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Suzanne P. Anderson

Joseph S. Walder

Robert S. Anderson

Erin R. Kraal

Michelle Cunico

See next page for additional authors

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Integrated hydrologic and hydrochemical observations of Hidden Creek Lake jökulhlaups, Kennicott Glacier, Alaska

Suzanne P. Anderson,[1,2] Joseph S. Walder,[3] Robert S. Anderson,[1,2] Erin R. Kraal,[1] Michelle Cunico,[4] Andrew G. Fountain,[4] and Dennis C. Trabant[5]

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[1] Hidden Creek Lake (HCL), an ice-marginal lake impounded by Kennicott Glacier, Wrangell Mountains, Alaska, fills annually to ~20 to 30 $\times$ 10^6 m^3 and then drains subglacially within 2 to 3 days. During the 1999 and 2000 jökulhlaups, we carried out a series of planned observations around the lake and in the Kennicott River, which drains the glacier. Approximately 20% of the lake volume was contained within a subglacial water "wedge" beneath the ice dam. The entire volume of the lake drains through the wedge; hydraulic head loss through this constriction may be responsible for the fairly symmetrical shape of the HCL outflow hydrographs, deduced from lake level records, basin hypsometry, and collapse of the ice dam. The flood hydrographs in the Kennicott River are similar in shape to the outflow hydrographs, and within error, lake volume matched the river flood volume in both years. Up to 12 $\times$ 10^6 m^3 of water was temporarily stored within the glacier during the 2000 jökulhlaup. During the 2000 jökulhlaup the background flow in the Kennicott River shifted to a dilute chemical composition. As the HCL jökulhlaup progressed, Donoho Falls Lake filled with water whose chemistry was closer to that of the background flow in Kennicott River than to HCL water. Comparison of these chemical signals with typical summer variations in Kennicott River chemistry suggests that the jökulhlaup created high subglacial water pressure that impeded normal drainage of solute-rich water from a distributed drainage system into a conduit system at the glacier bed and even caused flow direction locally to reverse. *INDEX TERMS:* 1827 Hydrology: Glaciology (1863); 1821 Hydrology: Floods; 1836 Hydrology: Hydrologic budget (1655); 1860 Hydrology: Runoff and streamflow; 1871 Hydrology: Surface water quality; *KEYWORDS:* outburst flood, glacier hydrology, water chemistry, glaciers, ice, ice-dammed lakes

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1. Introduction

[2] Jökulhlaups, or glacial outburst floods, are caused by sudden releases of water impounded by a glacier. Jökulhlaups represent a severe perturbation to the subglacial drainage system: water under pressure may erupt as geysers on the ice surface [*Knudsen and Theakstone*, 1988; *Rickman and Rosenkrans*, 1997; *Roberts et al.*, 2000], while at the terminus, discharge may increase by as much as an order of magnitude [*Post and Mayo*, 1971] with river channel morphology downstream largely controlled by the effects of jökulhlaups [*Desloges and Church*, 1992]. Jökulhlaups may pose significant hazards to people and structures, and tools for evaluating hazards—for example, predicting peak discharge as a function of lake characteristics—have been developed [*Clague and Mathews*, 1973; *Clarke*, 1982; *Walder and Costa*, 1996]. Nonetheless, data about outburst floods are sparse, owing to the generally unpredictable nature of these events. Available data are of highly variable quality and commonly consist of no more than rough estimates of lake volume, along with either stream gauge records (sometimes at locations far from the glacier terminus) or simply after-the-fact estimates of peak flood discharge [*Walder and Costa*, 1996]. The outstanding exception has been the data set for jökulhlaups from Grímsvötn, a lake formed in a subglacial caldera in Iceland [*Björnsson*, 1992, 1998].

[3] The goal of the work reported here was to study the jökulhlaup "system" at Kennicott Glacier, Alaska. We determined the volume and rate of release of water impounded in an ice-dammed lake, the magnitude and timing of the resulting flood at the glacier terminus, and the response of an additional basin along the presumed flow path. Jökulhlaups from Hidden Creek Lake, which is

[1]Center for Study of Imaging and Dynamics of the Earth, Department of Earth Sciences, University of California, Santa Cruz, California, USA.
[2]Now at Institute for Arctic and Alpine Research, University of Colorado, Boulder, Colorado, USA.
[3]U.S. Geological Survey, Cascades Volcano Observatory, Vancouver, Washington, USA.
[4]Department of Geology, Portland State University, Portland, Oregon, USA.
[5]U.S. Geological Survey, Fairbanks, Alaska, USA.

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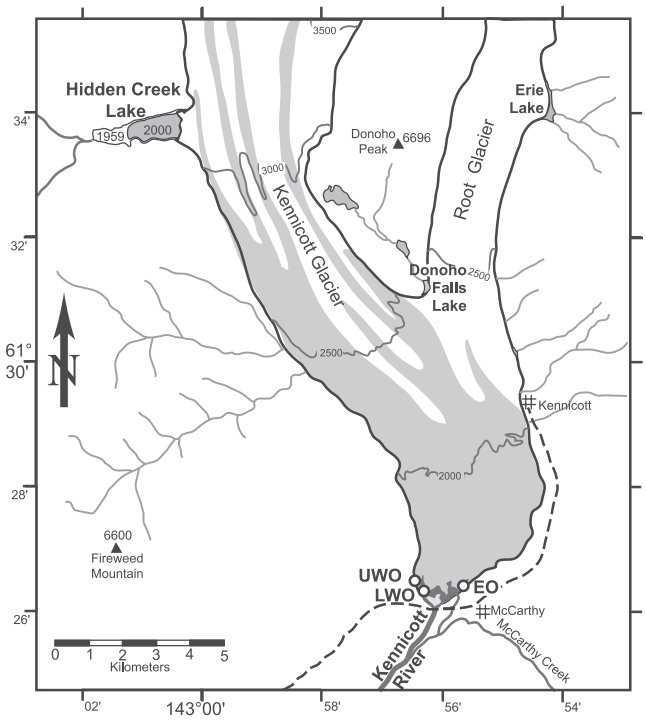

Figure 1. Map of lower half of Kennicott Glacier. Approximate maximum extent of Hidden Creek Lake in 1959 and in 2000 shown. Open circles identify outlet sampling sites: UWO, upper west outlet; LWO, lower west outlet; EO, east outlet. Dashed line shows road; Kennicott River was gauged where road crosses the river. Shading on glacier represents debris-covered ice. Topographic contours shown on ice only; contour and peak elevations (triangles) in feet.

impounded by Kennicott Glacier, occur annually. The present paper is intended primarily to be a presentation of our hydrologic observations; separate papers will deal with the mechanical response of the glacier to lake filling and drainage, as well as possible implications of our observations for explaining how jökulhlaups are triggered. Hydrochemistry of the Kennicott Glacier has been discussed by *Anderson et al.* [2003].

2. Field Site

[4] We investigated jökulhlaups from Hidden Creek Lake (HCL), an ice-dammed lake that forms within the largest deglaciated tributary to the Kennicott Glacier, Wrangell Mountains, south central Alaska (Figure 1). Below dramatic ice falls, the average surface slope of Kennicott Glacier and its two main tributaries, Gates Glacier and Root Glacier, is ~2° (U.S. Geological Survey McCarthy (C-6), Alaska, quadrangle topographic map). Kennicott Glacier is ~40 km long and is largely debris-covered over its lowermost 10 km, with several medial moraines being prominent in the vicinity of HCL. The lake itself is located 16 km from the glacier terminus, in the ablation zone. Ice thickness near HCL, determined from radar measurements to be described elsewhere, reaches ~350 m.

[5] Kennicott Glacier has retreated only slightly from its 1860 Little Ice Age terminus position [*Viereck*, 1967; *Wiles et al.*, 2002], but has thinned considerably (>30 m) in the terminus region over the last century [*Rickman and Rosenkrans*, 1997]. Incursion of Kennicott Glacier up Hidden Creek valley has decreased by at least 300 m since the Little Ice Age [*Rickman and Rosenkrans*, 1997]. Kennecott-type copper deposits in the lower Chitistone Limestone were mined from the east side of the Kennicott valley from 1908 until 1938 [*MacKevett et al.*, 1997]. Observations of Hidden Creek Lake jökulhlaups, extending back to this period of mining operations, show the annual HCL jökulhlaup has tended to occur progressively earlier in the summer. At present, HCL drains ~50 days earlier in the year on average than it did in circa 1920 [*Rickman and Rosenkrans*, 1997].

[6] HCL is the largest of several ice-dammed lakes found along the margins of Kennicott Glacier and its tributaries. Present-day HCL, when at maximum stage, has a surface area of ~0.8 to 1.0 km^2, and a depth near the ice dam of at least 100 m. Geomorphic evidence and historical records show that the maximum stage reached by lake waters has declined by at least 30 m over the last 90 years [*Rickman and Rosenkrans*, 1997]. The main water source to the lake is Hidden Creek, which drains a 27 km^2 valley with several small glaciers in its headwaters.

[7] Weather in the Wrangell Mountains is transitional between that of the wet, temperate coastal region and the dry continental interior. Data from the McCarthy 3 SW cooperative weather station compiled by the National Climate Data Center (http://lwf.ncdc.noaa.gov/oa/ncdc.html) show a mean annual temperature (MAT) of -1.4 ± 0.8°C and mean annual precipitation of 504 ± 102 mm for the period 1985–2000 (omitting seven years with incomplete records). In 1999, the MAT was -2.4°C and total precipitation was 568 mm. In 2000, the MAT was -0.8°C, and total precipitation was 542 mm.

3. Field Campaigns and Methods

[8] Our objectives were to monitor Hidden Creek Lake, the "ice dam"—meaning the part of the Kennicott Glacier adjacent to the lake—and the Kennicott River during a period of several weeks roughly centered on the beginning of lake drainage, during the summers of 1999 and 2000. The known dates of jökulhlaups in recent years [*Rickman and Rosenkrans*, 1997] provided a guide to scheduling our field campaign.

[9] In 1999, lake drainage began a few days after we arrived on site, and the only precursory data we obtained are for background flow and water chemistry of the Kennicott River. In the 2000 our observations started 3 weeks prior to the jökulhlaup. In both years, Kennicott River monitoring continued for more than a month after the jökulhlaup. Investigations in 2000 extended to two other small ice-dammed lakes: Erie Lake, an ice-marginal lake alongside Root Glacier (Figure 1) that drains annually near the time of the HCL jökulhlaup, and Donoho Falls Lake, a small ice-marginal basin at the confluence of the Root and Kennicott Glaciers (Figure 1) that commonly fills with water as HCL drains [*Rickman and Rosenkrans*, 1997].

3.1. Kennicott Glacier

[10] In both years, a surveying total station was set up on a bedrock knob north of the ice dam, with its location

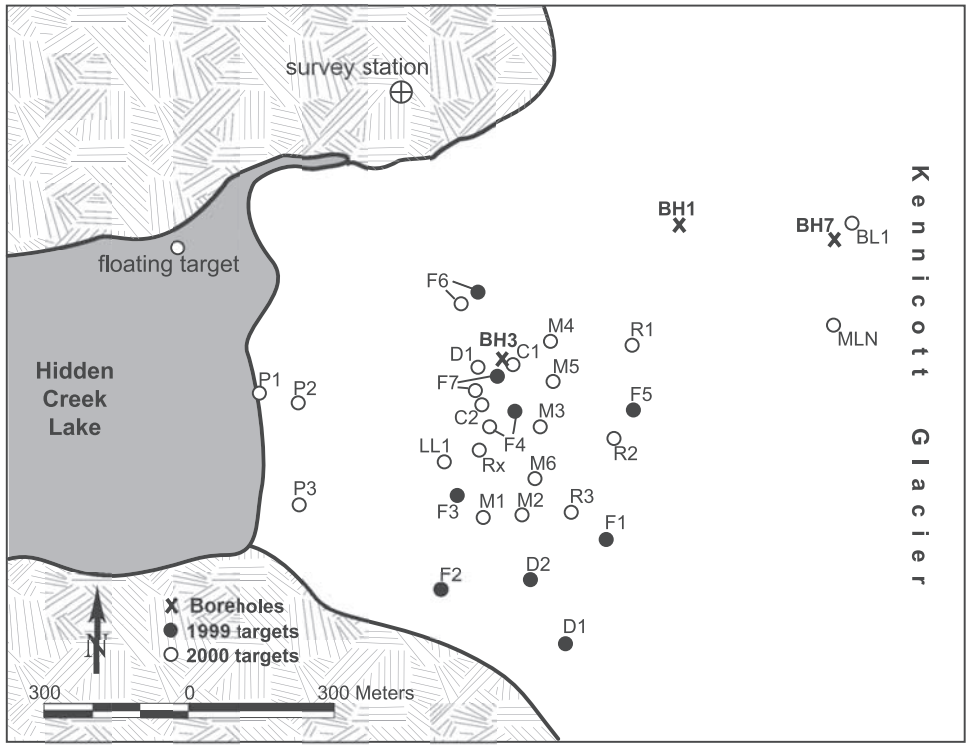

Figure 2. Map of Hidden Creek Lake, showing relative locations of lake, survey station, pertinent survey targets, and boreholes. Targets P1, P2, and P3 were lost when a section of the ice dam calved.

referenced to the North American geodetic datum NAD-1983 by GPS. Survey targets were set up on the glacier near the lake (Figure 2). Crevasses precluded access to some parts of the ice dam. The entire target array was surveyed typically four times per day, or more frequently once the lake began to drain. The probable error in survey measurements is ~10 mm; total displacement of most targets was at least several m in both the horizontal and the vertical. We used vertical angle measurements to determine the length of the exposed survey pole and calculate daily average ablation rate at each stake.

[11] Boreholes were drilled in the glacier using standard hot water-drilling methods [*Taylor*, 1984]. None of our boreholes reached the base of the glacier. We placed pressure transducers (70 m maximum range) in the boreholes and collected one useful record in 1999 and three records in 2000. Spot measurements of ice thickness of the ice dam were made with ground-penetrating radar. Glacier surface conditions on the ice dam made continuous radar profiling infeasible.

3.2. Hidden Creek Lake

[12] Lake level in HCL was monitored differently during the two field seasons. In 1999, lake level peaked and then fell by ~0.5 m, before we managed to deploy three pressure transducers in the lake. All three transducers functioned well until they went dry two days later. Total lake drawdown recorded was 56.1 m. In 2000, we placed four pressure transducers in the lake ~3 weeks before lake drainage, but they were all destroyed when a large piece of the ice dam calved off. We subsequently deployed additional transducers near the shore and supplemented these with repeated optical surveys to the lake surface in 2000. The diverse transducer and survey data were combined to produce a lake level history for 2000. The methods are described more fully in the auxiliary material[1].

[13] Lake basin topography is based on data from *Rickman and Rosenkrans* [1997], who surveyed accessible parts of the basin following lake drainage in 1994 and 1995. We supplemented their measurements with surveying in 1999 and spot sonar soundings in 2000. The combined data were used to construct a topographic map of the lake basin, from which we could compute the hypsometric function $A(h_L)$, the area of the lake basin as a function of elevation [*Cunico*, 2003].

[14] Several temperature soundings of the lake were performed in both 1999 and 2000 by lowering an encapsulated thermistor on a cable. Water samples were collected with a Van Dorn style sampler at 4 m intervals from 4 to 16 m depth and at 18 m depth in 1999 and at 10 m intervals from 0 to 80 m depth in 2000. The water samples were taken approximately at the centerline of the lake and within ~100 m of the floating ice tongue.

3.3. Hidden Creek

[15] In 2000, we measured the input to HCL from Hidden Creek, the only surface stream that feeds the lake. A stage gauge, consisting of an acoustic sensor attached to a cantilever over the channel [see also *Anderson et al.*,

[1] Auxiliary material is available via Web browser or via Anonymous FTP from ftp://agu.org/apend/jf/2002JF000004. Information on searching and submitting auxiliarry material is found at http://www.agu.org/pubs/esupp/about.html.

1999], was set up several hundred m upstream of the lake. Stage was measured at a frequency of 1 Hz, and the average and standard deviation were recorded automatically at 15-min intervals. Six direct measurements of discharge were performed, once with a current meter and five times by the salt-dilution method [*Kite*, 1993], either shortly after the gauge was installed or just before it was dismantled. More than 90% of the automated stage measurements fell within the calibrated range from 3 to 8 $m^3\ s^{-1}$.

3.4. Kennicott River

[16] We installed a stream gauge on a metal footbridge that spans the west fork of the Kennicott River ~500 m downstream from the glacier terminus (Figure 1), at which point the channel is ~80 m wide. (The east fork of the Kennicott River, which carries flow only during the Hidden Creek Lake jökulhlaup was not gauged. Spot measurements during the jökulhlaup show it carried no more than 1% of the total flow.) The gauge consisted of two acoustic look-down sensors (to provide redundancy) fixed to the bridge ~4 m above the water surface. As at Hidden Creek, the sensors measure the distance to the water surface at 1 Hz; the average and standard deviation of readings over 15-min intervals were recorded with a data logger. On average, the standard deviation was 2.3% of the measured stage value in any interval, but it increased with increasing stage, such that the standard deviation at the highest stages was almost 7%.

[17] To develop a rating curve, we measured stream discharge at a range of stage heights. We measured water depth and mean velocity at 15 to 30 points across the ~80 m wide channel, following standard U.S. Geological Survey discharge measurement techniques [*Buchanan and Somers*, 1969]. Measurement points were spaced to minimize the variance in discharge between sections. Measured discharge ranged from 80 to 415 $m^3\ s^{-1}$, including measurements made within a few hours of the flood peak in both years. As there was little or no change in channel bed geometry from 1999 to 2000 [*Kraal*, 2001], we used a single rating curve developed from measurements in both years to convert stage to discharge (see auxiliary material). Errors arising from uncertainty in the rating curve are greatest at high discharges, where the rating curve is steepest.

[18] Turbidity, electrical conductivity, and water temperature were measured at 15-min intervals with an instrument package suspended in a stilling well fixed to a downstream bridge pier. The instruments were suspended a fixed distance down the stilling well, so their submergence varied with stage; on average they were ~0.5 m below the surface in a location where water depth varied between 2.1 and 3.5 m. Measurements in this stilling well, which had an open bottom, holes in the side, and was located in the river thalweg, should be representative of the river as a whole.

[19] Water samples were collected from the river at least twice daily from 12 July to 6 August in 1999, and from 30 June to 5 August in 2000. The Kennicott River was sampled ~15 m upstream of the stream gauge on the footbridge. In 2000 we also sampled the three main outlets along the terminus of the Kennicott Glacier (Figure 1) before and during the HCL outburst, as well as rainwater, and water flowing on the surface of the Root Glacier. Suspended sediment concentration was determined from the mass of sediment retained on 0.45 μm filters from 0.5 L samples. These data were used to develop a rating curve relating turbidity to suspended sediment concentration.

3.5. Donoho Falls Lake and Erie Lake

[20] In 2000, we placed a pressure transducer ~50 m above the deepest part of Donoho Falls Lake basin. A waterproof temperature sensor was placed beneath a small rock cairn in the deepest portion of the basin to identify the time when the basin began to fill with water, an event we expected to produce a shift from the diurnal temperature fluctuations in air to a relatively constant (near 0°C) temperature in the lake water. In addition, five sample bottles were affixed to rocks embedded in cairns on the basin floor to collect samples of the lake water. A right-angle pipefitting attached to the bottle top allowed water to enter but prevented sediment settling out of the water column from entering the bottle. We placed these samplers and instruments in the dry basin in early July 2000, and retrieved them 2 days after the Hidden Creek Lake jökulhlaup.

[21] We collected surface water samples from Erie Lake before it drained in 2000. We relied on reports of local flightseeing pilots, who fly the Kennicott area daily, for information on the drainage of Erie Lake in both years.

3.6. Water Chemical Analysis

[22] All water samples were collected in high-density polyethylene bottles (HDPE), rinsed with sample, and were vacuum filtered through 0.45 μm filters (Gelman Metricel) in the field, generally within 24 hours of collection. One aliquot of each sample was acidified to <pH 2 with ultrapure concentrated nitric acid, while a second aliquot was not acidified. Samples were stored in acid-washed HDPE bottles under ambient conditions in the field, and then were refrigerated in the lab until analyzed.

[23] Cations and silica were analyzed by inductively coupled plasma optical emission spectrometry (Perkin Elmer Optima 4300 DV). Anions were analyzed by ion chromatography (Dionex DX-100). Alkalinity was determined by titration with 0.01N HCl using the Gran method [*Drever*, 1997; *Gran*, 1952] with an Orion 950 pH meter and digital titrator. Cation, anion, and silica concentrations on replicate analyses were better than 0.1 mg L^{-1}, except for calcium, for which replicates were within 0.3 mg L^{-1}. Charge balance averaged $-0.7 \pm 1.8\%$.

4. Results

[24] The day of year and maximum stage at which HCL drained varied considerably during the years we have observations (Figure 3). Maximum lake water level varied by 10 m between 1999 and 2002, and drainage in 2001 was 31 days earlier in the year than in 2002.

4.1. Hidden Creek Lake Water Balance

[25] The rate of change of lake volume V is given by

$$\frac{dV}{dt} = Q_i - Q_o \tag{1}$$

where $Q_i(t)$ is the sum of the inputs and $Q_o(t)$ is the sum of the outputs. $Q_i(t)$ can be further split into components as follows

$$Q_i(t) = Q_{\mathrm{HC}}(t) + Q_{\mathrm{melt}}(t) + Q_{\mathrm{precip}}(t) \tag{2}$$

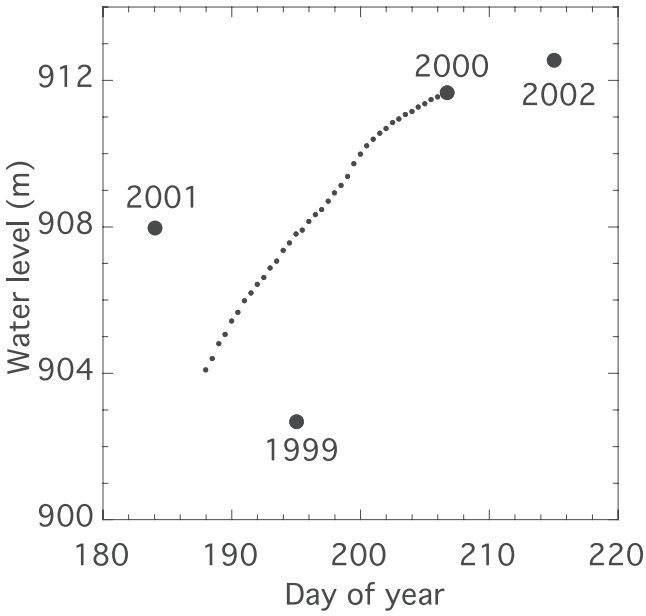

Figure 3. Maximum water level in Hidden Creek Lake and date of jökulhlaup for period 1999–2002. For 2000, series of small dots shows the rise in lake level preceding the jökulhlaup. HCL level in 2001 obtained by differential GPS survey of shoreline in a snow bank, 10 days after drainage; level in 2002 determined by hand level survey to a bench mark 1 day after drainage (M. Loso, personal communication, 2002). All measurements are referenced to the NAD-1983 geodetic datum.

where $Q_{\mathrm{HC}}(t)$ is inflow to the lake from Hidden Creek, $Q_{\mathrm{melt}}(t)$ is meltwater input from the glacier surface, and $Q_{\mathrm{precip}}(t)$ is runoff owing to rain that fell either directly into the lake, on the local valley walls, or on the glacier surface. Any stable leak from the lake prior to the jökulhlaup is not directly measurable, and we proceed by assuming that $Q_o = 0$ during the period of rising lake stage—an assumption that in fact proves to be entirely reasonable. The total rate of change of lake volume is given by

$$\frac{dV}{dt} = \frac{dV_v}{dt} + \frac{dV_w}{dt} \tag{3}$$

where the subscript v denotes the "visible" portion of the lake and subscript w denotes water stored in a "wedge" beneath the ice margin (Figure 4). The existence of V_w is made manifest by survey data (summarized below), which indicate that parts of the ice dam went afloat as the lake rose and subsequently fell by as much as 15 m over a broad domain as the lake drained. This sort of ice dam deformation associated with lake filling and drainage has been recognized before [*Kasper*, 1989; *Kasper and Johnson*, 1991], but its importance for water balance has not been appreciated.

[26] Change in lake level h_L in 2000 and the hypsometric function $A(h_L)$ are shown in Figure 5. The rate of change of volume V_v is related to lake stage $h_L(t)$ and the hypsometric function $A(h_L)$ by

$$\frac{dV_v}{dt} = [1 - f(h_L)]A(h_L)\frac{dh_L}{dt} \tag{4}$$

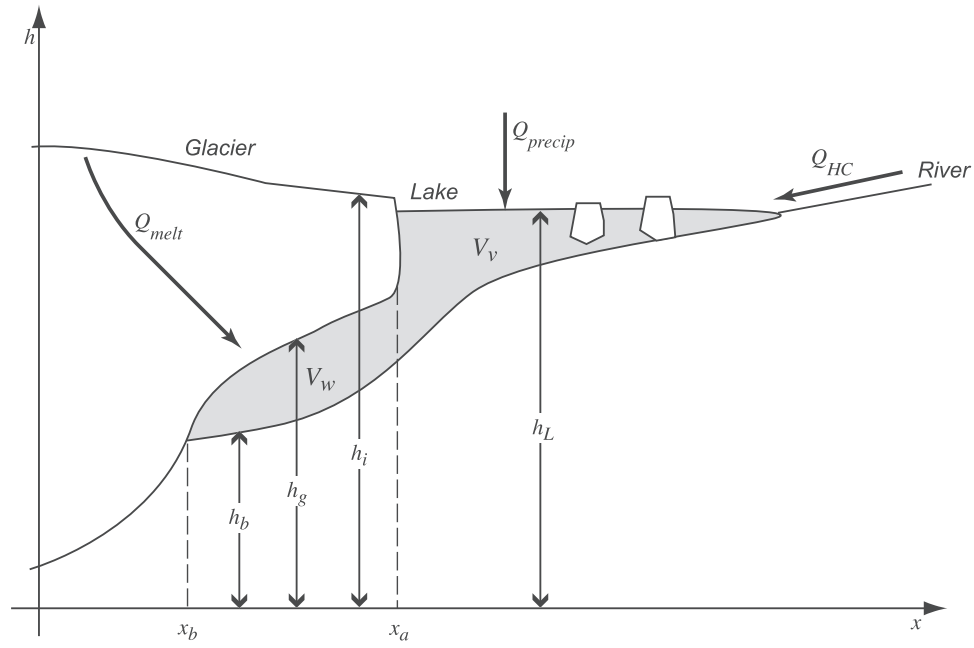

Figure 4. Sketch of water balance in Hidden Creek Lake. During lake level rise, prior to development of any subglacial outlet, the temporally integrated water inputs must match the volume history of the lake. Lake volume includes that in the visible basin (V_v) and that in a subglacial wedge (V_w), greatly exaggerated in this sketch. The former is deduced from a history of lake level, $h_L(t)$, and the basin hypsometry, $A(h_L)$; the latter from history of the deflection of the ice documented by repeat surveys of markers on the ice surface. Inputs are discharge of Hidden Creek, Q_{HC}, the melt of surface ice in the region near the lake, Q_{melt}, and precipitation on the area not drained by Hidden Creek, Q_{precip}.

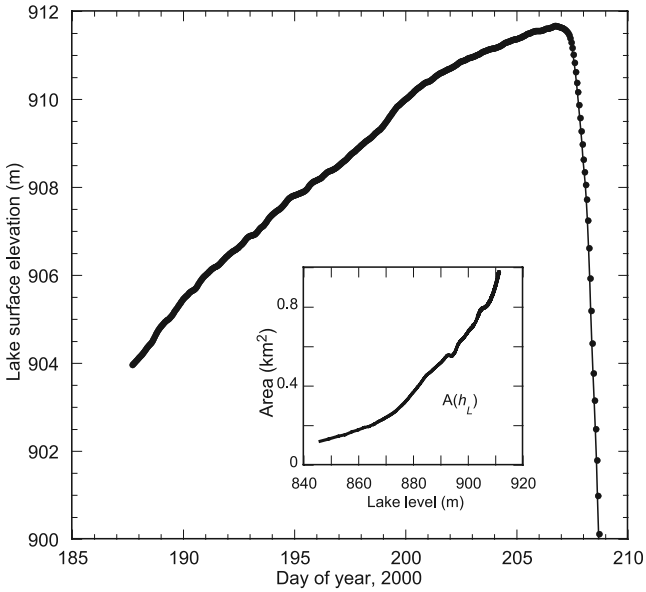

Figure 5. Measured level of Hidden Creek Lake in 2000 during filling and early stages of drainage of the lake. Inset shows hypsometric function $A(h_L)$.

where $f(h_L)$ is the fraction of lake surface area covered by grounded ice blocks. We have no direct measurements of $f(h_L)$, but note that during rising lake stage, roughly 20% of the lake surface was covered by icebergs, nearly all of which seemed to drift owing to wind; accordingly, we have set $f(h_L) = 0$ for the period of rising stage. The hypsometric function and lake stage record provide no information, however, about V_w. To determine the rate of change of V_w we use as a proxy measure the rate of change of $h_i(x, y, t)$, the ice surface elevation within the surveyed "ice dam" area. The justification for this method is as follows: Let $h_g(x, y, t)$ and $h_b(x, y, t)$ denote, respectively, the elevation of the glacier bottom and the elevation of the bed as a function of easting x, northing y, and time t. We then have

$$\frac{dV_w}{dt} = \frac{d}{dt}\int_S \int \left[h_g(x,y,t) - h_b(x,y,t)\right] dxdy \qquad (5)$$

where the integral is performed over the domain S affected by filling of the lake. If the ice thickness and the domain S were not changing with time and if the ice were everywhere afloat within S, then $dh_i/dt = d(h_g - h_b)/dt$ and the ice surface displacements would constitute a complete proxy data set. Implicit in equation (5) is the assumption that the ice adjusts instantaneously to rising lake level. If the ice dam were responding by viscoelastic flexure, as an ice shelf [e.g., *Vaughan*, 1995], target motions would lag lake level change. However, the ice within the domain of interest is pervasively fractured, and (as will be shown in a future manuscript) vertical target movement seems to be almost entirely due to fault motion.

[27] Surface displacement profiles from the Kennicott Glacier ice dam during lake filling in July 2000, plotted as a function of x (easting) only, can be fit to a sigmoidal curve (Figure 6). These displacement profiles are reasonably consistent with the idea that the part of the ice dam nearest the lake was floating, or nearly so, and only weakly coupled to the ice farther away. The 2000 data do not resolve any systematic variation in h_i with y (northing), but we did not have any targets near the north and south margins of the ice dam and cannot rule out the possibility that the ice was grounded in those areas. Indeed, there is circumstantial evidence of marginal grounding: (1) We did not observe formation of any "moats" between the glacier and the rock walls, as might have been expected if the ice were lifting off its bed at the ice dam margins, and (2) survey data collected during lake drainage in 1999 showed practically no surface drop at a target within ~150 m of the southern margin of the ice dam. Taking these considerations into account, we approximate equation (5) by

$$\frac{dV_w}{dt} \approx cW \int_{x_a}^{x_b} \frac{\partial h_i}{\partial t}(x, y_m, t)dx \qquad (6)$$

where x_a is the average easting of the free face, and x_b is the average easting of the grounding line, y_m is the northing of the ice dam midline, W is the average width of the ice dam, and c is a factor that would equal 1 if $\partial h_i/\partial t$ were independent of y, but a bit less if $\partial h_i/\partial t$ fell off toward the edges. We choose $c = 0.8$, a very rough estimate. We have also assumed in equation (6) that x_a and x_b do not change with time. Although we have no direct information about changes in x_b through time, the vertical displacement profiles (Figure 6) certainly do not support any interpretation that x_b changed much. The "free face" x_a of the ice dam moved lakeward, but as long as ice dam deformation involved no net volumetric strain, then the enlargement of the domain S, and the increase in V_w associated with that enlargement, would be exactly offset by a decrease in the volume V_v. For simplicity of bookkeeping we ignore the

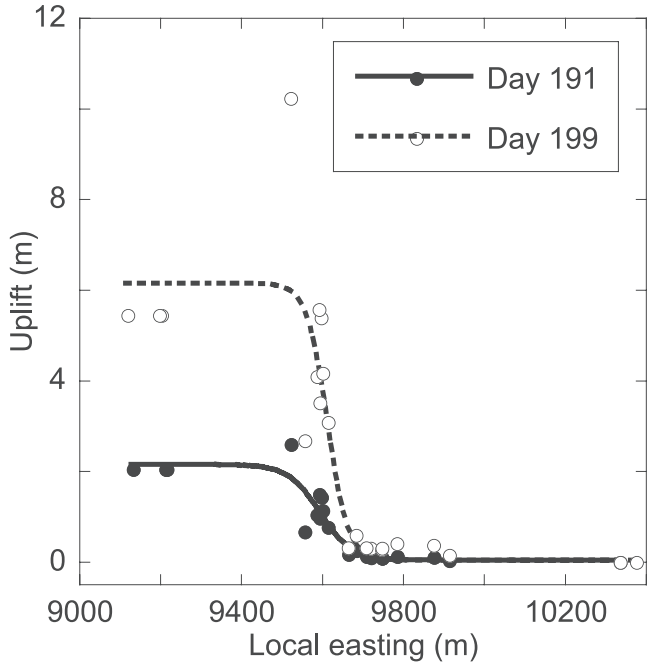

Figure 6. Ice surface displacement profiles from Kennicott Glacier ice dam as the lake filled in 2000. Uplift relative to target positions on day 186.75 is shown as a function of easting; the ice edge is located at ~9150 m. Data for the two dates shown are fit to a sigmoidal curve.

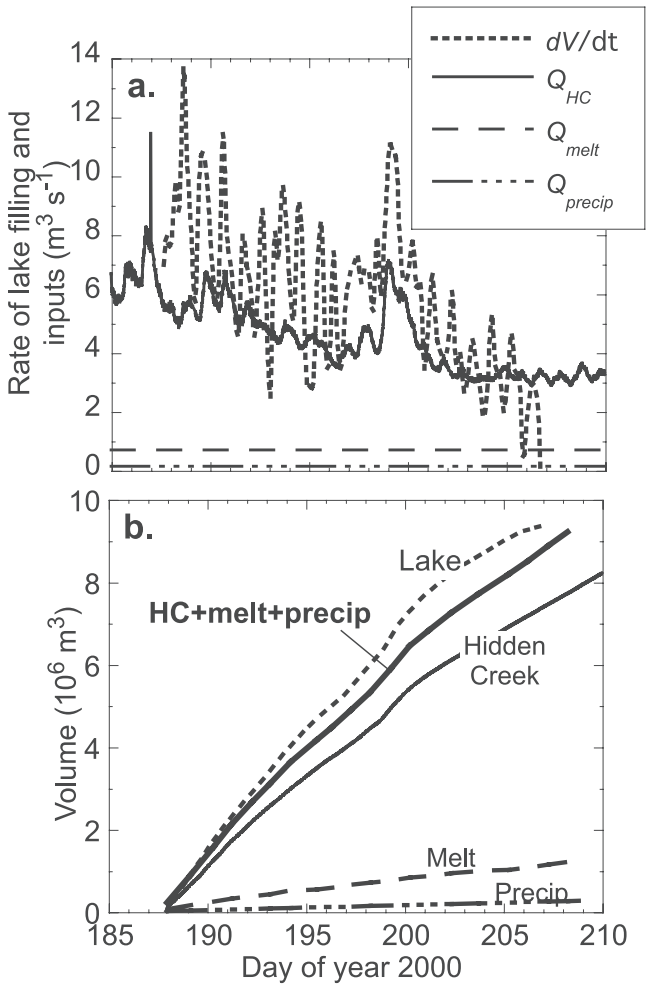

Figure 7. (a) Rates of visible lake and wedge volume increase with time and inputs from measured discharge from Hidden Creek (Q_{HC}) and approximate contributions (shown as averages over the observation period) owing to ablation runoff from Kennicott Glacier (Q_{melt}) and precipitation on the area not drained by Hidden Creek (Q_{precip}). (b) Integrated volume changes in HCL and from the sources illustrated in Figure 7a. Heavy solid curve shows sum of inputs from Q_{HC}, Q_{melt}, and Q_{precip}. Mismatch between lake volume and sum of inputs (HC + melt + precip) amounts to only ~10% of the total inflow into the lake, which we attribute to measurement errors.

enlargement of V_w owing to ice deformation, as the net effect on the total volume $V = V_v + V_w$ vanishes.

[28] Net rates of volume increase (dV_v/dt and dV_w/dt) and rates of recharge owing to various sources (Q_{HC}, Q_{melt}, and Q_{precip}) are shown in Figure 7, along with corresponding plots of accumulated water volumes. The contribution due to ablation, Q_{melt}, was calculated from

$$Q_{melt} = \left(\frac{\rho_i}{\rho_w}\right) \int \dot{m} dA \qquad (7)$$

where ρ_i is the density of ice, ρ_w is the density of water, $\dot{m}$ is the ice ablation rate, and the integral is taken over A_l, the area of the glacier that contributes meltwater to the lake. We used the average measured value of $\dot{m}$ over the target array, which during the period of investigation ranged from 29 to 75 mm d^{-1}, and averaged 68 mm d^{-1}. No supraglacial streams existed in the ice dam region, as all runoff quickly entered one of the many crevasses. The surface area A_l from which melt was routed toward the lake rather than into the main subglacial drainage system is not known a priori, but may be estimated by appealing to *Shreve*'s [1972] argument about the likely path of englacial water flow. Details are presented in the auxiliary material. In essence, A_l is that area for which water entering the glacier is routed to the bed lakeward of the putative "seal", or drainage divide on the bed (Figure 4). The location of the seal was roughly delineated with the aid of radar data (which are discussed more fully in a separate manuscript in preparation). We estimated $A_l = 0.95$ km^2 and took $\rho_i/\rho_w = 0.917$. During the 20-day period of measured lake level rise, ablation accounted for 1.2×10^6 m^3 of water (with a probable error of ~25%). This figure is ~60% of the difference between lake volume change and input from Hidden Creek.

[29] The rainfall contribution, Q_{precip}, included the amounts that fell either directly into the lake, on to the glacier surface area A_l that contributed ablation runoff, or on slopes that drain directly to the lake rather than to Hidden Creek. The first two components constitute a total catchment area of ~2 km^2. The third component is not sharply defined, but using topographic maps (U.S. Geological Survey McCarthy (C-6), Alaska, and McCarthy (C-7), Alaska quadrangles), we estimate the pertinent area as ~6.4 km^2, bringing the total source area for storm runoff to ~8.4 km^2. We did not measure rainfall at the lake or on the glacier, and use as proxy data from the nearest National Weather Service station located near McCarthy. Total precipitation there was 28 mm during the period of measured lake level rise (5 to 24 July 2000), and thus we estimate the additional rainfall input to the lake during this period as $\sim 0.24 \times 10^6$ m^3, with an error of perhaps 25%.

[30] Variability in dV_v/dt reflects both the intrinsic noisiness of numerical differentiation as well as uncertainty in the hypsometric function $A(h_L)$. Error in dV_w/dt arises from numerical differentiation and assumptions made about the geometry of the subglacial wedge. The calculated accumulated volumes are probably more accurate, as numerical differentiation is no longer an issue, and the errors have largely to do with uncertainties in $A(h_L)$ and the geometry of the subglacial wedge.

[31] The net sinks (V_v and V_w) exceed the net volume from sources (Q_{HC}, Q_{melt}, and Q_{precip}) for the period of measurement by $\sim 0.6 \times 10^6$ m^3, a discrepancy of only ~10% of the total inflow to the lake (Figure 7). In flux terms, this discrepancy is only 0.3 m^3 s^{-1}. Despite the approximations, we are confident that within measurement error, inputs and change in lake volume were indeed in balance. Our water balance calculations provide no basis for concluding that a stable leak existed prior to the jökulhlaup.

[32] Note that melting of icebergs within the lake, or indeed of the floating part of the ice margin, does not affect the lake level, although it does transfer water from solid state to liquid. However, the iceberg volume at the time of initiation of lake drainage is important, as it reduces the total water volume in the lake, and hence the volume available to drain from the lake, from that inferred with basin hypsom-

etry. With the fractional coverage of icebergs roughly 20%, and estimating from the freeboard of most floating ice that the mean iceberg thickness was ~10 m, the volume of water displaced by icebergs was roughly 2×10^6 m^3. Considering our poor knowledge of both iceberg thickness and their coverage of the surface, we suggest using a range from 1.5 to 2.5×10^6 m^3 for water volume displaced by icebergs at the start of lake drainage. A value in this range should be subtracted from the estimated total volume of water released during the jökulhlaups in both 1999 and 2000.

4.2. Drainage of Hidden Creek Lake

[33] Hidden Creek Lake drained completely in both 1999 and 2000, and anecdotal reports indicate that the lake always drains completely. We wish to compute dV/dt, the outflow hydrograph at the lake, and compare it to the flood hydrograph measured on the Kennicott River. The outflow from the lake is given by

$$Q_0 = Q_i - \frac{dV_v}{dt} - \frac{dV_w}{dt}. \tag{8}$$

This formal mathematical procedure should not be taken to mean that we envisage two separate reservoirs, one subaerial and one subglacial, draining in parallel. The subglacial wedge is clearly an extension of the lake beneath the glacier margin, and we have treated it separately because its history is decipherable by measuring glacier surface motion rather than by measuring lake stage. If the jökulhlaups indeed involve escape of water through a small number of discrete subglacial conduits, the usual physical picture of the process [*Björnsson*, 1992; *Clarke*, 1982; *Nye*, 1976], then in fact the entire volume of the lake escapes by passing through the zone of glacier/bed separation that we have called the wedge.

[34] Methods for determining dV_v/dt and dV_w/dt were worked out first for the more robust data collected in 2000 and then applied retrospectively to the data for 1999.

4.2.1. HCL Drainage in 2000

[35] Hidden Creek Lake reached peak stage in the evening of 23 July, or day of year (hereafter abbreviated day) 206.75. The drainage rate from the subaerial lake, dV_v/dt, was computed using equation (4) for a hypothetical iceberg grounding function $f(h_L)$, the fraction of the lake area occupied by grounded icebergs, which increases smoothly from zero at maximum lake stage ($h_{L,\ \max}$), as given by

$$f(h_L) = 1 - \exp\left[\frac{-(h_{L,\max} - h_w)}{\lambda}\right] \tag{9}$$

where the parameter λ is determined by the integral constraint:

$$V_v(h_{L,\max}) - V_v(h_{L,\min}) - V_{\text{icebergs}} = \int_{h_{L,\min}}^{h_{L,\max}} [1 - f(h_L)]A(h_L)dh_L. \tag{10}$$

$V_v(h_{L,\ \max})$ and $V_v(h_{L,\ \min})$ are the nominal lake volumes corresponding to $h_{L,\ \max}$ and $h_{L,\ \min}$, which are, respectively, the peak stage attained and the minimum measured lake stage. Values of these parameters for 2000 are given in Table 1. V_{icebergs} is the estimated volume displaced by icebergs, taken as 2.0×10^6 m^3, in line with the discussion above. The functional form of $f(h_L)$ reflects our observations that there seemed to be no grounded icebergs as the lake filled, and that the fraction of the lake surface occupied by grounded icebergs increased as the lake drained. One may interpret the quantity λ as the characteristic value of drawdown over which grounding occurs; λ is likely to depend on the distribution of iceberg sizes and the hypsometry of the basin. Iterative evaluation of the integral in equation (10) yielded the value $\lambda = 200$ m for 2000.

Table 1. Computation of the Lake Outflow Hydrograph

	1999	2000
Time of $h_{L,\ \max}$, day of year	196.6[a]	206.75
$h_{L,\ \max}$, m	902.25	911.68
$h_{L,\ \min}$, m	846.44	866.0
$V_v(h_{L,\ \max})$, 10^6 m^3	20.77[b]	28.25[b]
$V_v(h_{L,\ \min})$, 10^6 m^3	2.22[b]	5.24[b]
V_{icebergs}, 10^6 m^3	2.0	2.0
V_v, iceberg corrected, 10^6 m^3	16.9 ± 2.5	26.2 ± 2.5
V_w, 10^6 m^3	4.4 ± 1.5	5.7 ± 2.0
V_{HC} during drainage, 10^6 m^3	0.8 ± 0.2[c]	0.8 ± 0.1
V_{total}, 10^6 m^3	22.1 ± 3.0	32.7 ± 3.3

[a]Maximum lake level occurred as much as 20 hours earlier; this is the start of our instrumental record.
[b]Values of $V_v(h)$ determined from *Cunico*'s [2003] hypsometric relations.
[c]Hidden Creek discharge assumed to be the same in 1999 as measured in 2000.

[36] Our lake stage measurements end at t_0 = day 209.45, before the lake was fully drained. We extrapolated dV_v/dt to t_f = day 211.0 (the time at which, as we shall see, dV_w/dt is effectively zero) by assuming a functional form

$$\left.\frac{dV_v}{dt}\right|_{\text{extrapolated}} = Q_0\left(\frac{t_f - t}{t_f - t_0}\right)\exp\left[\frac{-(t - t_0)}{\tau}\right] \tag{11}$$

where Q_0 is the discharge at t_0 and the parameter τ is determined using an iterative scheme requiring that the total extrapolated outflow volume equal $V_v(h_{L,\ \min})$. For the 2000 outflow record we found $\tau = 0.75$ day. The total water volume released from the subaerial lake was 26.2×10^6 m^3. The probable error in this figure is $\sim 2.5 \times 10^6$ m^3 owing to uncertainties in lake hypsometry (especially for the lowest part of the basin) and the volume of icebergs.

[37] The rate at which water is released from the subglacial wedge is assumed to mimic the rate of drop of the ice surface within the ice dam region (Figure 8). There were clearly two mechanically distinct domains: the eight targets on the ice dam within ~500 m of the lake began to fall at about day 207.5, ~18 hours after lake level began to fall. Targets farther away from the lake did not begin to drop until about day 209.0. Likely reasons for the delay will be discussed in a future manuscript. We calculated dV_w/dt by supposing that the subglacial wedge comprised two domains, each of which drained everywhere at the same rate. The wedge in the domain nearest the lake was assumed to have a uniform initial thickness H of 15 m, about the average net vertical decline of the pertinent survey targets

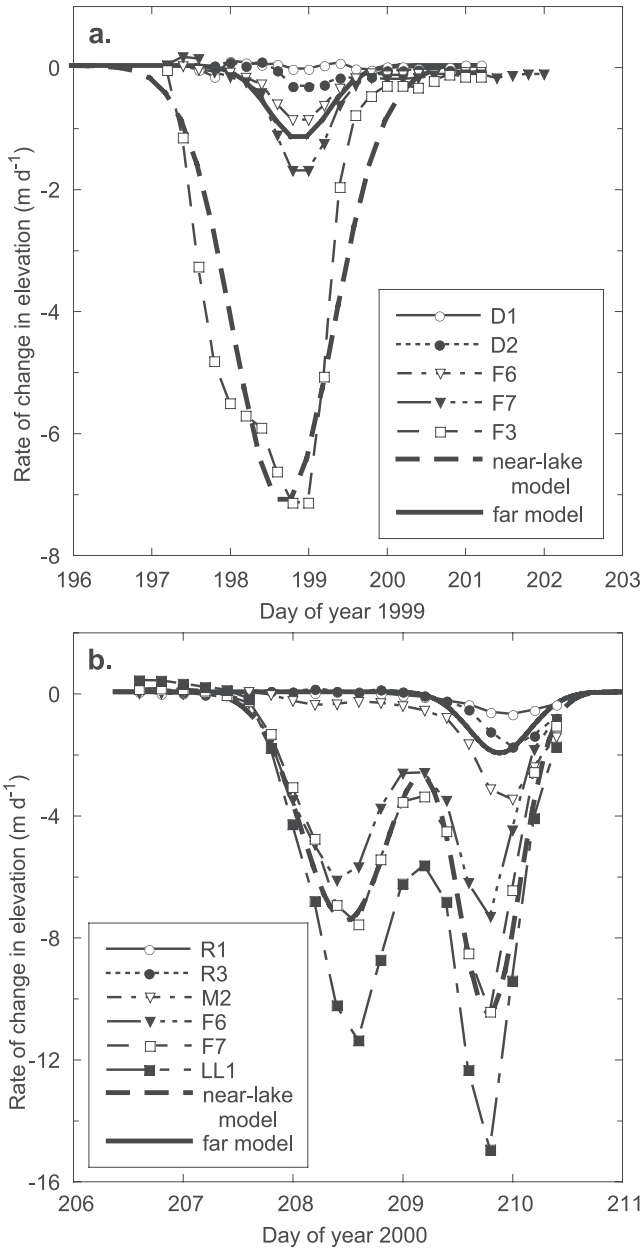

Figure 8. Vertical displacement rates on the ice dam surface in (a) 1999 and (b) 2000. Data from subsets of representative targets shown (locations found in Figure 2), along with model fits to the near-lake targets and the targets far from the lake, as described in the text.

(see auxiliary material), and to thin at a rate given by the empirical function

$$\left.\frac{dH}{dt}\right|_{\text{nearlake}} = -\eta_1 \exp\left[-\left(\frac{t-t_1}{\delta_1}\right)^2\right] - \eta_2 \exp\left[-\left(\frac{t-t_2}{\delta_2}\right)^2\right] \tag{12}$$

where $\eta_1 = 7.5$ m d^{-1}, $\eta_2 = 10.5$ m d^{-1}, t_1 = day 208.5, t_2 = day 209.8, $\delta_1 = 0.56$ day, $\delta_2 = 0.41$ day. The constants were chosen so that the assumed thinning rate mimics the average target motions and the total decline in H over time is 15 m. The wedge domain farther from the lake was assumed to have a uniform thickness of 1.4 m, again, consistent with the net vertical decline of the pertinent survey targets (see auxiliary material), and to thin at a rate given by the empirical function

$$\left.\frac{dH}{dt}\right|_{\text{far}} = -\eta_3 \exp\left[-\left(\frac{t-t_3}{\delta_3}\right)^2\right] \tag{13}$$

where $\eta_3 = 2.0$ m d^{-1}, t_3 = day 209.9, and $\delta_3 = 0.40$ day. Again assuming that ice surface motions mimicked withdrawal of water, this domain of the subglacial wedge had a breadth of ~300 m (see Figure 2). The net rate of withdrawal of water from the subglacial wedge was finally calculated using equation (6) with dH/dt substituting for $\partial h_i/\partial t$, again with c = 0.8 reflecting the likely lack of ice/bed separation near the ice dam margins. The estimated total volume released from the subglacial wedge was 5.7×10^6 m^3; as the method of arriving at this figure involved several simplifying assumptions and approximations, we assign an error of 2.0×10^6 m^3.

[38] The reconstructed total outflow hydrograph for 2000 is shown in Figure 9. The total water volume released,

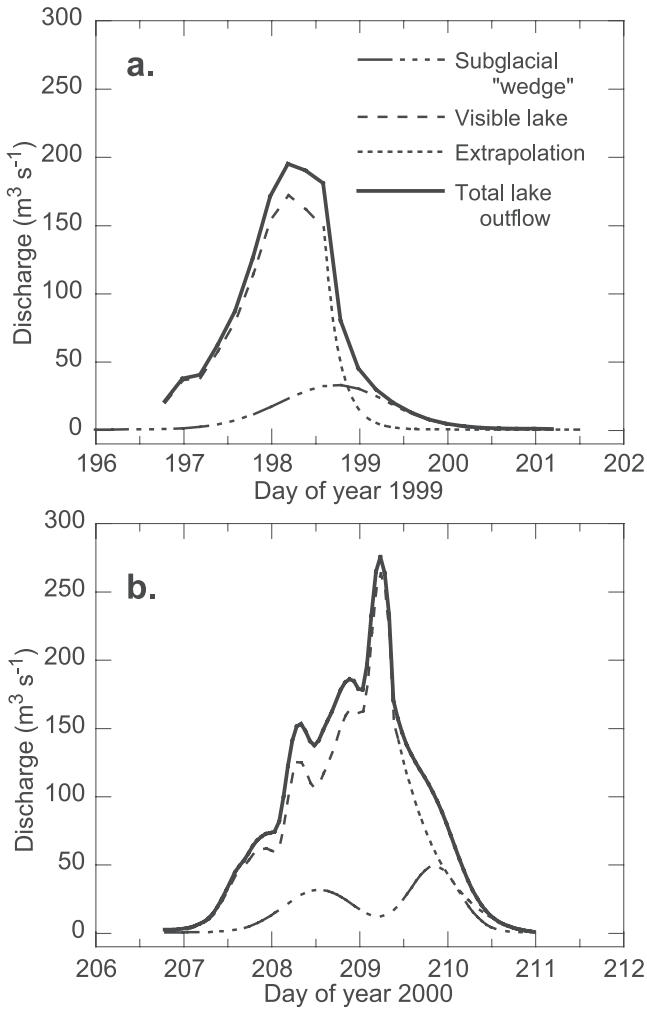

Figure 9. Outflow hydrographs from Hidden Creek Lake in (a) 1999 and (b) 2000. The hydrographs represent the rate of water escape from both the subaerial part of the lake, V_v, (approximately corrected for iceberg volume) and the subglacial wedge, V_w. Part of the falling limb of the hydrograph involves extrapolation, as detailed in the text.

including the recharge from Hidden Creek during the period of lake drainage, was 32.7 $\times$ 10^6 m^3, with an estimated error of 3.3 $\times$ 10^6 m^3 (Table 1). About 18% of the total lake volume was stored beneath the ice dam.

4.2.2. HCL Drainage in 1999

[39] Peak stage occurred sometime between the evening of 14 July (day 195) and the morning of 15 July (day 196). The measured lake stage record began at about day 196.6; our instrumental record missed about the first 0.3 to 0.4 m of stage drop, corresponding to a water volume of $\sim$0.3 $\times$ 10^6 m^3.

[40] The drainage rate from the subaerial lake was computed using equation (8) with the hypothetical "iceberg grounding" function $f(h_L)$ given by equation (9) and the values in Table 1. We again found λ = 200 m; the extrapolated tail of the subaerial hydrograph was fit by equation (11) with τ = 0.19 day. The total water volume released from the subaerial lake was 16.9 $\times$ 10^6 m^3 (including our estimate of the small fraction that drained before our transducers were set up) with a probable error again of $\sim$2.5 $\times$ 10^6 m^3 owing to uncertainties in lake hypsometry and the volume of icebergs.

[41] The rate of water release from the subglacial wedge was calculated in the same way as for the drainage of 2000. Figure 8 shows the rate at which targets on the ice surface moved vertically. One target (F3) dropped by a much greater amount than the others, and considering its location (Figure 2), we suggest that as in 2000, there were two mechanically distinct domains within the ice dam. On the basis of the total downdrop of target F3, we supposed that the thick part of the wedge nearest the lake had an effective thickness H = 11.4 m and that the thin part of the wedge farther from the lake had an effective thickness H = 1.2 m, with thickness histories given by

$$\left.\frac{dH}{dt}\right|_{\text{nearlake}} = -\eta_4 \exp\left[-\left(\frac{t-t_4}{\delta_4}\right)^2\right] \tag{14}$$

and

$$\left.\frac{dH}{dt}\right|_{\text{far}} = -\eta_5 \exp\left[-\left(\frac{t-t_5}{\delta_5}\right)^2\right] \tag{15}$$

where η_4 = 7.2 m d^{-1}, η_5 = 1.2 m d^{-1}, t_4 = day 198.7, t_5 = day 198.9, δ_4 = 0.9 day, δ_5 = 0.55 day. As for the drainage of 2000, the net rate of withdrawal of water from the subglacial wedge was finally calculated using equation (6) with dH/dt substituted for $\partial h_i/\partial t$ and c = 0.8. The estimated total volume released from the subglacial wedge was 4.4 $\times$ 10^6 m^3, and we assign a probable error of 1.5 $\times$ 10^6 m^3.

[42] The reconstructed complete outflow hydrograph for 1999 is shown in Figure 9. The total water volume released was 22.1 $\times$ 10^6 m^3, with a probable error of 3.0 $\times$ 10^6 m^3 (Table 1). About 21% of the total lake volume was stored beneath the ice dam.

4.3. Lake Temperature

[43] All our temperature soundings were done in the eastern end of the lake, relatively near the ice dam. Except for a thin layer near the surface, water in approximately the uppermost 20 m was nearly isothermal at $\sim$0.2°C, with temperature below that point increasing monotonically with depth (Figure 10). Considering the basin hypsometry, the uppermost 20 m of the lake at the time of our measurements would have accounted for about half of the lake volume at that time, so if the isothermal "cap" were widespread, then about half the lake water would have been at a temperature of $\sim$0.2°C or less. Data reported by *Friend* [1988], also shown in Figure 10, display similar characteristics to our own, although he found a zone of relatively large temperature gradient closer to the surface than did we. If we extrapolate to the lake bed (that is, to a depth of around 100 m in the deepest parts of the basin) it seems unlikely that the temperature at the bed exceeded $\sim$1.5°C. Although these data by no means constitute a thorough lake temperature survey, we suggest that the volume-averaged lake temperature was certainly no more than 1°C. Knowledge of lake temperature is important if one wishes to compare predictions of physically based models of jökulhlaup hydrographs [*Clarke*, 1982] to measured hydrographs.

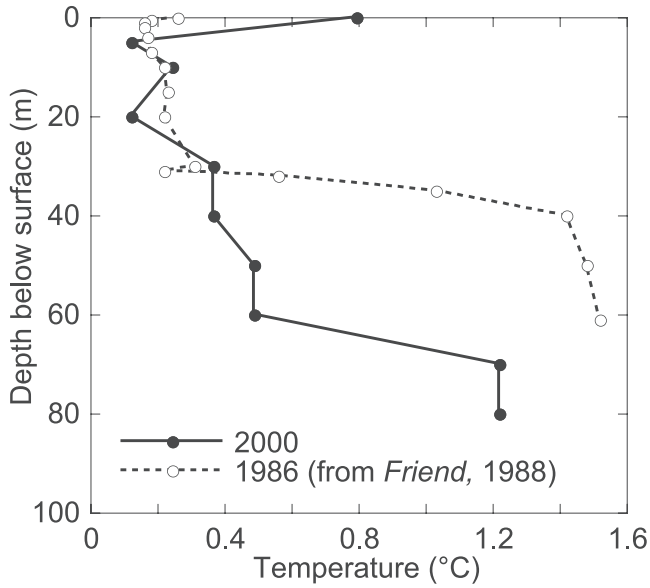

Figure 10. Water temperature profiles in HCL.

4.4. Borehole Water Levels

[44] Reliable records were obtained from three boreholes in 2000 (Figure 11). Two of these boreholes were drilled within ice probably overlying the subglacial water wedge: borehole 3 was drilled within a part of the ice dam that underwent large vertical motion, around 10 to 15 m, as the lake filled and drained, while borehole 1 was drilled a bit north of targets that underwent moderate (around 1 m) vertical motion during lake filling and drainage. In both of these boreholes, water level relative to the glacier surface was nearly constant as the lake rose, with no diurnal fluctuations; water-level drop in both of these boreholes fairly closely mimicked lake level decline as the lake drained. In contrast, borehole 7 was drilled near one of the two survey targets whose motions did not seem to be driven by lake filling and drainage. The water level in this borehole showed strong diurnal fluctuations continuing into midday on day 208, that is, $\sim$48 hours after the lake began to drain. The water level in borehole 7 then dropped more or

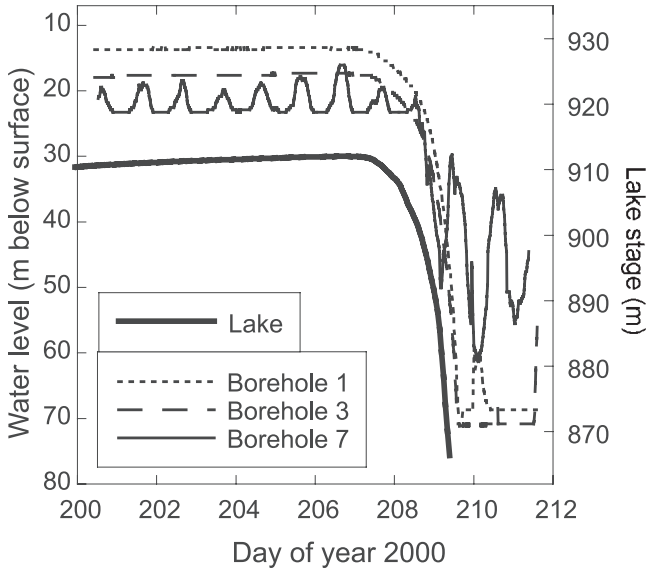

Figure 11. Borehole water level records in 2000. Concurrent Hidden Creek Lake level plotted at the same scale.

less monotonically, at about the same rate as the lake, for ~12 hours before resuming diurnal fluctuations, now with a considerably greater amplitude than before the lake drained.

4.5. Kennicott River

[45] Discharge in the Kennicott River undergoes both diurnal oscillations and long-period (10–14 day) oscillations over the course of the summer (Figure 12; data in the auxiliary material). In both 1999 and 2000, the highest discharge and the highest suspended sediment concentration in our observations occurred during the Hidden Creek Lake jökulhlaup. To compute a flood hydrograph for comparison with the lake outflow hydrograph, the background flow (which is ~50% of the peak flood discharge) must be identified. To this end, we used a geometric approach to hydrograph separation (Figure 13). After masking the period of clearly flood-influenced discharge, a double sinusoid was fit to the discharge for a period of 11 to 19 days around the time of flood. The calculated background was then subtracted from the total discharge to obtain the flood discharge. The precise choice of the flood-influenced period makes little difference in the calculation. The error in discharge, shown as gray bands in Figure 13, is computed from the standard deviation of the stage measurements. Error increases with discharge because the standard deviation of the stage measurement increases with stage. The integrated flood discharge in the Kennicott River in 1999 was 18.5×10^6 m^3 with error of $\sim 5 \times 10^6$ m^3, and in 2000 was $\sim 38.0 \times 10^6$ m^3 with error of $\sim 10 \times 10^6$ m^3 (Table 2).

[46] In both 1999 and 2000, peak flood discharge occurred ~0.5 day after peak lake outflow (Figure 14 and Table 2). This implies a mean travel time for the flood peak of ~0.4 m s^{-1}. Peak flood discharge is nearly the same in both years, despite a factor of 2 difference in lake volume. In both years, suspended sediment concentration reached a peak earlier than discharge in the Kennicott River. Although the exact time of the sediment peak is not known in 1999 because of problems with our turbidity sensor, our hand samples show that the sediment peak occurred ~8 hours before the discharge peak. In 2000, peak suspended sediment concentration was lower than in 1999 by nearly a factor of 2, and the lag between the sediment peak and the discharge peak was ~12 hours. Curiously, peak suspended sediment concentration coincided with peak discharge in the lake outflow hydrograph for both years.

4.6. Other Lakes

[47] In 1999, Erie Lake drained on 20 July (day 201), ~3 days after the HCL jökulhlaup. In 2000, Erie Lake drained on 24 July (day 206), a day before HCL. Because of its relatively small volume ($\sim 5 \times 10^6$ m^3), the Erie Lake jökulhlaup was not perceptible in the Kennicott River discharge.

[48] Our water level record in Donoho Falls Lake (DFL) in 2000 (Figure 14b) confirmed anecdotal reports that this basin fills during the HCL jökulhlaup. The basin began to fill on 26 July (day 208), when the Kennicott River was nearly 50 m^3 s^{-1} above its background discharge. Total Kennicott River discharge at this time was ~190 m^3 s^{-1}, a level that had been reached, but not exceeded, earlier in the summer (Figure 12). The onset of filling of DFL also coincided with a marked increase in suspended sediment concentrations in the Kennicott River. Water in DFL rose to ~60 m within a few hours and stayed nearly constant until the time of the discharge peak in the Kennicott River, after which the water level fell, at first slowly and then

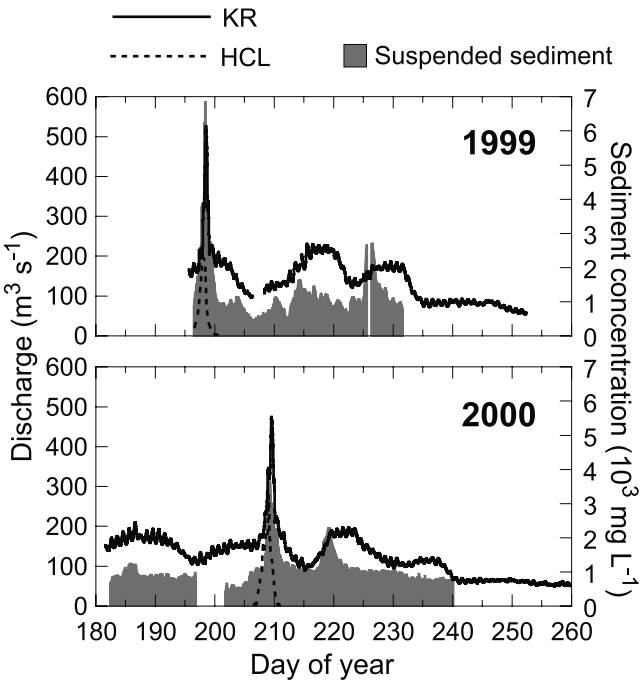

Figure 12. Kennicott River (KR) hydrographs and suspended sediment concentrations in summer of 1999 and 2000. The 2000 record extends from 29 June (day 181) to 21 September (day 260), while the 1999 record is shorter. Dashed lines show the Hidden Creek Lake (HCL) outflow hydrographs from Figure 9. Sediment records terminate before discharge records and have breaks due to sensor malfunctions.

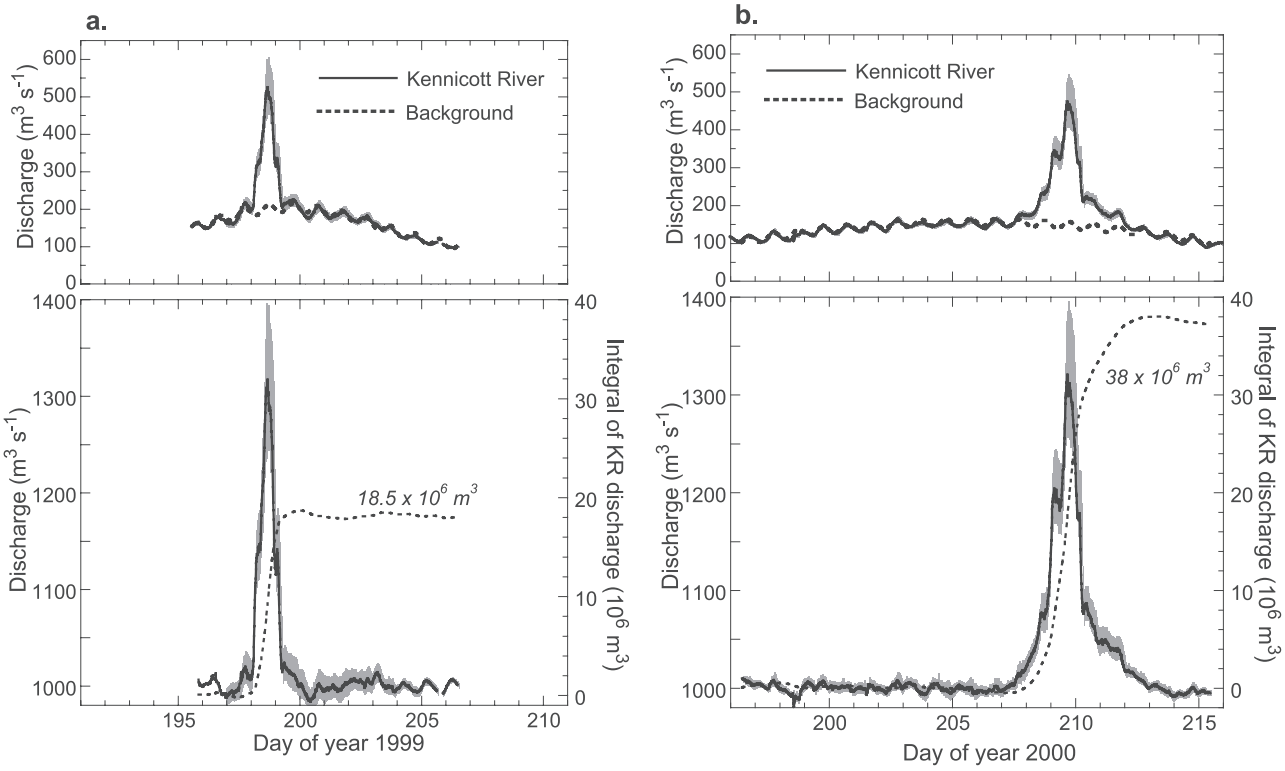

Figure 13. Hydrograph separation analysis of the Hidden Creek Lake jökulhlaups in the Kennicott River for (a) 1999 and (b) 2000. (top) Total Kennicott River discharge, with 1-sigma error computed from standard deviations of stage measurements shown in gray, and the background river discharge obtained by fitting a sinusoid to the region outside the obvious flood-enhanced discharge. (bottom) Kennicott River flood discharge obtained by differencing the curves in the top panel. Integral of the Kennicott River flood hydrograph shown with dashed line. Scales are identical in Figures 13a and 13b.

more rapidly than it had risen. Six hours after the discharge peak in the Kennicott River, the DFL basin was again empty.

4.7. Solute Chemistry

[49] We present data from 2000, which encompass pre-jökulhlaup conditions as well as the flood and aftermath. Similar patterns were seen in the records from 1999 (for full data sets, see the auxiliary material). The chemical composition of water in HCL and Erie Lake differed substantially from that of the Kennicott River at the time of the jökulhlaup (Table 3 and Figure 15). Both lakes had high Ca and alkalinity (HCO_3) concentrations compared with the average composition of Kennicott River water, consistent with the predominance of limestone bedrock within their drainage basins [*MacKevett*, 1978]. The lakes also have high SO_4 and Si concentrations, and low Cl, Na and K concentrations compared with Kennicott River water. Under normal melt season conditions, concentrations of Ca, HCO_3, Na, Cl and K in the Kennicott River vary by a factor of 2 or more (nearly an order of magnitude in the case of Cl) in a pattern that is lagged relative to discharge variations [*Anderson et al.*, 2003]. The Hidden Creek Lake jökulhlaup disrupts these patterns of solute variation (Figure 15), while drainage from the much smaller Erie Lake has unclear effects on Kennicott River chemistry.

4.7.1. Comparison of Outlets

[50] Although most of our water sampling was concentrated in the Kennicott River, in 2000 we also sampled at the three main individual outlets near where they emerge from the glacier (Figure 1). These outlets differed from each other

Table 2. Hidden Creek Lake Jökulhlaup Statistics

Year	Date of Maximum Lake Level	$h_{L,\ max}$, m	Lake Volume, 10^6 m^3	Date of Peak Outflow From Lake	Peak Lake Outflow, m^3 s^{-1}	Date of Peak River Discharge	Peak River Total Discharge, m^3 s^{-1}	Peak River Flood Discharge, m^3 s^{-1}	Integral of River Flood Discharge, 10^6 m^3
1999	196.6[a]	902.25	22.1 ± 3.0	198.2	190	198.71	520	315	18.5 ± 5
2000	206.7	911.7	32.7 ± 3.3	209.25	275	209.71	470	320	38 ± 10
2001		907.98				184			
2002		912.56				215			

[a]Maximum lake level in 1999 occurred as much as 20 hours earlier than this, the start of our instrumental record.

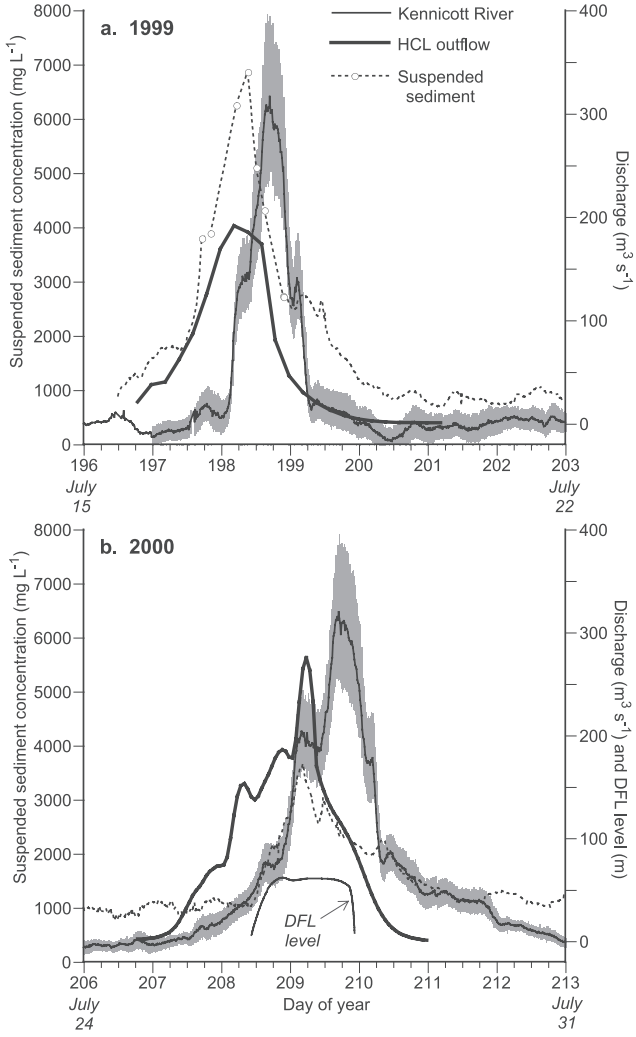

Figure 14. Comparison of the lake outflow hydrograph and the river flood hydrograph for the Hidden Creek Lake jökulhlaups of (a) 1999 and (b) 2000. Suspended sediment concentration in the Kennicott River based on turbidity, except for period with hand samples only (data points) in 1999. Water level in Donoho Falls Lake (DFL; see Figure 1 for location) shown for 2000.

in chemical composition and in response to the jökulhlaup. The upper west and the east outlets were substantial in size (with discharges a few tens of $m^3\ s^{-1}$), but the greatest discharge was through the lower west outlet. Water from the lower west outlet emerges into the bottom of a small ice-marginal lake, visibly roiling the lake surface. Although we did not measure discharge at any of the outlets, flow in the upper west and east outlets did not appear to change during the jökulhlaup, while flow out of the lower west outlet lake did visibly increase. The chemistry of water from the lower west outlet varied in concert with variations in the Kennicott River, while the chemistry of water from the other two outlets did not (Figure 15). We conclude that floodwater emerged only at a single outlet and thus, at least in some portion of the lower glacier, floodwaters appeared to be confined to one conduit.

4.7.2. Background River Chemistry

[51] In general terms, variations in solute concentrations in the Kennicott River during the jökulhlaup are consistent with the idea that Hidden Creek Lake water simply mixed with the normal background flow in the subglacial drainage system. Na, Cl, and K concentrations plummeted during the flood, approaching the low concentrations of these solutes in Hidden Creek Lake water. Si, SO_4, and NO_3 concentrations in the river increased to broad peaks (the highest for these solutes in our observations in the Kennicott River) during the flood, reflecting the relatively high concentrations of these solutes in Hidden Creek Lake water. Ca, Mg, and HCO_3 changed little in the river during the jökulhlaup, as one might expect given the similar concentrations of these solutes in Hidden Creek Lake and Kennicott River at the time of the jökulhlaup. To test this qualitative description we use a two-component mixing model.

[52] We assume that negligible solute is gained in the ~1 km long reach between the glacier terminus and our gauging station. The net flux of a solute species past our gauging station at time t is then given by

$$C_{i,\mathrm{tot}}(t)Q_{\mathrm{tot}}(t) = C_{i,\mathrm{bg}}(t)Q_{\mathrm{bg}}(t) + C_{i,\mathrm{fld}}(t)Q_{\mathrm{fld}}(t) \qquad (16)$$

where C_i refers to the concentration of species i and Q to discharge, Subscript bg indicates background flow, subscript fld indicates flood flow above background (assumed

Table 3. Average Chemical Composition of Selected Waters Around Kennicott Glacier in 2000[a]

Source	Number of Samples	K	Ca	Mg	Na	Si	Cl	NO_3	SO_4	HCO_3	Total Dissolved Solids
Hidden Creek[b]	4	0.1 ± 0.1	25.1 ± 1.8	3.8 ± 0.1	0.4 ± 0.1	0.9 ± 0.1	0.2 ± 0.1	0.1 ± 0.1	15.7 ± 1.1	75.1 ± 4.7	121.6 ± 5.7
Hidden Creek Lake[c]	9	0.1 ± 0.0	28.2 ± 0.4	4.6 ± 0.1	0.5 ± 0.0	1.4 ± 0.0	0.1 ± 0.0	0.2 ± 0.0	18.8 ± 0.2	86.1 ± 0.3	140.4 ± 0.4
Erie Lake[d]	1	0.2	29.7	4.0	0.7	0.9	0.2	0.2	36.8	60.2	113.0
Donoho Creek[e]	1	0.4	24.3	2.1	0.7	1.2	0.1	0.2	6.9	75.7	111.7
Donoho Falls Lake[f]	5	0.6 ± 0.1	23.7 ± 0.7	3.3 ± 0.2	1.3 ± 0.2	1.9 ± 0.4	0.1 ± 0.0	0.1 ± 0.0	18.5 ± 2.1	74.7 ± 2.3	124.4 ± 4.5
Rain[g]	4	0.6 ± 0.3	2.8 ± 1.8	0.2 ± 0.1	0.2 ± 0.1	0.0 ± 0.0	0.4 ± 0.2	0.1 ± 0.2	0.3 ± 0.2	7.2 ± 4.3	11.9 ± 6.9
Glacier surface streamg[h]	1	0.0	1.4	0.0	0.0	0.0	0.0	0.0	0.3	4.1	5.9

[a]All concentrations in mg L^{-1}.
[b]Collected 1–2 and 30 July 2000.
[c]Average of samples collected 4 July 2000 at depths ranging from 0 to 80 m.
[d]Collected 13 July 2000.
[e]Collected 4 July 2000.
[f]Samples retrieved 29 July 2000, two days after lake drained.
[g]Samples collected between 14 and 28 July 2000.
[h]Collected on Root Glacier on 4 July 2000.

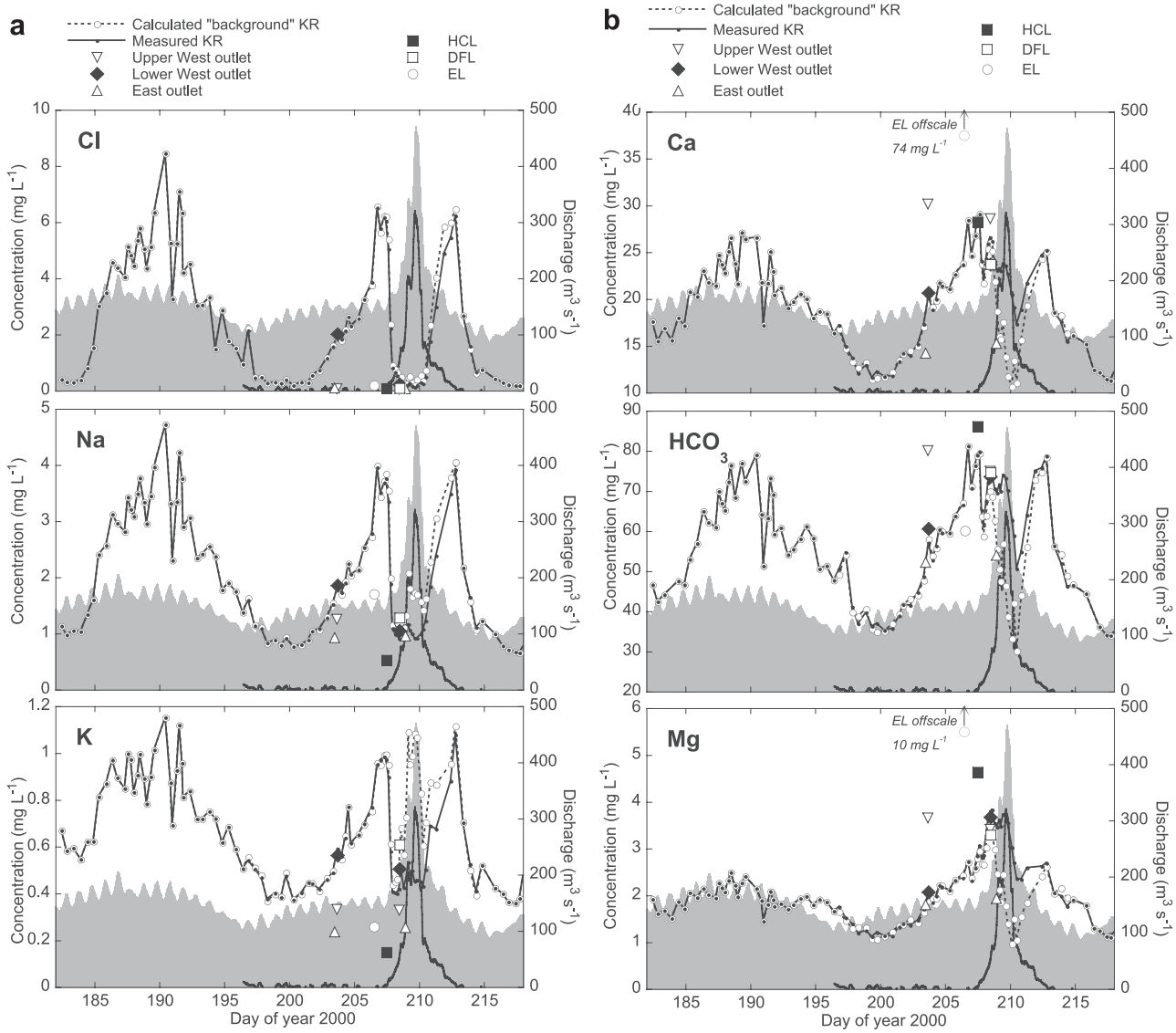

Figure 15. Water chemistry within the Kennicott Glacier basin during summer 2000. Measured solute concentrations in the Kennicott River (KR) plotted with heavy line and solid circles, and calculated "background" Kennicott River concentrations (see text) during the Hidden Creek Lake jökulhlaup shown with open circles and dotted line. Where these values coincide, data points plot as a "bull's-eye" pattern. The average solute concentrations in Hidden Creek Lake (HCL) and Erie Lake (EL) are plotted on the date they drained, while the average composition of Donoho Falls Lake (DFL) is plotted on the date that lake basin filled. Samples from the three main glacier outlets (upper west, lower west, and east) are plotted at the two times they were collected: before HCL drainage began and during its jökulhlaup. Kennicott River total discharge, in gray, forms background of each plot; calculated Kennicott River flood hydrograph (from Figure 13) is shown with thin solid line.

here to come entirely from Hidden Creek Lake), and subscript tot refers to the net outflow from the glacier, that is, the gauged values. $Q_{\mathrm{fld}}(t)$ and $Q_{\mathrm{bg}}(t)$ are the background discharge and flood discharge, respectively, as determined by hydrograph separation (Figure 13). Q_{tot} is simply the sum of Q_{bg} and Q_{fld}. Water samples from HCL show no evidence that the lake is chemically stratified, so we take $C_{i,\ \mathrm{fld}}$ as constant for every species i. The only unknown is then $C_{i,\ \mathrm{bg}}\ (t)$, so we rearrange equation (16) to find

$$C_{i,\mathrm{bg}} = \frac{C_{i,\mathrm{tot}}Q_{\mathrm{tot}} - C_{i,\mathrm{fld}}Q_{\mathrm{fld}}}{Q_{\mathrm{bg}}}. \qquad (17)$$

It would be tempting to suppose that $C_{i,\ \mathrm{bg}}$ during the flood must be the same as $C_{i,\mathrm{tot}}$ just before the flood began. Using equation (17), however, we can actually infer what the background concentration of every solute species must have been during the flood. The tempting assumption turns out to be wrong for almost every solute.

[53] Calculated chemistry of Kennicott River background flow, which is equivalent to the nonflood part of the subglacial flow, is shown in Figure 15. Solute concentrations in the Kennicott River were high when HCL began to drain, typical of the late phase of a high-discharge cycle in the river, such as that around day 190 in the preceding high

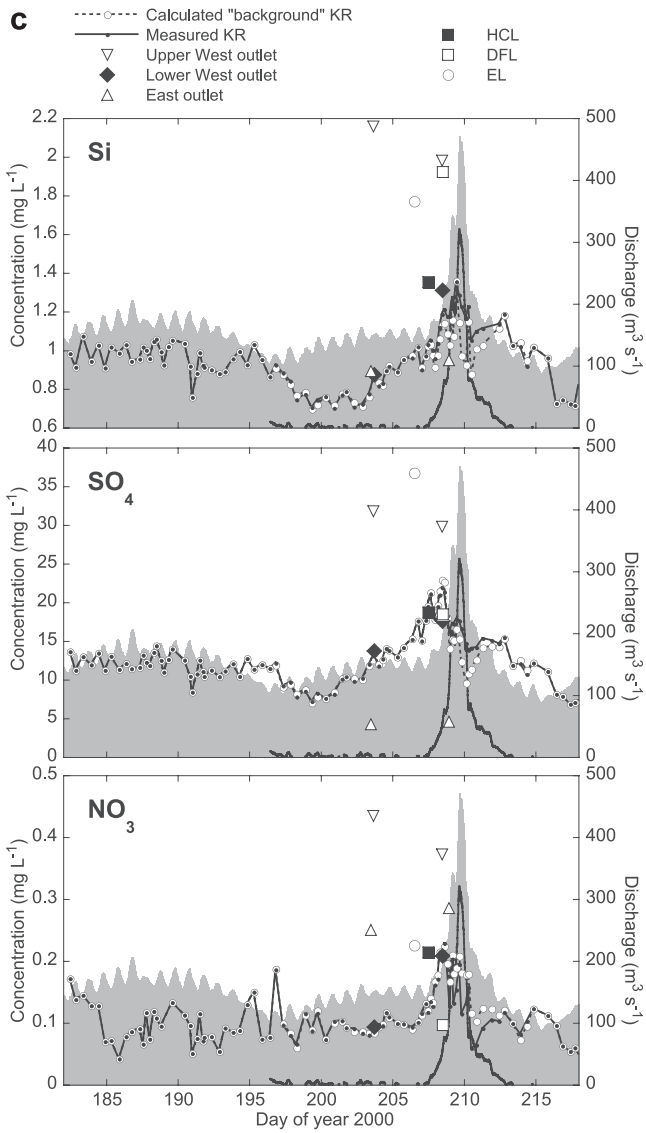

Figure 15. (continued)

discharge cycle. Calculated $C_{i,\,bg}(t)$ for most of the solutes fell during the jökulhlaup, however, rather than remaining at this high concentration. Patterns of solute concentration variations can be divided into three groups, which we will describe in turn.

[54] The first group consists of Cl, Na, and K, for which the concentrations in Hidden Creek Lake were much lower than ever observed in the Kennicott River. During the flood, concentrations of these solutes fell to such low values in the Kennicott River that mass balance (equation (16)) could only be satisfied if the background concentrations of these solutes also fell. The computed background concentrations during the flood are similar to concentrations in the river at about day 200 (Figure 15a). Note, however, that the concentration minima at about day 200 occurred shortly after a pronounced low in discharge, quite different from conditions during the flood.

[55] The next group of solutes comprises Ca, HCO_3, and Mg, which were present in high concentrations in Hidden Creek Lake water. Total concentration (C_{tot}) of each of these solutes varied little early in the jökulhlaup, but then fell conspicuously during the falling limb of the flood to values lower than found in HCL or in the river just before the flood. This pattern is amplified in the calculated C_{bg} values, all of which fell markedly during the falling limb of the jökulhlaup, to values quite close to those seen around day 200. We note again (just as for Cl, Na, and K) that these concentration minima at about day 200 occurred shortly after a pronounced low in discharge, quite the opposite of what obtained during the flood. Concentrations of Ca and Mg were even higher in Erie Lake than in Hidden Creek Lake, so drainage of Erie Lake on day 206 may be the reason that the C_{bg} values remained high until the falling limb of the Hidden Creek Lake jökulhlaup.

[56] The last group of solutes consists of Si, SO_4, and NO_3, which are present in moderately high concentrations in Hidden Creek Lake water. These solutes show patterns that are similar to the second group, but the decline in C_{bg} during the falling limb is less distinctive. Si and SO_4 were present in very high concentrations in Erie Lake, further obscuring the relationships.

[57] The picture that emerges from our calculations of C_{bg} can be summarized as follows. During the jökulhlaup, solute concentrations in the nonflood component of the subglacial flow fell dramatically: the nonflood component became much more dilute than it had been just prior to and after the jökulhlaup. As solute concentrations reflect routing of water and residence time at the bed [*Tranter et al.*, 1993, 1997; *Sharp et al.*, 1995; *Brown et al.*, 1996], the clear implication of the hydrochemistry is that the jökulhlaup radically affected the contributions of different subglacial flow paths to the glacier discharge. We will return to this point later.

4.7.3. Donoho Falls Lake

[58] Solute concentrations in the five samples of DFL water were greater than in Donoho Creek (Table 3), which prior to the jökulhlaup flowed into a subglacial tunnel in the deepest part of DFL basin. This, together with a disparity between the rapid filling rate of the basin and low discharge ($2-3$ m^3 s^{-1}) of the creek demonstrate that DFL did not fill from the creek. Most solute concentrations in DFL were identical to Kennicott River water at the time the basin filled and, in most cases, substantially lower in concentration than in Hidden Creek Lake (Figure 15). Si and K are the most notable exceptions. The Si concentration in DFL was almost twice as great as in HCL or in the Kennicott River (at any time) but was comparable to the concentration in Erie Lake and the upper west outlet.

5. Discussion

[59] Our data on Hidden Creek Lake jökulhlaups represent an unusually broad picture of the process of lake filling and drainage and routing of the floodwaters through a glacier. We now consider several aspects of the hydrology as revealed by our data.

5.1. Do Ice-Dammed Lakes Leak?

[60] Our water balance calculations for 2000 showed that recharge to the HCL basin was balanced by lake filling, as long as the subglacial wedge of water beneath the ice dam was taken into account. Failure to account for wedge storage would have led to the erroneous conclusion that

the lake was leaking at the rate of ~3 m^3 s^{-1}, on average, prior to the jökulhlaup. It is interesting to compare our conclusion with that of *Kasper* [1989], who investigated an unnamed ice-dammed lake alongside Kaskawulsh Glacier, Yukon Territory, Canada. She concluded from a water balance argument that the lake leaked, at a rate increasing from zero to ~8 m^3 s^{-1} during a 6-week period prior to the jökulhlaup that emptied the lake. The Kaskawulsh Glacier situation is strikingly similar to what we investigated at Kennicott Glacier in terms of topographic setting, lake size and recharge rate. *Kasper* [1989] surveyed targets on the ice dam near the lake and showed that large vertical motions occurred as the lake filled and drained. We have examined Kasper's survey data and estimated that water was going into wedge storage at a rate of at least 2.5 m^3 s^{-1} during the period of her observations. *Kasper*'s [1989] conclusion about leakage rate may need correction.

[61] The other outstanding claim about existence of a long-term leak prior to a jökulhlaup comes from work at Summit Lake, British Columbia, Canada, which is commonly ~10 times larger than HCL [see *Mathews and Clague*, 1993]. *Fisher* [1973] summarized results of three experiments undertaken in summer of 1968 involving Rhodamine B dye dropped into the lake near the ice margin. He measured dye in the Salmon River, which drains Summit Glacier, and estimated that Summit Lake was leaking at a rate of at least 0.2 m^3 s^{-1}. *Gilbert* [1971] had previously argued on the basis of water balance calculations that Summit Lake leaked at a rate of ~3 to 5 m^3 s^{-1} for several months prior to the jökulhlaup of November 1968 but did not consider possible storage of water beneath a deforming ice dam.

[62] Considering our data and the other investigations mentioned above, our tentative conclusion is that small, stable leaks from ice-dammed lakes are indeed possible, but such leaks may be quite small and hard to measure with much precision. Water balance calculations aimed at estimating leakage rates should certainly factor in storage beneath the ice dam. Direct measurements using dye may be the least equivocal means of assessing leaks.

[63] We have emphasized the word "stable" in discussing the concept of lake leakage, after all, an "unstable leak" would simply be the onset of the jökulhlaup itself. How, then, could a stable leak even exist? Clearly the leakage path could not have the characteristics of a Rothlisberger channel, as the leak would then be unstable, as shown by *Nye* [1976]. Presumably a stable leak would exploit drainage paths such as linked cavities [*Kamb*, 1987; *Walder*, 1986] or porous, permeable sediment [*Walder and Fowler*, 1994].

5.2. Hydrographs

[64] Our measurements permitted reconstruction of both the lake outflow hydrograph and the flood hydrograph. Differences between the two hydrographs should reflect how the flood is routed through the glacier. In this context, it is notable that the two hydrographs are, in fact, not very differently shaped. In 1999 (Figure 14), the peak of the river flood hydrograph is higher than the peak of the lake outflow hydrograph, but the durations are about the same. The difference in peak discharge values can be attributed to measurement error, but we cannot rule out the possibility that the flood hydrograph indeed had a higher peak than the outflow hydrograph. In 2000, the river flood hydrograph is practically the same as the lake outflow hydrograph aside from being translated in time. In comparison, for a flood in a subaerial channel, we expect the flood peak to decrease, and the hydrograph to broaden, with distance downstream.

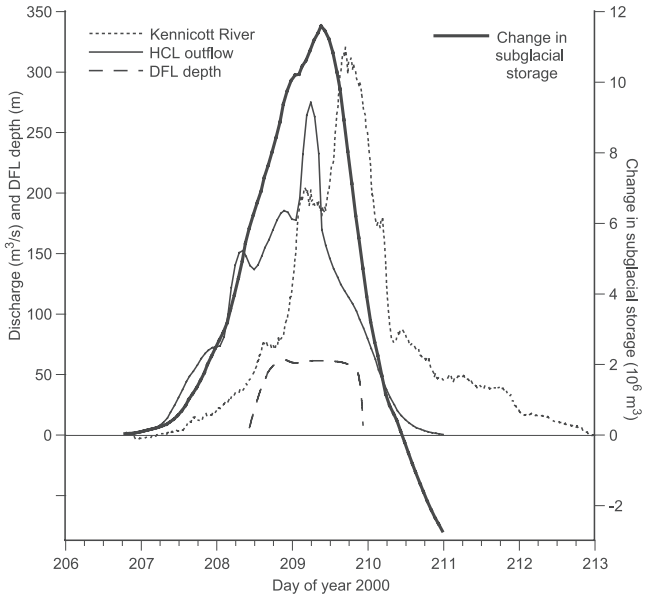

Figure 16. Calculated change in subglacial water storage associated with the 2000 HCL jökulhlaup. Hidden Creek Lake (HCL) outflow hydrograph, Kennicott River flood hydrograph, and Donoho Falls Lake (DFL) water depth shown for comparison.

[65] Within the bounds of probable error, the volume of water leaving the lake and the flood volume in the Kennicott River match in both 1999 and 2000. Although one might speculate that some HCL water could go into long-term storage beneath the glacier [see *Björnsson*, 1998], or that the jökulhlaup might trigger the release of water from subglacial storage, there is no compelling evidence to support either view. Inspection of the hydrographs in Figure 14 shows that there is a transient increase in water storage during the jökulhlaup, as flow into the glacier from the lake initially exceeds flood discharge out of the glacier. In 2000, the volume of water stored, calculated from the difference between the integrals of the two hydrographs, reached a maximum of nearly 12 × 10^6 m^3, roughly a third of the lake volume (Figure 16). Much of this stored water may have simply backed up into large voids within the glacier system. Approximately 10^6 m^3 was temporarily stored in Donoho Falls Lake, which filled during the period of maximum subglacial storage. We also saw evidence, in the form of wash lines on debris-covered ice, that water rose high in moulins and other basins during the jökulhlaup.

[66] Another striking aspect of the hydrographs for both 1999 and 2000 is that they display a tail that is nearly as drawn out as the rising limb. These hydrographs do not much resemble the canonical jökulhlaup hydrograph, with a relatively gentle rise to a peak followed by an abrupt fall, so often described in the literature [e.g., *Paterson*, 1994]. Hydrograph shapes for HCL jökulhlaups cannot be mimicked using *Clarke*'s [1982] physically based model

unless creep closure of the exit tunnel is implausibly rapid. We believe that the long hydrograph tail more likely has to do with two phenomena not considered in Clarke's model: transient storage of water, and the existence of the subglacial water wedge beneath the ice dam. Water will tend to be stored subglacially or in other voids when pressure in the drainage conduit is high; stored water will then be released as discharge and pressure fall. Furthermore, flow out of the lake will be impeded during the late stages of lake drainage, as a consequence of the entire outflow being forced through the subglacial wedge. The associated head loss means that water pressure at the entrance to the drainage tunnel will be less than what it would be otherwise.

[67] We envisage that the subglacial wedge comprises broad, irregular passages, with an average opening height that declines with distance away from the lake. As a first approximation, flow through the wedge can be treated as flow between (locally) parallel plates. It is easy to show that the Reynolds number will be so large that flow is turbulent, and we then treat the wedge as a so-called turbulent flow resistor as described by *Clarke* [1996]. Let H denote the local wedge opening and w denote the local wedge breadth. Assuming $H \ll w$, we find from *Clarke*'s [1996] equations (16), (17) and (24) that the local hydraulic head gradient in the wedge is proportional to $f\,(Q_{\text{out}})^2/gw^2H^3$, where f is a dimensionless roughness. Initially, w is about equal to the width of the glacier/lake contact (~900 m for HCL) at the beginning of drainage, but decreases as the lake shrinks. H also decreases with time, as evidenced by downdrop of the ice dam surface. Moreover, roughness is likely to increase as the ice collapses and the wedge narrows. We conclude that the head lost as water passes through the wedge will initially be negligible but will become progressively more important as the lake drains. There will also be head loss associated with the fact that the ever increasing density of grounded icebergs forces the water to follow increasingly tortuous paths. The net effect of these additional head losses will be to reduce water pressure in the outlet tunnel, thereby increasing the rate of tunnel closure by creep. Compared to a hypothetical lake identical to HCL save for the existence of the wedge, the outlet tunnel will be smaller, the discharge will be less, and the hydrograph will be broader and will have a tail. The role of the wedge in outflow hydraulics may also explain why peak flood discharge was nearly the same in 1999 and 2000, despite the lake volume in 1999 being only about half that in 2000.

[68] A more quantitative exploration of the phenomena mentioned above awaits development of a modified model that accounts for the effects of transient storage (which is likely to be universal) and the water wedge (which may not be important for all glacier-dammed lakes).

5.3. Effects of the Jökulhlaup on Hydrology of the Glacier

[69] Jökulhlaups have commonly been treated as hydrologic transients restricted to a single subglacial channel, most particularly in theoretical models of the phenomenon [*Nye*, 1976; *Spring and Hutter*, 1981; *Clarke*, 1982]. Our data show that the HCL jökulhlaups, which are actually rather modest in size, in fact broadly perturb the hydrology of Kennicott Glacier. We now synthesize our data to develop a conceptual picture of how the glacier's drainage system operates under normal conditions, and then how it is then disrupted by an HCL jökulhlaup.

[70] Both discharge and hydrochemistry in the subglacial drainage system of Kennicott Glacier (as revealed by measurements of Kennicott River) display relatively coherent, long-period (10–20 day) variations during the melt season, with solute variations lagged relative to discharge variations. All of the solutes involved in these concentration swings can be attributed to subglacial weathering processes; several are associated with subglacial abrasion. High K concentrations are commonly reported in glacial runoff, and attributed to abrasion of biotite [*Blum et al.*, 1994; *Brown*, 2002; *Drever and Hurcomb*, 1986]. Unusual, and perhaps unique to the Kennicott Glacier, are high Cl and Na concentrations. In most settings, Cl behaves as a conservative tracer of meteoric water [*Hem*, 1992]. However, Cl concentration in Kennicott River water is much greater than that measured in samples of rainwater and surface melt collected on the Kennicott Glacier (see Table 3). *Anderson et al.* [2003] proposed that high Cl concentration is due to abrasion of trace quantities of halite that plausibly may be present in sabkha (evaporite) facies of the Chitistone Limestone or saline fluid inclusions in the Nikolai Greenstone [*MacKevett*, 1978; *MacKevett et al.*, 1997]. Since halite would not persist in slowly eroding surface exposures in the climate of the Wrangell Mountains, Cl in the Kennicott River may be an extremely sensitive tracer of subglacial water, in particular from sites of active abrasion, that is, from the distributed flow system.

[71] We attribute the long-period swings in solute concentration to variations in the mean subglacial residence time (τ_{res}) of water emerging from the glacier, as follows. We suppose that the subglacial drainage system, at least in the ablation area, comprises an arborescent channel network fed by surface input, either directly (at a rate Q_f) via englacial conduits or indirectly (at a rate Q_s) via patches of linked cavities and permeable sediment that constitutes the ''distributed'' basal drainage system [*Fountain and Walder*, 1998]. As the mean subglacial residence time of a parcel of water increases, so does its solute concentration [*Tranter et al.*, 1993]; moreover, we expect that $\partial\tau_{\text{res}}/\partial(Q_s/Q_f) > 0$ [*Raymond et al.*, 1995]. During times of rapidly rising surface input, water pressure rises more in the channel network than in the distributed system; thus the hydraulic head gradient drives water out of the channel network and into the distributed system, both Q_s/Q_f and τ_{res} decrease, and solute concentrations are low in water emerging at the terminus. This describes the situation at, say, about day 184 in 2000 (Figure 15). Water that is meanwhile ''backed up'' within the distributed system becomes relatively enriched in solutes. After surface input begins to fall, water pressure drops more in the channel network than in the distributed system, the solute-rich water that had been backed up in the distributed system escapes, Q_s/Q_f and τ_{res} increase, and thus solute concentrations are high in water emerging at the terminus. This describes the situation at, say, about day 190 in 2000 (Figure 15).

[72] We now consider the hydrochemistry of the Hidden Creek Lake jökulhlaup within this context. The fact that the background flow in the Kennicott River during the jökulhlaup (Figure 15) is similar in composition to the low solute periods in the summer (for example, at about day 184 or day

200) implies that the jökulhlaup overwhelms and pressurizes subglacial conduits and diminishes flow of water out of the distributed flow system. The precipitous decline in Cl and Na concentrations in the background Kennicott River shows that this disruption to the system occurs abruptly and early during the jökulhlaup. The rapid filling of Donoho Falls Lake also implies high water pressures within subglacial conduits, as this lake basin was obviously fed by reversal of flow in the small conduit that normally accepts discharge from Donoho Creek. Although we have argued based on the outlet chemistry that the jökulhlaup was confined to a single (admittedly large) outlet at the terminus, the geochemistry of the river shows that the influence of the jökulhlaup on water exchange between high pressure conduits and the distributed system was widespread: there is no evidence of long residence time water emerging during most of the jökulhlaup.

[73] After the flood hydrograph peak, the background chemical composition of the Kennicott River began to climb from low concentrations up to concentrations typical of high or waning discharge (Figure 15). Solute concentrations in the river reached a peak early on day 213, at the very end of the flood hydrograph tail. This pulse of water with high solute concentration must reflect a release of long residence time water out of the distributed system after pressure in the conduit system subsided. In this context, it is interesting to note that in both 1999 and 2000, the lowest midsummer discharge in the Kennicott River occurred a few days after the HCL jökulhlaup (Figure 12). We suggest the following interpretation: as water pressure in the drainage tunnel plummeted during the waning phase of the jökulhlaup (an inevitable circumstance owing to the fact that the drainage tunnel would still have been greatly enlarged) flux from the distributed system into the channels would have risen rapidly. Thus water stored within the distributed system, which otherwise would have only slowly leaked into the channel system, was rapidly depleted and incorporated into the tail of the flood hydrograph. The precipitous decline in solute concentrations after the jökulhlaup suggests that the volume of high solute concentration, long residence time subglacial water is not very large.

5.4. Triggering Mechanism for Jökulhlaups

[74] The long-term trend at HCL of decreasing maximum lake stage with time [*Rickman and Rosenkrans*, 1997], as Kennicott Glacier has thinned, is consistent with the idea of jökulhlaup triggering being associated with ice dam flotation or with a critical effective pressure [*Björnsson*, 1992]. However, the data are harder to explain in this context when we consider the variability in maximum lake stage during the years 1999 to 2002 (Figure 3). Our survey measurements, all of which were referenced to a single geodetic datum, show a negligible change in ice thickness in the ice dam region from 1999 to 2000, yet the maximum lake stage in 2000 exceeded the maximum lake stage in 1999 by nearly 9 m. These observations are pretty clearly inconsistent with a simple lake level trigger hypothesis. Neither does the modified model of *Fowler* [1999] seem to provide an explanation. Fowler analyzed, with reference to jökulhlaups from Grímsvötn, the detailed hydraulics of the seal, and concluded that drainage will typically occur at a lake level less than the flotation value, but that the faster the lake fills, the closer is the final maximum stage to the flotation value. This seems to provide a reasonable explanation for why the 1996 Grímsvötn jökulhlaup, which coincided with a nearby subglacial volcanic eruption, occurred with the lake at the flotation level. At Hidden Creek Lake, however, there is no reason to suppose that filling rate varies greatly from year to year, and we need to look elsewhere for an explanation of the variability in maximum lake stage. We have no ready answer at this point, but speculate that the lake drains only when two conditions are satisfied: first, the lake level must rise enough for the seal at the bed to be breached; second, the basal drainage system beneath the main part of the glacier must then ''capture'' the outflow from the lake [see also *Whalley*, 1971]. In other words, and here we diverge from *Fowler*'s [1999] conception of the phenomenon, we do not assume that a tunnel necessarily extends to the region of the seal prior to the seal being breached. In our picture of the phenomenon, then, year-to-year variability in maximum lake stage over periods of a few years (Figure 3) reflects corresponding year-to-year variability in the configuration of the basal drainage system. We note that in our two years of river monitoring, lake drainage occurred during periods of maximum exchange of water at the glacier bed, as indicated by the chemistry of the river water. Precise timing of outbursts may be dictated by the response of the glacier hydrologic system to weather.

6. Conclusions

[75] Planned field observations during two jökulhlaups from Hidden Creek Lake, impounded by the Kennicott Glacier, Alaska, have allowed us to analyze these events in unusual detail. Water balance calculations do not support the notion that a stable ''leak'' from the lake existed prior to the jökulhlaup. Approximately 20% of the lake, by volume, was stored in a thick wedge beneath the ice dam. The outflow hydrograph from the lake is nearly symmetrical, rather dissimilar to the canonical jökulhlaup hydrograph so often mentioned in the literature. The prolonged tail of the hydrograph appears to result from lake water having to pass through the collapsing wedge underneath the ice dam before entering a conduit that routes the lake water into the main subglacial drainage system. The hydrograph is relatively little altered during its passage through the glacier.

[76] The jökulhlaup radically perturbed subglacial water routing through the entire glacier by pressurizing main drainage channels and shutting off flow into those channels from the distributed drainage system at the glacier bed. This produced an abrupt shift in the chemistry of the background flow in the Kennicott River. As the flood waned, solute-rich water that had been backed up within the distributed system was rapidly discharged into the enlarged subglacial conduit system, causing first a peak in solute concentrations in the Kennicott River followed by a pronounced seasonal low in discharge.

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R. S. Anderson and S. P. Anderson, INSTAAR, University of Colorado, Campus Box 450, Boulder, CO 80309-0450, USA. (Robert.S.Anderson@colorado.edu; Suzanne.Anderson@colorado.edu)

M. Cunico and A. G. Fountain, Department of Geology, Portland State University, Portland, OR 97207-0751, USA. (andrew@pdx.edu)

E. R. Kraal, Center for Study of Imaging and Dynamics of the Earth, Department of Earth Sciences, University of California, Santa Cruz, CA 95064-1088, USA.

D. C. Trabant, U.S. Geological Survey, 3400 Shell Street, Fairbanks, AK 99701-7245, USA. (dtrabant@usgs.gov)

J. S. Walder, U.S. Geological Survey, Cascades Volcano Observatory, 1300 Southeast Cardinal Court, Bldg. 10, Ste. 100, Vancouver, WA 98683, USA. (jswalder@usgs.gov)