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An estimate of spatial ratios for preferential water flow in a melting snowpack

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Abstract

Observations were made on the diurnal variations of discharge solute concentrations of a melting snowpack. The eluted solute amount, from the pack in one day in the late melt season, accounts for less than 23% of the total amount in the waters held in the pack before discharge increasing in the day. Based on the preferential flow model, this ratio can be interpreted into spatial ratios of the preferential flow. It is estimated that the spatial ratio of the preferential flow channels to the whole pore volume of the snowpack is below 32%, which includes 13% for the main channels and 19% for the non-main channels.

1. Introduction

Snow is one of the important components of environment at high latitudes and altitudes. It exerts great impact on the environment in its deposit, metamorphosis and melt. For example, when a snowpack begins to melt in the spring, solutes within the pack are released in a way what scientists call as fractionation process and preferential elution (Goto-Azuma, 1998). These two processes lead to a very high acidity in the first melt water, and its magnitude is not affected by snow depth (Harrington and Bales, 1998a). The acid flush may cause severe physiological stress to aquatic organisms, fish kill (Leivestad and Muniz, 1976) and many other environmental problems. Therefore, it is very important to explore the physical, chemical and hydrological processes involved in the snow metamorphism, especially in the melting snow metamorphism.

Similar to water flow in other porous media, when a snowpack melts, meltwater is considered to flow only in partial pores, which is termed the "preferential flow". This part of the water is called mobile water, and another part is the immobile water which includes water in dead-end pores, thin liquid films around solid particles, and immobile intra-aggregate water or water in isolated regions associated with unsaturated conditions (Kutilek and Nielsen, 1994). The formation of preferential flow paths has been observed in both arctic (Marsh and Woo, 1984) and temperate (Kattlemann, 1985) snowpacks. Preferential flow develops heterogeneous distribution of solute concentrations in the profile of a snowpack, which has also been confirmed by both experiment (Harrington *et al.*, 1996) and field work (Ishikawa *et al.*, 1997). Following models for other porous media and with adaptations for processes unique to snow, Harrington and Bales (1998b) developed an ionic solute transport model. The model considered metamorphic processes involving melt-freeze episodes, as well as the dual-velocity nature of the snowpack. The solute behavior of the mobile phase is described by an advection-dispersion equation, and the mobile and immobile waters exchange is described with a fixed rate constant. Feng *et al.* (2001) applied a similar model using a series of artificial rain-on-snow experiments. Rare earth elements, sprayed onto the snowpack surface at intervals, were used as chemical tracers. It is found that the model, with a fixed exchange rate constant, can produce frequently observed decreases in solute concentrations with increasing discharges but does not explain the concentration increases with increasing water flux seen in their experiments. Hence the mobile-immobile exchange rate coefficient is assumed to increase exponentially with the effective water saturation, and then the solute concentration variations of the discharge are successfully simulated. These observations suggest that solute transport is largely dominated by fast flow channels. This paper is to explore the spatial ratio of preferential flow channels via analyzing the diurnal solute concentration variations of discharge of a melting snowpack.

2. Observations

The field work was conducted at the Snow Melting Research Station of Institute of Low Temperature Science, Hokkaido University, Moshiri (44°23′N, 142°17′E), in the northern part of Hokkaido, Japan from March 10, when the snowpack began to melt and its depth was 150 cm, to the end of April 1989, when the snowpack disappeared (Suzuki, 1991). The data used in the present analysis were obtained between April 20 and April 26. Discharge water was collected using a 3.6×3.6 m^2 lysimeter placed at the bottom of the snowpack. Discharge was measured by a tipping-bucket gauge and a pulse counter. Meltwater was sampled hourly or every two hours using an automatic water sampler. Meltwater percolating in the snowpack was also collected by a porous cup and a vacuum syringe. A trench was dug and a plastic tube with the porous cup was installed in the snowpack horizontally. Percolating water surrounding the porous cup was sucked up into the vacuum syringe. Air temperature was recorded by an automatic station. Precipitation was collected and sampled.

These samples were filtered and then subjected to chemical analysis for ion concentrations by an ion chromatography (Dionex-2020i/SP).

3. Diurnal Variations of Discharge and its Solute Concentration

In a melting season, the snowpack surface generally undergoes diurnal melt-freeze cycles and thus the discharge also shows daily variations. It has been shown (Zhou *et al.*, 2001a) that there is still faint discharge flow even if the melting has ceased for one or two days. Since meltwater is produced in the surface layer within several centimeters (Zhou *et al.*, 2001a; Ohno and Nakawo, 1998) the discharge of snowpack can be partitioned into two parts: the meltwater from the surface produced in the day, and the water held in the pack since before. Thus the solute concentrations of the discharge depend on the concentration contrast, mixing and exchanging of the two parts. Figure 1 shows some data obtained in around April 21. It includes variations of the air temperature, discharge and its Cl^- and SO_4^{2-} concentrations. It can been seen, from Fig. 1, that high discharges are associated with low solute concentrations as a whole, which indicates that solute concentrations in surface meltwater are lower than in the interstitial water since before. From 0:00 to 6:00 on April 21, the Cl^- concentration increased in a small range, and decreased slightly after 6:00. Then, from 7:00 to 10:00, it remained a concentration of about 156 μeql^{-1}. After 10:00, the concentration decreased as the flow rate increasing dramatically, then reached the lowest value of 80.9 μeql^{-1} at 13:00 and kept it constant (actually varied in about 1 μeql^{-1}) till 15:00. After that, the concentration increased again. In the variations, if the lowest concentration stands for the value for the surface meltwater, as suggested by Suzuki (1993) and Suzuki *et al.* (1994) then we can further explore some solute transport mechanisms.

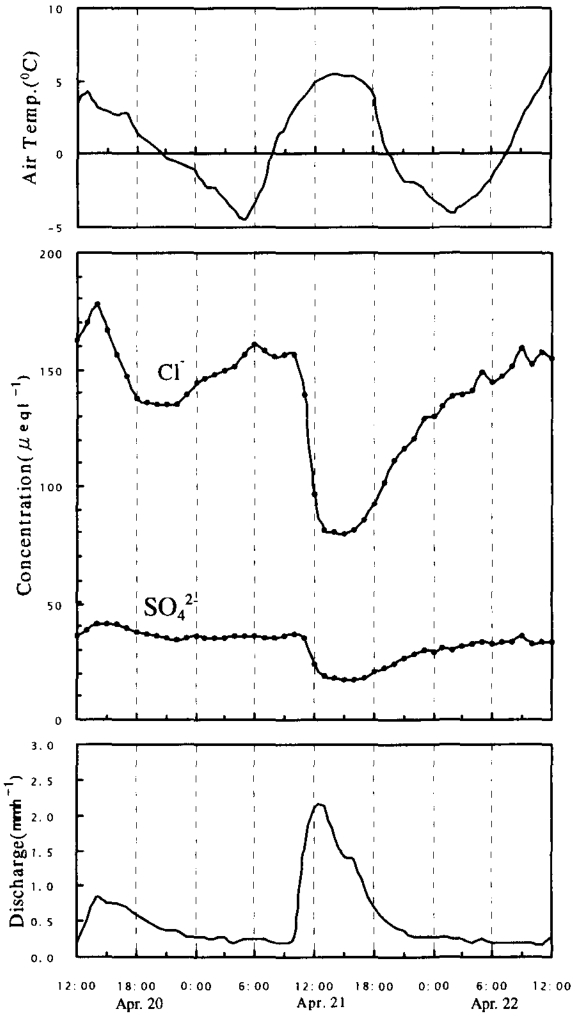

Fig. 1. Variations of discharge, its solute concentrations and air temperature around April 21.

Since the measured snow thickness was only 45 cm on April 21, the shape of the discharge curve (Fig. 1) indicates that the surface melt curve is an approximate parabola (Zhou *et al.*, 2001b) which can be partly seen from the air temperature variations (Fig.1). When air temperatures become above 0 °C in the morning, the surface layer melts and then meltwater begins to percolate downward. According to the theory of preferential flow, the meltwater would preferentially enter the most easily accessible water channels and mix with the water having existed in the channels since before, and thus the snow water content would go up. As the channels consist of pores and meanwhile connect around with numerous pores of which some are holding immobile water, the rising of water content is not just an increase in water fluxes in these channels; some water would enter the surrounding pores and keep linked with the channels. For the water leaving the channels, some would still be mobile water, and thus increase channels; some would mix and mobilize some immobile water; some other water would become immobile water as thin liquid films, water in dead-end pores or in other forms. As the surface melting intensifies, the water fluxes in the already developed channels increase and new channels are developed. Since the new channels have not been eluted yet, solute concentrations in these channels are higher than in the old ones. Therefore, solute concentrations of the discharge depend on the contrast of water fluxes and concentrations between the old and the new channels, supposing the solute concentration of the surface meltwater is constant in a same day and the solute exchange between snow grains and percolating water can be ignored. As the water flux is related to the effective water saturation in a power function (Colbeck, 1972) and then the increasing rate for the former is much larger than for the latter, the growth of water fluxes is mainly in the old channels. Since solute concentrations of the water in the old channels become lower and lower as the elution proceeds, the solute concentration of the discharge decreases gradually before the discharge peak. After the peak, an opposite process occurs, namely, the portion of the water in the early developed channels gets smaller and smaller with the portion in the late developed channels becoming larger, thus the solute concentration of discharge returns to the high value gradually. Figure 2 shows the concentration-discharge variations. It can be seen from the figure that both curves for Cl^- and SO_4^{2-} are consistent with the above-explained model and each forms a roughly clockwise loop. Differences are still shown on the flow-falling limb of the hydrograph. With the discharge having dropped in a large range after the outflow peak, the lowest concentration remains for about 4 hours rather than increases instantly. Considering a constant concentration supposed for the surface meltwater, the remaining can only be explained by saying that concentrations for the waters in the relatively fast flow channels have reached the same level and the waters from other channels can be ignored. This level must be the concentration value for the surface meltwater, namely, solutes in these relatively fast flow channels are eluted out, because water in the early developed channels should always have higher velocities and lower concentrations than in the late developed channels except that all the concentrations have reached the lowest level. Therefore, the lowest value 80.9 μeql^{-1} is regarded as the surface meltwater concentration.

Nevertheless, in the above analysis, the solute exchange between mobile and immobile water is ignored, and so the lowest concentration value could still be higher than the surface meltwater concentration. According to Harrington's model (Harrington and Bales, 1998b), the exchange rate coefficient is constant, and equals 4×10^{-6} s^{-1}. Feng *et al.* (2001) related it to the effective water saturation in a power function. However, the mobile-immobile solute transfer is a very complicated problem. If immobile water is truly quiescent even when water flux is changed, then the transfer is only by molecular diffusion. Nevertheless, the solute transfer coefficient depends not only on the molecular diffusion coefficient but also on the distance and the cross-sectional area for diffusion (Smedt and Wierenga, 1984) and is a function of not only the diffusion rate but also the solute retardation factor, column length, and the fluid velocity (Young and Ball, 1995). Many experiments show a dependence of the mass transfer rate on pore water velocity (Griffioen *et al.*, 1998) because the water content is greater at high pore water velocities, probably resulting in a larger cross-sectional area (Smedt and Wierenga, 1984). In fact, there is no clear line to divide the mobile and immobile phases, and the immobile water is not really immobile. When a sudden increase in water flux occurs at the surface, this disturbance propagates down the column faster than the water itself (Feng *et al.*, 2001). This means that the flow velocity or the effective water saturation at a given depth would increase before the water introduced at the time of disturbance reaches that depth, or in short, the immobile water can become movable. In the modeling, while hydrodynamic dispersion and interphase mass transfer are distinct physical processes, their effect on the concentration signal is similar; therefore it is probable that the pair of parameters used in the calculations is nonunique (Harrington and Bales, 1998b). Therefore, it is very hard to calculate the diffusion effect involved in the lowest concentration of the discharge. However, the diffusion should be very limited because the snow thickness is only 0.45 m through which, according to our observation (Zhou *et al.*, 2001b), the meltwater flux peak passes in less than 50 minutes (The molecular diffusion coefficient for chloride is 0.054 cm^{2}h^{-1} (Griffioen *et al.*, 1998)). Even though Harrington's exchange rate constant of 4×10^{-6} s^{-1} is adopted, and assuming a Cl^- concentration of 70 μeql^{-1} for surface meltwater, 70 minutes for percolating time, a lowest water saturation of 7% and 100% of both the mobile and immobile water involved in the exchange, the solute exchange amount equals only 1.26‰ of the amount eluted in one hour at the time of around discharge peak. Furthermore, the concentration does not change for a long time after the discharge peak. Therefore, we ignore the diffusion and treat the lowest concentration as that for the surface meltwater.

4. Estimate of the spatial ratios for preferential flow

On the basis of above analysis and assumptions, the solute amount eluted from the snowpack in the day can be partitioned. In Fig. 2, A is the start point for discharge increase and F is the point for the discharge falling to the start level, supposing both the discharge and concentration curves (Fig.1) are continuous functions, then the eluted Cl^- amount Q_1 for the flow-increasing ABC segment (Fig. 2) is:

$$Q_1=\iint_{ABC} c(u,t)dudt-c_0\int_A^C u(t)dt, \qquad (1)$$

where u is water flux of discharge, t is time, c is concentration and c_0 is the lowest concentration. Similarly, adopting constant c_0 for CD segment, the eluted Cl^- amount Q_2 for the flow-falling DEF segment is:

$$Q_2=\iint_{DEF} c(u,t)dudt-c_0\int_D^E u(t)dt. \qquad (2)$$

Using the hourly data shown in Fig. 1, we got: $Q_1=1.51\times10^{-2}$ μeqcm^{-2}, and $Q_2=7.65\times10^{-3}$ μeqcm^{-2}. It has been known, from the above conceptual model, that Q_1 is the solute amount from the fast flow channels, and that Q_2 is from the slow ones as well as those far from the mains. There should be some solute exchange between the fast and the slow channels, and on the rising limb of the hydrograph (Fig.1) the net balance for the slow ones should be positive, namely, Q_2 should contain some amount from the fast channels. This amount could probably flow in at a time approaching the discharge peak, which means Q_2 may mainly come from the finally developed channels because only these channels would have lower water saturations and be far from the main channels. Since the concentrations near the discharge peak are lower, this exchange amount is ignored in the following calculations.

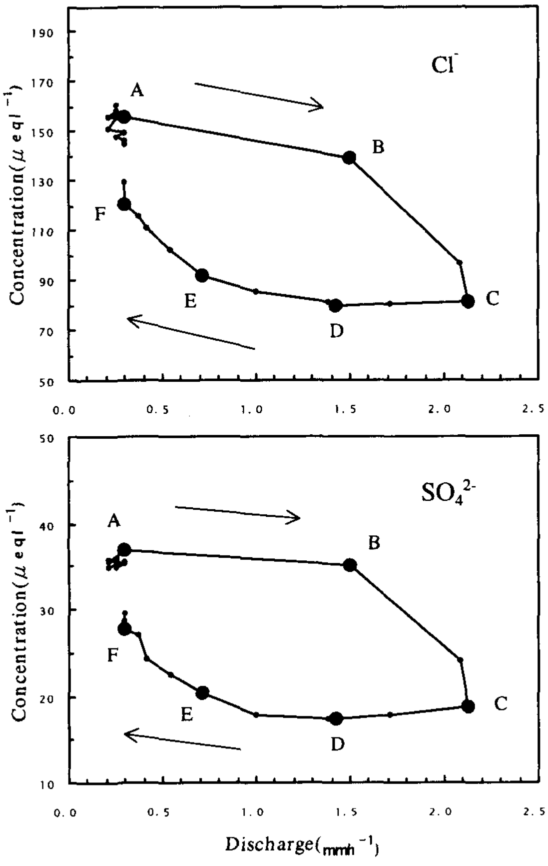

Fig. 2. Concentration vs discharge for April 21.

Figure 1 shows that the flow rate of the discharge was 8.1×10^{-8} ms^{-1} (0.293 mmh^{-1}) before it began to increase in the morning. According to the field measurements by Zhou *et al.* (2003) the flow rates correspond to an averaged water saturation of 0.05-0.07 for the snowpack. The averaged snow density is 470 kgm^{-3} with the snow depth being 0.45 m in the day. As the calculated melt thickness of the day is 0.03 m, a thickness of 0.42 m is adopted for the solute amount calculations. Based on these data and using the discharge concentration of 156 μeql^{-1} before the increase, the total solute amount for Cl^-, in the different pools or flow paths of the pack before discharge increasing, is calculated to be 0.1688-0.2419 μeqcm^{-2}. Thus the ratios of the eluted solutes to the total amount are 6.24% to 8.95% for Q_1, 3.16% to 4.53% for Q_2, and 9.40% to 13.48% for the whole. Suppose a uniform distribution of the waters in the pack before the outflow increase and the differences of the solute exchange between grain melting and water freezing can be neglected, the solute amount ratio for Q_1 can be regarded as the spatial ratio of the main channels (the fast flow channels) to the whole pore volume of the snowpack, but the ratio for Q_2 can't be interpreted so simply because the slow and far-reaching flow channels are just partially eluted. According to the above model, the farther the channels from the main ones are, the less the solutes in them are eluted, and vice versa. Thus we assume that the eluted ratio decreases linearly from 100% in the main channels to 0% in the farthest channels, then the spatial ratios of the non-main channels (the slow and far-reaching flow channels) to total pores are two times as large as the ratios for the solute amount, being 6.32-9.06%. A sum of the two ratio ranges yields a range of 12.56-18.01% which stands for the ratio of the pores for water percolation to the total pore space of the snowpack (P).

As shown in Fig. 2, the shape of the concentration-discharge plot for SO_4^{2-} is the same as the loop for Cl^-. Following the above calculations, we obtained: $Q_1(SO_4^{2-})=8.75\%\sim12.53\%$, $Q_2(SO_4^{2-})=3.05\%\sim4.37\%$, and a P value of $14.85\%\sim21.27\%$. The $Q_1(SO_4^{2-})$ ratios are higher, but the $Q_2(SO_4^{2-})$ ratios are almost the same as compared with those for Cl^-, which might come from the distribution differences of the two kinds of solutes in the snowpack.

Figure 3 presents a data set similar to the one shown in Fig. 1, for April 24. Figure 4 is the concentration-discharge relations of the day. Compar-

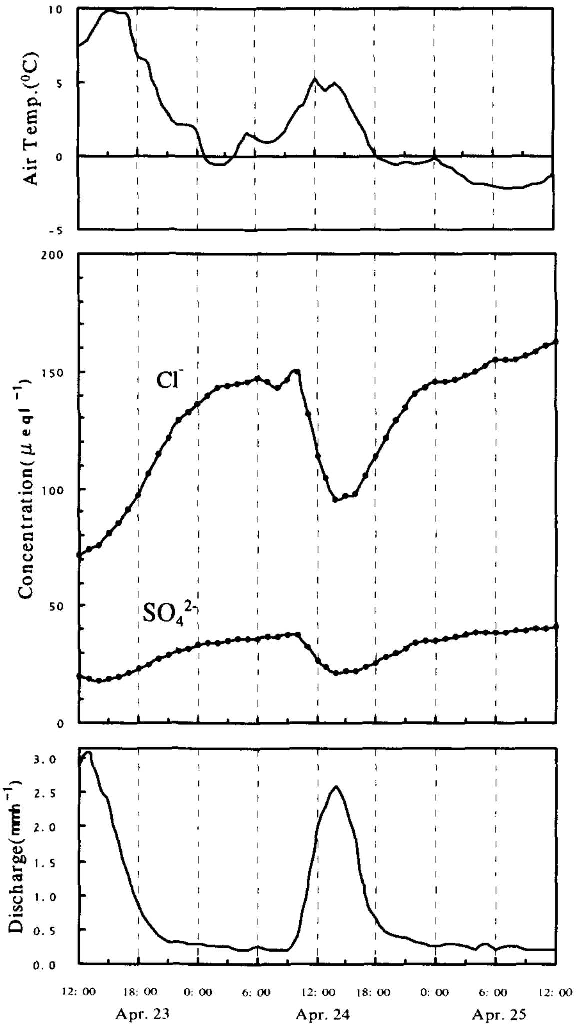

Fig. 3. Variations of discharge, its solute concentrations and air temperature around April 24.

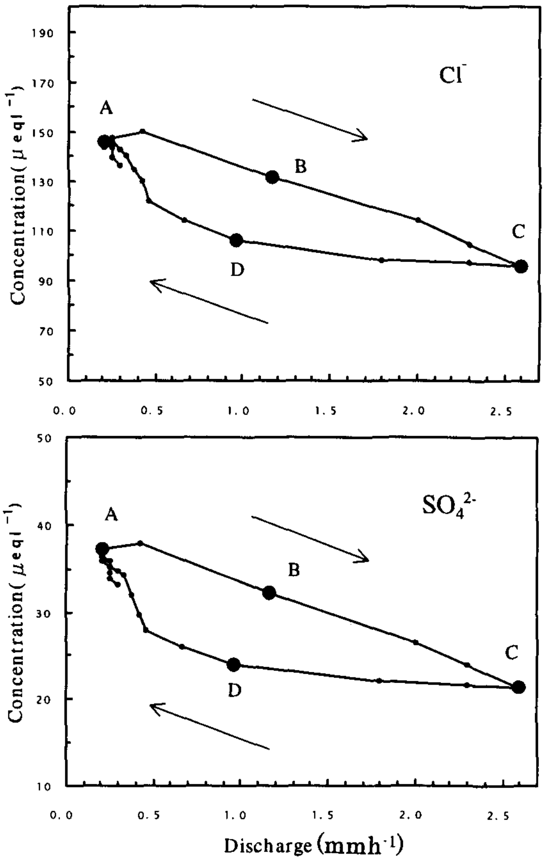

Fig. 4. Concentration vs discharge for April 24.

ed with Fig. 2, Fig. 4 is different in three aspects: first, on the flow-rising limb of the graphs, the concentration decreases almost linearly with discharge increasing, yet the drop in Fig. 2 is initially slow and finally fast. Second, once the discharge peak passes, unlike a constant value kept for four hours in Fig. 2, the concentration rises at once as exactly as described in the model. Third, when the discharge returns to the start level, the concentration also falls exactly back to original level, but in Fig. 2 it does not until the flow rate reaches a level lower than at the beginning (Fig. 1). Comparing Fig. 1 with Fig. 3, it can be seen that the air temperatures were lower in the early morning of April 21, and the snowpack surface froze, but on April 24, the lowest air temperature was about 0 °C and its duration was very short, so the freeze, if any, would be very few. The freezing process excludes and pushes the impurities to the surface of the newly formed ice (Colbeck, 1981) which then accelerates the elution. Moreover, there was about 1.5 mm (w.e.) snowfall in the early morning of April 21, and its average Cl^- concentration was 130 μeql^{-1}. Therefore, the concentrations in the outflow were higher and their decreasing was slower in the early period. The differences of restoration in concentration may be related to the elution of the day before. If the discharge was larger and the elution was stronger in a first day than in a second day, then the spatial range of flow channels in the second day would be smaller, and thus the concentration can return back to the original level when or even before the flow rate restores; and vice versa. This agrees well with the situations in the two days (Fig.1 and Fig.3). The discharge on 21st was larger than on 20th, but that on 24th was smaller than on 23rd.

Although the lowest concentration of discharge water might be higher than the concentration in surface meltwater on 24th, the discharge concentration, after the lowest, went up slowly, only a gain of less than 3 μeql^{-1} in two hours. Hence we treat it as the concentration for the meltwater of the day and make similar calculations as made for April 21. The results are: 7.76%~11.37% for $Q_1(Cl^-)$ (ABC segment in Fig. 4) 6.27%~9.18% for $Q_2(Cl^-)$ (CDA segment in Fig.4) and 20.30%~29.73% for P; 9.03%~12.94% for Q_1 (SO_4^{2-}) (ABC segment in Fig.4) 6.47%~9.28% for Q_2 (SO_4^{2-}) (CDA segment in Fig.4) and 21.97%~31.50% for P. These values are higher than for 21st, and the actual differences should be larger than these because, in the morning of 21st, there was about 1.5 mm (w.e.) new snow which had a high concentration. These differences should arise from the melting and then the discharge differences between the two days. According the above model, a larger total outflow volume, together with a higher peak flow rate, would result in a larger ratio of the eluted solutes to the total amount in a pack. This is truly the case for 24th versus 21st with the former having higher values (Fig.1 and Fig.3) and furthermore, the snow thickness on 24th was 0.1 m thinner than on 21st (The peak flow rate, on 24th, would be larger if the snow were 0.1 m thicker than the observed one (Colbeck, 1972)). Therefore, although there are some differences in the ratios between the two days, the ratios of the main-channel space to the whole pore volume of the pack are within 13%, and the P values are within 32%.

5. Discussion

In the above model and calculations, two assumptions are essential. One is assuming a constant concentration for the surface meltwater in one day; and another, for the calculation of total solute amount in the waters within the snowpack, is presuming the discharge concentration, before flow increasing in the morning, being the average concentration of the waters at that time.

As melting occurs in the surface layer, its water content is the highest, and its grain coarsening is the most fastest (Zhou *et al.*, 2002) among all the layers of a snowpack. These features, together with the diurnal melt-freeze cycles, lead to a low concentration in the surface layer. Observations (Ishikawa *et al.*, 1997) have revealed a very homogeneous and low concentration in the upmost 0.2 m around of a pack, which is especially true in a late melting season. Since the daily melt thickness is generally not larger than 0.05 m (Zhou *et al.*, 2002) and it is actually about 0.03 m on April 21, thus the assumption of a constant concentration for the meltwater produced in one day is tenable.

As mentioned above, different flow channels have different flow rates, elution degrees and solute concentrations. Following the recession and stop of surface melting, part of the waters in the non-main channels may begin to flow back to the main channels because of the quicker flow decrease in the mains, and mix with the existed water, therefore, the longer the time from melt ceasing, the more homogeneous the distribution of solute concentrations in a snowpack. For April 21 and 24, before flow rising, the discharge concentrations varied only within 5 μeql^{-1} for 6 to 8 hours, thus solute distributions in the pack should have been rather homogeneous except some isolated waters far from the flow region. However, the heterogeneity would not change our conclusion because the water, from the main channels and having a lower concentration, constitutes a larger portion of the discharge, so taking the discharge concentration as the average value for the waters held in the pack would just produce a smaller total solute amount value and then a larger P value.

Surface freeze also affects the estimation of the total solute amount because the solute amount in the frozen part, like that in the unfrozen part, is also calculated using the pack-uniformed water saturation and solute concentration. The low concentration of the surface snow produces meltwater with low concentrations, but freeze concentrates solutes onto surfaces of snow grains (Colbeck, 1981) and thus enhances the meltwater concentrations, hence the actual concentrations depend on these two competing processes. Meltwater, at the depth of 0.05 m within the pack, was collected by the porous cup and the vacuum syringe on April 21. The average concentration (Cl^-) of the water collected in the initial 1.5 hours (10:30 -12:00) was 135 μeql^{-1}. Since the outflow rate almost reached its maximum at 12:00 (Fig. 1) the initial concentration at the water saturations of 5-7% could probably be higher than the adopted value (156 μeql^{-1}). This would also result in a larger P value and wouldn't make the conclusion untrue.

In the above analysis, we also neglected the solute exchange in the melt-freeze process between percolating water and ice grains. As the freezing process enriches the infiltrating meltwater (Colbeck, 1981) thus the lowest solute concentration in the outflow is still higher than the source concentration at the surface even if the mobile-immobile effects are ignored. This process occurs simultaneously with both the dispersion and the diffusion, hence it is rather complicated to distinguish them. Feng *et al.* (2001) ignored this process in their modeling as well. The coarsening rate of snow grains increases with increasing water content (Zhou *et al.*, 2002). A large flow rate corresponds to a high water content value. Thus the concentration from the melt-freeze effect would increase with increasing flow rates, which is similar to the diffusion effect. Yet the increasing rate for flow rates is much more rapid than for water saturations (Colbeck, 1972) and in Fig. 2, after the flow peak, we don' t see a decrease in the concentration for about 4 hours. It is unlikely to consider that these two effects, in the 4 hours, were offset by the flow rate decreasing effect on the mixing of the surface meltwater with the pore water, because the grain coarsening rate is very slow (Zhou *et al.*, 2002) and the ice crystals should be very clean after experiencing the long-time metamorphism (Colbeck, 1981).

There are some differences between P values calculated by Cl^- and by SO_4^{2-} respectively. These differences may result from solute distribution differences in the snowpack, or from the preferential elution of which the mechanism has still not been well understood.

The largest outflow rates in the two days (April 21, 24) were above 2.00 mmh^{-1}, belonging to the high levels. Especially, the largest rate on 24th was 2.59

mmh^{-1} beyond which only two days' largest rates (3.20, 3.50 mmh^{-1}) were in the whole melt season (Suzuki, 1991). Therefore, our P values may be applied to the snowpacks in late melt seasons, and may not, to the packs in early periods because the flow channels, especially the main ones, should have been developed gradually and at the initial stage the snow density, grain size and stratigraphy all are quite different (Zhou *et al.*, 2002). Most probably, early P values are larger than late P values, which may partly explain the mechanism of preferential fractionation.

6. Conclusion

On the assumptions that snow surface meltwater has a uniform solute concentration in one day and that the solute concentration, in the outflow before the discharge increase in the morning, equals the average concentration of the waters within the snowpack at the same time, the eluted solute amount, from the pack in one day in the late melt season, accounts for less than 23% of the total amount in the waters held in the pack before discharge increasing in the day. Based on the preferential flow model, this ratio can be interpreted into spatial ratios of the preferential flow. It is estimated that the spatial ratio of the preferential flow channels to the whole pore volume of the snowpack is below 32%, which includes 13% for the main channels and 19% for the non-main channels.

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