Response of glacier basal motion to transient water storage

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Basal motion of glaciers is responsible for short-term variations in glacier velocity[1–6]. At the calving fronts of marine-terminating outlet glaciers, accelerated basal motion has led to increased ice discharge and thus is tightly connected to sea level rise[1,7]. Subglacial water passes through dynamic conduits that are fed by distributed linked cavities at the bed, and plays a critical role in setting basal motion[8]. However, neither measured subglacial water pressure nor the volume of water in storage can fully explain basal motion[2–6,8,9]. Here, we use global positioning system observations to document basal motion during highly variable inputs of water from diurnal and seasonal melt, and from an outburst flood at Kennicott Glacier, Alaska. We find that glacier velocity increases when englacial and subglacial water storage is increasing. We suggest that whenever water inputs exceed the ability of the existing conduits to transmit water, the conduits pressurize and drive water back into the areally extensive linked cavity system. This in turn promotes basal motion. Sustained high melt rates do not imply continued rapid basal motion, however, because the subglacial conduit system evolves to greater efficiency. Large pulses of water to the bed can overwhelm the subglacial hydrologic network and incite basal motion, potentially explaining recent accelerations of the Greenland Ice Sheet[3], where rapid drainage of large surficial melt ponds delivers water through cold ice[10].

Glaciers and ice sheets transfer ice from accumulation to ablation zones by internal deformation and by basal motion, comprising sliding of ice over bedrock, regelation and deformation of basal till. Basal motion is implicated in the recent acceleration of many Greenland outlet glaciers[1], whereas sliding causes erosion of glacier beds[11]. Ice deformation is relatively steady; basal motion varies on sub-daily to annual timescales, modulated by the highly dynamic hydrologic system of a glacier[1–6,8,9]. Steady-state theory of subglacial hydraulics describes water transport through high-discharge, low-pressure conduits, low-discharge, high-pressure linked cavities and subglacial canals within a till cover[8,12]. However, as water inputs are rarely steady, the subglacial hydrologic network must adapt to the changing water supply. Every component of the glacial plumbing system is dynamic: canals in till erode and deform; tunnels grow by viscous heating of the walls and collapse by ice creep; cavities enlarge by sliding and collapse by creep; and the orifices between them enlarge by both sliding and viscous dissipation, and close by creep[13,14]. Current models of basal motion are limited in that they do not consider hydrologic transience; none of them successfully span the range of timescales encountered on glaciers and ice sheets. Owing to the lack of a physical model, the 2007 Intergovernmental Panel on Climate Change used a simple projection of recent ice-sheet flow rates in its predictions of sea level change[15]. Recent predictions of the response of ice sheets to warming[16] have modelled enhanced ice flow as being proportional to the modelled rate of melt input, in effect directly predicting basal motion from meteorology. The system is more complex and more interesting than this[17]. Our data suggest that understanding basal motion requires that both the meteorologically driven inputs and the dynamism of the subglacial plumbing system be acknowledged.

We use several features of Kennicott Glacier, a 40-km-long, land-terminating, temperate glacier in the Wrangell Mountains, Alaska (Fig. 1), to study connections between basal motion and hydrology. As observed elsewhere, the surface velocity of Kennicott Glacier varies on diurnal and seasonal timescales[1–5]; in addition, the annual outburst flood of ice-marginal Hidden Creek Lake (HCL) distinctly perturbs the glacier hydrology[18–21]. HCL empties $20–30 \times 10^6$ m^3 of water beneath the lower 15 km of glacier over approximately two days every summer. Owing to the halite-bearing bedrock beneath Kennicott Glacier, electrical conductivity of outlet river water increases with time spent at the glacier bed[18]. Water chemistry therefore offers a proxy for subglacial water residence time and hence hydrologic efficiency. During the 2006 melt season, we measured global positioning system (GPS) positions continuously at five locations distributed over 12.5 km of the ablation area centreline, water pressure every 10 min in HCL and Donoho Falls Lake (DFL), air temperature every 10 or 30 min at the three lowest GPS receivers, and stage and electrical conductivity every 15 min at the Kennicott River glacier outlet (Fig. 1).

The instrumented reach of the glacier switched between two modes of ice motion. Mode A is marked by diurnally varying horizontal velocity at each GPS receiver (Fig. 2b), ice-surface uplift relative to surface-parallel motion (similar to observations in refs 2,4) (Fig. 2c) and low electrical conductivity in outlet water (Fig. 2e). The HCL filling rate increased during Mode A before lake drainage (Fig. 2d). Mode B is characterized by lower mean speeds, the absence of diurnal velocity variations, surface-parallel or bed-convergent ice-surface trajectories and high electrical conductivity. Whereas horizontal speeds and uplift calculated at the five GPS receivers vary, with higher speeds associated with thicker ice, the timing and nature of mode changes are remarkably coordinated between sites. Hereafter, we focus on our longest record (GPS3).

The rate of change of storage is the difference between water inputs to the glacier and water outputs at the terminus: $\partial S/\partial t = Q_{in} - Q_{out}$. We calculate melt inputs using a positive

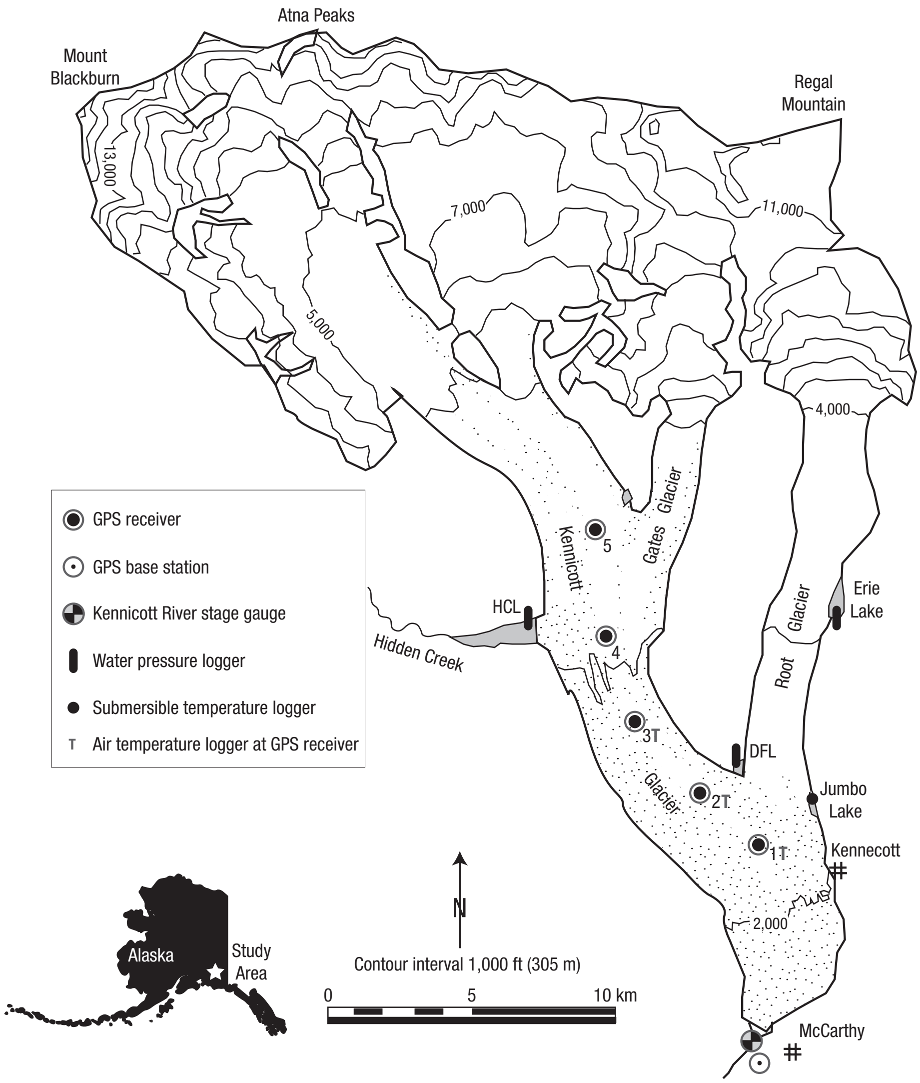

Figure 1 Map of Kennicott Glacier showing instrumentation installed in 2006 to document the connection between basal motion and glacier hydrology. Five GPS receivers document surface ice speeds. One-hour GPS position files were processed statically relative to the fixed base station near McCarthy. Pressure and submersible temperature sensors document levels of four side-glacier lakes and the presence of lakes, respectively. The stage gauge and conductivity probe in the glacier outlet river allow calculation of river discharge and chemistry. Air temperatures logged 0.5–0.7 m above the glacier surface and measured emergence of the GPS support poles constrain calculation of the melt inputs to the glacial hydraulic system. The locations of the towns of McCarthy and Kennecott are represented by a cross-hatch pattern.

degree-day model[22]. Degree-day factors (DDFs), calculated on the basis of ablation and the cumulative air temperature at GPS1–GPS3 over the intervals between ablation measurements, were nearly constant. Using a DDF similar to that deduced from the GPS3 site, and hourly average temperatures from GPS3 (Fig. 2a), adjusted by an adiabatic lapse rate, we calculated melt inputs from the entire Kennicott Glacier watershed. Although others have reported variability of the DDF when used in sub-daily calculations[23], the integrated daily melt is well constrained by the use of temperatures from on-glacier sites, and the timing of maximum melt rate is probably well captured. We approximate flood discharge from HCL into Kennicott Glacier with a normal-distribution time series[19] constrained by lake volume and drainage timing. Precipitation is measured 4.5 km south-west of the glacier terminus[24]. Discharge from streams draining non-glacial hillslopes does not contribute significantly to the glacier-wide water balance and is therefore not included. The Kennicott River hydrograph (Q_{out}, Fig. 2e) is constructed from our stage record by assuming logarithmic velocity profiles[25] in channel segments, with channel roughness chosen to match the rating curve developed during the summers of 1999 and 2000[19]. Channel geometry did not change appreciably until the 2006 outburst, the largest in 10 yr, which scoured $\sim$3 m in the thalweg.

We found that $\partial S/\partial t$ correlates with ice-surface velocity at GPS3 on diurnal, seasonal and outburst-flood timescales (Fig. 3). Mode A diurnal variations in basal motion are particularly clear in day of year (DOY) 180–184 (Fig. 3a). Peaks in basal speed occur shortly after the time of the greatest rate of increase in the calculated water storage; adding a 2.2 h lag to $\partial S/\partial t$ maximizes the correlation between positive $\partial S/\partial t$ and ice speed ($r^2 = 0.64$). Two hours is a reasonable time for supraglacial melt to reach the subglacial hydrologic system, given the $\sim$2.5 km mean travel length to the nearest moulin and a mean supraglacial channel flow speed of 0.5 m s^{-1}.

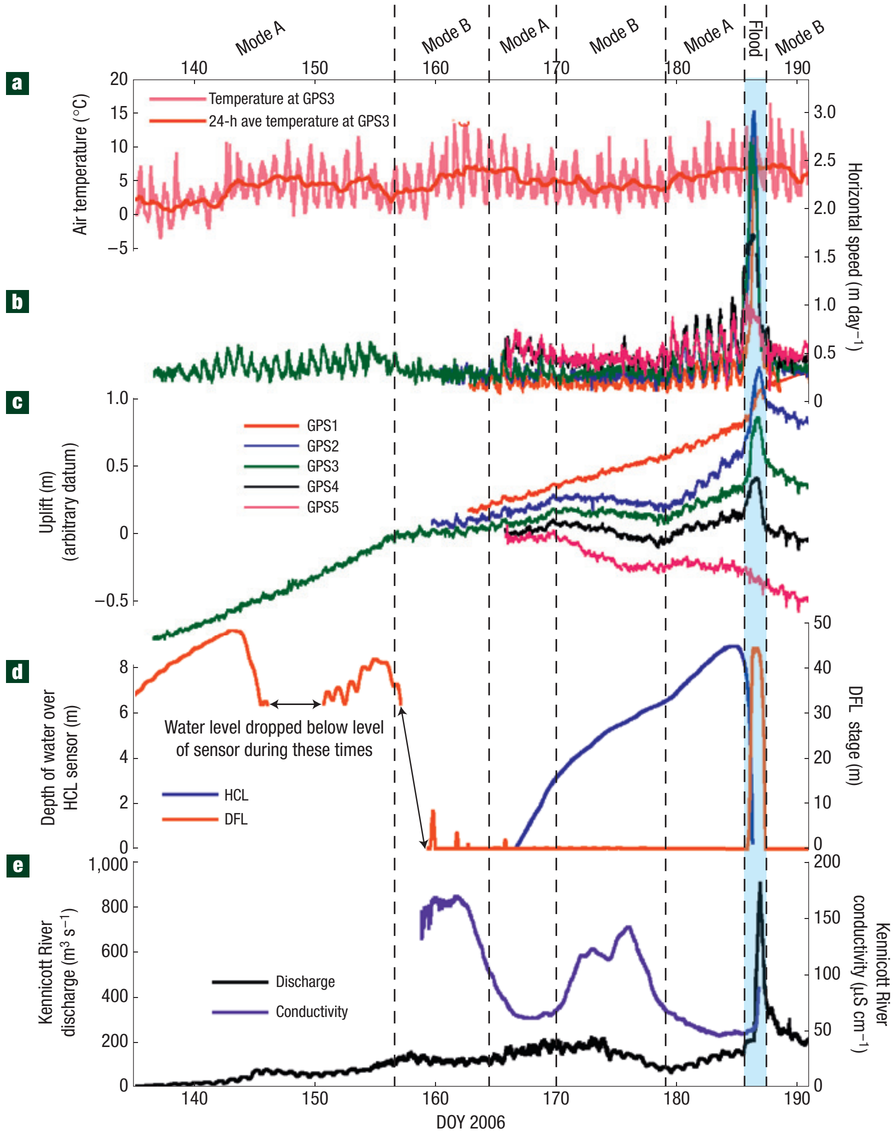

Figure 2 Melt season 2006 record of glacier motion and associated meteorological and hydrological histories. Vertical dashed lines identify two distinct glacier modes, A and B. The period of the HCL outburst flood is shaded light blue. **a**, Half-hour and 24-h averaged air temperatures at GPS3. **b**, Four-hour averaged horizontal ice speeds at each GPS receiver. **c**, Uplift (vertical motion minus the surface-parallel trajectory) at each GPS receiver. **d**, Lake level record at HCL and DFL. The DFL stage is relative to the lake basin floor, whereas the HCL record captures the uppermost 9 m (of ~100 m) of lake filling and draining. **e**, Kennicott River discharge and electrical conductivity.

On the seasonal timescale, before the HCL flood, periods of positive 24-h mean $\partial S/\partial t$ correspond with Mode A episodes (Fig. 3b), whereas periods of negative $\partial S/\partial t$ correspond with Mode B motion. In contrast, no Mode A basal-motion events occurred after the HCL outburst flood (record not shown). Although GPS base-station malfunction soon after the HCL flood interfered with GPS data processing, average daily speeds are still resolvable and remained at or below the pre-flood Mode B average speeds at all five receivers. After the HCL flood, the electrical conductivity of the outlet river was more stable, at a level intermediate between its Mode A lows and Mode B highs.

The abrupt outburst of HCL represents a third type of transient event (Fig. 3c). The ice-marginal DFL and Jumbo Lake basins (Fig. 1) along the flood path—empty at flood initiation—filled with water during this period of maximum water storage[19] (Fig. 2d). Again, the greatest basal speeds coincide with the greatest $\partial S/\partial t$: the rapid rate of increase in water storage between DOY 185.3 and 186.5, driven by HCL drainage into the glacier, correlates remarkably well ($r^2 = 0.67$) with basal speeds nearly sixfold above non-flood maxima.

We suggest a transient hydrology–basal-motion cycle as follows. Surface water is delivered through moulins and fractures to the subglacial, arborescent conduit system[8,26]. When inputs are greater than conduits can transmit, the conduits pressurize relative to the linked cavity system. The resulting pressure gradient—opposite to that of the steady condition[8]—drives some water out of conduits and into the linked cavity system and any basal till. Water that emerges at the glacier terminus has low electrical conductivity because it has passed quickly through conduits while drainage from the linked cavity system is blocked. The water flux from conduits increases the pressure within cavities, temporarily backing water up into available voids within the glacier[26]. This rise in stored englacial

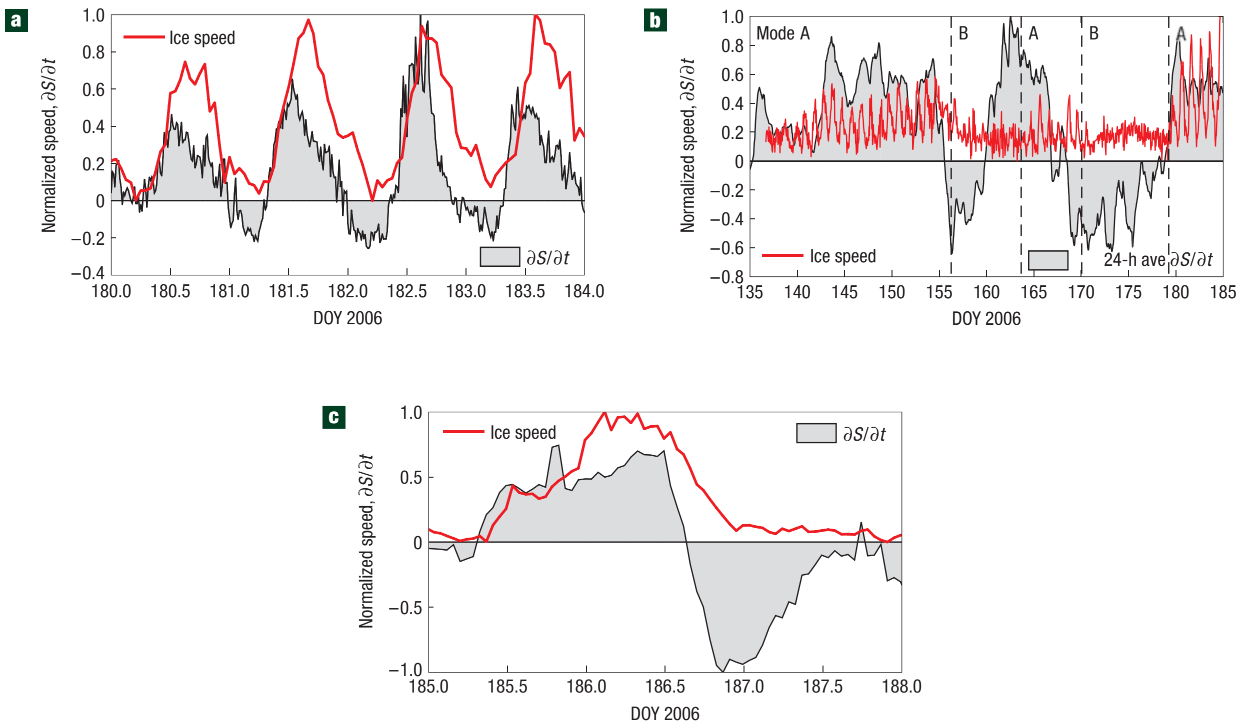

Figure 3 Rate of change of water storage ($\partial S/\partial t$) and ice speed at GPS3. a–c, For diurnal (**a**), seasonal (**b**) and outburst-flood (**c**) timescales. Horizontal ice speed is normalized onto [0, 1] by first subtracting the minimum value over the chosen time interval, then dividing by the maximum value. Rate of change of storage is normalized to lie within [−1, 1] by dividing by the maximum of $\partial S/\partial t$ in the interval, thereby preserving the sign of the change. In **b**, the 24 h running averaging of $\partial S/\partial t$ smoothes diurnal fluctuations from the storage record. At each timescale, sliding coincides with times of increasing water storage, $\partial S/\partial t > 0$.

water increases the pressure head at the bed, which reduces the effective pressure at the bed and promotes basal motion[2,4–6,9]. In the absence of continued high water supply, as might occur in the winter, any enlargement of cavities by sliding would reduce englacial water pressure and serve as a negative feedback.

Eventually, the conduit system recovers its ability to transmit the water supplied to it, either by increases in the size of conduits due to frictional melting of walls, or by a drop in the rate of water input. As the backpressure dissipates, water in the linked cavity system will once again drain towards the conduits, and outlet water will exhibit higher electrical conductivity, reflecting its greater residence time at the bed[18]. As subglacial water moves from the linked cavity system towards the conduits, englacial water is drawn down, reducing the pressure in the areally extensive linked cavity system, which in turn reduces the rate of basal motion[2,9]. Collapse of cavity and conduit roofs[27] due to ice creep towards these low-pressure sites then re-primes the glacier system for another basal-motion cycle.

This link between transient increases in storage and/or water inputs and basal motion is suggested on other alpine glaciers[4] and ice sheets[3], and is associated with rapid drainage of other ice-dammed lakes[6,28]. The few data available from the ablation area of the Greenland Ice Sheet (Fig. 3 of ref. 3) imply that average ice speeds are correlated with average melt production, although evidence to the contrary also exists[29]. We note that in these studies the temporal resolution of the data (typically weeks) is likely to have missed the individual episodes of enhanced basal motion that illuminate the physical processes involved.

Although the melt inputs can perhaps be both documented and modelled in response to climate change scenarios[16], our data indicate that it is inadequate, especially on the short timescales over which basal motion increases, to model the system as responding to water inputs alone; basal motion is not a state variable simply slaved to climate. For instance, drainage of water from the system, dependent on the spatial and temporal evolution of the conduit and cavity systems, must be considered to quantify the rate of change in water storage and thus sliding speed (as in ref. 13). Yet, on marine-terminating glaciers and outlet glaciers of the great ice sheets, water outputs cannot be directly measured, preventing quantitative documentation of the evolution of storage.

Subglacial hydraulic systems are rarely, if ever, in steady state. Models that make this assumption will misrepresent a critical component of ice dynamics. We argue that basal motion increases when subglacial hydraulic efficiency is insufficient to handle the water inputs. This hydraulic efficiency is dynamic: it declines when conduits and cavities collapse and re-establish flow restrictions that backpressure the system when faced with renewed high meltwater input. The lack of high correlation between $\partial S/\partial t$ and surface speed at GPS3 between DOY 157 and 170 is probably the result of such dynamism in the subglacial hydraulic network. One explanation for the lack of diurnal basal-motion response at GPS3 in this period is that an efficient conduit extending to at least GPS3 drained the lower glacier; the high electrical conductivity in this period supports this view. This conduit may not have extended to GPS3 until about DOY 155, reflecting the yearly up-glacier insertion of glacial conduit systems[13]. The conduit would then have collapsed significantly during the extended period of low melt input from DOY 172–179, priming the system for significant diurnal response to renewed high melt rates after DOY 180.

Theoretically, the timescale for collapse is the inverse of the ice thickness cubed when the water pressure is low[27]. On 150-m-thick Bench Glacier, the time needed for cavities to collapse to

$1/e$ of their original height was eight days[4,5], whereas on Kennicott Glacier, which is approximately twice as thick as Bench Glacier, it is roughly one day (Fig. 2). Beneath a 1-km-thick ice sheet, this response time would be much shorter still. Natural lulls of meltwater delivery to the bed of an ice sheet should almost ensure that subsequent high meltwater inputs will encounter a constricted flow system and hence incite basal motion; conversely, if melt delivery is steady or slowly changing, the conduit system can adjust and basal motion may not be induced[17]. As we have shown in the outburst flood at Kennicott, discrete pulsed water inputs generate the strongest transients in the subglacial water system and promote the strongest basal-motion response. If increased melt on Greenland[30] results in greater frequencies, durations or magnitudes of pulsed water delivery to the bed[10], the glacier should respond with more enhanced basal-motion events, and hence greater mean rates of ice discharge towards the sea.

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