**Rapid last-deglacial thinning and retreat of the marine-terminating southwestern
Greenland ice sheet**

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Abstract

Marine-terminating outlet glaciers are a major source of modern ice loss from the
Greenland Ice Sheet (GrIS), but their role in GrIS retreat during the last deglaciation is
not well constrained. Here, we develop deglacial outlet glacier retreat chronologies for

four regions in southwest and south Greenland to improve understanding of spatial variations in centennial- to millennial-scale ice loss under a warming climate. We calculate ^{10}Be surface exposure ages of boulders located in fjords near the towns of Qaqortoq, Paamiut, Nuuk, and Sisimiut. Our northernmost study site, Sisimiut, deglaciated earliest at ~ 18 ka to ~ 15 ka with an average thinning rate of $0.1\text{--}0.3\text{ m yr}^{-1}$. Inland retreat from Sisimiut to the modern ice margin took ~ 7 ka at an average retreat rate of $15\text{--}20\text{ m yr}^{-1}$. A ^{10}Be -dated moraine ~ 25 km from the modern GrIS margin deposited at ~ 8.0 ka suggests a possible ice-margin still-stand, but this does not change overall retreat rates. After retreat from the small coastal Sisimiut fjords, the GrIS margin was mainly land-terminating in this region. In contrast, earliest exposure occurred at ~ 12 ka near Qaqortoq, and $11\text{--}10$ ka near Nuuk and Paamiut, with ice thinning at rates of $0.2\text{--}0.3\text{ m yr}^{-1}$ to instantaneous within measurement uncertainty. Ice retreat inland through the extensive Nuuk, Paamiut, and Qaqortoq fjord systems to near modern ice margins occurred in <1 ka, resulting in minimum retreat rates of $25\text{--}60\text{ m yr}^{-1}$ and maximum retreat rates of $\sim 60\text{ m yr}^{-1}$ to instantaneous within the uncertainty of our measurements. This rapid thinning and retreat of marine-terminating southwest GrIS margins is contemporaneous with an incursion of relatively warm ocean waters into the Labrador Sea and toward the southwest Greenland coast, suggesting that a warming ocean may have contributed to the more rapid retreat of marine GrIS termini in the Nuuk, Paamiut, and Qaqortoq fjord systems relative to the slower ice retreat inland from Sisimiut. Our results highlight past outlet glacier-ocean interaction as a potentially important driver in rapid GrIS retreat.

1. Introduction

Modern marine-terminating margins of the Greenland Ice Sheet (GrIS) are thinning, accelerating, and retreating more quickly than land-based margins (e.g., Holland et al., 2008; Rignot et al., 2010; Joughin et al., 2012; Moon et al., 2012). As ocean waters melt the base of a marine-terminating glacier, basal shear decreases due to reduced ice load and increased basal lubrication (Joughin et al., 2012). This leads to acceleration, enhanced crevassing and calving, further thinning, and ice margin retreat (Joughin et al., 2012). How do the present rates of thinning and retreat compare to rates during previous warming episodes? Determining the magnitude of past thinning and retreat rates for both terrestrial- and marine-based margins could provide important context for predicting the long-term behavior of the GrIS in an anthropogenically warming world.

The southwestern portion of the GrIS extended beyond the modern coastline during the Last Glacial Maximum (LGM) (Kelly, 1985; Weidick et al., 2004; Fleming and Lambeck, 2004; Funder et al., 2011). Moraine systems on the inner continental shelf indicate that the LGM ice margin may have terminated 10-50 km offshore of the modern shore (Kelly, 1985; Funder et al., 2011). New core data and modeled southwest Greenland isostatic rebound, on the other hand, suggests that LGM ice extended much farther, to the continental shelf break (Bennike et al., 2002; Weidick et al., 2004; Ó Cofaigh et al., 2013; Lecavalier et al., 2014). Following the local LGM, the southwestern GrIS began to retreat inland from the shelf, eventually exposing glacially scoured coastal terrain. Radiocarbon dates along the southwest Greenland coast generally indicate that ice retreated inside of the modern coastline prior to 11-10 ka (e.g., Bennike and Björck,

2002; Larsen et al., 2014). ^{10}Be surface exposure ages from the coast near 66°N from Rinterknecht et al. (2009) and Roberts et al. (2009) suggest earlier ice retreat by ~18 ka and ~20 ka, respectively. Farther to the south at ~60°N, a radiocarbon age suggests first coastal land exposure by ~14.1 ka (Bennike et al., 2002).

Many datasets for last deglacial GrIS retreat rely on minimum-limiting radiocarbon ages. These ages can lead to underestimation of the responsiveness of outlet glaciers to climate forcings. They are also restricted in their ability to precisely constrain past thinning and retreat rates (Balco, 2011). Here, we present a suite of ^{10}Be surface exposure ages from four fjords in southwestern Greenland, sampled along vertical profiles of coastal hillslopes. This builds off of our recent reconstruction of retreat chronologies adjacent to the modern ice margin (Carlson et al., 2014). When the new coastal dates are combined with our and other data from the GrIS margin (Levy et al., 2012; Carlson et al., 2014; Larsen et al., 2014), we are able to calculate thinning and retreat rates of the ice margins. The purpose of this study is twofold: to construct retreat chronologies that do not rely on minimum-limiting ages, and to identify mechanisms that may explain the thinning and retreat rates of different fjords in the region.

2. Geologic Setting and Sampling Locations

^{10}Be surface exposure samples were collected from boulders in four fjord valleys in southern and southwestern Greenland, near the coastal towns of Sisimiut, Nuuk, Paamiut, and Qaqortoq (Fig. 1). A moraine was also sampled near the town of Kangerlussuaq, inland from Sisimiut. In general, the southern and southwestern Greenland coast consists of glacially smoothed hills gaining in elevation as they extend

inland. The terrain is punctuated by fjords that vary in length, with some extending to the present GrIS margin (inland ice), and others terminating distal of the ice margin.

The northern Labrador Sea abuts southern and southwestern Greenland. Surface and near-surface waters of the Labrador Sea are sourced from the Arctic and North Atlantic, bringing a mix of cooler and fresher water from the north and warmer and more saline water from the south. Modern observations of southeastern and western Greenland fjords demonstrate that the incursion of warm waters—on both seasonal and annual timescales—provides significant heat to fjord systems and has likely contributed to recent glacier acceleration and resultant thinning (e.g., Rignot et al., 2010; Mortensen et al., 2011; Joughin et al., 2012; Straneo and Heimbach, 2013). The influence of open ocean waters on southwestern Greenland fjord systems varies with bathymetry of the fjords, as many fjords possess sills that restrict entry of deeper water (e.g., Sparrenbom et al., 2006; Mortensen et al., 2011). This influence also varies with continental shelf depth and width, with some locales having particularly shallow offshore shelves.

2.1 Terrestrial Study Locales

2.1.1 Sisimiut to Kangerlussuaq

Sampling locations near Sisimiut are located ~150 km from the modern ice margin to the east, and ~120 km from the nearest shelf break to the west (Fig. 1, 2A). LGM ice extent near Sisimiut is uncertain, and may (e.g., Bennike and Björck, 2002; Lecavalier et al., 2014), or may not (e.g., Kelly, 1985; Fleming and Lambeck, 2004; Roberts et al., 2010) have reached the continental shelf edge. The marine limit at this location is ~140 m above sea level (asl) (Bennike et al., 2011). To the south of the town

and sampling area, the modern fjord extends only ~35 km inland, well distal of the current inland ice margin. To the north of Sisimiut, a small valley connects the coast to the hills and mountains inland, with ~300 m of elevation gain over less than 10 km. Here, a stream dissects proglacial sediment into terraces. Between the two valleys lies a ridge extending to ~590 m asl, where previous workers have observed a weathering boundary between smoothed bedrock below, and grussified, felsenmeer-like surfaces above (e.g., Kelly, 1985; Fig. 3). This ridge and the proglacial terraces to the north host the sampled boulders (Fig. 2A), with five sampled boulders above the trimline (~590 m asl), and four sampled boulders at lower elevations (~160 m asl). Low elevation samples were located above the local marine limit. Due to the steepness of the slope leading to the ridge, no samples were collected from middle elevations at this site.

We also collected samples from three boulders on the Keglen moraine where it abuts the base of Mount Keglen, which is also referred to as Mt. Sugarloaf (Fig. 4) (van Tatenhove et al., 1996). The moraine is ~25 km west of the present ice margin, is low-relief and clast-supported, and does not contain collapse features. We chose the three sampled boulders because they were propped up by smaller clasts and located on the moraine crest, in order to lessen the likelihood that post-depositional exhumation or settling would affect the results (Balco, 2011).

2.1.2 Nuuk

The town of Nuuk is ~70 km from the shelf break (Fig. 1). Like other coastal southwest Greenland sites, the landscape is that of smoothed, glacially scoured bedrock, which in this location is primarily gneiss. Inland of the town, a network of fjords extends

to modern outlet glaciers of the GrIS, located ~100 km to the east (Fig. 2B). The main fjord of this network, termed Nûp Kangerdlua or Godthåbsfjord, has an outer sill at ~170 m water depth, although the shelf shallows seaward to ~50 m below sea level (Fig. 2B) (Mortensen et al., 2011). The two smaller fjords of the network have outer sills at ~350 m below sea level to the south but shallower sills to the north (Mortensen et al., 2011). Although these sills limit the depth of water masses that can influence GrIS outlet glaciers, modern observations show that relatively warm and saline pulses of seawater entrained within the West Greenland Current do enter Nûp Kangerdlua (Mortensen et al., 2011).

Sampling locations near Nuuk were focused on the hilly, narrow strip of land to the south of Nûp Kangerdlua (Fig. 2B). Four high-elevation samples at ~420 m asl were located on a hilltop just outside of town, two middle-elevation samples (~330 m asl) were on the adjacent hill slope, and another four were taken from low elevations (~140 m asl). Here, the local marine limit is at ~90 m asl (Kelly, 1985).

2.1.3 Paamiut

The coastal town of Paamiut sits adjacent to Kuannersooq (Fig. 1), a branching fjord that extends to the modern inland GrIS (Fig. 2C). Smaller fjords that terminate distal of the outlet glacier terminus lie between Paamiut and the Frederikshåb ice cap (Fig. 1). Low-elevation, scoured, and jointed gneissic bedrock stretches from the coast eastward to the sampling sites, which are located where the land rises from ~50 m asl to ~550 m asl. The continental shelf extends ~60 km offshore. The region near Paamiut is

relatively unstudied, but LGM ice extent may have reached the shelf break (e.g., Weidick et al., 2004).

Just to the east of town, we sampled two erratics near the highest accessible elevation (~500 m asl), and another three from ~70 m below that (Fig. 2C). Two samples were collected from middle elevations (~250 m asl), and five from low elevations (~65 m asl). The lowest samples from Paamiut were from above, but very close to, the ~50 m asl marine limit (Kelly, 1985; Woodroffe et al., 2014).

2.1.4 Qaqortoq

Our southernmost study site is located just northeast of the town of Qaqortoq (Fig. 1) on a ridge rising to ~400 m asl from the shoreline. Fjords in this area are parallel and NE-SW oriented (Fig. 2D). Beyond the mainland coastline are numerous glacially smoothed bedrock islands. The continental shelf extends ~75 km from the westernmost land near Qaqortoq—the town itself is ~20 km inland of the coast. LGM ice may have extended near to, or reached, the shelf edge (e.g., Bennike et al., 2002; Weidick et al., 2004), and contributed to overdeepening of fjords in the area, which reach depths of ~600 m below sea level (Sparrenbom et al., 2006).

Three samples were collected from the highest elevation peak (~400 m asl), and an additional four from an adjacent, slightly lower elevation ridge top (~320 m asl) (Fig. 2D). Two sets of three samples each were also taken from low-middle (~180 m asl) elevations and low (~100 m asl) elevations. Like at the first three sites, Qaqortoq samples were all above the marine limit, which is estimated to be ~50 m asl near Qaqortoq (Kelly, 1985).

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189 **3. Methods and Age Calculation**

190 *3.1 Field and Laboratory Methods*

191 Our sampling for ^{10}Be surface exposure ages targeted boulders that rested directly
192 on bedrock or on a thin soil covering bedrock, with the one exception being the Keglen
193 moraine (see section 2.1.1). We avoided sampling bedrock directly because of evidence
194 for ^{10}Be inheritance in deglacial samples from southwest Greenland (e.g., Roberts et al.,
195 2009; Larsen et al., 2014). All boulders were chosen based on the following
196 characteristics: little to no visible weathering on the boulder surface and on the
197 surrounding substrate, a height >1 m, a horizontal to slightly dipping upper surface, a
198 high quartz content, and a location that would preclude post-depositional rolling (e.g., on
199 high points rather than on steep slopes). Topographic shielding and GPS coordinates
200 were measured in the field.

201 At the University of Wisconsin-Madison Cosmogenic Nuclide Laboratory,
202 samples were crushed and sieved to 480-825 μm to obtain grains of single mineral
203 species, then exposed to a Frantz magnetic separator to remove the more magnetic
204 minerals. Non-magnetic sample components were repeatedly acid-etched to isolate quartz
205 grains and remove meteoric ^{10}Be contamination. Quartz purity was verified by
206 inductively coupled plasma optical emission spectroscopy. Pure quartz samples were
207 fully dissolved with a ^9Be carrier and $\text{Be}(\text{OH})_2$ was isolated through column separation.
208 After ignition, the final BeO was measured by accelerator mass spectrometry (AMS) at
209 the Purdue Rare Isotope Measurement (PRIME) Laboratory and at Lawrence Livermore
210 National Laboratory (LLNL) (Rood et al., 2010; 2013).

Laboratory blanks using one of either a ^9Be carrier from the commercial Claritas standard or a custom-made low-level carrier from Oregon State University (OSU Blue) were processed with each sample set. Average $^{10}\text{Be}/^9\text{Be}$ blanks for the Claritas carrier were $1.316 \times 10^{-14} \pm 9.937 \times 10^{-16}$ (n=5) and for the OSU Blue carrier were $1.63 \times 10^{-15} \pm 3.05 \times 10^{-16}$ (n=14). PRIME and LLNL measurements were normalized to the 07KNSTD standard, which has a reported $^{10}\text{Be}/^9\text{Be}$ of 2.85×10^{-12} (Nishiizumi et al., 2007).

3.2 ^{10}Be Exposure Age Calculation

^{10}Be surface exposure ages were calculated using the CRONUS online calculator version 2.2, with constants version 2.2 (Balco et al., 2008). We use the Arctic sea-level high-latitude ^{10}Be production rate of 3.93 ± 0.15 atoms $\text{g}^{-1} \text{yr}^{-1}$ (Young et al., 2013b). This production rate has been used in Greenland and is similar to that found for Scandinavia and the Canadian Arctic (Young et al., 2013b). We use the time-varying spallation production rate-scaling scheme of Lal (1991) and Stone (2000) to account for temporal changes in geomagnetic field strength. When other spallation scaling schemes are used (Balco et al., 2008), final calculated ages differ by only ~2 percent (Supplementary Table).

No correction for snow cover was made due to the size of the boulders (>1 m) (Balco, 2011; Young et al., 2013a). Similarly, we do not correct for post-LGM isostatic uplift or late-Holocene subsidence. The marine limit at Qaqortoq, Paamiut, and Nuuk is between 50 and 90 m asl (Kelly, 1985; Bennike et al., 2002; Sparrenbom et al., 2006; Woodroffe et al., 2014), and taking an average of elevations experienced by samples since exposure yields very small changes from the samples' modern elevations (Young et

al., 2013a; b). Sisimiut, with a marine limit of ~140 m (Kelly, 1985), has experienced more isostatic rebound, but the effects are still small relative to the uncertainty in the ^{10}Be measurements (Rinterknecht et al., 2009; Hughes et al., 2012; Young et al., 2013a; b). Furthermore, the Young et al. (2013b) Arctic production rate was determined using ages that were not uplift corrected. We identify outliers as present when error bars do not overlap for a given sample set. For each elevation where we have multiple samples, we calculated an error-weighted mean age and uncertainty to account for varying analytical uncertainties (Balco, 2011).

We use our coastal ^{10}Be surface exposure ages along elevation transects to calculate ice thinning rates at four study locales along southwest Greenland. We also compare our lower elevation coastal ages to ^{10}Be surface exposure ages from samples inland (Levy et al., 2012; Carlson et al., 2014; Larsen et al., 2014) of each of our four sites to calculate ice retreat rates from the coast to the position of the modern margin. We also consider what effect an ice margin still-stand at the Keglen moraine would have had on the retreat rate of ice along our Sisimiut transect. Data from Roberts et al. (2009), Rinterknecht et al. (2009), Levy et al. (2012), Carlson et al. (2014), and Larsen et al. (2014) were recalculated using the Young et al. (2013b) Arctic production rate to allow inter-study comparison. We measure retreat distance between coastal and inland locales for the three southern sites using along-fjord lengths measured in Google Earth. For Sisimiut-Kangerlussuaq-inland ice, we measure retreat distance linearly between the two sites. Maximum, minimum, and mean thinning and retreat rates are provided in Table 3.

4. Results

¹⁰Be surface exposure ages (Fig. 5, 6) for our four sample locations (Fig. 2, 4) are presented in Table 2. At our northernmost site, Sisimiut, samples at ~590 m asl (n=5) date to 17.8 ± 1.1 , 33.2 ± 1.8 , 36.8 ± 1.8 , 37.4 ± 1.8 , and 37.6 ± 2.1 ka, with all but one sample dating to well before the LGM (Fig. 5A, 6). At lower elevations (~160 m asl), Sisimiut samples (n=4) date to 10.1 ± 0.6 , 13.5 ± 1.0 , 14.4 ± 1.5 , and 15.2 ± 0.9 ka. The 10.1 ka age is likely too young, and does not overlap with the other three ages within uncertainty. Thus, these lower elevation samples yield an error-weighted mean and uncertainty of 14.5 ± 0.6 ka if the 10.1 ka outlier is excluded. Our three boulder ¹⁰Be surface exposure ages from the Keglen moraine are 7.9 ± 0.2 ka, 7.9 ± 0.2 ka, and 8.1 ± 0.4 ka (Fig. 5A) with an error-weighted mean of 8.0 ± 0.2 ka.

Near Nuuk, samples at ~420 m asl (n=4) date to 9.7 ± 1.4 , 10.7 ± 1.0 , 10.9 ± 0.9 , and 11.4 ± 2.6 ka (Fig. 2B, 5B, Table 2). We calculate an error-weighted mean and uncertainty of 10.7 ± 0.6 ka. Samples at ~330 m asl (n=2) date to 9.6 ± 0.6 ka and 9.9 ± 0.9 ka, with an average of 9.7 ± 0.5 ka. Samples at ~140 m asl (n=4) yield dates of 9.7 ± 1.1 , 10.6 ± 1.1 , 11.7 ± 1.9 , and 14.9 ± 2.6 ka. We calculate an error-weighted mean of 11.0 ± 0.7 ka for the lowest elevation samples as Nuuk.

The Paamiut sampling location yielded a deglacial age of 11.7 ± 0.6 ka at 515 m asl and an early Holocene date of 8.9 ± 1.2 ka at 499 m asl (Fig. 2C, 5C, Table 2). We exclude the ~8.9 ka age as it is younger than inland ages for this region (Carlson et al., 2014) and not overlapping with the other deglacial age at this locale. The next sample site downhill at ~445 m asl dated to 10.2 ± 1.5 ka, 10.6 ± 1.6 ka, and 11.0 ± 2.0 ka, with an error-weighted mean of 10.6 ± 1.0 ka. Elevations at ~250 m asl (n=2) deliver dates of 9.9 ± 0.7 ka and 10.0 ± 0.6 ka, averaging 9.9 ± 0.4 ka. Elevations at ~60 m asl (n=5) date to

9.7 $\pm$ 1.5 ka, 10.4 $\pm$ 0.7, 12.0 $\pm$ 1.0, 12.3 $\pm$ 2.1, and 14.3 $\pm$ 3.3 ka, with an average of 11.2 $\pm$ 0.5 ka.

Qaqortoq samples at ~400 m asl (n=3) date to 9.8 $\pm$ 2.2, 11.7 $\pm$ 0.6, and 11.9 $\pm$ 0.8 ka (Table 2, Fig. 2D, 5D) with an error-weighted mean of 11.7 $\pm$ 0.5 ka. Samples at ~320 m asl (n=4) date to 8.8 $\pm$ 2.6, 12.1 $\pm$ 3.9, 12.6 $\pm$ 1.4, and 13.7 $\pm$ 4.2 ka. For this site, we report an error-weighted mean of 12.3 $\pm$ 1.1 ka. Samples at ~180 m asl (n=3) date to 10.5 $\pm$ 1.1, 12.6 $\pm$ 0.7, and 12.7 $\pm$ 0.6 ka. We suggest the youngest date of 10.5 ka is an outlier as it is not overlapping with the older two dates. This sample set thus possesses an error weighted mean of 12.7 $\pm$ 0.4 ka. Low elevation ages at ~100 m asl (n=3) date to 7.2 $\pm$ 1.0 ka, 8.8 $\pm$ 0.9, and 9.3 $\pm$ 1.2 ka.

5. Timing and Rates of Deglaciation

5.1 Sisimiut & Kangerlussuaq

Four out of the five high-elevation Sisimiut boulder samples from above the regional trimline (Kelly, 1985) cluster around 36.5 $\pm$ 1.0 ka, which pre-dates the global LGM, with only one sample dating from the last deglaciation (Table 2, Fig. 6). Kelly (1985), Roberts et al. (2009), and Rinterknecht et al. (2009) suggested that the trimline below our high-elevation samples represents an LGM ice-sheet boundary, with ice-free conditions above the weathering limit during the LGM (Fig. 3, 6). Alternatively, the pre-LGM dates at the high elevations near Sisimiut may be due to the presence of cold-based, less erosive ice during the last glacial period, with associated samples containing inherited nuclides.

Roberts et al. (2009) reported bedrock exposure ages at high elevations near Sisimiut of 24.2 ± 1.3 ka to 165.7 ± 8.2 ka (Figs. 2A, 6). Rinterknecht et al. (2009) reported a boulder date from above the trimline of 21.7 ± 4.1 ka (Figs. 2A, 6), which overlaps with our deglacial-age boulder sample. Given the occurrence of bedrock ^{10}Be ages above the trimline that pre-date the LGM but that are much younger than the penultimate deglaciation (Rinterknecht et al., 2009; Roberts et al., 2009), and that at least two boulder samples from above the trimline are deglacial in age (Table 2; Rinterknecht et al., 2009), we suggest that the trimline likely represents an englacial thermal boundary. Above this demarcation cold-based ice cover during the LGM only partly removed inherited nuclides in the bedrock and boulder surfaces. This inference of cold-based ice agrees with other studies in Greenland where bedrock surfaces above an erosional trimline contained significant inheritance while boulder surfaces from above the trimline sometimes contained minimal inheritance (e.g., Roberts et al., 2013; Lane et al., 2014; Larsen et al., 2014). Our single early deglacial date likely exhibits the least inheritance. Therefore, we use only the younger deglaciation date when calculating thinning from high elevation near Sisimiut. Using this single high-elevation date of 17.8 ± 1.1 ka and the low elevation error-weighted mean of 14.5 ± 0.6 ka, ~440 m of thinning occurred at this location over the course of ~3 ka (Fig. 6), thinning between $0.1\text{--}0.3$ m yr^{-1} (Table 3). If the 17.8 ± 1.1 ka sample did in fact contain inherited nuclides, these estimated thinning rates would be slower than the actual thinning rates.

Our ^{10}Be surface exposure ages, and those of Rinterknecht et al. (2009) and Roberts et al. (2009), suggest an early onset of deglaciation from the coast near Sisimiut, around the onset of Northern Hemisphere deglaciation (e.g., Carlson and Winsor, 2012).

While limited, there is evidence for retreat of southeast and east GrIS margins before ~17 ka (Nam et al., 1995; Andrews et al., 1997; Carlson et al., 2008). Our timing of the pull back of the GrIS from the Sisimiut coast at ~14.5 ka is earlier than, but not inconsistent with, reconstructions based on minimum-limiting ^{14}C dates (Fig. 2A) (Bennike et al., 2011), and is concurrent with the initial retreat of the west GrIS margin from the continental shelf break west of Disko Bugt (Ó Cofaigh et al., 2013). ^{10}Be surface exposure ages on GrIS erratic boulders and valley glacier moraines in Scoresby Sund also show that the much larger LGM outlet glacier had retreated from the shelf before ~14 ka (Håkansson et al., 2007; Kelly et al., 2008). Therefore, the GrIS retreat recorded around Sisimiut may not be an isolated phenomenon as there is limited evidence elsewhere on Greenland for such early deglacial thinning and retreat.

While exposure at the coast occurred in the early part of the last deglaciation, the GrIS margin retreated to its present extent by the middle Holocene. The Carlson et al. (2014) and Levy et al. (2012) ^{10}Be surface exposure ages from near the inland ice margin average 6.7 ± 0.3 ka ($n=16$), when recalculated using the Arctic sea-level high-latitude production rate (Young et al., 2013b). Using our low elevation error-weighted mean ^{10}Be surface exposure age of 14.5 ± 0.6 ka, retreat rates from Sisimiut to the inland ice are ~20 m yr^{-1} . This retreat rate from Sisimiut is also reconstructed by Rinterknecht et al. (2009), who place the ice margin ~20 km west of its present extent at ~9 ka ($n=3$). Both our timing of retreat and our retreat rates from Sisimiut are in contrast to those calculated for Disko Bugt, ~200 km to the north. There, the modern coastline was exposed 11-10 ka and ice retreated at a rate of 50-450 m yr^{-1} (Kelley et al., 2013).

Our three ^{10}Be surface exposure ages from the Keglen moraine of 8.0 ± 0.2 ka agree with a minimum-limiting ^{14}C constraint of the moraine being older than ~ 7.3 ka (van Tatenhove et al., 1996). Unpublished dates by Levy et al. (2013) ($n = 6$) give a ^{10}Be surface exposure age of the Keglen moraine of 7.3 ± 0.1 ka, which is slightly younger than our average. Our Keglen moraine ages indicate a still-stand of the ice margin, possibly during the 8.2 ka event (Alley et al., 1997). An 8.2 ka event moraine has been dated further north in the Disko Bugt region (Young et al., 2013a), but similarly dated moraines are absent elsewhere in western Greenland (Carlson et al., 2014; Larsen et al., 2014). The rate of GrIS margin retreat from the coast to the Keglen moraine was ~ 20 m yr^{-1} and the rate of retreat from the moraine to the present ice margin was also ~ 20 m yr^{-1} . We thus suggest that this still-stand did not greatly affect the overall ice-margin retreat rate we document here.

There are no crosscutting relationships to suggest a major readvance at the Keglen locale (van Tatenhove et al., 1996). Additionally, the relatively warm conditions associated with the Holocene Thermal Maximum (e.g., Kaufman et al., 2004) would have made a significant readvance during this time unlikely. While Young et al. (2011) document a rapid response of Jakobshavn Isbræ during both the 8.2 ka and 9.3 ka events, those authors conclude that it is the marine-terminating nature of Jakobshavn that permitted this behavior. The land-terminating ice at Keglen would likely not have had time to undergo significant readvance during the centennial-scale cooling of the 8.2 ka event within the generally warm early Holocene.

5.2 Nuuk

¹⁰Be surface exposure ages from Nuuk indicate that ice thinned ~280 m between
 10.8 ± 0.7 ka and 10.4 ± 0.7 ka (Fig. 5B). Our middle elevation average age at Nuuk of
 9.7 ± 0.5 ka overlaps within uncertainty with the overall thinning history. The minimum
 estimated thinning rate is ~0.2 m yr⁻¹. The maximum thinning rate is essentially
 instantaneous within the uncertainties of our measurements. The timing of coastal retreat
 is consistent with minimum-limiting ¹⁴C dates in the fjord west of our Nuuk samples of
 ~10.7 to ~11.4 ka (Weidick, 1976; Larsen et al., 2014). Likewise, three coastal ¹⁰Be
 surface exposure ages on boulders several fjords south of Nuuk range from 10.4 ± 0.4 ka
 to 10.8 ± 0.7 ka (Larsen et al., 2014), in excellent agreement with our dates.

We calculate a minimum retreat rate from Nuuk of ~90 m yr⁻¹ (Table 3, Fig. 7C),
 using an error-weighted mean age of 10.1 ± 0.3 ka from ¹⁰Be surface exposure ages (n=3)
 just to the south of the head of Nûp Kangerdlua (Larsen et al., 2014), and our low
 elevation date of 10.4 ± 0.7 ka (Fig. 5B). The maximum retreat is essentially
 instantaneous within the uncertainty of these dates. This range agrees well with that
 estimated by Larsen et al. (2014) of 90-210 m yr⁻¹ using minimum-limiting ¹⁴C dates.

5.3 Paamiut

At Paamiut, ~450 m of ice thinning occurred between 11.7 ± 0.6 ka and 11.2 ± 0.5
 ka (Table 2, Fig. 5B). Our highest elevation site is only dated by one deglacial boulder
 sample, but another three boulder samples from ~70 m lower are 10.6 ± 1.0 ka in age.
 The middle elevation site error-weighted mean age of 9.9 ± 0.4 ka is almost within
 stratigraphic order. We estimate thinning rates between infinite and ~0.2 m yr⁻¹. Our

timing of retreat from the coast at ~ 11.2 ka agrees with the oldest nearby minimum-limiting ^{14}C date of 11.0 ka (Woodroffe et al., 2014).

Carlson et al. (2014) showed that the GrIS margin neared its current extent by Paamiut at 10.4 ± 0.4 ka ($n=4$), which overlaps with our low elevation date of 11.0 ± 0.5 ka. Another five boulder dates from high elevations (~ 740 m asl) near the inland GrIS margin averaged 10.7 ± 0.1 ka. (Carlson et al., 2014). Therefore, simultaneous deglacial exposure at the coastal Paamiut site and the modern outlet glacier terminus, ~ 40 km away, is possible within uncertainty (Table 3, Fig 7B). We calculate a minimum retreat rate of ~ 45 m yr^{-1} . The timing of retreat from the coast up-fjord is in general agreement with the last pulse in ice-rafted debris north of Paamiut ~ 11 ka to ~ 9 ka (Fig. 1) (Knutz et al., 2011), where icebergs would be transported in the West Greenland Current.

5.4 Qaqortoq

The highest elevation samples (~ 400 m asl) at Qaqortoq date to 11.8 ± 0.5 ka while the lowest samples (~ 60 m asl) are between 8.8 ± 1.0 ka and 9.3 ± 1.3 ka (Table 2, Fig. 2D). Minimum-limiting ^{14}C ages show ice-free conditions in fjords near Qaqortoq by ~ 11 ka (Sparrenbom et al., 2006). Therefore, we suggest that the three lowest elevation Qaqortoq boulders are erroneously young and may have experienced post-depositional rolling. As a result, we calculate thinning over ~ 220 m between our highest elevation site (11.7 ± 0.5 ka) and our second lowest elevation site (~ 180 m asl; 12.7 ± 0.4 ka). Thinning rates are infinite within the uncertainty of our measurements (Table 3). Our ^{10}Be ages agree with the oldest minimum-limiting age from the outermost island southwest of Qaqortoq suggesting ice-free conditions by ~ 14.1 ka (Bennike et al., 2002).

Inland of Qaqortoq near the present ice margin near the town of Narsarsuaq, surface exposure ages ($n=4$) date to 11.1 ± 0.2 ka (Carlson et al., 2014), consistent with other nearby ^{10}Be ages (Nelson et al., 2014). This boulder data and our lowest *in situ* elevation Qaqortoq boulder data (12.7 ± 0.4 ka) suggest retreat rates from Qaqortoq inland of $30\text{--}80$ m yr^{-1} (Table 3, Fig. 7A). Using mainly minimum-limiting ^{14}C dates, Larsen et al. (2011) estimated a minimum retreat rate of ~ 30 m yr^{-1} . A GrIS runoff record from south of Qaqortoq on the Eirik Drift (Fig. 1) shows the major peak in southern GrIS deglaciation occurring ~ 13.0 ka to ~ 10.5 ka (Carlson et al., 2008; Fig. 7H), concurrent with our ^{10}Be dated ice retreat in southernmost Greenland.

5.5. Summary of thinning and retreat rates

We have calculated thinning rates at the coast, and retreat rates inland of these sites, using a combination of new ^{10}Be ages and previously published ^{10}Be ages (Fig. 7, Table 3; Levy et al., 2012; Larsen et al., 2014; Carlson et al., 2014). Coastal thinning occurred slowly at Sisimiut (0.1 to 0.3 m yr^{-1}), more rapidly at Nuuk and Paamiut (0.2 m yr^{-1} to instantaneous exposure), and even more rapidly at Qaqortoq (instantaneous exposure within measurement). Retreat inland from Sisimiut was also relatively slow, at a rate of ~ 20 m yr^{-1} . Retreat from Qaqortoq was faster, at $30\text{--}80$ m yr^{-1} . From coastal Paamiut, retreat was at least 30 m yr^{-1} , and the maximum retreat rate implies instantaneous exposure within our uncertainties. Retreat from Nuuk was the fastest of our four retreat estimates, with a minimum value of 90 m yr^{-1} and a maximum that again implies instantaneous exposure within uncertainty (Table 3, Fig. 7C).

These retreat rates are similar to, though potentially faster than, those estimated for other GrIS margins. Roberts et al. (2013) estimated deglacial retreat of the Uummannaq ice stream (71.0°N, 53.7°W; Fig. 1), west Greenland, at ~30 m yr⁻¹. This is slower than all but one of our retreat rates for Qaqortoq, Paamiut, and Nuuk, and faster than maximum rates at Sisimiut (Table 3, Fig. 7). Kelley et al. (2013) report retreat from outer to inner Disko Bugt as 50-450 m yr⁻¹ for the central Jakobshavn outlet glacier, and 50-70 m yr⁻¹ for the surrounding glacier margins. Overlapping with these rates are those proposed by Ó Cofaigh et al. (2013), who calculate a 22-250 m yr⁻¹ retreat rate in the Disko Trough.

South of Nuuk, Larsen et al. (2014; Fig. 1) calculated a retreat rate of 90-210 m yr⁻¹. Our minimum retreat rate from Nuuk is identical to their minimum value, with our maximum rate of infinite-within-uncertainty being greater than the Larsen et al. (2014) maximum estimate (Table 3, Fig. 7C). In southeastern Greenland, Hughes et al. (2012) document late deglacial retreat rates of ~80 m yr⁻¹ for Helheim Glacier in Sermilik Fjord between ~11 and ~10 ka (65.9°N, 37.8°W; Fig 1). Similar retreat rates were also found at other southeastern outlet glaciers (Fig. 1; Dyke et al., 2014). These are comparable to our minimum retreat rate estimates at Nuuk, our maximum estimate at Qaqortoq, and between our minimum and maximum values at Paamiut (Table 3, Fig. 7). The southeast GrIS retreat rates are, however, much faster than at Sisimiut (Fig. 7D; Hughes et al., 2012; Dyke et al., 2014).

6. Environmental Influences on southwest Greenland Deglaciation

Why did retreat inland from Sisimiut occur at a slower average pace than at our three more southern sites? The Sisimiut fjord system is less extensive than the fjords around Qaqortoq, Paamiut, and Nuuk (Fig. 2). This means that the ice margin near Sisimiut would have become largely isolated from oceanic influences before reaching Kangerlussuaq, even with a marine limit of ~140 m asl (Kelly, 1985). Just to the west of Kangerlussuaq (Fig. 2), Kelly (1985) observed a marine limit of ~65 m asl, and at Mt. Keglen (~25 km distal of the modern margin) a marine limit of ~50 m asl. Modern minimum elevations in the valleys between Sisimiut and Kangerlussuaq are on the order of 100-150 m asl, permitting 250-300 m of maximum water depth at the retreating ice margin close to the modern coastline, and 100-150 m maximum depth close to Kangerlussuaq. However, low elevation valleys are very narrow and winding in this region, with intervening ridges that would have limited the areal contact of the ice margin with the penetrating marine system. Considering this topography inland of our Sisimiut study location, the ice-sheet margin in this region would likely lose mass primarily via surface melting as opposed to by calving and submarine basal melting. We thus use the retreat rates from Sisimiut to the inland ice as an example of how fast an ice-sheet margin can retreat when it is largely driven by surface ablation.

The outer sill at Nuuk is ~170 m below modern sea level (Mortensen et al., 2011), and with a coastal marine limit of ~90 m asl (Kelly, 1985), this restricted deglacial seawater deeper than ~260 m from entering the fjord (Fig. 2B). Unlike the fjords near Sisimiut, the fjord inland of Nuuk deepens. The two southernmost outlet glaciers entering Nûp Kangerdlua (Fig. 2B) are grounded below sea level near their modern margin (Morlighem et al., 2014). Although fjord bathymetry is not constrained at Paamiut,

Morlighem et al. (2014) indicate that subglacial bed topography upstream from the modern outlet-glacier terminus is below sea level. The bed of outlet glacier Kiagtût Sermiat at the head of our Qaqortoq transect, (Fig. 2D) is above modern sea level, although other nearby glacier termini are grounded below sea level and fjords do extend inland close to the Kiagtût Sermiat terminus (Morlighem et al., 2014). This relatively shallow bathymetry may help explain the slower maximum retreat rate estimated at Qaqortoq, compared to those at Nuuk and Paamiut (Table 3, Fig. 7). Considered as a whole, the fjords of Nuuk, Paamiut, and Qaqortoq are wider and deeper than the fjords near Sisimiut and extend inland close to or at modern GrIS margins. It is likely that calving and submarine melting played a larger role for GrIS mass loss near Nuuk, Paamiut and Qaqortoq relative to near Sisimiut-Kangerlussuaq. These additional ablation mechanisms may explain the more rapid thinning and retreat inland near Nuuk, Paamiut, and Qaqortoq relative to Sisimiut. Similar to our Sisimiut-Kangerlussuaq transect, ice retreat from land-terminating Paakitsoq, just east of Disko Bugt, occurred into the Holocene. There, ^{10}Be ages of ~ 7.9 ka and a minimum-limiting ^{14}C date of ~ 6.5 ka constrains latest deglaciation (Carlson et al., 2014; Håkansson et al., 2014). The relatively slow retreat from Sisimiut and Paakitsoq, geographically bracketed by the faster retreat at both Disko Bugt to the north (Kelley et al., 2013) and Nuuk to the south, lends support to a marine driver of retreat for the marine-terminating margins.

Indeed, deglacial warming of the eastern Labrador Sea began ~ 13 ka and reached maximum warmth by ~ 10 ka, likely due to the influx of warm Irminger Current waters (Solignac et al., 2004; Knutz et al., 2011; Winsor et al., 2012; Fig 7F). This warming is concurrent with up-fjord ice retreat at our three southern locations. The strong contrast

between the timing and rate of thinning and retreat at Sisimiut and the sites with extensive fjord systems could therefore be explained by the relative strength of marine influence on local ice margins.

Another possibility is that the abrupt retreat of southwest GrIS margins was a response to warming at the end of the Younger Dryas cold event (Fig. 7H). However, such abrupt atmospheric warming should affect both the terrestrial and marine ice margins equally and would thus not alone explain the different retreat rates. In this case, initial warming could have initiated retreat into deeper fjord waters inland of the grounding line, triggering sustained and rapid retreat. Climate models and new ice-core reconstructions show that the cooling during the Younger Dryas may be overestimated in the ice-core $\delta^{18}\text{O}$ records (Fig. 7H), with much of the cooling focused in the winter when it would not affect surface ablation (Denton et al., 2005; Liu et al., 2012; Buizert et al., 2014).

Ice-sheet models of the GrIS forced with ice-core $\delta^{18}\text{O}$ -inferred climate generally lag the deglacial timing we have dated (Tarasov and Peltier, 2002; Lecavalier et al., 2014). These models do not deglaciade Sisimiut until well after our ^{10}Be ages, nor do they reproduce the rapidity of ice-margin retreat we document from Nuuk southwards. These ice-sheet models do not explicitly include a submarine ablation component from ocean temperature changes, nor do they possess the resolution to simulate behavior of small outlet glaciers. We suggest that this model-data mismatch on ice-margin retreat rates could therefore be due in part to the lack of marine temperature forcing, which would support our inference that rapid marine margin retreat was in response to ocean warming.

We conclude that the difference in retreat rates along southwest Greenland is likely due to the ice-margin setting. Marine ice margins with the additional heat from the ocean (Rignot et al., 2010; Joughin et al., 2012) are prone to significantly more rapid retreat than terrestrial ice margins that have only an atmospheric forcing.

7. Conclusions

We present a new deglacial thinning and retreat chronology of four locations in southwestern Greenland using ^{10}Be surface exposure dating. Our data show that ice retreat in three of the fjords occurred at 12.0-10.0 ka at rates $>30\text{-}90\text{ m yr}^{-1}$. Ice retreat in a fourth region with relatively little marine influence (having shallow and short fjords) initiated comparatively early ($\sim 17.8\text{ ka}$ to $\sim 14.5\text{ ka}$), but at a rate of only $\sim 20\text{ m yr}^{-1}$, including several ice-margin still-stands as evidenced by local moraine systems. Together, our fjord retreat chronology suggests that ocean warming—and transfer of ocean heat to the outlet glacier systems—may have played an important role in the timing and rate of southwestern GrIS retreat. Alternatively, our chronology could reflect initial retreat driven by atmospheric warming and sustained by ice dynamics unique to marine-terminating glaciers (e.g., surface meltwater production lubricates basal ice, followed by acceleration, thinning, and retreat). A marine influence on ice retreat is consistent with modern observations of accelerated outlet glacier thinning and retreat coeval with the incursion of warmer waters into Greenland fjord systems (e.g., Straneo and Heimbach, 2013). Interestingly, modern ice thinning rates for marine GrIS margins are in some places $>0.2\text{ m yr}^{-1}$ (Pritchard et al., 2009). Our deglacial ^{10}Be chronology suggests that

such rates could be sustained for centuries in marine-terminating settings, with attendant significant mass loss from the GrIS.

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761 FIGURE CAPTIONS

762 Figure 1. Index map from Bamber et al. (2013) of Greenland study locales. Shown are:
 763 previous southern Greenland ^{10}Be surface exposure ages (yellow circles), ^{10}Be surface
 764 exposure ages from this study (orange circles), GISP2 ice core location (blue square), and
 765 mapped modern elevations (m) above sea level. Previous studies are numbered
 766 accordingly: ¹Carlson et al. (2014), ²Dyke et al. (2014), ³Håkansson et al. (2014),
 767 ⁴Hughes et al. (2012), ⁵Kelly et al. (2008), ⁶Larsen et al. (2014), ⁷Levy et al. (2012),
 768 ⁸Rinterknecht et al. (2009), ⁹Roberts et al. (2009), ¹⁰Roberts et al., (2013), ¹¹Young et al.
 769 (2013a).

770

Figure 2. Close-up satellite imagery showing the study sites of A. Sisimiut, B. Nuuk, C. Paamiut, and D. Qaqortoq. Blue circles indicate sample locations for ^{10}Be dating published in Rinterknecht et al. (2009, medium blue), Roberts et al. (2009, light blue), and Larsen et al. (2014, medium blue). Orange circles mark our coastal locales, with nearby towns (pink circles), and yellow circles mark inland locales used to calculate retreat rates. Sites of relative sea level data are shown in green circles, with ages representing the start time of rapid deglacial sea level rise (Sparrenbom et al., 2006; Woodroffe et al., 2014). North is oriented page-up.

Figure 3. Geomorphic difference between high-elevation Sisimiut and other coastal locales. (A) Highly weathered boulders in an environment resembling felsenmeer on the ridge-top sampling site at Sisimiut, and (B) glacially smoothed and exposed bedrock at the high elevation Nuuk site, exhibiting glacial scouring and little post-glaciation weathering.

Figure 4. Kangerlussuaq study site showing Keglen moraines. (A) Satellite imagery showing positions of mapped Keglen moraines (yellow dashes) and younger Ørkendalen moraines (white dashes). Orange dots indicate sampling areas of both Carlson et al. (2014) and Levy et al. (2012). (B) Photo of Mt. Keglen, stream terraces, and sampling location. North is oriented page-up.

Figure 5. New ^{10}Be ages with internal errors for each boulder at (A) Sisimiut (maroon) and Keglen (orange), (B) Nuuk (brown), (C) Paamiut (green), and (D) Qaqortoq (purple).

Outliers are included in this figure (gray circles), but pre-LGM Sisimiut dates are not.

Figure 6. All ^{10}Be data from the Sisimiut region, including new ages from this study (red circles) and data from Rinterknecht et al. (2009; black diamonds) and Roberts et al.

(2009; gray diamonds), recalculated using the Young et al. (2013b) Arctic production rate for spallation. Error-weighted mean age for this study's low elevation samples is shown with orange square.

Figure 7. Paleoclimate context for our ^{10}Be surface exposure ages. (A-D) the error-weighted mean of our new coastal ^{10}Be dates (squares) and published inland ^{10}Be dates (open circles) (Levy et al., 2012; Carlson et al., 2014; Larsen et al., 2014), (E) June insolation at 60°N (Berger and Loutre, 1991), (F) Ca-corrected Ti concentrations from the northeastern Labrador Sea representing GrIS runoff (Carlson et al., 2008), (G) Mg/Ca-derived calcification temperatures of *Neogloboquadrina pachyderma* (s) from the northeastern Labrador Sea (Winsor et al., 2012), and (H) Greenland ice core GISP2 $\delta^{18}\text{O}$ (Grootes and Stuiver, 1997). Bars indicate intervals of surface exposure at coastal and inland sites.

811 Table 1. Sampling and lab processing data for ^{10}Be surface exposure dating.

812

813 Table 2. AMS results and age calculation of boulder samples using the CRONUS online

814 calculator (Balco et al., 2008). All ages calculated with the Arctic high latitude/sea level

815 production rate (Young et al., 2013b), zero erosion, a rock density of 2.65 g cm^{-3} , and the

816 Lal/Stone time-varying spallation scaling scheme (Lal, 1991; Stone, 2000). Excluded

817 outliers are represented in italics. Site averages and uncertainties are indicated below

818 each sample group.

819

820 Table 3.

821 Maximum and minimum calculated thinning (this study) and retreat (this study; Carlson

822 et al., 2014¹; Larsen et al., 2014²; Levy et al., 2012³) rates. Elevation change between

823 high and low elevation sites, and distance between coastal and inland sites, is also listed.

824

825 Supplementary Online Table. Results using alternative production rate scaling schemes

FIGURES

Figure 3

Figure 4

Figure 5

Figure 6

Figure 7

Tables 1, 2, 3

Sample Name	Latitude (N)	Longitude (E)	Elevation (m asl)	Shielding correction	Sample Thickness (cm)	Quartz mass (g)
QAQORTOQ						
QQ-6-08	60.733	-46.018	97	0.991	1.0	28.401
QQ-7-08	60.733	-46.018	97	0.991	2.0	24.585
QQ-8-08	60.733	-46.018	97	0.991	2.0	26.268
QQ-4-10	60.731	-46.023	175	0.997	1.0	28.303
QQ-5-08	60.732	-46.022	176	0.994	2.5	35.946
QQ-5-10	60.732	-46.023	185	0.996	2.0	27.219
QQ-3-08	60.735	-46.031	305	1.000	2.0	21.299
QQ-4-08	60.735	-46.031	305	1.000	1.0	11.220
QQ-1-08	60.736	-46.033	313	1.000	2.0	12.002
QQ-2b-08	60.735	-46.033	326	1.000	2.0	46.976
QQ-1-10	60.744	46.046	399	1.000	1.0	21.769
QQ-2-10	60.744	-46.046	399	1.000	1.0	28.246
QQ-3-10	60.744	-46.046	399	1.000	1.0	28.937
PAAMIUT						
PA-4-10	62.005	-49.554	57	0.999	2.3	20.126
PA-5-10	62.005	-49.553	59	0.998	2.8	29.977
PA-1-10	62.004	-49.550	65	0.997	3.0	29.041
PA-2-10	62.004	-49.550	65	0.997	1.8	28.777
PA-3-10	62.004	-49.550	66	0.997	1.0	22.805
PA-12-10	62.007	-49.540	251	0.994	3.5	26.043
PA-11-10	62.007	-49.538	257	0.994	1.5	23.717
PA-10-10	62.011	-49.529	439	0.988	3.3	32.242
PA-8-10	62.011	-49.530	448	0.990	2.3	28.774
PA-9-10	62.011	-49.530	448	0.990	3.4	29.376
PA-7-10	62.012	-49.528	499	0.985	2.0	29.481
PA-6-10	62.012	-49.528	515	0.959	2.0	30.233
NUUK						
NU-1-10	64.178	-51.674	126	0.995	2.6	20.824
NU-3-10	64.178	-51.673	144	0.996	1.9	20.113
NU-7-2008	64.178	-51.672	146	0.997	1.5	22.739
NU-2-10	64.178	-51.672	149	0.997	2.0	17.984
NU-5-2008	64.182	-51.657	333	0.998	2.0	21.456
NU-4-2008	64.183	-51.657	334	0.998	1.5	23.837
NU-1-2008	64.187	-51.646	419	1.000	2.5	22.578
NU-2-2008	64.187	-51.646	422	1.000	2.0	19.258
NU-4-10	64.187	-51.645	427	1.000	2.7	23.071
NU-5-10	64.187	-51.645	427	1.000	3.8	20.305