



Nitrogen dynamics in the hyporheic zone of a forested stream during a small storm, Hokkaido, Japan

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Abstract. Water and dissolved nitrogen flows through the hyporheic zone of a 3rd-order mountain stream in Hokkaido, northern Japan were measured during a small storm in August 1997. A network of wells was established to measure water table elevations and to collect water samples to analyze dissolved nitrogen concentrations. Hydraulic conductivity and the depth to bedrock were surveyed. We parameterized the groundwater flow model, MODFLOW, to quantify subsurface flows of both stream water and soil water through the hyporheic zone. MODFLOW simulations suggest that soil water inflow from the adjacent hill slope increased by 1.7-fold during a small storm. Dissolved organic nitrogen (DON) and ammonium (NH_4^+) in soil water from the hill slope were the dominant nitrogen inputs to the riparian zone. DON was consumed via mineralization to NH_4^+ in the hyporheic zone. NH_4^+ was the dominant nitrogen species in the subsurface, and showed a net release during both base and storm flow. Nitrate appeared to be lost to denitrification or immobilized by microorganisms and/or vegetation in the riparian zone. Our results indicated that the riparian and hyporheic system was a net source of NH_4^+ to the stream.

Introduction

Nitrate (NO_3^-), ammonium (NH_4^+) and dissolved organic nitrogen (DON) lost from the soil–vegetation system in terrestrial portions of watersheds are transported to streams via subsurface flows of soil water and groundwater. However, this water must first flow through riparian zones which are often critical locations for biogeochemical transformations of nitrogen, and thus can regulate both form and concentration of nitrogen found in stream water. Riparian zones may also influence the amounts of nitrogen lost from forested catchments (Cirimo and McDonnell 1997). However, the effect of riparian zones on stream and catchment

nitrogen cycles depends on the biogeochemical environment within the saturated zone beneath the riparian zone (Triska et al. 1989; McDowell et al. 1992; Hedin et al. 1998). That biogeochemical environment, in turn, is affected by: (1) the mixing of waters from different sources. These may include groundwater discharged from local or regional aquifers, soil water draining from adjacent hillslopes and stream water, sometimes also called advected-channel water (Triska et al. 1989; Wondzell and Swanson 1996a,b); (2) the type and amount of dissolved organic compounds in each water source; (3) the composition of micro-organism communities (Vervier et al. 2002); and (4) residence time of water in the riparian zone (Findlay 1995). These factors influence the transformation and movement of dissolved nitrogen through riparian zones and determine if riparian zones will be net-sources or net-sinks of nitrogen for stream ecosystems. However, due to the uncertainty in quantifying the amounts of water flowing through riparian zones, it is difficult to estimate fluxes of dissolved constituents. Therefore, the net effect of riparian nitrogen dynamics on stream ecosystems remains unclear. In this study, we combined measurements of dissolved nitrogen concentrations with numerical modeling of subsurface flows to quantify the fluxes of dissolved nitrogen and to clarify the role of the hyporheic zone in the nitrogen dynamics of a forest stream in Hokkaido, Japan.

Materials and methods

Site description

This study was conducted at the Karuushinai River, located in the Hokkaido University's Uryu Experimental Forest (UREF), northern Hokkaido (44°22'N; 142°15'E), Japan (Figure 1). The nearest meteorological station is located approximately 7 km from the study site, at the headquarters of the UREF. Climatic records are available from 1973 to 1997. Precipitation is relatively evenly distributed throughout year, with approximately 50% falling as rain between May and November and the remainder as snow (water equivalent). Annual precipitation averages 1375 mm year⁻¹, but has ranged between 1120 and 2200 mm year⁻¹ over the period of record. Winters are cold and snowy while summers are hot and humid. Annual average temperature is 2.5 °C, and has ranged between 0.3 and 4.7 °C over the period of record. The forest is dominated by deciduous species, especially *Ulmus davidiana* var. *japonica*, and *Fraxinus mandshurica* var. *japonica*, although coniferous species are present. *Alnus hirsuta* and *Salix* spp. dominate the riparian forest. The understory vegetation of both the uplands and the riparian zone is dominated by the dwarf bamboo, *Sasa senanensis*.

The Karuushinai River is a 3rd-order stream draining a 10.5 km² watershed. At the study site, the channel is 4–5-m wide during low flow (Figure 2) and the active channel is approximately 30 m wide. The active channel is bounded by steep, well defined banks approximately 3-m high cut into the toe of the hillslope and alluvial fan on the right side (facing downstream) of the valley floor, and cut into a high

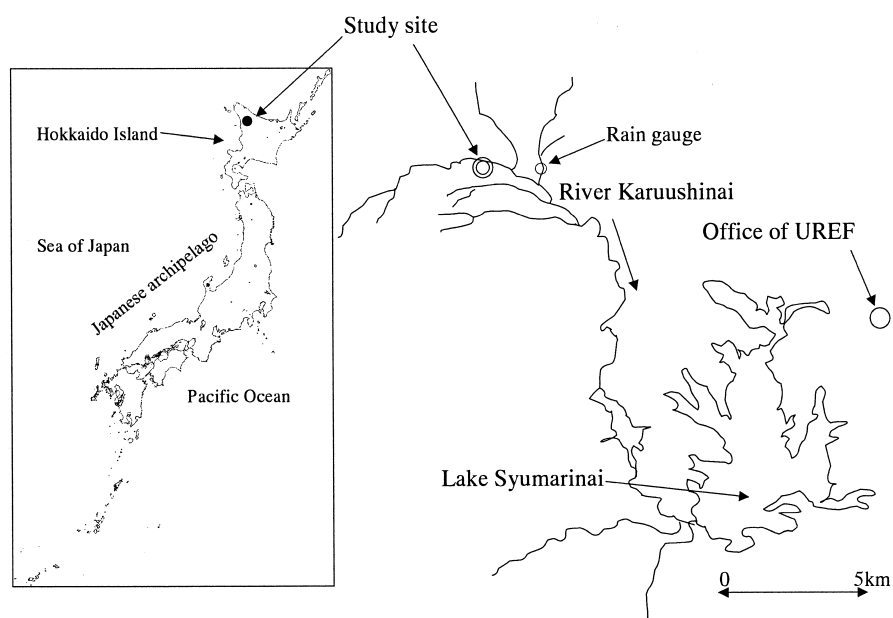


Figure 1. Location of study site.

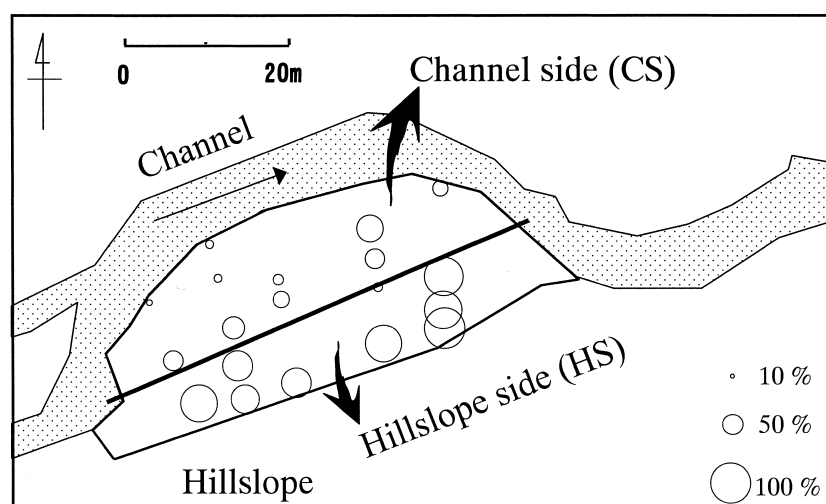


Figure 2. Relative contribution of subsurface water from the adjacent hillslope to the groundwater estimated by the single mixing model (Eqs. (1–3)) and the definitions of channel and hillslope sites in the hyporheic study zone. See text for details.

terrace on the left side. At low flow, the wetted channel meanders within the active channel. The study site is on a 60-m long, 20-m wide point bar within one of these meander bends. The depth of sediment forming the gravel bar is quite shallow. The local bedrock, a Tertiary andesite, is exposed at some locations in the streambed. Elevations of bedrock exposures were mapped and surveyed at 24 locations. Also, an electrical resistivity survey was conducted at 12 locations (Shin 1989), using Wenner's configuration, to identify the depth to the bedrock under the gravel bar. The distance from the ridge to the valley floor at the studied plots ranged approximately 400–500 m.

Well networks and field measurements

Twenty-five wells were installed on the gravel bar in the summer of 1995. The well casings were made from 8.8 cm diameter PVC pipe screened by drilling holes (0.5 cm diameter) on four sides at 10-cm intervals over the bottom 50–100 cm. Wherever possible, the well casings were placed in holes dug at least 50 cm below the baseflow water table elevation. Well locations were mapped and the elevations of both the wellhead and the ground level were surveyed.

Saturated hydraulic conductivities (K) were estimated for each well using falling head slug tests (Bouwer and Rice 1976). This method is appropriate for partially-penetrating, small-diameter wells in unconfined, heterogeneous, anisotropic aquifers (Dawson and Istock 1991), but the estimated values of K apply only to a small area immediately adjacent to each well (Bouwer 1989). The geometric mean of hydraulic conductivity was $1.4 \times 10^{-5} \text{ cm s}^{-1}$, and ranged from 3.6×10^{-7} to $1.4 \times 10^{-4} \text{ cm s}^{-1}$.

Water table elevations were measured and water samples were collected from wells during a small storm from 1 to 3 August 1997, to measure storm-related changes in subsurface water flows and dissolved nitrogen concentrations. Water table elevations were measured manually at 6–24-h intervals. Stream stage at the study site was measured automatically at 1-h intervals using a pressure transducer and a KADEC-MIZU[®] data logger (Kona Inc.¹, Sapporo, Japan). Discharge was calculated using an empirical relationship between stage and discharge developed from estimates of discharge made using a flow meter. Water samples were also collected from wells at 24 h intervals using a hand-powered suction pump and grab samples were collected from the stream at the same time.

Chemical analysis

Dissolved oxygen (DO) concentration and redox potential (Eh) were measured on a single date in July 1997. Well diameters were too small to measure DO and Eh *in*

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situ, so measurements were made from water samples immediately after they were collected. However, the hand pump used for collecting samples did not prevent contact with air, so we probably over estimated both DO and Eh. Measurements were made using a DO meter (TOA Inc.¹, Tokyo, Japan) and Eh meter (Horiba Inc.¹, Kyoto, Japan), respectively.

Water samples for nutrient analysis were immediately transported to the laboratory where they were refrigerated ($<4^{\circ}\text{C}$) until analyzed. We measured concentrations of dissolved total nitrogen (TDN), NO_3^- and NH_4^+ in samples collected from the stream and from observation wells. Water samples were filtered (TOA Inc.¹ No. 5C, Tokyo, Japan) and TDN concentrations was measured using the ultra-violet absorption method (220 nm) following wet digestion with a mixture of potassium peroxodisulfate and sodium hydroxide in an autoclave (The Japan Society for Analytical Chemistry, Hokkaido Branch 1994). NO_3^- and NH_4^+ concentrations were measured using the capillary electrophoresis method (Waters Inc.¹, Tokyo, Japan; Jandik and Bonn, 1993) after filtering the water through a membrane filter ($0.2\mu\text{m}$). The detection limits of the NO_3^- and NH_4^+ by the capillary electrophoresis method were 0.079 and 0.11mgNL^{-1} , respectively. DON concentration was calculated by subtracting the sum of NO_3^- and NH_4^+ from TDN in each sample. If calculated concentration of DON was negative, it was regarded as being zero.

Definition of hill-side versus channel-side compartments

Water samples collected from wells at noon on 1 August were analyzed for Cl^- concentrations using the capillary electrophoresis method (Waters Inc., Tokyo, Japan). We used the simple mixing model (Eqs. (1) and (2)), with chloride (Cl^-) as an inert tracer, to estimate the relative contribution of the hillslope water to the subsurface, and to characterize mixing of hillslope- and stream-source waters in each compartment:

$$\alpha_{\text{stream}} + \alpha_{\text{hillslope}} = 100, \quad (1)$$

$$\alpha_{\text{stream}} \cdot C_{\text{stream}} + \alpha_{\text{hillslope}} \cdot C_{\text{hillslope}} = C_i, \quad (2)$$

where: α_{stream} , $\alpha_{\text{hillslope}}$ are relative contributions (%) of stream water and hillslope water, respectively; C_{stream} , $C_{\text{hillslope}}$ are the Cl^- concentrations in stream water and soil water (10.9 and 4.9mgL^{-1} , respectively); and C_i is the concentration of Cl^- in the water sampled from each well. The long residence time of stream water supplied from the upper watershed, where wetlands are abundant, might allow evapotranspiration to concentrate Cl^- in the water, whereas the soil water from the adjacent hillslope has relatively short residence time so that Cl^- concentration reflects the ambient concentration in rainfall.

We divided the studied gravel bar into two compartments: a channel-side compartment and a hill-side compartment (Figure 2) based on the mixing of different source waters, so that areas with wells dominated by hillslope-source water ($>50\%$)

were grouped into a hill-side compartment. Areas dominated by stream-source water were grouped into a channel-side compartment. The hill-side compartment appears to have been a former back channel. It has lower surface elevation, the water table is closer to the surface, and surface soils were usually wetter than the adjacent channel-side compartment.

Groundwater modeling

Groundwater flow simulations were run using MODFLOW (McDonald and Harbaugh 1988). MODFLOW enables numerical evaluation of the partial differential equations for groundwater flow (Eq. (3)) using a block-centered finite difference method:

$$\frac{\partial}{\partial x} \left(K_x \frac{\partial h}{\partial x} \right) + \frac{\partial}{\partial y} \left(K_y \frac{\partial h}{\partial y} \right) + \frac{\partial}{\partial z} \left(K_z \frac{\partial h}{\partial z} \right) = S_s \frac{\partial h}{\partial t} - R, \quad (3)$$

where K_x , K_y , and K_z are hydraulic conductivities; h is the hydraulic head; S_s is the specific storage; R is the source/sink term; x , y and z are spatial directions; and t is the time step. Since we assume the aquifer was isotropic, we used the same K values for K_x , K_y and K_z in our simulation.

We used a 2-dimensional steady-state model because the wells were shallow so that calibration data were available only for the upper 1–2 m of the aquifer. To apply the 2-dimensional model, it was necessary to set boundary conditions to define the model domain. We assumed that there was no leakage through the bottom of the modeled aquifer because the bedrock would tend to restrict vertical flow to or from deeper layers. Therefore, the bottom of the model domain was a no-flow boundary. We assumed that subsurface flow in the valley converges (or diverges) along the channel. Hence, the stream was defined by a specified head cells located at the center of the channel, which defined the left boundary to the model domain. Upstream and downstream boundaries were defined to include a geomorphic surface consisting of gravel bar and floodplain. The right side of the model domain was the border between the floodplain and adjacent hillslope/alluvial fan. Water fluxes were calculated for each compartment, namely from the stream into the stream-side compartment and from the hillslope into the hill-side compartment. We assumed that evapotranspiration losses could be ignored over the short period of time during which we made observations (30 h). All models were run at state model, even during the storm event, during which we changed the head in each specified-head boundary cell to match changes in stream water heights observed during the storm.

Saturated hydraulic conductivity varied substantially among wells at the site and values estimated from slug tests pertain only to a small area surrounding each well. We used a homogeneous model for our initial simulations with the site-averaged hydraulic conductivity ($1.4 \times 10^{-5} \text{ cm s}^{-1}$). The water table elevations predicted from this model matched elevations observed in wells on the hillslope-side of the gravel bar, but did not match elevations observed in near-stream locations. Some of

the directions of the estimated water flow did not match the observed gradient of the groundwater table. Therefore, we made repeated calibration runs, changing the saturated hydraulic conductivity in each run, until the model solution provided a reasonable match to the water table elevations observed in the well network. The best-fit model was achieved by dividing the model domain into two zones with different saturated hydraulic conductivity: $5.4 \times 10^{-5} \text{ cm s}^{-1}$ in the hillslope-side and $2.4 \times 10^{-5} \text{ cm s}^{-1}$ in the channel side. These values were averaged within each site. The determined coefficient (R^2) between the observed and modeled values of the groundwater table was 0.86. Once calibrated, the model was used to estimate fluxes of both advected-channel water and groundwater through the subsurface. We used the effective porosity as 0.01 (Morris and Johnson 1967).

Budget calculation

Flows across the boundary of each compartment, estimated from the MODFLOW flow-budget module, were summed over each sampling period to calculate the total subsurface flux of water through each compartment. Dissolved nitrogen fluxes (ng N h^{-1}) for each sampling period were obtained by multiplying the mean nitrogen (DON , NO_3^- and NH_4^+) concentration measured for each water source (hillslope, stream, hill-side and channel-side compartment) by the total flow of water across the corresponding boundary of that compartment. To estimate nitrogen fluxes from the hillslope, we averaged nitrogen concentrations from the wells located at the hillslope end of each well transect. The storage of water and nitrogen in each compartment was calculated as the difference between inputs and outputs. For example, a negative value of storage change means that the output exceeds the input and the compartment was a net source of water or nitrogen during that sampling period.

Results

Differences between hill-side and channel-side compartments

During a low flow (12:00 on 1 August 1997, Figure 3), Cl^- concentrations in the stream water at the upper and lower reaches were 10.8 and 10.9 mg L^{-1} , respectively. In contrast, Cl^- concentrations in well-water samples varied from 3.9 to 9.9 mg L^{-1} and were lowest closest to the hillslope. We used Cl^- concentration in stream water to define one end member and minimum Cl^- concentration observed in wells closest to the hillslope 3.9 mg L^{-1} to define the hillslope end member. The relative proportions of hillslope water in the channel-side compartment and in the hill-side compartment (Figure 2) were 0.36 ± 0.17 and 0.86 ± 0.11 , respectively. The mean DO concentrations in the channel-side and hill-side compartments under baseflow conditions in July 1997 ranged from 2 to 3 mg L^{-1} (Figure 4). The mean Eh at these sites ranged from 350 to 450 mV (Figure 4). Although these differences

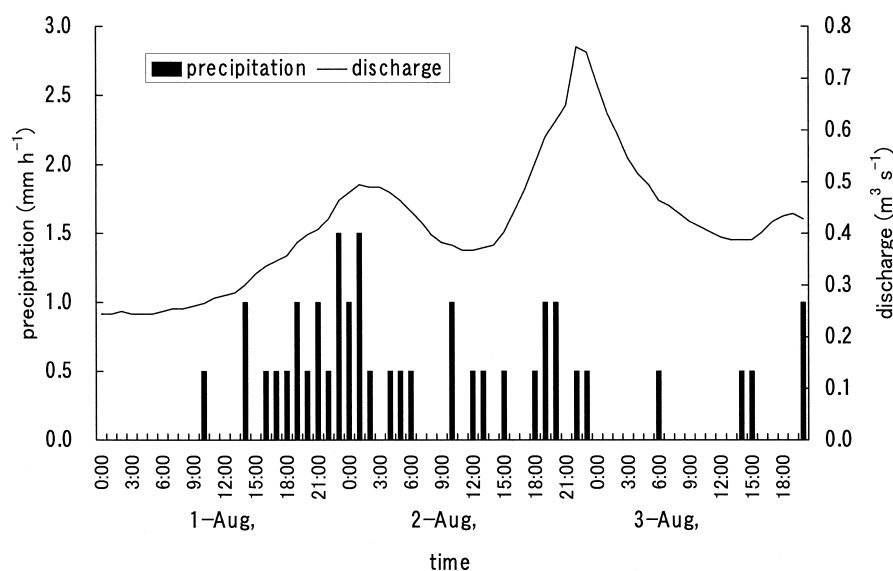


Figure 3. Temporal fluctuation of channel discharge and hourly precipitation during a small storm from 1–3 August, 1997.

were not statistically significant, DO and Eh in the hill-side compartment tended to be lower than those in the channel-side compartment. Also, DO concentration in the subsurface was unsaturated.

Changes in hydrologic flow paths and water chemistry during the storm

The total amount of precipitation received between 1 and 3 August 1997 was 20.5 mm. Rainfall started at 10:00 on 1 August. Light rain fell steadily throughout the day, with a few hours of more intense rainfall near midnight. Stream discharge increased noticeably late on 1 August, reaching an initial peak at 02:00 on 2 August (Figure 3). Late in the evening of 2 August, stream discharge increased sharply, with peak discharge for this storm occurring at 23:00, 2 August.

Water table elevations across most of the gravel bar increased only slightly, in response to the storm. However increased input of hillslope water near the head of the gravel bar resulted in sharply increased water table elevations in the early morning of 2 August (Figure 5), after which water table elevations decreased and by 3 August elevations across the gravel bar were quite similar to the pre-storm conditions. Peak discharge and the highest elevation of water table did not correspond to the peak period of rainfall, suggesting that the heavy rain in upper parts of the watershed accounted for the peak flow.

Nitrogen concentrations in the stream water sampled on 1 August were very low, but both NO_3^- and DON concentrations had increased by 12:00 on 2 August.

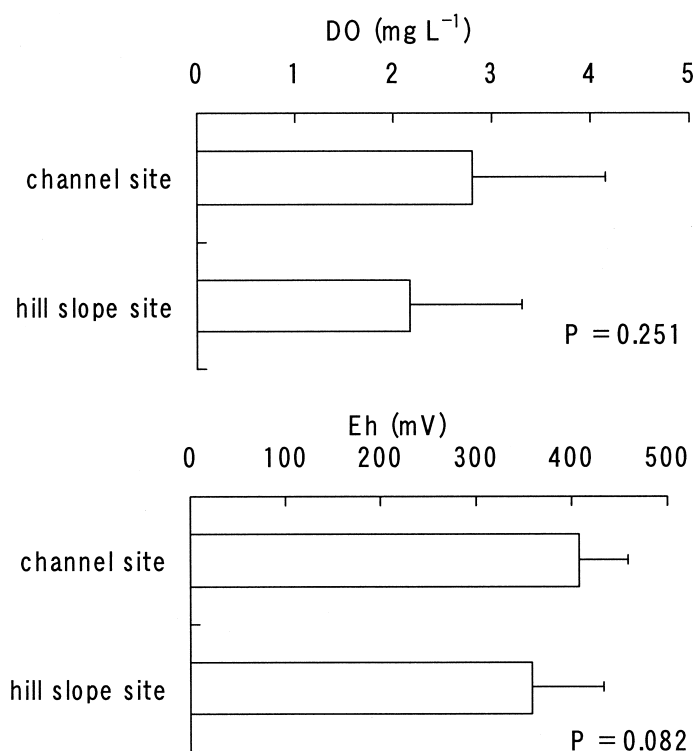


Figure 4. DO concentration (upper) and redox potential (Eh, lower) in groundwater at each site in the base flow. The bar represents standard deviation.

Dissolved nitrogen (NH_4^+ , NO_3^- and DON) concentrations were spatially heterogeneous within the gravel bar. Within the channel-side compartment, NH_4^+ was the dominant form of dissolved nitrogen at noon on 1 August and did not change substantially throughout the storm (Figure 6). NH_4^+ was also the dominant form of dissolved nitrogen within the hill-side compartment, although DON concentration had increased markedly by noon on 2 August. Water samples collected from wells located at the toe of the hillslope (FH) showed the same tendency as did other wells in hill-side compartment.

Water and nitrogen fluxes

Using the MODFLOW model, we calculated the water-flow budget during each measurement period from 1 to 2 August, 1997. Water supply to the gravel bar was dominated by stream water inputs during all measurement periods, however, stream water inputs decreased from $16.3 \text{ cm}^3 \text{ s}^{-1}$ on 1 August to $15.0 \text{ cm}^3 \text{ s}^{-1}$ on 2 August (Figure 7). In contrast, subsurface inflows from the hillslope increased during the storm (12:00, 2 August). Total outflows from the gravel bar to the stream were

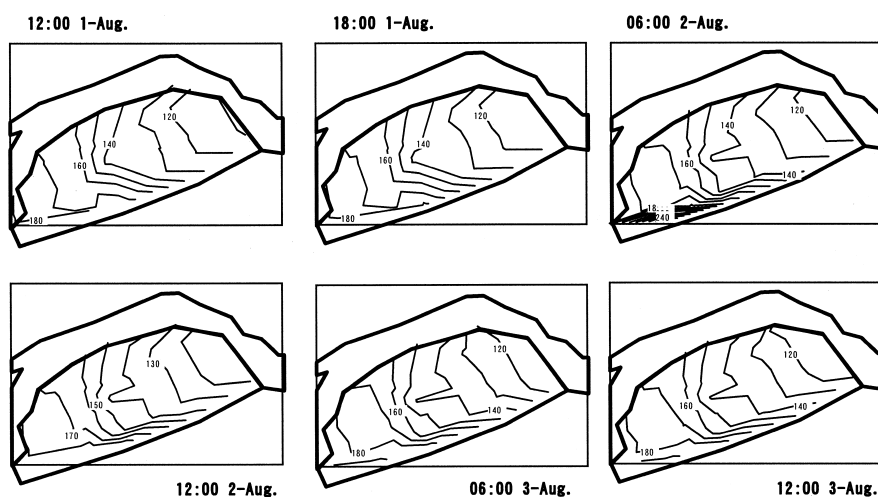


Figure 5. Temporal and spatial fluctuations in the groundwater table from 1–3 August, 1997. Contour line was drawn on the basis of actual observations of the groundwater table in each well at each time.

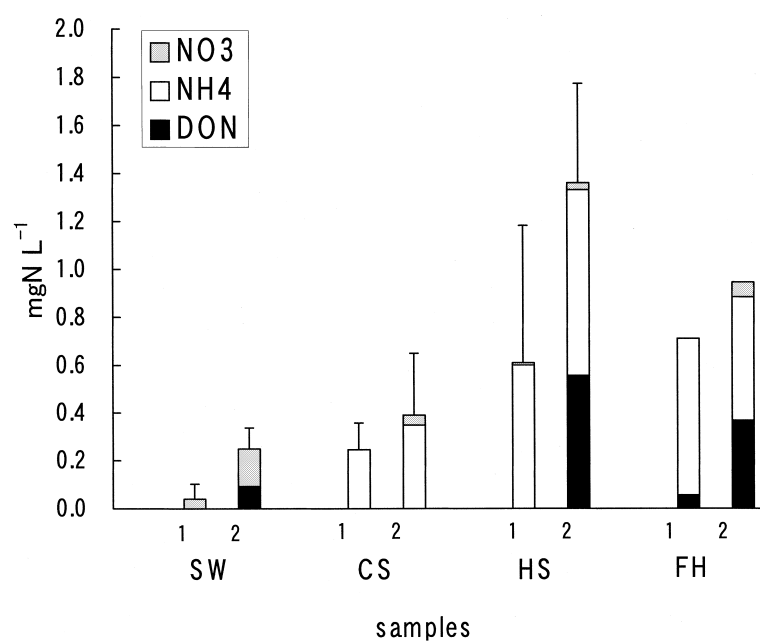


Figure 6. Mean concentrations of NO₃, NH₄ and DON in groundwater, subsurface water and channel water at noon on 1 August (1) and 2 August (2), 1997. CS: groundwater at the channel-side, HS: groundwater at the hillslope-side, FH: subsurface water from the adjacent hillslope, SW: channel water. The bar represents standard deviation.

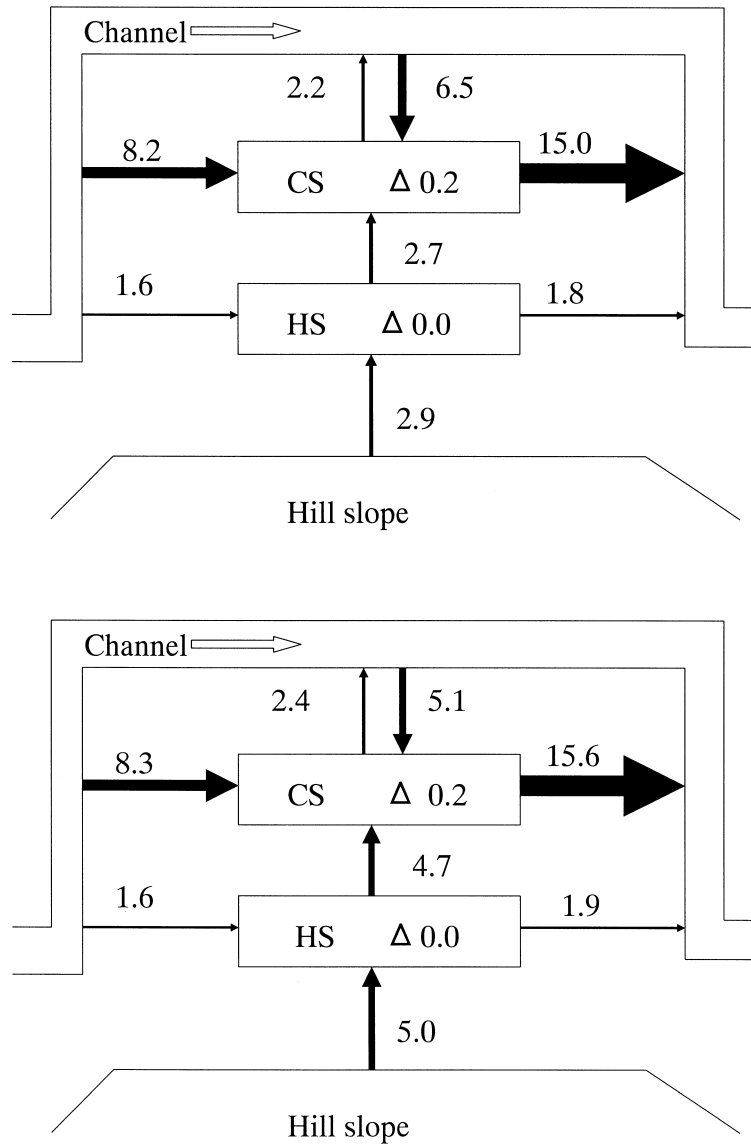


Figure 7. Water flow ($\text{cm}^3 \text{s}^{-1}$) at 12:00 on 1 August (upper) and 2 August (lower), 1997. The values of delta in each box represent the each budget of water (input minus output). CS: channel-side, HS: hillslope-side.

relatively unchanged between the two periods. Water inflow from the hillslope was the major input to the hill-side compartment, and subsurface water flowed diagonally through this compartment and into the channel-side compartment. Our

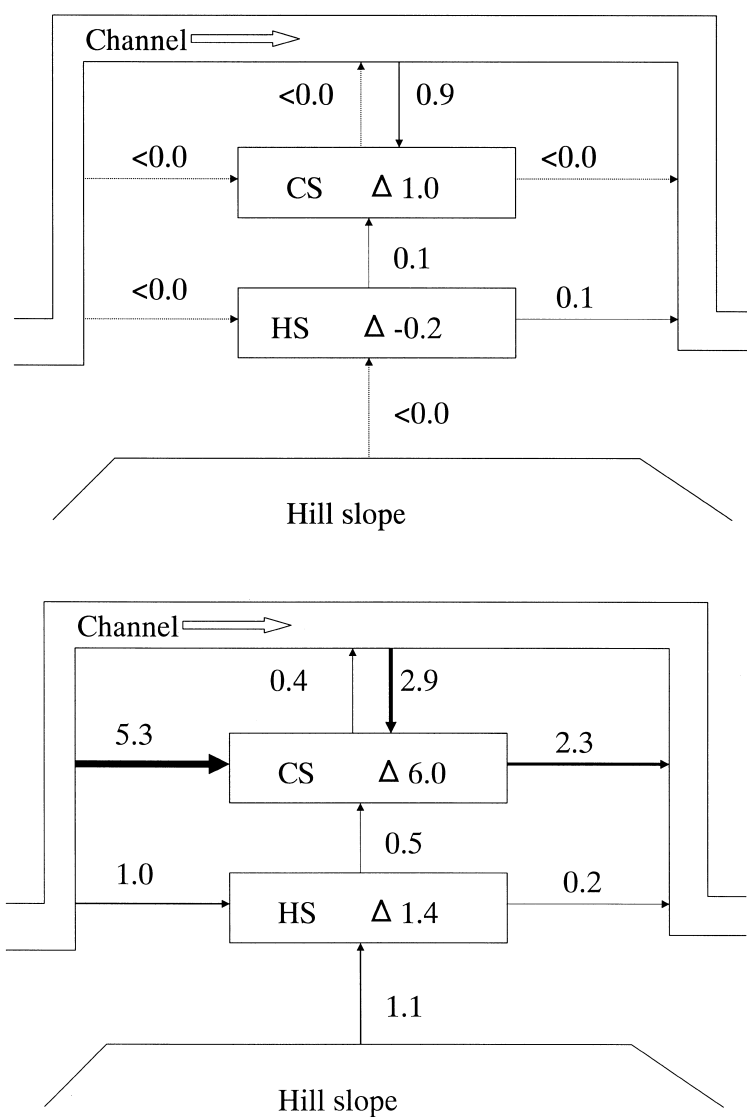


Figure 8. NO_3^- flow (ng N h^{-1}) at 12:00 on 1 August (upper) and 2 August (lower), 1997. The values of delta in each box represent each budget of NO_3^- (input minus output). CS: channel-side, HS: hillside-side.

budget calculations suggest that inflows balanced outflows from the hill-side compartment during both measurement periods (Figure 7). Similarly, under base-flow conditions, inflows to the channel-side compartment balanced outflows.

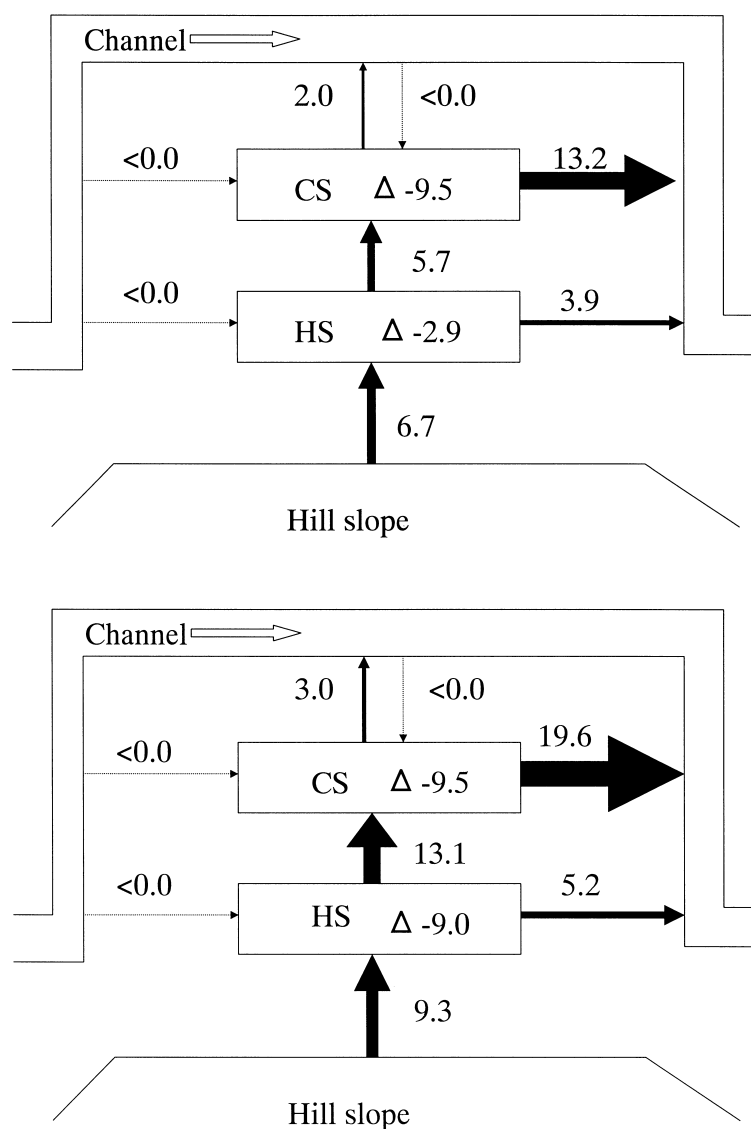


Figure 9. NH_4^+ flow (ng N h^{-1}) at 12:00 on 1 August (upper) and 2 August (lower), 1997. The values of delta in each box represent each budget of the NH_4^+ (input minus output). CS: channel-side, HS: hillslope-side.

Prior to peak discharge, all NO_3^- fluxes were small ($<0.1 \times 10^{-6} \text{ ng N h}^{-1}$), however, inputs of NO_3^- from both the channel and hillslope had increased substantially by noon, 2 August (Figure 8). The net storage of NO_3^- in both the channel-side and hill-side compartments was positive, indicating a net loss of NO_3^- .

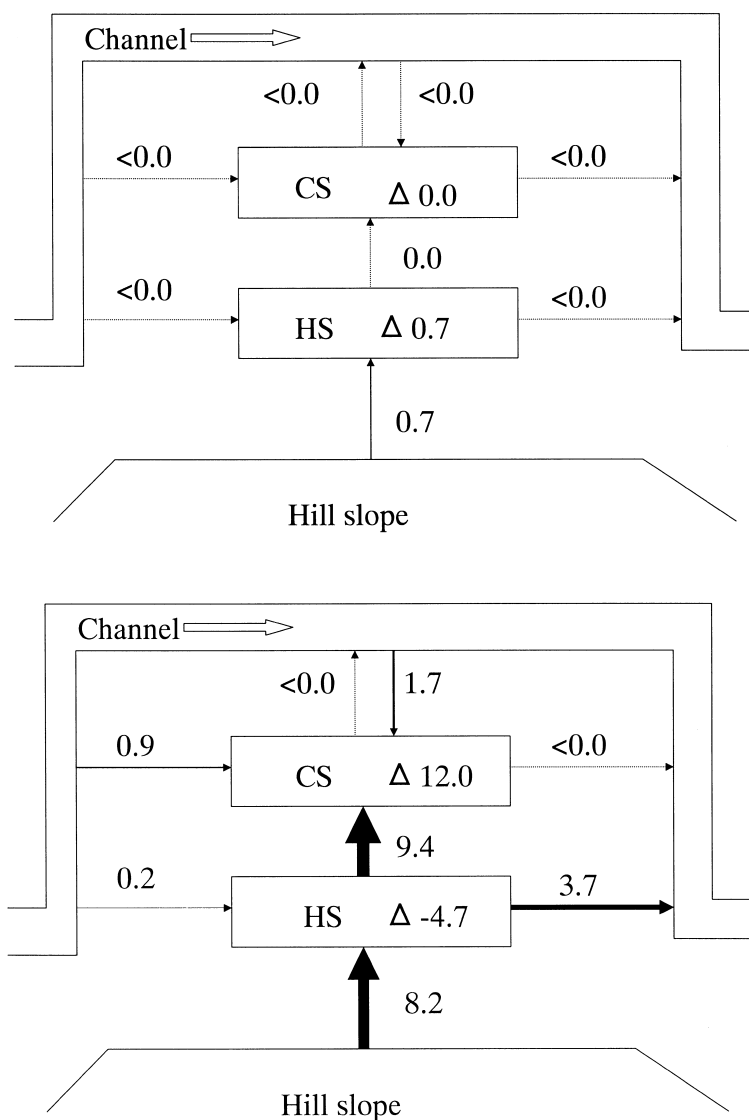


Figure 10. DON flow (ng N h⁻¹) at 12:00 on 1 August (upper) and 2 August (lower), 1997. The values of delta in each box represent each budget of NO₃⁻ (input minus output). CS: channel-side, HS: hillslope-side.

from water flowing through each compartment. Hillslope-source water was the dominant source of NH₄⁺ to the gravel bar and both the hill-side and channel-side compartments were net sources of NH₄⁺ to the stream (Figure 9). Flux of DON was negligible on 1 August, for all flow paths except from the hillslope. Flux of DON from the hillslope had increased greatly by 2 August, after peak discharge (Figure

10). Net storage of DON in the hill-side compartment after peak flow was negative, indicating that it was a net source of DON to both the stream and the channel-side compartment. Relatively little DON was transported from the channel-side compartment to the stream, indicating that this compartment was a net sink for DON. The relative contribution of each dissolved nitrogen species to the total nitrogen budget in each compartment after peak flow was different from that before peak flow. Before the peak flow, NH_4^+ was the dominant form of nitrogen transported through the system. The fluxes of DON and NO_3^- were substantially greater after the peak flow, with NO_3^- as the dominant input from stream water and DON the dominant input from the adjacent hill slope. Fluxes of nitrogen to the stream were still dominated by NH_4^+ after peak flow.

Discussion

Sensitivities and uncertainties of the model

We used a steady-state groundwater flow model to simulate subsurface flows during both base flow and storm periods. There are several potential large sources of uncertainty in the predictions of subsurface water fluxes, and these directly impact our estimates of nitrogen fluxes. First, flux estimates are linearly related to the saturated hydraulic conductivity, and the hydraulic conductivities measured from the well network ranged over three orders of magnitude. The model was calibrated to match observed water table elevations by varying the hydraulic conductivity in the stream-side and hill-side compartments (Eq. (3)), but without an independent measure of subsurface water fluxes, we cannot validate our model predictions. Further, predictions of both water table elevation and subsurface fluxes are sensitive to the spatial distribution of hydraulic conductivity values. For example, when the site-average hydraulic conductivities were applied to the entire model domain, predicted water table elevations averaged some 16% lower than those predicted from a simple heterogeneous model in which different hydraulic conductivities were assigned to each of the two compartments. More detail analyses of the spatial variability in hydraulic conductivities are needed to reduce the uncertainties of the model simulations.

During storms flows of water through shallow, unconfined aquifers respond dynamically to changes in inputs from rainfall, from hillslope drainage, and from changes in stream water elevations (Wondzell and Swanson 1996a). We did not simulate these dynamic responses. We only measured changes in water table elevations from the well network at 24-h intervals. In our opinion, these measurements do not provide sufficient temporal resolution to calibrate a transient simulation. Therefore, we used a steady-state simulation with boundary conditions during the storm even parameterized to measurements made at peak flow. We recognize that the use of the steady-state model runs to estimate subsurface fluxes during the storm creates an additional source of uncertainty in our flux predictions. As the water level of the stream goes up, or down, during a storm, it takes some time for the water to flow into the gravel bar, and during this time the height of the water in

the gravel bar will be out of equilibrium – that is, transient – with respect to the water level in the stream. Because the model was run as steady state for each time period, the fluxes are calculated as if subsurface flows were in equilibrium with the water level in the stream. Notwithstanding these uncertainties, our results show that fluxes of water and nitrogen through this gravel bar are dynamic, especially the inputs of DON which changed rapidly during a small storm.

Role of riparian zone on the nitrogen budgets

There were distinct spatial patterns in subsurface flow paths through the gravel bar. Hyporheic exchange flows dominated water fluxes in the channel-side compartment whereas fluxes through the hill-side compartment were dominated by soil-water inflows from the adjacent hillslope. Inputs of soil water increased during the small storm, but the biggest change in water flux was a substantial decrease in stream–water inputs to the channel-side compartment after peak flow. A similar pattern was observed at McRae Creek by Wondzell and Swanson (1996a). In their case, stored rain water raised the water table and decreased head gradients from the stream to the gravel bar, thereby decreasing hyporheic exchange flows. At McRae Creek, the hyporheic zone was isolated from inputs of hillslope or groundwater by preferential drainage through an abandoned channel so that changes in groundwater or soil-water inputs from the adjacent hillslopes had little effect on hyporheic exchange flow. At the Karuushinai study site, flow paths are not spatially isolated. Thus, increased inputs of soil water from the adjacent hillslopes appeared to raise the water table in the gravel bar, relative to the elevation of water in the stream channel, and thereby reduced the exchange flow of stream water through the hyporheic zone during this storm. In this study, precipitation inputs were not included in the model, but we calculate rainfall during this small storm would raise the water table by as much as 7 cm if all the precipitation (2.0 cm) was stored in the aquifer, assuming a soil porosity of 30%.

The gravel bar/hyporheic zone was a net source of nitrogen to the stream during both base flow and during a small storm (Figure 11). Both NO_3^- and DON were retained in the hyporheic zone and nitrogen exports were dominated by NH_4^+ . Many processes are related to the transformation of dissolved nitrogen in the saturated zone: mineralization, nitrification, denitrification and immobilization by microbes and uptake by vegetation. We summarized the dominant dissolved nitrogen dynamics based on our budget analysis of dissolved nitrogen during the base and peak flow periods (Figure 12).

Fluxes of NO_3^- were very small during base flow periods (Figure 12(a)) but NO_3^- input from both stream water and soil water from the adjacent hillslope increased during the storm. Nearly all the NO_3^- that entered the hyporheic zone via exchange flows during the storm was lost from the system, probably by denitrification and/or immobilization. Consequently, there virtually no NO_3^- was exported to the channel (Figure 12(b)).

DON was supplied to the hyporheic zone, mainly from soil water from the adjacent hillslope. Inputs were small under baseflow conditions prior to the storm,

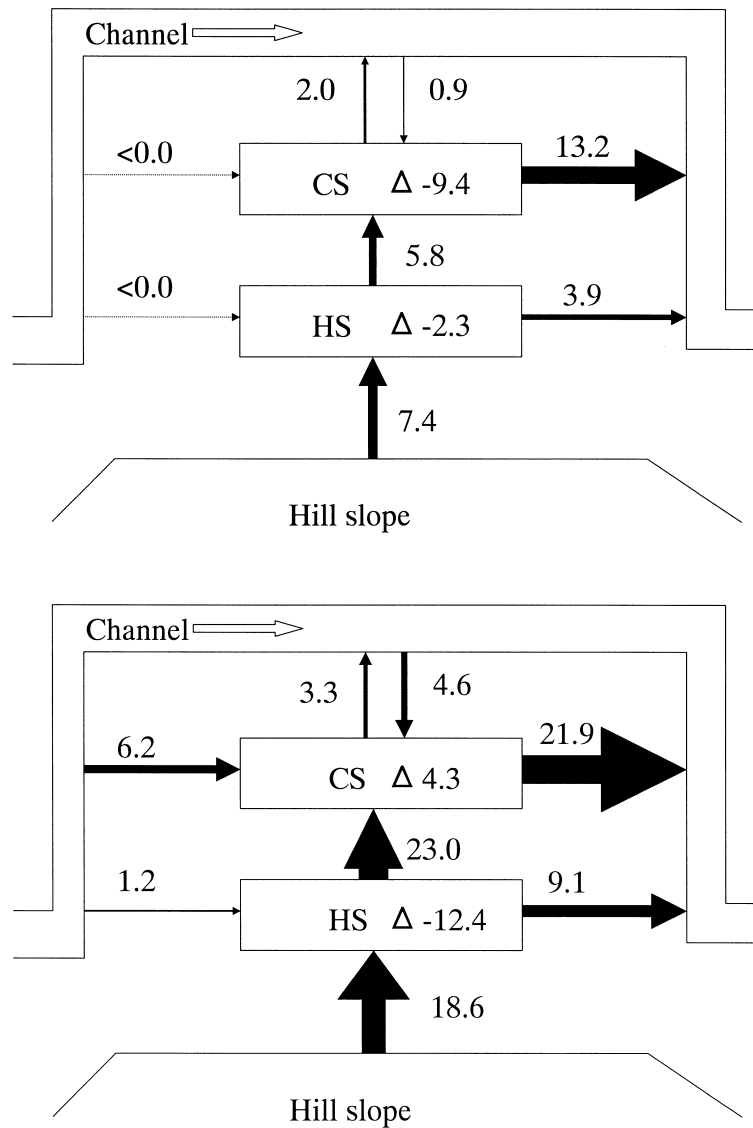
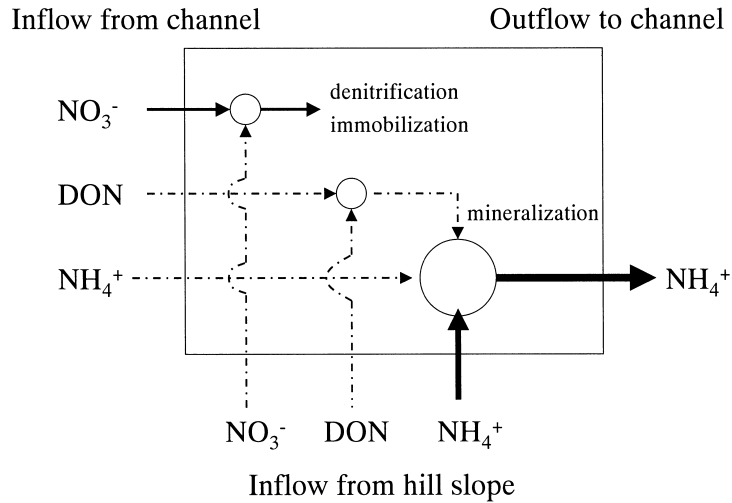
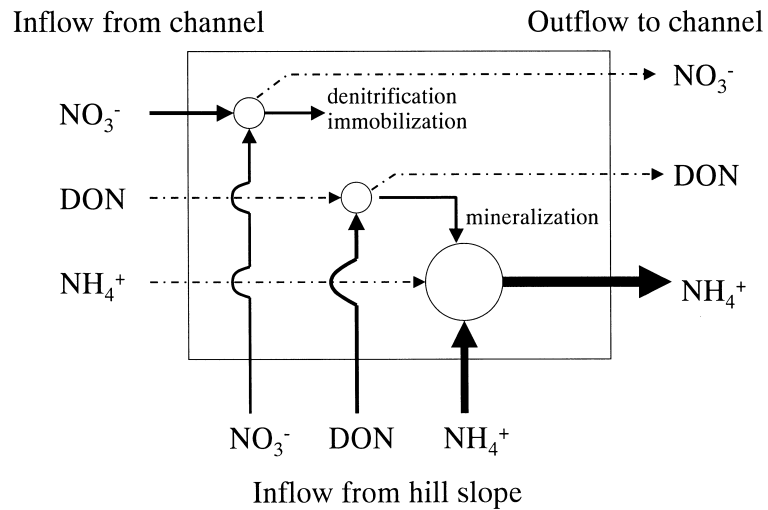


Figure 11. TDN flow (ng N h^{-1}) at 12:00 on 1 August (upper) and 2 August (lower), 1997. The values of delta in each box represent each budget of TDN (input minus output). CS: channel-side, HS: hillside-side.

and increased 10-fold during the storm. However, little DON was lost from the hyporheic zone even when hillslope inputs increased. Nitrogen export from the hyporheic zone to the stream was dominantly as NH_4^+ , suggesting that the DON supplied to the hyporheic zone in subsurface flows was mineralized. NH_4^+ was



(a) base flow period



(b) small storm period

Figure 12. Summary of nitrogen cycling in the hyporheic zone of a small stream. Transformations are inferred from net changes in nitrogen fluxes. Size of arrows represent relative magnitude of each flux, size of circles represents relative pool size.

present in the hyporheic zone in relatively high concentrations at all times, apparently maintained by DON mineralization and continuous NH_4^+ input via sub-surface inflows (Figure 12).

Exchange flows are often important in nitrogen transformations occurring in the hyporheic zone because they supply oxygen for nitrification (Triska et al. 1990, 1993; Wondzell and Swanson 1996b). However, nitrification was not a dominant process in this study. Rather, the soil was silty with relatively low saturated hydraulic conductivity, especially near the adjacent hillslope. Available oxygen appears to have been consumed rapidly, leading to conditions that favor mineralization of organic nitrogen to NH_4^+ , but inhibit the nitrification of the NH_4^+ .

NH_4^+ was exported from the hyporheic zone to the stream, with peak flux occurring during the storm. However, NH_4^+ concentration in stream water was negligible during both base flow and the storm, but both DON and NO_3^- increased during the storm. These data suggest that most of the nitrogen supplied to the stream during base flow is consumed by plant (algae) uptake or immobilization, although the amount that is first nitrified is unknown. In contrast, during the storm, total export of nitrogen to the stream must exceed rates of uptake and immobilization. The excess NH_4^+ must be nitrified, either as it passes through the stream bed or during in-channel processing.

During base flow and during the storm, NH_4^+ export from the hill-side compartment exceeded the combined influxes of NH_4^+ and DON (Figure 12). This suggests that there was an additional source of nitrogen, most likely either leached from the overlying riparian soils or input as fine-particulate organic nitrogen (FPON). During the storm, inputs of DON to the hill-side compartment appear to exceed mineralization rates and substantial quantities of DON are exported to the channel-side compartment (Figure 12(b)). Despite large inputs of DON to the channel-side compartment during the storm, we did not observe a measurable export of DON to the stream, suggesting that mineralization rates in the channel-side compartment were much higher than in the hill-side compartment. A more detailed study would be needed to confirm these speculations.

The riparian and hyporheic zones at this site are a net source of nitrogen to the stream. The source of nitrogen is probably the riparian soils, where the forest overstory is dominated by *A. hirsuta*, a nitrogen fixing species. Overall, the nitrogen budget was dominated by the throughput of NH_4^+ transported from hillslopes to the stream. The hyporheic zone also appears to be a location where DON dissolved in soil water draining from the adjacent hillslopes is transformed to NH_4^+ . The nitrogen budgets for the riparian and hyporheic zones suggest that little nitrogen is lost to denitrification, apparently because anoxic conditions limit the conversion of NH_4^+ to NO_3^- . These budgets also suggest that neither uptake nor immobilization are sufficient to provide a net sink for nitrogen.

The results of this study generally agree with the results from several other studies of streams in temperate forest catchments. Triska et al. (1990, 1993) showed that nitrification was inhibited in anoxic areas connected to the stream via long-residence time flow paths, and that these areas were dominated by NH_4^+ . Further, denitrification in these anoxic areas was limited by NO_3^- supply (Triska et al. 1993). This study contrasts with studies by Triska et al. (1989, 1990, 1993) and Wondzell and Swanson (1996b) that showed that oxygen in advected channel water penetrated many meters into the hyporheic zone, maintaining oxidizing subsurface conditions. In this study,

oxygenated water was not present, even short distances into the hyporheic zone, probably because low head gradients and fine textured sediment (low hydraulic conductivity) limit the supply of oxygenated water and slow subsurface flow rates. Consequently, the oxygen in advected channel water is depleted close to the stream–subsurface interface. Our study showed that the riparian and hyporheic zones of the Karuushinai River are potentially net sources of nitrogen to stream ecosystems, as has been shown for other temperate forest streams (Wondzell and Swanson 1996b) and tropical forest streams (Chestnut and McDowell 2000). However, neither our study, nor the other studies document increased nitrogen concentrations in stream water, suggesting that added nitrogen is retained at the interface between water upwelling from the subsurface and the stream channel.

Conclusions

There were distinct spatial patterns in subsurface flow paths through the gravel bar at our study site. Hyporheic exchange flows dominated water fluxes on the channel side of the gravel bar whereas fluxes through the hillslope side of the bar were dominated by soil-water inflows from the adjacent hillslope. Inputs of soil water increased during a small storm event, but the biggest change in water fluxes was a substantial decrease in stream water inputs to the channel-side compartment during the storm. Our nitrogen budget suggests that the hyporheic zone was a sink for both NO_3^- and DON. NO_3^- was probably immobilized or denitrified, whereas DON was probably mineralized to NH_4^+ . The gravel bar/hyporheic zone was a net source of nitrogen to the stream during both base flow and during the small storm event. Nitrogen exports from the hyporheic zone to the stream were dominated by NH_4^+ . Further, exports exceeded inputs, suggesting that the riparian forest soils were an additional source of nitrogen for the stream.

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