Supplementary Information

Acceleration of Jakobshavn Isbræ Triggered by Warm, Subsurface Irminger Waters

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Background

Considerable progress has been made over the last decade in constraining the present mass budgets of the large ice sheets, Greenland and Antarctica, largely by airborne and satellite techniques S1,S2, with both sheets presently considered to be contributing to observed global sea-level rise. The mass budgets are a function of atmospheric and oceanographic conditions near the sheets. While the atmospheric conditions have been relatively well observed over past decades S3, the oceanographic conditions are poorly known, having suffered from much sparser observations. The presence of ice and ice-calving makes such observations difficult to obtain. We here provide some brief discussion of the factors that influence glacial dynamics with an emphasis on the circulation of subsurface ocean waters along the periphery of the ice sheets, and of the Jakobshavn Isbræ (JI) in particular.

Glaciology

The present mass balance of the ice sheets reveals a net gain of mass over the vast interior surface areas, and a net loss at the edge. Some loss is caused by increased summer melting, particularly in Greenland, and some by recent acceleration of specific glaciers and ice streams that transport ice from the slow moving interior to the coast. There, it is discharged into the ocean, either by calving of icebergs or by melting from beneath floating ice tongues.

Until recently, JI terminated as a floating ice tongue. The JI is unique in having a detailed time series of glaciological measurements since the early 1990s, spanning the breakup of the floating tongue and subsequent glacier acceleration. It is also known that the JI calving front retreated
30 km between 1850 and 1964 and subsequently occupied approximately the same location until the late 1990s. Balance calculations suggest that total loss by surface and basal melt and ice discharge was slightly less than total accumulation within the drainage basin. Estimates of anomalous surface melting using positive-degree-day methods are small compared to total thinning, suggesting that a large part of the thinning is dynamic, associated with increased longitudinal creep rates, and therefore velocity.

A curious fact is that while the JI is now flowing faster than prior to 1998, and thus now transporting more land ice towards the sea, it is at the same time retreating; the ice is thinning more rapidly than can be balanced by the thickening effects of increased advection of thicker ice from upstream. Extremely rapid thinning results because the ice is so deep at the grounding line, where removal of the ice tongue allows creep thinning rates to be roughly proportional to the forth power of the ice thickness. Continued rapid thinning inevitably causes additional grounding-line retreat over a bed that deepens further inland. Consequently, continued retreat and probably further acceleration is likely unless a new ice tongue develops, and “jams up” part of the fjord. This is unlikely, because the heavily-crevassed glacier fragments into small icebergs close to the calving front, and because continued subsurface warm waters in the fjord favor rapid melt rates.

**Oceanography**

Relative to other factors influencing the mass balance of the ice sheets, the temperature of the subsurface waters near the ice edge is, as mentioned, difficult to observe as the ocean there is frequently covered by ice shelves, icebergs, or sea ice. Nonetheless measurements have been made using either traditional oceanographic instrumentation or autonomous underwater vehicles. Available data show the presence of relatively warm, saline waters several hundred meters below the ocean surface, well separated from the cool, fresh surface waters. Indeed, such warm, subsurface waters are quite common in the polar oceans. They originate in the extra-polar oceans and although warmer than the surface waters of the polar oceans, they are saltier and denser and consequently as they move poleward they sink beneath the cold, fresh polar surface waters and generally spread laterally at depths of between about 200 and 800 m. Where such waters come directly into contact with the periphery of an ice sheet, enormous melt rates result. Floating glacier tongues and the ice shelves into which most Antarctic glaciers flow are particularly susceptible to increased basal melting, as recent observations show that some of these ice shelves are already thinning.

In the particular instance of the oceanography of West Greenland, changes in the ocean temperature regime are traditionally described on the basis of measurements made at oceanographic standard stations, such as SJA3 as discussed in the main text (Fig. 3). In turns out that from a fisheries-science perspective, these data are too sparse and do not necessarily reflect the conditions faced by demersal species. In contrast to standard oceanographic stations, bottom temperatures from fishery trawls sites show the conditions encountered by species on a finer spatial scale. Interestingly, these data are the most complete set available to demonstrate the presence of subsurface warm water on the continental shelf. These are precisely the waters that are critical to monitor to assess the degree of ocean-glacier interaction. Only by serendipity did the demersal fisheries surveys record a deep ocean temperature jump for the JI in 1997 (Fig. 4).
In general, it must be acknowledged that the monitoring of the subsurface ocean waters near the ice sheets and beneath ice shelves remains an acute observational problem.

Supplementary References

Figure S1. Schematic of water masses near the continental shelf break, West Greenland\textsuperscript{512,513}. The waters on the continental shelf here consist of two distinct water masses. The surface layer consists of Polar Water (PW), a cold, fresh water mass originating in the Arctic Ocean, but with local input due to freshwater runoff from coastal glacier melt. The subsurface layer consists of Irminger Water (IW), a relatively warm, saline water mass originating in the Irminger Sea of the North Atlantic. As IW advects around the southern tip of Greenland, it subducts beneath the PW. Along the west coast of Greenland the core of the subducted IW is typically found at depths of about 400-600 m, and located at the continental shelf break.
Figure S2. Terminus location for the JI during various years over the period 1851 - 2006\textsuperscript{14}. The glacier has undergone a nearly continuous retreat over the entire period, but at rates that have varied over time. Most notable, the glacier front essentially stagnated during the period 1964 to 2001, and then during 2001-2006 recorded its largest retreat rate.
Figure S3. Time series from standard oceanographic sections, West Greenland south of Disko Bay\textsuperscript{S15}. a Map of Greenland showing (as red dotted lines) some of the standard oceanographic sections taken annually in early summer. The data from the northernmost section shown in panel b is ‘Hol5’ (near Holsteinborg), and the second northernmost section data shown in panel c is ‘Suk5’ (near Sukkertoppen). b Temperature and salinity collected at station Hol5 during the period 1950-2005. The data clearly shows a steady increase in Irminger Water temperature in the 400-600 m depth range since 1996, and in the intermediate waters from 150-400 m between the Polar water core (\textasciitilde 50 – 150 m) and the Irminger Water core (\textasciitilde 400 – 600 m). c Same as panel b, except for standard section Suk5, further south.
Figure S4. Observations of warm, subsurface waters at Jakobshavn during summer 2007.

a Blue curve is an AXCTD temperature profile taken in the fjord August 21, 2007, at the location marked by blue square icon in Fig 2a. Red curves illustrate one dozen CTD temperature profiles taken June 7, 2007, in the vicinity of the mouth of the fjord, marked by red circle icon in Fig 2a.

b Same as panel a, except for salinity.

c Thickness (in meters) of the observed bottom layer of warm water (defined as having temperature above 1.5 °C) near the mouth of the fjord derived from the CTD data in panel a (red curves). The gray shading indicates a mask for land areas in the vicinity of the mouth of the fjord. The relatively thick, warm water layer sits on the floor of Disko Bay at depths above the nearby sill depth of 350 m at the mouth of the fjord, and thus has direct access to the ocean fjord.
Figure S5. Schematic of positive and negative phases of the North Atlantic Oscillation (NAO). The North Atlantic marine climate is largely influenced by the so-called North Atlantic Oscillation (NAO), driven by the pressure difference between the Azores-High and the Icelandic-Low pressure cells. This is the dominant mode of variability in the North Atlantic winter climate. 

a Positive NAO index is associated with a more northeasterly storm track. 
b Negative NAO index is associated with a more southerly, and weaker, storm track.
Figure S6. Schematic of North Atlantic Ocean subpolar gyre during positive and negative phases of the atmospheric NAO\(^{518}\). a The NAO is known to impact the ocean circulation and hydrologic conditions in the North Atlantic region including the western Greenland waters. The North Atlantic Current (NAC) is narrowed and strengthened during periods of high NAO. This is caused by an increase in westerly wind stress, and the oceanic current response is sketched in panel a. b In periods of low NAO the NAC is widened and the westward branching of currents is increased, including a noticeable increase in the Irminger Current.
Figure S7. Time series of mean temperature at Fylla Bank. a Map of Greenland showing (as red dotted lines) some of the standard oceanographic sections taken annually in early summer. The data from the section shown in panel b is station ‘Fylla2’ (near Nuuk). b Temperature collected at Fylla2 over depths 0–40 m during the middle of June for the period 1950–2007. The red curve is the 3-year running mean value, and the purple curve extends the time series back to 1875 using Smed’s climatology. While these are direct measurement of the surface temperature, these waters are nonetheless modified by Irminger Water en route to Fylla Bank, thus partly reflecting the temperature of the Irminger Water as originating in the North Atlantic.
Figure S8. Northward advance of cod along West Greenland in the early 20th century, during a previous warming period\textsuperscript{521}. During the 1920s and 1930s there was a dramatic warming of the North Atlantic Ocean. Warmer-than-normal sea temperatures and enhanced Atlantic inflow in the northern oceans continued through to the 1950s and 1960s. Ecosystem changes associated with the warm period included a general northward movement of fish. The maximum recorded movement involved cod, which spread approximately 1200 km northward along West Greenland, their progression marked by the red lines along the coast.