelled stream water reaching each well varies between high and low baseflow (Table 1). Our calculations are normalized to 69 hours, the duration of the shorter (high-baseflow) injection experiment. Wells D4, E3, F3, and F4 are consistently well connected to the stream, as indicated by a very high percentage of tracer-labelled water present in these wells after 69 hours. Other wells, however, show very different dynamics between tracer experiments. The greatest observed change is in D3, where a shift from 1% stream water to 91% stream water is observed from high to low baseflow experiments. This is surprising because well D3 is located on the stream bank, only 1 meter away from the stream. Less substantial increases were observed in the other wells, such as D6. These data show that the flow net connecting the stream with the riparian aquifer is spatially complex and temporally dynamic. Neither travel time nor the proportion of stream water found at any given well location are simply predicted by distance from the stream, suggesting that spatial heterogeneity in sediment, especially the saturated hydraulic conductivity, significantly influences flow paths. Further, some locations within the floodplain appear sensitive to changes in flow conditions between high-and low baseflow where as others change little with baseflow recession.

Table 1. Average water table daily anomalies (a_i) during baseflow recession in 2004 and summary results from stream tracer experiments detailed by Wondzell (2006) for the same wells. Distance from channel of 0 indicates that the wells are located in the channel (see Fig. 1 for site map). The high and low baseflow tracer injections were run for 69 and 102 hr, respectively.

Well	Distance from channel (m)	Average a_i (mm)	% stream water at 69 hr	
			high baseflow	low baseflow
D3	1.1	14.2	1	78
D4	0	8.8	92	78
D5	1.4	25.2	12	30
D6	5.8	3.7	44	32
D7	8.8	7.0	9	22
E1	5.7	16.4	7	5
E2	3.8	20.2	22	5
E3	1.1	6.1	100	87
E7	7.4	13.3	Dry	Dry
F1	8.8	9.1	11	Dry
F2	5.2	20.0	16	9
F3	1.4	11.5	92	77
F4	0	22.6	101	98

4. Discussion & Conclusions

The obvious question posed by analyzing these data is: Does the stream control the daily fluctuations in water table heights or does the water table control the daily fluctuations in stream discharge? The answer is, unfortunately, not obvious. It is reasonably clear that the fluctuations in stream discharge cannot directly control the daily fluctuations in water table heights simply because the stage level fluctuations of water in the open channel are very small. Consider the diel fluctuations in stream stage measured at the stream gauge with a trapezoidal flume. Daily fluctuations in discharge are typically

largest in early summer – for example, between 20 and 28 June, the daily range in stage measured from the trapezoidal flume averaged 2.5 mm (Fig. 2). Over this same period, the daily range in water table heights was much larger, at least in some wells. For example, the daily range in water table heights for the stream bank well, D5, averaged 27 mm. Installation of the V-notch weir plate at the end of June greatly exaggerates the fluctuations of stream stage – making the fluctuations in discharge easy to measure. But a V-notch weir does not simulate the conditions for open channel flow. In the open channel, discharge can be viewed as the product of width, depth, and flow velocity, as influenced by roughness and gravity. At very low flows, changes in discharge are strongly controlled by changes in flow velocity (Wondzell et al., 2007). The result is that the range in the daily fluctuations in stage height in the open channel is much smaller than in the adjacent water table. Clearly, the stream cannot “drive” water table fluctuations with such large amplitude.

Well then, do the daily fluctuations in the riparian water table control the daily fluctuations in stream discharge? Maybe. Several lines of evidence suggest that ET from riparian vegetation generates daily fluctuations in stream discharge (Dunford and Fletcher, 1949; Bren, 1997). However, the fluctuations in stream discharge are a macro-scale attribute of the whole watershed’s response to ET. Undoubtedly, ET from riparian vegetation must remove water from the riparian aquifer. Therefore, for ET to influence stream discharge, the effect must be transmitted through the riparian aquifer. Other researchers have shown highly synchronous fluctuations in water table heights (Troxell, 1936; Butler et al., 2007). However, careful examination of daily variations in water table heights at WS1 during the summer of 2004 failed to show the expected temporal synchrony that would be expected if drawdown of the riparian water table caused wide-scale de-watering of the stream. Subtle changes in stream stage are not likely to drive observed changes in water table (i.e., in a ‘de-watering’ perspective), though they may contribute to local heterogeneity in the water table fluctuations. Even though the WS1 site we studied tended to lack close synchrony, we still expect that the overall behavior of the riparian aquifers along the stream network must control either subsurface inputs to the channel or the amounts of water lost from the stream to the riparian aquifer, with sufficient synchrony to generate the observed fluctuations in stream discharge.

Site differences may explain the lack of synchrony at the WS1 study site. The valley floor has a 13% longitudinal gradient, and the valley is narrow, and valley floor sediments are relatively coarse, but only a few meters thick, lying on top of relatively impermeable bedrock. The dominant direction of subsurface flow, both at high- and low-baseflow discharge, is down valley (Kasahara and Wondzell, 2003; Wondzell, 2006). With relatively steep potential gradients and relatively high saturated hydraulic conductivities, water table drawdown caused by ET demand is transmitted down valley. Thus, the fluctuations in water table heights observed in any given well result from both the local ET demand and the influence of ET transmitted to that well from all up-flow-path locations. In contrast, if potential gradients are shallow and saturated hydraulic conductivities are low, then the water table height observed at a given location will dominantly reflect the local conditions (ET demand, the specific yield of the sediment, and relative importance of unsaturated zone and the depth to the water table). These local factors will not change the timing of ET demands, which will be synchronous over large areas. However, subsurface spatial heterogeneity of soil hydraulic properties is likely to contribute to the spatial and temporal heterogeneity of water table responses, even over small areas. Thus it seems reasonable that both Troxell (1936) and Butler et al. (2007) would have observed reasonably synchronous fluctuations in water table heights over large riparian areas along the low gradient streams that they studied.

Overall, it seems that local-scale responses to both seasonal drying and daily evapo-transpiration are manifest in spatially complex patterns in high gradient mountain streams. At the local-scale, these patterns appear too complex to support simple cause-effect relationships between diel patterns in ET, and fluctuations in both water table height and stream discharge. At the scale of the whole watershed, however, these relationships must give rise to the observed behavior of streams during baseflow recession.

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