An approach to understanding hydrologic connectivity on the hillslope and the implications for nutrient transport

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[1] Hydrologic processes control much of the export of organic matter and nutrients from the land surface. It is the variability of these hydrologic processes that produces variable patterns of nutrient transport in both space and time. In this paper, we explore how hydrologic "connectivity" potentially affects nutrient transport. Hydrologic connectivity is defined as the condition by which disparate regions on the hillslope are linked via subsurface water flow. We present simulations that suggest that for much of the year, water draining through a catchment is spatially isolated. Only rarely, during storm and snowmelt events when antecedent soil moisture is high, do our simulations suggest that mid-slope saturation (or near saturation) occurs and that a catchment connects from ridge to valley. Observations during snowmelt at a small headwater catchment in Idaho are consistent with these model simulations. During early season discharge episodes, in which the mid-slope soil column is not saturated, the electrical conductivity in the stream remains low, reflecting a restricted, local (lower slope) source of stream water and the continued isolation of upper and mid-slope soil water and nutrients from the stream system. Increased streamflow and higher stream water electrical conductivity, presumably reflecting the release of water from the upper reaches of the catchment, are simultaneously observed when the mid-slope becomes sufficiently wet. This study provides preliminary evidence that the seasonal timing of hydrologic connectivity may affect a range of ecological processes, including downslope nutrient transport, C/N cycling, and biological productivity along the toposequence. A better elucidation of hydrologic connectivity will be necessary for understanding local processes as well as material export from land to water at regional and global scales. *INDEX TERMS:* 1615 Global Change: Biogeochemical processes (4805); 1860 Hydrology: Runoff and streamflow; 1866 Hydrology: Soil moisture; 1899 Hydrology: General or miscellaneous; *KEYWORDS:* carbon and nitrogen transport, hydrologic connectivity, TOPMODEL

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1. Introduction

[2] The study of integrated biogeochemical cycles is challenging because of the tremendous number of processes and linkages governing the dynamics of individual elements. A first approach to understanding these cycles was to examine land-water linkages, and specifically the movement of biochemical elements from terrestrial ecosystems to streams, lakes, and eventually the oceans [*Likens and Bormann*, 1974; *Likens et al.*, 1981; *Meybeck*, 1982]. It soon became evident that spatial heterogeneity creates distinct control points for the movement and processing of materials; this led to a second approach, the study of specific landscape components such as ecotones or riparian zones [*Peterjohn and Correll*, 1984; *Lowrance et al.*, 1984; *Van der Peijl and Verhoeven*, 2000]. A third approach has been to experimentally manipulate whole systems by, for example, clear-cutting or nutrient fertilization. Taken together, these approaches have led to the general understanding that material export is primarily a function of water flow within specific basins and of landscape heterogeneity, such as geologic setting, land use, and vegetation type

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among different basins [e.g., *Beaulac and Reckhow*, 1982; *Lewis and Saunders*, 1989]. Secondary controls are mainly functions of disturbance (e.g., clear-cutting), and the biological transformations [*Likens et al.*, 1981; *Aber et al.*, 1998] and anthropogenic inputs [*Howarth et al.*, 1996, 2000; *Vitousek et al.*, 1997] of organic matter and nutrients.

[3] Despite this extensive study of the amounts, forms, and timing of terrestrial exports, we still understand little about the complex dynamics and rates of production of dissolved materials on land, and their delivery to surface waters. For example, a recent review of 42 studies of DOC and DON concentrations and fluxes in temperate forests found that lab and field studies differed greatly in their results, and that site-specific controls such as temperature, C:N ratios, or litterfall were rarely evident at regional scales [*Michalzik et al.*, 2001]. The factor that emerged as most influential across varying spatial and temporal scales was the amount of precipitation. Water flow through the soil column has been found to influence nutrient fluxes in peatlands [*Judd and Kling*, 2002; *Tipping et al.*, 1999], as well as inorganic nutrient fluxes in many systems [e.g., *Moldan and Wright*, 1998; *McHale et al.*, 2000]. While water has both a flushing effect and an impact on soil moisture, which in turn controls oxygen status and microbial degradation, the flow paths and spatial distribution of water in the catchment can also be important in determining material export [*Kling*, 1995; *Kashulina et al.*, 1998; *Frank et al.*, 2000]. This effect of spatial heterogeneity extends to vegetation patterns and to the influence of different plant types on soil water nutrient concentrations and fluxes [*Shaver et al.*, 1991; *Giblin et al.*, 1991; *Scott et al.*, 2001; *Judd and Kling*, 2002], although the underlying controls may be related more to plant production and exudates than to decomposition. Clearly, however, the processes of organic matter production and decomposition, interacting with hydrological flows, are fundamental to our understanding of material production and export in all environments.

[4] Here we explore how hydrologic connectivity potentially affects catchment system flushing rates, as well as the spatial and temporal isolation of nutrients on the hillslope. Hydrologic connectivity is the condition by which disparate regions on the hillslope are linked via subsurface water flow, and is a key determinant of the movement of nutrients down a hillslope [*Hornberger et al.*, 1994; *Creed and Band*, 1998]. *Grayson et al.* [1997] proposed that two distinct hydrologic states predominate: (1) a dry state, in which hillslope regions are hydrologically unconnected, the dominant flow is vertical, and spatial patterns of soil moisture are unorganized, and (2) a wet state, in which hillslope regions are connected via lateral subsurface flow and soil moisture patterns are organized spatially.

[5] Our understanding of the nature, extent, and impacts of variable hydrologic connectivity on element export is incomplete, as is our understanding of how hydrological and biogeochemical processes interact on the landscape. Integrating state-of-the-art, process-based models of hydrology and biogeochemistry can make progress in these areas. The work presented here is a step toward developing such a linked, process-based model. It is notable that there are very few (if any) applications of mechanistic models that can predict the outflow of materials as a function of processes and interactions within the catchment [*Cosby et al.*, 1985; *Hornberger et al.*, 1994; *Fisher et al.*, 2000; *Lee et al.*, 2000; *Band et al.*, 2001].

[6] The focus of this paper is to (1) more generally define hydrologic connectivity on the landscape, (2) present numeric simulations that demonstrate the spatial evolution of connectivity on the landscape, and (3) examine observational evidence for these ideas and show the impact that hydrologic connectivity between ridges and valleys has on stream chemistry. For this study we use (1) model simulations at two catchments, one located in New York and the other on the North Slope of Alaska, and (2) in situ stream hydrographic and chemographic data as observed during the snowmelt season at a small headwater basin located near Boise, Idaho.

2. Site Descriptions

2.1. Cascade Brook Catchment, New York

[7] The Black Rock forest is a 1500-hectare preserve located in the Hudson Highlands region of New York. Elevations in the forest range from 110 to 450 m above mean sea level, with seasonal temperatures ranging from $-2.7°C$ to $23.4°C$. Soils in the lowland areas contain more organic matter than upslope soils, but bulk densities are not significantly different. Exposed bedrock is common throughout the preserve, and consequently the area was not extensively farmed during the period of European settlement. Lumber extraction ceased in 1927, and the forest has been managed as a preserve without significant disturbance since that time. The system is typical of the *Quercus*-dominated, secondary growth forests that have characterized the northeast United States over the past century.

[8] Within the Black Rock forest is the 135-hectare Cascade Brook catchment. A single stream, Cascade Brook, drains this catchment. Average hourly discharge from Cascade Brook is monitored continuously using a V notch weir installed in 1998. Hourly measurements of precipitation, air temperature, dewpoint temperature, incoming shortwave radiation, and wind speed are also taken. Hourly thermal radiation for the site is calculated following the methodology of *Anderson and Baker* [1967]. Three years of meteorological and hydrologic data, collected between 1998 and 2000, were used to drive the model.

2.2. Imnavait Creek Catchment, Alaska

[9] The Kuparuk River (the basin is $\sim$9,000 km^2) has its headwaters in the Brooks Range and drains through the foothills and coastal plain of northern Alaska to the Arctic Ocean. The Imnavait Creek subcatchment (2.2 km^2) lies within the headwaters of the Kuparuk Basin at 68°37′N, 149°19′W. Imnavait Creek is located in rolling piedmont hills, where the predominant soils include 15–20 cm of porous organic peat underlain by silt and glacial till [*Hinzman and Kane*, 1991; *Hinzman et al.*, 1991]. The climate, hydrology, and energy balance of Imnavait Creek have been studied continuously since the mid-1980s [*Cooper et al.*, 1996; *Hinzman and Kane*, 1991; *Hinzman et al.*, 1991, 1996,

1998; *Kane et al.*, 1991; *Michaelson et al.*, 1998; *McNamara et al.*, 1998]. Multiyear records of soil thaw, precipitation, and streamflow exist [*Everett et al.*, 1996; *Nelson et al.*, 1998], and a weather station with winter snowfall records has operated year-round since 1975. Imnavait Creek has also been the focus of several hydrological and linked hydrological-biological modeling efforts [e.g., *McNamara et al.*, 1997; *Stieglitz et al.*, 2000]. The topographic sequence of land cover ranges from wet sedge in the riparian zones to tussock tundra along the mid-slopes to dry heath near the ridge tops. Water tracks, regions of enhanced soil moisture that run down the hillslope at a spacing of $\sim$10–20 m, channel flow down the slope [*McNamara et al.*, 1999]. From 1985 through 1993 the mean annual precipitation, maximum snow water equivalent, and air temperature averaged 34 cm, 12 cm, and -7.4°C, respectively; 66% of the annual precipitation fell during the short summer season. Snowmelt accounted for 47% of annual discharge while runoff and evapotranspiration were 46 and 54% of the water budget, respectively. Snowfall is possible throughout the year; however, the snow season begins in earnest in September when soil freezing begins. Spring melt is in mid-May to early June. The maximum thaw depth in summer ranges from 25 to 100 cm depending on vegetation, aspect, slope, and soils. Hydrologic activity is limited to the near surface because of the shallow maximum thaw depth and the fact that saturated hydrologic conductivities fall off extremely rapidly below the porous organic layer.

2.3. Upper Dry Creek Catchment, Idaho

[10] The Dry Creek catchment is a 27 km^2 semi-arid mountain watershed located in the Boise National Forest just north of Boise, Idaho. The study sub-area (hereafter referred to as the Upper Dry Creek catchment) for this project is a small, ephemeral headwater stream oriented east-west with a drainage area of approximately 0.02 km^2. The average elevation in this small basin is 1640 m with approximately 50 m of relief on north and south facing aspects with slope angles approaching 20 degrees. Mean annual precipitation is 550 mm/yr, approximately half of which falls as snow in the cold season. The primary vegetation is sagebrush, forbs, and grasses. The soil is classified as a sandy loam with average depths of less than 1 m overlying Idaho Batholith granite, a Cretaceous age granitic intrusion ranging in age from 75 to 85 million years. Discharge and stream electrical conductivity were continuously recorded. Time domain reflectometers (measuring volumetric soil moisture) were installed horizontally into the undisturbed side of a mid-slope soil pit at depths of 5, 15, 30, 45, and 65 cm. The depth-to-bedrock in this mid-slope region is 75 cm.

[11] The study basin is characterized by winterlong snow cover and frequent snowmelt events in late winter and early spring, and may experience rain on snow events throughout the winter months. The soil remains unfrozen throughout the winter months due to snow cover. Streamflow typically begins in early winter and continues through the snowmelt period in March or April. While there are occasional summer and fall thunderstorms, typically no streamflow occurs after snowmelt. Hydrochemical studies indicate that there are no interactions with deep groundwater [*Yenko*, 2003] (see section 4).

[12] Important characteristics of this basin with respect to our study of connectivity are (1) the hydrology of the basin is driven by intermittent snowmelt; (2) the electrical conductivity (a proxy for the total dissolved ion content) of the snow is low; and (3) conductivity increases with meltwater-soil contact time.

3. Modeling Framework and Simulations: The Concept of "Connectivity"

[13] TOPMODEL [*Beven and Kirkby*, 1979; *Beven*, 1986a, 1986b; *Beven et al.*, 1994] is a conceptual rainfall-runoff model in which the impact of topography is accounted for using quasi-statistical techniques and which permits spatial-temporal prediction of the development and dissipation of surface waters. TOPMODEL formulations define areas of hydrological similarity, that is, points within a catchment that respond to meteorological forcing in similar fashion, saturate to the same extent, produce the same levels of discharge, etc. These points of hydrological similarity are identified by an index that is derived from analysis of catchment topography. This topographic index is often of the form $\ln(a/\tan\beta)$, where $\tan\beta$ is the local slope angle at a patch on the land surface, and a is the amount of upslope area draining through that patch. Lowland areas tend toward higher topographic index values, due to a combination of either low slope angle or large upslope area. Upland areas tend conversely toward lower topographic index values. Points within a catchment with the same topographic index value are assumed to respond identically to atmospheric forcing. Thus within a TOPMODEL framework, the topographic index provides the fundamental unit of hydrological response.

[14] The TOPMODEL-based hydrology model employed for this study has been previously described [*Stieglitz et al.*, 1997]. This quasi-statistical model is at once computationally efficient while still permitting dynamic representations of physical processes within the system. TOPMODEL assumptions (see *Ambroise et al.* [1996] and *Beven* [1997] for details) permit reconstruction of the spatial variability of catchment response to meteorological forcing solely from modeling of the response of the mean state (Appendix A). Base flow and surface runoff can be partitioned, and the spatial distribution of water table depth (WTD, also a proxy for soil moisture) across the landscape can be mapped at the resolution of a given digital elevation model data (DEM). Land surface models derived from these formulations have been applied to areas both large [*Ducharne et al.*, 2000] and small [*Beven and Kirkby*, 1979].

[15] The TOPMODEL framework does account for hydrologic connectivity; that is, the uplands and lowlands are dynamically linked via the main water table. Upland water supports and maintains lowland saturated regions via the relaxation of the hydraulic gradient under the influence of gravity and topography. During storm events, the hillslope hydraulic gradient and base flow increase, and lowland saturated regions expand. During dry down periods, the hydraulic gradient and base flow decrease, and lowland

saturated regions contract. However, as originally constructed, TOPMODEL does not account for storm flow, that is, shallow subsurface flow generated in the vadose zone by perched water tables (or lateral unsaturated flow) just subsequent to storm events. Nor does it account for any representation of bedrock; specifically, the spatial distribution of the depth-to-bedrock (DTB). We now show that by incorporating a storm flow mechanism and by implicitly representing the spatial variability of DTB, TOPMODEL can be used to gain an understanding of connectivity that takes place not at depth but in the shallow subsurface organic zone, a region dominated by high biological and biogeochemical activity.

[16] Despite the success of the TOPMODEL approach, simulations often perform poorly during drier conditions. The responses of catchments to wetting by spring snowmelt and to storms after an extended dry period have proven particularly difficult to represent. Specifically, *Shaman et al.* [2002] demonstrated that the water table in a TOPMODEL framework is relatively unresponsive to precipitation events during dry conditions. As originally formulated, TOPMODEL ignores shallow subsurface storm flow, defined here as downslope flow generated in the vadose zone by perched water tables (or lateral unsaturated flow). This negates the possibility for rapid flushing from the vadose zone during storm events, thus ignoring an important aspect of hydrologic connectivity as it relates the overall surface water balance and to material export from a catchment. To address these shortcomings, two new strategies were incorporated into the model framework: (1) we (as did *Datin* [1999] and *Scanlon et al.* [2001]) developed a methodology for representing the physical process of storm flow within a TOPMODEL framework [*Shaman et al.*, 2002]. The new method allows for discharge from regions of near saturation that develop in the vadose zone during storm events, thereby allowing for flushing of soil nutrients from the near-surface zones; (2) we introduce a method that makes use of calculated differences in the local depth-to-bedrock between upland and lowland areas (Figure 1). *Braun et al.* [2001], in a recent study of the variability of DTB and its geomorphic origins, found variability in soil depth to be a function of local slope, curvature, and drainage geometry. The authors were able to predict DTB using parameterizations of simple creep, depth-dependent creep, and transport by overland flow. Specifically, we use the topographic index to approximate the hillslope catena effect; that is, lowland areas with high topographic index values have deeper soils and upland areas with low topographic index values have more shallow soils. To implicitly incorporate the spatial distribution of the DTB, we reduce the porosity and field capacity with depth in the soil column. This results in a more realistic representation of soil depth differences between uplands and lowlands [see, *Cox and McFarlane*, 1995; *Webb and Burgham*, 1997; *Yanagisawa and Fujita*, 1999], and increases the responsiveness of groundwater flow (Figure 2) [*Shaman et al.*, 2002]. For example, Cascade Brook shows a tenfold increase in soil depth from ridge tops to lowlands while at Imnavait Creek there is a near-sevenfold increase in soil/peat depth from ridge tops to lowlands [*Walker and Walker*, 1996]. The impact of implementing these changes is an increased responsiveness of groundwater flow. Especially noticeable is the ability of the model to capture the transition from the extremely dry summer state to the rapid wetting resulting from hurricane Floyd to the subsequent storm recession. In addition, as shall be demonstrated below, by accounting for the differences in DTB between upland and lowland areas, we significantly modify our reconstructions of local WTDs and our view of the evolution hydrologic connectivity on the landscape. A detailed discussion on the catena effect and the details of the calculation of the local WTD for both a typical TOPMODEL reconstruction and this modified reconstruction are given in Appendix A.

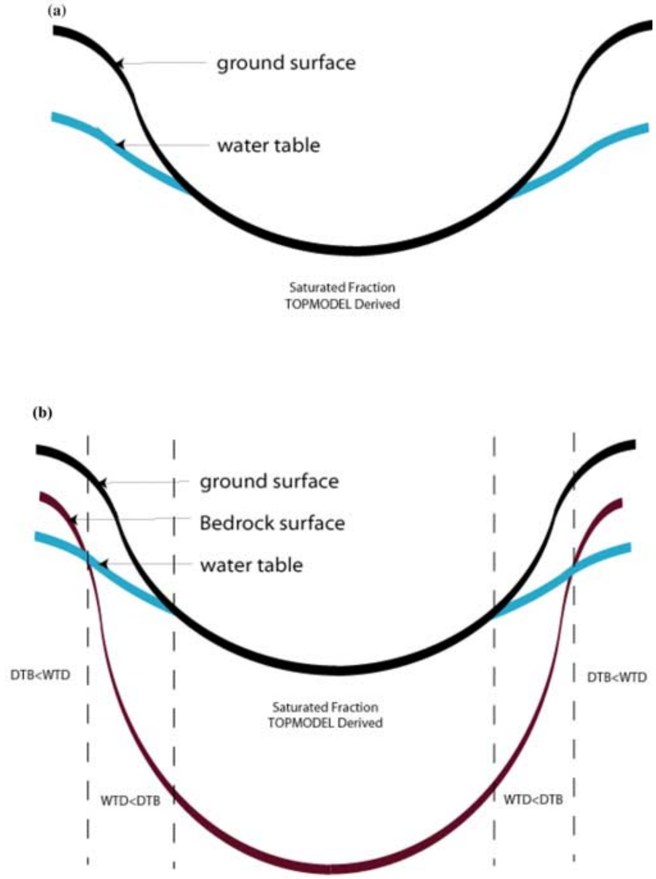

Figure 1. Schematic of the method that makes use of the differences in DTB between upland and lowland areas so as to modify the reconstruction of local WTDs (see Appendix A). (a) Depiction of water table outcropping in the lowlands and the fall off in the WTD as one progressively moves up the hillslope, (b) Relationship between the local surface topography, the water table, and the bedrock.

3.1. Cascade Brook

[17] Figure 3 depicts the simulated spatial reconstruction of the WTD at Cascade Brook at 10:00 PM during a storm event that lasted for 10 hours on 11 August 2000. In Figure 3a the topographic index ($\ln a/\tan\beta$) is shown for reference. The lowlands and streams (where the local water table outcrops) are best identified as the dark blue region.

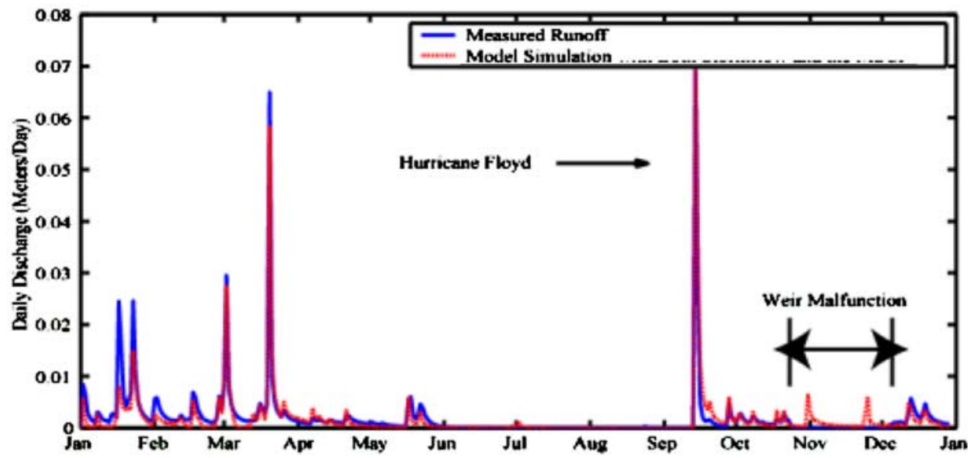

Figure 2. Discharge for Cascade Brook catchment in 1999. This simulation includes for storm flow and a decrease in soil porosity with depth (see section 3).

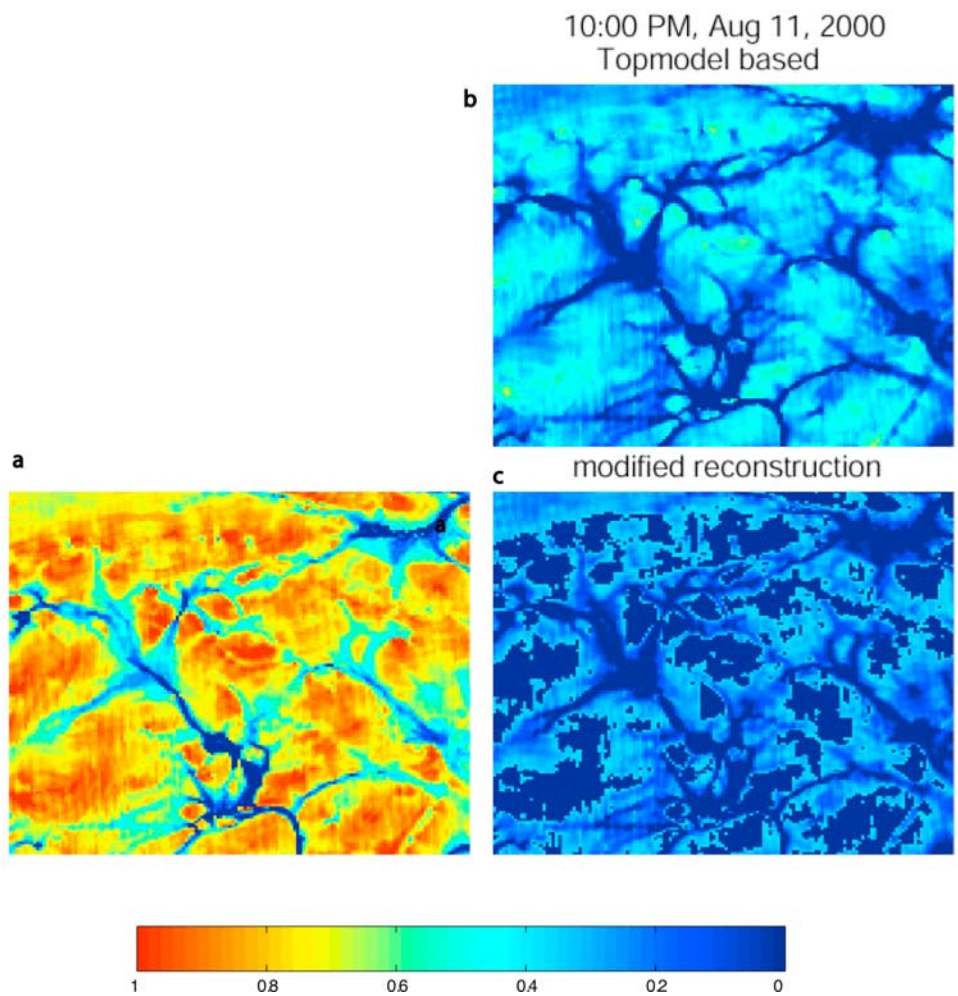

Figure 3. Spatial reconstruction of the WTD at Cascade Brook at 10:00 PM during a storm event on August 11, 2000. (a) For reference (a mapping of the topographic index), the lowlands and streams are best identified as the dark blue region. Moving progressively up the hillslope toward the ridges, the colors turn aqua, then yellow, then red. (b) Typical TOPMODEL WTD reconstruction. (c) WTD reconstruction that accounts for the differences in DTB between upland and lowland areas. The color scale relates to the WTD reconstructions and ranges from 0 (the water table at the surface) to 1 (the water table at or below a depth of 1 m).

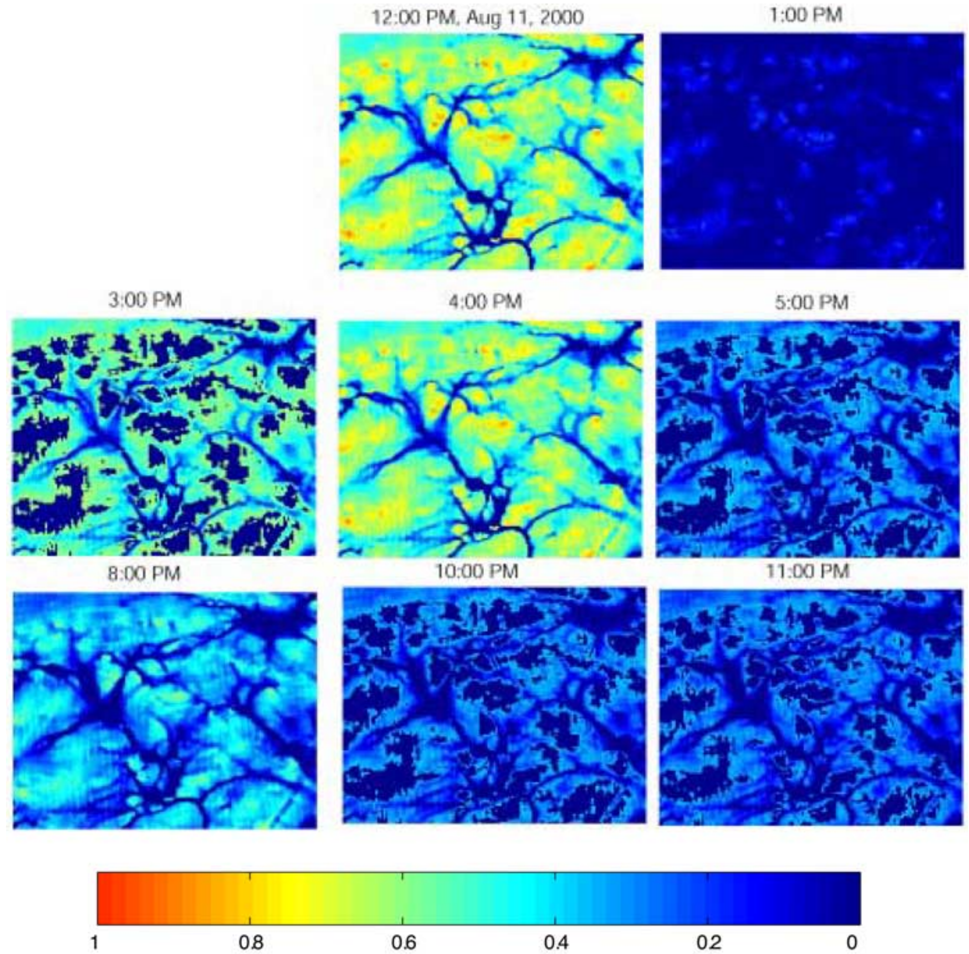

Figure 4. Cascade Brook water table depth reconstruction time series that accounts for the upland-lowland differences in the DTB. Antecedent conditions prior to the storm were relatively high. The color scale ranges from 0 (the water table at the surface) to 1 (the water table at or below a depth of 1 m). The degree of connectivity is indicated qualitatively. That is, two "potentially" connected regions have a higher relative degree of connectivity if their respective WTDs are closer to 0 than 1.

Moving progressively up the hillslope toward the ridges, the colors turn aqua, then yellow, then red. Two distinct depictions of the spatial distribution of the local WTD during a storm event are shown (Figures 3b and 3c).

[18] Figure 3b depicts a typical TOPMODEL WTD reconstruction. Wetness expands from the valleys up through the hillslopes. Upland regions remain dry and are in fact the last to wet and first to dry. For the WTD reconstruction in Figure 3b, no allowance is made for the fact that the ground surface is underlain by bedrock (Figure 1a). Therefore, progressing up the hillslope the WTD progressively deepens and the size of the vadose zone is assumed to increase. Consequently, more water is needed to saturate the hilltops.

[19] However, at Cascade Brook the medium texture soils are typically very thin, with parent material ranging in depth from less than 10 cm in the uplands to approximately 1 m below the surface in the depressional areas. Figure 3c depicts a modified TOPMODEL WTD reconstruction in which the inferred effects of local DTB are accounted for. In this case, wetness expands up from the lowlands and down from the ridge tops.

[20] Throughout this work we use the modified reconstruction of the local WTD as a proxy for understanding connectivity. No attempt is made to actually estimate laterally transported subsurface storm flow waters. Methodologies do exist by which the profile of vadose zone soil moisture could be estimated [*Koster et al.*, 2000], and such reconstructions could be used to infer pixel-to-pixel downslope lateral flow. However, at present, the existence of the necessary terrestrial validation is so sparse as to make this exercise of moving water laterally via pixel-to-pixel flow too speculative.

[21] Figures 4 and 5 show a time series of simulated WTDs in Cascade Brook for one storm event during a wet period with high antecedent soil moisture conditions, and another event during a relatively dry period. These time series reconstructions of surface wetness are generated from a model run that included storm flow and accounts for the spatial distribution of DTB, as well as the methodology for

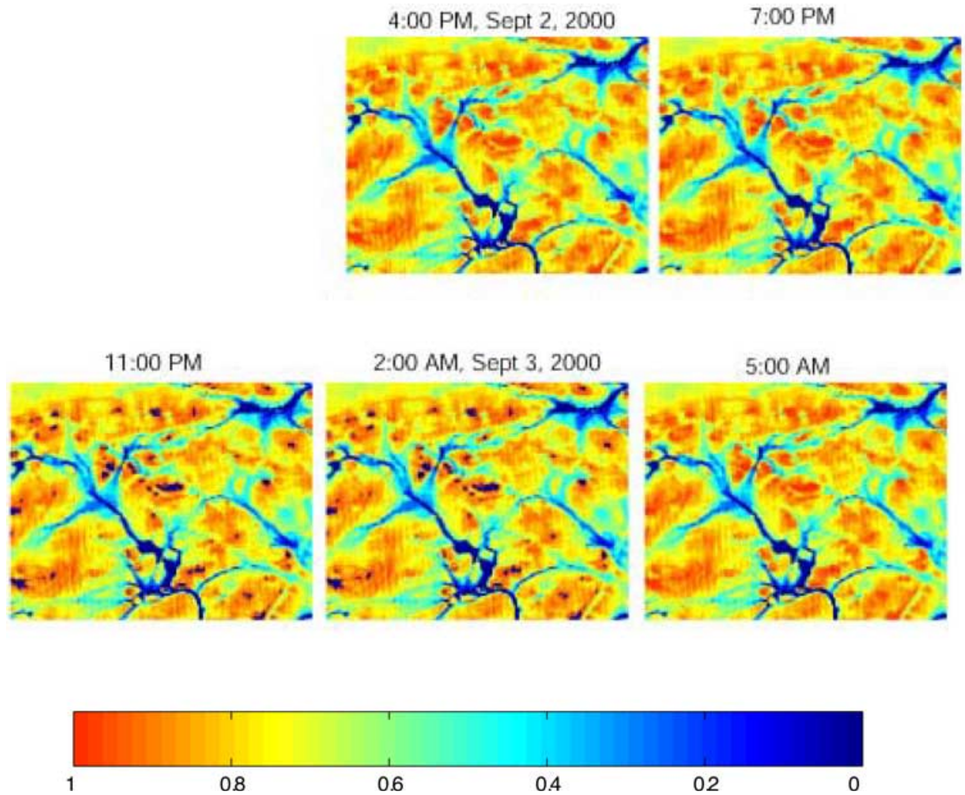

Figure 5. Cascade Brook water table depth reconstruction time series that accounts for the upland-lowland differences in the DTB. Antecedent conditions prior to the storm were relatively low. The color scale ranges from 0 (the water table at the surface) to 1 (the water table at or below a depth of 1 m). The degree of connectivity is indicated qualitatively. That is, two "potentially" connected regions have a higher relative degree of connectivity if their respective WTDs are closer to 0 than 1.

reconstruction of local WTD described above (and in Appendix A).

[22] The storm of 11 August 2000 occurred when antecedent moisture conditions were relatively high (Figure 4). The storm consisted of three separate events; an event (12:00–2:00 PM, August 11) that delivered 36 mm of precipitation, an event (4:00–5:00 PM) that delivered 2.6 mm of precipitation, and a final event (8:00–10:00 PM) that delivers 7.6 mm. Because antecedent moisture conditions are relatively high, just after the onset of precipitation, by 1:00 PM on 11 August, the entire catchment becomes hydrologically connected. (The degree of connectivity is indicated qualitatively. That is, two "potentially" connected regions have a higher relative degree of connectivity if their respective WTDs are closer to 0 than 1.0; see Figure 4.) By 2:00 PM the storm ceases and the catchment begins to dry, particularly in the mid-slope regions which neither benefit from the expansion of saturated lowlands nor have the shallow DTB as do the uplands. Still, regions of the upslope remain saturated and hydrologically connected through 3:00 PM. By 4:00 PM, however, the upper soil layers have dropped below field capacity and much of the catchment is hydrologically unconnected. Overall soil moisture conditions remain wet enough that the subsequent small rain event, delivering 2.6 mm after 4:00 PM brings the catchment back to full connectivity by 5:00 PM. At this point we not only see the downward expansion of saturated soils from the uplands, but also the upward expansion from the valleys to the hillslopes. Again, precipitation ceases and the upper reaches of the catchment dry out until the final rain event delivers 7.6 mm after 8:00 PM and brings the catchment to full connectivity by 10:00 PM. After 10:00 PM, all storm activity ends. During the next 2 hours, while the uplands progressively dry, the lowland saturated extent remains large and relatively unchanged, reflecting the slow groundwater recharge of the mean water table. It is interesting to note that the mid-slope region seems to behave as a ridge-valley cutoff switch as far as connectivity is concerned. That is, presumably when pore pressure (and soil moisture for that matter) in the mid-slope regions falls below a critical value, shallow ridge-to-valley connectivity ceases.

[23] We contrast this with the storm of 2–3 September 2000, which occurred when antecedent soil moisture conditions were relatively low (Figure 5). This storm consists of two small events (4:00–7:00 PM, September 2) that in total deliver 0.5 mm of rain, a larger event (10:00 PM on 2 September to 1:00 AM on 3 September) that delivers 2.6 mm, and a final small event (1:00–2:00 AM, 3 September) that delivers 0.6 mm. Because of high transpiration demand and low antecedent soil moisture conditions, the first two events deliver too little precipitation

to impact the WTDs as viewed by these reconstructions (Figure 5; 4:00 PM and 7:00 PM). The hillslopes remain dry and there is no apparent expansion of the lowland saturated zones. Only the large nighttime event, which occurs when transpiration demand is minimal, wets the uplands sufficiently to impact the reconstruction. Even so, WTD reconstructions throughout the period show that connectivity barely extends below the ridge tops. By 5:00 AM the next morning, the uplands have completely dried out, presumably due to vertical infiltration of vadose zone water. It should be said that during this storm when antecedent conditions are dry, ridge to valley connectivity is not precluded. If the ridges are sufficiently wet, it is possible that ridge to valley connectivity can occur via flow along the soil-bedrock interface. However, with low hydraulic conductivities at deeper levels in the mid-slope and the fact that the mid-slope remains significantly dry throughout the storm, discharge will most likely reflect localized activity near the riparian zone and not ridge to valley connectivity.

[24] The hillslope response during the wet and dry period storm events differs radically. These dissimilarities reflect differences in antecedent soil moisture conditions, total precipitation, storm intensity, and hydrologic connectivity. At the same time, the WTD reconstructions do show common features. In both cases, the ridges wet up fastest, followed by the upper hillslopes. This feature is the consequence of having the least soil moisture storage associated with these upper hillslope regions (low $\ln(a/\tan B)$). The mid-slopes neither benefit from the expansion of saturated lowlands nor had shallow soils, as do the uplands. Also, in both cases, the ridges and upper hillslopes dry out relatively quickly and nearly simultaneously, while the valleys remain wet for a prolonged period. In our simulations this is due to the fact that after the cessation of precipitation the lowlands receive downslope transported water for a prolonged period at the expense of upland soil moisture, which most probably occurs via the slow and deeper lateral transport of saturated water along and above the soil-bedrock interface.

3.2. Imnavait Creek

[25] In our previous work we demonstrated that a TOPMODEL-based approach could successfully simulate the dynamics of thermal, hydrological, and biological processes operating in a permafrost-dominated tundra setting [*Stieglitz et al.*, 1999, 2000, 2001]. In order to conduct successful simulations, we modified our model to account for two processes important at high latitudes, blowing snow and a seasonally evolving active layer (thaw zone). Because snow accumulation in valleys can be substantial [*Liston*, 1986; *Kane et al.*, 1991; *Liston and Sturm*, 1998], it takes longer to melt a deep snowpack over a reduced surface area compared to a pack that is uniformly distributed. Using snow heterogeneity data from Imnavait Creek [*Hinzman et al.*, 1996; *Liston and Sturm*, 1998], we developed a sub-grid-scale parameterization of snow that accurately predicts snowpack ablation and subsequent runoff (Figure 6a) [*Déry et al.*, 2003]. Our second major modification was to account for freeze-thaw processes and the seasonally evolving active layer, as is shown by the soil temperature reconstructions in Figure 6b. This scheme has been tested for seasonal evolution of ground temperatures in regions ranging from New England [*Lynch-Stieglitz*, 1994] to the Arctic [*Stieglitz et al.*, 1999, 2000, 2001], where permafrost dynamics play a large role in the seasonal hydrologic cycle.

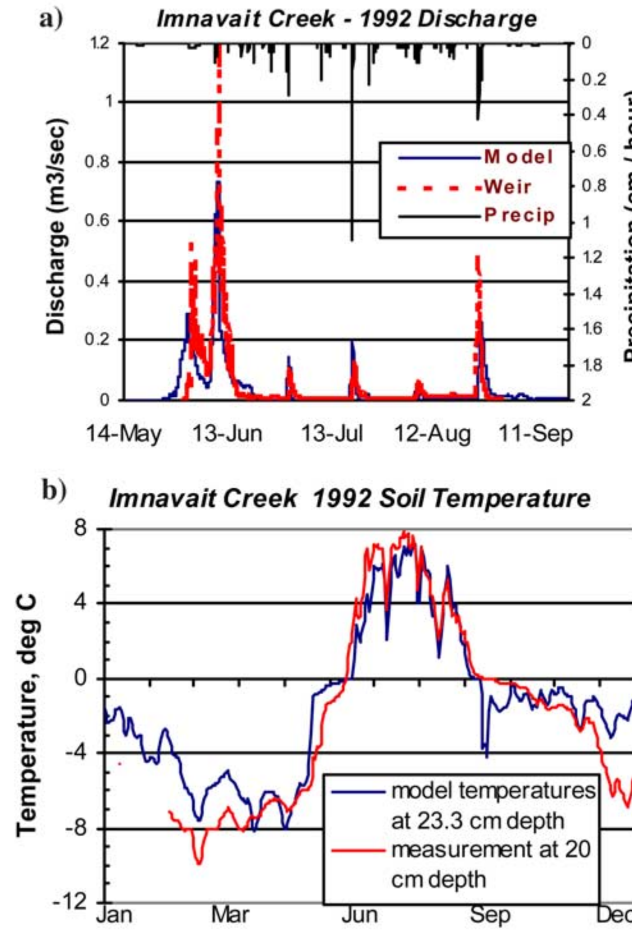

Figure 6. (a) By accounting for snow distribution, our three-layer snow model can properly simulate the timing of the snowpack ablation as well as the subsequent impact this has on the timing and amount of melt related discharge. By accounting for the impact that storm flow has on the timing of discharge during relatively dry periods, we can better simulate storm recession [*Shaman et al.*, 2002]. (b) To simulate ground freeze-thaw processes and the seasonally evolving active layer, a multilayer ground scheme is used in which heat transport is physically modeled via transport along the thermal gradient [*Abramopoulos et al.*, 1988; *Bonan*, 1996; *Lynch-Stieglitz*, 1994; *Stieglitz et al.*, 2001].

[26] Changing hydrological connectivity at Imnavait Creek is shown through the spatial dynamics of the WTD during a storm event in August 1992 (Figure 7). The storm began at ~2:00 PM on 25 August 1992, delivered a total of 28 mm of rain, and ended by 11:00 PM the next day. Prior to the onset of precipitation, simulations show a clear moisture gradient along the toposequence (at Imnavait Creek, measurements show DTB ranging from 0.1 m near the ridge tops to 0.75 m near the valley bottoms [*Hinzman*

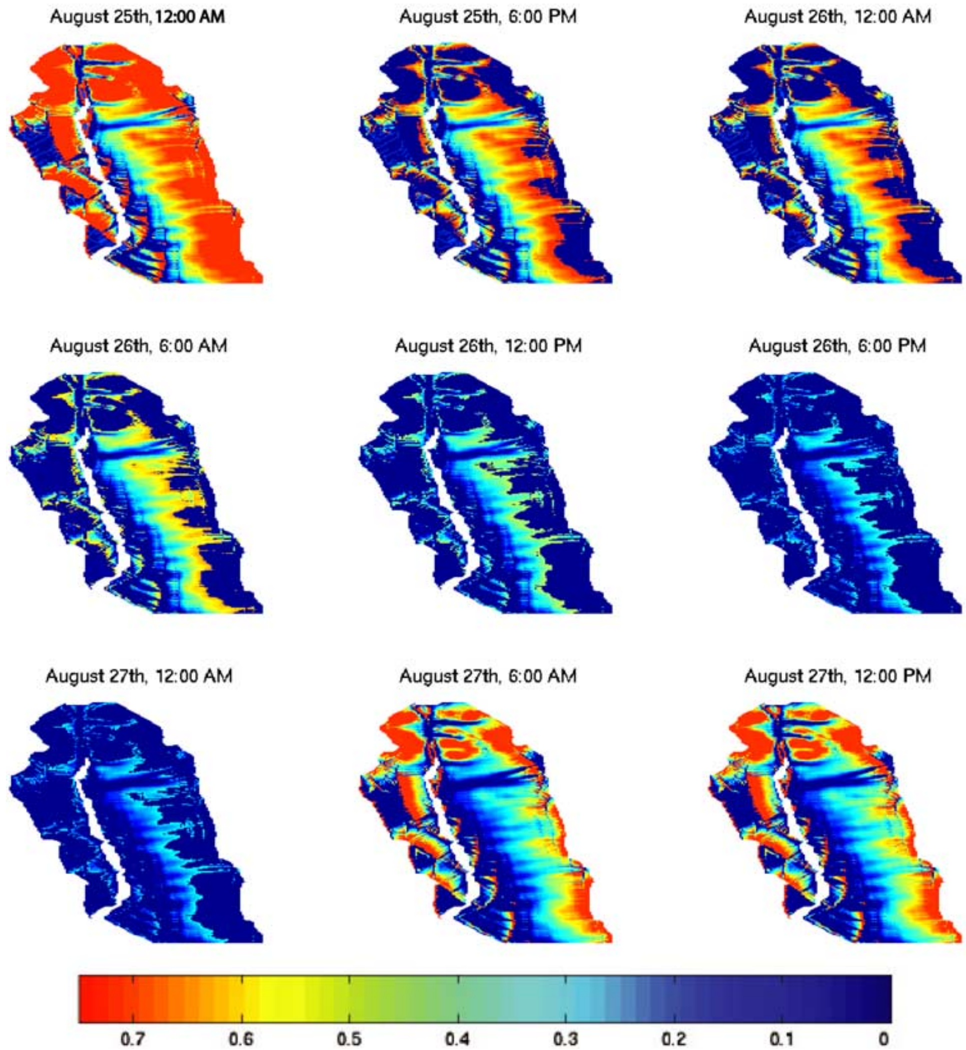

Figure 7. Changing hydrological connectivity is shown through the spatial dynamics of WTD changes at the Imnavait Creek catchment during a storm event in August 1992. The color scale ranges from 0 (the water table at the surface) to 0.75 (the water table at or below a depth of 0.75 meter). The degree of connectivity is indicated qualitatively. That is, two ‘‘potentially’’ connected regions have a higher relative degree of connectivity if their respective WTDs are closer to 0 than 0.75.

et al., 1991, 1996; *Kane et al.*, 1991; *McNamara et al.*, 1997; *Walker and Walker*, 1996]). In this simulation the stream channel is delineated in white. These simulations depict a clear pattern of soil moisture increase first in lowlands and ridge tops, but not the mid-slope, which delays complete hydrological connectivity of the full hillslope. Although by 6:00 PM on 25 August enough rain had fallen to saturate the shallow-soil ridge tops, they were still hydrologically disconnected from the lowlands. As the storm proceeded, there was a downward expansion of saturation from the uplands as well as upward expansion from the valleys. By 6:00 PM on 26 August the catchment was hydrologically connected. At 11:00 PM the storm ceased and the catchment began to dry in the mid-slope and upper reaches of the hillslope. As the uplands continued to dry the lowlands remained largely saturated which reflects the slow groundwater recharge of the mean water table.

4. Is There Observational Evidence for Hydrological and Chemical Connectivity?

[27] The modeling studies and interpretation described above present a hypothesis about how hydrologic connectivity may operate. To test this hypothesis, we need to know if these same mechanisms are responsible for hydrologic connectivity in the real world as in our model. At this point the detailed spatial measurements of soil moisture and water chemistry are not available to confirm or refute this hypothesis. However, we can explore some relevant, available hydrologic and chemical data from a small headwater catchment in Idaho.

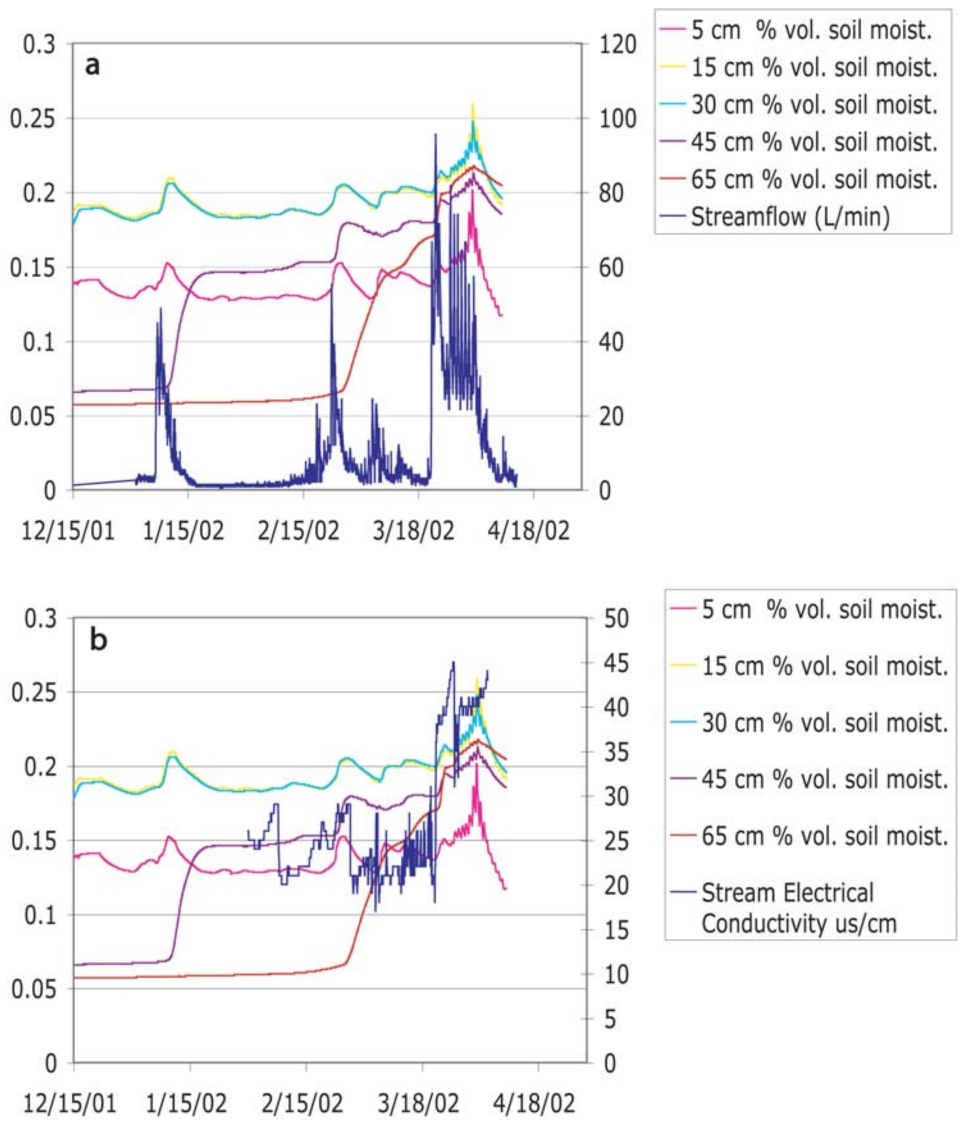

Figure 8. (a) Stream discharge (liters per minute; right hand axis) for the 2002 snowmelt season at the Upper Dry Creek Catchment, and (b) stream water electrical conductivity (micro Siemens per cm; right hand axis). Both panels show the volumetric soil moisture at depths of 5, 15, 30, 45, and 65 cm (left hand axis). The TDR probes were placed horizontally on the north facing mid-slope.

[28] The Upper Dry Creek catchment is oriented east-west, and thus is characterized by north-south facing slopes. As such, the winter snowpack on the south-facing slope generally melts prior to the pack on the north-facing slope. Furthermore, redistribution of snow by wind provides the north-facing slope with more snow than the south-facing slope (not shown). The problem this poses for modeling the hydrograph within the TOPMODEL framework is imposing. More imposing is the associated remapping of WTD throughout the catchment during the melt season. We have yet to modify our model to account for these difficulties. Instead, we will use the snowmelt hydrograph-chemograph data to better understand the relationship between the initiation of hydrologic connectivity and downslope transport of materials in this catchment, and to understand the simulations at Cascade Brook and Imnavait Creek in a broader perspective.

[29] High-resolution stream chemistry measurements taken during the snowmelt period or during storm events offer a unique opportunity to observe nonlinear initiation of lateral flow on the landscape. For example, stream water electrical conductivity is relatively low in Upper Dry Creek during the winter and then rises dramatically with streamflow at the beginning of the final snowmelt event in 2002 (Figures 8a and 8b). We argue that the rise in electrical conductivity at Upper Dry Creek occurs when the hillslopes become hydrologically connected to the stream by lateral subsurface flow. An alternative explanation for the step-rise in stream electrical conductivity is that older, relatively concentrated groundwater may be contributing to the

stream. However, snowmelt hydrograph separations, end-member mixing analysis (EMMA), and water budget calculations conducted at the site indicate that there are no contributions from deep groundwater to the stream [*Yenko*, 2003]. Measurements do indicate that the catchment loses about 30% of its water to deep groundwater recharge per year. However, this water emanates much further downstream from the weir of this headwater catchment [*Yenko*, 2003]. In any case, catchments that have base flow sustained by groundwater typically experience a chemical dilution during the snowmelt period, but in Upper Dry Creek we see the opposite pattern, which is consistent with our observation that groundwater sources are minimal. We proceed with the argument that the rise in stream electrical conductivity during the snowmelt event likely results when previously disconnected hillslope sources of solutes becomes hydrologically connected to the stream.

[30] We interpret the observed hydrograph-chemograph relationship as follows. During midwinter, flow events are characterized by low electrical conductivity and driven by near-stream melt (near-stream probe data not shown); the mid-slope is not involved as the soil moisture response at mid-slope clearly lags the stream water response (Figures 8a and 8b). Because upland soils were relatively dry from evapotranspiration during the preceding year, it takes a significant amount of time before these soils are recharged to the point where field capacity is surpassed and the necessary pore pressure gradient is achieved to initiate downslope transport of water (Figure 8a). The low conductivity of stream water in midwinter suggests that there is no lateral flow at this time and no hydrologic connectivity between the hillslope and stream. As the melt progresses, hillslope soil pore water becomes increasingly concentrated with solutes formed by weathering reactions because upland meltwater-soil contact is prolonged. Beginning in January (at 45 cm) and February (at 65 cm), meltwater recharge associated with midwinter melt events initiates a gradual but sustained increase in soil moisture that preconditions the mid-slope profile. Near March 21, soil moisture increases dramatically at all depths, concurrent with a sharp increase in streamflow (Figure 8a) and in stream water conductivity (Figure 8b). The sudden and simultaneous increase in soil moisture and flow represents the volumetric soil moisture at which sufficient pore pressure is achieved for lateral flow to occur. At this point in time, the mid-slope region is sufficiently wet that water moving laterally from the saturated ridges toward the mid-slopes, as well as infiltration meltwater entering the mid-slopes, will move laterally (downslope) toward the stream rather than be impeded by unsaturated conditions. Note that at the end of snowmelt in early April, soil moisture decreases at all depths, particularly at 5 cm, timed with a sharp decrease in stream discharge to near zero. While discharge falls, however, deep soil moisture remains high. These patterns suggest that “top-down” wetting of the deeper soils is a necessary pre-condition to initiate connectivity and lateral downslope flow (and nutrient flushing), but that the main meltwater pulse is from water higher up in the soil column where the hydraulic conductivities are the greatest.

[31] From our modeling studies, we expect that when mid-slope saturation (or near saturation) occurs, the catchment can be connected from ridge to valley. The results observed at Upper Dry Creek are consistent with our modeling: a simultaneous increase in flow, and higher electrical conductivity stream water reflecting the release of the water from the upper reaches of the catchment once the mid-slope has been sufficiently preconditioned (wetted). During previous discharge episodes in January and February, where the mid-slope soil column does not saturate, the electrical conductivity in the stream remains low, reflecting the local (lower slope) source of the meltwater. However, while we have explored simulations at two catchments, as well as observations at another catchment, full validation of our ideas will most likely require sampling programs specifically targeted toward addressing these issues. On the other hand, observations at Upper Dry Creek seem to support the mechanisms we propose in our modeling studies.

5. Conclusions

[32] Our model simulations are consistent with the *Grayson et al.* [1997] proposition that two distinct hydrologic states predominate: (1) a dry state, in which hillslope regions are hydrologically unconnected, the dominant flow is vertical, and spatial patterns of soil moisture are unorganized, and (2) a wet state, in which hillslope regions are connected via lateral subsurface flow and soil moisture patterns are organized spatially. Also consistent with *Grayson et al.*’s [1997] results, we find a high correlation between the TOPMODEL topographic index and the spatial distribution of WTDs when the soils are wet (and much less so during dry conditions). We see that the mid-slope regions behave as a ridge-valley cutoff switch as far as shallow subsurface connectivity is concerned. That is, when pore pressure, and wetness, in the mid-slope regions falls below a critical value, shallow ridge-to-valley connectivity ceases and the reduction of runoff is quite substantial.

[33] The simulations presented here suggest that for much of the year the water draining through a catchment is spatially isolated. That is, only rarely, during storm events when antecedent soil moisture is high, are the uplands and lowlands actually connected hydrologically. For example, in the catchments discussed above, hydrologic connectivity among the different regions of the hillslope is most pronounced during the spring snowmelt, but is more sporadic in the summer when it is dependent on antecedent conditions and storm activity. Summertime transpiration may reinforce this tendency toward a hydrological disconnect between upper slopes and the mid-slope-riparian zones. Consequently, the nutrients and other elements flushed from the catchment during the growing season mostly originate in near-stream riparian zones.

[34] At the catchment level, this seasonal timing of hydrologic connectivity may have significant ramifications for a range of ecological questions. Such questions include how the spatial heterogeneity of vegetation and variable hydrological connectivity impacts C-N cycling and turnover, for example, the balance between water flow increas-

ing nutrient supply and waterlogging depleting local oxygen and decreasing nutrient production from decomposition. From a plant's perspective, waterlogging slows mineralization rates of N, but water movement from upslope provides new nutrients. We hypothesize that the balance between these two effects may result in maximum production somewhere in the middle of the slope. Consequently, the complexity of variable spatial hydrologic connectivity may critically control the cycling rates and export of nutrients.

[35] We propose that time-variable spatial heterogeneity (connectivity) accounts for much of the variation in biological productivity along a toposequence, and for much of the variability at daily to seasonal timescales in the export of materials from a catchment. These simulations and observations provide evidence of the variable hydrological connectivity occurring in catchments of different size and characteristics. While these conclusions are consistent with the available data, additional experiments are needed to better understand the mechanisms leading to hydrologic connectivity and downslope movement of water and nutrients along hillslopes. Our simulations provide testable hypotheses concerning these mechanisms and point the way toward the types of experiments that need to be conducted. This in turn will be critical to our understanding of the general controls on export of materials from land to water at regional scales and for the entire globe.

[36] Finally, if these types of hypotheses are to be adequately tested much work is still needed:

[37] 1. Assuming equilibrium morphology, *Heimsath et al.* [2001] found simple relationships between topographic curvature (from DEMs) and soil depth that are supported by data. However, these relationships do not apply to all landscapes, especially those undergoing regular disturbance. At this point in time, adequate experimental data are not available to definitively relate the hillslope location and local DTB. Therefore, while we believe that our WTD reconstructions based on limited site DTB information yield a more realistic spatial soil moisture distribution during dry periods, more data are clearly needed.

[38] 2. While the mechanisms invoked above for flushing of dissolved and particulate materials from the hillslope portray water and material moving downslope in unison, this is not quite accurate. It is important to distinguish the mechanisms by which water moves (kinematic wave) (as used in the *Beven and Freer* [2001] TOPMODEL framework) from the mechanisms by which material is transported. Once the mid-slope no longer impedes lateral flow, further water infiltrating the upslope region most probably establishes the necessary pressure wave movement to flush downslope solutes to the stream system. This most likely results in dissolved and particulate material located in the riparian zone being flushed first [*Boyer et al.*, 1997].

[39] 3. Along these lines, we have focused on the flushing of high conductivity water from a pre-conditioned hillslope. Ridge-to-valley connectivity was assumed. Flushing, however, can occur via piston action or when connectivity is localized to the near-riparian zone. For example, during dry down periods, especially in summer months, dissolved organic carbon (DOC) may preferentially build-up in the organic rich, deep near-riparian areas where, just months earlier, anaerobic conditions prevailed. A subsequent storm may not establish ridge-to-valley connectivity, but local riparian connectivity may be established, flushing the high concentration DOC water [*Schiff et al.*, 1990, 1997]. This type of flushing is only qualitatively different from that presented in the body of this work; connectivity is still invoked. However, ridge-to-valley connectivity is not a pre-requisite for downslope movement of dissolved or particulate material in soil waters.

[40] 4. The hydrograph-chemograph relationship is dependent on other factors; the dissolved or particulate materials being transported have a wide range of adsorption properties and bio-chemical reactivity rates. Given identical flushing rates, different dissolved or particulate material will move downslope at different rates.

[41] 5. Finally, it is our hope that studies such as this serve to establish a dialogue among the modelers and the experimentalists in order to better coordinate efforts with the goal of better understanding hydrological connectivity and the subsequent implications for downslope nutrient transport, C/N cycling, and biological productivity along the toposequence. As such, both model simulations and experimental field campaigns in the future will need to be conducted in coordination and at the same catchments.

Appendix A

[42] Within a TOPMODEL framework, it is possible to model surface wetness across individual plots of land as small as 25 m^2 (e.g., 5-m grid cells), and to monitor how the surface wetness of these plots changes through time. Our model uses gridded digital maps of land surface topography and a dynamic numerical framework that accounts for the movements of water, i.e., TOPMODEL [*Beven and Kirkby*, 1979; *Beven*, 1986a, 1986b; *Beven et al.*, 1994], and energy within the soil and at the surface [*Stieglitz et al.*, 1997]. The topographic index $\ln(a/\tan\beta)$ is first calculated for each pixel cell. At each point in time, the mean water table depth (WTD) and a probability density function of the topographic index values are then used to compute the saturated areas of the watershed and the shallow groundwater flow that supports these areas. We can also use the mean WTD and the topographic index value of each cell to calculate the local WTD for each cell,

$$z_x = \bar{z} - 1/f\left[\ln(a/\tan\beta)_x - \lambda\right], \tag{A1}$$

where $\ln(a/\tan\beta)_x$ is the local topographic index at location x, z_x is the local water table depth at location x, $\bar{z}$ is the mean water table depth, WTD, λ is the mean watershed value of $\ln(a/\tan\beta)$, and f is the rate of decline of the saturated hydraulic conductivity with depth in the soil column. Thus at any point in time we can create a mosaic of cells, each denoted by a local WTD, which, taken as a whole, represents the surface conditions of the entire watershed. This mosaic of WTDs depicts the spatial variability of conditions at the land surface that result from terrain (topography, vegetation, and soil type) and integrated weather forcing (meteorological conditions) (Figure 3b, for example). Saturated areas include all pixel cells for which local WTD is at or above the surface. As TOPMODEL was originally formulated, DTB is not accounted for. As such, drier upland areas have pixel cells with local WTD well below the surface.

[43] However, during and immediately after rain events, we expect two regions of a catchment to wet up first; near-saturated lowlands, and uplands, which are characterized by a shallow DTB and therefore have limited water-holding capacity. However, while TOPMODEL-based reconstructions of local WTD depict saturated lowlands, no such wetting of the uplands is provided (Figure 3b). Instead, upland regions remain dry, and are in fact the last to wet and the first to dry. This unrealistic representation of upland wetness is a consequence of the TOPMODEL framework, which does not account for variability in DTB and the impact that such variations can have on local rates of saturation.

[44] We introduce a method that makes use of calculated differences in DTB between upland and lowland areas to modify the reconstruction of local WTDs. *Braun et al.* [2001] in a recent study of the variability of DTB and its geomorphic origins, found variability in soil depth to be a function of local slope, curvature, and drainage geometry. The authors were able to predict DTB using parameterizations of simple creep, depth-dependent creep, and transport by overland flow. At Cascade Brook and Imnavait Creek we lacked the comprehensive measurements of soil depth needed to fit such a parameterization. Alternatively, we use transport by overland flow, controlled by local slope and upslope area, (i.e., $\ln(a/\tan\beta)$) as the sole determinant of DTB. This choice provides one measure of the variability in soil depth between upland and lowland regions, and permits use of the topographic index, which is similar in form to the overland flow term used by *Braun et al.* [2001], to calculate local DTB.

[45] Specifically, we use the topographic index to approximate the hillslope catena effect; that is, lowland areas with high topographic index values have deeper soils and upland areas with low topographic index values have more shallow soils. The topographic index is also used to determine upland saturated areas during and immediately after rain events. In the Cascade Brook experimental catchment, the topographic index ranges from 3 to 30. Soil core measurements taken at 11 sites throughout Cascade Brook yield a DTB range from 0.1 to 1 m (William Schuster, unpublished data). A linear regression relating our ranges for the topographic index to measured DTB is then determined. We then use this $\ln(a/\tan\beta)_x$-DTB_x relationship in our modified reconstruction of local WTD.

[46] At Imnavait Creek the topographic index ranges from 4 to 17 and measurements show DTB ranging from 0.1 m near the ridge tops to 0.75 m near the valley bottoms [*Hinzman et al.*, 1991, 1996; *Kane et al.*, 1991; *McNamara et al.*, 1997; *Walker and Walker*, 1996]. A linear solution was fit to these endpoints to produce a modeled $\ln(a/\tan\beta)_x$-DTB_x relationship for Imnavait Creek.

[47] For the reconstructions shown here we only recalculate the local WTD based on depth-to-bedrock when there is significant water in the soil column (top 10 cm, the first two model layers of our 10-layer soil column model [*Stieglitz et al.*, 1997; *Shaman et al.*, 2002] are above field capacity). At each point in time for which the top two model layers are above field capacity, and for pixel cells in which the local DTB is shallower than the mean WTD (see schematic Figure 1b), all the water in the vadose zone (area above the model mean WTD) is compressed into the soil column pore space defined by the local DTB. If this water exceeds this local pore space, the pixel is considered saturated at the surface. Otherwise, the local WTD is calculated as per equation (A1).

[48] We stress that our inclusion of the spatial heterogeneity of DTB is by no means the only or best solution. Both within the TOPMODEL hydrologic framework and using more explicitly based models, others have attempted to deal with the question of the impacts that landscape spatial heterogeneity within a watershed (e.g., soil transmissivity, soil depth, vegetation) have on hydrologic [*Beven and Freer*, 2001; *Calver and Wood*, 1991; *Kirkby*, 1998; *Watson et al.*, 1998; *Wigmosta et al.*, 1994; *Wigmosta and Lettenmaier*, 1999; *Tague and Band*, 2001] and biogeochemical transport [*Band et al.*, 2001]. What we have attempted to do in this paper is to use a simple parameterization of DTB as a means for exploring and understanding hydrologic connectivity and the implications this has on downslope nutrient transport.

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References

Aber, J. D., W. H. McDowell, K. J. Nadelhoffer, A. Magill, G. Bernston, M. Kamakea, S. G. McNulty, W. Currie, L. Rustad, and I. Fernandez, Nitrogen saturation in temperate forest ecosystems: Hypotheses revisited, *BioScience*, *48*, 921–934, 1998.

Abramopoulos, F., C. Rosenzweig, and B. Choudhury, Improved ground hydrology calculations for global climate models (GCMs): Soil water movement and evapotranspiration, *J. Clim.*, *1*, 921–941, 1988.

Ambroise, B., K. Beven, and J. Freer, Toward a generalization of the TOPMODEL concepts: Topographic indices of hydrological similarity, *Water Resour. Res.*, *32*, 2135–2145, 1996.

Anderson, E. A., and D. R. Baker, Estimating incident terrestrial radiation under all atmospheric conditions, *Water Resour. Res.*, *3*, 975–988, 1967.

Band, L. E., C. L. Tague, P. Groffman, and K. Belt, Forest ecosystem processes at the watershed scale: Hydrological and ecological controls of nitrogen export, *Hydrol. Processes*, *15*, 2013–2028, 2001.

Beaulac, M., and K. Reckhow, An examination of land use-nutrient export relationships, *Water Resour. Bull.*, *18*, 1013–1024, 1982.

Beven, K. J., Hillslope runoff processes and flood frequency characteristics, in *Hillslope Processes*, edited by A. D. Abrahams, pp. 187–202, Allen and Unwin, Concord, Mass., 1986a.

Beven, K. J., Runoff production and flood frequency in catchments of order ***n***: An alternative approach, in *Scale Problems in Hydrology*, edited by V. K. Gupta, I. Rodriguez-Iturbe, and E. F. Wood, pp. 107–131, D. Reidel, Norwell, Mass., 1986b.

Beven, K., TOPMODEL: A critique, *Hydrol. Processes*, *11*, 1069–1085, 1997.

Beven, K., and J. Freer, A dynamic TOPMODEL, *Hydrol. Processes*, *15*, 1993–2011, 2001.

Beven, K. J., and M. J. Kirkby, A physically-based variable contributing area model of basin hydrology, *Hydrol. Sci. J.*, *24*, 43–69, 1979.

Beven, K., P. Quinn, R. Romanowicz, J. Freer, J. Fisher, and R. Lamb, TOPMODEL and GRIDATB, A user's guide to the distribution versions

(94.01), report, Cent. for Res. on Environ. Syst. and Stat., Lancaster Univ., Lancaster, England, 1994.

Bonan, G. B., A land surface model (LSM Version 1.0) for ecological, hydrological and atmospheric studies: Technical description and user's guide, report, 150 pp., Natl. Cent. for Atmos. Res., Boulder, Colo., 1996.

Boyer, E. W., G. M. Hornberger, K. E. Bencala, and D. M. McKnight, Response characteristics of DOC flushing in an alpine catchment, *Hydrol. Processes*, *11*, 1635–1647, 1997.

Braun, J., A. M. Heimsath, and J. Chappell, Sediment transport mechanisms on soil-mantled hillslopes, *Geology*, *29*(8), 683–686, 2001.

Calver, A., and W. L. Wood, Dimensionless hillslope hydrology, *Proc. Inst. Civ. Eng., Part 2*, *91*, 593–602, 1991.

Cooper, L. W., I. L. Larsen, C. Solis, J. M. Grebmeier, C. R. Olsen, D. K. Solomon, and R. B. Cook, Isotopic tracers for investigating hydrological processes, in *Landscape Function and Disturbance in Arctic Tundra*, edited by J. F. Reynolds and J. D. Tenhunen, pp. 165–182, Springer-Verlag, New York, 1996.

Cosby, B. J., G. M. Hornberger, J. N. Galloway, and R. F. Wright, Modeling the effects of acid deposition: Assessment of a lumped-parameter model of soil water and stream chemistry, *Water Resour. Res.*, *21*, 51–63, 1985.

Cox, J. W., and D. J. McFarlane, The causes of waterlogging in shallow soils and their drainage in southwestern Australia, *J. Hydrol.*, *167*, 175–194, 1995.

Creed, I. F., and L. E. Band, Exploring functional similarity in the export of nitrate-N from forested catchments: A mechanistic modeling approach, *Water Resour. Res.*, *34*, 3079–3093, 1998.

Datin, R., The development of tools for flash flood forecasting: The introduction of spatial variability in TOPMODEL, *Houille Blanche*, *54*(7–8), 13–20, 1999.

Déry, S. J., W. T. Crow, M. Stieglitz, and E. F. Wood, Modeling snowcover heterogeneity over complex Arctic terrain for regional and global climate models, *J. Hydrometeorol.*, in press, 2003.

Ducharne, A., R. D. Koster, M. J. Suarez, M. Stieglitz, and P. Kumar, A catchment-based approach to modeling land surface processes in a general circulation model: 2. Parameter estimation and model demonstration, *J. Geophys. Res.*, *105*, 24,823–24,838, 2000.

Everett, K. R., D. L. Kane, and L. D. Hinzman, Surface water chemistry and hydrology of a small arctic drainage basin, in *Landscape Function and Disturbance in Arctic Tundra*, edited by J. F. Reynolds and J. D. Tenhunen, pp. 185–202, Springer-Verlag, New York, 1996.

Fisher, T. R., D. Correll, R. Costanza, J. Hollibaugh, C. Hopkinson, R. Howarth, N. Rabalais, J. Richey, C. Vorosmarty, and R. Wiegert, Synthesizing drainage basin inputs to coastal systems, in *Estuarine Science*, edited by J. E. Hobbie, pp. 81–106, Island, Washington, D. C., 2000.

Frank, H., S. Patrick, W. Peter, and F. Hannes, Export of dissolved organic carbon and nitrogen from Gleysol dominated catchments: The significance of water flow paths, *Biogeochemistry*, *50*, 137–161, 2000.

Giblin, A. E., K. J. Nadelhoffer, G. R. Shaver, J. A. Laundre, and A. J. McKerrow, Biogeochemical diversity along a riverside toposequence in arctic Alaska, *Ecol. Monogr.*, *61*, 415–436, 1991.

Grayson, R. B., A. W. Western, F. H. S. Chiew, and G. Bloschl, Preferred states in spatial soil moisture patterns: Local and nonlocal controls, *Water Resour. Res.*, *33*, 2897–2908, 1997.

Heimsath, A. M., W. E. Dietrich, K. Nishiizumi, and R. C. Finkel, Stochastic processes of soil production and transport: Erosion rates, topographic variation and cosmogenic nuclides in the Oregon Coast Range, *Earth Surf. Processes Landforms*, *26*, 531–552, 2001.

Hinzman, L. D., and D. L. Kane, Snow hydrology of a headwater arctic basin: 2. Conceptual analysis and computer modeling, *Water Resour. Res.*, *27*, 1111–1121, 1991.

Hinzman, L. D., D. L. Kane, R. E. Gieck, and K. R. Everett, Hydrologic and thermal-properties of the active layer in the Alaskan Arctic, *Cold Reg. Sci. Technol.*, *19*(2), 95–110, 1991.

Hinzman, L. D., D. L. Kane, C. S. Benson, and K. R. Everett, Energy balance and hydrological processes in an arctic watershed, in *Landscape Function and Disturbance in Arctic Tundra*, edited by J. F. Reynolds and J. D. Tenhunen, pp. 131–154, Springer-Verlag, New York, 1996.

Hinzman, L. D., D. J. Goering, and D. L. Kane, A distributed thermal model for calculating soil temperature profiles and depth of thaw in permafrost regions, *J. Geophys. Res.*, *103*, 28,975–28,992, 1998.

Hornberger, G. M., K. E. Bencala, and D. M. McKnight, Hydrological controls on dissolved organic-carbon during snowmelt in the Snake River near Montezuma, Colorado, *Biogeochemistry*, *25*, 147–165, 1994.

Howarth, R. W., et al., Regional nitrogen budgets and riverine N and P fluxes for the drainages to the North Atlantic Ocean: Natural and human influences, *Biogeochemistry*, *35*, 75–139, 1996.

Howarth, R., N. Jaworski, D. Swaney, A. Townsend, and G. Billen, Some approaches for assessing human influences on fluxes of nitrogen and organic carbon to estuaries, in *Estuarine Science: A Synthetic Approach to Research and Practice*, edited by J. E. Hobbie, pp. 17–41, Island, Washington, D. C., 2000.

Judd, K. E., and G. W. Kling, Production and export of dissolved C in arctic tundra mesocosms: The roles of vegetation and water flow, *Biogeochemistry*, *60*, 213–234, 2002.

Kane, D. L., L. D. Hinzman, C. S. Benson, and G. E. Liston, Snow hydrology of a headwater arctic basin: 1. Physical measurements and process studies, *Water Resour. Res.*, *27*, 1099–1109, 1991.

Kashulina, G., C. Reimann, T. E. Finne, P. de Caritat, and H. Niskavaara, Factors influencing NO_3 concentrations in rain, stream water, ground water and podzol profiles of eight small catchments in the European Arctic, *Environ. Pollut.*, *102*, 559–568, 1998.

Kirkby, M., Hillslope runoff processes and models, *J. Hydrol.*, *100*, 315–340, 1998.

Kling, G. W., Land-water interactions: the influence of terrestrial diversity on aquatic ecosystems, in *Arctic and Alpine Biodiversity*, edited by T. Chapin and C. Korner, pp. 296–308, Springer-Verlag, New York, 1995.

Koster, R. D., M. J. Suarez, A. Ducharne, M. Stieglitz, and P. Kumar, A catchment-based approach to modeling land surface processes in a general circulation model: 1. Model structure, *J. Geophys. Res.*, *105*, 24,809–24,822, 2000.

Lee, K.-Y., T. R. Fisher, T. E. Jordan, D. L. Correll, and D. E. Weller, Modeling the hydrochemistry of the Choptank River Basin using GWLF and Arc/Info: 1. Model calibration and validation, *Biogeochemistry*, *49*, 143–173, 2000.

Lewis, W. M., and J. F. Saunders, Concentration and transport of dissolved and suspended substances in the Orinoco River, *Biogeochemistry*, *7*, 203–240, 1989.

Likens, G. E., and F. H. Bormann, Linkages between terrestrial and aquatic ecosystems, *Bioscience*, *24*, 447–456, 1974.

Likens, G. E., F. H. Bormann, and N. M. Johnson, Interactions between major biogeochemical cycles in terrestrial ecosystems, in *Some Perspectives of the Major Biogeochemical Cycles*, edited by G. E. Likens, pp. 11–93, John Wiley, New York, 1981.

Liston, G. E., Seasonal snowcover of the foothills region of Alaska's arctic slope: A survey of properties and processes, M.S. thesis, Univ. of Alaska, Fairbanks, Fairbanks, Alaska, 1986.

Liston, G. E., and M. Sturm, A snow-transport model for complex terrain, *J. Glaciol.*, *44*, 498–516, 1998.

Lowrance, R., R. Todd, J. Fail, O. Hendrickson, R. Leonard, and L. Asmussen, Riparian forests as nutrient filters in agricultural watersheds, *BioScience*, *34*, 374–377, 1984.

Lynch-Stieglitz, M., The development and validation of a simple snow model for the GISS GCM, *J. Clim.*, *7*, 1842–1855, 1994.

McHale, M. R., M. J. Mitchell, J. J. McDonnell, and C. P. Cirmo, Nitrogen solutes in an Adirondack forested watershed: Importance of dissolved organic nitrogen, *Biogeochemistry*, *48*, 165–184, 2000.

McNamara, J. P., D. L. Kane, and L. D. Hinzman, Hydrograph separations in an arctic watershed using mixing model and graphical techniques, *Water Resour. Res.*, *33*, 1707–1719, 1997.

McNamara, J. P., D. L. Kane, and L. D. Hinzman, An analysis of stream flow hydrology in an Arctic drainage basin: A nested watershed approach, *J. Hydrol.*, *206*, 39–57, 1998.

McNamara, J. P., D. L. Kane, and L. D. Hinzman, An analysis of arctic channel networks using a digital elevation model, *Geomorphology*, *29*, 339–353, 1999.

Meybeck, M., Carbon, nitrogen, and phosphorus transport by world rivers, *Am. J. Sci.*, *282*, 401–450, 1982.

Michaelson, G. J., C. L. Ping, G. W. Kling, and J. E. Hobbie, The character and bioactivity of dissolved organic matter at thaw and in the spring runoff waters of the arctic tundra, North Slope, Alaska, *J. Geophys. Res.*, *103*, 28,939–28,946, 1998.

Michalzik, B., K. Kalbitz, J. H. Park, S. Solinger, and E. Matzner, Fluxes and concentrations of dissolved organic carbon and nitrogen: A synthesis for temperate forests, *Biogeochemistry*, *52*, 173–205, 2001.

Moldan, F., and R. F. Wright, Changes in runoff chemistry after five years of N addition to a forested catchment at Gardsjon, Sweden, *For. Ecol. Manage.*, *101*, 187–197, 1998.

Nelson, F. E., K. M. Hinkel, N. I. Shiklomanov, G. R. Mueller, L. L. Miller, and D. A. Walker, Active-layer thickness in north central Alaska: Systematic sampling, scale, and spatial autocorrelation, *J. Geophys. Res.*, *103*, 28,963–28,974, 1998.

Peterjohn, W. T., and D. L. Correll, Nutrient dynamics in an agricultural watershed: Observations on the role of a riparian forest, *Ecology*, *65*, 1466–1475, 1984.

Scanlon, T. M., J. P. Raffensperger, and G. M. Hornberger, Modeling transport of dissolved silica in a forested headwater catchment: Implications for defining the hydrochemical response of observed flow pathways, *Water Resour. Res.*, *37*, 1071–1082, 2001.
Schiff, S. L., R. Aravena, S. E. Trumbore, and P. J. Dillon, Dissolved organic-carbon cycling in forested watersheds: A carbon isotope approach, *Water Resour. Res.*, *26*, 2949–2957, 1990.
Schiff, S. L., R. Aravena, S. E. Trumbore, M. J. Hinton, R. Elgood, and P. J. Dillon, Export of DOC from forested catchments on the Precambrian Shield of Central Ontario: Clues from C-13 and C-14, *Biogeochemistry*, *36*, 43–65, 1997.
Scott, C. A., M. F. Walter, G. N. Nagle, M. T. Walter, N. V. Sierra, and E. S. Brooks, Residual phosphorus in runoff from successional forest on abandoned agricultural land: 1. Biogeochemical and hydrological processes, *Biogeochemistry*, *55*, 293–309, 2001.
Shaman, J., M. Stieglitz, V. Engel, R. Koster, and C. Stark, Representation of subsurface storm flow and a more responsive water table in a TOPMODEL-based hydrology model, *Water Resour. Res.*, *38*(8), 1156, doi:10.1029/2001WR000636, 2002.
Shaver, G. R., K. J. Nadelhoffer, and A. E. Giblin, Biogeochemical diversity and element transport in a heterogeneous landscape, the North Slope of Alaska, in *Quantitative Methods in Landscape Ecology*, edited by M. G. Turner and R. H. Gardner, pp. 105–126, Springer-Verlag, New York, 1991.
Stieglitz, M., D. Rind, J. Famiglietti, and C. Rosenzweig, An efficient approach to modeling the topographic control of surface hydrology for regional and global climate modeling, *J. Clim.*, *10*, 118–137, 1997.
Stieglitz, M., J. Hobbie, A. Giblin, and G. Kling, Hydrologic modeling of an arctic tundra watershed: Toward pan-arctic predictions, *J. Geophys. Res.*, *104*, 27,507–27,518, 1999.
Stieglitz, M., A. Giblin, J. Hobbie, M. Williams, and G. Kling, Simulating the effects of climate change and climate variability on carbon dynamics in Arctic tundra, *Global Biogeochem. Cycles*, *14*, 1123–1136, 2000.
Stieglitz, M., A. Ducharne, R. Koster, and M. Suarez, The impact of detailed snow physics on the simulation of snow cover and subsurface thermodynamics at continental scales, *J. Hydrometeorol.*, *2*, 228–242, 2001.
Tague, C. L., and L. E. Band, Evaluating explicit and implicit routing for watershed hydro-ecological models of forest hydrology at the small catchment scale, *Hydrol. Processes*, *15*, 1415–1439, 2001.
Tipping, E., C. Woof, E. Rigg, A. F. Harrison, P. Ineson, K. Taylor, D. Benham, J. Poskitt, A. P. Rowland, R. Bol, and D. D. Harkness, Climatic influences on the leaching of dissolved organic matter from upland UK Moorland soils, investigated by a field manipulation experiment, *Environ. Int.*, *25*, 83–95, 1999.
Van der Peijl, M. J., and J. T. A. Verhoeven, Carbon, nitrogen and phosphorus cycling in river marginal wetlands: A model examination of landscape geochemical flows, *Biogeochemistry*, *50*, 45–71, 2000.
Vitousek, P. M., J. D. Aber, R. W. Howarth, G. E. Likens, P. A. Matson, D. W. Schindler, W. H. Schlesinger, and D. G. Tilman, Human alteration of the global nitrogen cycle: Sources and consequences, *Ecol. Appl.*, *7*, 737–750, 1997.
Walker, D. A., and M. D. Walker, Terrain and vegetation of the Imnavait Creek watershed, in *Landscape Function and Disturbance in Arctic Tundra*, edited by J. F. Reynolds and J. D. Tenhunen, pp. 73–108, Springer-Verlag, New York, 1996.
Watson, F. G. R., R. B. Grayson, R. A. Vertessy, and T. A. McMahon, Large-scale distribution modelling and the utility of detailed ground data, *Hydrol. Processes*, *12*, 873–888, 1998.
Webb, T. H., and S. J. Burgham, Soil-landscape relationships of downlands soils formed from loess, eastern South Island, New Zealand, *Austral. J. Soil Res.*, *35*, 827–842, 1997.
Wigmosta, M. S., and D. P. Lettenmaier, A comparison of simplified methods for routing topographically driven subsurface flow, *Water Resour. Res.*, *35*, 255–264, 1999.
Wigmosta, M. S., L. W. Vail, and D. P. Lettenmaier, A distributed hydrology-vegetation model for complex terrain, *Water Resour. Res.*, *30*, 1665–1679, 1994.
Yanagisawa, N., and N. Fujita, Different distribution patterns of woody species on a slope in relation to vertical root distribution and dynamics of soil moisture profiles, *Ecol. Res.*, *14*, 165–177, 1999.
Yenko, M., Determination of cold season streamflow source areas using two- and three- component hydrograph separation models in the semi-arid Dry Creek Watershed, Master's thesis, Dep. of Geosci., Boise State Univ., Boise, Idaho, 2003.

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