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THE UPLIFT OF UNTERAARGLETSCHER AT THE BEGINNING OF THE MELT SEASON—A CONSEQUENCE OF WATER STORAGE AT THE BED?

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Abstract. Results of systematic movement studies carried out by means of an automatic camera on Unteraargletscher since 1969 are discussed together with supplementary theodolite measurements made at shorter intervals and over a longer section of the glacier. In addition to the typical spring/early summer maximum of velocity known from other glaciers, an upward movement of up to 0.6 m has been recorded at the beginning of the melt season. It was followed, after a few fluctuations of the vertical velocity, by an equal but slower downward movement which continued at an almost constant rate for about three months. Possible explanations of the uplift are discussed, the most satisfactory explanation being water storage at the bed. The observations then suggest that this storage system is efficiently connected with the main subglacial drainage channels only during times of very high water pressure in the channels. Detailed measurements showed that the times of maximum horizontal velocity coincided with the times of maximum upward velocity rather than with the times when the elevation of the surveyed poles had reached a maximum. On the basis of the hypothesis of water storage at the bed this finding means that the sliding velocity is influenced mainly by the subglacial water pressure and the actual, transient stage of cavity development, while the amount of stored water is of lesser influence.

Résumé. *Le mouvement vertical de l'Unteraargletscher enregistré au début de la fonte des neiges est-il la conséquence d'une accumulation d'eau au niveau du lit glaciaire?* Les mouvements de l'Unteraargletscher sont enregistrés systématiquement depuis 1969 à l'aide d'une caméra automatique. Les valeurs de ces mouvements, complétées par des mesures au théodolite qui sont faites sur des intervalles de temps plus courts et sur de plus grandes surfaces, sont analysées. En plus du maximum de vitesse, typique à la fin du printemps et connu sur d'autres glaciers, un mouvement ascendant pouvant atteindre 0,6 m a été enregistré au début de la fonte. Il a été suivi, après quelques fluctuations de la vitesse verticale, par un mouvement descendant égal, beaucoup plus lent, avec un gradient constant pendant environ trois mois. Les explications plausibles de ce mouvement sont discutées, la plus satisfaisante étant une accumulation d'eau sur le lit glaciaire. Les observations laissent supposer que le système d'accumulation possède seulement lors de hautes pressions d'eau une bonne liaison avec les canaux importants du réseau subglacial d'écoulement. Des mesures détaillées montrent que le moment de la vitesse horizontale maximum du glacier coïncide avec celui de la vitesse verticale maximum et non avec l'élevation maximum du glacier. Ces résultats signifient, sous l'hypothèse de l'accumulation d'eau sur le lit glaciaire, que le glissement est principalement influencé par la pression d'eau subglaciale et par l'état momentané de formation de cavernes plutôt que par la quantité d'eau accumulée.

Zusammenfassung. *Die Hebung des Unteraargletschers zu Beginn der Schneeschmelze—eine Folge von Wasserspeicherung am Gletscherbett?* Am Unteraargletscher werden seit 1969 systematische Bewegungsmessungen mit einer automatischen Kamera durchgeführt. Ergebnisse dieser Messungen werden zusammen mit ergänzenden Theodolitmessungen, welche in kürzeren Zeitabständen und in einem grösseren Gebiet des Gletschers durchgeführt wurden, analysiert. Während der Schneeschmelze wurde neben dem typischen Geschwindigkeitsmaximum, das auch von anderen Gletschern bekannt ist, eine Aufwärtsbewegung bis zu 0,6 m beobachtet. Dieser Aufwärtsbewegung folgte—nach einigen Auf- und Abbewegungen—eine gleichgrosse, aber langsame kontinuierliche Abwärtsbewegung, die etwa drei Monate andauerte. Mögliche Erklärungen der Hebung werden diskutiert; die befriedigendste ist Wasserspeicherung am Gletscherbett. Die Beobachtungen lassen vermuten, dass das Speichersystem nur während Zeiten sehr hohen Wasserdruckes eine gut funktionierende Verbindung zu den grösseren Kanälen des subglazialen Abflusssystems besitzt. Detaillierte Messungen zeigten, dass der Zeitpunkt der maximalen Horizontalgeschwindigkeit des Gletschers mit dem der maximalen Vertikalgeschwindigkeit übereinstimmte und nicht mit dem Zeitpunkt, zu welchem die Messmarken im Gletscher

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die grösste (Meeres)höhe erreicht hatten. Unter der Annahme von Wasserspeicherung am Gletscherbett bedeutet dieses Ergebnis, dass die Gleitbewegung des Gletschers durch den subglazialen Wasserdruck und den jeweiligen Entwicklungszustand der Hohlräume an der Gletschersohle beeinflusst wird, während die Menge des gespeicherten Wassers von geringerer Bedeutung ist.

1. Introduction

Detailed movement studies of Unteraargletscher have been carried out for many years (Flotron, 1973). They have revealed a strong seasonal variation not only of the horizontal, but also of the vertical velocity component. The upward movement of the glacier surface, referred to in this paper as *uplift*, has from the beginning been suspected of indicating water storage at the bed. This idea was presented at the Symposium on glacier beds held at Ottawa in 1978 (Iken and others, 1979). The main point in question was that the notion of a direct dependence of the rate of ice movement on the amount of water stored at the bed is only partly correct. Instead, the highest horizontal velocity was observed when the *rate* of vertical displacement was largest, rather than at the time when the uplift reached its maximum. The subject of the present paper is the uplift and its various characteristics. Explanations, as presented at the Ottawa symposium, are discussed, supplemented by more recent considerations. The results of a numerical model study closely related to the views expressed here, which were also discussed at Ottawa, have been published as a separate paper (Iken, 1981).

For presentation a typical year (1975) is chosen, during which an additional detailed survey by theodolite was also carried out. So far the records of glacier movement have been similar each year.

2. Former studies relevant to the uplift phenomenon

Unteraargletscher, the common tongue of the two tributaries Finsteraar- and Lauteraargletscher (Fig. 1), is a valley glacier about 5 km long and gently sloping at approximately 4° except near the terminus, where it is steeper. It is one of the most comprehensively studied glaciers in the Alps. Already in 1841 L. Agassiz and his co-workers had started a variety of investigations on this glacier which still look quite modern. Of particular interest are a precise map and detailed measurements of the glacier movement, carried out even during winter (Agassiz, 1847). On two transverse profiles the vertical movement of poles ("pieux") was also measured over a period including the winter of 1842/43. The upward movement of several poles in the central part of the glacier tongue was remarkably large and clearly overcompensated ablation. Possibly these measurements already revealed an "uplift".

Since 1928 volume change and movement measurements have been carried out by the Vermessungsbüro Flotron, father and son, at the request of the Kraftwerke Oberhasli. The results are stated annually (Commission des Glaciers de la Société Helvétique des Sciences Naturelles, [1929ff.]). Seismic investigations have been carried out by Knecht and Süsstrunk (unpublished) for the Swiss Gletscherkommission (Mercanton and Renaud, 1948). This long profile of the glacier, as obtained from these measurements, has been compared with theory by Nye (1952). The maximum depth at the confluence of Finsteraar- and Lauteraargletscher is presently about 380 m. At the site where Flotron (1973) has installed an automatic camera (Fig. 1) the glacier is 310 m thick. Seismic soundings revealed two reflecting boundaries in some areas of the snout. Their possible meaning has been discussed by Röthlisberger and Vögtli (1967) in

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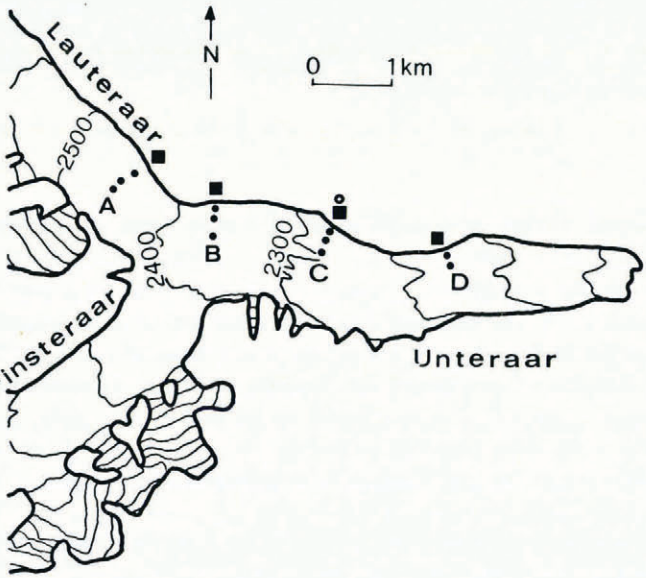

Fig. 1. Map of Unteraargletscher. ○ *automatic camera,* ■ *theodolite station,* ● *surveyed pole.*

relation to electrical resistivity soundings. The lower boundary is considered by us to represent the bedrock.

According to the early bore-hole measurements of Agassiz (1847) and temperature measurements in other glaciers of the Alps (Haeberli, [1976]), it can be assumed that the glacier is temperate below the thin surface layer of seasonal temperature variations. The changes of glacier movement and geometry in the time between 1840 and 1965 have been discussed by Haefeli (1970). Based on estimates of the sliding velocity obtained by calculating the contribution of shear deformation within the glacier to the total surface velocity, Haefeli concluded that the observed decrease of velocity since the last century is predominantly due to a reduction of the sliding velocity. A closer inspection of the problem shows that considerable scatter of the calculated values of the contribution due to sliding is introduced by uncertainties about the choice of flow-law parameters (cf. the compilation in Hodge, unpublished, p. 270) and of adequate techniques for averaging geometrical parameters in the calculation of basal shear stresses in alpine glaciers (Raymond, 1980; Haeberli, unpublished). This makes it impossible to draw a definite conclusion even about the sign of the change of the relative contribution due to sliding.

3. Methods for movement studies 1974–76

3.1. *Survey by automatic camera*

In 1969 a Hasselblad 500 EL camera with motor-driven shutter and film transport was placed in a watertight, insulated metal box on the steep slope at the left glacier margin (Fig. 1). Five reference points on the opposite rock wall were photographed together with a marker on the

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glacier. The procedure of photogrammetric evaluation and assessment of accuracy has been explained in detail elsewhere (Flotron, 1973). The mean errors of position of the target on the glacier are given in that paper as ± 14 mm in the horizontal direction and ± 25 mm in the vertical direction. Photographs were usually taken at intervals of 4 d, or daily during certain periods. Records are sometimes missing due to fog or technical failure.

3.2. *Theodolite survey*

During three weeks in June 1975 the camera records were supplemented by theodolite measurements at four transverse profiles, A–D (Fig. 1). Profile C was surveyed two to four times a day, profiles A, B and D at intervals of 1 to 3 d. For the short-time measurements a DKM-2A theodolite with automatic levelling has been used. This instrument was placed on a cairn, built with stones and held together by cement. Further measures taken in order to keep errors of the short-time surveying small are described elsewhere (Iken, [1978]). Movements were calculated from angular displacement and the horizontal distance from the instrument to the surveyed target. Accuracies of horizontal angles were estimated as one half of the difference of the angle measured in the direct and in the reversed theodolite position. Vertical angles which were measured repeatedly differed usually by less than 5^{cc}, corresponding to 2 mm for pole C_3. One would generally expect large errors caused by vertical refraction varying diurnally with temperature. However, the results show no corresponding diurnal variations in the vertical velocity. This suggests that, in the present case, vertical refraction had no significant effect.

4. Characteristic features of the movement pattern

4.1. *Presentation of data*

The results of the survey carried out by means of the automatic camera are plotted as movements in space, i.e. in a vertical plane. Figure 2 shows in the upper part the displacement of a point fixed in the ice at the glacier surface with a fivefold vertical exaggeration, while the horizontal velocity is plotted versus horizontal distance x at the bottom of the figure. Time is indicated by the dates written above some of the points. The calendar year of 1975 is primarily illustrated, but additional data are plotted for the winter months of 1974 and 1976 above and below the main curve.

A different method of presentation is chosen for the detailed study by theodolite of one of the uplift events (E, Fig. 2) in Figure 3. The horizontal velocity component of three stakes C1 to C3 at distances of 85 m, 135 m, and 255 m from the margin (Fig. 1) is given. Air temperature is plotted at the top to give a clue to the melt-water influx.

4.2. *Main features*

As far as concerns the uplift phenomenon, two major periods can be discerned, an active summer period of seven months duration and a quiet winter period some five months long. Both periods show a characteristic movement pattern. Referring to Figures 2 and 3, we can list the main features as follows;

(a) There is a remarkably *stable downward movement* at almost constant velocity through winter (A_0–A–B; G–H–I).

(b) The straight line representing the winter motion of 1974/75 is interrupted during the

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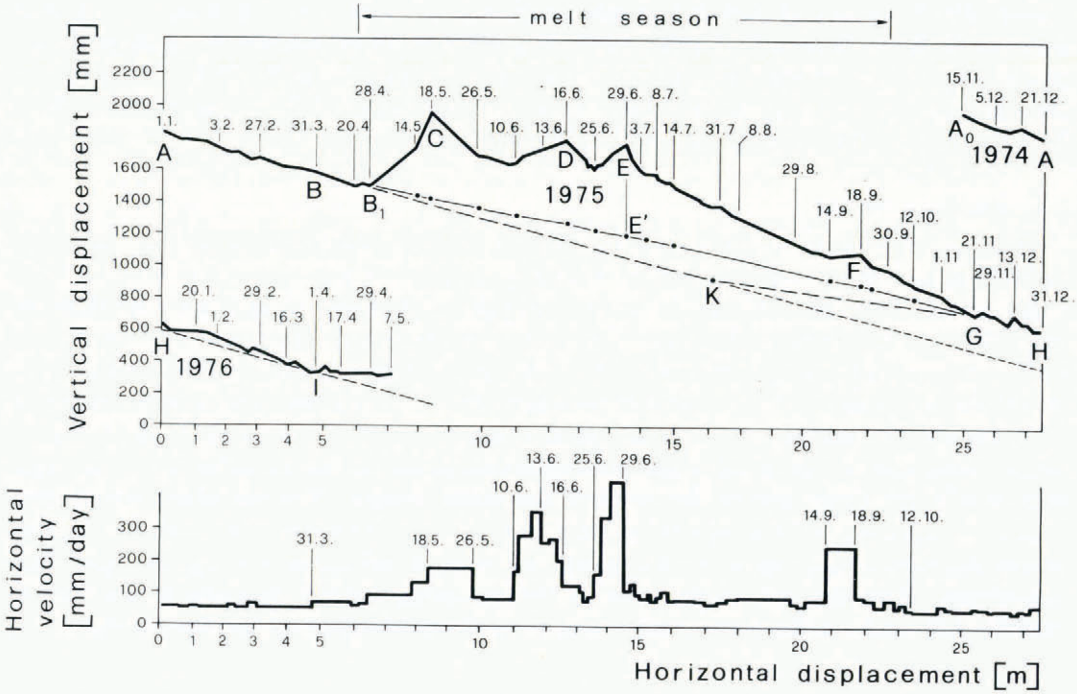

Fig. 2. Camera record of glacier movement. Top: Trace of flow line in a vertical plane. (Vertical scale fivefold exaggeration). Bottom: Plot of horizontal component of velocity versus horizontal displacement.

active summer period. It continues *at the same slope* in the following winter of 1975/76, but is *offset* to the right (upward) as compared to the extension of line A–B.

(c) The active summer period is characterized by an *extended uplift* also referred to as a "bulge".

(d) Superimposed on the bulge are three *short-term major uplift events*, referred to as the peaks C, D and E. Also present are one lesser peak F and a number of minor fluctuations; the latter may be real or may represent scatter of the measurements.

(e) The velocity is generally higher during the extended uplift than in winter, in particular from 31 March (B) till about 12 October.

(f) The velocity fluctuates, showing high peaks which are related to the uplift events. Note that the *velocity maxima* occur during the upward motion only (events D–F). For peak E this is illustrated in detail in Figure 3. (Event C is an exception discussed later).

(g) A typical asymmetry appears in the variation of vertical displacement with time (centre of Fig. 3): the upward movement is fast, the downward movement slow. This holds both for the short-term uplift events and for the bulge as a whole.

Apart from the features listed above, it is important to notice the fact that the uplift events are related to periods of strong melt, and that the uplift events do not appear to be a local phenomenon. Events D and E, the only ones measured at different cross-sections, occurred almost simultaneously at all four cross-sections A–D (Fig. 1). However, there was a slight time delay—less than one day, just detectable in the survey—between the uplift peaks at different profiles. This time delay increased down-glacier.

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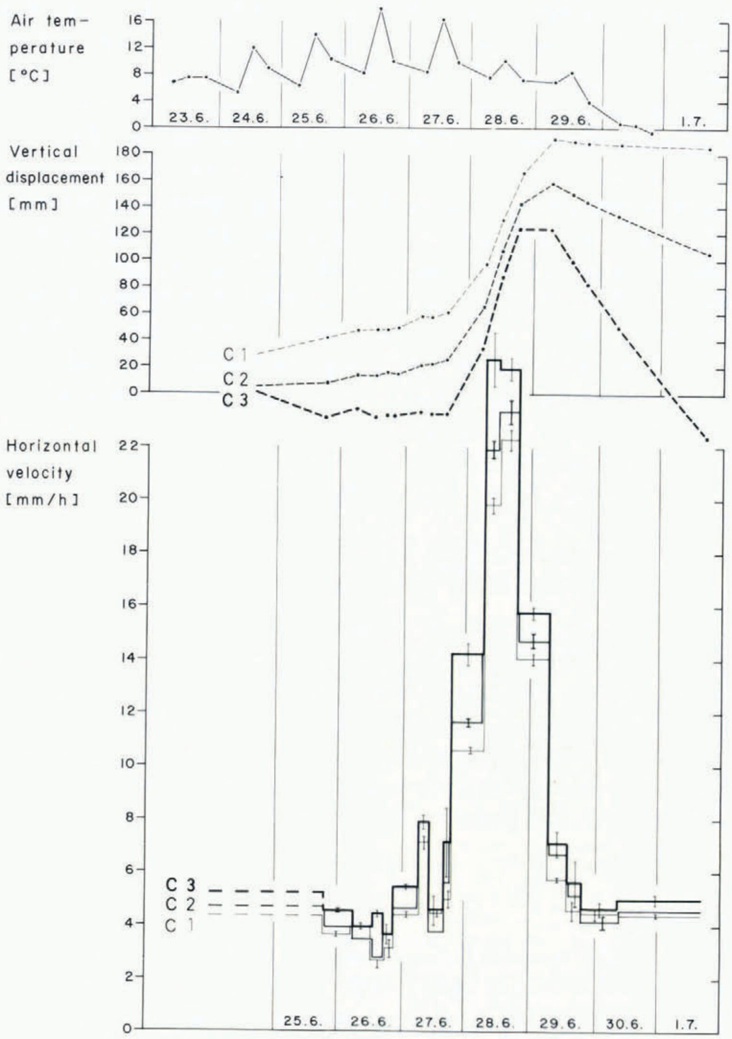

Fig. 3. Detailed theodolite measurements of glacier movement indicating an "uplift" which followed a brief period of strong melt. Top: Air temperature. Mid: Vertical displacement of 3 poles, C1 to C3; C1 is near the glacier margin, C3 on the medial moraine. Bottom: Horizontal velocity of the same poles.

5. Interpretation

5.1. *Concept of water storage at the bed, sliding motion and uplift*

The explanation we are suggesting for the uplift in the following paragraphs is based on the assumption, confirmed on various occasions by observation, that the vertical displacement of the glacier surface is not a local phenomenon, but is representative for an appreciable section of the glacier. The rise then is equivalent to an increase in volume. Since the volume expansion takes place in the melt season, it is a reasonable assumption that it is primarily caused by melt water.

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i.e. *water storage*. Glacial water storage has also been proposed by Stenborg (1970) and by Tangborn and others ([1975]) in order to explain a deficit in run-off apparent in water-balance studies. As the primary discontinuity below the glacier surface is the glacier bed, this is also the most likely place where water storage can occur, either in an extended sheet or layer, or else in discrete cavities. The first possibility can, under most conditions, be ruled out because of the implied thickness of the layer necessary to account for the considerable amount of water storage, and hence the water is expected to occur in cavities. Large openings between ice and rock are often observed at the edge of sliding glaciers, and, given sufficient water pressure, there is no reason to doubt that they may extend in a similar fashion all the way down to the largest depths.

One of the authors (Iken, 1981) has studied in a simplified numerical model the mechanism of sliding on a uniform wavy bed with and without water-filled cavities, and during the transient stages when cavities grow or shrink. The most important finding was that the influence of the water on velocity is much greater during the transient stages than under steady-state conditions. In particular the sliding velocity is largest at the early stage of cavity growth and is smallest, or even negative, when cavities shrink. In the same paper, a critical pressure where sliding becomes unstable is derived and discussed. It is a water pressure well below the ice overburden pressure:

$$P_{\text{crit}} = P_0 - \frac{\tau}{\tan \delta_{\max}},$$

where P_0 is the ice overburden pressure, τ the basal shear stress, and $\delta_{\max}$ the angle between the steepest tangent on the stoss faces of bed obstacles and the mean bed slope. This holds for an idealized glacier bed, where no tangential stresses occur at the ice–rock interface and where the water has access to the lee faces of all bed undulations. The equation has been derived for two-dimensionally rough beds ($X-Z$ plane), but also holds for three-dimensionally rough beds under certain conditions. If the subglacial water pressure exceeds the critical value, the glacier is pushed upward as a rigid body along the stoss faces of the bed undulations. This type of sliding is illustrated in Figure 4.

5.2. *Steady winter flow*

The first feature (a) listed in section 4 is the steady downward movement in winter. From the time of the year and the lack of any significant fluctuations we conclude that the glacier is not influenced by changes of the water regime during this period. The surface movement then represents primarily the flow of the glacier by internal deformation. (We neglect the sliding at the bed under steady-state winter conditions, which is probably small). The average horizontal and vertical displacements and velocities are given in Table I together with the inclination of the flow vector. The two winter periods show consistent results, and so they are averaged for further consideration. In addition, the table includes summer data.

Fig. 4. Extreme case of upward sliding. The dashed line shows a later stage. The gap bounded between the two sine waves shown by solid and dashed lines, respectively, represents the transient shape of a fast growing cavity.

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TABLE I. VELOCITY COMPONENTS IN DIFFERENT SEASONS

Time interval	*Horizontal displacement* m	*Vertical displacement* m	*Horizontal velocity* mm/d	*Vertical velocity* mm/d	*Inclination of movement vector* deg
15 November 1974 to 31 March 1975	7.31	−0.40	53.7	−2.9	−3.1
21 November 1975 to 1 April 1976	6.89	−0.35	52.2	−2.7	−2.9
Mean over the two winter periods:			53.0	−2.8	−3.0
31 March to 21 November 1975	20.53	−0.88	87.4	−3.7	−2.5
Difference: summer velocity (31 March to 21 November 1975) minus mean winter velocity			34.4	−0.9	−1.5

5.3. *Total bed-slip in summer*

We make the preliminary assumption that during the melt season the rate of internal deformation of the bulk of the glacier is the same as in winter. In that case the movement due to ice flow can be extrapolated from winter. Continuation of winter flow, plotted with a dashed line in Figure 2, would bring a reference point which is at B on 31 March to K on 21 November, but the observed position for that date lies at G instead. Under the assumption that the additional displacement from K to G is caused by bed-slip alone, a value of 8.08 m is found for the horizontal sliding distance. If we visualize bed-slip as the movement of a block of ice moving along like a sled on the bed, then the connecting line K–G should be parallel to the glacier bed (allowing for the vertical exaggeration in the figure). From Table I the mean inclination of the bed-slip vector is found to be 1.5°. Note that although water storage is essential for the actual sliding process as discussed later, it does not intervene when we compare positions of the glacier from before and after the uplift period. According to the seismic survey (Knecht and Süsstrunk, unpublished) the bed is essentially horizontal, but as the survey lacks detail in the actual area of the movement studies, a direct verification of the 1.5° slope is not yet possible.

The inclination of the flow vector at the glacier surface amounts to 3°, thus differing significantly from the dip of the bed of about 1.5°. This probably indicates lateral spreading of the high medial moraine on which the camera target was placed. The convergence of the surface displacement-vector and bed is the reason for the shift to the right of the movement lines in Figure 2 from one winter to the next during enhanced sliding in summer. This causes point G to be located about 0.2 m above the extension of line A–B and explains feature (b).

In Figure 2 the dashed line B–K–G shows the effects of steady ice deformation and overall bed-slip separately, one after the other. An additional thin solid line B_1–E′–G gives the simultaneous effects of deformation flow and bed-slip when these are combined for some intermediate time increments. This line serves as a base against which the uplift is measured, e.g. E′–E.

5.4. *Water-induced uplift and sliding velocity*

Although the total dislacement by summer sliding is directed parallel to the bed, the actual motion goes alternately up and down at variable inclinations; it is this particular behaviour that suggests the existence of water storage, variable with time. The extended uplift, feature (c),

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implies long-term storage, whereas the short-term uplift events, feature (d), suggest a more rapidly changing water storage.

5.4.1. *Effects of cavities and cavity growth on the sliding velocity*

The water storage is closely linked to velocity variations. During the period of water storage, small irregularities of the bed (obstacles) become flooded, and the bed becomes effectively smoothed; the result is reduced friction (Weertman, 1962; Lliboutry, 1968, 1979; Kamb, 1970). This explains why the velocity is generally higher in summer than in winter, feature (e). The strong velocity variations, in particular the sharp peaks and their relation to the uplift effect, feature (f), however, cannot satisfactorily be explained by the quoted steady-state theories: the velocity maxima indeed occur while the glacier is rising and not when water storage is at a maximum, as is clearly shown in the case of peaks D, E, and F. This unexpected finding has been confirmed by further observations at other profiles, in other years, and on other glaciers. It agrees well with the results of the model study on transient phases summarized in paragraph 5.1, thus providing one of the strongest arguments in favour of the water-storage concept.

5.4.2. *Relation between uplift characteristics and water-drainage conditions*

A characteristic feature of the uplift peaks is the asymmetry in a diagram of vertical displacement versus time (Fig. 3), feature (g). The upward movement sets in abruptly, while the downward movement is much slower and gradual. From this trend can be inferred a characteristic trait of the subglacial water-carrying system. In the latter we include all the passageways supplying water to the opening-up cavities, or draining water from the closing ones. Unlike the model study, in which an instant unlimited water exchange between the bed and (external) reservoirs or sinks has been assumed, under natural conditions the water exchange is time dependent and limited. In relation to the uplift asymmetry we now postulate that the performance, i.e. capacity of the water carrying system is controlled by the subglacial water pressure: During uplift the subglacial water pressure is high and opens up rapidly a network of interconnecting passageways along the ice sole by spontaneous ice separation. Once the water pressure drops, the majority of these passageways closes off and the stored water essentially is trapped. The subsequent slow drainage corresponding to the gradual downward movement of the ice is probably controlled either by drainage through a porous substratum, or seepage through tiny grooves in the rock. During this stage, variations of water pressure in the main drainage channels should have practically no effect on the sliding velocity as long as the pressure stays low. Indeed, during many weeks of the melt season the camera record shows no velocity variations in spite of variations of melt-water input to the glacier.

A seasonal variation of the degree of interconnection of subglacial cavities has also been proposed by Lliboutry (1978), not in relation to storage of surface melt water, however. Lliboutry assumed a gradual slow growth of isolated, water-filled cavities due to melt of ice at the base by geothermal heat and due to melt in the interior of the glacier by energy dissipation in the deformation process ("autonomous system"). He proposed that the isolated cavities would become interconnected in the melt season, when the subglacial water pressure is high, and that then the cavities would drain into the main drainage channels ("interconnected system").

5.4.3. *Special conditions at the beginning of the melt season.*

The concept of limited (and irregular) water supply to the bed also helps to a better

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understanding of the initial uplift phenomena at the beginning of the melt season. At first sight, event C shows a velocity peak from 18 to 26 May concurrent with a *downward* movement, in contrast to the later peaks. Since one measurement is missing (the time interval being eight days instead of the standard four days), this finding should be treated with reservation—a higher uplift is conceivable which takes place concurrently with a high velocity peak in the early part of the interval and which is followed by a steep downward movement at very low velocity later on—but there is another significant difference in peak C as compared to later events: The initial rise from 28 April till 18 May is not only exceedingly steep, but it also occurs at a relatively low horizontal velocity. The explanation we would like to put forward is that with the first input of melt water the water pressure rises locally to extremely high levels (Röthlisberger and Iken, 1981). In this case no gradual, tangential cavity growth occurs, but spontaneous ice separation where the water reaches overburden pressure. In consequence the average uplift becomes considerable. The horizontal velocity, however, is not much affected, because the glacier is still anchored in those areas not already reached by the water.

5.5. *Size and shape of* roches moutonnées

The base line for the uplift B_1–E′–G of Figure 2 is according to us at the same time the "reference line for water storage". It is the line showing the movement at the glacier surface which would occur if sliding were to take place without separation from the bed. The difference in height of the heavy and light solid lines connecting B_1 with G, the uplift, represents the mean thickness of the water storage. At the uplift peaks C, D and E, the maximum storage reaches values of 0.55, 0.53, and 0.48 m, respectively. Between 14 May and 1 August storage is never less than about 0.3 m.

The large quantity of storage water raises the question of how this water can be accommodated at the glacier sole without starting a catastrophic advance of the glacier. Obviously, large bed undulations are a prerequisite. Based on that part of the total uplift which decayed only slowly, we conclude that long-term water storage in 1975 amounted to about 0.3 m, and under the supposition that all the water is stored in cavities of uniform size and shape we can estimate their volume, assuming a certain simplified type of bed, for instance sinusoidal undulations. In this case the largest possible *steady* cavities extend from the crest of the undulations to the inflexion point on the next stoss face down-glacier. Figure 5 shows a simplification of such a cavity (the sharp corner at its down-stream edge is not realistic). The cross-section of that cavity is very closely $F = \lambda a(3 + 4/\pi)/8 = 0.534\lambda a$, where λ is the wave length and a the amplitude (defined as one half of the difference between the maximum and the minimum values of the sine function). The corresponding net uplift is $\Delta h = F/\lambda = 0.534a$. Thus, if $\Delta h \approx 0.3$ m, the amplitude a of the undulations has to be equal to or greater than 0.56 m. Such an amplitude, of even 1 to 2 m, would be an acceptable size; polished rock ridges of this size are now exposed at the margin near the glacier tongue. However, it appears highly improbable that such long cavities as shown in Figure 5—they cover 75% of the bed—could be maintained

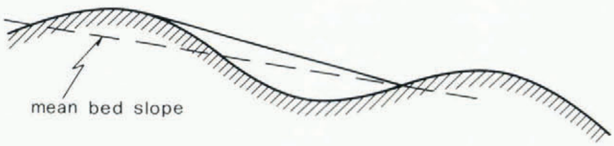

Fig. 5. Maximum possible steady cavity (cavity shape simplified).

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without sudden drainage long enough to match the observed small rate of downward movement. Furthermore, when cavities cover 75% of the bed, one would expect a strong reduction of the sliding friction and consequently a large increase of the sliding velocity during the period of storage. This has not been observed: during the melt season the velocity was, apart from the major peaks, only approximately 30–40% higher than in winter. It is therefore sugested that the cavities were more compact (short and thick) than that depicted in Figure 5. Thus we are lead to assume that either the amplitude of the undulations was substantially larger than 0.56 m, or that the bed consisted of asymmetrical humps with very steep lee faces, i.e. *roches moutonnées*. In the more compact cavities the water pressure would be well below the critical pressure. Even if a large water pressure is acting during a short time at the beginning of cavity growth, and if the connections between the cavities and the subglacial drainage system then become shut off before a steady state is reached, the pressure in the isolated cavities will decrease as cavities become more compact. The final, steady value is determined by the volume of entrapped water.

Additional information on the shape of undulations at the glacier bed can speculatively be drawn from observed angles of upward sliding: the larger the subglacial water pressure, the steeper upward is the sliding motion during the process of cavity growth (a result of the numerical modelling quoted). In the limit, under ideal conditions, a rigid-body motion parallel to the steepest tangent on the stoss faces of bed undulations does take place as illustrated in Figure 4. In Nature this situation may be approximated during short time intervals of fastest sliding. The highest velocity peak of 1975 has been evaluated in this respect: Between 27 June, 15.30 h and 28 June, 20.00 h the horizontal component of surface velocity amounted to 451 mm/d and the vertical component to 116.5 mm/d. Assuming that the sliding velocity is zero during winter and that the rate of internal deformation of the ice is constant throughout the year (as assumed earlier for the entire summer period), the sliding vector can be inferred. In our particular case it is found that sliding occurred at an angle of 16.5° to horizontal, or 18° to the mean bed slope. These figures may give an indication of the slope of the steepest stoss faces numerous enough to determine the direction of overall sliding. For the special case of a sinusoidal bed, a wavelength of 11 m would follow from the values of amplitude and maximum slope of stoss faces derived above, and the roughness parameter would amount to 0.05. In the case of asymmetrical *roches moutonnées*, the wave length would be less and the roughness parameter larger.

6. Alternative explanations

6.1. *Variations of longitudinal strain-rate*

In principle, variations of vertical velocity can also be understood by assuming variations of straining of the glacier with time. Clearly, changes of straining can be induced by changes of the sliding velocity.

Consider, for instance, a zone of the glacier just up-stream of a narrow pass in the valley. There the flow is compressive. When the horizontal velocity increases, the longitudinal compression at this site is likely to increase as well and consequently the ice expands vertically which will appear as an "uplift". However, when the horizontal velocity reassumes the value which it had before the "uplift", the strain-rate should also attain its former value (not a lower one) and thus the observed downward movement of the glacier, following the actual uplifts, cannot be understood in this way. Another possible cause of variations of longitudinal strain-rates is local variations of velocity. For instance, if a zone of increased velocity travels down the

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glacier, a compressive zone moves in front of it and an extending zone in the rear. In this case, measurements of surface velocity carried out at a certain location would show an upward movement during the time of increasing, and a downward movement during the time of decreasing, horizontal velocity.

Surface velocities measured in June 1975 at three transverse lines (B–D, Fig. 1) have been analysed in order to estimate large-scale average strain-rate variations and related variations of vertical velocity at line C; this assessment is given in the Appendix. In the analysis the slight widening of the valley width down-glacier of the C-profile and the deviation of the mean bed slope from the horizontal are also taken into account. The result is that the calculated variations of vertical velocity are not nearly in accord with the observed ones. From this we conclude that the variations of vertical velocity, observed in June 1975, cannot be explained in their major part by temporary straining of the ice.

For a similar analysis of the long-term uplift (May to November 1975) insufficient data exist. However, it appears unlikely that the upward movement in May and the gradual downward movement till November can be explained by straining of the ice. What one usually observes is that the melt season starts at the glacier tongue and triggers the first velocity increase there. Higher up on the glacier, snow melt and the first seasonal velocity increase, take place later (e.g. Iken, 1974). This sequence of events should cause an extra downward movement rather than an uplift in spring.

To sum up, it is very unlikely that the uplift phenomenon can be explained by variable strain-rate. A very complicated alternation of positive and negative strain-rates with time would be needed for each uplift peak. This can be justified neither by theoretical considerations nor by existing observations. When and to what degree straining may contribute to (or diminish) the uplift is nevertheless still an open question.

6.2. *Crevasse formation*

The opening-up of crevasses undoubtedly increases the volume of the glacier, tending to cause an upward movement of the glacier surface. Since the glacier was still snow-covered in 1975 when the uplift occurred, volume changes by crevasse formation could not be assessed. A survey of crevasse volume was however carried out in August 1977, i.e. at a time when the uplift had already begun to fade, between profiles B and C. Crevasse width at the surface was measured together with the depth where the width had decreased to 0.03 m. The total crevasse depth was assumed to be 40 m which is in general probably an overestimate. The intensity of crevassing varied locally. In the marginal zone it was in some places sufficient to account for an increase of surface elevation by 0.16 m, in other places only by 0.001 m. In the central part of the glacier crevassing was negligible. The uplift, on the contrary, was most pronounced at the central pole, C_3. This suggests that crevassing was not a major cause of the uplift.

The effect of crevassing is probably somewhat larger at the beginning of uplift, when fresh crevasses open up in large numbers, some even extending all the way down to the bed due to water pessure (Deichmann and others, 1979; Röthlisberger, 1980). Although crevassing may contribute to the uplift, we do not expect that it causes a major part of it. The water-filled crevasses seem to open up rapidly and drain subglacially without getting very wide. They are, in our opinion, directly responsible for a minor part of the increase in volume only, but are very important for supplying at the bed the water that causes the uplift. Different conditions may

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nevertheless occur close to the bed in zones of strong local extension, where bottom crevasses might possibly form.

6.3. *Swelling of veins at grain edges*

Water-filled veins, situated at three-grain intersections in the ice, have been investigated by Lliboutry (1971), Nye and Mae (1972), Nye and Frank (1973), and Raymond and Harrison (1975). If a permeable network of such veins were to exist throughout the glacier—and this has been questioned by Lliboutry (1971) and Nye and Mae (1972)—an increase of glacier thickness as a consequence of a swelling of the veins is in principle conveivable. If vein cross-sections can adjust quickly enough to a transient increase of melt-water supply on the glacier surface, this can cause the propagation of a kinematic wave of constant flux downwards. The microscopic observations on core samples taken from depths below 7 m from Blue Glacier have, however, revealed a vein network which accommodated at most a flux per unit area of 0.1 m a^{-1} and thus could not respond to the availability of surface melt water (Raymond and Harrison, 1975). In two of their cores, Raymond and Harrison found large-scale, tubular "conduits" with diameters of a few millimetres, which could provide more efficient water access if they were numerous enough. This is not known. More, it is doubtful whether these tubular features could adjust their size sufficiently fast to account for an uplift. A further hindrance to the functioning of such small-scale drainage mechanisms in the present problem is the fact that the ice temperature at the top layer was undoubtedly still below the freezing point during the time of uplift over most of the glacier surface.

6.4. *Shearing and related volume change of subglacial sediments*

Without knowledge of the type and thickness of possible subglacial sediments only certain general deformation characteristics can be outlined and compared to observations.

The discussion is restricted to coarse granular sediments since the existence of a large clay (or silt) content can be ruled out: it would impede subglacial water flow. The fact that the uplift events occurred within a few days after periods of strong melt implies that subglacial sediments lying on an impermeable bed would have to be quite permeable.

Typical values of friction factors, f, defined as the ratio of shear stress and normal stress on the failure plane at a failure, range from 0.8 to 1.1 for densely packed granular sediments and from 0.6 to 0.7 for loose packing. Here "dense" packing corresponds to void ratios between 0.3 and 0.55, "loose" packing to void ratios between 0.6 and 0.85. Once shear deformation is initiated the dense sediment will increase in volume (a very loose one would contract) until a certain critical void ratio is reached which is typical of the particular soil. There will be no further volume change when shearing of the sediment continues. The critical void ratio is nearly independent of normal effective stress. (Roscoe and others (1958) report a decrease of the critical void ratio by only 3 to 6% when, in shear tests with steel balls or glass beads, the normal stress was increased from 0 to 7 bar. The corresponding volume changes are 1 to 2.5%.) Critical void ratios of granular substances quoted by Lambe and Whitman (1979) and by Roscoe and others (1958) are in the range of 0.6 to 0.85.

On the basis of these facts we will now examine whether the uplift features can be understood as being caused by sediment deformation. From the friction factors given above it is obvious that sediment deformation can only take place when the subglacial water pressure is quite high; given a basal shear stress of 1 bar the effective stress $\sigma_e = \tau/f$ ought to be less than approximately

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$1/0.8 = 1.25$ bar if the sediment is dense, and less than $1/0.6 \approx 1.7$ bar if the void ratio is at the critical value. The high water pressures corresponding to such low effective stresses do not continuously exist in the subglacial drainage system during summer and hence, at best, the short-term events of very high velocity might be due to sediment shearing. A further difficulty arises when trying to explain the substantial downward movement following the uplift peaks: In general the compaction of a soil requires a series of repeated loadings and unloadings and it is doubtful whether this can be accomplished by a few variations of subglacial water pressure. Perhaps very small density changes are possible. This implies that significant volume changes of the sediment can occur only—if at all—where the sediment layer is quite thick. In no way, however, can the typical pattern of the uplift events be explained on the basis of this hypothesis: The observations have revealed that sliding velocities are highest when the rate of uplift is largest and not when the maximum vertical displacement is established. In contrast, for the shearing of sediments the smallest shear stress is required when the maximum volume has been attained.

In summary, the characteristics of sediment deformation do not well match the observed uplift features.

7. Conclusions

The variation with time of the movement of a surface point has been studied in detail on Unteraargletscher and a complicated pattern of displacement has been found when analysing both the horizontal and vertical components. Various characteristic trends and features of the movement variations can satisfactorily be explained with a conceptual model based on the temporary formation of water-filled cavities on the lee-side of bed undulations, apart from the mere fact that water storage at the bed automatically implies an *uplift* of the glacier surface. The main aspects of the conceptual model are as follows: The process of water storage depends primarily on surface melt water and rain. This water is stored at the glacier bed when the subglacial water pressure is sufficiently high to open up a branching network of flat, interconnecting passageways along the ice sole. The surface melt water, which arrives at the bed via moulins and fresh crevasses, is then transferred through the passageways towards the growing cavities. When the subglacial water pressure drops, the majority of the interconnecting passageways become blocked and the water is trapped in the cavities. This concept is compatible with a lasting water storage in spring, and corresponds to the observed upward movement during periods of melt-water abundance. It also explains why the influence of water pressure variations in the main drainage channels on the sliding velocity was limited to a few events, presumably with very high water pressure. During these events the velocity peaks occur simultaneously with the upward motion only, i.e. while cavities are opening up. This behaviour is in good agreement with numerical results obtained for idealized cavities which are growing and shrinking, but it deviates from the commonly held view of a direct dependence of velocity on the amount of stored water. Such a dependence also exists, but it is of much less importance than the sliding events caused by the water-pressure peaks. From the large amount of water storage of up to 0.6 m we conclude that the bed is formed by large-size *roches moutonnees* (or boulders) and that the bed is probably quite rough. Water storage begins abruptly at the beginning of the melt season while the release of water is slow and ends in November. During the intervening time span the glacier moves up and down on the bed irregularities, depending on the changing water storage, but the total displacement by sliding occurs along the mean bed.

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The strongest argument in favour of the above ideas derives from the fact that a variety of observed details of the glacier movement can be explained by bed processes which are likely to occur. This seems far more difficult with alternative concepts. Nevertheless, other processes, such as crevassing (including water-filled bottom crevasses?) temporary strain, deformation of unconsolidated sediments, etc., may contribute. Modifications of the model itself are indicated because a real bed consists of both bedrock and unconsolidated sediments. The topography is rough in three dimensions and the irregularities are not monochromatic. The amount of uplift certainly poses a problem since very large irregularities are needed to render possible the total amount of observed vertical displacement as a consequence of cavity formation.

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APPENDIX

Variations of the vertical component of velocity deduced from the measured variations of the horizontal component

Changes of vertical ice velocity can be caused by changes of longitudinal strain-rate and changes of the sliding velocity in a channel of variable width and with a bed slope differing from the horizontal direction. The following analysis is based on continuity considerations. We assume incompressibility of the ice, neglecting here volume changes due to the formation of crevasses. From the assumption of incompressibility it follows that

$$-\frac{\partial v_z}{\partial z}=\frac{\partial v_x}{\partial x}+\frac{\partial v_y}{\partial y} \tag{A–1}$$

where v_x, v_y, and v_z are velocity components; the x-axis is horizontal and points in the direction of flow, the z-axis points vertically upward (see Fig. A–1). Integration of Equation (A–1) between $z=z_0$ (glacier bed) and $z=z_h$ (glacier surface) gives

$$-\{v_z(z_h)-v_z(z_0)\}=\left(\overline{\frac{\partial v_x}{\partial x}}+\overline{\frac{\partial v_y}{\partial y}}\right)h \tag{A–2}$$

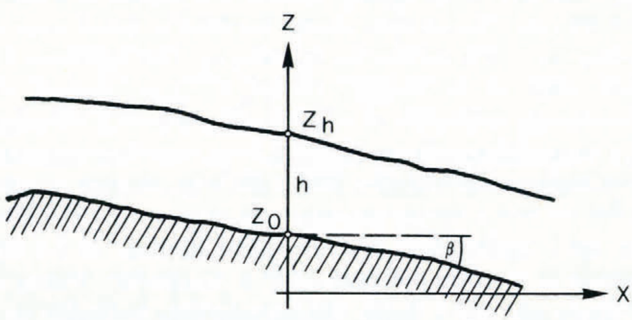

Fig. A–1. Sketch of glacier showing direction of coordinate axes.

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where the bars denote averages over the thickness h. $v_z(z_0)$, the vertical component of the sliding velocity, is related to the bed slope β, by

$$v_z(z_0) = -v_x(z_0)\tan\beta; \tag{A–3}$$

β is taken positive where the bed slopes downward towards the terminus. Equation (A–3) holds if the sliding velocity is parallel to the bed, i.e. if cavity formation is not considered. The term "sliding velocity" stands for the mean value of sliding velocities taken over an area of the bed with a diameter of, say, 200 m; β is a mean value for the same area. Near the margins the variation of the sliding velocity along a transverse profile can be very large (Raymond, 1971) but we restrict the analysis to an area at the medial moraine.

In general, the strain-rates averaged over the thickness are not equal to the surface strain-rates, however, we only need to consider temporal changes of strain-rates due to changes of the sliding velocity. We assume that, near the centre-line of the glacier, short-term changes of horizontal velocity are essentially independent of depth and equal to changes of the horizontal component of the sliding velocity,

$$\Delta v_x(z_h) = \Delta v_x(z_0) \tag{A–4}$$

(here the symbol Δ refers to a time interval Δt). It is plausible then to assume that the changes of longitudinal strain-rate $\Delta(\partial v_x/\partial x)$ are independent of depth so that

$$\Delta\,\overline{\frac{\partial v_x}{\partial x}} = \Delta\,\frac{\partial v_x(z_h)}{\partial x}. \tag{A–5}$$

During the considered short time intervals, the ice thickness h does not change, and thus, from Equations (A–2) to (A–5), it follows that

$$\Delta v_z(z_h) = -\Delta v_x(z_h)\tan\beta - h\Delta\,\frac{\partial v_x(z_h)}{\partial x} - h\Delta\,\overline{\frac{\partial v_y}{\partial y}}. \tag{A–6}$$

The first term on the right-hand side is the change of vertical velocity due to a change of the sliding velocity on the inclined bed, the two other terms are the contributions of the changes of longitudinal and transverse strain-rates.

Evaluation of Equation (A–6) for pole C_3 on the medial moraine

(a) Changes of longitudinal strain-rate $\Delta(\partial v_x/\partial x)$

Approximate longitudinal strain-rates at pole C_3 were calculated from velocity data of three poles on a flow line, B_3, C_3, and D_2, by fitting a parabola $(v_x - v_{x0})^2 = 2p(x - x_0)$ through the points (v_{xi}, x_i). v_{x1} is the velocity of pole B_3, v_{x2} of pole C_3 and v_{x3} of pole D_2. The x_i correspond to the positions of the three poles on a flow line, v_{x0} and p are suitable constants. (Numbering of poles along a transverse line starts with 1 at the margin.)

The times of movement measurements were different at the different sites. Interpolation was necessary to derive mean velocities over the same time intervals at the three sites x_i. Errors due to this procedure were estimated and added to the errors of surveying. For all combinations of velocity values (v_{xi}, v_{xi} + tolerance, v_{xi} − tolerance) polynomials were determined, and from these the longitudinal strain-rates at pole C_3 were calculated. The largest of the results is given as "upper limit" in column 3 of Table A–I, the smallest as "lower limit" in column 4. The values calculated from $v_{xi} \pm 0$ are listed in column 2. A comment concerning the validity of this assessment of the strain-rates is due: the distances of the three profiles may be too large to permit conclusions on the mean strain-rate over a distance of, say, the order of the ice thickness at the H-profile. However, two facts in support of the procedure can be quoted:

1. The mean yearly velocity (1973/74) along this flow line, inferred from air photographs, decreased gradually with increasing x (measurements by A. Flotron 1974 archived as Vermessung der Aargletscher, Aufnahmen 1974. Archiv der Kraftwerke Oberhasli AG, Innertkirchen). Thus, for the velocity averaged over one year, the interpolation procedure is not unreasonable. This need not be so for the short periods considered here.

2. Together with pole C_3, the pole with the camera target at a distance of 72 m in flow direction was also surveyed. From these measurements the mean strain-rate over the section of 72 m can be calculated. For the period of 12 to 26 June the result is well above the error limit and in good agreement with the strain-rate value based on interpolation (see Table A–I). For the shorter periods considered the accuracy of surveying was insufficient for this particular purpose.

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TABLE A–I: VARIOUS TERMS OF EQUATION (A–6) FROM WHICH VERTICAL VELOCITY VARIATIONS AT POLE C_3 WERE CALCULATED (COLUMN 9) AND THE MEASURED VERTICAL VELOCITY VARIATIONS OF THE SAME POLE (COLUMN 10)

1	2	3	4	5	6	7	8	9	10
Period	$10^5 \times \frac{\partial v_x}{\partial x}$ d^{-1}	*upper limit* d^{-1}	*lower limit* d^{-1}	$-h\Delta \frac{\partial v_x}{\partial x}$ mm d^{-1}	$v_{x2}(z_h)$ mm d^{-1}	$\Delta v_z(z_0)$ mm d^{-1}	$-\left(h\Delta \frac{\overline{\partial v_y^{\star}}}{\partial y}\right)$ mm d^{-1}	$\Delta v_z(z_h)$ *calculated*† mm d^{-1}	$\Delta v_z(z_h)$ *measured* mm d^{-1}
5 June, 15.00– 9 June, 18.00	−3.03	−3.37	−2.89		81.7 ± 2.2				
				+4		−3	−2	+1	+30
9 June, 18.00– 12 June, 11.00	−4.23	−4.54	−3.69		207 ± 3				
				−1		+1	+1	0	−35
12 June, 11.00– 26 June, 09.00	−3.81*	−3.86	−3.76		150.9 ± 0.1				
				+23		−1	−1	+22	+43
26 June, 09.00– 28 June, 08.30	−11.1	−12.5	−10.1		197 ± 6				
				−27		−7	−4	−34	+21
28 June, 08.30– 29 June, 10.00	−2.5	+1.2	−6.8		469 ± 16				

* This value can be compared with a mean strain-rate measured over a distance of 72 m in flow direction, amounting to $(-3.8 \pm 0.2) \times 10^{-5}\ d^{-1}$ for the period 12 June, 07.40 h to 26 June, 05.50 h.

† Values in column 9 were calculated assuming $\Delta(\overline{\partial v_y/\partial y}) = 0$.

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(b) Changes of transverse strain-rate $\Delta(\overline{\partial v_y/\partial y})$

No measurements have been made regarding the transverse strain-rate. If the valley walls were vertical, the glacier surface plane, the channel width constant, and the lateral variation of the longitudinal strain-rate zero, there would be no flow in y-direction and $\Delta(\overline{\partial v_y/\partial y})$ would be zero. Actually the valley walls slope at 45° and the channel width is not constant but, at the Camera Profile, slightly increasing in the direction towards the glacier terminus. While the effects of the inclination of the valley walls and of changes of the convexity of the glacier surface are negligible, the widening of the channel cross-section could cause, together with fluctuations of the velocity in x-direction, significant variations of lateral strain-rate. In order to obtain a rough estimate of this part of $\Delta(\overline{\partial v_y/\partial y})$, labelled in the following by a star, we simplify the problem as follows: We replace the channel by one of equal cross-section to that in reality, but of rectangular shape. Further, we assume that the ice always fills the total width of the channel and moves everywhere in the x-direction with a velocity approximately equal to three-quarters of the surface velocity measured on the medial moraine. $(\overline{\partial v_y/\partial y})^\star$ is presumably larger near the margins than at the centre-line, but we assume it to be independent of y. We then obtain

$$\Delta\left(\overline{\frac{\partial v_y}{\partial y}}\right)^\star = \frac{3}{4}\,\frac{\mathrm{d}W}{\mathrm{d}x}\,\frac{\Delta v_{x2}}{W_c}. \tag{A–7}$$

For the gradient of channel width $\mathrm{d}W/\mathrm{d}x$, a mean value, taken over a distance of 500 m, was determined from the bedrock map by Knecht and Süsstrunk (unpublished). W_c is the channel width at the considered cross-section.

Numerical values of the terms of Equation (A–6) are assembled in Table A–I.

Comments on Table A–I

Column 1, Dates of measurements.

Columns 2–4, Longitudinal strain-rate and tolerances; assessment explained above.

Column 5, $(-h\Delta(\partial v_x/\partial x))$: $\Delta(\partial v_x/\partial x)$ is the difference between two successive values in column 2; $h = 310$ m is the thickness of the glacier at pole C_3.

Column 6, (v_{x2}, horizontal velocity of pole C_3): Mean velocity during the periods given in column 1. The calculation of the velocity means over exactly the considered time intervals required linear interpolation of velocities measured over shorter periods near the interval ends.

Column 7, $\Delta v_z(z_0)$: Numerical values were calculated from Δv_{x2}, the differences of two successive values of column 6 and the bed slope, $\beta = 1.5°$. This figure was obtained from the analysis in section 5.3. It is probably not very accurate, but of little influence.

Column 8, $(h\Delta(\overline{\partial v_y/\partial y}))$: Estimate of a contribution to the change of vertical velocity which results from the change of transverse strain-rate caused by a change of the sliding velocity in a channel of variable width (Equation (A–7)).

Column 9, $\Delta v_z(z_h)$: Calculated changes of vertical velocity at the surface according to Equation (A–6) and assuming $\Delta(\overline{\partial v_y/\partial y}) = 0$. If, instead of this assumption, the values of column 8 are included, no significantly different results are obtained.

Column 10, measured changes of vertical velocity of pole C_3.

Result: Since various simplifying assumptions and interpolations had to be made, the changes of vertical velocity calculated from continuity considerations can be quite inaccurate. However, the discrepancy between these results and the measured variations of vertical velocity is so large that it appears justified to conclude that the vertical velocity variations cannot be explained by this kind of continuity consideration alone.

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