Earth Surface Processes and Landforms
Earth Surf. Process. Landforms (in press)
Published online in Wiley InterScience
(www.interscience.wiley.com) **DOI:** 10.1002/esp.1374

Hydrology and dynamics of a polythermal (mostly cold) High Arctic glacier

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Received 28 February 2005; Revised 20 January 2006; Accepted 9 February 2006

Abstract

To improve our understanding of the interactions between hydrology and dynamics in mostly cold glaciers (in which water flow is limited by thermal regime), we analyse short-term (every two days) variations in glacier flow in the ablation zone of polythermal John Evans Glacier, High Arctic Canada. We monitor the spatial and temporal propagation of high-velocity events, and examine their impacts upon supraglacial drainage processes and evolving subglacial drainage system structure. Each year, in response to the rapid establishment of supraglacial–subglacial drainage connections in the mid-ablation zone, a 'spring event' of high horizontal surface velocities and high residual vertical motion propagates downglacier over two to four days from the mid-ablation zone to the terminus. Subsequently, horizontal velocities fall relative to the spring event but remain higher than over winter, reflecting channelization of subglacial drainage but continued supraglacial meltwater forcing. Further transient high-velocity events occur later in each melt season in response to melt-induced rising supraglacial meltwater inputs to the glacier bed, but the dynamic response of the glacier contrasts with that recorded during the spring event, with the degree of spatial propagation a function of the degree to which the subglacial drainage system has become channelized.

Keywords: glacier motion; subglacial drainage; polythermal glacier; High Arctic; Arctic Canada

Introduction

Coupling between glacier hydrology and ice flow may hasten glacier melting in response to projected 'greenhouse' warming by accelerating the transfer of ice to lower elevations where surface melting and/or calving remove mass (Meier, 1993; Fountain and Walder, 1998; Zwally *et al*., 2002; Parizek and Alley, 2004). The issue is especially pertinent in the High Arctic (>75 °N) where recent climate models predict exceptional warming over the next century (Manabe *et al*., 1991; Cattle and Crossley, 1995; Houghton *et al*., 2001). The many small- to medium-sized glaciers that exist throughout the region (covering *c*. 275 000 km^2) are likely to respond much more rapidly to climate warming than the major ice sheets, and may contribute significantly to eustatic sea level rise over the next century (Prowse, 1990; Dowdeswell *et al*., 1997; Munro, 2000). However, the coupling between hydrology and dynamics has not been considered adequately in recent models of High Arctic glacier response to climate change (Houghton *et al*., 2001), owing largely to limited understanding of such coupling in the mostly cold polythermal glaciers that characterize the High Arctic.

Outside the major ice sheets and ice caps, most High Arctic glaciers have a mostly cold thermal regime, with only a relatively small proportion of basal ice raised to the pressure-melting point (polythermal structure types d and e of Blatter and Hutter, 1991, their figure 1). The predominance of cold ice in such glaciers suggests that supraglacial–subglacial hydrological connections are limited, and therefore that the hydrologically induced flow instabilities commonly observed on temperate glaciers (e.g. Hodge, 1974; Iken and Bindschadler, 1986; Willis, 1995) and mostly warm polythermal glaciers (e.g. Etzelmüller *et al*., 1993; Jansson, 1996; Kavanaugh and Clarke, 2001) are unlikely. However, many mostly cold glaciers exhibit evidence that supraglacial meltwater penetrates to their subglacial drainage systems (Hodson *et al*., 1998; Skidmore and Sharp, 1999; Wadham *et al*., 2001; Boon and Sharp, 2003; Rippin *et al*., 2003) and many display significant short-term and/or seasonal variations in dynamics (e.g. Müller and Iken, 1973; Iken, 1974, 1977; Andreasen, 1985; Blatter and Kappenberger, 1988; Rabus and Echelmeyer, 1997; Copland *et al*., 2003a,b; Rippin *et al*., 2006). These observations suggest that hydrologically forced variations in glacier flow may be common even in mostly cold glaciers, but the details of the hydrology–dynamics coupling in such glaciers are currently lacking. In particular, it is not known whether, or how, a mostly cold thermal regime might impact upon the propagation of high-velocity events through a glacier, nor whether, or how, the subglacial drainage system evolves at such glaciers and how this might impact upon glacier dynamics.

This study contributes towards an improved understanding of the relationship between hydrology and dynamics in mostly cold glaciers by analysing short-term (two-day) variations in glacier dynamics and subglacial hydrology observed at a high spatial resolution over two melt seasons across the lower ablation zone of John Evans Glacier, eastern Ellesmere Island, Nunavut, High Arctic Canada. The objectives of the paper are to: (i) examine the short-term distributions and variations of horizontal (*xy*) and vertical (*z*) surface velocities across the predominantly warm-based lower 4 km of the glacier during two consecutive melt seasons; (ii) determine whether any *xy*- and *z*-velocity anomalies propagate spatially and/or temporally; (iii) evaluate whether spatial and/or temporal variations in glacier dynamics can be related to contemporaneous changes in the structure of the subglacial drainage system; and (iv) assess the extent to which the observed short-term variations in dynamics may impact on long-term mass balance.

Field Site

John Evans Glacier (79°40′ N, 74°00′ W; hereafter JEG) covers *c*. 75 per cent of a 220 km^2 catchment, is *c*. 15 km long, and spans an elevation range of 100–1500 m (Figure 1). Ice is almost 400 m thick near the equilibrium line, but is typically 100 to 200 m thick in the lower 4 km of the glacier. However, it thins to *c*. 40 m over a large bedrock riegel *c*. 4 km upstream from the glacier snout in the mid-ablation zone (hereafter MAZ) (Copland and Sharp, 2001). The glacier consists primarily of cold ice except in its lower 4 km, where ice reaches the pressure-melting point at the bed and in a basal layer up to 25 m thick (Copland and Sharp, 2001). Ice at the margins and terminus is entirely cold. The thermal regime is significant because it enables the existence of subglacial drainage and is geographically widespread across the High Arctic (Blatter and Hutter, 1991).

The climate at JEG is cold and dry. From 1997 to 2002 mean annual air temperature at 824 m a.s.l. was −14·6 °C and mean annual precipitation was <0·15 m. The short summer melt season typically lasts for *c*. 60 days (early June to early August). Subglacial meltwater outflow is prevented during winter by a ‘thermal dam’ of cold ice and/or frozen sediments at the terminus (Skidmore and Sharp, 1999), and is typically restricted to a period of *c*. 40 days during summer. An initial meltwater outburst occurs from the terminus in late June or early July each year, one or two days after the sudden onset of supraglacial drainage into a series of moulins (h1–h5 on Figure 1) located over the bedrock riegel in the MAZ, 4 km upglacier from the terminus (Skidmore and Sharp, 1999) (Figure 1). Following the outburst, outflow typically continues until early August, although it occasionally ceases for short periods during cold weather (Skidmore and Sharp, 1999). Dye tracing has shown that the sediment-free, low electrical conductivity (EC) meltwaters ($<10^{-6}$ S cm^{-1}) that enter these moulins during summer are released at the terminus as sediment-rich, high EC meltwaters ($>0{\cdot}04–0{\cdot}06$ S cm^{-1}) within a period of hours to days (Bingham *et al*., 2003, 2005). It therefore seems likely that these meltwaters flow along the warm basal interface, where they can potentially influence basal motion.

Glacier dynamics at JEG: previous findings

Measurements along JEG’s centreline from July 1998 to early June 2000 revealed that horizontal (*xy*) velocities were, on average, 62 per cent higher in summer (July) than in winter (early August to late May) in the lower ablation zone (hereafter LAZ; the 8 km^2 area between moulins h1–h5 and the terminus, Figure 1), although locally, summer velocities were more than double winter velocities (Copland *et al*., 2003b). Upstream from the bedrock riegel differences between summer and winter velocities were less pronounced. Copland *et al*. (2003b) suggested that drainage into moulins h1–h5 enhanced basal motion (expressed as an increase in surface motion) in the LAZ throughout summer, and that longitudinal stress–gradient coupling (cf. Kamb and Echelmeyer, 1986) increased velocities up to 2 km upglacier. Bingham *et al*. (2003) analysed variations in surface dynamics over the entire glacier over seven- to ten-day periods during the melt seasons of 2000 and 2001. They found that, during summer, weekly surface velocity fluctuations varied considerably across the entire 175 km^2 glacier (excluding the uppermost, cold-based 3 km), with the greatest velocity fluctuations occurring in the LAZ. Concurrent dye-tracing studies revealed that these weekly fluctuations were linked to changes in the structure of the subglacial drainage system which underlies the LAZ (Bingham *et al*., 2005). Thus, the previous

DOI: 10.1002/esp

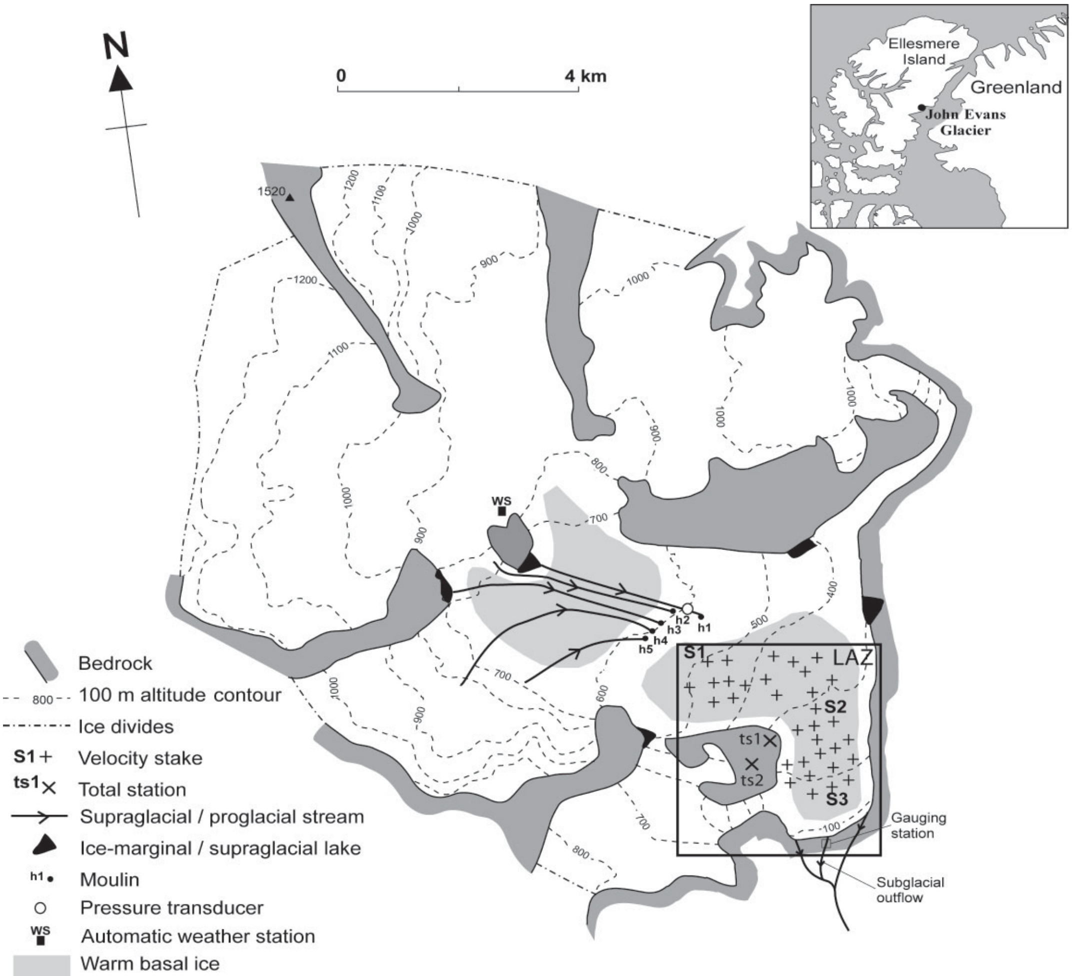

Figure 1. The John Evans Glacier field site. The boxed area labelled LAZ denotes the lower ablation zone study domain analysed in Figures 4–8. Inset: location of John Evans Glacier.

studies of the dynamics of JEG indicate that most of its variability is driven by hydrological processes in the warm-based LAZ. The dynamics and hydrology of this lower 4 km of the glacier thus merit particular focus.

Daily surveys of glacier surface velocities across the LAZ were undertaken throughout the melt seasons of 1998 and 1999 (Copland *et al*., 2003a). Measurement errors precluded analysis of the data at daily resolution, but subsampling at a two-day resolution uncovered significant variations in *xy*-velocities during summer, associated with variations in vertical (*z*) velocities. Transient periods of high (up to 400 per cent overwinter) *xy* and positive *z* (i.e. upward) surface velocities lasting for two to four days were associated with (i) the sudden onset of drainage into moulins h1–h5 in the MAZ and the initial emergence of subglacial outflow in each melt season ('spring events'), and (ii) a period of rapidly increasing supraglacial meltwater input to the MAZ moulins (h1–h5) following a mid-summer spell of cold weather ('mid-summer event'). During the spring events *xy* velocities rose over most of the LAZ, but high positive *z* velocities were limited to a small area near the terminus. During the mid-summer event, *xy*-velocity anomalies were highly localized over a predicted subglacial channel, whilst *z*-velocity anomalies were widely distributed over the lower 2 km of the glacier. Following all of these events, *xy*-velocities dropped (but remained higher than overwinter velocities) and *z*-velocities changed sign.

Thus, Copland *et al*.'s (2003a) findings highlight considerable short-term variability both in *xy*- and *z*-surface motion throughout the warm-based LAZ of JEG, and provide strong evidence that this is linked to subglacial hydrology. However, information on the structural evolution of the subglacial drainage system was not available in 1998 and 1999, hence the relationship of the observed surface dynamics to changes in the subglacial drainage system

DOI: 10.1002/esp

remained unclear. Moreover, it is unknown whether the high-velocity events observed in 1998/99 propagated spatially and/or temporally. Measurements conducted in 2000 and 2001 were designed to investigate these issues.

Measurement and Analysis Methods

Surface velocities

Thirty-three velocity stakes were established over the LAZ (boxed area in Figure 1) at the end of July 1999. The stakes froze naturally into the ice due to low surface ice temperatures. A board with a reflecting prism was attached to each stake and the position of each prism was monitored from survey stations at either ts1 or ts2 (Figure 1). All stake positions were surveyed on 30 July 1999 and then periodically (approximately every two days, weather permitting) between May and August in both 2000 and 2001, using a Geodimeter System 500 total station. Two survey stations were required because not all stakes could be viewed from a single location. To counteract errors associated with wind- or meltout-induced stake motion, stakes were periodically redrilled or replaced. Surveys were not, in general, undertaken daily because previous analyses showed that errors at this resolution were usually greater than calculated velocities (Copland *et al*., 2003a). However, during the 2001 'spring event' (28 June–1 July 2001), when rapid motion was predicted, surveys were undertaken every day. Occasional gaps in the surveys resulted from periods of poor weather.

Directly following procedures outlined in detail in Mair *et al*. (2001, appendix I), repeat surveys to two or three 'reference' targets located on the surrounding bedrock allowed for determination of, and compensation for, errors associated with inherent instrument drift and changes in temperature and pressure during each survey. Each velocity stake was surveyed at least twice during a survey; the x, y and z coordinates were taken as the mean, and the positional error associated with each coordinate value was calculated as the standard deviation of this mean value. Over a typical two-day summer measurement period, this gave a mean xy-velocity error of 6·19 m a^{-1} (1·69 cm day^{-1}) (24 h) and a mean z-velocity error of 4·48 m a^{-1} (1·23 cm day^{-1}), although on occasions individual velocity errors in both dimensions differed significantly from these mean values.

Residual vertical motion

To determine whether vertical motion at the surface over any given period reflected vertical motion at the base of the glacier, a parameter known as the 'residual vertical motion', $\dot{c}$, was derived. We derived $\dot{c}$ using the method described in Copland *et al*. (2003a). Parameter $\dot{c}$ describes the vertical velocity at the surface after correction for the vertical velocity due to vertical strain and vertical motion due to sliding along an inclined surface. This parameter is commonly interpreted as indicative of vertical velocities at the glacier base relative to the bed, although opening and closing of englacial cavities cannot be discounted (Hooke *et al*., 1989). High values of $\dot{c}$ have been assumed to reflect either the opening of basal cavities or an increase in the dilatancy of subglacial till, and commonly coincide with high subglacial water pressures (Hooke *et al*., 1989; Mair *et al*., 2002; Copland *et al*., 2003a). Hence determination of $\dot{c}$ can help to identify vertical motion at the surface resulting from basal hydrological forcing. We identify basal uplift in response to basal hydrological forcing wherever the vertical strain rate averaged over the ice thickness ($\dot{\varepsilon}_{zz}$) increases with depth relative to the vertical strain rate at the surface, and $\dot{c} > 1$ cm day^{-1}. Full details of this method with specific reference to John Evans Glacier can be found in Copland *et al*. (2003a).

Spatial interpolation of velocities

To analyse the spatial distribution and propagation of surface velocity anomalies, spatial plots of xy- and z-velocities and residual vertical motion over the LAZ were produced by interpolating over a 100 m^2 grid from the point measurements at the stakes. All interpolations were undertaken using the kriging interpolation routine in the Golden Software graphics program 'Surfer'. Interpretations of velocity patterns from the resultant spatial plots are made only for areas where stakes are present.

Supraglacial–subglacial hydrology

Investigations of subglacial drainage system structure beneath the LAZ during each melt season were undertaken using dye-tracer tests in conjunction with measurements of discharge into and out of the system. Full details of these experiments are presented in Bingham *et al*. (2005); here, only a brief summary is provided for comparison with the glacier velocity analyses.

DOI: 10.1002/esp

During each melt season, following the initiation of drainage into h1–h5 (19–21 June 2000; 28 June 2001) and the onset of subglacial outflow (22 June 2000; 29 June 2001), known quantities of the fluorescent dye Rhodamine-WT were injected periodically into moulin h1 over the bedrock riegel (Figure 1). Immediately prior to each dye injection, the discharge, Q_S, of the supraglacial stream flowing into the moulin was measured using the velocity–area method. Dye emergence at the glacier snout occurred via a single subglacial outflow and was determined by fluorometric analysis of water samples taken from the proglacial stream near the subglacial outlet portal (Figure 1). Two parameters were determined from the resultant breakthrough curves:

(i) a minimum estimate of the mean water flow velocity, u (m s^{-1}), during each experiment, given by

$$u = \frac{x}{t}$$

where x (m) is the distance between the injection and detection sites, assuming flow along a flow-line that parallels the glacier margins, and t is the time between dye injection and peak dye concentration at the snout; and

(ii) an estimate of the mean cross-sectional area, A_M (m^2), of the subglacial drainage system between h1 and the subglacial outflow, determined by

$$A_M = \frac{Q_S}{u}$$

where Q_S is the discharge entering moulin h1 (after Willis *et al.*, 1990; Hock and Hooke, 1993). It should be noted that if discharge increases between h1 and the subglacial outflow, this method will underestimate the mean cross-sectional area of the subglacial drainage system. Values of A_M therefore provide an index, rather than a true measure, of subglacial drainage system cross-sectional area. Large A_M is indicative of distributed drainage; low A_M implies channelized drainage.

Discharge into and out of the subglacial drainage system (Q_S and Q_B respectively) was monitored continuously throughout each melt season using pressure transducers placed in supraglacial stream SS1 above moulin h1, and in the subglacial outflow, each calibrated with periodic on-site measurements of discharge. However, frequent channel aggradation and migration in the subglacial outflow meant that a reliable record of Q_B could not be attained in either summer, hence only the records of Q_S are used in the analysis presented here. However, qualitative observations indicate that Q_B was typically of the order of five times greater than Q_S in SS1, with additional supraglacial inputs over the MAZ riegel accounting for the difference.

To gauge the influence of weather conditions on possible hydrologically driven velocity variations at JEG, meteorological variables were measured throughout both melt seasons at a weather station located at 824 m a.s.l. (Figure 1).

Results

Temporal variations in glacier dynamics

Seasonal variations in *xy*-velocities and *z*-displacements for three selected stakes across the LAZ (stakes S1, S2 and S3 on Figure 1) are shown for summer 2000 (Figure 2) and summer 2001 (Figure 3). Motion at these three stakes broadly reflects glacier motion in the upper, central and lower sectors of the LAZ. The plots are annotated with information about significant hydrological events that occurred during each melt season. Plots of minimum dye flow velocities (u), apparent drainage system cross-sectional area (A_M), mean daily air temperature (T), and supraglacial discharge (Q_S), are also presented in parts g and h of Figures 2 and 3.

During both melt seasons, two-day variations in horizontal and vertical motion were broadly, although not exactly, coincidental across the LAZ. Periods of high *xy*-velocities and positive (upward) *z*-displacements often occurred one or two days earlier upglacier (represented by stake S1) than centrally (stake S2) or further downglacier (stake S3) (Figures 2 and 3). Prior to 18 June in 2000, and 26 June in 2001, all *xy*-velocities closely matched mean annual (and overwinter) velocities, and *z*-displacements were negative or only slightly positive (Figures 2 and 3). However, two short-term 'events' of anomalously high *xy*-velocities and positive *z*-displacements occurred subsequently in both melt seasons. In 2000, *xy*-velocities increased by *c*. 2·5 times at stake S1 from 18 to 20 June, and were associated with a

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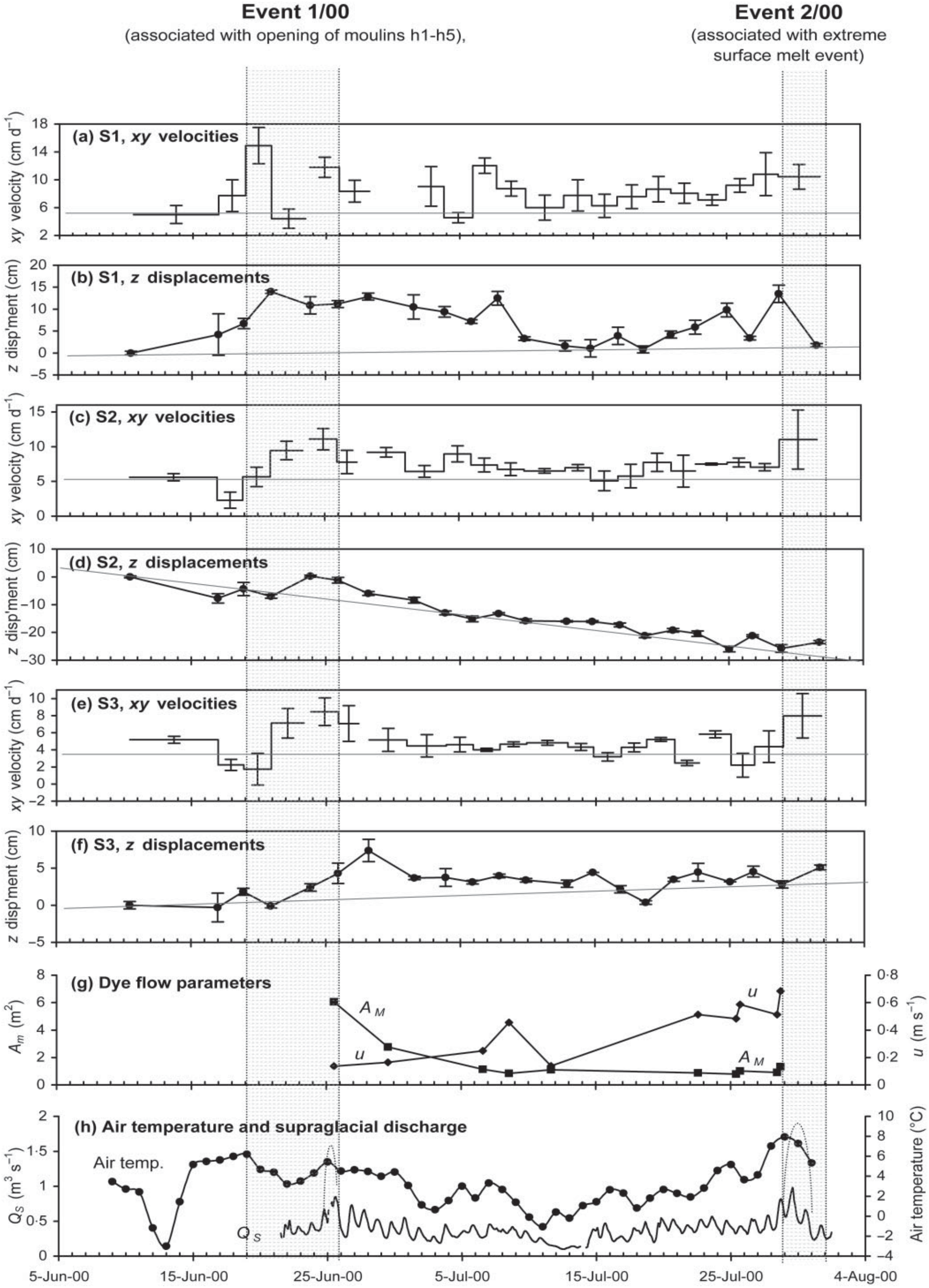

Figure 2. Time series of glacier surface dynamics during summer 2000: (a, c, f) horizontal surface velocities, and (b, d, f) vertical displacements (faint lines represent mean summer trends) for stakes S1 (relatively upglacier), S2 (central) and S3 (relatively downglacier) respectively (locations on Figure 1). The main high-velocity events are highlighted. (g) Concurrent variations in mean subglacial drainage system throughflow velocities u, and apparent drainage system cross-sectional area, A_M, in 2000. (h) Concurrent variations in mean daily air temperature, T, at 824 m a.s.l. and supraglacial discharge, Q_S, entering moulin h1 (see Figure 1 for position of transducer) in 2000.

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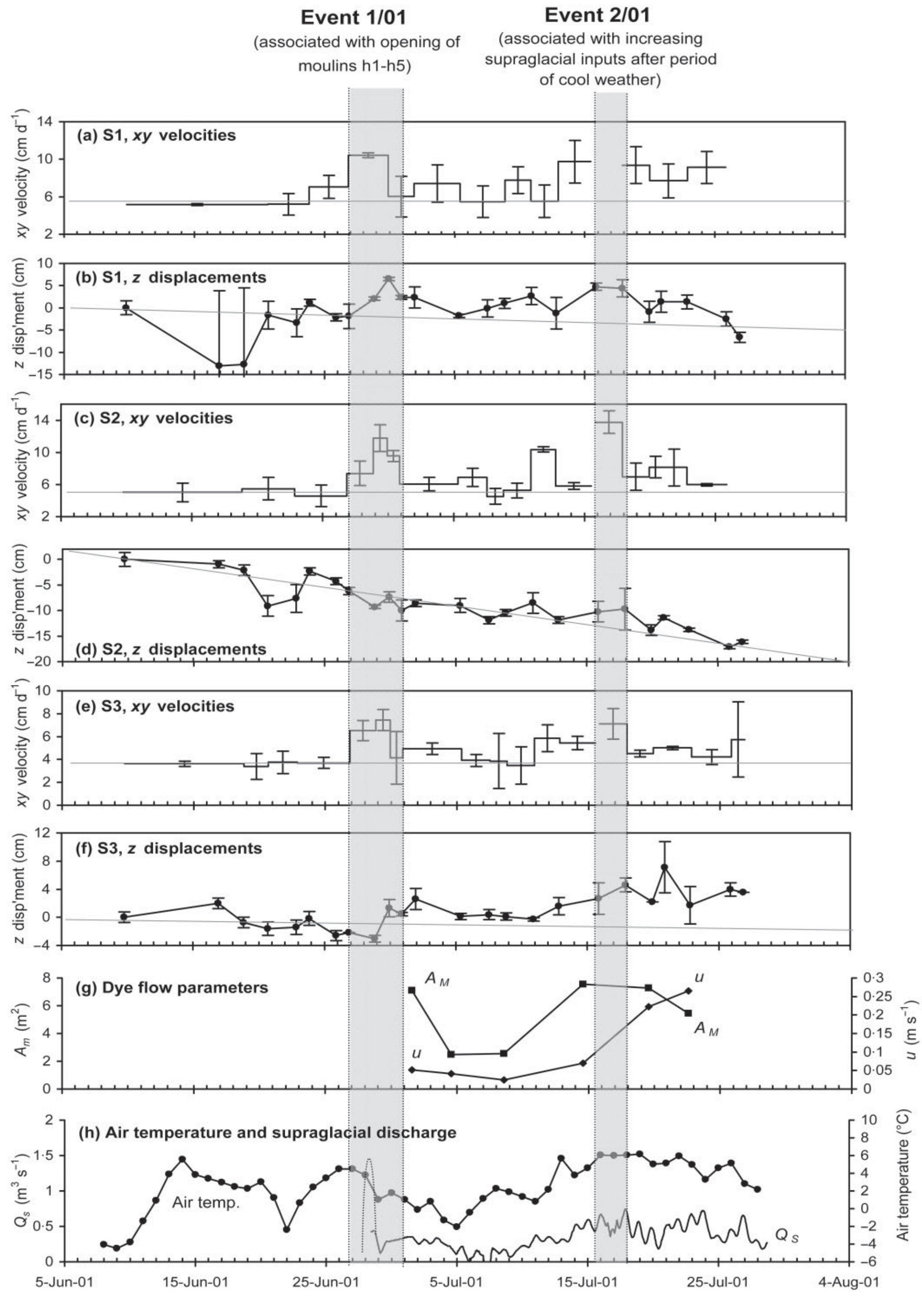

Figure 3. As for Figure 2, but relating to summer 2001.

positive z-displacement of the glacier surface by c. 7 cm (Figure 2a, b). This was followed on 20–25 June by an increase in xy-velocities and a positive z-displacement at stakes S2 and S3 (Figure 2c–f). Relatively high xy-velocities were also recorded at stake S1 from 23 to 25 June, although these were not associated with a significant change in z-displacement (Figure 2a, b). For the week after this 'spring event' (hereafter 'Event 1/00'), xy-velocities decreased

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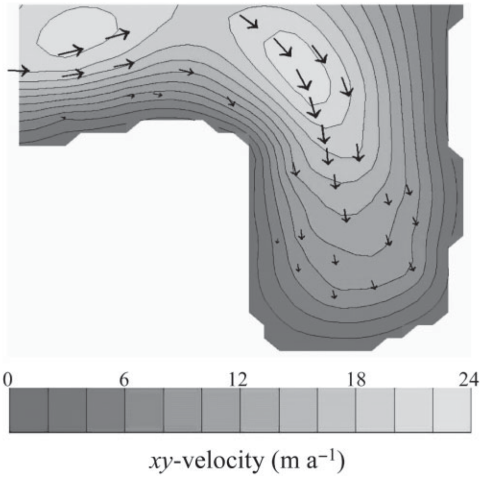

Figure 4. The distribution of mean annual horizontal surface velocities over JEG as measured from 30 July 1999 to 26 July 2001. Vectors are scaled to glacier speeds and show the direction of flow at each marker stake. The location of the study domain within JEG is shown in Figure 1.

(although they stayed higher than winter levels until the end of the melt season), and stakes dropped vertically (Figure 2). Later in the summer, anomalously high *xy*-surface velocities and positive *z*-displacements were observed on 28–31 July at stakes S2 and S3 (Figure 2). This 'mid-summer event' (hereafter 'Event 2/00') again initiated earlier (24 July) at S1 (Figure 2).

In 2001, a 'spring event,' which resulted in a >100 per cent increase in *xy*-velocities at each of the three stakes and vertical uplift at S1 and S3, occurred between 26 and 30 June ('Event 1/01') (Figure 3). From 30 June until 10 July *xy*-velocities dropped to a mean summer level, while *z*-displacements were negative or slightly positive (Figure 3). A further 'mid-summer event' occurred between 15 and 17 July ('Event 2/01') (Figure 3). This peak was not recorded at S1 because data are unavailable for 15–17 July, but it was recorded at neighbouring stakes (not displayed in Figure 3). From 17 to 20 July, *xy*-velocities decreased and *z*-displacements were negative; after 20 July *xy*-velocities remained steady and *z*-displacements were minimal.

Spatial variations in glacier dynamics

We now examine the spatial pattern of the glacier's dynamic response to hydrological forcing and the propagation of high-velocity events in 2000 and 2001. Figure 4 shows the distribution of mean annual surface velocities across the LAZ measured between 30 July 1999 and 26 July 2001, and acts as the benchmark velocity distribution against which subsequent velocity fields are compared. Because each melt season at JEG is so short (*c*. 60 days), and the period during which supraglacial meltwaters access and perturb the subglacial drainage system is even shorter (*c*. 40 days), mean annual horizontal velocities approximate to overwinter velocities, and reflect the 'background' surface motion due to internal deformation and long-term, average basal motion (Figure 4). However, over short periods during summer, the surface velocity distribution deviates significantly from this mean annual distribution, and these differences are best highlighted by mapping 'horizontal surface velocity anomalies', defined as percentages of mean annual horizontal surface velocities.

Horizontal surface velocity anomalies are mapped across the LAZ for each *c*. two-day period in 2000 (Figure 5) and 2001 (Figure 6). Anomalies <100 per cent and >100 per cent indicate surface velocities lower and higher than background rates, respectively, but to discount effectively any anomalies that might be derived from positional errors during the measurement of velocity stakes we only interpret anomalies >130 per cent as clear indicators of hydrologically forced basal motion.

Figures 5 and 6 expose considerable short-term variability in surface dynamics during each melt season. The evolution of surface velocity fields before and during each spring event (Events 1/00 and 1/01) was similar in both 2000 and 2001, but dynamic behaviour after the spring event in 2001 contrasted significantly with that observed in 2000. Thus, prior to each spring event (10–18 June 2000, Figure 5; 9–26 June 2001, Figure 6) horizontal surface velocities closely resembled mean annual velocities in overall magnitude, direction and spatial distribution. Each spring event was signalled by increasing velocities across parts of the LAZ (18–20 June 2000, Figure 5; 26–28 June

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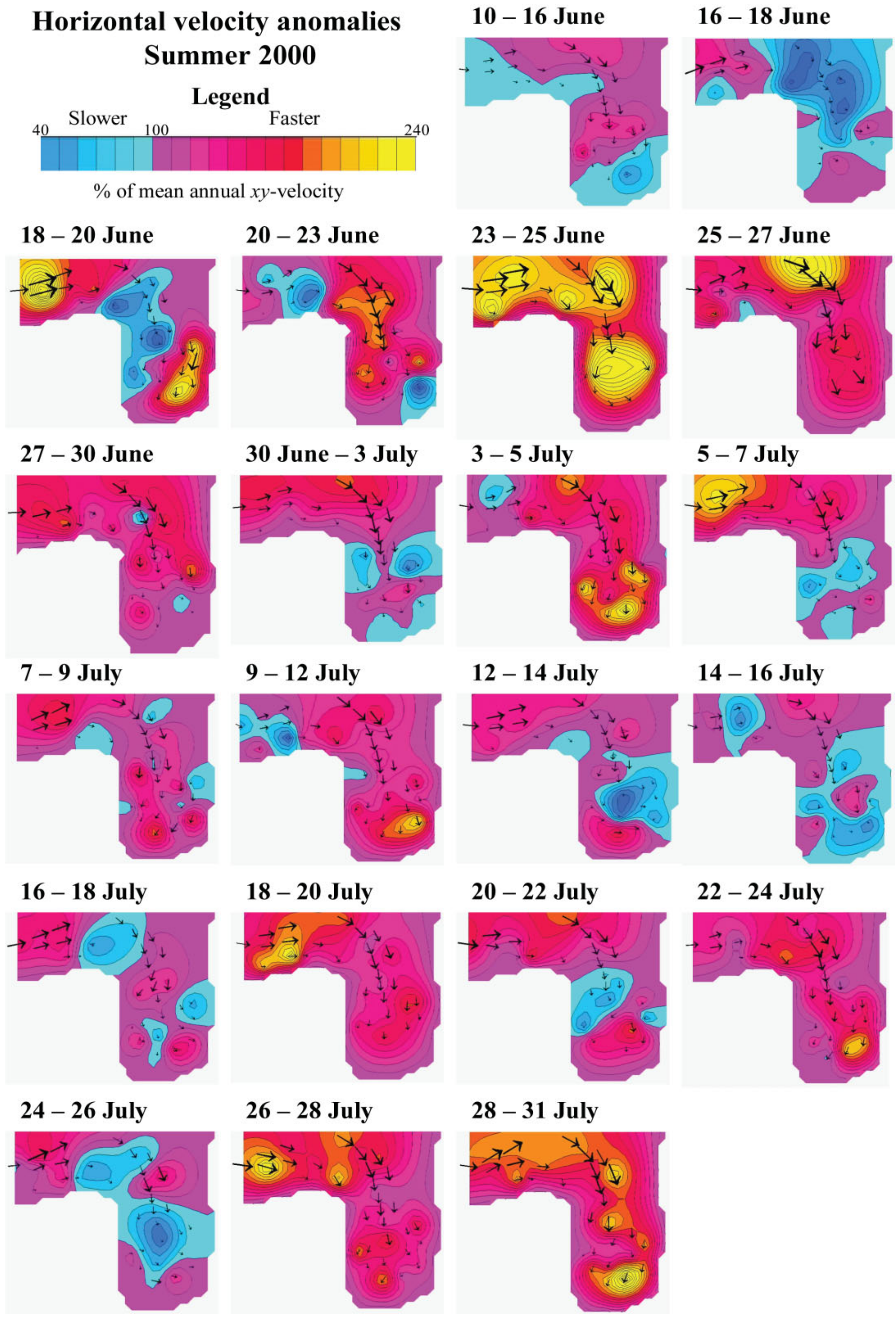

Figure 5. Spatial distributions of horizontal surface velocity anomalies (percentages of mean annual velocities as shown in Figure 4) between each survey in 2000. Flow vectors are scaled and oriented to recorded velocities at each stake. This figure is available in colour online at www.interscience.wiley.com/journal/espl

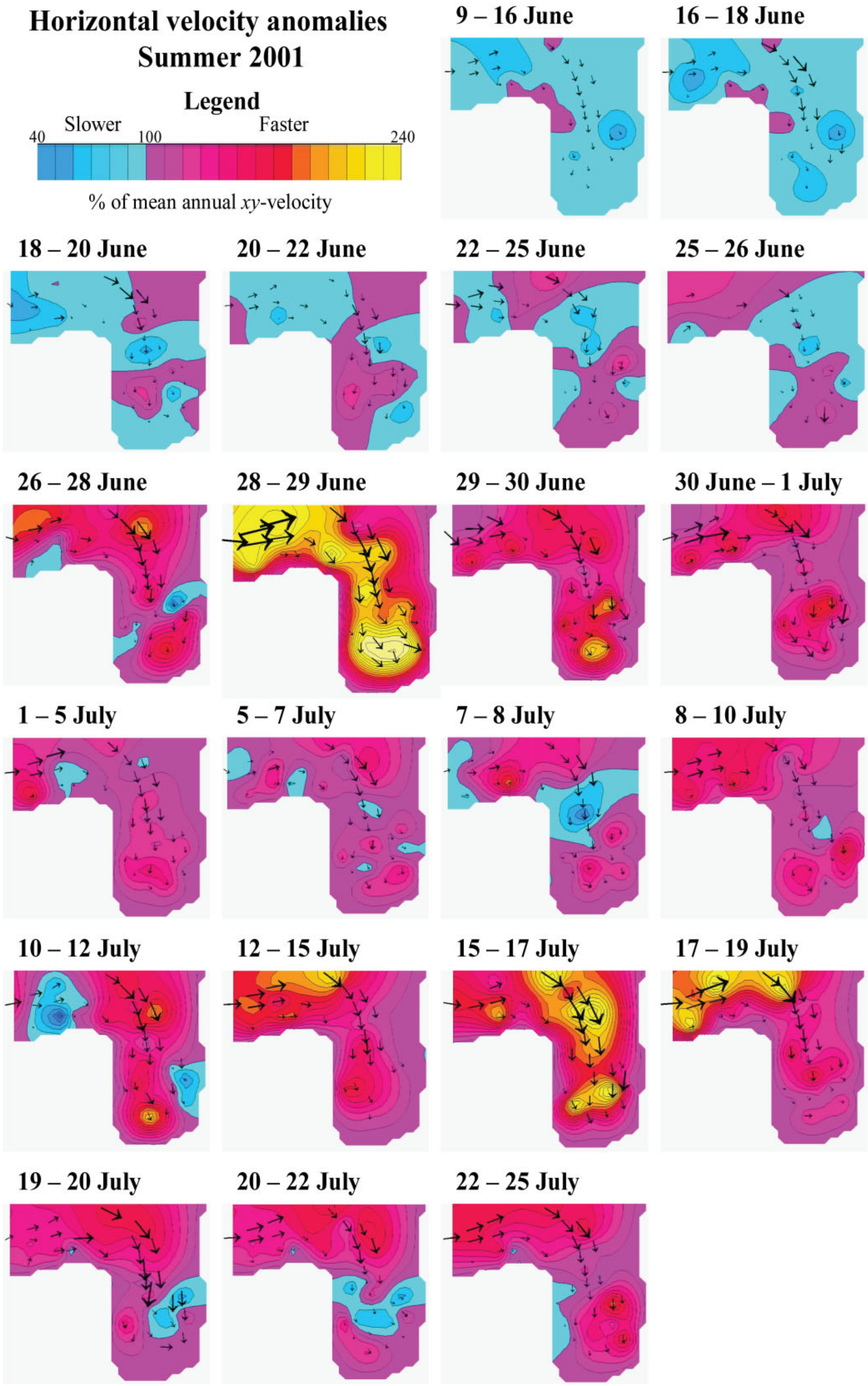

Figure 6. As for Figure 5, but relating to summer 2001. This figure is available in colour online at www.interscience.wiley.com/journal/espl

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2001, Figure 6), followed shortly afterwards by a peak in velocities spread across the entire LAZ (23–25 June 2000, Figure 5; 28–29 June 2001). In 2000, the build-up to peak velocities took six days longer than in 2001 (cf. 18–27 June 2000 (Figure 5) with 26–29 June 2001 (Figure 6)). In both years, the velocity anomaly persisted upglacier after it had disappeared downglacier (25–27 June 2000, Figure 5; 30 June–1 July 2001, Figure 6).

In the days after each spring event (27 June–26 July 2000, Figure 5; 1–10 July 2001, Figure 6), horizontal velocities remained lower than during Events 1/00 and 1/01 but higher than mean annual levels. In 2000, occasional periods of higher surface velocities occurred locally during the periods 3–7 July, 7–12 July, 18–20 July and 22–24 July, and over the course of the melt season the velocity anomalies became increasingly concentrated along the glacier centreline. A particularly large velocity event was then observed towards the end of the melt season, on 26–31 July (Event 2/00), during which surface velocities across the entire LAZ, but especially along the glacier centreline, were characterized by considerably higher than mean summer velocities and, as during Event 1/00, the zone of high velocities propagated downglacier. After the spring event in 2001, only one notable further period of high horizontal surface velocities occurred in that melt season, between 12 and 19 July. This event was signalled by a high surface velocity anomaly upglacier (12–15 July 2001, Figure 6), and was followed by a widespread surface velocity anomaly across much of the LAZ (15–17 July 2001, Event 2/01). From 17 to 19 July 2001, as during the spring events, the velocity anomaly persisted later in the upglacier reaches of the study area than downglacier (Figure 6). At no time during this event was the velocity anomaly concentrated along the glacier centreline, as observed during Event 2/00. From 19 to 25 July 2001 (Figure 6), after Event 2/01, horizontal surface velocities stayed lower than during Event 2/01 but remained higher than mean annual values.

Vertical motion during high-velocity events

Analysis of vertical motion variations was limited to the three- to five-day periods before, during and after the spring and mid-summer high-velocity events in each melt season, because outside these periods and with a smaller temporal resolution, vertical velocity errors precluded the identification of clear vertical velocity trends.

Plots of vertical uplift and residual vertical motion before, during and after each of the two high-velocity events are presented for 2000 (Figure 7) and 2001 (Figure 8), although the 'after' segment is missing from Event 2/00 (Figure 7) because field surveys ceased in early August 2000. Vertical uplift and residual vertical motion phenomena were similar during Events 1/00, 1/01 and 2/01, but different during Event 2/00. Thus, before Events 1/00, 1101 and 2/01, residual vertical motion was generally low; during these three events residual vertical motion was high and widespread across the LAZ; and in the three to five days after these events residual vertical motion again became negligible (Figures 7 and 8). However, in the four days before the peak of Event 2/00, residual vertical motion was moderately high in isolated areas but largely negligible during the event itself (28–31 July 2000, Figure 7).

Discussion

Our data demonstrate that summer ice motion across the entire warm-based LAZ of JEG is highly variable over at least a two-day timescale. Horizontal surface velocities increased during each melt season relative to mean annual (*c*. overwinter) velocities, and the overall summer increase was significantly enhanced by the occurrence of spring and mid-summer velocity events which persisted for two to four days. These high-velocity events were often accompanied by vertical uplift of the glacier surface and a widespread increase in residual vertical velocities, implying that the predominant cause of the surface uplift was enhanced vertical motion at the base relative to the bed. However, the spatial and temporal distribution of horizontal and vertical velocities differed between and during events, and there was evidence that zones of high velocity propagated downglacier and persisted upglacier during some, if not all, of such events. In this discussion we interpret the observed short-term variations in glacier dynamics with reference to concurrent variations in supraglacially driven subglacial hydraulics.

Spring events

In both 2000 and 2001, short-term horizontal velocity patterns and vertical displacements at the glacier surface deviated little from their mean annual equivalents until the onset, in late June, of supraglacial drainage into moulins h1–h5, 4 km upglacier from the terminus. This drainage was initiated in response to the rapid propagation of crevasses to the base of the glacier by hydrofracture after supraglacial water ponding over the riegel during late May and early June overcame a pressure threshold (Boon and Sharp, 2003). Dye-tracer experiments conducted in the days shortly after this supraglacial–subglacial connection was made each year yielded subglacial water flow velocities

DOI: 10.1002/esp

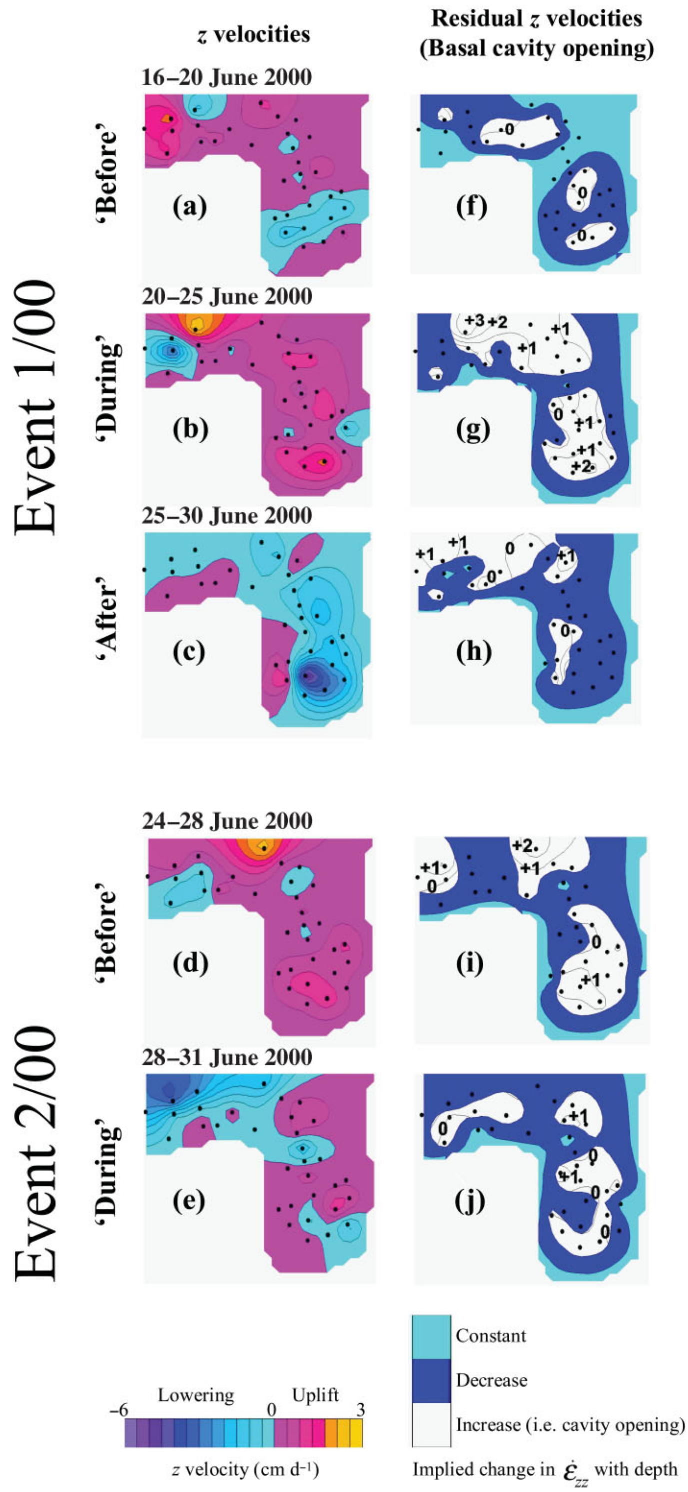

Figure 7. (a–e) Measured vertical uplift at the surface, and (f–j) residual vertical velocities in the three to five day periods before, during and after the spring and mid-season high-velocity events in 2000. Motion at the glacier bed relative to the base is identified where vertical strain rate averaged over the ice thickness ($\dot{\varepsilon}_{zz}$) increases with depth relative to the vertical strain rate at the surface, and $\dot{c}$ (contoured) > 1 cm day^{-1}. Black dots indicate marker stake locations. This figure is available in colour online at www.interscience.wiley.com/journal/espl

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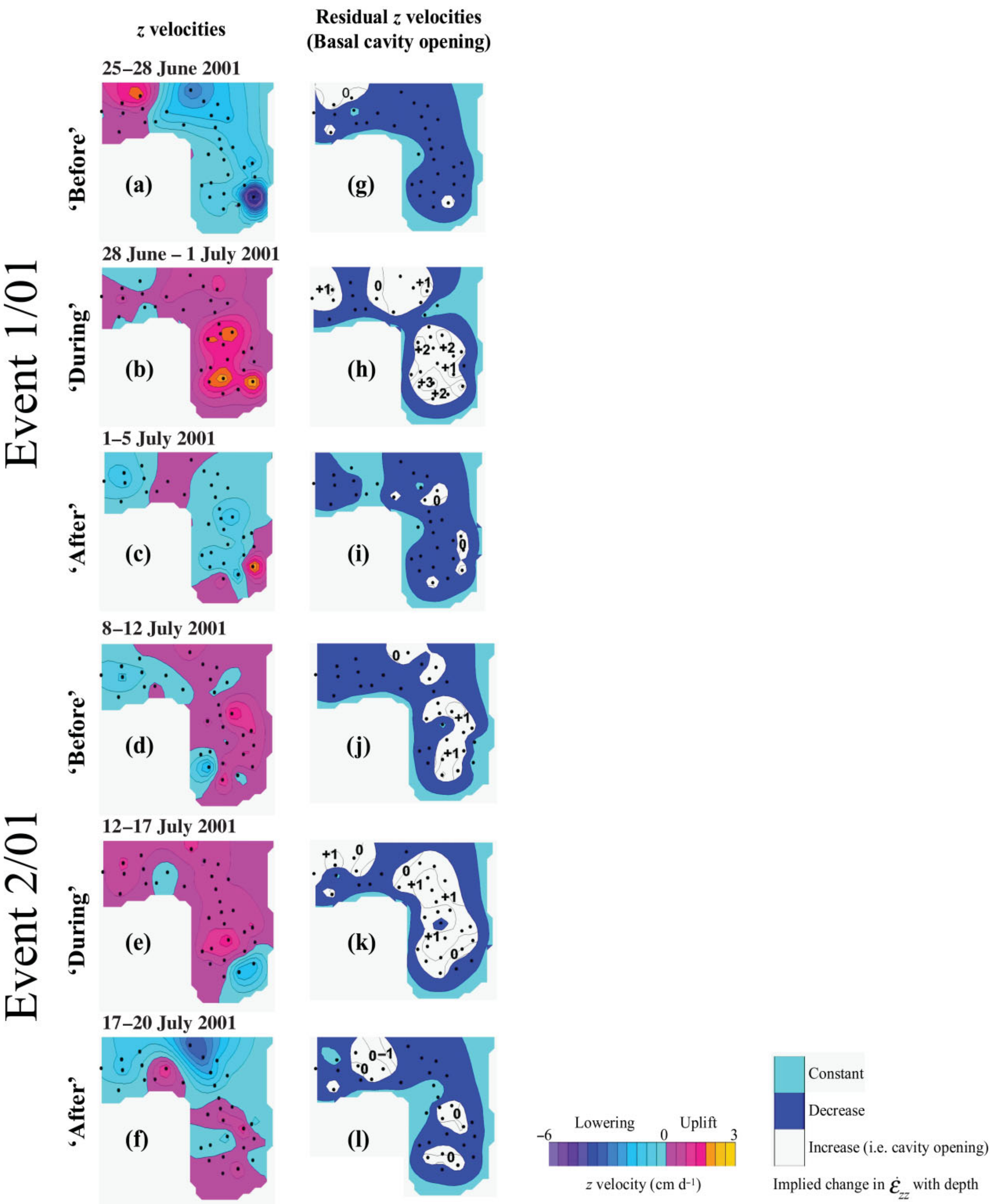

Figure 8. As for Figure 7, but relating to the spring and mid-season high-velocity events in 2001. This figure is available in colour online at www.interscience.wiley.com/journal/espl

DOI: 10.1002/esp

$u < 0{\cdot}15$ m s^{-1} and subglacial drainage system cross-sectional areas $A_M > 6$ m^2 (Figures 2g and 3g), suggesting a distributed subglacial drainage system configuration in which basal water pressures are high (Bingham *et al.*, 2005). Hence, the large volumes of supraglacial meltwaters penetrating rapidly to the glacier bed (see supraglacial stream discharge estimates, Figures 2h and 3h) would have quickly generated high subglacial water pressures, initially in the vicinity of the bedrock riegel. Over the next two to four days, the zone of high subglacial water pressures propagated downglacier, inducing the observed downglacier propagations of high horizontal surface velocities during the spring events each year. High residual vertical velocities during both Events 1/00 and 1/01 (Figures 7 and 8) suggest that high subglacial water pressures were inducing widespread basal cavity opening (or subglacial till dilatancy) as the supraglacially derived meltwaters were routed downglacier along the basal interface.

It is notable that in both melt seasons, surface velocities over the lower ablation zone increased significantly one or two days prior to the first observations of surface-melt drainage into the MAZ moulins (h1–h5). Thus, in 2000, although drainage into h1–h5 was first observed between 19 and 21 June, and the subglacial outburst occurred on 22 June, anomalously high surface velocities and vertical uplift were already observed across the upper-western and lower-eastern sectors of the lower ablation zone between 18 and 20 June (Figure 5). Similarly, in 2001, high surface velocities were observed on 26–28 June (Figure 6), although drainage into h1–h5 only began on 28–29 June and the subglacial outburst occurred on 29 June. One explanation for this perhaps surprising finding might be that supraglacial meltwater initially accessed the glacier base on 19–20 June 2000 and 28 June respectively, thus the measurement periods 18–20 June 2000 and 26–28 June 2001 partly captured their resultant impacts on glacier dynamics. Alternatively, or perhaps in addition, small volumes of supraglacial meltwater may have leaked into the subglacial drainage system in the one or two days prior to the initiation of major drainage into h1–h5, perturbing the basal drainage system and raising subglacial water pressures sufficiently to induce basal motion. Although these initial water inputs would not have been sufficient to cause breaching of the thermal dam at the terminus, it is possible that their resultant effects on basal dynamics may have fed back into and enhanced crevasse opening over the riegel, leading up to a threshold at which a significant amount of the remaining ponded supraglacial meltwaters rapidly accessed the subglacial system. Subsequent observations from 2002 relating to the opening of the moulins by hydrofracture support the latter hypothesis, showing that the connection to the base is established by small multiple fracture events prior to the larger principal fracture event (Boon and Sharp, 2003).

The subsequent escape of the high-pressure meltwaters through the thermal dam of cold ice and frozen sediments to the proglacial zone, one or two days after the establishment of the supraglacial–subglacial connections, apparently enabled subglacial water pressures (hence horizontal surface velocities) to fall gradually across most of the LAZ over the weeks following both Events 1/00 and 1/01 (27 June–1 July, Figure 5; 1–8 July 2001, Figure 6). The virtual elimination of residual vertical velocities after each spring event (Figures 7 and 8) also reflects decreasing meltwater inputs to the subglacial system as the surface stores drained, allowing basal cavities to reduce in size (or basal till to compress) due to lower subglacial water pressures. The persistence of high surface velocities in the upper LAZ in the days following the main event (25–27 June 2000, Figure 5; 30 June–1 July 2001, Figure 6) probably results from the presence of thicker ice, which would have counteracted subglacial channel opening more effectively than in other locations where ice is thinner.

In terms of the behaviour of horizontal and vertical velocities, Events 1/00 and 1/01 were similar to Events 1/98 and 1/99 reported in Copland *et al.* (2003a). It seems probable, therefore, that such spring events are annual phenomena at JEG, and are driven primarily by the annual build-up of supraglacial meltwater stores over the upper ablation zone, and its subsequent drainage into a hydraulically inefficient distributed subglacial system two to three weeks into each melt season.

Post-spring-event behaviour and mid-summer events

Following the 2000 and 2001 spring events, flow fields generally differed little from the mean summer pattern; horizontal surface velocities slightly exceeded mean annual velocities, but otherwise experienced relatively little variability. This implies that after the spring events, supraglacial meltwater continued to access the bed and influence basal motion, but subglacial water pressures had reduced. This was almost certainly due to a combination of reduced surface melt inputs after the drainage of the surface stores (see Q_S, Figures 2h and 3h) and increasing channelization of subglacial drainage with time through each melt season (see u and A_M, Figures 2g and 3g). These horizontal velocity patterns over mid- to late summer resemble summer velocity responses to hydrological forcing at many temperate glaciers, where fluctuations in weather conditions and subglacial drainage system evolution influence summer flow variability (Fountain and Walder, 1998; Nienow *et al.*, 2003).

Remarkable responses were, however, observed during the 'mid-summer' high-velocity events (Events 2/00 and 2/01) which differed in nature and timing between the two melt seasons. Neither Event 2/00 nor Event 2/01 was

DOI: 10.1002/esp

associated with further drainage of ponded supraglacial waters; instead both events were associated with rapidly rising meltwater inputs to all moulins (see, for example, Q_S in Figures 2 and 3) in response to a significant increase in air temperatures (Figures 2h and 3h). Event 2/00 followed a month-long period during which meltwater inputs had been consistently high but were supplemented considerably during the event itself by extra surface meltwaters generated by extreme rates of surface melting brought about by exceptionally strong, warm winds (Boon *et al.*, 2003). Event 2/01 occurred when supraglacial meltwater inputs rose dramatically after a two-week period of cool weather and low (at times zero) surface melting (Figure 3h) following the spring event.

Event 2/00 probably represents a relatively rare event. At that time, the subglacial drainage system had rationalized into an efficient, channelized configuration (evidenced by high u and low A_M, Figure 3g), which would typically be less sensitive to supraglacial hydrological forcing than the distributed configuration earlier in the melt season. However, from 28 to 30 July 2000 exceptionally warm, strong winds more than doubled rates of surface melting over background summer rates (Boon *et al.*, 2003). This unusually large increase in melt inputs in late July (see Q_S, Figure 2h) probably exceeded that which could be accommodated solely by channel widening over such a short period. Hence some meltwater must have been forced into the distributed system adjacent to the channel, initially in the vicinity of the riegel, then propagating downglacier as the increased melt volumes flowed downglacier along the subglacial channel (26–31 July 2000, Figure 5). The existence of a large subglacial channel during this event also explains why, in contrast to the other three high-velocity events identified in 2000/01, the horizontal surface velocity was localized along the glacier centreline (28–31 July 2000, Figure 5) and little concurrent residual vertical motion was observed (Figure 7). In conclusion, Event 2/00 occurred because exceptionally high supraglacial meltwater inputs exceeded the threshold at which the hydraulically efficient subglacial system was able to cope with rising input volumes. Such a sudden rise in inputs is probably uncommon, because the synoptic conditions which brought about the late-July extreme melt event are rare, occurring on only 0·1 per cent of days between 1948 and 2000 (Boon *et al.*, 2003).

By contrast, Event 2/01 may represent a more common mid-summer phenomenon at JEG. In 2001, cool weather conditions (see T, Figure 3h) and patchy snowfall in the two weeks following the formation of the supraglacial–subglacial links in the MAZ precluded channelization of subglacial drainage (as reflected by variations in u and A_M in Figure 3g). Consequently, when air temperatures and surface melting increased on 10 July (about two weeks after Event 1/01), and supraglacial meltwater inputs into h1-h5 rose correspondingly (Figure 3h), the increased volumes of meltwater accessing the bed probably encountered a distributed subglacial drainage system similar in form to that encountered by supraglacially derived meltwaters two weeks earlier. As a result, the occurrence and subsequent downglacier propagation of high glacier surface velocities (12–17 July 2001, Figure 6), coincident with vertical uplift and high residual vertical velocities across a wide area of the glacier tongue (Figure 8), probably occurred in response to a similar downglacier propagation of high subglacial water pressures. The processes operating during Event 2/01 were therefore equivalent to those reported for the spring events, except that they were induced by a sharp rise in surface melting rather than a sudden release of stored supraglacial meltwater. As during the spring events, the velocity anomaly persisted upglacier after its propagation downglacier (17–19 July 2001, Figure 6), reflecting greater resistance to conduit opening and greater conduit closure beneath thicker ice. Mid-summer high-velocity events, such as Event 2/01, may be relatively common at JEG, because previous observations in the summer of 1994, 1996 (Skidmore and Sharp, 1999) and 1999 (Copland *et al.*, 2003a) have all shown that supraglacial–subglacial connections can close down during brief (about one week) spells of cool weather.

Implications for mass balance response to climate change

The coupling between supraglacial–subglacial hydrology and glacier flow as observed at JEG has profound implications for the mass balance response of High Arctic glaciers to climate change. Surface ice in the LAZ flowed, on average, 5·3 per cent further downglacier in 2000, and 2·7 per cent further downglacier in 2001, than it would have done had it flowed continually at overwinter (August–May) velocities. (The greater transfer of ice in summer 2000 reflects a longer melt season and higher overall melt inputs to the base throughout summer 2000 relative to summer 2001.) These figures indicate that greater volumes of ice are being transferred annually to lower elevations – where surface melting and runoff are enhanced – than would be the case assuming flow due to internal deformation and long-term, average basal motion alone. Given that many mostly cold glaciers and ice sheets flow faster in summer than over winter (e.g. Müller and Iken, 1973; Andreasen, 1985; Blatter and Kappenberger, 1988; Zwally *et al.*, 2002), such summer increases in motion should be incorporated in models of High Arctic glacier mass balance response to climatic warming. Thus, where glacier velocities for use in such models are derived from remote measurements, for example, care must be taken to ensure that both ‘winter’ and ‘summer’ modes of motion are measured and incorporated into the dynamics’ parameterizations.

DOI: 10.1002/esp

Conclusions

Annual-scale dynamics at polythermal John Evans Glacier are dominated by internal deformation and long-term, average basal motion, but superimposed onto this background behaviour during summer are significant short-term variations in glacier dynamics induced by subglacial hydrological forcing. The principal driver of these short-term variations is supraglacial hydrological forcing, but the spatial and temporal propagation of each high-velocity event across the lower ablation zone is mediated by the efficiency of the subglacial drainage system at the time of supraglacial melt inputs.

The initial supraglacial forcing in each melt season involves the drainage of large volumes of supraglacially stored meltwaters to the glacier bed in the mid-ablation zone, 4 km above the terminus, after a threshold is overcome to permit englacial drainage through cold ice. The glacier's flow response to this event is broadly similar each year, and consists of a downglacier propagation of high horizontal surface velocities and vertical uplift over the full width of the lower ablation zone. Such spring events typically last for two to four days, and reflect the laterally widespread downglacier propagation of a zone of high subglacial water pressures through a spatially extensive distributed subglacial drainage system, inducing enhanced basal motion and vertical uplift at the ice–bed interface. Subglacial water pressures subsequently fall with subglacial drainage system rationalization once the thermal dam at the terminus is breached and subglacial meltwaters escape from the previously confined subglacial reservoir.

After each year's spring event, the nature of supraglacial hydrological forcing, and the degree to which the subglacial drainage system becomes channelized each melt season, is determined by the summer weather conditions, in a similar manner to coupling between hydrology and glacier flow at temperate glaciers. In melt seasons during which warm weather and high surface melting follow the spring event and are sustained throughout summer (e.g. 2000; 1998, Copland *et al*., 2003a), the creation and persistence of efficient subglacial channels may suppress the widespread generation of high subglacial water pressures. On rare occasions exceptionally high meltwater inputs may induce high subglacial water pressures along, and in the regions immediately neighbouring, subglacial channels, thereby inducing high surface velocities along the glacier centreline (e.g. Event 2/00); but mostly in such summers surface velocities will diverge little from mean summer velocities after the spring event as subglacial channels adjust to varying melt inputs. In contrast, where melt seasons are punctuated by at least week-long periods of cool air temperatures and low surface melting (e.g. 2001; 1999, Copland *et al*., 2003a; 1996 and 1994, Skidmore and Sharp, 1999), the subglacial drainage system may remain in, or revert to, largely distributed form, leaving it vulnerable to the further widespread generation of high subglacial water pressures after the spring event. Such summers are much more likely to witness secondary 'mid-summer' high-velocity events, as in 2001 and 1999 (Copland *et al*., 2003a), which are similar in form to the earlier spring events.

Acknowledgements

This research was supported by NERC ARCICE grant GST/02/2202 (P.W.N.) with tied studentship 24/99/ARCI/16 (R.G.B.), NSERC Discovery Grant 155194-99 (M.J.S.), and research grants from the University of Alberta Circumpolar Institute and Northern Science Training Programme, and the Geological Society of America. Field support and logistics were provided by the Polar Continental Shelf Project, Natural Resources Canada (PCSP/EPCP contribution number 001-06). We would like to express our thanks to the Nunavut Research Institute and the communities of Grise Fjord and Resolute Bay for permission to work at John Evans Glacier; J. D. Barker, S. Boon, K. E. Heppenstall, D. H. Lewis and T. M. H. Wohlleben for field assistance; and A. J. Hodson and I. C. Willis for insightful reviews.

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