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EFFECT OF SNOW AND FIRN HYDROLOGY ON THE PHYSICAL AND CHEMICAL CHARACTERISTICS OF GLACIAL RUNOFF

ANDREW G. FOUNTAIN

US Geological Survey, PO Box 25046, MS-412, Denver, CO 80225, USA

ABSTRACT

Near-surface processes on glaciers, including water flow over bare ice and through seasonal snow and firn, have a significant effect on the speed, volume and chemistry of water flow through the glacier. The transient nature of the seasonal snow profoundly affects the water discharge and chemistry. Water flow through snow is fairly slow compared with flow over bare ice and a thinning snowpack on a glacier decreases the delay between peak meltwater input and peak stream discharge. Furthermore, early spring melt flushes the snowpack of solutes and by mid-summer the melt water flowing into the glacier is fairly clean by comparison. The firn, a relatively constant feature of glaciers, attenuates variations in water drainage into the glacier by temporarily storing water in saturated layer. Bare ice exerts opposite influences by accentuating variations in runoff by water flowing over the ice surface. The melt of firn and ice contributes relatively clean (solute-free) water to the glacier water system.

KEY WORDS glacial runoff; ice; snow; firn; streamflow; stream chemistry

INTRODUCTION

Prediction of the rate and quantity of water flow through a glacier is complicated by the number of different hydrological processes that need to be considered. Surface processes include overland flow on bare ice and both unsaturated and saturated flow in snow and firn. Water in these surface layers generally drains into the body of the ice via crevasses. The englacial and subglacial drainage systems then route the water through and eventually out of the glacier. Englacial systems are largely unknown, whereas subglacial systems have been the subject of considerable attention (e.g. Weertman, 1972; Röthlisberger, 1972; Walder, 1986; Kamb, 1987). This paper addresses the physical processes of water flow though snow and firn on temperate glaciers and assesses their effects on runoff in glacial streams. In addition, snowmelt chemistry and chemical processes in the snow are reviewed with respect to their effect on stream waters flowing from glaciers. The effect of snow chemistry on basal waters and its influence on basal weathering is briefly reviewed.

PHYSICAL PROCESSES

Snow-free ice surfaces

In the bare ablation zone of a glacier, surface meltwater and rain flow across the ice surface to nearby moulins and crevasses where the water enters the body of the glacier (Stenborg, 1973). Water can accumulate along grain boundaries and within cracks in the surface ice. Little storage is expected in dynamic regions of the ablation zone because such ice is not long exposed at the glacier surface and subjected to solar radiation, thus the glacier ice exhibits narrow grain boundaries. As the surface ice is ablated, it is continually replenished as ice is advected from the glacier's interior by the emergence velocity (Meier and Tangborn, 1965; Paterson, 1981). In stagnant ice, however, such storage can be significant because the ice has been subjected to long periods of solar radiation that widens the grain boundaries, as often observed in melting lake

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ice (Ashton, 1980). Measurements by Larson (1977; 1978) show a water-table in the stagnant ice of Burroughs Glacier, Alaska (Figure 1). The maximum daily fluctuations of the water level are about 2 m and the hydraulic transmissivity of this active layer, derived from pump tests, is about $7{\cdot}7 \times 10^{-5}\,m^2\,s^{-1}$. However, it is unlikely that the storage volume is large, as indicated by the relatively quick response of the stream flow from Burroughs Glacier to diurnal variations in meltwater input (Larson, 1978). The quick response, typical for most glaciers, is indicative of the rapid flow of meltwater across the bare ice in the ablation zone and through the glacier's interior to the glacial streams (Fountain, 1992a).

Water flow in snow

When the snowpack is below freezing, rain and meltwater refreeze in the snowpack and release latent heat. This is the primary mechanism that warms a snowpack (Marsh and Woo, 1984; Conway and Benedict, 1994). The depth of water penetration depends on the snowpack temperature and structural characteristics

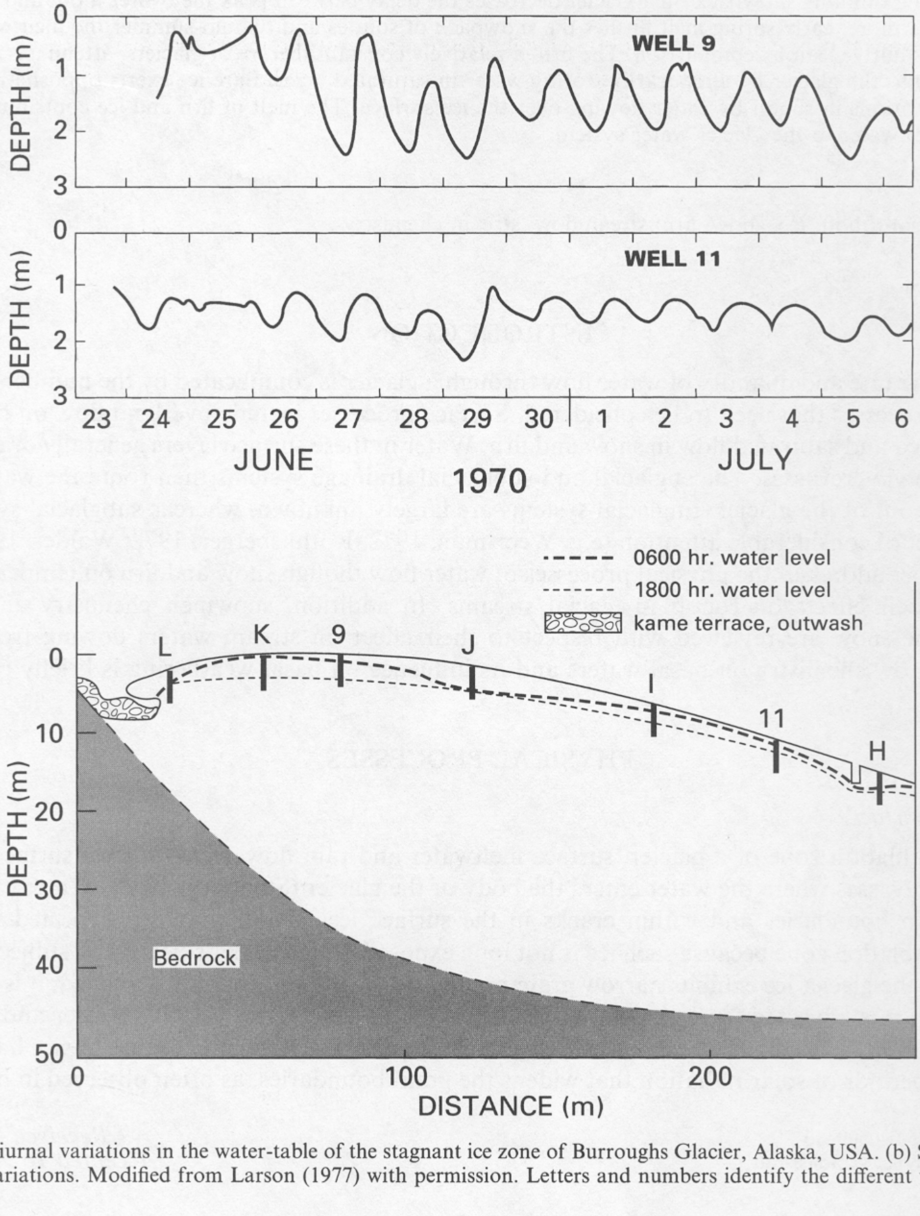

Figure 1. (a) Diurnal variations in the water-table of the stagnant ice zone of Burroughs Glacier, Alaska, USA. (b) Spatial water level variations. Modified from Larson (1977) with permission. Letters and numbers identify the different wells

and on the flux of water into the snow (Pfeffer *et al.*, 1990). Clearly, the refreezing of meltwater delays the penetration of water into and runoff from the glacier. In fact, for dry, cold snow, more than half of the surface melt is used to supply the irreducible water saturation (water retained by capillary forces) from the time the snowpack starts to melt until the time that the water reaches the base of the snowpack (Marsh and Woo, 1984).

For glaciers with large elevation ranges, the snowpack will warm earlier and stay warm longer at lower elevations than at higher elevations. If the elevation range of the glacier is large, snow may not melt in the upper reaches. Conversely, if the elevation range of a glacier is small, as for small alpine glaciers, there is little change in snow temperature. In either instance, the relation between snowpack temperature and elevation must be considered in assessing melt and runoff. Subsurface temperatures must also be understood given that meltwater may freeze in the lower layers of snow or in the firn. Surface ablation measurements may suggest a mass loss to runoff when in fact a redistribution of mass from the surface to the interior of the glacier has occurred (Trabant and Mayo, 1985). Neglecting the effects of snow and firn temperature can result in underestimates of the glacier mass balance (Trabant and Mayo, 1985) and overestimates of the predicted runoff.

The rate of flow through the snowpack is largely determined by the ice layers in the snow. Water stops flowing vertically when it encounters an ice layer and ponding above the ice layer occurs, which then saturates the snow. Water then flows laterally in the saturated snow to a gap in the ice layer before continuing downwards (Colbeck, 1973). Ice layers may disintegrate quickly, perhaps within hours when subjected to large meltwater fluxes (Gerdel, 1954, as interpreted by Male, 1980). The vertical flux of water in homogeneous snow is related to the intrinsic permeability of the snow and its effective saturation (Colbeck and Davidson, 1973)

$$u = \alpha k S^{*3} \tag{1}$$

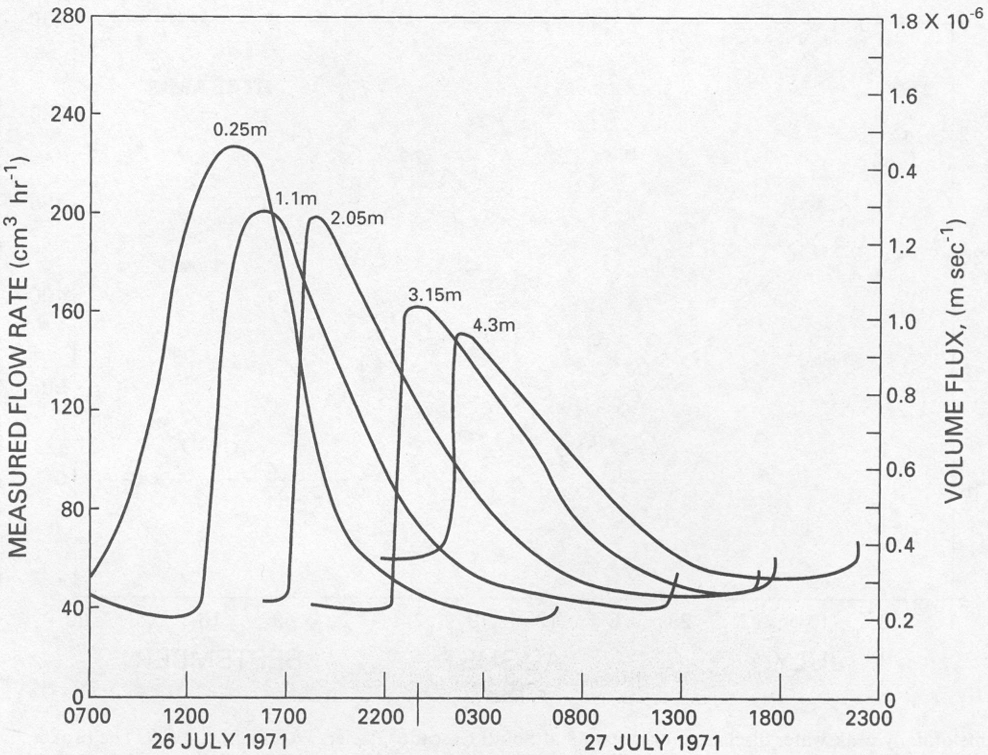

Figure 2. Diurnal meltwave at different depths below the snow surface. Note the steepening front of the wave. Modified from Colbeck and Davidson (1973) with permission

where u is the vertical liquid flux (volume of water flowing per unit area per unit time) in m s^{-1}, α is a constant equal to density multiplied by gravity divided by water viscosity, k is the intrinsic permeability and S^* is the fraction of pore volume containing moving water (effective saturation). It can be shown (Colbeck and Davidson, 1973) that the rate of vertical infiltration is related to the water flux

$$\left.\frac{\mathrm{d}z}{\mathrm{d}t}\right|_u = 3\alpha^{1/3}k^{1/3}u^{2/3}\phi_\mathrm{e}^{-1} \tag{2}$$

where $\mathrm{d}z/\mathrm{d}t$ is the rate of downward movement for flux u and ϕ_e is the effective porosity. This equation indicates that the rate of movement is dependent on the vertical flux and implies that larger meltwater fluxes will catch up with smaller fluxes and will form a wave characterized by a sharp front (Figure 2), like a shock front, followed by a slow recession. If Figure 2 is typical of a mature and homogeneous snowpack, such experiments indicate a wave speed of about 0·3 m h^{-1}. This speed is roughly equivalent to the lateral flow speeds in the saturated zone of a ripe snowpack on a sloping glacier (Fountain, 1992b).

Once the water reaches the base of the snowpack, it either meets the surface of the underlying ice or percolates into the firn. Part of the water drains directly into crevasses, but this occurs under a small fraction of the total snow-covered area. If the water meets the ice surface, as in the case of an ice layer, it will saturate the snow to form a shallow water-table. The water then moves downslope and drains into the nearest crevasse. It is not uncommon to encounter slushy snow when walking in the ablation zone of a glacier in late spring. The water may saturate the full thickness of the snow, such as at the base of a steep slope. Under the right circumstances of water saturation and surface slope, the snowpack may fail, creating a slush avalanche (Onesti, 1987; Elder and Kattelmann, 1993). Also, ice crystals eroded from the ice surface or entrained from a nearby snowpack often form temporary ice jams in the small streams flowing over a glacier's surface (pers. obs.). Although both situations result in pulses of water entering crevasses or moulins, they are

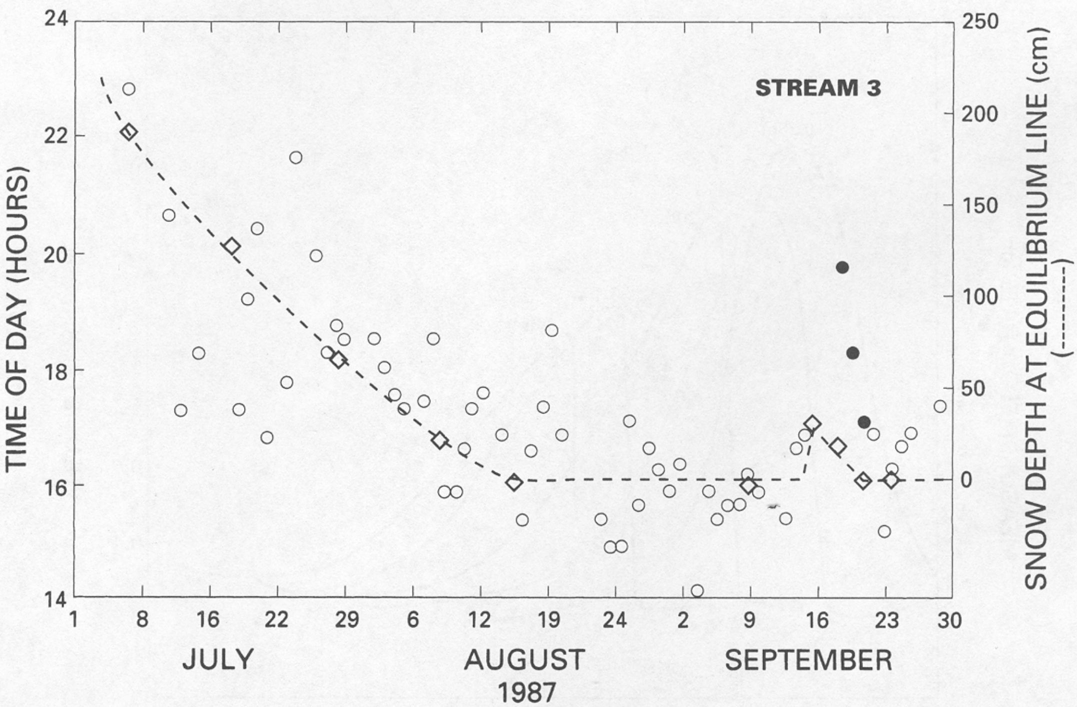

Figure 3. Time of daily peak water discharge for stream 3 at South Cascade Glacier, Washington, USA. The broken line is the interpolated snow depth at the equilibrium line; open diamonds indicate measured thickness; closed circles are times of peak daily discharge following a snowfall (adapted from Fountain, 1992b). Reproduced courtesy of the International Glaciological Society from the *Journal of Glaciology*, 1992, **38** (128), 191, figure 2

either rare in frequency or small in magnitude and probably do not appreciably affect either the rate at which water flows over the glacier or the discharge of water flowing from the glacier.

Snow-covered ice

The presence of snow on ice in the ablation zone significantly decreases the speed of water flow compared with that of flow over a bare ice surface. The delay is equal to the vertical transit time through the snowpack to the saturated zone plus the lateral transit time through the saturated zone to the nearest crevasse. This effect is shown in Figure 3. The time of day of peak flow in a proglacial stream decreases as the season progresses until mid-season, after which no further change is apparent (Fountain, 1992b). The earlier appearance of peak flow correlates with decreasing snow depth, measured at the glacier equilibrium line. Snow depth, in this instance, is a proxy measure for the elevation of the snowline on the glacier. Decreasing depth implies that the snowline is moving up-glacier and revealing more ice. As more bare ice is exposed, a greater amount of water moves quickly from the surface to the interior of the glacier. The trend of earlier peak flows terminates when the snowline reaches the firn line. Above the firn line, the firn itself stores water, as will be discussed later, precluding fast surface flow.

The effect of snowcover on the timing of peak flow was particularly well demonstrated by the 14 September snowfall (Figure 3), which covered the bare ice surface with a layer of snow 0·25 m deep. When the timing of peak flow could be reliably determined, four days later, the snow was 0·16 m deep and the peak was delayed by 3·75 h, compared with pre-snowfall values. As the snow depth decreased and the snowline retreated up-glacier, the time of peak flow appearance returned to pre-snowfall times. Both the seasonal change and short-term change caused by the late summer snowfall can be explained by vertical and lateral transit times through the snow (Fountain, 1992b).

The increasing amplitude of the diurnal variation in discharge from spring to mid-summer (Figure 4) can be partly explained by meteorological conditions and snowcover. In spring, compared with summer, the skies are cloudier, resulting in less solar insolation, the air temperatures are cooler and the albedo of the glacier is higher (Figure 5). These conditions produce less meltwater than in the summer and reduce the amplitude of diurnal variations in discharge. The amplitude is further attenuated by the presence of a snowpack over much of the ablation zone. The path length for vertical percolation increases with elevation on the

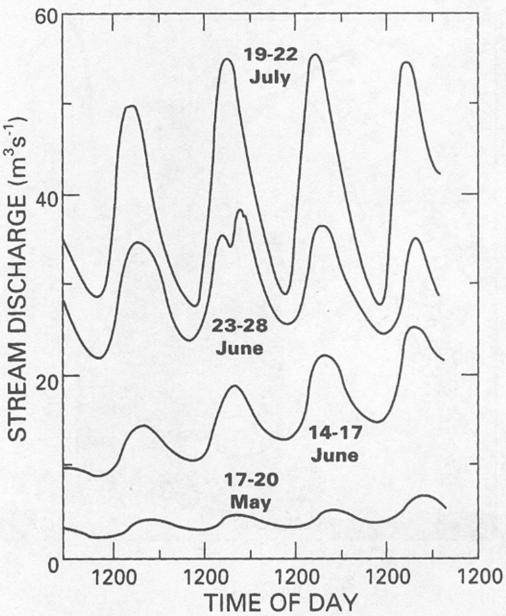

Figure 4. Daily variations in glacial stream discharge in the Matter-Vispa river basin, Switzerland. Modified from Elliston (1973) with permission

glacier because of the average increase in snow depth with elevation; thus the meltwater reaches the base of the snowpack at a different time. At the base, water flows laterally through the saturated layer and the path length depends on the distance between crevasses, which can vary significantly. If the path length, including the vertical and lateral components, were constant, then the diurnal variations of meltwater input would be delayed by some constant time lag. However, the path length is not constant because of variations in snow depth and in the separation distance between crevasses. The aggregate effect of these variations is to attenuate the amplitude of the meltwater wave entering the snowpack by the time it passes into the glacier via the crevasses. In contrast, when the ablation zone is snow-free, water is routed rapidly across the surface to the nearest crevasse and the delay between generation and penetration into the glacier is relatively small, which increases the amplitude of the diurnal variation of streamflow.

This explanation contrasts with the commonly held opinion (Elliston, 1973; Bezinge, 1981; Röthlisberger and Lang, 1987) that the decreasing delay results from the enlarging subglacial hydraulic system. Although some delay results from flow through any hydraulic system, the exact subglacial causes of the observed large delays have not been identified. Fountain (1992b) has argued that if the subglacial hydraulic system is envisioned as a pre-existing network of conduits, then we cannot explain the decreasing delay. Alternatively, the enlarging hydraulic system may be viewed as increasing its spatial coverage by connecting with subglacial regions previously isolated hydraulically during the winter. In this instance, the effect on the storage and delay of water flow is uncertain.

Under some circumstances, the thinning snowpack and increasing area of bare ice in the ablation zone favour development of large subglacial conduits. Hooke (1984) calculated a range of values of water discharge, glacier slope and thickness for which the rate of conduit enlargement exceeds that of closure. In this situation, conduits will most often be partly full at atmospheric pressure. To keep the ice walls from closing,

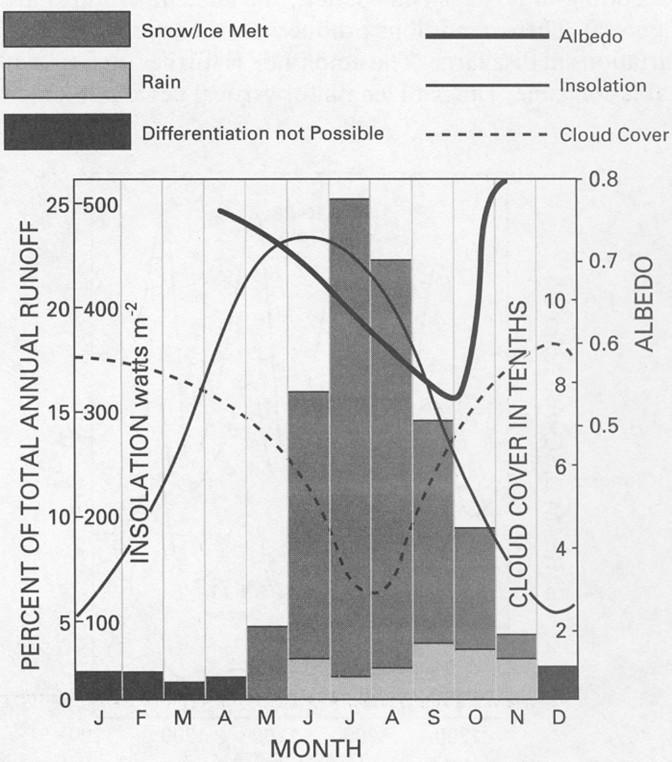

Figure 5. Seasonal variation in components of runoff from South Cascade Glacier with average glacier albedo, cloud cover and insolation (modified from Meier, 1969). Reprinted from *Journal of the American Water Works Association*, **61** (1) (Jan 1969), by permission. Copyright© 1969, American Water Works Association

the conduit needs to be pressurized from time to time. Calculations (Fountain, 1992a) show that for relatively thin ice (70 m) and a 10% slope, only a few hours of full-conduit flow at atmospheric pressure are required to compensate for several days of closure. These conditions often exist in the ablation zone of glaciers, particularly near the terminus where the ice is thin. For these situations, the damped variations of water flow in the glacier, caused by the presence of the snowpack, will minimally enlarge the conduit. Later in the season when the ablation zone is generally free of snow, larger discharge variations will enlarge the conduits to a greater extent than expected for large, but steady, discharges. The implication is that the position of the snowline may significantly influence the size of the conduits.

Firn

Once the water has percolated through the snow in the accumulation zone, it encounters the firn, the metamorphic transition between snow and glacier ice. Firn is porous and water percolates downwards through the unsaturated zone by presumably the same process as that for unsaturated snow. In temperate glaciers, percolating water forms a saturated layer at the base of the firn (Sharp, 1951; Schommer, 1977; Ambach *et al.*, 1978; Oerter and Moser, 1982; Fountain, 1989). Water is found from 0 to 40 m below the surface and drains to crevasses (Lang *et al.*, 1977; Schommer, 1977; Fountain, 1989). The water-table forms in the spring when water percolates into the firn and is depleted in the autumn when surface melt stops (Figure 6) and the firn water drains into crevasses. The maximum range in seasonal water levels at South Cascade Glacier, from which the saturated thickness is inferred, is 2 m (Fountain, 1989).

The presence of another porous medium below the snow layer indicates that the routing of water flow from the surface of the accumulation zone to the interior of the glacier is further delayed. For reasons similar to those discussed earlier — increasing snow and firn thickness with elevation and the variation in distance between crevasses — it is likely that spatially averaged drainage into the body of the glacier does not exhibit appreciable diurnal variations. This conclusion agrees with that of Humphrey *et al.* (1986). Furthermore, the volume of the slowly varying (baseflow) component of the streamflow can be explained by the volume of daily meltwater input into the accumulation zone (Fountain, 1992b).

Seasonal storage of water in the firn, which depends on the thickness of the saturated layer and the effective porosity of the firn, was estimated for South Cascade Glacier. The results indicated that about 11 cm of water averaged over the accumulation zone of the glacier, or $1{\cdot}78 \times 10^5\,m^3$ (Fountain, 1989),

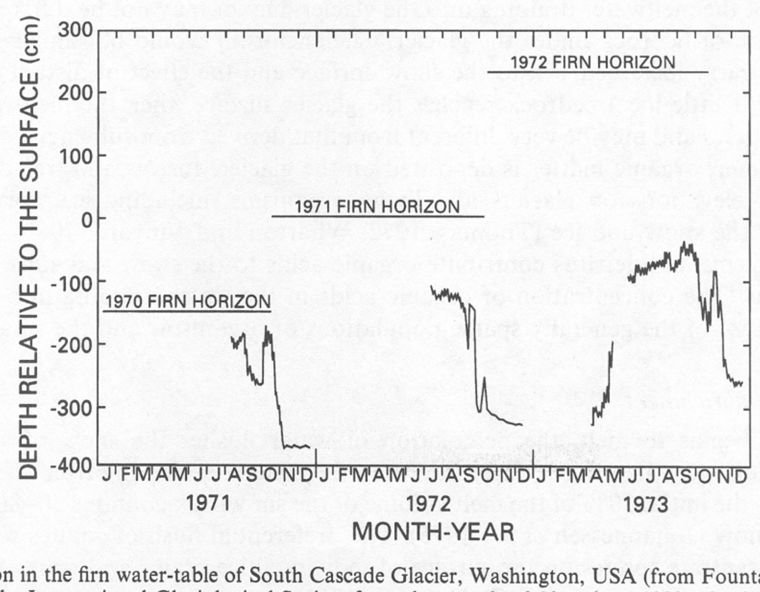

Figure 6. Seasonal variation in the firn water-table of South Cascade Glacier, Washington, USA (from Fountain, 1989). Reproduced courtesy of the International Glaciological Society from the *Annals of Glaciology*, 1989, **13**, 191, figure 2

were stored. This volume of water represents about 12% of the total spring storage measured at South Cascade Glacier by Tangborn *et al.* (1975). The drainage of the stored water contributes to the runoff from the glacier in the early winter months. Generally, the saturated layer completely drains by late November.

SNOW CHEMISTRY

Origin

The solute content of a snowpack, in part, reflects the chemistry of the atmosphere in which the snow formed and through which it falls. Ice crystals form by a number of different processes that involve either homogeneous or heterogeneous nucleation (Hobbs, 1974). The latter process requires a nucleus on which water vapour condenses. As the snowflake descends through the atmosphere it scavenges atmospheric contaminants, either by gas adsorption or particle adhesion (Junge, 1977). Adhering particles may include cloud droplets contaminated with acidic gases and solid particulates such as dusts or ashes (Magono *et al.*, 1979). Acidic gases, such as SO_2 and NO_2, can be absorbed by aerosol particles, rain drops and snow crystals and subsequently form acids such as sulphuric and nitric acids (Winkler, 1980). When the snowflake lands on a glacier, these contaminants are carried with it. In addition, when the snowpack forms, it traps local atmospheric gases within the pack. Gas exchange can continue in the pack as air is pumped through the snow by the topographically induced pressure variations as the wind blows over the surface (Clarke and others, 1987; Colbeck, 1989; Clarke and Waddington, 1991). Not only is gas exchanged, but the snow filters and traps airborne particulates (Cunningham and Waddington, in press). The effect of wind of pumping is a relatively near-surface phenomenon, so subsequent snow accumulation reduces or eliminates the air and particulate exchange between the atmosphere and interior of the snowpack.

The solute species expected in the snowpack depends on the trajectory of atmospheric flow and the location of a glacier relative to source regions. For example, snow in the Himalayas exhibits higher concentrations of sodium and chloride than the snow in the Karakoram, indicating a greater influence of monsoon moisture in the Himalayas (Wake *et al.*, 1990). Proximity to local pollution sources increases the snow acidity and related solutes such as nitrate and sulphate (Wagenbach and Münnich, 1988). Long-range transport of dust, such as Saharan dust to the Col du Dôme, French Alps, increases the concentration of aluminium and calcium (Maupetit, 1994). Therefore, the snow on any given glacier or group of glaciers may have a unique chemical signature depending on their location relative to natural and anthropogenic source regions, on the chemical characteristics of the source regions and on the patterns of local and long-range atmospheric transport. The chemical content of the meltwater draining into the glacier may or may not be different from that obtained from the dissolution of bedrock under the glacier. The chemistry would be similar if rockfalls and wind-borne particulates carry local bedrock to the snow surface and the effect of distant sources was relatively small. Conversely, if little local bedrock reaches the glacier surface, then the meltwater chemistry is controlled by other sources and may be very different from that derived from subglacial dissolution of bedrock.

During the summer, organic matter is deposited on the glacier. Insects and tree detritus are commonly observed at lower elevations on glaciers and living organisms, including ice worms (Goodman, 1971) and algae, inhabit the snow and ice (Thomas, 1972; Wharton and Vinyard, 1983). The by-products and decay of the organisms and detritus contribute organic acids to the snow and subsequently to the glacier and glacial streams. The concentration of organic acids in the water draining into and from a glacier is probably low because of the generally sparse populations of organisms and the small amount of detritus.

Processes of solute enrichment

Once the snow begins to melt, the percolation of water flushes the snow of solutes (Skartveit and Gjessing, 1979). However, the solute concentration in the meltwater flowing from the snow is not constant (Figure 7). Rather, the initial 30% of the melt volume of the snowpack contains 50–80% of the total solutes contained in the snow (Johannessen *et al.*, 1975). The preferential flush of solutes results from microscale processes that concentrate solutes on the surface of individual ice grains and from the macroscale vertical distribution of solutes in the snowpack (Bales *et al.*, 1989).

a frozen layer. Such redistribution can produce increased solute concentrations near the base of the snowpack and will enhance the effect of a concentrated flush of solutes after the snowpack warms to the freezing temperature (Colbeck, 1981).

Field investigations have indicated a preferential flushing of certain solute species, particularly sulphate and nitrate, in contrast with chloride (Davies *et al.*, 1982; Tsiouris *et al.*, 1985; Tranter *et al.*, 1986). Controlled laboratory experiments have shown that ice grains do not preferentially absorb any ion species from meltwater (Bales *et al.*, 1989; Cragin *et al.*, 1993). Rather, the apparent preferential flushing results from the original, non-homogeneous distribution of different solutes in the snowpack (Bales *et al.*, 1989). Cragin *et al.* (1993) showed that less soluble ions, such as sulphate, are more efficiently excluded during grain growth, by vapour transfer, than more soluble ions such as chloride. However, nitrate did not show preferential exclusion relative to chloride, indicating that the initial distribution of solutes in snow may be the primary cause for the preferential flushing of solutes.

Effect of concentrated solute flushing

The solute concentration entering the glacier is largely controlled by processes in the recent seasonal snow rather than the firn. Davies *et al.* (1982) showed that solutes from melting snow do not concentrate in the firn or ice, but rather are flushed out of the surface layers and presumably away from the glacier. However, the firn does have the effect, already described for diurnal meltwater variations, to delay and attenuate the solute pulse. In contrast, once the solutes leave the snowpack in the ablation zone they will be routed quickly to the glacial hydraulic system and hence to the glacial streams. Together, these processes will probably result in a strong initial pulse followed by several days of high solute concentration, perhaps similar to Figure 7 for snow. One effect of a concentrated flush of solutes is to change the chemistry of alpine lakes and streams, and the effect is particularly significant in basins that have a limited buffering capacity to neutralize acids (Melack *et al.*, 1985). These changes can also adversely affect aquatic biota (Hagen and Langeland, 1973).

The preferential flush of ions from the snow affects glacier erosion and inferences about the subglacial hydraulics based on stream chemistry. At the base of the glacier, the increased solute concentration increases the acidity and promotes chemical erosion (Reynolds and Johnson, 1972; Raiswell, 1984). The magnitude and significance of this increased erosion is uncertain. Two techniques used to infer the hydraulics of subglacial water flow — measurements of stream water chemistry (Raiswell, 1984; Raiswell and Thomas, 1984; Thomas and Raiswell, 1984) and of electrical conductivity in glacial stream water (Collins, 1978; 1979a; 1979b; Humphrey *et al.*, 1986; Fountain, 1992b) — make assumptions about the chemistry of the water entering the glacier drainage system. For such investigations, particularly if taking place in the spring, ignoring the preferential flushing of solutes may invalidate interpretations gained from the studies. For example, Raiswell (1984) proposed using stream water chemistry to infer the hydraulic condition (pressurized or partly full) of subglacial passages with the assumption that atmospheric inputs of sulphate, via meltwater, are negligible and practically all the sulphate is derived from subglacial weathering. This assumption is significantly violated in the spring during the onset of melt when solutes are flushed from the snow. Ignoring the increased solute concentration, including sulphate, in the snowmelt would result in the misinterpretation of the hydraulic condition of the subglacial system. Similarly, the electrical conductivity of stream water has been used to separate the hydrograph of glacial streams into relatively clean englacial waters and comparatively solute-rich subglacial waters (Collins, 1979b; Vaughn, 1994). However, it is clear that such separations are useful only for those periods when the englacial water is largely devoid of solutes and cannot be applied during the early spring melt period.

CONCLUSIONS

Rapid changes in the seasonal snowpack greatly affect the flow-rate of water from glaciers. Interpretation of the variations in glacial runoff without considering the changing surface snow conditions can lead to mistaken conclusions about englacial and subglacial conditions. The slow passage of water through snow strongly contrasts with the rapid overland flow on bare ice. Therefore, as the snowline retreats up the glacier

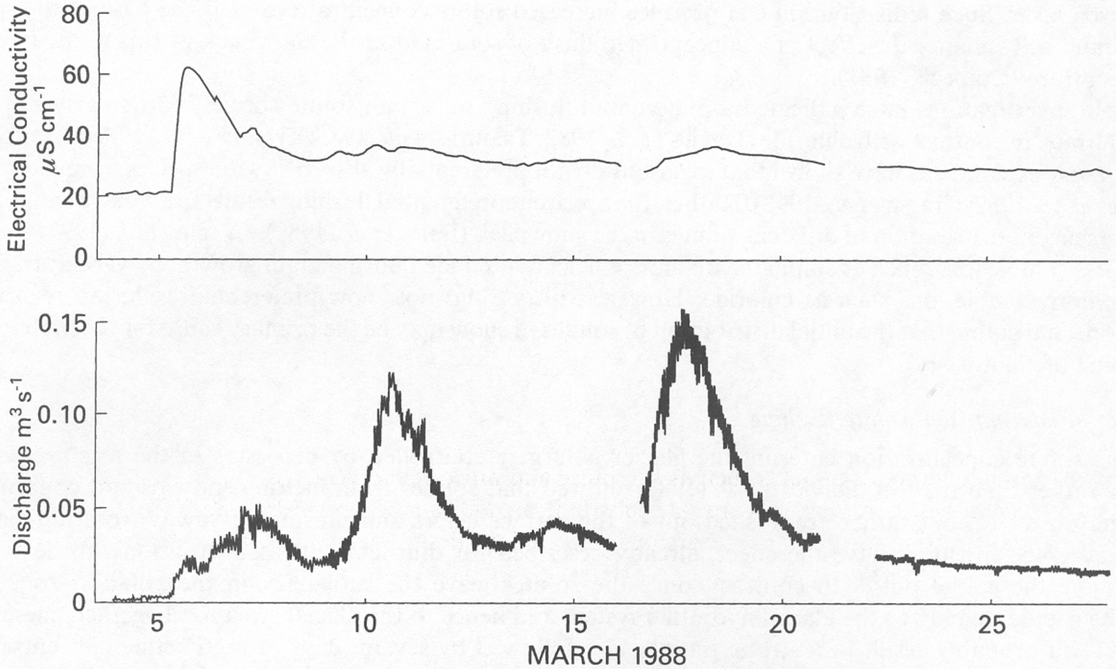

Figure 7. Electrical conductivity and water discharge in a snow-fed stream in Scotland during the first spring melt (modified from Morris and Thomas, 1985) Reproduced courtesy of the International Glaciological Society from the *Journal of Glaciology*, 1985, **31** (108), 191, figure 1

The microscale processes that concentrate the solutes are melt–freeze cycles (Bales *et al.*, 1989) and vapour transfer (Cragin *et al.*, 1993). These metamorphic changes to the crystal structure are concurrent with the chemical changes. When a snow grain begins to melt, it forms a water layer. The solute concentration in the water layer is equal to the sum of the solutes in the melting snow grain and on the grain surface. When the water refreezes, it rejects most of the solutes as the ice crystallizes, forming a rind of solutes on the outside of the grain. The solutes are excluded from the ice crystal lattice because of their inability to become incorporated into the ice lattice (Cobb and Gross, 1969; Gross *et al.*, 1987). The remaining trapped impurities are surrounded by ice rather than in the ice lattice itself. Repeated cycling of melting and freezing purifies the ice grain and increases the concentration of solutes on the grain surface.

Alternatively, solutes may concentrate on a grain surface through vapour transfer without an intervening liquid phase (Cragin *et al.* 1993). Vapour transfer results from temperature gradients produced by meteorological changes and by microscale vapour pressure differences. Temperature changes the vapour pressure such that surfaces with warmer temperatures have higher vapour pressures. Consequently, warmer ice grains lose mass to cooler ice grains. Microscale vapour pressures are also produced by the geometry of the ice grains themselves (Hobbs, 1974). A smaller grain, with a smaller radius of curvature, has a higher vapour pressure than a larger grain with a larger radius of curvature, thus smaller grains lose mass to larger grains. As an ice grain sublimates it leaves the solute behind, thus increasing the solute concentration on the grain surface. The solutes on the surface of the larger grain, on which the vapour is deposited, are maintained on the surface because they too are rejected from the growing ice front.

Processes that determine macroscale solute distribution include the initial solute content of the snow layers deposited during different storms and the redistribution of the solutes by meltwater flow. The initial concentration of solutes in the snowpack is determined by scavenging of the snowflakes, by the filtration of air moving through the snow and by dry deposition onto the snow surface, as previously discussed. The redistribution results from meltwater or rain acquiring the solutes as they wash over the snow grains. If the snow temperature is well below freezing the meltwater will refreeze, trapping the acquired solutes in and exposes larger areas of bare ice in the ablation zone, the meltwater flow increases in speed and volume. In response, glacial streams increase their diurnal variation in discharge and reach larger peak daily flows earlier in the day. The position of the snowline is thought to greatly affect the size of englacial and subglacial conduits.

The firn also affects the runoff of meltwater, but its effect is more subtle than that of the transient snow. The firn stores a significant fraction of the spring melt water. In the autumn, when surficial melt is reduced or absent, the firn drains water to the glacier. In addition, water flow through firn is significantly delayed and its variations are attenuated such that it probably provides much of the baseflow in glacial streams.

The chemistry of the water flowing into the glacier is controlled by the chemistry of the snowpack because the firn and ice have been leached of solutes. Most of the solutes are flushed from the snow early in the melt season because of microscale processes that concentrate the solutes on the exterior of the ice grains and macroscale processes that redistribute the solutes towards the base of the snowpack. The concentrated solutes increase chemical erosion at the base of the glacier, change the stream chemistry and may adversely affect the aquatic biota in the streams. Inferring subglacial processes during the early melt season based on interpretation of the electrical conductivity or geochemistry of the glacial streamflow must take into consideration the chemical processes in the snowpack.

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