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An analysis of alternative conceptual models relating hyporheic exchange flow to diel fluctuations in discharge during baseflow recession

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Abstract:

Diel fluctuations in stream flow during baseflow have been observed in many streams and are typically attributed to water losses from evapotranspiration (ET). However, there is no widely transferable conceptual model that explains how ET results in diel fluctuations in streamflow at the watershed outlet. For fluctuations to occur, two factors must be present: (1) some process must generate the fluctuations and transfer them to the stream channel, and (2) fluctuations must be accumulated and transported down the stream network in such a way that they arrive at a stream gauge as a coherent signal. We have previously shown how stream flow velocity affects the transport of diel fluctuations in discharge through a stream network. Here, we examined how riparian ET and hyporheic exchange flows generate diel fluctuations in discharge. We hypothesized that ET would cause a slight drawdown of riparian aquifers during the day, slightly increasing head gradients away from the stream and slightly reducing head gradients back to the stream. Thus, slightly more water would flow into the hyporheic zone than is returned to the stream, gradually reducing stream discharge. The process would be reversed at night. Using stream-tracer experiments and riparian water-level data, we tested two hypotheses related to this conceptual model—that the amplitude (H1) and time lag (H2) of diel aquifer drawdown would be constant over the summer. Neither hypothesis was supported by our data. We conclude that the processes that link watershed ET with streams include both local- and watershed-scale effects. Conceptual models attempting to explain diel fluctuations need to include the combined effects of ET on lateral inputs and hyporheic exchange flows, the redistribution of water within riparian aquifers, and the transport of ET signals from the whole stream network to the stream gauge. Copyright © 2009 John Wiley & Sons, Ltd.

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INTRODUCTION

Diel variations in both stream discharge and water table elevations in shallow aquifers have long been observed (Godwin, 1931; Hoyt, 1936 as cited by Bren, 1997; Wicht, 1941) and are usually attributed to evapotranspiration (ET) from riparian vegetation (Godwin, 1931; Troxell, 1936; Dunford and Fletcher, 1947; Meyboom, 1964; Hewlett, 1982, p. 97; Kobayashi *et al*., 1990). Several experimental studies support this conclusion. For example, complete destruction of a transpiring forest from an entire watershed eliminated diel fluctuations (O'Loughlin *et al*., 1982) as did removal of only the riparian forest (Dunford and Fletcher, 1947), whereas removal of forest vegetation from adjacent hillslopes, but with retention of a 30-m wide forested riparian buffer, actually increased the amplitude of diel fluctuations (Bren, 1997).

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Although the literature consistently identifies ET from riparian vegetation as the primary cause of diel fluctuations in discharge, there is no widely transferable conceptual model that explains how ET results in diel fluctuations in streamflow at the watershed outlet. For these fluctuations to occur, two factors must be present. First, some process must generate the fluctuations and transfer them to the stream channel. However, ET is widely distributed throughout the watershed. For diel fluctuations to be observable at the mouth of a watershed, a second process must accumulate the effects of ET and transport those effects down the stream network in such a way that they arrive at a stream gauge as a coherent signal. We have previously shown how stream flow velocity affects the transport of ET-induced fluctuations in discharge through the whole stream network (Wondzell *et al*., 2007). We treated ET as a distributed impulse function in a network-scale advection model and showed that when flow velocity was high, ET-generated signals from the stream network tend to arrive at the stream gauge 'in phase' so that constructive interference resulted in strong diel fluctuations at the mouth of the watershed. Conversely, when flow velocity was low, ET-generated signals tended to be out of phase so that destructive interference masks these signals at the gauging station. While flow velocity can influence the time lags and amplitudes of diel fluctuations observed at the mouth of a watershed, it does not answer the question of how diel fluctuations are generated.

Two somewhat different conceptual models have been proposed to explain the generation of ET-induced fluctuations in stream discharge: (1) ET from riparian trees captures some portions of the lateral inputs of water flowing from adjacent hillslopes, across the riparian zone and into the stream channel (Bren, 1997), or (2) ET from riparian trees captures water from shallow stream-side aquifers that are linked to stream discharge via hyporheic exchange flows (Bond *et al*., 2002). Bren (1997) used a groundwater flow model to examine hillslope–riparian–stream connectivity in gaining reaches where water tables sloped towards the stream. Simulations showed that ET demands of near-stream vegetation reduced water flux from the adjacent hillslopes during the day which would account for the observed diel fluctuations in discharge. Furthermore, hillslope vegetation removal increased lateral inputs, raising the water table throughout the riparian area, resulting in more ET and thereby explaining the increased amplitude of fluctuations observed after forest harvest. Bren (1997) worked in a strongly gaining stream reach. In many mountainous catchments, however, hillslope–riparian–stream connectivity is spatially limited to the largest hillslope hollows during baseflow (Jencso *et al*., 2009). Because the area of the riparian zone at the base of the hillslope hollows is small, whole watershed ET losses from lateral inputs in these locations are also likely to be small. Thus, ET losses from lateral inputs in gaining reaches may be insufficient to generate strong fluctuations in stream discharge at the mouth of the watershed.

The hyporheic zone is defined by the presence of stream water that has recently flowed out of the stream and into shallow riparian aquifers, and will return to the stream in a relatively short period of time. These flows of water from the stream, into the shallow riparian aquifer, and back into the stream are called hyporheic exchange flows. During baseflow, when lateral inputs from adjacent hillslopes are minimal, hyporheic exchange maintains shallow aquifers throughout the valley floors of mountainous stream networks (Wondzell, 2006). Because those aquifers are spatially extensive, and because ET losses of water along hyporheic exchange flow paths should reduce the flux of water returned to the stream, interactions between ET and hyporheic water have the potential to generate strong fluctuations in stream discharge.

Bond *et al*. (2002) developed a conceptual model linking ET losses of hyporheic water to diel fluctuations in discharge. They examined ET from trees growing in the riparian zone and lower hillslopes of a mountainous watershed. Their conceptual model related hypothesized changes in water table elevations, lateral groundwater inputs, and hyporheic exchange flows to changes in the time lags and amplitudes in diel fluctuations observed at the mouth of the watershed. We have worked at the same site as Bond *et al*., so we begin by examining their conceptual model, using data documenting patterns of hyporheic exchange flow at this study site (Kasahara and Wondzell, 2003; Wondzell, 2006). We show that critical features of Bond *et al*.'s conceptual model are not supported by the observed patterns of hyporheic exchange. Thus we develop and test an alternative conceptual model.

The objective of this paper is to examine linkages between riparian ET and hyporheic exchange flows. We propose that ET causes a slight drawdown of riparian aquifers over the course of the day, which will slightly increase head gradients driving flow away from the stream and slightly reduce head gradients driving return flows back to the stream. Thus on each exchange flow path, slightly more water would flow from the channel into the hyporheic zone, and slightly less water would be returned to the stream at the distal end of each of these flow paths, gradually reducing stream discharge. The process would be reversed at night. As the water table recovers, slightly less water would flow from the channel into the hyporheic zone along each exchange flow path leading to a gradual increase in stream discharge. If these slight changes in the water table generate diel fluctuations in discharge, we hypothesize:

H1: Given that diel patterns in ET demand are relatively constant over the summer, *the amplitude in daily variations of water table elevation should not change as stream discharge declines over the period of summer baseflow recession.*

H2: Given that the flow net linking the hyporheic zone and the stream changes little over the summer, *there should be no change in the time lags between the time of greatest ET demand and minimum water table elevations as stream discharge decreases over the period of summer baseflow recession*.

METHODS

The sapflow (Moore *et al*., 2004), stream discharge (Bond *et al*., 2002), and hyporheic (Kasahara and Wondzell, 2003; Wondzell, 2006) studies that provide the foundation for this article were all conducted in the lower portion of Watershed 1 (WS1; Figure 1) 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 100-ha catchment. The valley floor in the study reach of WS1 averages nearly 14 m wide and the longitudinal gradient averages 13%. Annual low flows occur at the end of the summer dry season with discharge less than 1 l/s. Baseflows during the wet winter season range from 10 to 20 l/s and the flood of record generated discharge of nearly 2·4 m^3/s.

We used topographic analysis of WS1, using a 10-m digital elevation model (DEM) to quantify the size and shape of the stream network and the location and size of upslope inputs to the stream network. Upslope

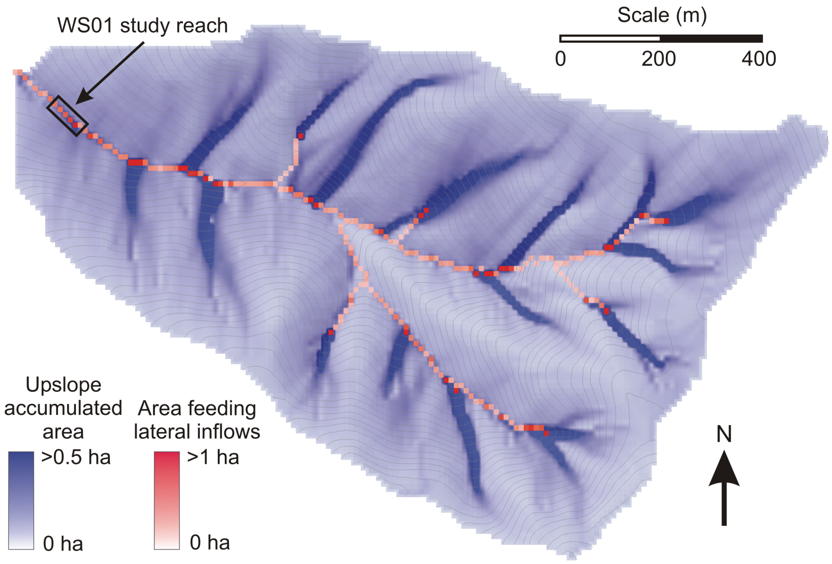

Figure 1. Topographic analysis of WS1 showing upslope accumulated area and the area contributing lateral inflows to each stream cell along the stream network. The rectangle near the mouth of the watershed shows the location of the WS01 study reach used by Bond *et al.* (2002) and used in this study

accumulated area contributing to each DEM cell was computed from DEMs with a multiple-flow-direction algorithm utilizing triangular facets (Seibert and McGlynn, 2007). Once the accumulated area exceeded a threshold of 2·5 ha, the area was routed downslope as stream cells and all subsequent cells along the downslope flow path were flagged as stream cells. Lateral inflows for each stream cell along the stream network were computed from the upslope accumulated area draining directly into each stream cell.

The hyporheic studies in WS1 used well networks to make direct observations of the water table and stream-tracer experiments to evaluate hyporheic exchange fluxes within the study reach. The well network (comprised both wells and piezometers) was installed along a 30-m reach of the WS1 stream in the summer of 1997. Wells were made from 1 to 2-m lengths of polyvinyl chloride pipe 'screened' over the bottom 50 cm whereas piezometers were only screened over the bottom 5 cm. Wells were located in closely spaced transects to provide high-spatial resolution of subsurface flows. Transects typically had one piezometer located in the centre of the stream channel and six wells which were located on stream banks, at mid-valley floor locations and at the toe slopes of adjoining hills, on both sides of the stream. Well heads, valley-floor crosssections along each transect, and the longitudinal profile of the stream channel were surveyed using a level and stadia rod. A subset of nine wells and two piezometers was instrumented with water-level recording TruTrack (the use of trade or firm names in this publication is for reader information and does not imply endorsement by the US Department of Agriculture of any product or service) capacitance rods in late summer, 2003 (Van Verseveld, *unpublished data*). Measurements were collected at 10-min intervals through early August, 2004, from which hourly measurements of water table elevations over the period of baseflow recession in the summer of 2004 were saved for further analysis. The stream channel within the study reach has been changing since the well network was first established and the channel surveyed and mapped. Consequently, the location of the wetted channel and the shape of water table equipotentials during the summer of 1997 were somewhat different from those of summer 2004.

Stream-tracer experiments were conducted at low- and high-baseflow discharge to compare the influence of varying discharge on the patterns of exchange flows coupling the stream to the riparian aquifer. The injection experiment for low-baseflow discharge ($Q = 1{\cdot}2$ l/s) was conducted from 4 to 8 August, 1997. The high-baseflow injection ($Q = 4{\cdot}7$ l/s) was conducted from 30 June to 3 July, 1998. The study reach for the tracer injection was 99·7 m, roughly centred around the well network. A concentrated solution of NaCl was injected at a constant rate until tracer concentrations reached a constant (or plateau) concentration at the bottom of the study reach. Tracer concentrations were measured at the bottom of the study reach using electrical conductivity (EC) as a surrogate because EC was highly correlated to Cl^- ($r^2 = 0{\cdot}995$, $n = 21$). Stream flow velocities were estimated from both the time at which the initial break through of tracer was observed and the time at which median tracer concentration was observed. True median flow velocity likely lies between these two estimates because hyporheic exchange flow was large enough to significantly retard advection of the tracer pulse through the stream reach, thereby distorting the estimates of median flow velocity. Water table elevations were measured from the well networks immediately before each stream-tracer experiment (see Wondzell, 2006 for additional details) and the median arrival times of stream-tracers reaching each well were used to calculate median travel times of hyporheic exchange flows between the stream and each well.

The examination of ET effects on stream discharge used hourly climate and stream gauge records that are publicly available through the H. J. Andrews Experimental Forest data bank (http://www.fsl.orst.edu/lter). Cross-correlation analysis was used to evaluate changes in time lags between the time of maximum ET and the time of both minimum stream discharge and maximum water table drawdown. We used vapour pressure deficit (VPD; defined as the saturation vapour pressure minus the actual vapour pressure) as a surrogate measure of the time of maximum ET demand rather than the direct measurement of sap flow used by Bond *et al.* (2002). We were concerned that using VPD would bias our results. Consequently, we repeated the Bond *et al.* (2002) analysis for summer 2000, but with VPD rather than sap flow, and obtained results identical to theirs. We examined the period of summer time baseflow recession during which discharge decreased from approximately 10 l/s in June to approximately 1 l/s in mid-August at the WS1 stream gauge. For each day over this period, hourly discharge was lagged behind hourly VPD by 0–23 h and the correlation coefficient (r) between VPD and discharge was calculated for each time lag ($n = 24$ h observations for each day). The time lag with the minimum correlation was saved for each day (giving the lag between the time of maximum VPD and minimum discharge). Seven-day averages and 95% confidence intervals were calculated for both mean daily discharge and the daily time lag. The diel fluctuation in discharge was also calculated by subtracting the minimum from the maximum discharge. (see Wondzell *et al.*, 2007 for additional details).

RESULTS AND DISCUSSION

Examining conceptual models linking riparian ET and hyporheic exchange

Bond *et al.* (2002) developed a conceptual model to explain two important observations in the time series of stream flow fluctuations observed in WS1: over the period of summer baseflow recession (1) the time lag between the time of maximum ET and minimum stream discharge increased, and (2) the amplitude in diel fluctuations decreased. Wondzell *et al.* (2007) showed that decreases in stream flow velocity over the period of baseflow recession can explain both increased time lags and decreased amplitude. Even so, ET-induced fluctuations in water table elevations in shallow streamside aquifers are likely to contribute to diel fluctuations in discharge. Thus, we first examine critical aspects of this conceptual model to see if it could also explain the observed changes in time lag and amplitude.

Change in type of dominant subsurface flow path

Bond *et al.* (2002) hypothesized that changes in subsurface flow paths could account for the increased time lag. In their conceptual model, 'active exchange along short-hyporheic flow paths' (flow path r2, Figure 2A) and 'shallow hillslope flow paths' (flow path h1, Figure 3A) should decrease with baseflow recession over the course of summer because of the decreases in stream discharge, water table drawdown, and moisture depletion from hillslope soils. They further speculated that the reduction in the relative dominance of near-stream, fast exchange flow paths and shallow hillslope inputs in late summer results in ET demands being transmitted to the stream via slower flow paths (flow paths r3 and h2, Figures 2A and 3A, respectively) which accounts for the increase in time lag. Data from tracer experiments showed that there is no change in the relative dominance of short- versus long-time scale subsurface flow paths over the period of baseflow recession (Figure 2B; Wondzell, 2006). Also, lateral inflows were small and changed little from high- to low-baseflow discharge. Finally, measurements from the well networks showed little change in water table elevation (Figure 3B; Wondzell, 2006). These observations show that subsurface flow paths changed little at this study site over the period of baseflow recession and thus hypothesized changes of Bond *et al.* in flow path are unlikely to account for the observed increase in the time lag between maximum ET and minimum daily stream flow.

Change in water table elevation

Bond *et al.* (2002) observed that amplitude of diel fluctuations in discharge decreased over the summer. They speculated that changes in near-stream water table elevation over the period of baseflow recession associated with the progression of the summer drought (Figure 3A) caused the water table to fall below the depth of most roots by late summer. They further speculated that the decrease in connectivity from the falling water table would reduce total ET losses from the riparian aquifer which would explain the decreased amplitude in diel fluctuations. Our data from 1997 to 1998, however, do not show substantial drops in water table elevations over the summer (Figure 3B). Measurements from the well network ($n = 30$) in the summer of 1997 showed that the depth to the water table averaged 70 cm and that the water table elevations averaged for all wells in the network changed by less than 1·0 cm between high- and low-baseflow injection experiments (Wondzell, 2006). In the summer of 2004, automated measurements from the nine wells showed water table drawdown averaged 7·8 cm (max = 13·3 cm; min = 1·2 cm) over the period of baseflow recession during which discharge decreased from 10·0 l/s to less than 1·0 l/s. It seems unlikely that changes of a few centimeters in water table elevations would dramatically change coupling between trees and their sources of soil moisture. These observations show that water table elevations changed only slightly at this study site over the period of baseflow recession and thus hypothesized changes of Bond *et al.* in water table elevation are unlikely to account for the observed decrease in the amplitude of diel fluctuations in discharge.

An alternative explanation. Wondzell *et al.* (2007) showed that transport of ET-induced fluctuations in discharge through the whole stream network was greatly

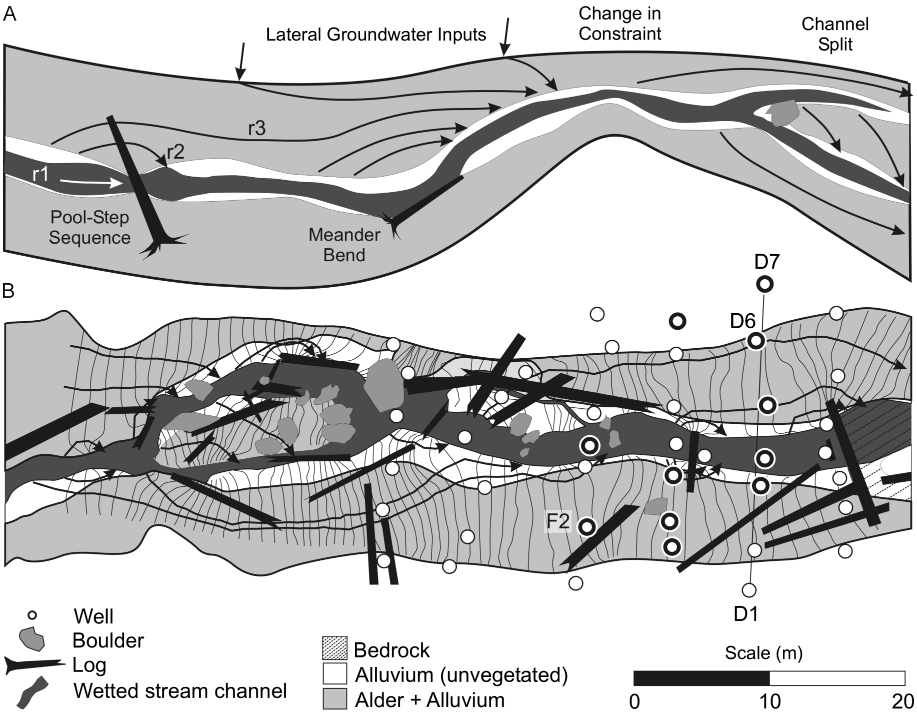

Figure 2. (A) Idealized flow paths through a shallow, surficial aquifer underlying the riparian zone of a small, steep, forested stream styled after Figure 1c of Bond *et al.* (2002). The figure illustrates typical channel-morphologic features driving exchange flow between the stream and the shallow aquifer. Labels denote very short-residence time surface stream channel flow paths (r1), short-residence time near-stream flow paths (r2) and longer-residence time flow paths (r3). (B) Map of a 60-m reach of WS1 in 1997 showing location of boulders and logs controlling channel morphology, locations of wells (on floodplain) and piezometers (in wetted stream channel), and water table equipotentials (contour interval = 0·1 m) from Kasahara and Wondzell (2003). Wells drawn in bold outline indicate those instrumented with capacitance rods to record water-level fluctuations over the summer of 2004. Some subsurface flow paths are drawn to illustrate the flow net linking the stream and the shallow riparian aquifer. Though more complex than the idealized flow paths shown in (A), the general patterns resulting from various channel-morphologic features are clearly present. The location of the cross-sectional profile of the 'D' well transect is shown in Figure 3B (from D1 to D7), as are the locations of wells D6 and F2 that are shown in Figure 4B

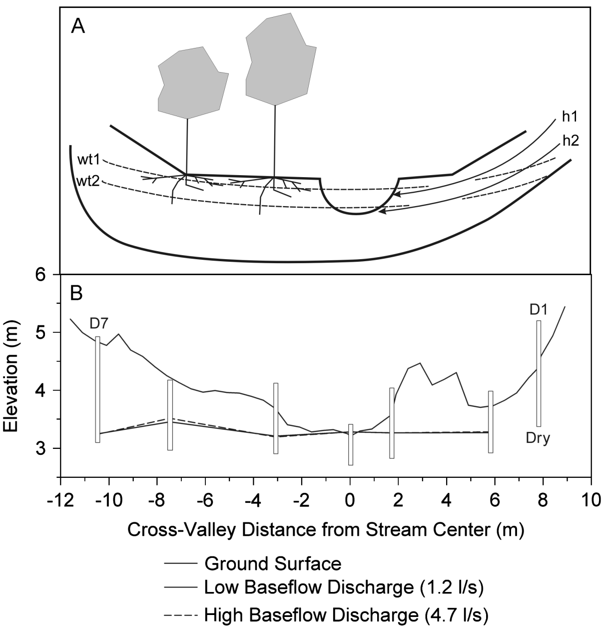

Figure 3. (A) Conceptual model of lateral groundwater inflows, water table elevations, and rooting depths at a time of high stage (inflow h1; water table elevation wt1) and low stage (inflow h2; water table elevation wt2) redrawn from Figure 1d of Bond *et al*. (2002). *Reproduced with permission from Bond et al., Hydrological Processes. Copyright 2002, John Wiley and Sons Limited.* (B) Cross section of a well transect 'D' (shown in Figure 2B) showing the cross-sectional profile of the floodplain surface, the depth of penetration of wells, and the cross-sectional profile of the water table at high- and low-baseflow discharge

affected by stream flow velocity. The effects of ET on discharge are accumulated over the 1·4-km long stream network, from the most distal channel head to the stream gauge near the mouth of the watershed (Figure 1). Estimated whole network travel times for WS1 ranged from as little as 7 h in early summer to more than 24 h by late summer (Wondzell *et al*., 2007). With high-flow velocity at high-baseflow discharge in early summer, ET-generated signals tended to be in phase so that constructive interference resulted in strong diel fluctuations at the gauging station at the mouth of the watershed. Also, as flow velocity was high, time lags were short. By late summer, when both discharge and flow velocity were low, time lags were very long and ET-generated signals tended to be out of phase so that destructive interference led to substantial decrease in amplitude. Thus decreasing flow velocity over the period of baseflow recession can account for both the decreased amplitude of the diel fluctuations in stream discharge and the increased time lag between the time of maximum ET demand and the time of daily minimum stream discharge.

We have closely examined hyporheic exchange in a well field established on a 30-m long reach of the stream channel. Is the studied stream reach representative of the WS1 stream network? Jencso *et al*. (2009) examined hillslope–riparian–stream connectivity. They showed that large hillslope hollows with convergent flow were necessary to accumulate sufficient hillslope area to maintain a saturated water table from the lower hillslope, across the riparian zone to the stream during low-flow periods when soils within the watershed were dry. Over most of their stream network, hillslopes were disconnected from the riparian zone throughout the low-flow period. Patterns of hillslope accumulated area in the WS1 catchment (Figure 1) are similar to those observed by Jencso *et al.* (2009). Our topographic analysis of WS1 shows that the area of riparian forest located at the base of hillslope hollows and thus able to intercept and transpire lateral inputs of hillslope water is relatively small. The remainder of the stream network only receives lateral drainage from relatively small hillslope areas (Figure 1), similar to the WS1 study reach, suggesting that large portions of the stream network will have minimal lateral inputs of hillslope water during baseflow. In these areas, shallow riparian water tables must be maintained by hyporheic exchange flows throughout the summer. We conclude that the WS1 study site is likely to be a characteristic of much of the WS1 stream network. That is, hyporheic exchange flows over the period of baseflow recession throughout much of the stream network are likely to be similar to those observed in our 30-m long study reach.

In the WS1 study site, critical elements of Bond *et al.*'s (2002) conceptual model that hypothesizes that hyporheic exchange flows link ET with discharge fluctuations do not occur. We find neither evidence for changes in the subsurface flow net, nor evidence that substantial water table drawdown occurs. Cross-valley flow paths may be present where headwater hollows drain to the stream and in these locations, changes in water table elevations may substantially effect the generation of diel fluctuations as described by Bren (1997) and as one component of Bond *et al.*'s conceptual model. We have not examined subsurface flow paths in such locations so we do not know if the seasonal drawdown of water table elevations is larger in these locations. Nor do we know if time lags of hillslope inputs change with baseflow recession in these locations.

Testing specific hypotheses related to riparian ET and hyporheic exchange

Signal generation. In our conceptual model, we expect that the drawdown of the riparian aquifer (Figure 4B) due to transpiration during the day should lead to losses of water from the stream, contributing to reductions in stream flow and in minimum daily discharge in late afternoon, or later in the night. Using Bond *et al.*'s (2002) average estimate of early summer transpiration rates (0·3 cm/day) and a specific yield of 25% for the valley-floor sediments (mid-range value for sands and gravels; Domenico and Schwartz, 1990, p. 118, Table IV.2), we estimate that transpiration could drawdown water table elevations by as much as 1·2 cm over the course of the day. This estimate agrees closely with the 1–2 cm average daily drawdown in water table elevations observed in the nine automated monitored wells during the summer of 2004. Relating the observed amplitude in aquifer

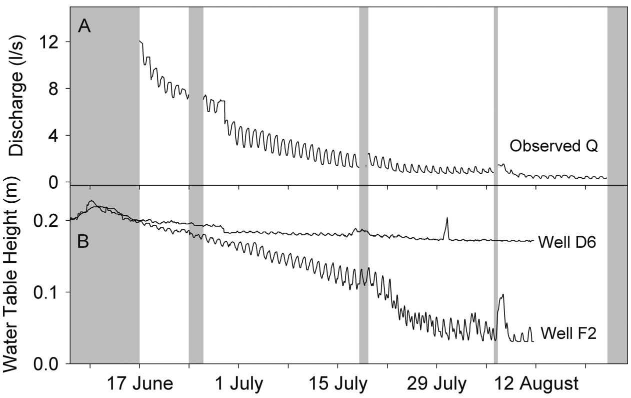

Figure 4. Temporal trends over the period of baseflow recession, summer 2004, in Watershed 1. (A) Summer baseflow recession in discharge during 2004. (B) Water table elevations in wells showing the smallest (Well D6) and largest (Well F2) changes in the diel fluctuations in water table elevation with baseflow recession. Water table elevations are reported with an arbitrary datum with the elevation in each well at midnight, 7 June 2004, adjusted to 0·2 m to facilitate comparisons between the wells. Shaded zones denote time periods excluded from the analysis due to very high or very low discharge, rainstorms, or missing data

drawdown with diel fluctuations in stream discharge was problematic, however.

If water table drawdown is driven by ET from riparian vegetation, we expected that neither the amplitude of the diel fluctuation in water table elevation (H1) nor the time of maximum drawdown (H2) would change over the summer. Little change should be expected because diel patterns of ET demands from alders (*Alnus rubra* Bong.) changed relatively little during the period of baseflow recession (Moore *et al.*, 2004) and because the forest growing on the valley floor was dominated by alder. However, measurements of diel fluctuations in water table elevations made during summer 2004 were counter to our expectations. A few wells did show relatively constant diel fluctuations in water table elevations (Figure 5A). For most wells, however, the amplitude in water table fluctuations increased with baseflow recession (Figure 5A). Similarly, most wells showed substantial changes in time lags. Of the nine automated wells and two automated piezometers, six closely tracked changes in time lags of the surface stream while the other five showed patterns substantially different to that of the stream (Figure 5B). The minimum water table elevations in some wells did precede the time of minimum stream discharge as would be expected if water table drawdown was the physical mechanism leading to daily reductions in stream discharge. However, the minimum water table elevations in other wells occurred later than the period of minimum discharge (Figure 5B). There was no apparent pattern related to the distance of the well from the stream, or with the median travel time of water from the stream to each well. Furthermore, we cannot ascribe the failure of the data to fit the expected pattern to changes in the subsurface flow net because previous data from stream-tracer studies showed that subsurface flow paths did not change between periods of high- and low-baseflow discharge.

The fundamental question still remains—where in the watershed, and through what physical mechanisms, are

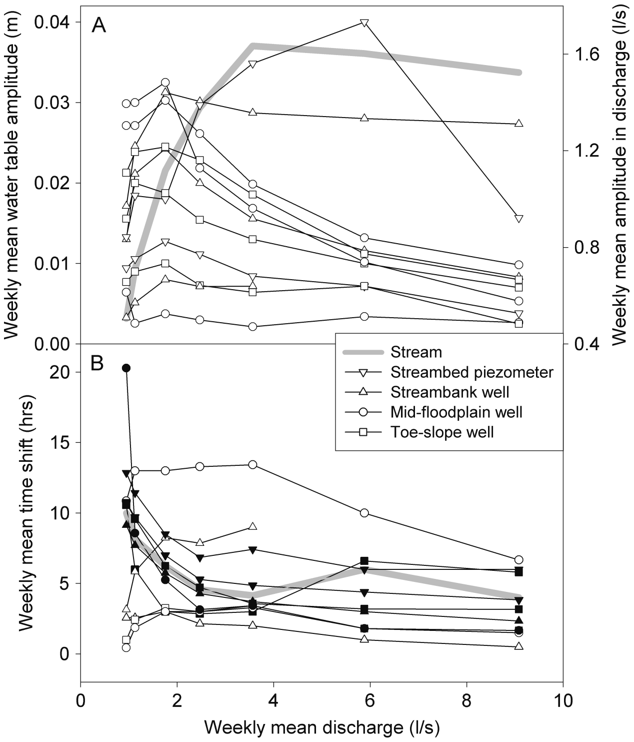

Figure 5. (A) Changes in the weekly average amplitude of diel fluctuations in stream discharge (heavy grey line) and water table elevation (symbols) over the period of baseflow recession in the summer of 2004. (B) Changes in the weekly average lag between the time of maximum VPD versus the time of either minimum stream discharge (heavy grey line) or minimum water table elevation observed in each well (symbols). Closed symbols denote wells where time-lag patterns were similar to the stream; open symbols denote wells where time-lag patterns were different from the stream. Wells in which the minimum water table elevation occurred before the minimum stream discharge plot below the heavy grey line. Conversely, wells in which the minimum water table elevation occurred after the minimum stream discharge plot above the heavy grey line

the diel fluctuations generated? Many authors have proposed that diel fluctuations in streamflow are caused by ET from riparian vegetation because of the obvious and ready access of riparian trees to water in the near-stream zone (Troxell, 1936; Dunford and Fletcher, 1947; Meyboom, 1964; Kobayashi *et al.*, 1990; Bren, 1997). As the area of riparian forest in WS1 is large, and because hyporheic exchange flows maintain shallow water tables beneath these forests throughout the summer, it seems that riparian forests could easily generate the observed diel fluctuations in discharge. However, our analysis of water table response in a riparian zone dominated by hyporheic exchange flows did not provide evidence for a simple causal linkage between riparian ET and diel fluctuations in discharge as we originally hypothesized. We believe that our hypotheses were overly simplistic because they failed to account for the fact that stream discharge is an integrated response to inflows and losses to the stream occurring over the entire stream network.

On the need for an integrated conceptual model. Despite our earlier work on flow velocity and diel fluctuations (Wondzell *et al.*, 2007), we failed to account for the influence of stream network transport times on the relative timing of fluctuations in discharge versus the water table at our site near the mouth of the watershed. As stated above, the effects of ET on discharge are accumulated over the 1·4-km long stream network. Estimated whole network travel times for WS1 ranged from as little as 7 h in early summer to more than 24 h by late summer (Wondzell *et al.*, 2007). Our well network was within 100 m, or so, of the stream gauging station in WS1. Thus, it should not be surprising that minimum daily drawdown of the water table in wells near the mouth of the watershed occurred before the effect of ET-induced reductions in stream discharge generated along the whole stream network was observed at the stream gauge.

The fact that there was substantial variability among wells in the exact timing of minimum water table elevations, relative to stream discharge, suggests that individual wells respond to multiple forcing factors, including (1) direct ET losses of water from the aquifer immediately surrounding the well; (2) transport of ET-induced effects through the aquifer via ground water flow paths; and (3) interactions between the stream and the adjacent aquifer. If diel fluctuations in water table elevation around each well resulted *only* from riparian ET withdrawals through the overlying soil, it might be reasonable to expect that the time of minimum water table elevations should not change substantially over the summer. Conversely, if the water table was controlled *only* by the water elevation in the adjacent stream, then the water table would be expected to closely follow changes in stream stage at both daily and seasonal time scales. Clearly, many wells were strongly influenced by stream discharge, as the seasonal pattern of increasing time lags observed in the wells paralleled changes observed in stream discharge (Figure 5B, filled symbols). However, time lags for these wells varied greatly, with minimum water table elevations occurring as much as 3 h before, and more than 3 h after, minimum daily discharge. Minimum water table elevations in all wells occurred after the estimated time of maximum ET demand, but the time lag varied widely, from as little as half an hour for some wells to as much as 10 h in other wells. The variability in the response of individual wells suggests that the length and travel time along ground water flow paths between the stream and the aquifer, and within the aquifer, may be important in determining the water table response to ET-caused withdrawals.

The patterns of increasing amplitudes in diel fluctuations of water table elevation over the course of the summer (Figure 5A) suggest to us that the properties of the overlying unsaturated zone that links the aquifer with the trees may also influence water table response to ET. If water table drawdown causes the diel fluctuations to occur in sediment with dramatically different specific yield, then constant ET demand could result in different amplitudes of water table fluctuations over the period of baseflow recession. However, we did not observe layering of distinct sediment textures within the valley floor nor was there substantial drawdown of the water table, so changes in specific yield are an unlikely explanation for the increased amplitude of fluctuations. Perhaps ET demands in early summer are preferentially met with water from the unsaturated zone above the water table, which averages 70-cm thick within our well network. Alternatively, the ET demands from the trees might be transmitted more directly to the aquifer in late summer, as the overlying soil dries out, thereby creating steeper moisture gradients between the aquifer and the tree roots. However, if the water table was not closely connected with roots located near or in the capillary fringe above the aquifer, soil drying should decrease hydraulic conductivity which could potentially reduce ET-caused withdrawals from the aquifer.

The network-scale transport model proposed by Wondzell *et al.* (2007) can incorporate multiple factors that may contribute to diel fluctuations in stream discharge. Here, we have focused on the relation between riparian ET and stream discharge, because these are linked through hyporheic exchange flows. However, other factors may also contribute to diel fluctuations in discharge. For example, data presented by Dunford and Fletcher (1947) suggested that weak diel fluctuations persisted, even after complete removal of riparian vegetation. Similarly, WS3, in the H. J. Andrews Experimental Forest showed diel fluctuations in discharge, even though debris flows in 1996 removed all riparian vegetation from a large portion of the stream network. These observations would be consistent with the explanation suggested by Bren (1997)—that ET in zones of hillslope discharge are the critical locations for generating diel fluctuations. Alternatively, diel changes in water viscosity caused by changes in water temperature may also explain diel fluctuations in discharge (Constantz and Zewelleger, 1995; Constantz, 1998) as could direct evaporation from the stream channel. All of the possible mechanisms proposed to explain diel fluctuations are physically related and dependant upon the daily solar cycle—and these effects would be additive, leading to maximum water losses during the afternoon.

CONCLUSIONS

The hydrologic network that connects the soil–plant–atmosphere continuum to the stream, and the linkages and feedbacks among the hillslopes, stream and the shallow riparian aquifer is complex. But the combined behaviour of these component parts of the watershed, and the linkages among them, must explain observed watershed behaviour. Simple conceptual models attempting to link local-scale drawdown of riparian aquifers such as the hypotheses initially proposed here, or the conceptual model proposed by Bond *et al.* (2002), do not appear sufficient to explain how diel fluctuations in discharge can be generated from small watersheds. At a minimum, a more comprehensive conceptual model attempting to explain diel fluctuations in discharge during summer baseflow needs to include the combined effects of ET-induced withdrawal of water from both lateral inputs in gaining reaches as well as shallow riparian aquifers maintained by hyporheic exchange flows, the redistribution of water occurring at local scales within riparian aquifers, transport of ET signals generated from riparian zones along the whole stream network, and possibly the effect of hillslope processes.

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