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HYPORHEIC FLOW IN A LOW ORDER MOUNTAIN STREAM

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Introduction

In low order mountain streams the topmost layers of the hyporheic gravel, the bedsediments sensu Bretschko (1991a, 1994, 1998), are densely colonized by the epigeic fauna. Abundances can exceed 100.000 individuals per square metre for macrozoobenthos (Williams and Hynes 1974, Bretschko 1991 a, b). The densities of meiobenthos and microorganisms are several orders of magnitude higher (Bott et al. 1984, Schmid-Araya 1994). On account of the lack of autochthonous production below the light discontinuity, the energetic base of the system is organic matter, mainly allochthonic. Organic matter is imported by surface-runoff, throughfall (Bretschko & Moser 1993) and subsurface imports via soil- and groundwater (Fiebig 1995). Surface and interstitial water are interacting and transport organic matter, nutrients and oxygen through the interstices between the sediment particles, which are colonised by biofilms, functioning as an effective biological filter (Triska et al. 1993). The interstitial flow is not only a transport mechanism (Munn & Meyer 1988) but also an important physical factor for the invertebrates (Statzner & Higler 1986, Statzner 1988). Exchange processes between groundwater, surface water and the hyporheic zone are well documented (review in Brunke & Gonser 1997), but little is known about flow velocities and flow patterns within the hyporheic zone. Flow velocities are measured either indirectly by calculating the filtration rate, using the Law of Darcy (Kovacs 1991) or directly by measuring the dilution rate of dye or salt in a stand-pipe (Pollard 1955, Terhune 1958, Peter 1985) or the dilution of plaster (Panek 1990, Angradi & Hood 1998). These methods measure the velocity in one spatially defined point; it is not possible to measure flow direction or the mean flow velocity over a longer distance.

The aim of the present study is the measurement of the flow of water through the bedsediments. With salt as tracer we tracked the direction and measured the velocity of hyporheic flow within a distance of one to 10 metres. Our hypothesis is, that the saline solution injected into the sediment is transported downstream and diluted slowly along its flow through the hyporheic zone. Due to this dilution process a a saline plume is expected with decreasing salt concentrations in increasing distance from the injection point.

Study site and methods

The investigation is carried out within the "Ritrodat" study area (altitude 615 m; 47°51'N, 15°04'E) of the Biological Station Lunz. A detailed description is given by Bretschko (1991a, 1991b, 1998).

In the hyporheic zone the actual (effective) flow length is longer than the straight flow length between two points, because the water flows around sediment particles. In the study area the interstitial water of several litres sediment is replaced by a mixture of cement and a fast-bonding agent (Bretschko pers. comm.). Because of ist great viscosity the cement mixture flows preferentially into the sediment layers with the highest permeability. From these sediment-concrete complex dissections are made (slices with a diameter of appx. 20 cm and thickness of 5 cm). Meshes of 6 straight lines are placed at random angles over 9 sediment sections. Measurements are taken from the lengths of the straight lines and the corresponding lengths of flows around the sediment particles (effective flow length) on this two-dimensional plane (Fig. 1).

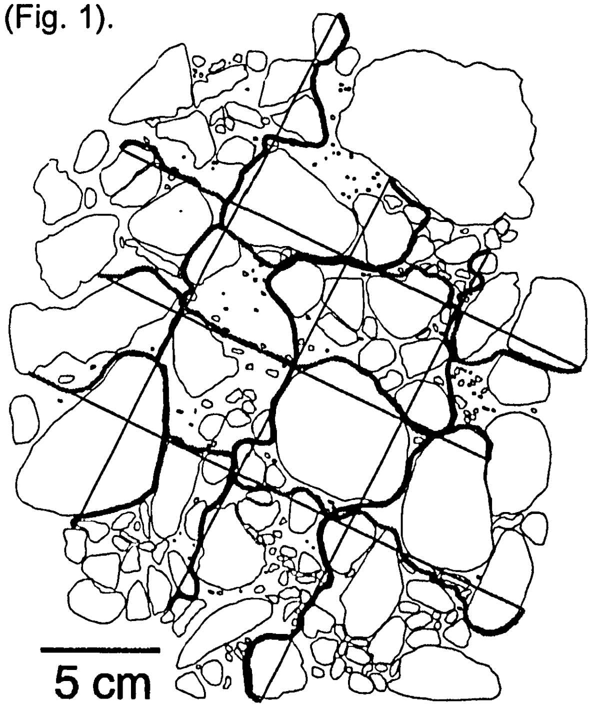


Fig 1: Example for a dissection through hyporheic sediment (interstitial water has been replaced by concrete). The thin lines are straight flow lengths; the bold, black and grey lines running along the straight lines are the theoretical traces of the flow around the sediment particles (effective flow lengths).

The sizes of the interstices are determined simply by measuring the part of the straigth line over an interstitial space on the two-dimensional plane. Since the water is flowing through a three-dimensional space the measurements on the plane are underestimating.

The hyporheic flow velocities are measured by tracer injections. For 2 hours a 10% NaCl solution is injected in 30 cm sediment depths with a steel pipe with an infiltration rate of 3 litres per minute. Several studies have shown that chloride acts as a conservative tracer (Bencala et al. 1984, Munn & Meyer 1988, Zellweger 1993). However, it turns out by colouring the saline solution that a part of it is immediately pressed along the steel pipe to the surface. Consequently an infiltration rate of 5 litres per hour and a longer infiltration time (48 hours) is used for further investigations (Wagner and Bretschko in prep.). 15 pipes of the same type are inserted in 30 cm sediment depth, distributed within 10 m upstream and downstream from the injection point (Fig. 2). From these pipes interstitial water is sampled with piston pumps (Bretschko and Klemens 1986, Fig. 2) at 10 minute intervals from the start of the injection until 2 hours after the termination of the salt infiltration.

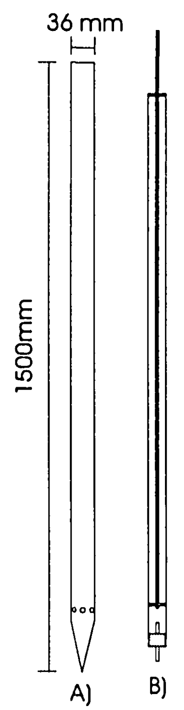


Fig. 2: Pipe for sampling interstitial water: a steel tube (A) with a tip and holes on the end is inserted down to a sediment depth of 30 cm. With a piston pump (B), the interstitial water is sampled.

The conductivity in the samples is measured with a HANNA HI 9033 conductivity meter. From the salinity pattern in time we calculate the interstitial flow velocities, using the time after which the increase of conductivity in the samples exceeds the standard deviation from the measurement of natural conductivities. The travel-distances result from the straight distances between injection point and sampling points and are corrected using the ratio factors between straight flow and the flow around sediment particles. Velocities calculated in this way are taken as a more realistic approximation (effective velocities). 10 experiments are carried out at different water levels (gauge between 115 and 135 cm) in the time between 3.11.1997 and 2.6.1998.

For all combinations of interstices and flow velocities Reynolds numbers are calculated using the formula for flow in pipes (Rp in Tab. 2), as suggested by Kovacs (1991). The dependence on the respective water temperature is considered for fluid density and the dynamic viskosity. For statistical analysis data are log(x+1) transformed to achieve independency of means and variances. Analyses were carried out in SPSS 6.0.

Results

The natural conductivities of surface water and sediment water of different depth layers are uniform (ANOVA, $p > 0.05$, $n = 203$) and amount 221 µS cm^{-1} ± 17 SD. The interstices have a mean diameter of 1 cm and the mean effective flow length in the interstitial is 27 % longer than the straight flow length. Effective flow velocities range between 0.01 and 1.32 cm s^{-1}, based on conductivity patterns from 76 pipes (51 % of all measurements, Tab. 1). Surprisingly, the infiltrated saline solution is not only transported downstream in the sediment but also against the direction of the surface

Table 1: Diameter of the interstices und effective flow length, measured from the sediment dissections. Straight flow velocities are based on the linear distances between infiltration point and sampling points; effective flow velocities are calculated using the ratio between straight flow length and effective flow length.

	width of interstices (cm)	effective flow length (% of straight flow)	straight flow velocity (cm s^{-1})	effective flow velocity (cm s^{-1})		
				mean	min	max
mean	1.0	127	0.16	**0.21**	0.17	0.27
variance	0.5	133	0.02	0.03	0.02	0.05
mean min	0.3	105	0.01	0.01	**0.01**	0.02
mean max	3.1	164	0.81	1.02	0.85	**1.32**
n	71	72	76	76	76	76

flow. Increased conductivities are measured downstream up to a distance of 12 m and upstream up to 6 m. The flow velocities of the two directions do not differ significantly (t-test, $p > 0.05$, $n = 76$). Discharge has no significant influence on the flow velocities (ANOVA, $p > 0.05$, $n = 76$).

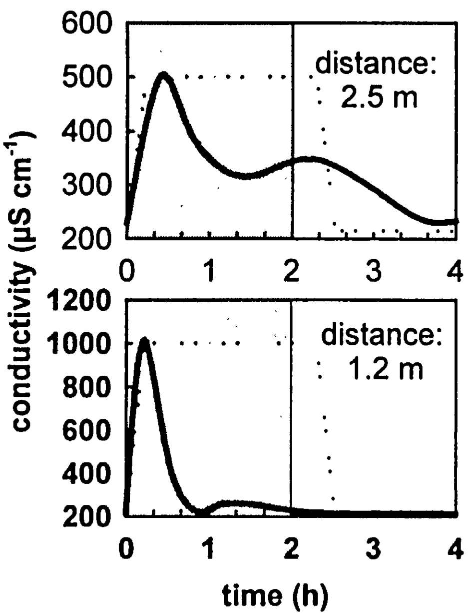


Fig. 3: Conductivities from samples of 2 pipes. Continuous line: measured values, punctured line: theoretically expected conductivities. Shaded area: period of infiltration.

Moreover, a remarkable phenomenon occurres frequently: during the infiltration the conductivity increases distinctly in the sampling pipes at the beginning, but decreases before the infiltration ends (Fig. 3).

Under the studied conditions the interstitial flow is always definitely laminar, as shown by the range of Reynolds numbers in Table 2.

Table 2: Reynolds numbers for the flow in the hyporheic interstitial for all combinations of interstices with effective flow velocities.

$$Rp = (V \times L \times \rho) / \mu$$

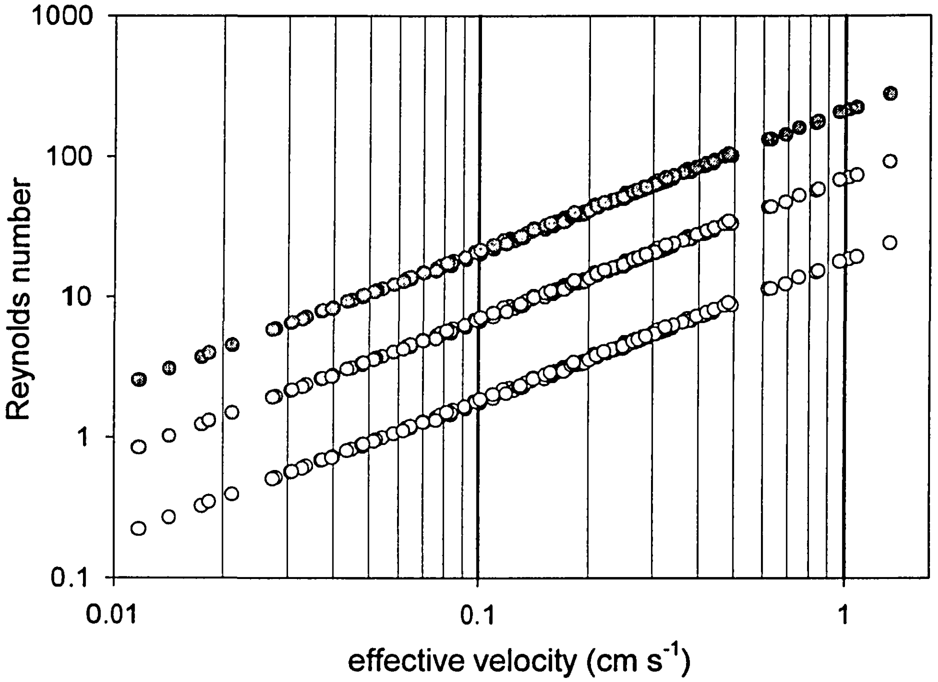
V = velocity (ms^{-1})

L = characteristic length (m)

ρ = fluid density (kg m^{-3})

μ = dynamic viscosity (Ns m^{-2})

		size of interstices		
		mean	min	max
effective flow velocity	mean	**14.3**	3.8	43.6
	min	0.8	**0.2**	2.3
	max	91.1	23.9	**275.8**



○ mean interstices (1.03 cm)	$y = -0.0534 + 69.3466\,x$	$R^2 = 1$	n = 227
● max. interstices (3.1 cm)	$y = -0.1611 + 210.055\,x$	$R^2 = 1$	n = 227
○ min. interstices (0.3 cm)	$y = -0.0144 + 18.1786\,x$	$R^2 = 1$	n = 227

turbulent (Re > 2000) if velocity > 29 cm s^{-1} (for mean interstices)
> 10 cm s^{-1} (for max. interstices)
> 110 cm s^{-1} (for min. interstices)

Fig 4: Correlations between flow length and Reynolds numbers. Limits of turbulent flow are calculated from the regression formulae.

The Reynolds numbers are increasing linear with the effective velocity and correlations are significant for different sizes of the interstices (Fig. 4). As calculated from the regressions, the flow would start to be turbulent (R > 2000) above a velocity of 10 cm s^{-1}, by far more than the greatest velocities measured in this study.

Discussion

The use of dye or salt to trace water flow is common for investigating infiltration rates (Bencala et al. 1984, Munn & Meyer 1988, Triska et al. 1993, Zellweger 1993), but rarely used to measure subsurface flow. Williams & Hynes (1974) carried out a comparable study with fluorescin dye to measure interstitial flow between two pipes in the Speed River, Ontario. They reported interstitial velocities of about 0.08 cm s^{-1} in 10 cm depth to 0.02 cm s^{-1} in 40 cm sediment depth. Using the plaster dilution method, Panek (1990) and Angradi & Hood (1998) found higher velocities. Angradi & Hood (1998) reported velocities up to 3.5 cm s^{-1} in a first order Appalachian stream, but they measured mainly in the upper layer in 0 to 10 cm sediment depth in which velocities are higher on account of the influence of the surface flow (Peter 1985, Panek 1990). Panek (1990) measured in the RITRODAT study area with the plaster dillution method interstitial velocities up to 3.8 cm s^{-1} with a mean velocity of 0.4 cm s^{-1}, which matches good with the findings of the present study. The plaster dilution methods give higher velocities because they register spatially very located peak values, whereas measurements with tracers integrate the velocities over a distance. Furthermore, good agreement of our data is given with the results of Schwoerbel (1966), Tilzer (1967) and Richter & Lillich (1975).

Generally, hyporheic water velocity is controlled by the flow regime and the sediment structure (Bretschko 1992, Vervier et al. 1992, Brunke & Gonser 1997). The often described increase of the interstitial flow velocity with the discharge (Panek 1994, Angradi & Hood 1998) is not found. Probably the discharge alterations during the investigation are too small. Also the measurements are made below a sediment depth of 20 cm where the water velocity is not directly influenced by the surface flow (Peter 1985). However, the variability of the measured flow velocities is high within a sampling date: pipes located in a distance of 1 m or nearer show very different responses. This demonstrates that the salt solution develops no saline plume in the hyporheic zone but flows in spatially defined layers or channels in the sediment. These channels are areas of higher hydraulic connectivity in which the interstitial water flows faster. The drop of the salt concentration in the samples during infiltration, which happened frequently after the initial rise, could be taken as an indication of high temporal variability.

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