Water flow through a temperate glacier

R.M. Krimmel, W.V. Tangborn, and M.F. Meier

ABSTRACT: Dye and salt tracers were used to determine the rate at which meltwater travels from the surface and margin of South Cascade Glacier to the terminus. Good results were obtained using large volumes of Rhodamine B dye when the path required percolation through snow. Salt and dye were used in free-flowing streams disappearing under the glacier margins and into moulins. Velocities in the snowpack were about 0.1 m/hr. Average velocities from the snow surface to terminus were 6 to 27 m/hr, and from moulins and marginal streams 266 to 2450 m/hr. Travel times of dye injected just above the equilibrium line at three different times suggest that the speed of flow is influenced by snow depth and the time of season. The latter influence may be due to seasonal changes of the water storage in the glacier. Experiments in free-flowing streams on the ice surface and at the margins gave velocities in the same order of magnitude and large enough to suggest open conduit flow within or under the glacier.

RESUME: Des sels et des colorants ont servi à déterminer la vitesse à laquelle l'eau de fonte s'écoule de la surface et des bords du South Cascade Glacier au terminus. De bons résultats ont été obtenus avec des volumes importants de Rhodamine B pour l'écoulement lent à travers la neige. Le sel et les colorants ont été utilisés pour les cours d'eau disparaissant sous les bords du glacier. Les vitesses moyennes de la neige ont été d'environ 0.1 m par heure. Les vitesses moyennes de la surface au terminus ont été de 6 à 27 m par heure, et des moulins et les rivières de bord de 266 à 2450 m par heure. Les colorants injectés juste au-dessus de la ligne d'équilibre à trois moments différents conduisent à penser que la vitesse de l'écoulement est influencée par l'épaisseur de la neige et l'époque de l'année, c'est-à-dire par les variations saisonnières de l'eau renfermée par le glacier. Les expériences faites sur des cours d'eau libres à la surface du glacier et sur les bords ont donné des vitesses du même ordre de grandeur et assez élevées pour que l'on puisse considérer qu'il existe un écoulement du type à conduites ouvertes sous le glacier et en son sein.

INTRODUCTION

The speed and direction that a water molecule takes from the snow or ice surface to the outlet stream of a temperate glacier is an elusive, yet essential, parameter in many hydrological analyses of glacier basins. The relative amounts of water passing through rapidly or slowly, or in temporary storage in a glacier has been a matter of controversy [1]. The importance of resolving this problem is critical in regard not only to glacier hydrology but also to glacier dynamics because of the important role of a water film or

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water pockets at the bed in controlling the rate of glacier sliding [2].

The obvious aspects of water flow in and from a glacier are the observable end points: the upper surface and the stream discharging from the terminus. Water on the surface infiltrates a porous medium (snow or firn) or runs into discrete openings in a karst-like medium (ice). The input varies with changes in the rate of ablation and liquid precipitation. Although a large number of experiments have been conducted in the laboratory and in the field, few have properly duplicated the thick stratified snowpack found on many glaciers, and few have treated the movement of water through the complete snow-firm-ice system of a glacier. Colbeck [3] has analyzed the one-dimensional motion of water through snow according to the theory of two-phase flow in porous media. These results seem to be generally confirmed by the field observations of Sharp [4] on the upper Seward Glacier and S. Colbeck and G. Davidson (1972, written communication) on South Cascade Glacier.

Much less is known about the flow of water through the ice portion of a glacier. Tracer experiments in the ablation zone [5] suggest rapid flow in open conduits, and the morphology and structures of large vertical conduits have been described [6]. Theoretical studies indicate the possibility of slow flow along a network of three-grain intersections [7]. Whether these large or small vertical conduits feed a thin film of slowly moving water to the bed is a matter of dispute [8].

The experiments reported here were conducted on South Cascade Glacier, Washington. This is a typical small temperate valley glacier with well-defined areas of ice, snow and sometimes firn exposed on the surface in mid or late summer [9] (Fig. 1). The thickness of snow overlying firn or ice is well known at all times in the summer. However, the contact between ice and overlying firn is gradational and becomes increasingly hard to define as the head of the glacier is approached. The discharge of water from South Cascade Lake, which is adjacent to South Cascade Glacier and receives water from it, has been recorded accurately since 1958 with a special weir and gauging station [10] at MG on Figure 1. No evidence of glacier outbursts (jökulhlaups) appear in the record.

EXPERIMENTS CONDUCTED

A large number of experiments were undertaken on South Cascade Glacier during the ablation seasons of 1970 and 1971 in an attempt to clarify some of the problems of time delay of water through a glacier and storage mechanisms of water within a glacier. Three types of experiments were performed: some dealt with the snowpack only, some dealt with water travel from a snow surface on the glacier to the terminus, and some dealt with the time taken by water flowing freely in open channels on ice or entering the glacier at the sides to travel to the terminus.

In these experiments attempts were made to measure the time of travel to the terminus of water originating on the surface at many different locations on the glacier, including both the ablation and accumulation areas, and of water entering the glacier at the margin at several points in the middle and upper reaches. In addition, several experiments were repeated at different times of the year or in different years to investigate variations in transit time as a function of snowpack thickness and season.
One tracer used—salt—caused an increase in the conductivity of water and was easily recorded. Because of the complex thermodynamic consequences of adding salt to wet snow which interferes with the simple movement of water, this tracer was successful only when used in free-flowing streams. A more successful tracer was Rhodamine B, a fluorescent dye. This dye appeared to behave as water when diluted in the snowpack; the slight changes in freezing point caused by the diluted dye solution probably did alter the detailed flow paths between ice grains, but we believe that this had an extremely small effect, if any, on our results. The fluorometer used was capable of detecting about 0.01 parts per $10^9$ of Rhodamine B, so very small concentrations could be detected. It was impossible to continuously record the fluorescence in the field and hence frequent and laborious sampling of water at the terminus was required. The fluorescence of the samples was then measured in the laboratory. In earlier years some simple experiments were done with fluorescein dye in which arrival of the dye cloud at the terminus was detected visually.

**WATER FLOW THROUGH SNOW**

The interior of a glacier is not accessible, so direct observations of water flux except at the surface and the outlet are generally impossible. An exception is in the snowpack, where flow can be examined to depths of several metres with some difficulty. In 1970 cores and pits through areas of small dye injections on the snow were used to determine percolation rates of meltwater containing dye through the snow and firn.

The cores and pits through dye patches indicate that large variations in the percolation rate of the dye through snow and firn can be expected. On August 19, 1970, 10 ml of Rhodamine B were spread over six similar areas of about 0.6 m of snow of density about 500 kg/m$^3$. At the time of injection, 0800 hours, there was a 12-cm radiation crust at the surface of the snow. Five patches were in the centre of runnels, a sixth was between runnels. The underlying one-year-old firn had a density of about 650 kg/m$^3$. After 2, 5, 11, 25, and 35 hours, pits or cores were cut through the snow under the patches in the runnels. These showed dye penetrations of 0.30 m (0.14 m/hr), 0.59 m (0.12 m/hr), 1.09 m (0.10 m/hr), 1.81 m (0.073 m/hr), and 1.76 m (0.051 m/hr), respectively. The patch between runnels showed 0.88 m of penetration in 11 hours (0.08 m/hr). The effect of the crust is difficult to analyze. Variations in the snowpack is the probable reason for greater penetration at one place after 25 than after 35 hours at another location.

The flow of water into pans buried at various depths was recorded in 1971, but these results were erratic and are not reported here.

**WATER FLOW FROM SNOW SURFACE TO TERMINUS**

It is relatively difficult to measure the delay time for a tracer moving through the complete snow-firn-ice system because large tracer concentrations may affect the thermodynamics of the motion of water through snow, and yet the transit times are long, so that a considerable amount of tracer is required to produce a measurable concentration at the terminus. Success was achieved by spreading
relatively large amounts of slightly diluted dye over a considerable area of snow surface with a sprinkling can.

The first experiment was the injection of 19 litres of 40 per cent Rhodamine B over 100 m² near P1 (Fig. 1) on August 5, 1970. A second experiment was a similar injection at P3 on August 23, 1970. Two more Rhodamine B tests were made in 1971 near P1: the first was injected on June 30 and the second on August 4.

All the experiments that required sampling at the terminus were complicated by the number and varying discharges of the streams issuing from under the ice. Figure 1 shows the streams emerging at the terminus, the total discharge of which is equal to the discharge measured at the lake outlet (MG), less a minor amount of runoff from the slopes northeast and directly south of the lake. The terminus discharge can be calculated from the continuously measured lake discharge by accounting for the lake storage:

\[ Q_t = Q_w + A_L (\Delta h / \Delta t) \]

where \( Q_t \) is the terminus discharge, \( Q_w \) is the lake outlet discharge (at MG), \( A_L \) is the area of the lake, \( \Delta h \) is the stage change of the outlet, and \( \Delta t \) is the time increment [11]. The terminus discharge can then be divided into percentages contributed by each stream based on instantaneous streamflow measurements. In 1970 a temporary stage recorder was in operation at TG, and during both 1970 and 1971 discharge measurements of all streams were made several times. Generally, stream ME contributed about 60 per cent of the total flow at the terminus, MW about 20 per cent, A about 10 per cent, and the remainder was contributed by B, C, D, and E. These ratios varied slightly during each experiment.

For direct comparisons of experiments it is necessary to make adjustments for diurnal and longer term variations in discharge. For each tracing experiment two illustrations are given. In the first (Figs. 2, 3), the data are given as fluorometer units versus time, as we feel the subsequent corrections made for discharge and a somewhat questionable fluorometer calibration may introduce subjective errors. In the second set of illustrations (Figs. 4, 5), the calculated flow of dye in cm³/hr against time is plotted. The sampling interval was usually one-third to one-half day.

The area under the total curve of Figure 5, which received an injection of Rhodamine in a free-flowing moulin, indicates only 73 per cent of the dye injected was discharged. This could be due to several possibilities: the stream discharges were greater than determined, dye remained trapped in the drainage system either frozen to the ice or adsorbed on silt particles, the dye deteriorated, or the calibration curve used was incorrect. We feel that a faulty calibration curve is the primary reason the moulin injection showed less dye recovered than was injected. The fluorometer was calibrated in the laboratory using tap water to dilute the dye. One calibration curve was used to analyze all the data, so any inaccuracies would not affect the results relative to each other. In the case of injections onto a snow surface, the snow remained discoloured for several days, indicating that some of the dye was held up on the snow grains.

The following summary compares the results of the snow surface to terminus-tracing experiments (Table 1). The experiments on August 5, 1970, and August 4, 1971 (Figs. 2 and 4), were similar in all respects except that in 1971 there were about 2.8 m more snow at the injection area. In both experiments dense firn (700-800 kg/m³) was
<table>
<thead>
<tr>
<th>Location</th>
<th>Date</th>
<th>Tracer</th>
<th>Distance to terminus (m)</th>
<th>Time to peak concentration (sec x 10^3)</th>
<th>Velocity (m/sec)</th>
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covered with granular, isothermal snow of density 580-600 kg/m³. Surface input was similar, usually 30 to 60 mm water equivalent per day of snow ablation with about 9 mm of rain on August 8-9, 1970. The resulting curves have similar shapes; they began rising and peaked at times that are almost identical. Smoothed curves drawn through the points indicate that the peak dye concentration in 1971 was reached 7 hours later than in 1970 (Table 1). The similar results of the August 5, 1970 and August 4, 1971 experiments show that the greater snow depth on August 4, 1971 had only a slight retarding effect on the dye passage.

The experiments on August 5, 1970 and August 23, 1970 (Figs. 3 and 5), were similar in all respects except for the location with respect to the firm line and the resulting firm thickness. The August 5, 1970 experiment was near the equilibrium line (P1), the August 23, 1970 experiment was in an area that usually has a positive balance of about 1 m of water equivalent (P3). The 0.95 m (density of 500 kg/m³) of snow at the August 23, 1970 experiment site covered firm that was distinguishable from the snow only by a dirt horizon. Although a density depth profile was not made at P3 at that time it is assumed the density increases relatively slowly with depth as compared to the P1 density depth profile. From an analysis of earlier P3 density depth profiles it is estimated that the density of firm under P3 reaches that of ice (840 kg/m³) at about 10 m. The dye concentration curve resulting from the August 23, 1970 experiment shows the first indication after 3 days and a peak around 8 days.

Comparison of the P1 and P3 experiments also indicates a separation or change of subglacier drainage channels. Stream ME discharged the most dye in the P1 injections; MW discharged almost all the dye in the P3 injection.

The flat curve (Figs. 2 and 4) from the June 30, 1971 experiment indicates that the dye of that experiment was released extremely slowly. The amount of dye injected was approximately half that in the other P1 and P3 experiments. Even so, it is felt that we did not miss the peak as samples were taken at least every other day until the August 5, 1971 experiment. Early July 1971 was unseasonably cool, periods of sun alternated with snowstorms, and uninterrupted ablation did not begin until July 10. Also, liquid water storage curves for 1970 and 1971 (Fig. 6) indicate that the June 30 experiment was the only one started before the seasonal storage release was well established. The technique used to calculate the storage curves is given in [12].

WATER FLOW FROM THE ICE SURFACE OR MARGIN TO TERMINUS

Fluorescein dye was used to visually determine the velocity of water flowing in small streams on the ice surface. A stream approximately 0.15 m wide required 600 seconds to travel 110 m (0.18 m/sec) at which point it had diverged into a system of streams some 15 m wide. In another case, dye was put into a moulin and issued from a spring in the ice about 100 m away after 4900 seconds (0.02 m/sec).

A stream about 5 m wide flowing on the ice surface was gauged regularly in 1961. Mean velocities over the cross-section ranged from 0.17 to 0.42 m/sec for corresponding discharges of 0.012 to 0.074 m³/sec.

Rhodamine B dye was injected into a moulin (M1 on Fig. 1) and subsequently sampled at the terminus, the result of which is given

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in Figures 3 and 5. Essentially all the dye had been discharged 3 hours after the injection.

Table 1 summarizes results of the ice, snow, and glacier margin injections. In the case of fluorescein injections in which only the time of the first indication was noted, the time to peak concentration was based on the Rhodamine B experiment on September 27, 1970. Results of several salt injections on the snow surface were omitted as it is suspected the salt interfered with the percolation process.

CONCLUSIONS

Figure 7 shows mean velocity of the tracer flow as a function of distance from the terminus. The average of velocities of the nine moulin (M1) injections is 0.29 m/sec and of the three ice margin injections is 0.22 m/sec, close enough to suggest similar flow mechanisms and within the velocity range of surface streams on the ice (0.17 to 0.42 m/sec). This similarity indicates that water flow beneath the glacier, at least in parts of the ablation area, and from the glacier margin above the equilibrium line, is in large open conduits.

Velocities from the centre of the glacier above the equilibrium line (P1 and P3) are two orders of magnitude lower than those of the moulin and margin injections. Water with a velocity of 0.29 m/sec would require 1.3 hours to traverse the straight line distance from P1 to the terminus (1370 m) and 2.5 hours from P3 (2600 m). If it is assumed that the flow mechanism and velocities of water below or within the glacier were approximately the same over most of the length of the glacier, the retardation effects of the overlying snow and firn for the August 5, 1970, August 4, 1971, and August 23, 1970, experiments would be respectively 49, 56, and 197 hours. The difference between the two P1 injections, 7 hours, is probably due to the additional 2.8 m of snow on August 4, 1971 (Fig. 2). A rate of 0.4 m/hr (2.8 m/7 hr) through the snowpack is considerably greater than that shown by the smaller dye patches of August 19, 1970, but other variables such as higher water and dye concentrations at the P1 location may increase the rate. It is impossible to tell from these data whether the P1 results indicate a 49-56 hour delay by percolation through snow and firn followed by rapid (0.29 m/sec) flow in a large conduit to the terminus, or a more complex path involving percolation, slow flow in tiny conduits or through a water table, followed by rapid flow in a main stream. By using 0.1 m/hr as the dye percolation rate through the snow (a rather arbitrary value from small injections on the snowpack) and by assuming primarily vertical downward flux of the dye cloud at P3, a depth of 15 m of snow and firn would have to be present at P3. This depth seems reasonable when compared with the density profiles made at P3.

These experiments indicate that water flow through glaciers is complex. There are many variables that are difficult to separate but which play an important role in the net travel time. The data from these experiments have been subjected to some additional manipulation. For instance, one might expect that rate of movement of a dye cloud is also related to the total discharge of water at that time. A plot of these two variables shows a rough but inconclusive relationship. All pertinent data have been included here to allow the reader to do his own analyses.

The major delay factors appear to be the time of season, thickness of snow, and the distance of travel. The comparison of the
June 30, 1971, to the August 5, 1970, August 4, 1971, and August 23, 1970, curves (Figs. 2-5) suggests a major seasonal factor in the release rate. The August 23, 1970, versus August 5, 1970, and August 4, 1971, comparison suggests that the distance travelled is also a major controlling factor, but in this case the path that was followed was also known to be different as dye exited from separate streams.

Although the bedrock contours of Figure 1, derived by gravity observations, do not show any apparent difference in bedrock channels from P3 to P1 to the terminus, yet dye tracers emerged in different streams. Gravity analysis results in a smoothing of the real bedrock contours and the actual subglacier topography may be very complex [13]. More experiments will be necessary to make definite conclusions. These results show that season and location primarily determine the time taken for water to pass through a glacier. Further attempts to separate variables are needed.

REFERENCES


Fig. 1. South Cascade Glacier basin and inset of terminus area. The terminus outline and stream locations are approximately correct for the fall of 1970. Dotted area is rock in late season; clear area is glacier. Long dashes are bedrock by gravity survey [13]. MG and TG indicate water level recorders; A, B, C, D, E, K, D, T, P, MW, and ME are streams; M1, M2, and M3 are moulins; P1 and P3 indicate locations of dye injection on the snow. Contour interval = 50 m.
Fig. 2. P1 dye experiments. Fluorometer units as a function of time after injection of the dye on the glacier surface.
Fig. 3. P3 and M1 dye experiments. Fluorometer units as a function of time since injection. M1 is a moulin located in the ablation area of the lower glacier.
Fig. 4. P1 dye experiments. Cubic centimeters of dye discharged per hour as a function of time since the dye was injected. The reasons for the discrepancy between the injected and recovered dye could be one or all of the following:

a) A faulty fluorometer calibration.

b) Erroneous discharges of the terminus streams.

c) Part of the dye was not released from within the glacier.
Fig. 5. P3 and M1 dye experiments. Cubic centimeters of dye discharged per hour as a function of time since injection.
Fig. 6. Liquid water storage curves for 1970 and 1971. The P1 experiment on June 30, 1971 (Fig. 2, 4) was performed during a period of increasing storage while the P1 and P3 experiments on August 5 and 23, 1970 and August 4, 1971 (Figs. 2-5) were performed during periods of decreasing storage. 1970 data taken from [12].
Fig. 7. Average velocities of the peak dye discharges as a function of the straight line discharge between the point of injection and the terminus. Distance variations are due to terminus recession. Arrows indicate range of velocities of streams on the ice surface.

There was no discussion of this paper.