Supplementary Methods

We recognise that uncertainty is introduced at each of the stages of erosion rate estimation described in this paper. Combined with the differing sediment fluxes in 2009 and 2010, this produces a range of possible values of erosion rate at Leverett Glacier. For ease of interpretation, we have amalgamated this into a single figure of 4.8 ± 2.6 mm a^{-1}. This process by which we have attained this figure is detailed below. We do this so as to ensure the transparency of our methods, and because we believe that our conclusions remain robust even when all possible sources of error are taken into account.

Sources of error

1. Discharge

Uncertainty in the estimates of river discharge is the result of error in the discharge measurements themselves and error introduced by the rating curve between the discharge measurements and water stage in the bedrock cross-section.

For the discharge record calculated through application of a rating curve, uncertainty is introduced by interpolation and extrapolation of the modeled values, unsteady flow conditions and changes in river roughness throughout the survey period. Given the lack of vegetation and the use of a bedrock cross-section, changes in roughness during the season are likely small (Richards, 1982) and this uncertainty is therefore negligible. In addition, discharge was measured across the full range of observed stage meaning that we have no need to extrapolate discharge values. Stage was measured using a Druck pressure transducer connected to a Campbell CR1000 datalogger and errors are consequently small (± 1-2 cm on a stage depth ranging up to 6 m). Therefore, uncertainties due to application of the rating curve are primarily due to a combination of interpolation errors and the effect of unsteady flow conditions.

Calculation of the different components of error induced by the application of the rating curve would require the use of a hydrological model and is beyond the realistic scope of this paper. An estimate of the error is therefore achieved by calculating the root mean square deviation (RMSD) between the observed and fitted discharge estimates. This effectively lumps the components discussed in the previous paragraph into a single term. We express this uncertainty as a normalised root mean square deviation (NRMSD) between the modeled and measured discharge. The RMSD between the discharge measurements and the rating curve is 40.1 m^3s^{-1}, and the NRMSD (expressed as a percentage) is 10.44 %.

Discharge estimates were made using the dye-dilution method (e.g. Kilpatrick and Cobb, 1985; Rantz, 1982). A known quantity of fluorescent Rhodamine WT dye was manually injected into the stream in a single pulse. Dye concentration was then measured at a downstream location using a Turner Designs CYCLOPS-7 Submersible Fluorometer attached to a Campbell CR800 datalogger and used to calculate discharge.

The main sources of error in dye dilution gauging are due to:

a. Assumption of complete mixing of dye within the channel. The tests were made in a reach with a single turbulent channel of approximately 1 km. Dye was injected above, and sampled below, a waterfall to ensure complete mixing. This was confirmed by a test where dye was injected from each side of the channel within a short time period, which produced the same dye concentrations (and 'area under the curve') at the sample site.

b. Fluorescence of suspended sediment at a similar wavelength to the dye. This background fluorescence was recorded for 10 minutes before and after each test and removed in subsequent processing. Repeat calibrations of river water with known concentrations of dye on different days (and therefore with different sediment concentrations) showed that the calibration slope was not affected by turbidity and that an offset is effective in minimising this background signal.

c. Loss of dye along the reach due to sorption onto or reaction with material in the river. These errors were minimised by use of Rhodamine WT, which is known to adsorb onto suspended sediment less than other fluorescent dyes (Smart and Laidlaw, 1977).

Other sources of error include measurement of the volume of tracer used and calibration and resolution of the fluorometer (which was repeated at intervals throughout the season using water from the river with standards of known concentration).

Although dye-dilution gauging is an established technique for measuring streamflow, there are very few estimates for its accuracy in the literature. Herschy (1995) suggests a figure of ± 5 % while information from the fluorometer manufacturer suggests that uncertainties are ± 2 % (Turner Designs).

Without an independent measure of discharge we are unable to verify the accuracy of our discharge measurements. However, based on repeatability of traces which were done within a short time of each other (less than an hour) at high rates of discharge, a conservative estimate is ±10 %.

We assume that the error due to uncertainty in the discharge measurements (E_{QM}) and from the rating curve (E_C) are independent (Di Baldassarre and Montanari, 2009) and compute the discharge error (E_Q) from the quadratic sum as follows:

$$\begin{aligned} E_Q &= \pm\sqrt{E_C^{\,2} + E_{QM}^{\,2}} \\ &= \pm\sqrt{10.44^2 + 10.00^2} \\ &= \pm 14.45\,\% \end{aligned}$$

2. Suspended sediment concentration

Turbidity was measured continuously at the gauging site (~2.2 km downstream of the portal) using a Partech IR15C turbidity meter connected to a Campbell CR1000 datalogger. A second turbidity sensor, placed at the portal between July 11 and August 16, 2010, generated the same pattern of SSC as the gauging site (r = 0.98, Supplementary Fig. 1) with values on average only ~4% lower. Erosion and deposition along this 2.2 km stretch of river is therefore considered negligible compared to the mass of sediment in transit.

Suspended sediment samples were collected manually on 80 occasions from the vicinity of the turbidity probe using a USDH-48 depth-integrating suspended sediment sampler. The samples were filtered in the field with 0.45 μm filter papers following the procedure laid out by Hubbard & Glasser (2005), and the volume of filtrate was measured in a measuring cylinder. The samples were stored and returned to a lab for drying and weighing in order to calculate suspended sediment concentration.

Given the high precision and accuracy of the balance and the routine nature of the field sampling, errors in the suspended sediment measurement (E_{SM}) are likely to be small (taken here to be less than 2%).

The greatest error (E_T) is that induced by application of a relationship between measured turbidity and suspended sediment concentration. Uncertainty is minimised by calibration of the turbidity record with field samples rather than in a laboratory. Following a similar procedure to that for the discharge rating curve, the RMSD for the measured SSC against that produced by the calibrated turbidity values is 0.93 kg m^{-3}, and the NRMSD is 5.44 %. The error in SSC (E_S) is therefore quantified as:

$$
\begin{aligned}
E_S &= \pm\sqrt{E_T{}^2 + E_{SM}{}^2} \\
&= \pm\sqrt{5.44^2 + 2.00^2} \\
&= \pm 5.80\,\%
\end{aligned}
$$

3. Catchment area

Temperature index models are a well established technique in glaciology, 'often on a catchment scale outperforming energy balance models' (Hock, 2003). We have a comprehensive data set from which to calculate degree day factors and run the model. Temperature data is available at 15 minute intervals from 7 stations in a transect commencing ~ 2 km from the snout of Leverett Glacier and reaching to beyond the extent of the catchment area (Fig. 1). Ultrasonic distance gauges (UDGs) at stations 1, 3, 5 and 6 record surface lowering at 15 minute intervals, and spring and autumn ablation stake measurements are available from which to constrain total melting at all sites. As such, we feel errors introduced by the model should be small.

The storage of meltwater in supraglacial lakes and crevasses adds a lag between melt and runoff, generating errors in the modeled catchment extent (Supplementary Fig. 2). This is particularly noticeable at times when runoff exceeds total melt due to lake drainage events, or the continued release of meltwater following a sudden cooling. These occasions are easily recognised and ignored, but lesser errors may be introduced when the discrepancy is less striking, which could account for the smaller fluctuations in modeled catchment area following its upglacier expansion in the spring. Coupled with uncertainty in the exact location of moulins, this makes defining a precise upper boundary for the catchment unrealistic.

The greatest source of uncertainty in the catchment area is the lack of high resolution bed data. Bed topography has the potential to change both the distribution of the catchment, and by changing the hypsometry, the surface area of this catchment required to provide the observed runoff. The general distribution of the catchment area is well supported by ice velocity patterns (Palmer et al., 2011), but it is difficult to test its precise form.

In light of these uncertainties, we recognise that defining the catchment area is necessarily approximate, and we do not claim to procure a precise figure. The value used (600 km^2) represents our best estimate from the available evidence. We are confident however that our findings are relatively insensitive to catchment area, and the uncertainty in this value is not sufficient to affect our conclusions. To demonstrate this, we have applied what we feel is a generous estimate of error at ± 25 % (E_A) to the catchment area.

4. Incomplete record in 2010

The extrapolation of the discharge and SSC record following the cessation of monitoring in 2010 indicates that total suspended sediment flux in this year may be ~ 10 % greater than observed (see *Results*). To account for this, suspended sediment flux is increased by 10 %. Because 10 % is only an estimate of the additional flux, we add an uncertainty of ± 9 % (E_R) to the new total flux. This is equal to placing the additional flux at between 0-20 % of the observed flux, or the new total flux at 110 ± 10 % of the observed flux.

5. Bed load

Conventional models for estimating the bed load component of total load prove unreliable in proglacial settings (Bogen and Bonsnes, 2003), and so we estimate bed load based on comparison with previous studies of glacierized catchments. As noted in the *Discussion*, bed load has been observed to constitute 30 – 60 % of the total load of proglacial rivers (Gurnell, 1987). We have opted to assume a bed load component of 50 %, based primarily on the record presented by Bogen and Bonsnes (2003) for Nigardsbreen, Norway. This particular study was chosen for two reasons. Firstly, it is the longest record available, spanning 32 years, which serves to reduce interannual variation (on any individual year, bed load is equal to anything between half and double the suspended load in the Nigardsbreen proglacial river). Secondly, the bedrock at Nigardsbreen is of broadly comparable strength to that at Leverett Glacier. This is relevant, as more resistant bedrock may lead to a greater proportion of material being transported as bed load (Bogen and Bonsnes, 2003). To estimate the total sediment flux, we therefore multiply the suspended sediment flux by 2, making bed load equal to 50 % of total load. These new values are given an uncertainty of ± 25 % (E_B). Assuming that this uncertainty lies entirely within the bed load component, this range is equivalent to bed load constituting between ~ 33 – 60 % of the total load, or 100 ± 50 % of suspended load.

Total error

We assume errors introduced by uncertainty in discharge, suspended sediment concentration, sediment source area, missing data and bed load to be independent. Total error for 2009 (E_{09}) and 2010 (E_{10}) is therefore calculated as follows:

$$\begin{aligned} E_{09} &= \pm\sqrt{E_Q{}^2 + E_S{}^2 + E_A{}^2 + E_B{}^2} \\ &= \pm\sqrt{14.45^2 + 5.80^2 + 25.00^2 + 25.00^2} \\ &= \pm 38.63\ \% \end{aligned}$$

$$\begin{aligned} E_{10} &= \pm\sqrt{E_Q{}^2 + E_S{}^2 + E_A{}^2 + E_R{}^2 + E_B{}^2} \\ &= \pm\sqrt{14.45^2 + 5.80^2 + 25.00^2 + 9.00^2 + 25.00^2} \\ &= \pm 39.66\ \% \end{aligned}$$

Erosion rate

In 2009, the observed suspended sediment flux is 4.50 x 10^6 t a^{-1}. When estimated bed load is included, this becomes 9.00 x 10^6 t a^{-1}. Incorporating uncertainty at ± 38.63 %, this produces an erosion rate of 5.36 ± 2.07 mm a^{-1}.

In 2010, the observed suspended sediment flux is 2.75 x 10^6 t a^{-1}. This is first increased by 10 % to account for the missing record at the end of the season, then increased further to include bed load. Total sediment flux therefore becomes 6.05 x 10^6 t a^{-1}. Incorporating error at ± 39.66 %, this equates to an erosion rate of 3.60 ± 1.43 mm a^{-1}.

Taking both years together, the lowest value facilitated by this range is 2.17 mm a^{-1}, while the highest is 7.43 mm a^{-1}. If the midpoint of these is taken, the total range of erosion rates permitted by our calculations can be expressed as 4.8 ± 2.6 mm a^{-1}. By expressing the value in this manner, we hope to facilitate relatively easy comparison with studies of erosion rate in glacial and non-glacial catchments around the world.

Supplementary References

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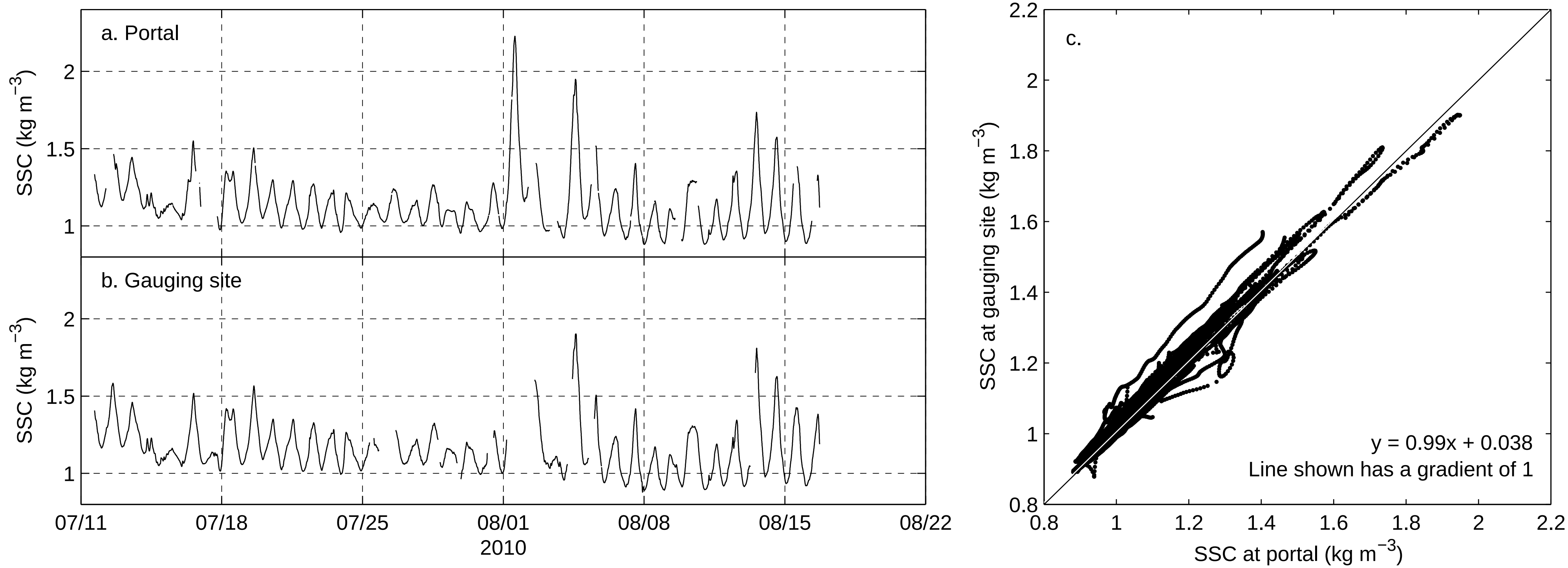
Supplementary Figure Captions

Supplementary Figure DR1. Concurrent records of suspended sediment concentration from the portal (a) and gauging site (b). c. Scatter plot of suspended sediment concentration at the portal and gauging site at five minute intervals. Gauging site values are lagged by 15 minutes.

Supplementary Figure DR2. Maximum elevation within the Leverett catchment from which melt water is required to produce the observed runoff, based on a temperature index model of melt. The

1560 m limit on the y-axis represents the uppermost extent of this catchment. Spikes which exceed this limit indicate periods in which there is insufficient melt within the catchment area to account for runoff, most likely due to lake drainage or continued release of meltwater following a sudden cooling.

Tom Cowton, Supplementary Figure 1, .pdf



Tom Cowton Supplementary Fig. 2, .pdf