ENTRAINMENT AND TRANSPORT OF SUBGLACIAL SOILS AND ROCK IN WESTERN GREENLAND

A Progress Report Presented

by

Joseph A. Graly

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The Following members of the Thesis Committee have read and approved this document before it was circulated to the faculty:

______________________________ Advisor
Paul Bierman

______________________________ Chair
George Pinder

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Tom Neumann

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Andrea Lini

Date Accepted:______________________

Introduction

The work presented in this progress report differs substantially from the scope of work laid out in my May, 2008 thesis proposal. There, I proposed to use the three-dimensional thermomechanical ice sheet model Glimmer [Rutt et al. 2009] to interpret cosmogenic isotopes in glacial detrital clasts from Western Greenland in terms of ice sheet mechanics and subglacial erosion rates. The implementation of this proposed work became difficult for two reasons. Delays in the reconstruction of Paul Bierman's laboratory deprived me of a dataset with which to control model parameters. And, Thomas Neumann's departure from the UVM faculty left me primarily working with Paul Bierman, who lacks expertise in ice dynamics. Furthermore, we have found interesting and unexpected meteoric ^{10}Beryllium values in sediment sampled from the western Greenland Ice Sheet. The analysis and interpretation of these data is now the primary focus of my research. The modelling component of the project has not been abandoned, but simpler two and one dimensional models are being employed.

In the summer of 2008, I traveled to Greenland with Paul Bierman, Tom Neumann, and Lee Corbett. We collected ice-marginal samples at three locations in western Greenland: Kangerlussuaq (67.1º N), Ilulissat (69.4 º N), and Upernavik (72.5º N) [Figure 1]. Our primary purpose was to sample glacial detrital clasts. But ice and ice-bound sediment were also collected. Cosmogenic ^{10}Be was measured in meteoric particles adhered to the ice-bound sediment grains. δ^{18}O and δ^{2}H were measured in the corresponding ice.

Background

Due to its size and sensitivity to changing climate conditions, the Greenland Ice Sheet has likely had a major role in sea level fluctuations over previous glacial /interglacial cycles [Huybrechts 2002]. Modelled arctic temperatures from the Eemian interglacial (~130-116 ka BP) closely resemble those forecasted for c. 2100 in global warming models [Overpeck et al. 2006]. A detailed understanding of the Greenland Ice Sheet's behavior during the Eemian would therefore aid in the prediction of future sea level change.

While the Eemian is the most recent interglacial period with warmer temperatures than present, the marine oxygen isotope record indicates six periods during the past two million years when the global temperature was at or exceeded present levels [Lisiecki and Raymo 2005]. Thermomechanical ice sheet models show substantial Eemian Greenland ice sheet retreat, with contribution to global sea level ranging from 3 to 5.5 meters [Cuffey and Marshall 2000; Huybrechts 2002; Lhomme et al. 2005; Otto-Bliesner et al. 2006; Tarasov and Peltier 2003]. These models show a near total melting of Greenland's southern dome and a more modest reduction in the main dome's volume. Earlier interglacial periods may have seen even larger reductions in ice cover. Cosmogenic isotopes measured in a rock core beneath the GISP2 ice core shows burial ages of 0.5 ± 0.2 million years [Nishiizumi et al. 1996]. As the GISP2 core was taken at Summit Station on the ice divide in central Greenland, this could indicate a near-total melting of the Greenland Ice Sheet consistent with OIS-11, approximately 400 ka.

During these warm interglacial periods, the ice sheet's retreat exposed the present ice-covered surface of Greenland to the flux of galactic cosmic rays. These galactic cosmic rays split molecules in the upper atmosphere creating ^{10}Be and other cosmogenic isotopes that fall out in precipitation or are delivered to the land surface as dry deposition [Lal and Peters 1967]. This meteoric ^{10}Be accumulates in soils, where it adheres to sediment particles. [Pavich et al. 1984]. The galactic cosmic rays also interact with surface materials, causing cosmogenic isotopes to form *in situ* [Lal 1988]. The concentration of both meteoric ^{10}Be in soils and *in situ* ^{10}Be in rock both relate to the duration of surface exposure.

During subsequent glacial periods, both rock and soil will retain cosmogenic isotopes. ^{10}Be is particularly long-lived, with a half-life of 1.36 my [Nishiizumi et al. 2007]. Rock clasts can be entrained by subglacial quarrying of bedrock surfaces [i.e. Hallet 1996] or by regelation of subglacial sediment such as tills [i.e. Iverson 1993]. Soils are primarily entrained through regelation processes. When this detrital material reaches the ice margin, the record of its past exposure to galactic cosmic rays or to ^{10}Be enriched rainfall and dryfall is recorded in its concentration of cosmogenic isotopes.

Methods

273 detrital clasts were collected, 100 from Upernavik and Kangerlussuaq, and 73 from Ilulissat. We sought to sample three sorts of clasts: clasts bound in ice within the basal layer of the ice sheet, clasts resting on top of an inclined basal slope that likely melted out in place as the ice ablated, and clasts situated at the mouths of subglacial outwash streams. We likewise sampled a diverse range of angularities and rock types [Table 1]. We were not able to find an outwash stream with cobble-sized sediment in Ilulissat.

Rock samples were crushed and quartz isolated through a series of HCl and HF etches. Several samples had to be rejected due to insufficient quartz content. Beryllium is being isolated from this purified quartz and the concentration of ^{10}Be measured by accelerator mass spectrometry. Thus far, six samples have been so analyzed and several dozen more will be measured in early December. By the end of February, ^{10}Be will have been extracted from about 100 clasts.

In addition to rock samples, 41 ice and sediment samples were collected, 12 from Kangerlussuaq, 8 from Ilulissat, and 21 from Upernavik. Typically, 3-5 samples were taken at each field location, approximately corresponding to the locations were specific detrital clasts were sampled. At one Upernavik field site, 15 samples were collected along a transect up the ice face, allowing for more robust interpretations of paleo ice dynamics at this site.

The ice samples were allowed to melt; with the assistance of Andrea Lini, I measured $\delta^{18}O$ through equilibration with CO_2 gas [Socki et al. 1992]; δ^2H was measured through H_2 extraction by elemental zinc [Coleman et al. 1982]. The sediment from the ice was filtered out, dried, pulverized, and its ^{10}Be extracted via the KHF_2 flux method [Stone 1998]. Concentrations of ^{10}Be have been measured by accelerator mass spectrometry in 13 of the 41 samples. 21 additional samples will be measured in early December.

Results and Discussion

Regelation signal in stable isotope data

$\delta^{18}O$ values in Central Greenland ice cores generally vary between -30‰ and -40‰ [Dansgaard et al. 1982; Stuiver and Grootes 2000]. These values can be distinctly divided into Holocene, and Pleistocene cold and mild phases. Our measured $\delta^{18}O$ values also fall within this range, but are generally below Holocene values from the southern Dye 3 site and above Pleistocene cold phase values found in the Summit cores. Deuterium excess is calculated by subtracting 8 times the $\delta^{18}O$ value from the δD value. The excess values in our samples vary from around 10‰, to below 0‰. Several of our excess values are well below those recorded in central Greenland ice cores [Johnsen et al. 1989; Jouzel et al. 2007] [Figure 2].

The process of regelation, by which basal ice melts and refreezes around particles or obstructions in the bed, can enrich $\delta^{18}O$ and δD values when it occurs in an open-system where some of the melted water escapes into subglacial flow [Iverson and Souchez 1996]. Because this enrichment effect is stronger in δD than in $\delta^{18}O$, open-system regelation lowers deuterium excess values in the regelated ice. At locations where open-system regelation has occurred, we expect a consistent trend with increasing $\delta^{18}O$ and decreasing deuterium excess [Sugden et al. 1987]. The deuterium excess values that plot below ice core value ranges are therefore explained by regelation enrichment.

Kangerlussuaq samples have $\delta^{18}O$ values within the range of the Dye 3 ice core Pleistocene mild phase, though many samples have deuterium excess values indicative of regelation enrichment. Two samples (not shown in figure 2) plot along an enrichment trend from Dye 3 ice core Holocene values.

Ilulissat samples are divided between the two sampling locations, one along the lateral margin of the Sermeq Avannarleq Glacier and one at the terminus of a small outlet branch thereof. The lateral location data show a regelation enrichment trend that may relate to the Summit ice core Pleistocene mild phase.

Upernavik samples are divided between the transect location and two northern locations. The transect location data show a trend originating from Summit ice core Holocene values. This could also be interpreted as the Pleistocene mild phase of a more coastal location. The data from the northern locations are more variable. They include some of the lowest $\delta^{18}O$ and excess values in the dataset and may indicate regelation enrichment from Pleistocene cold phases.

Meteoric ^{10}Be, regelation and erosion rates

The 13 measured meteoric ^{10}Be values vary from $3.2*10^6$ to $2.1*10^8$ atoms/gram. While all three sites have similar lowest measured values, the highest measured value varies significantly by latitude, the higher latitude sites having progressively greater maximum values. It is unlikely that the northern latitude sites experienced longer exposure during past interglacial periods, so this variation is likely explained by differences in erosion rates.

In comparing meteoric ^{10}Be concentration and deuterium excess, we find that the highest ^{10}Be values are found in samples where deuterium excess is also high and open-system regelation is therefore unlikely [Figure 3]. Significant loss of deuterium excess through open-system regelation is unlikely where deuterium excess exceeds 8.25‰. These values are at or above the deuterium excess values found in all periods of the Greenland ice core records. Lowering of deuterium excess due to open-system regelation is likely below values of 3.25‰, as ice core deuterium excess values are seldom this low. Between values of 3.25‰ and 8.25‰, open-system regelation may have occurred depending on whether the ice was deposited during a Pleistocene mild phase.

Open-system regelation is most likely to occur where there is abundant subglacial water. As subglacial streams are one of the most effective agents of subglacial erosion [Alley et al. 1997], sediment entrained by open-system regelation is more likely to be sourced from an area where erosion rates are high. As subglacial water is often derived from surface melt [Das et al. 2008], lower latitudes may have higher rates of erosion.

Sediment entrained through closed-system regelation, in which all the melted water refreezes, is more likely to be sourced from the interior of the ice sheet, where subglacial water is scarce. The time necessary to transport the sediment from the interior to the margin also reduces the erosional signal in such sediment, the sediment retaining the ^{10}Be signal of the top layer of the soil prior to transit. If the sediment is sourced far enough from the margin, the transport times may equal most of the current glacial cycle [Figure 4].

In order to estimate these potential transport times, I use model ice velocities provided by Wei Li Wang of the Goddard Space Science Institute at NASA. These data are derived from geophysical data for the present Greenland Ice Sheet and physical equations that describe ice behavior [Wang et al. 2002]. Ice velocity is interpolated from horizontal data spaced along 1 km nodes and 100 vertical grid points. For each horizontal node, a total sediment transport time is calculated for sediment originating from that node. This method assumes that the basal layer had similar horizontal velocities during the Pleistocene and ignores the possibility that sediment might be deposited and later re-entrained by changing ice dynamics. Generally, sediment sourced within the first 100 km will arrive at the margin within 10,000 years. Possibly, much, if not all, of our sediment is so sourced. But these results nevertheless indicate the possibility that sediment was in transit for most of the last glacial cycle.

Soil ages and inheritance

In order to evaluate the potential ages of the soils from which our sediments were sourced, I compiled a global database of meteoric ^{10}Be soil profiles; a total of 86 profiles from 24 papers. ^{10}Be is mobile in the soil profile, and the maximum ^{10}Be concentration is often not in the top layers [Pavich et al. 1986]. I conservatively assume that the highest measured ^{10}Be value at each of our sites represents the maximum ^{10}Be concentration in source soils for that site's sediments. I compared peak ^{10}Be concentration to the total ^{10}Be inventory in 48 appropriate soil profiles from the global database [Barg et al. 1997; Bouchard and Pavich 1989; Brown et al. 1988; Harden et al. 2002; Maejima et al. 2004; McKean et al. 1993; Monaghan et al. 1983; Monaghan et al. 1992; Pavich et al. 1986; Pavich et al. 1984; Pavich et al. 1985; Pavich and Vidic 1993; Reusser and Bierman In Press; Shen et al. 2004; Stanford et al. 2000; Tsai et al. 2008]. I then use this correlation to estimate the total ^{10}Be inventory in the source soils of each of our 3 sites [Figure 5].

To convert a ^{10}Be inventory into a soil age, a ^{10}Be deposition rate must be known. ^{10}Be fallout in the interior of Greenland is well-constrained by measurements of ^{10}Be in ice cores [Finkel and Nishiizumi 1997]. To estimate the deposition rate in the coastal areas, I applied the deposition per unit of precipitation in the ice cores to the local annual rainfall at each of our field sites [Table 2]. The values generated by this method approximate ^{10}Be fluxes modelled through general circulation models for the coastal areas of Western Greenland [Field et al. 2006].

The implied soil age of 50-100 ka at the northernmost site may have been acquired over multiple glacial cycles. A recycled ^{10}Be component is found in North American dusts that date from the Pleistocene-Holocene transition [Harden et al. 2002], and ^{10}Be concentrations presently in ice-bound sediment would be incorporated as an inherited signal in soil forming in newly-exposed terrain. Likewise, the less than 2 ka age in the southernmost site presumably represents a deeply eroded soil. If soils were consistently so deeply eroded over several glacial cycles, the initial inherited signal in southern Eemian soils would be small to non-existent.

Alternatively, the northern sediments may be sourced from erosion-resistant Tertiary regolith. Cratonic regolith can be tens of meters deep. However, in extant ^{10}Be saprolite profiles, concentrations sufficient to produce the highest Upernavik values are only found within the top 5 meters of the soil [Brown et al. 1988; Pavich et al. 1985; Stanford et al. 2000].

Timeline of Future Work

Twenty-one meteoric ^{10}Be analyses are forthcoming. In total, we expect 10 measurements from Kangerlussuaq, 8 from Ilulissat, and 16 from Upernavik (11 of which are from the transect field location). The other 7 samples contain too little sediment to process. The meteoric ^{10}Be samples will be analyzed at Livermore National Laboratory during the first week of December. These additional data will allow for more robust interpretations and further test our hypothesis that an open-system regelation signal corresponds to a higher erosion rate and therefore lower meteoric ^{10}Be concentrations. The initial latitudinal signal will likewise be tested.

I am presently working on review papers relating to the meteoric ^{10}Be system. In one, I will present my global completion of meteoric ^{10}Be soil profiles. This paper will examine the relationships between profile shape, clay content, pH, and other geochemical factors in the extant ^{10}Be soil literature. In another, I will both present new data and review and synthesize ^{10}Be deposition data, including both measurements in rainfall and long-term records in ice and sediment. I will compare these deposition rates to deposition rates estimated from ^{10}Be inventories in soil profiles. The method of assessing long term deposition rates from soil profiles was developed by Luke Reusser and Paul Bierman in their work in New Zealand. A soil profile from Waipaoa in New Zealand will be analyzed, as will selected published ^{10}Be soil profiles. Within the next two months, I intend to submit these papers to Geochemica et Cosmochemica Acta and Earth and Planetary Science Letters, respectively.

These two papers are critical for robust interpretation of the Greenland data. The global soil profile comparison provides the relationship between peak ^{10}Be concentration and total ^{10}Be inventory discussed above. A well-constrained ^{10}Be deposition rate is also essential in estimating the sediment-source soil ages. Once we have the complete meteoric ^{10}Be dataset for Greenland and these papers are submitted, I will begin work on a third paper that analyzes the Greenland meteoric ^{10}Be and stable isotope data. I hope to submit the Greenland paper in January.

To further analyze the entrainment of relict soils in the Greenland Ice Sheet, we are expecting to receive ice samples from the sediment-rich basal layer of the GISP2 core, drilled at Summit, Greenland. This winter we will analyze the meteoric ^{10}Be and stable isotopes in this basal ice as well. This will provide broader context for the marginal samples.

In late winter / early spring we expect to have at least 100 *in situ* ^{10}Be measurements from the detrital clasts. The analysis of these clasts will primarily be the subject of Lee Corbett's thesis. However, I plan to create two-dimensional models of the retreat of the Greenland Ice Sheet during past interglacial periods, the isostatic response of Greenland, and the subglacial entrainment of detrital clasts during the subsequent glacial period advance. These models will allow us to estimate the extent of interglacial retreat and subsequent glacial erosion, based on the observed cosmogenic isotope data.

In late spring, I intend to defend my masters thesis. I have funding to remain working over the summer. We will likely then be working on academic papers emerging from the *in situ* cosmogenic isotope data.

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Table 1: Rock type and angularity sampled by location.

Kangerlussuaq	Banded Gneiss	Granitoid	Foliated Granitoid	Meta Sedimentary	Other	Total
	50	27	17	3	3	100
Ilulissat	Banded Gneiss	Granitoid	Foliated Granitoid	Meta Sedimentary	Other	Total
	26	15	29	1	2	73
	36%	21%	40%	1%	3%	100%
Upernavik	Banded Gneiss	Granitoid	Foliated Granitoid	Meta Sedimentary	Other	Total
	7	61	4	19	9	100

Kangerlussuaq	Angular	Sub-Angular	Sub-Rounded	Rounded	Total
	20	51	19	10	100
Ilulissat	Angular	Sub-Angular	Sub-Rounded	Rounded	Total
	37	29	7	0	73
	51%	40%	10%	0%	100%
Upernavik	Angular	Sub-Angular	Sub-Rounded	Rounded	Total
	44	37	18	1	100

Table 2: The table below provides the highest measured ^{10}Be value from each of our 3 field sites. Using the correlation in Figure 5, we derived the total inventories used below to calculate soil ages. The use of the correlation for the Kangerlussuaq site is questionable, as the highest measured Kangerlussuaq value is well below the lowest peak values of the 48 profiles. Soil ages are calculated using a deposition rate based on local precipitation and an alternative deposition rate based on data from central Greenland ice cores that is treated as a proxy for interior conditions.

Location	Highest measured 10Be Value (atoms/gram)	Correlation -Derived Inventory (atoms)	Local Deposition Rate (atms/cm2/yr)	Soil Age (ka)	Inland Deposition Rate (atms/cm2/yr)	Soil Age (ka)
Upernavik (n=6)	$2.1*10^{8}$	$3.25*10^{10}$	$6.17*10^{5}$	53.4	$3.50*10^{5}$	95.0
Ilulissat (n=4)	$6.4*10^{7}$	$7.65*10^{9}$	$6.43*10^{5}$	11.9	$3.50*10^{5}$	22.0
Kangerlussuaq (n=3)	$6.7*10^{6}$	$4.92*10^{8}$	$3.50*10^{5}$	1.4	$3.50*10^{5}$	1.4

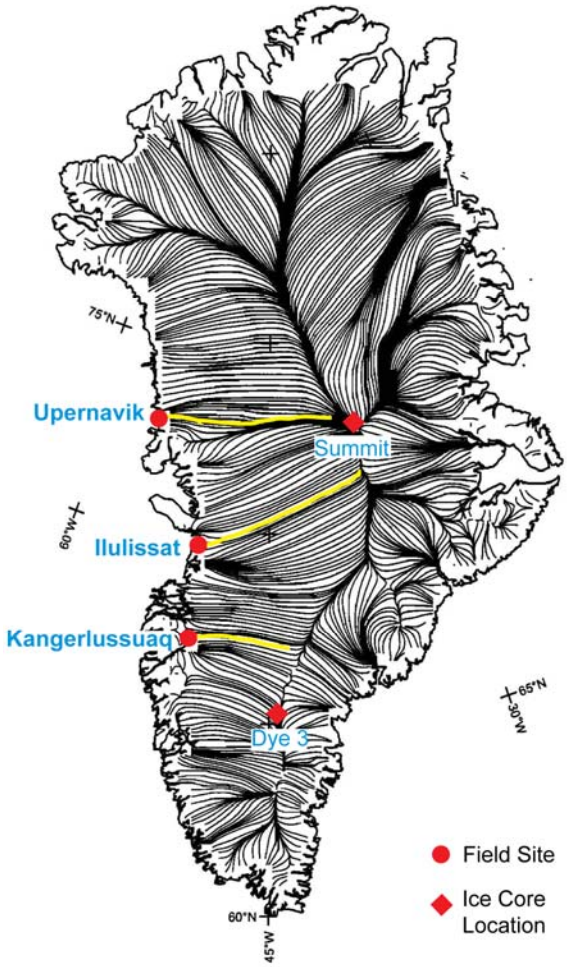

Figure 1: Modelled glacial flowlines of Greenland (modified from Zwally and Giovinetto, [2001]) showing the locations of our three field sites and the flowlines along which sediment may be sourced to these sites. The locations of Dye 3 and Summit Station are also shown.

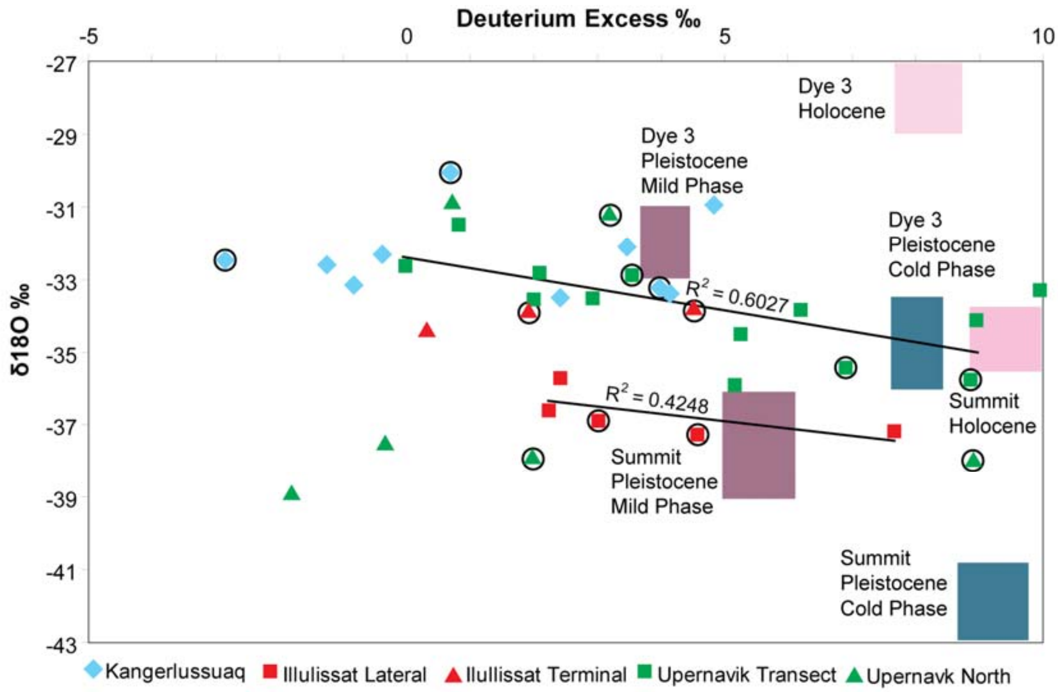

Figure 2: Deuterium excess versus $\delta^{18}O$ in ice water samples. The general value ranges for the Summit and Dye 3 ice cores are also shown. Circles indicate meteoric ^{10}Be was measured in the sample's sediment.

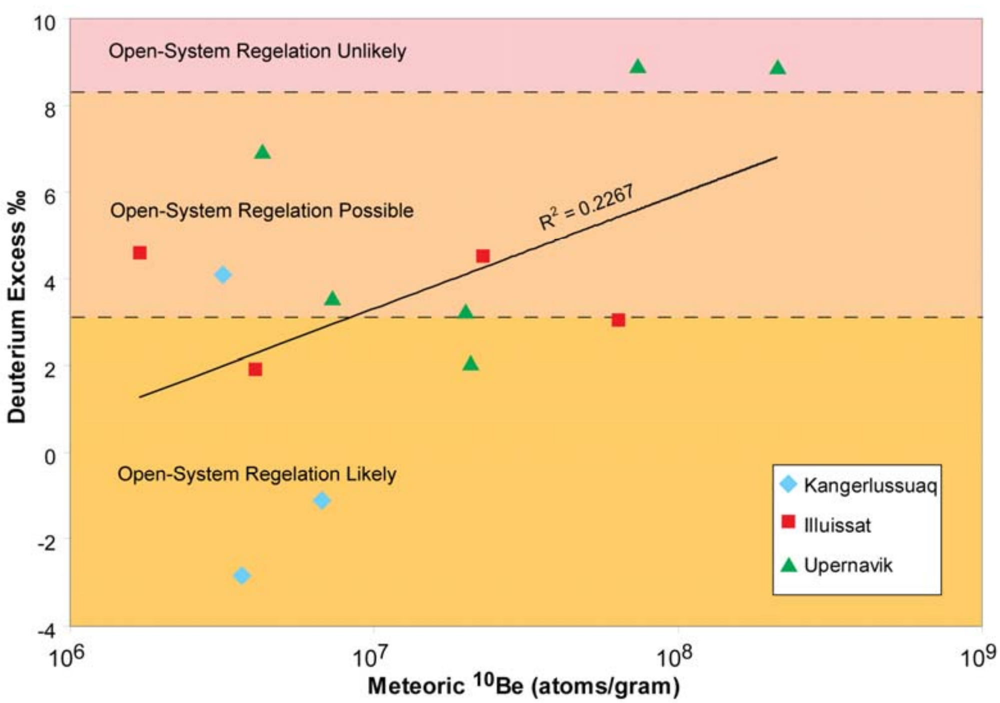

Figure 3: Relation between meteoric ^{10}Be in sediment and deuterium excess ice in samples where both were measured. The highest meteoric ^{10}Be concentrations come from sediment bound in ice where open-system regelation is unlikely to have occurred. Where open-system regelation is possible or likely, a range of values is found.

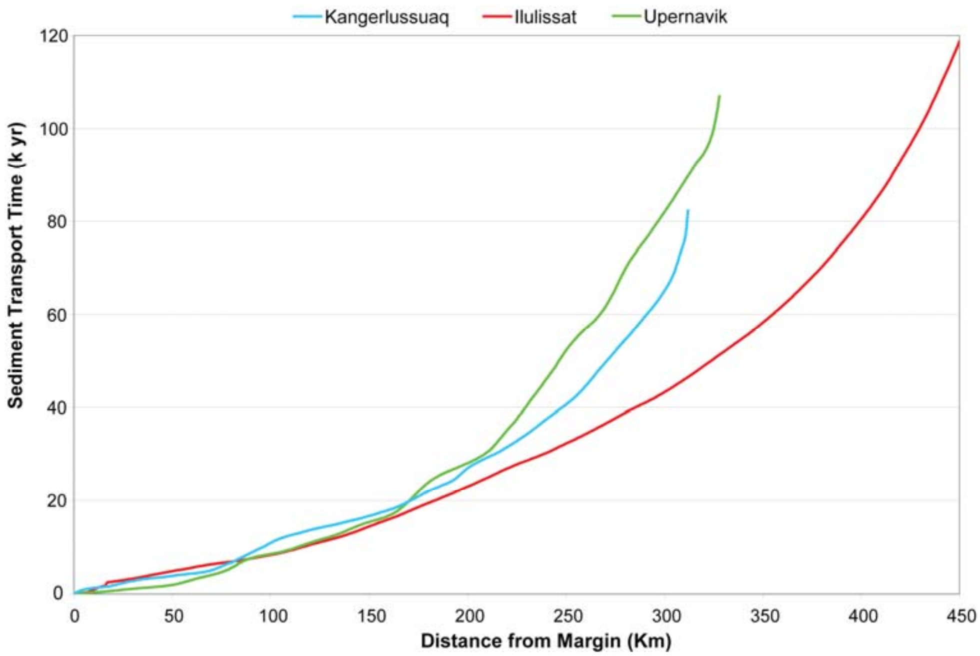

Figure 4: Modelled transport times for sediment entrained in the basal layer of the Greenland Ice Sheet along flowlines providing sediment to our three sampling sites. For Kangerlussuaq and Upernavik, data from the entire flowline length is shown. Nodes closest to the ice divide are excluded, as the near-zero velocities modelled there are unlikely to be consistent over time. The Ilulissat flowline is 550 km long. The last 100 km has modelled transport times increasing exponentially to over 300,000 years.

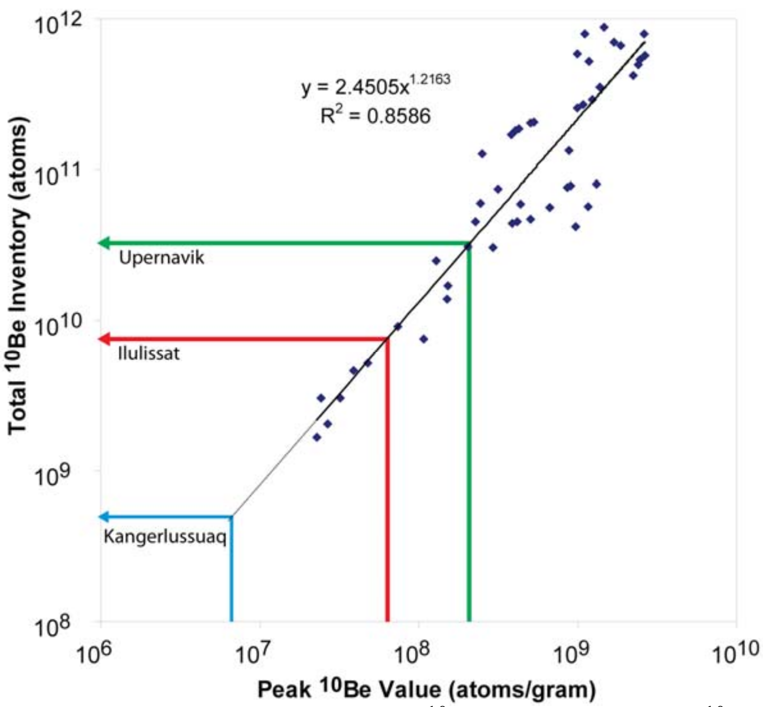

Figure 5: Relation between peak ^{10}Be values and total ^{10}Be inventories in 48 soil profiles from diverse global sources. This correlation is used to estimate a total ^{10}Be inventory at our three field sites.