Warming of the Polar Water Layer in Disko Bay and Potential Impact on Jakobshavn Isbrae

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ABSTRACT

A number of recent studies have shown enhanced retreat of tidewater glaciers over much of southern and western Greenland. One of the fastest retreats has occurred at Jakobshavn Isbrae, with the rapid retreat linked to the arrival of relatively warm and saline Irminger water along the west coast of Greenland. Similar links to changes in ocean water masses on the coastal shelf of Greenland were also seen on the east coast. This study presents hydrographic data from Disko Bay, additionally revealing that there was also a significant warming of the cold polar water entering Disko Bay from the mid-to-late 1990s onward. This layer, which lies at a depth of ~30–200 m, warmed by 1°–2°C. The heat content of the polar water layer increased by a factor of 3.6 for the post-1997 period compared to the period prior to 1990. The heat content in the west Greenland Irminger water layer between the same periods increased only by a factor of 2, but contained more total heat. The authors suggest that the changes in the polar water layer are related to circulation changes in Baffin Bay.

1. Introduction

The high latitudes have been warming recently, at a rate faster than the rest of the globe. Several studies have examined changes in the Greenland Ice Sheet (GrIS) with the general conclusion that it is losing mass (Box et al. 2006; Hanna et al. 2008; van den Broeke et al. 2010). Church et al. (2011) reported that the observed sea level rise from 1972 to 2008 was $1.8 \pm 0.2\,\mathrm{mm\,yr^{-1}}$ based on tide gauges alone and $2.1 \pm 0.2\,\mathrm{mm\,yr^{-1}}$ when altimeter observations were included. They also estimated that $0.12 \pm 0.17\,\mathrm{mm\,yr^{-1}}$ of that rise came from the Greenland Ice Sheet, with the Greenland contribution increasing to $0.31 \pm 0.17\,\mathrm{mm\,yr^{-1}}$ over 1993–2008. The International Panel of Climate Change (IPCC) Fourth Assessment Report (AR4) projected that Greenland could contribute 1–12 cm to sea level rise by the end of the twenty-first century, while Graversen et al. (2011) estimated a contribution of 0–17 cm. The provision of freshwater from any such melt may also have major implications for the large-scale oceanic meridional overturning circulation and poleward heat transport (Stouffer et al. 2006; Smith and Gregory 2009; Hu et al. 2009).

A number of recent studies have focused on tidewater glaciers and how their retreat has quickened over much of southern and western Greenland (Krabill et al. 2004; Joughin et al. 2004; Howat et al. 2008; Rignot et al. 2010). At Jakobshavn Isbrae, which sits in a fjord extending off Disko Bay (Fig. 1), it was shown (Holland et al. 2008; Motyka et al. 2011) that the glacier rapidly started to thin and retreat, beginning in 1997 and it was inferred that these changes were synchronous with a significant input of warm Atlantic origin Irminger water (IW) to the glacier front. Three other glaciers in the Disko Bay area were shown, for a given summer, to have rates of submarine melting two orders of magnitude larger than surface melt rates (Rignot et al. 2010). Similarly, Straneo et al. (2010) found that warm Irminger waters in the Sermilik Fjord in eastern Greenland led to enhanced submarine melt rates. Johnson et al. (2011) found that the oceanic heat flux into the Petermann Fjord in northern Greenland is sufficient to explain the observed basal

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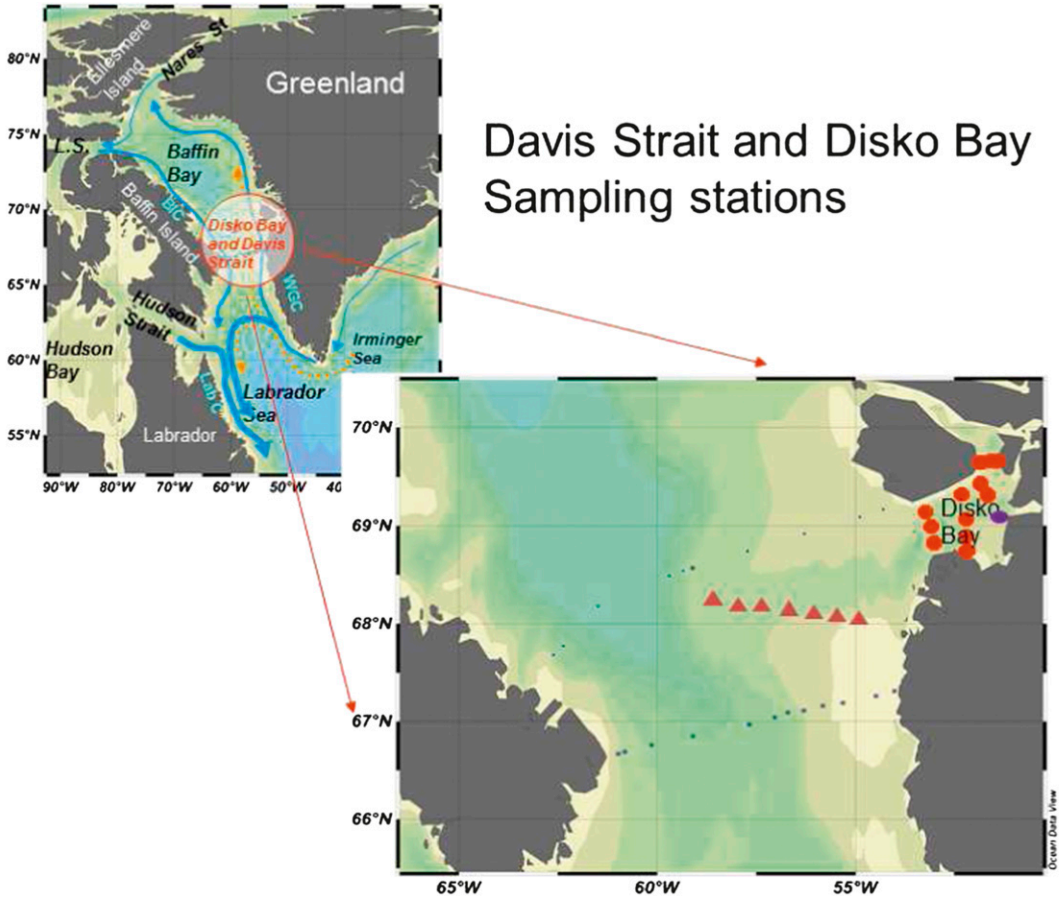

FIG. 1. Map of the study area, showing cold and fresh surface currents (solid blue lines) and warmer and saltier subsurface currents associated with the transport of Irminger Sea water (dotted orange lines). Abbreviations include the following: West Greenland Current (WGC), Baffin Island Current (BIC), and Labrador (LabC) and Lancaster Sound (L.S.). Red dots indicate the locations of the stations we use in Disko Bay, with Ilulissat st. 3, the closest to Jakobshavn Isbrae, indicated by the purple circle. The brown triangles indicate the locations of the stations along the Aasiaat section. This figure is based on and redrawn from one used by Azetsu-Scott et al. (2012) and plotted using the Ocean Data View software (Schlitzer 2007).

melting (Rignot and Steffen 2008). Mortensen et al. (2011) examined heat sources to Godthaabs Fjord in western Greenland. They (Mortensen et al. 2011) suggested a significant role of heat from the input of warm Irminger waters from the West Greenland Current, but also presented evidence for the existence of a local heat source caused by deep tidal mixing (Mortensen et al. 2011). A record of calving activity at Helheim Glacier in eastern Greenland was examined (Andresen et al. 2012) and showed a strong link between high calving activity and proxies for both Atlantic and polar water influence on the offshore shelf.

Greenland is bounded to the west by two major water bodies: the Labrador Sea to the south of Davis Strait and Baffin Bay to the north of Davis Strait (Fig. 1). The Labrador Sea circulation is cyclonic, with the West Greenland Current as the boundary current along the Greenland coast. Although the West Greenland Current is dynamically represented as a jet with a single core at (or near) the shelf break, it carries two water masses (Fratantoni and Pickart 2007). These include a cold and fresh component on the shelf as well as warmer and saltier Irminger water offshore (Fratantoni and Pickart 2007). The majority of the Irminger water, as well as the cold and fresh shelf water, is exchanged offshore into the Labrador Sea (Myers et al. 2009) but some continues north to flow through Davis Strait into Baffin Bay. Based on 1 year of data from mooring array at Davis Strait, Curry et al. (2011) observe the northward flow of both west Greenland slope water (WGSW; where temperature $\theta < 7°C$ and salinity $S < 34.1$) as well as west Greenland Irminger water (WGIW; where $\theta > 2°C$ and $S > 34.1$), which is a cooler and fresher product of the Irminger water formed in the Irminger Sea (Brambilla and Talley 2008; Thierry et al. 2008) that has been mixed and diluted along its path to Baffin Bay.

Within Baffin Bay, the circulation is cyclonic. Thus, the water masses that enter at Davis Strait continue north with the West Greenland Current along the west coast of Greenland (Tang et al. 2004). To the north, Arctic water enters, mainly through Nares Strait and Lancaster Sound. Being of Arctic origin, these waters are cold and fresh, and as discussed by Tang et al. (2004) can be called Arctic, cold, or polar water. We shall use the term polar water (PW) throughout this paper. These waters ($\theta < 1$ and $S \leq 33.7$) flow south in the Baffin Island Current before leaving Baffin Bay through Davis Strait (Tang et al. 2004; Curry et al. 2011).

Disko Bay is a large semienclosed region on the west coast of Greenland and north of Davis Strait (Fig. 1). A broad mouth connects Disko Bay with Baffin Bay and the West Greenland Current. This includes a deep channel that leads from the center of the bay to the shelf break offshore. There is also a narrow and shallower connection to the north through the Vaigat. The circulation in Disko Bay is not that of a traditional estuarine circulation, but instead involves inflow on the southern side of the bay and outflow to the north (Hansen et al. 2012).

2. Methods

Data were collected along a number of sections during ''once a year'' annual summer cruises conducted by the Greenland Institute of Natural Resources and processed by the Danish Meteorological Institute. Depending on the year, measurements were taken as early as 24 June and as late as 26 August. The observations are documented in a series of annual reports (Ribergaard 2012) that detail the data handling, which were based on CTD data post-1989 and bottle data prior to this year. In addition to measurement error, errors occur for bottle data because of the angle on the wire. The measurement being recorded may be shallower than the real depth. This error is not assumed to be large and definitely cannot explain the change we will discuss in polar water temperatures as this is the coldest water in the entire water column. The older bottle data are expected to be accurate to within 0.1°C, with an accuracy of at least 0.01 for the CTD data. Further details are provided in an analysis of similar data for the West Greenland Current (Myers et al. 2009).

Within Disko Bay, the data were collected along four sections repeated most summers, with each having either three or four stations (Fig. 1). Because we do not have repeat profiles within a year, we do not carry out any time averaging. We show profiles for the station Ilullisat st. 3 (350 m deep), which is the station closest to Jakobshavn Isbrae. In addition, for each year we average all the profiles within Disko Bay forming one single ''profile'' (minimum 1 and maximum 13 stations used in a given year). Doing this, we are able to monitor waters found at greater depths down to 700 m. In addition by averaging we get a more robust representation of Disko Bay properties, as we reduce the noise due to, for example, a local eddy, frontal, or vertical tidal movement.

Ocean heat content HC measures the amount of heat within a volume of water compared to some reference temperature, normally 0°C. Mathematically, $\mathrm{HC} = \rho C_p \int_A \int_z^0 (T - T_{\mathrm{ref}})\, dz\, dA$, where HC is the heat content, ρ is the density, C_p is the heat capacity (computed as a function of temperature, salinity, and pressure), T is the observed temperature, T_{ref} is the reference temperature (here, 0°C), z is the vertical depth coordinate, and A is the area the calculation is performed over. We perform our calculations over 2-m bins in the vertical (linearly interpolating to this resolution when using sparser bottle data) and giving our results per square meter area. To estimate uncertainty, we use a Monte Carlo approach with 100 draws, assuming a normal distribution with widths of 0.1 for salinity and 0.2°C for temperature.

Additionally, to understand how changes in Baffin Bay circulation are linked to polar water temperature changes in Disko Bay, we examine f/H and Ertel's potential vorticity PV. Our calculations of f/H are made using the global 5-min (Smith and Sandwell 1997) bathymetry dataset, with H being the ocean depth. Ertel's potential vorticity is computed as $\mathrm{PV} = 1/\rho_0(f + \zeta)\rho_z$, where f is the Coriolis parameter, ζ is the relative vorticity, ρ_z is the vertical density gradient, and ρ_0 is a reference density (1000 kg m^{-3}).

We compute Ertel's potential vorticity based upon output from a 1/12° experiment using the Ocean Circulation and Climate Advanced Modelling Project (OCCAM) global general circulation model, ran over 1988–2006 (Aksenov et al. 2010). We use output from OCCAM given its high resolution, the fact it covers the time period in question, and because it has been used successfully in a number of high-latitude studies (e.g., Aksenov et al. 2010). Figure 2b in Aksenov et al. (2010) compares model salinity and velocities at Davis Strait from April 2004 to August 2005 with the mooring data of Curry et al. (2011) and shows some agreement in terms of salinity and velocity, although discrepancies include the model being everywhere too saline, and the modeled West Greenland Current being too baroclinic [a common issue in these types of models Penduff et al. (2005)] and possibly too weak. In Fig. 8 of Aksenov et al. (2010), the model's surface circulation in Baffin Bay shows the expected cyclonic circulation with greater velocities in

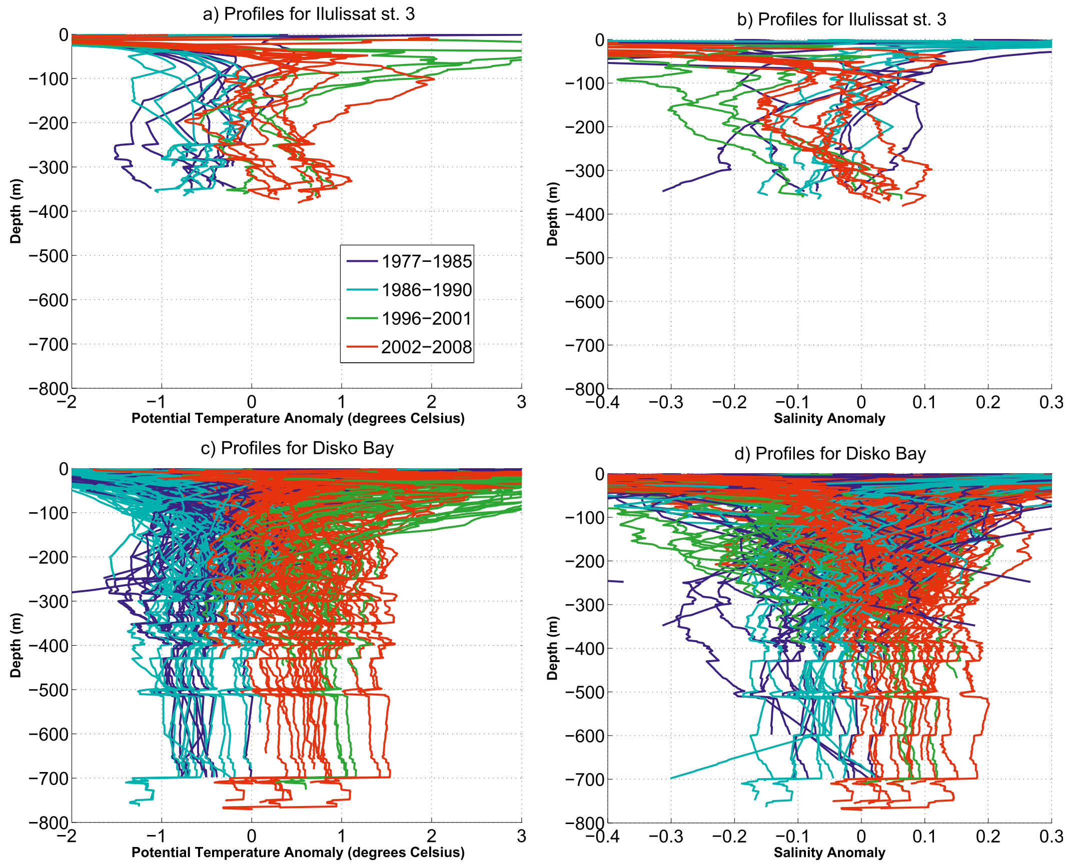

FIG. 2. (a),(c) Profiles of potential temperature anomalies (°C) and (b),(d) salinity anomalies for (top) all profiles available for Station Ilulissat st. 3, the closest to Jakobshavn Isbrae, and (bottom) all available profiles in Disko Bay. Anomalies are computed from the mean, at each depth, of all available profiles for the station(s) in question over 1977–2008. The station locations are shown in Fig. 1. Blue profiles are taken over 1977–85, cyan profiles are taken over 1986–90, green profiles are taken over 1996–2001, and red profiles are taken over 2002–08.

the Baffin Island Current (Tang et al. 2004). The scarcity of observations in Baffin Bay suggest that, beyond the basic comparisons discussed in Aksenov et al. (2010), OCCAM has not been fully validated against observations in this region. However, this lack of observations for validation is why we are forced to use the model fields if we wish to make computations of Ertel's potential vorticity in this region.

3. Results

The vertical structure in Disko Bay involves at least three layers or water masses (Figs. 2 and 3). At the surface there is a thin upper layer, which is very fresh from glacial melt and runoff that also undergoes significant seasonal temperature changes (Hansen et al. 2012). Polar water is found in a cold intermediate layer between ~30 and 200 m. At depth, the bay is filled with warmer and saltier waters advected into the bay, associated with the mixing of Atlantic-origin waters with the colder and fresher waters of Arctic Ocean origin modified by local runoff (Buch 2000). These are the Irminger waters discussed by Holland et al. (2008) and defined as WGIW ($\theta > 2°C$ and $S > 34.1$) by Curry et al. (2011).

Following previous work (Holland et al. 2008; Motyka et al. 2011), we examine the change in water properties pre- and post-1997. Breaking down the data into four time periods and examining anomalies compared to the 1977–2008 mean, the profiles (Fig. 2) show no clear signal at the surface. Below that thin ~30-m layer, significant warming with time can be seen with a shift from negative to positive anomalies. There is a significant increase in salinities below ~250–300 m post-1997 but in the 100–200-m layer, the salinities initially decrease post-1997 before increasing in the 2000s. These changes in the deeper layers of Disko Bay are consistent with observation of Holland et al. (2008) of an inflow of Irminger water into the region from 1997 onward.

An unfortunate gap in the observational record during the first part of the 1990s highlights the significant

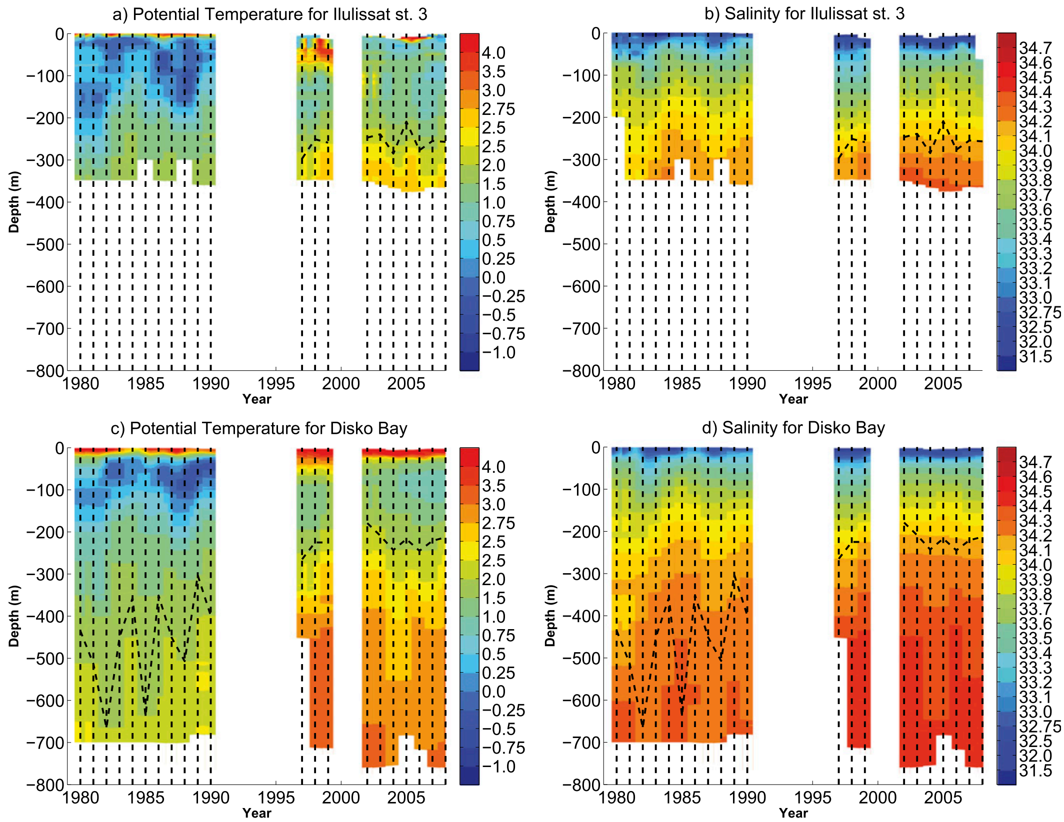

FIG. 3. Hovmöller plots of (a),(c) potential temperature (°C) and (b),(d) salinity for (top) Station Ilulissat st. 3, the closest to Jakobshavn Isbrae and (bottom) averaged over all of Disko Bay to show more detail of the water structure deeper in the water column. The station locations are shown on Fig. 1. The vertical dashed black lines show the years when observations were taken. Note that color bar scales are nonlinear for both fields. The horizontal black dashed line shows the upper boundary of WGIW in each year that it is present.

change in water properties in Disko Bay before and after the gap (i.e., pre- and post-1997; Fig. 3). We note that Motyka et al. (2011) present data from the mid-1990s in Disko Bay, but we have been unable to access that data and include them in our analysis. Examining Fig. 5 of Motyka et al. (2011) suggests that conditions during the mid-1990s were similar to the earlier period and it is only post-1996/97 that the significant changes discussed below occurred. This change in IW presence in 1996/97 was also reported by Holland et al. (2008) using trawl bottom temperature measurements over the west Greenland shelf upstream of Disko Bay.

During the entire time period under examination WGIW was present within Disko Bay (Figs. 3c,d), with the depth of the interface between this water mass and the cooler and fresher water above slowly rising through the decade of the 1980s. However, prior to 1997 (or at least 1990 from our data) WGIW was present at depths deeper than the water depth at Ilulissat st. 3. However, even though WGIW was present in Disko Bay prior to 1997, the profound changes in this water mass first pointed out by Holland et al. (2008) are clearly seen here, with the WGIW getting significantly warmer and saltier and occupying a much larger volume of Disko Bay, with its upper interface much closer to the surface.

The change in temperature from the 1980s to the late 1990s and 2000s is also very significant for the polar water (Fig. 3). Prior to 1990, and during the mid-1990s according to Motyka et al. (2011), this layer had a core temperature of below 0°C, when this isotherm in some years reached down to depths close to 200 m. In more recent years, water below 0°C is not seen in the observations. In fact, we do not use dashed lines to indicate the boundaries of this water mass in Fig. 3 like we do for the WGIW because no water satisfying the definition of polar water is present in Disko Bay in most years post-1997.

Such significant changes in temperature will have major impacts on oceanic heat content (Fig. 4). Disko Bay experienced more than a doubling of heat content (relative to a reference temperature of 0°C, the melting temperature of glacial ice) per square meter of surface

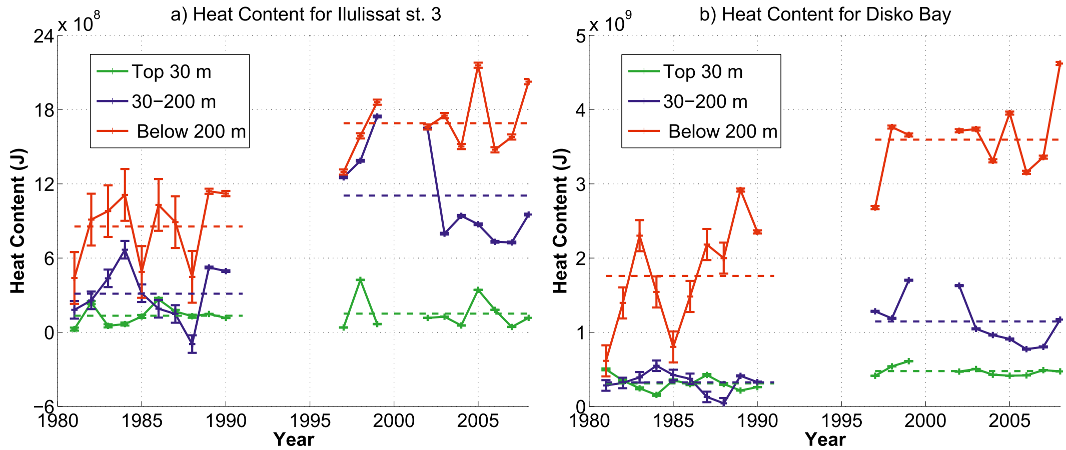

FIG. 4. Time series of HC, per square meter of ocean surface area. The results are averaged over three layers, the top 30, 30–200, and 200–700 m (or the bottom of the profile if shallower than 700 m). See the methods sections for details on how these are computed. (a) Station Ilulissat st. 3, the closest to Jakobshavn Isbrae, and (b) an average profile composed of all available profiles in Disko Bay are shown. The station locations are shown on Fig. 1. The horizontal dashed lines give the averages, for each layer, over 1981–90 and 1997–2008.

area post-1997 compared to years before 1990. The highest heat content occurs in the warm and thick WGIW layer below 200 m. However, in terms of change before and after 1997, the largest increase post-1997 relative to the prior period is in the polar water between 30 and 200 m. To be precise, the heat content in the polar water layer increases by a factor of 3.6 (from 0.32×10^9 to 1.15×10^9 J) compared to an increase by a factor of 1.8–2.0 in the west Greenland Irminger water layer (from 1.76×10^9 to 3.60×10^9 J considering the average Disko Bay profile integrated over the 200–700-m layer, or from 0.81×10^9 to 1.48×10^9 J considering the Ilulissat st. 3 profile integrated over the 200–350-m layer).

4. Discussion

We present evidence that Disko Bay prior to 1990 was relatively cold and fresh, with significant amounts of polar water present at depths between 30 and 200 m. We also present evidence that WGIW was present in Disko Bay prior to 1997, although it was fairly cold and less saline than in later periods. The upper interface of the WGIW was relatively deep during the 1980s, generally below 400 m, rising to almost 200 m after 1997.

Our analysis supports the contention (Holland et al. 2008) that there was a significant influx of oceanic heat around 1997 that may have played an important role in the acceleration of Jakobshavn Isbrae. We show that the WGIW becomes significantly warmer and saltier from 1997 onward. It also occupies more of the Disko Bay, with its interface rising more than 200 m upward (to ~200 m), which leads to easier access to Jakobshavn Isbrae given that the WGIW was thus found above sill depth to the glacial fjord. Using multibeam bathymetry, Schumann et al. (2012) showed that the Ilulissat ice fjord is separated from Disko Bay by a shallow sill of less than 200-m depth, with the deepest spot being a narrow canyon ~250 m deep. The sill depth is sufficient for the easy exchange of polar waters into the fjord. However, WGIW has only been shallow enough for easy exchange over the sill since 1997. It is also possible for deeper WGIW to cross over a shallow sill, such as the observed episodic winter intrusions into Godthaabs Fjord (Mortensen et al. 2011) farther south in west Greenland. Thus, our results go beyond previous analyses (Holland et al. 2008) in indicating that the enhancement in heat content within Disko Bay was not just associated with the changes in WGIW, as up to ~1/3 of the enhanced summer heat content came from a warming/disappearance of the intermediate-depth polar water layer.

It was argued (Buch 2000) that the presence of this very cold intermediate layer in Disko Bay may either indicate vertical convection due to winter cooling or that the cold water was coming from the Baffin Island Current and transported across Baffin Bay into Disko Bay. But as previously discussed (Holland et al. 2008), there was no sudden change in air temperature in the 1990s that could explain the rapid warming in this layer. Furthermore, the deepest convection normally occurs during winter, when the air temperatures over Disko Bay are still very cold. Hansen et al. (2012) report that the sea ice cover over Disko Bay has reduced by around 50%. A reduction of sea ice cover would suggest a larger heat loss to the atmosphere, an increased likelihood of winter convection, and the likely persistence of cold temperatures from newly ventilated winter waters through

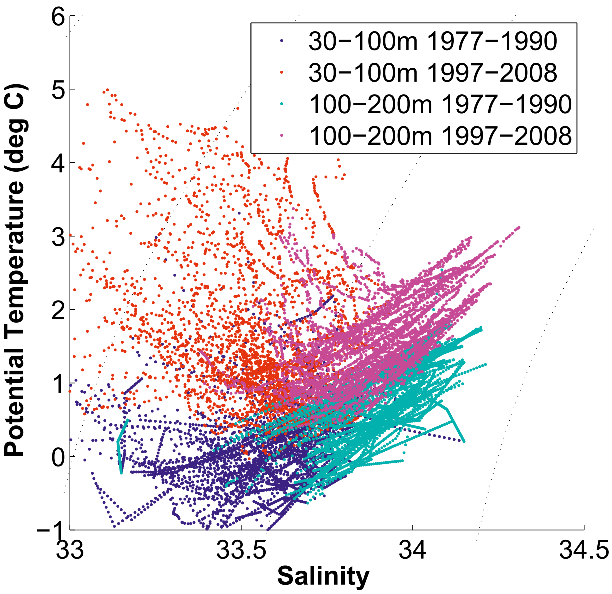

FIG. 5. Diagram of σ–S for all observations within Disko Bay in the 30–200-m-depth range. Observations shallower than 100 m taken before 1996 are shown in blue, while those shallower than 100 m taken from 1996 onward are shown in cyan. Observations deeper than 100 m taken before 1996 are shown in red, while those deeper than 100 m taken from 1996 onward are shown in maroon.

to the next summer. Thus, it is unlikely that a change/decrease in convection can explain the warming/disappearance of polar water within Disko Bay.

The only other potential source of heat within Disko Bay to warm the polar water is the warmer Irminger water below it. Mixing, whether by internal waves/tides, instabilities driven by flows over the sills, overturning after coastal upwelling processes, or other vertical exchange processes may contribute to diapycnal mixing and the warming of the polar water from below. However, given that the temperature anomalies in the 1990s and 2000s in the polar water depth range increase from 200 m upward (Figs. 2a,c), rather than being largest at the interface between the polar water and the WGIW layer, suggests that warming from below is not the main driver of our observed changes. A temperature–salinity diagram (Fig. 5) for all measurements in the 30–200-m layer also shows that this warming is not accompanied by any significant salinity change. Given that the WGIW is much saltier than the polar water, significant mixing between these water masses would require an increase in salinity in the polar water layer, which is not observed during the first years after 1997 (Figs. 2c,d).

Therefore, we believe that the changes to the polar water around 1997 are not local to Disko Bay and are more likely related to circulation changes in Baffin Bay. Disko Bay has a cyclonic circulation with inflow on its southern side (Sloth and Buch 1984; Buch 2000; Hansen et al. 2012) and is thus fed by the West Greenland Current. Below zero temperatures are seen flowing northward through Davis Strait on the west Greenland shelf during winter but are confined to the top ~100 m and are associated with low salinities (Curry et al. 2011). However, temperatures below −1°C are not seen at the Maniitsoq section at the southern edge of the Davis Strait during early summer [not shown; see figures in Myers et al. (2009) and Ribergaard (2012)], showing that the extremely cold polar water is not just a continuation of polar water flowing north into and through Davis Strait.

Cold polar water is carried south in the Baffin Island Current, formed by mixing and sinking in the North Water polynya region of northern Baffin Bay, and mixed with Arctic water flowing out of Lancaster Sound (Lobb et al. 2003; Rudels 2011). This Arctic-origin polar water has a low temperature (~−1.6°C; Rudels 2011), extends to depths of 200–300 m, and is overlaid by a thin (25–50 m deep) warmer and fresher summer layer (Fissel et al. 1982; Rudels 1986). The Baffin Island Current flows south along the east coast of Baffin Island, with the core of the current steadily following the continental slope, although some meanders occur offshore near the base of the slope (Fissel et al. 1982). There is then a broad outflow to the south through Davis Strait (Tang et al. 2004).

The question then is the exchange (and its variability) of polar water from the Baffin Island Current into the West Greenland Current north of Davis Strait to feed into Disko Bay. A schematic (Lobb et al. 2003) highlights that previous work suggested there is such exchange, which would be consistent with flow along lines of f/H and conservation of potential vorticity (Fig. 6a), as first suggested by (Buch 2000).

But, as a function of latitude and topography, this does not explain the temporal differences. So we take into account the stratification and plot Ertel's PV from a high-resolution 1/12° numerical model (OCCAM). We examine two time periods, 1988–92 and 1997–2003, associated with different amounts of polar water being present in Disko Bay (Figs. 6b,c). In the later period, the southward-flowing polar water can continue to flow south, exiting Baffin Bay while conserving PV, and thus minimizing exchange to the east. Meanwhile, this is not the case in the earlier period, when there is more exchange and mixing, making it easier for the polar water to be transported toward Greenland. We note that the water masses in OCCAM are more saline than the observations and thus may impact the stratification and our calculation of Ertel's PV. Still, this calculation suggests that there may have been an evolution of water masses

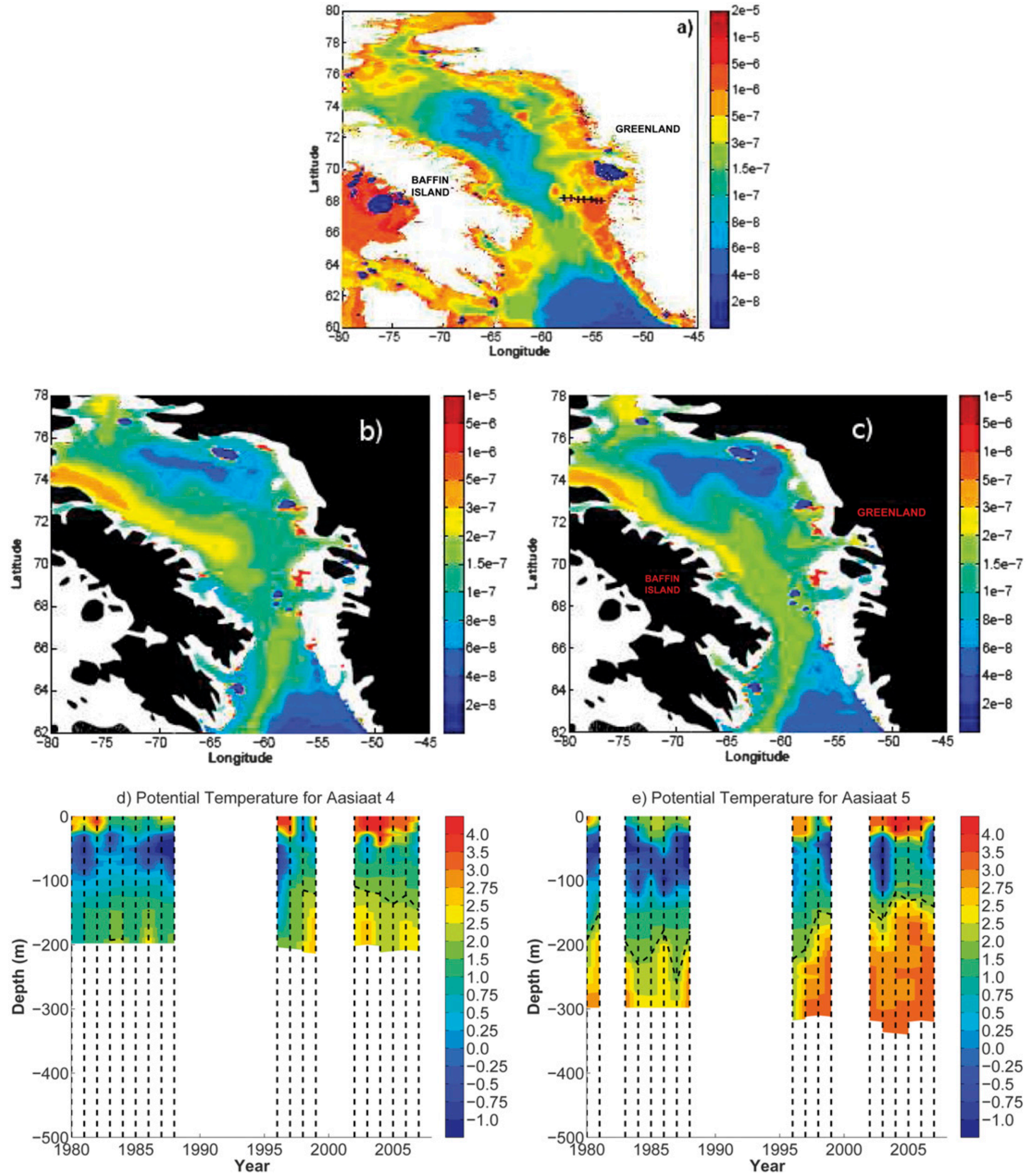

FIG. 6. (a) Contour plot of f/H using a nonlinear color scale and global bathymetry (Smith and Sandwell 1997). The "+" indicates the location of the stations along the Aasiaat section. (b),(c) Ertel's PV based on the 1/12° OCCAM general circulation model (see the methods section for further details on this computation) is shown, using a nonlinear color scale, for 1998–92 (b) and 1997–2003 (c). (d),(e) Hovmöller plots of potential temperature (°C) for the Aasiaat section from stations 4 and 5, respectively. The vertical dashed black lines show the years when observations were taken. Note that color bar scale is nonlinear. The horizontal black dashed line shows the upper boundary of WGIW in each year that it is present.

and circulation in southern Baffin Bay between the time periods in question.

To examine if such an evolution might be present in the observational record, we use data from the Aasiaat section (location given in Fig. 1) that crosses the West Greenland Current to the south of Disko Bay. First, a Hovmöller plot of temperature from station 4 (the middle station of the section—Fig. 6d) shows a similar evolution for the polar water as in Disko Bay, with a near disappearance of water below 0°C post-1997. Examining the next station offshore, whose profiles extend deeper into the water column (Fig. 6e) shows the arrival of the very warm and saline vintage of WGIW observed in Disko Bay, in 1997. Note that while data were lacking

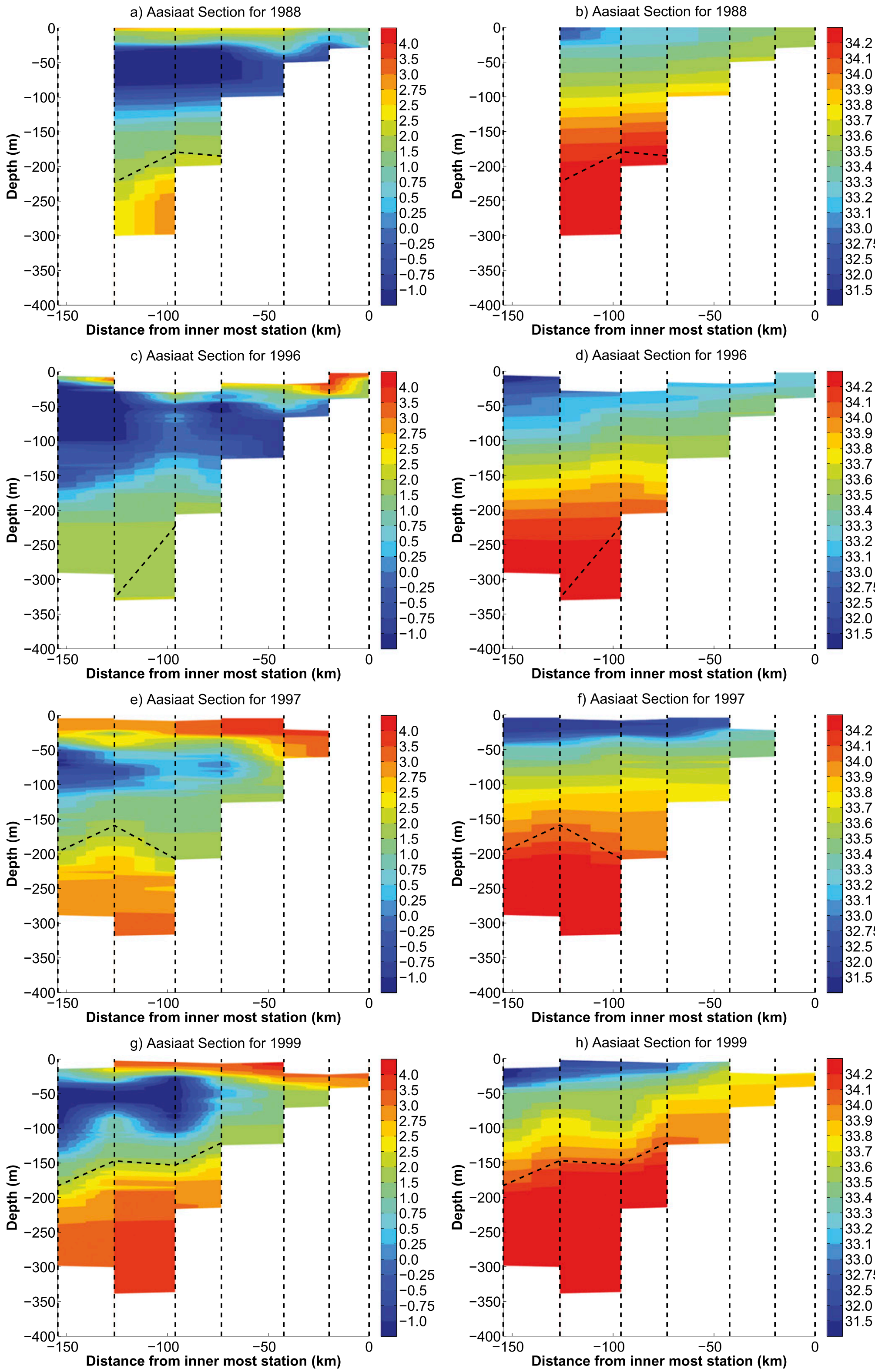

FIG. 7. Depth–distance plots for the Aasiaat section for the following years: (a),(b) 1988; (c),(d) 1996; (e),(f) 1997; and (g),(h) 1999, for potential temperature (°C; left) and salinity (right). The station locations are shown on Fig. 6a. Greenland is on the right. The vertical dashed black lines show the actual station locations. Note that color bar scales are nonlinear for both fields. The horizontal black dashed line shows the upper boundary of WGIW at each station where it is present.

for Disko Bay in 1996, this plot shows that this very warm and saline WGIW is not present high up in the water column at this station in 1996. Thus, the arrival of this vintage of WGIW in this region must have occurred between the summers of 1996 and 1997.

In Fig. 7 we present data from four individual years before (1988 and 1996), during (1997), and after (1999) the appearance of the very warm and saline vintage of WGIW at this location (and in Disko Bay). Although our data are summer snapshots that are not representative of the rest of the year, we show these sections to highlight the tremendous interannual change and variability occurring outside of Disko Bay.

Figures 7a and 7b show properties from the summer of 1988, with polar water extending all the way from the Baffin Island Current across onto the west Greenland shelf. WGIW can be seen as warmer water below 150 m. Fresh surface water being transported south by the Baffin Island Current can also be seen in the top 30 m at the westernmost 3 stations. In the summer of 1996 (Figs. 7c,d), the polar water core has thickened, although there is a region of warmer water around station 4 in the middle of the section. The WGIW that is present is barely above the temperature minimum for the definition of this water mass. Additionally, a thicker layer of low-salinity water can now be seen across the entire section at the surface, reaching ~50–100-m depth at the station farthest offshore from Greenland. A year later, in the summer of 1997 (Figs. 7e,f), the polar water has withdrawn to the west, no longer extending to the west Greenland shelf. A thick (~50-m thick) layer of very low-salinity (<33) water now extends across the entire section onto the west Greenland shelf. The WGIW is still found below 150–200 m, but is warmer and saltier than in previous years. In the summer of 1999 (Figs. 7g,h), the polar water remains confined to the stations on the Baffin Island side, while the fresh surface layer has withdrawn from the shallowest parts of the west Greenland shelf. The WGIW has continued to warm and occupy more of the section, with its upper interface reaching 100 m.

Given that we see significant changes in both the fresh surface layer, as well as with the polar water on the western side of the Aasiaat section, we speculate that this variability may be governed by upstream changes in Arctic water discharge through the Canadian Arctic Archipelago (CAA). There are no available transport measurements available within the CAA for this period. A number of studies have pointed out that the volume and freshwater transports through the CAA are linked to the dynamic height or sea surface height gradients across the archipelago (Kliem and Greenberg 2003; McGeehan and Maslowski 2012; Houssais and Herbaut 2011). Kliem and Greenberg (2003) and McGeehan and Maslowski (2012) show that the gradient is impacted as much by changes in Baffin Bay as in the Arctic Ocean. Based on their modeling study, McGeehan and Maslowski (2012) go on to suggest that one could also relate and monitor the CAA flow by measuring the sea surface height gradient between northern Baffin Bay and eastern Davis Strait. They (McGeehan and Maslowski 2012) also show from their simulations that there was a significant drop in their model sea surface height gradient anomaly between northern Baffin Bay and Davis Strait from high values in the early–mid-1990s to a minimum around 1999 (with the sign of the gradient anomaly flipping around 1997). They (McGeehan and Maslowski 2012) were unsure of the cause of this rapid decline, finding little correlation with the volume flux anomalies in their model and the major atmospheric modes (Arctic and North Atlantic Oscillations). Given that McGeehan and Maslowski (2012) find their gradient anomaly most sensitive to sea surface height changes at the eastern end of Davis Strait, maybe the significant lifting of the heights at this time was related to the influx of the warm Irminger water. This is speculation and is a question that needs further study (probably with numerical models because of the lack of available data), as well as the question of how changes in the circulation might feed into the provision of different water masses to Disko Bay.

Although WGIW was present in both Baffin and Disko Bays prior to 1997, there was a significant influx of this water mass into the region in 1997. We see this through the increase in temperature and salinity, as well as the lifting of the upper interface of this water mass. This influx is related to changes observed in the subpolar gyre outlined in numerous other studies (e.g., Thierry et al. 2008; Myers et al. 2007) and has been previously highlighted (Holland et al. 2008). Given the significant changes in polar water occurring at the same time, it suggests that the oceanic changes impacting on Jakobshavn Isbrae are related to an interplay and coupling of Arctic and subpolar processes. Consequently, a proper understanding of the underlying dynamics to allow for future predictability will require hemispheric models/studies, not just regional.

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