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Greenland Melt and the Atlantic Meridional Overturning Circulation

By Eleanor Frajka-Williams, Jonathan L. Bamber, and Kjetil Våge

Scientists and crew aboard the research vessel *Knorr* faced winds ranging from 60 knots up to 100 knots and 10 m to 12 m tall waves on an expedition to the Irminger Sea in October 2008. *Photo credit: Kjetil Våge*



“...the ocean’s response to varied forcings is complicated, occurring on a range of time scales and mediated by interactions with the cryosphere and atmosphere. Several possible responses (and time lags) have been identified through modeling or have been explained by theory, suggesting that observed circulation changes may result from several competing influences and from a range of past events.”

ABSTRACT. More than a decade of observations of the meridional overturning circulation in the subtropical North Atlantic show it to be highly variable on time scales of days to years and with an overall trend toward slowing down. Over the same time period, melting from Greenland (and elsewhere in the Arctic, including from sea ice) has been increasing, resulting in greater freshwater input to the northern North Atlantic. In this article, we examine evidence for the impact, if any, of this influx of freshwater on the large-scale ocean circulation and for potential changes.

INTRODUCTION

The Atlantic meridional overturning circulation (MOC) plays a key role in the climate system, transporting energy (e.g., heat) and matter (e.g., carbon and a variety of other substances) around the ocean. The overturning circulation in the Atlantic is composed of northward-flowing warm waters in the top 1,000 m that transport heat from near the equator toward the northern high latitudes, where it is then released to the atmosphere (Figure 1; Trenberth and Caron, 2001). In contrast, the southward-flowing deep limb is isolated from atmospheric ventilation and thus stores energy and matter for hundreds of years. In numerical modeling scenarios where the MOC is abruptly halted, anomalous patterns of temperature variability are seen across the globe, including dramatic cooling over Europe (e.g., Manabe and Stouffer, 1988; Jackson et al., 2015). Because the MOC has the potential to switch to a qualitatively different state (a reduction or shutdown) as a consequence of relatively small perturbations, it is considered

to be a potential tipping element in the Earth system (Lenton et al., 2008), with profound implications for Europe’s climate (Levermann et al., 2012).

One mechanism by which the MOC can be altered is through a change in stratification at high latitudes. Dense waters moving southward in the MOC’s lower layer are formed at high latitudes in the subpolar regions. Water near the surface becomes denser as it loses heat to the atmosphere or becomes saltier due to evaporation or ice formation. The dense waters sink and mix with the underlying waters to form deep layers of homogeneous water with intermediate temperatures and salinities. Increases in freshwater input at northern latitudes from precipitation and/or melting of ice decrease surface seawater density and increase stratification, slowing down or halting convection.

In the North Atlantic, dense water formation occurs through deep convection both in the open ocean in the subpolar gyre (Labrador and Irminger Seas) and in marginal seas to the north of Iceland

(whose waters enter the North Atlantic by flowing over relatively shallow sills). In the subpolar gyre, Labrador Sea Water (LSW), a class of dense, well-mixed water in the depth ranges of 1,000–2,500 m, is the product of deep convection. The overflow waters are typically deeper. While dense water from higher latitudes contributes to the southward-flowing MOC (Gebbie and Huybers, 2010), the sills may control the rate of dense water export, reducing its influence. Here, we focus on a more direct pathway by which freshwater from Greenland is entrained into the MOC—through open-ocean convection in the Labrador Sea region.

Paleoclimate records reveal a possible relationship between freshwater fluxes from the advance and retreat of ice sheets and abrupt shifts in climate. Near the end of the last ice age (8,200 years ago), glacial Lake Agassiz over North America drained nearly completely into Hudson Bay (e.g., Barber et al., 1999). This event flooded the North Atlantic with freshwater, and is believed to have resulted in a temporary shutdown of the Atlantic MOC and its northward heat transport. Temperature proxies show dramatic cooling over Greenland associated with the lake outburst. Numerical simulations called freshwater hosing experiments (e.g., Manabe and Stouffer, 1988; Gerdes et al., 2006) show how large inputs of freshwater can shift the state of the

Atlantic MOC toward a reduced (slowing down) or “off” state.

While such extreme events as the Lake Agassiz drainage in the paleoclimate record provide evidence that freshwater can influence the large-scale ocean circulation, they occurred following an ice age (rather than a warm period, which we are in at present) and involved enormous volumes of freshwater (200,000 km³; Barber et al. 1999)—an order of magnitude larger than the fluxes anticipated from present-day melting of the Greenland Ice Sheet (GrIS) or the Arctic (Curry and Mauritzen, 2005). Are the present-day changes in freshwater forcing too small to have an observable influence on ocean circulation? What magnitude of forcing is required to significantly influence the Atlantic MOC, and what other factors influence its strength?

We explore these questions by examining the evidence for recent variability in the MOC from both model simulations and observational data and consider whether melting of the GrIS could be (in part) responsible for this variability. Rather than providing a complete review of the literature over this broad range of topics, we point the reader to additional reviews in the relevant sections. In the next section, we consider evidence that freshwater from the GrIS makes its way to regions of deep convection and discuss

known influences on deep convection intensity, including freshwater, the atmosphere, and ice, and the idea that the deep water formed through convection is directly linked to the MOC’s strength. We then discuss updated time series of the strength of the Atlantic MOC and freshwater fluxes from Greenland, and consider a possible link between deepwater formation and the Atlantic MOC. Finally, we speculate that a link may yet be found between these varied processes.

INFLUENCES ON CONVECTION AND THE MOC

Freshwater (and Other) Influences on Convection

Freshwater input is one of several competing controls on the strength of convection and the density of waters formed. Convection occurs when surface waters lose heat to the atmosphere, become more dense, and sink, eroding surface stratification. Preconditioning—or the absence of strong stratification due to cyclonic circulation with doming isopycnals—can make a region more susceptible to convection. In contrast, buoyant surface waters from freshwater inputs can suppress convection.

In the Labrador Sea, surface heat fluxes tend to be very high (at times, exceeding 1,000 W m⁻²), with the strongest fluxes occurring in the southwest near the shelf

edge (Moore et al., 2014). These strong surface heat fluxes are a result of wintertime storms blowing cold dry air from Canada eastward over the ocean and the topography of Greenland (Moore et al., 2014; Schulze et al., 2016). Atmospheric circulation is steered by the Greenland landmass, which can generate narrow regions of intense winds called “tip jets” over the Irminger Sea (Doyle and Shapiro, 1999; Moore and Renfrew, 2005). These tip jets drive intense air-sea heat flux and have been shown to induce convection (Pickart et al., 2003; Våge et al., 2008). Sea ice distribution also modulates ocean heat fluxes. For example, during the winter of 2007/08, extensive sea ice on the western side of the Labrador Sea insulated the ocean from the atmosphere over the shelf, resulting in strong heat losses immediately downstream of the ice edge (Våge et al., 2009). The oceanic heat loss during that year was particularly strong over the open ocean, leading to the return of strong convection after several years of reduced convection.

Preconditioning can vary on interannual and decadal time scales. For example, during the decade following 1994, advection of buoyant waters from surrounding areas caused the top 2,000 m of water in the Labrador Sea to gradually restratify. The increase in stratification resists deep convection the following year because of the large amount of heat that must be removed from the buoyant surface waters by air-sea fluxes before they can sink to a particular depth. In contrast, after several years of anomalously intense air-sea fluxes, a weakly stratified surface layer may cap a well-mixed bolus of recently formed LSW. As a consequence, in the next year, the air-sea fluxes need only erode that weakly stratified surface layer before reaching the previous year’s deepwater layer. Decadal variations in temperature and salinity in the subpolar gyre can lead to more (less) buoyancy near the surface, resulting in weaker (stronger) convection (van Aken et al., 2011). Due to the range of influences on convection—intensity of air sea fluxes,



FIGURE 1. The Atlantic meridional overturning circulation (MOC). Red lines show warm currents near the surface; blue indicates deeper currents. The blue circles mark major regions of deep convection in the Labrador (west) and Irminger (east) Seas. The dashed blue line indicates overflow waters formed in the Greenland-Iceland-Norwegian Seas, north of Iceland. Dashed red lines show the direction of ocean gyres.

preconditioning, sea ice extent—it is difficult to isolate the influence of freshwater alone on deepwater formation.

Freshwater input from precipitation or glacial meltwater reduces deepwater formation by increasing the buoyancy of the surface layer. Although freshwater is one of several influences on deep convection, more recent freshening periods, in addition to paleoclimate records, suggest that freshwater input can strongly reduce or shut down the MOC (Lenton et al., 2008). A notable event of the 1960s to 1972, known as the “Great Salinity Anomaly,” was characterized by a net freshening in the top 800 m of the subpolar gyre (Dickson et al., 1988). In total, about 10,000 km³ of freshwater was released into the subpolar gyre, shutting down convection in the Labrador Sea for several years (maximal wintertime mixed layer depths were about 200 m; Lazier, 1980). During this period, the deep waters slowly increased in salinity and warmed until the winter of 1972 when deep convection resumed due to extreme surface heat fluxes associated with the most positive state of the North Atlantic Oscillation in decades (Dickson et al., 2000).

Pathways of Greenland Melt to Convection Regions

In hosing experiments, freshwater is typically distributed uniformly over the North Atlantic. For large volumes of water released rapidly (e.g., 200,000 km³ from the Laurentide Ice Sheet lakes at the end of the last ice age), this is likely a reasonable distribution of the freshwater input. In contrast, while the absolute freshwater flux from Greenland is presently large (>1,000 km³ yr⁻¹ since 1998; Bamber et al., 2012), the anomaly (i.e., the change in flux) is smaller, and freshwater entering the ocean in coastal waters does not necessarily invade key regions of convection (Böning et al., 2016; Luo et al., 2016). Hydrographic measurements from autonomous Argo floats show an increase in stratification in the Labrador Sea over the 2004–2012 period, with freshening in the top 200 m

but salinifying in the 200–1,000 m depth range (Schulze, 2016). Altogether, the top 1,000 m has only freshened slightly over the past decade, in spite of continued freshwater release from GrIS, suggesting that melt from the GrIS hasn’t yet reached the Labrador Sea central basin.

Several high-resolution simulations have investigated whether and how freshwater from the Arctic and Greenland reaches the central Labrador Sea, where it may suppress convection by increasing surface buoyancy. Using a ¼° numerical model, Myers (2005) simulated a tracer release from Baffin Bay, which connects to the Labrador Sea through Davis Strait at 67°N. In this simulation, the tracer—representing freshwater—did not enter the central Labrador Sea, showing the absence of a pathway from the Arctic through Baffin Bay to the Labrador Sea. In a second simulation, tracer was released at the southern tip of Greenland and did enter the central Labrador Sea. This simulation supports the expectation that eddy activity west of Greenland drives cross-shelf exchange. Using discrete particle trajectories, rather than a tracer dispersal, in an offline ½° global simulation, Schulze (2016) simulated freshwater pathways into the central Labrador Sea by tagging particles there and advecting them backward in time using the model velocities. The results show that particles in the central Labrador Sea did not originate in Baffin Bay, but instead, similar to the results of Myers (2005), made their way to the sea by way of the East Greenland Current. The timing of cross-shelf exchange diagnosed from the particles indicates that wind forcing, rather than eddy activity, plays a dominant role in the cross-shelf exchange. However, numerical investigations of freshwater exchange into the Labrador Sea can be sensitive to model choice and resolution (Dukhovskoy et al., 2016), resulting in uncertainty about the pathways of freshwater transport from Greenland and Labrador shelves into the central Labrador Sea.

To investigate the effect of Greenland meltwater on high-latitude convection, a

very high-resolution numerical simulation was forced with realistic GrIS melt patterns estimated from Böning et al. (2016). Böning et al. (2016) found that although the Labrador Sea is gradually freshening, the degree of freshening has not yet reduced the intensity of convection or the MOC. Instead, variations are consistent with variations in convection prior to applying freshwater fluxes. These authors conclude, however, that continued melting of the GrIS could dampen convection in the near future.

In contrast, Yang et al. (2016) suggest that freshening of surface waters in the Labrador Sea has already reduced convection, as evidenced by the reduction in LSW thickness. Their study shows that freshwater inputs to the Labrador Sea from GrIS mass loss (estimated from Gravity Recovery and Climate Experiment satellites, GRACE) and from Arctic sea ice melting (based on the annual minimum volume predicted by the Pan-Arctic Ice Ocean Modeling and Assimilation System – PIOMAS) exhibit a combined increase since the mid-1990s of about 20 mSv (1 Sv = 1 × 10⁶ m³ s⁻¹). They further estimate salt flux into the Labrador Sea from hydrographic sections along Southwest Greenland and argue that until the mid-1990s, the salt flux drove increases in convection intensity, as measured by LSW thickness that peaked at 1,300 m in 1994. Since the mid-1990s, increases in freshwater inputs have overwhelmed salt fluxes to reduce LSW thickness to a minimum of less than 600 m in 2013. However, in the 2013/14 and 2014/15 winters, while freshwater inputs from GrIS continued, deep convection returned in the Irminger Sea (de Jong and de Steur, 2016; Fröb et al., 2016) and in the Labrador Sea, in apparent contradiction with the findings of Yang et al. (2016).

Another study, conducted by Rahmstorf et al. (2015), considered the influence of Greenland melt on the MOC on decadal time scales, using a sea surface temperature (SST) index for the MOC and Greenland melt estimates from satellite and surface mass balance calculations.

These authors described the MOC slowdown from 1975 to 1995 as unprecedented ($p > 0.99$), and linked it to accumulated mass loss from GrIS between 1900 and 1970 (Box and Colgan, 2013). A further increase in GrIS melt since 2000 was related to the present reduction in the MOC (since 2010, in their estimate). While the timing of the freshwater fluxes and MOC change, as indicated by the SST proxy, is not clear, the freshening periods precede the MOC changes.

In summary, the results from numerical simulations are inconclusive regarding the volumes and pathways of freshwater from the shelves around Greenland to the central Labrador Sea. Two studies disagree on whether or not the melt from GrIS has already been reducing the intensity of deep convection. On a multidecadal time scale, a proxy study of the MOC using SSTs and accumulated mass loss from GrIS suggests that the MOC reduced in response to melt from Greenland, however, without particular attention to the timing of the response (MOC change) relative to the freshwater forcing (GrIS melt). These studies highlight some of the difficulties in linking freshwater sources to potential impacts on the MOC.

Relationship Between Deepwater Formation and Atlantic MOC

Ship-based hydrographic sections and tracer data support the conclusion that the MOC carries recently formed deep waters southward (Lozier, 2012). These deep waters can be identified by their temperature and salinity properties and also by traces of chemicals (e.g., CFCs, SF₆) entrained during interaction with the atmosphere, and they can be followed across many thousands of kilometers (Smethie et al., 2013). Chemicals introduced into the environment by humans, such as CFCs, show elevated concentrations in patterns consistent with the spread of recently ventilated water in the North Atlantic (Figure 2). Concentrations of these chemicals are higher near the source regions (the deepwater formation regions) and decrease as the water is transported away and the tracer diluted. The relatively higher concentrations along the western boundary of the Atlantic support the idea of a deep western boundary current (DWBC), where the southward flow of the overturning circulation is concentrated (Figure 2). However, observational evidence for a relationship between variations in convection and

deepwater formation with variations in the MOC strength is lacking (see review by Lozier, 2012).

Two investigations into the relationship between deepwater formation and MOC strength test the idea that more convection translates to stronger meridional circulation. Deep water formed during wintertime convection at high northern latitudes is expected to supply the southward-flowing limb of the MOC. Pickart and Spall (2007) tested the relationship between deep mixing and the MOC using float trajectories and repeat hydrographic sections during the period 1990–1997. For determining the formation of deep water, they calculated only 1 Sv of downward flow and 2 Sv of vertical mixing. When compared to the strength of the MOC (~17 Sv), this source of deep water is insufficient to supply the southward limb of the overturning. A separate study relied on velocity measurements from an array of moorings at 53°N along the western boundary of the Labrador Sea. These moorings were used to evaluate the strength of the DWBC during the periods 1993–1995 and 1999–2001 (Schott et al., 2004). The second period is known to have weaker convection and LSW production than the first, but the strength of the DWBC was not weaker in the second period. These results appear to defy the expectation that stronger convection and LSW production correspond to stronger southward flow in the MOC.

Numerical and observational studies now distinguish between the DWBC and the deep limb of the MOC, finding that the southward flow of the Atlantic MOC is not confined to the western boundary. Rather, the recently formed LSW spreads across the interior of the Atlantic Ocean (Bower et al., 2009; Lozier, 2012). The array used to measure the DWBC only measures the southward flow at the western boundary, and not this interior flow. The DWBC may also have additional variability due to local mesoscale eddies that do not necessarily contribute to net southward flow (Wunsch and Heimbach, 2013). While measured properties and

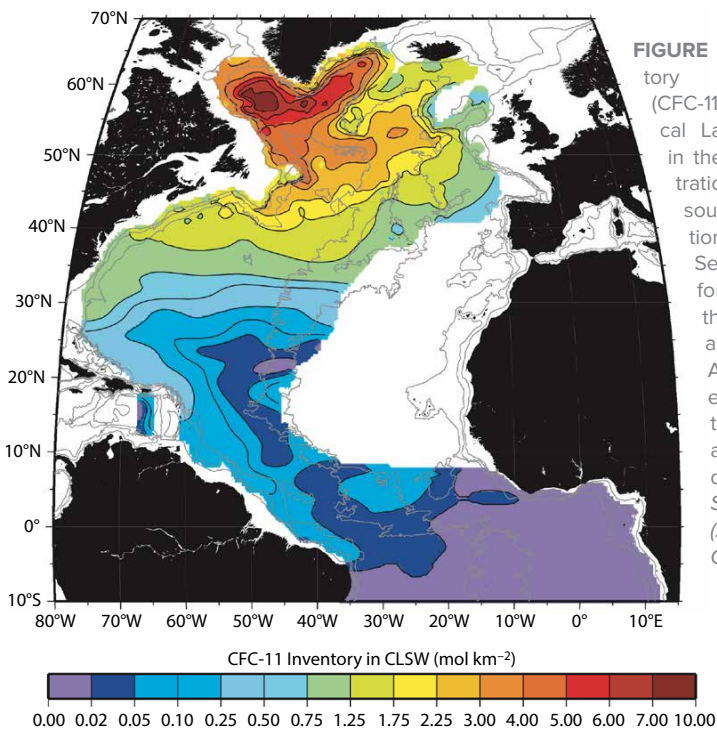


FIGURE 2. Map of the inventory of chlorofluorocarbons (CFC-11, mol km⁻²) in the classical Labrador Sea Water layer in the Atlantic. Higher concentrations are found near the source regions (deep convection region in the Labrador Sea and overflow waters formed north of Iceland). To the south, concentrations are higher in the western Atlantic than in the eastern, which suggests that the deep western boundary current is a major conduit for southward flow. Source: Smethie et al. (2013), ©2013 American Geophysical Union

tracers in deep waters confirm the southward flow of LSW away from deep convection regions in the north where they are formed, data linking the volume or intensity of deepwater formation and the strength of the overturning are lacking.

OBSERVING THE MOC AND GRIS MELTING

Overturning Observations

The strength of the Atlantic MOC is defined as the net northward transport (velocity \times area) in the surface layer of the ocean, down to the depth where the flow reverses southward. The overturning can be viewed as a zonal (east-west) average of the meridional (north-south) flow, where the overturning comprises surface waters moving northward and returning southward at depth (Figure 3). At a given latitude, overturning strength is computed by zonally integrating meridional velocities, then accumulating them vertically to produce an overturning stream function. The overturning strength (the maximum in the stream function) is approximately 17 Sv.

Overturning is measured by applying the geostrophic relationship between meridional velocities and zonal pressure gradients. In the Northern Hemisphere, northward flow balances a zonal pressure gradient with higher pressure in the east. Due to intrinsic drift in pressure sensors,

seawater density rather than pressure is used in the calculation of ocean volume transports. Although density varies very little, even small variations create pressure gradients that drive ocean circulation. The basin-wide Atlantic MOC is now being observed continuously in the subtropics through a moored array at 26°N and a combination of satellite and autonomous floats at 41°N (Figure 4).

At 26°N, the RAPID (Rapid Climate Change)/MOCHA (Meridional Overturning Circulation and Heat transport Array) program collects vertical profiles of temperature and salinity at the western and eastern boundaries of the Atlantic. These properties are used to calculate density profiles that can then be used to estimate vertical shear from zonal density gradients through the thermal wind relation. Vertically integrating velocity shear profiles gives velocity profiles, and integrating a second time vertically, and once horizontally, gives volume transports. These basin profiles of transport are combined with transport estimates of the Gulf Stream through the Florida Straits and surface wind-driven Ekman transport to provide daily estimates of MOC strength (Figure 5a). Full details of the observational method can be found in McCarthy et al. (2015), and recent results are highlighted in Srokosz and Bryden (2015).

At 41°N, a combination of

satellite sea surface height measurements and hydrographic (positioning) data from autonomous Argo profiling floats gives three-monthly estimates of the overturning and meridional heat transports (Willis, 2010). Geostrophic velocity is estimated from hydrographic profiles from the Argo floats; parking depth velocities and altimetric estimates of surface velocities are used to reference geostrophic velocity estimates. The hydrographic data are corrected for eddy activity using sea surface height altimetry, then the geostrophic relation is applied. These measurements provide an estimate of the time-varying northward transport in the top 2,000 m of the ocean to which a depth-invariant compensation is applied in order to construct an estimate of the overturning on a three-month time scale (Figure 5a). At both latitudes, an extended overturning estimate is available estimated from sea surface height altimetry alone (Willis, 2010; Frajka-Williams, 2015).

The 10-year record from the 26°N moorings shows striking variability in MOC transport on time scales of days to years (Cunningham et al., 2007; McCarthy et al., 2012). On interannual time scales,

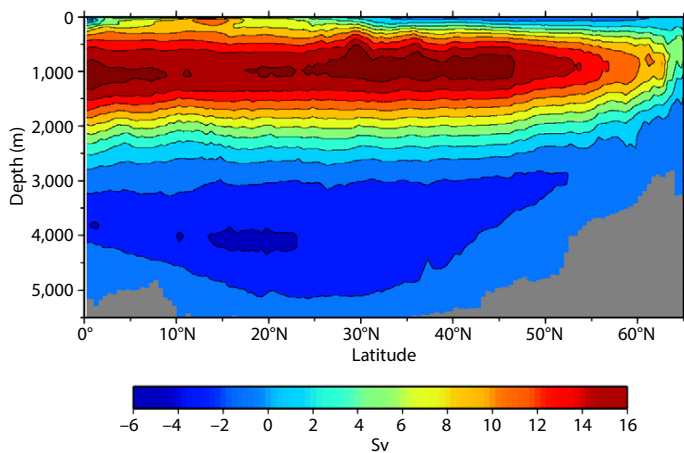


FIGURE 3. Overturning stream function for the Atlantic MOC as a function of depth and longitude. The circulation is northward near the surface, downward at northern high latitudes, and southward at depth. Source: Bingham and Hughes (2007), ©2007 by the American Geophysical Union

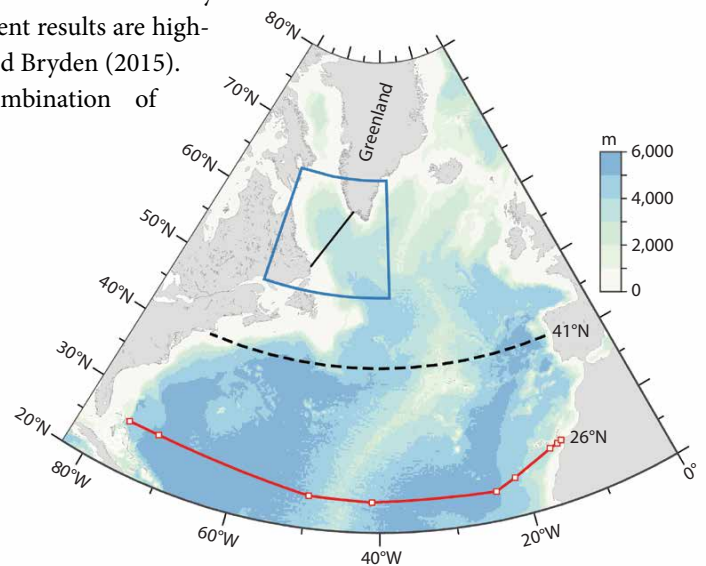


FIGURE 4. Map of the North Atlantic, showing the Labrador Sea (blue box) and the latitude of the transports estimated from altimetry and Argo floats (Willis, 2010) at 41°N (black dashed line) and RAPID (Rapid Climate Change)/MOCHA (Meridional Overturning Circulation and Heat transport Array) at 26°N (red line). Squares at 26°N denote mooring locations. The black line across the Labrador Sea tracks the Atlantic Repeat Hydrography Line 7 (AR7W). Bathymetry is shaded at 1,000 m intervals.

the transport divergence between 26°N and 41°N matched independent estimates of heat content changes in the subtropical Atlantic (Cunningham et al., 2014). Over the 10-year observational period at 26°N, the transports show a decline in the net northward MOC estimated at -0.4 ± 0.2 Sv yr⁻¹ (Smeed et al., 2014; updated in Frajka-Williams et al., 2016). This reduction is primarily seen in deep transports (3,000–5,000 m) rather than in intermediate layer transports (1,100–3,000 m). At 41°N, the trend over December 2001 to November 2014 is not significant, though the trend at 41°N over the RAPID period of April 2004 to March 2014 is -0.3 ± 0.1 Sv yr⁻¹. While the amplitude of the trend at the two latitudes varies, both show a substantial reducing trend over the recent decade. A more complete review of the observations and mechanisms dominating the Atlantic MOC can be found in Buckley and Marshall (2016).

Freshwater Fluxes from Greenland Mass Changes

Freshwater fluxes from Greenland include both surface runoff and solid ice discharge from the various marine-terminating glaciers. From a combination of observations and model output, freshwater fluxes were constructed for the period 1958–2010 (Bamber et al., 2012). Surface runoff can be estimated from a surface mass balance model forced by reanalysis data. Solid ice discharge across the land/ice boundary (termed the grounding line) can be estimated from ice motion determined by satellites combined with measurements or estimates of ice thickness (Bamber et al., 2012). Estimating freshwater flux as an anomaly relative to a more stable period (1960–1990) shows that mass loss from Greenland has increased in recent decades, with a present-day flux of about 1,100 km³ yr⁻¹, representing an increase of about 300 km³ yr⁻¹ over 1960–1990 levels (Bamber et al., 2012).

Here, we update these results to include the years 2011–2014 (Figure 6) using data and methods associated with

CryoSat-2 (Wouters et al., 2015). The freshwater flux anomaly (calculated with respect to the mean for the period 1960–1990) has now reached ~5,000 km³ since about 1995. This is around half the magnitude of the Great Salinity Anomaly. Since 2009, 84% of the increase in freshwater flux seen in Figure 6 was due to enhanced surface melting rather than ice discharge (Enderlin et al., 2014). As temperatures rise in the future, this enhanced surface melt is predicted to increase, resulting in further freshening and accelerating freshwater flux anomalies in the subpolar North Atlantic (Fettweis et al., 2013). At the same time, observational studies are investigating the role of enhanced melt of marine-terminating glaciers in Greenland fjords by the warm influx of Atlantic waters. Contrasting with surface melt, this meltwater may enter the Atlantic at mid-depth rather than at the surface (see review by Straneo and Cenedese, 2015). While the acceleration of ice melt from the northern high latitudes provides a source of freshwater proximate to regions

of dense water formation, it is important to note that not all of this freshwater gets to areas of overturning in the Labrador and Irminger Seas (see earlier discussion and Böning et al., 2016).

Linking Overturning and Convection Through Density

Time-series observations of convection are difficult to obtain, as convection is a spatially inhomogeneous process that occurs during the harsh winter months in the North Atlantic. With the Argo float program, vertical profiles of hydrographic properties are available—albeit distributed unevenly in space—year-round since approximately 2004. While past observational attempts to relate convection to overturning have been inconclusive, they have tended to focus on the volume or depth of convection as a measure of its intensity. An alternate measure of the integrated effect of convection is the density of the deep waters formed. Like other measures of deep convection, the density of deep water formed can

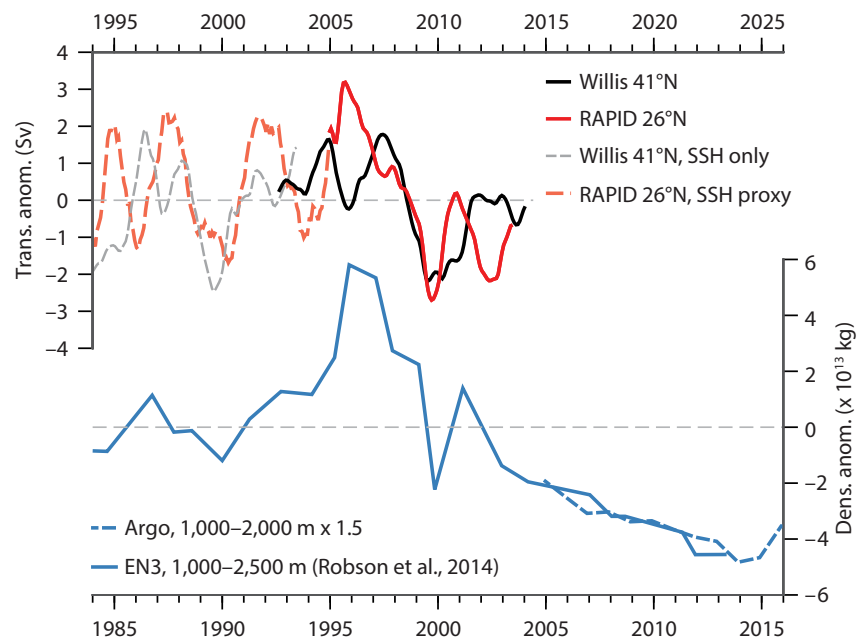


FIGURE 5. Overturning transport anomalies at 26°N and 41°N after removing Ekman transports. At 41°N, transports are estimated using Argo and altimetry (black line) and altimetry only (dashed gray line). At 26°N, transports are estimated from the RAPID array (solid red line) and altimetry (dashed pink line). The blue line shows the density anomaly integrated in the Labrador Sea between 1,000 m and 2,500 m from Robson et al. (2014), and the dashed blue line shows that anomaly estimated from Argo over 1,000 m to 2,000 m, then scaled by 3/2. The x-axis on the two plots is offset by 10 years to show a possible lagged response of transports to density changes.

vary due to the air-sea fluxes encountered during the convective winter, but it is also modulated by the density of the water that was there prior to convection. In this way, multiple years of strong air-sea fluxes and deep convection (e.g., 1990–1995) result in a continued densification of the deep water layer even if the depth or volume of deep water does not change (Lazier et al., 2002; Yashayaev, 2007). Multiple years with an absence of deep convection (e.g., since 1995) result in gradual restratification of the deep layer by lateral fluxes from lighter surrounding waters (van Aken et al., 2011).

The relationship between the strength of the overturning and the density of water in the North Atlantic has been investigated in a range of models of different complexity, from simple box models (Stommel, 1961) to global simulations (Butler et al., 2015). While the strength of the overturning at a given latitude is measured by the twice-integrated zonal density gradient (as described above), a possible driver of the MOC is through buoyancy forcing establishing meridional density gradients (see a review in Kuhlbrodt et al., 2007). While these idealized investigations typically consider long time scale variations in the MOC, they find that meridional gradients of twice-integrated density anomalies show a positive relationship with overturning strength on time scales of 10 years and longer (Butler et al., 2015). These meridional density gradients are controlled by density in the northern North Atlantic rather than in the Southern Hemisphere.

On interannual time scales, several studies have investigated MOC variability and the response of the MOC to perturbations or fluxes at high latitudes. Robson et al. (2014) showed that deep density anomalies in the Labrador Sea (50°N–65°N and 38°N–65°N between 1,000–2,500 m water depth) provided a footprint of MOC strength in a numerical simulation. When the density of water between 1,000 m and 2,500 m depth in the Labrador Sea was greater, the Atlantic MOC was stronger. Using historical

observations of density anomalies in the Labrador Sea, they further showed that deep density anomalies in the Labrador Sea are presently declining and suggested that the observed slowing trend of the Atlantic MOC at 26°N is due to the density decreases. Also on interannual time scales, the Atlantic MOC in the subtropics has been found to respond to variations in surface freshwater or heat fluxes in the subpolar regions with a time lag of nine years (Pillar et al., 2016). The time lag and sensitivity were determined using a linear adjoint model to identify optimal perturbations to the MOC at 26°N. The time lag was associated with the decadal time scale of a thermal Rossby mode in the subpolar regions. Considering temperature and salinity anomalies from hydrographic sections at 26°N, van Sebille et al. (2011) found a similar time lag of nine years between hydrographic anomalies in the Labrador Sea and at 26°N. In this case, the time scale was suggested to be advective, with nine years being the time it takes for the bulk properties of the LSW to physically move from their formation region to subtropical latitudes. Advective time scales are generally longer than wave-propagation time scales. In studies of the meridional coherence of the MOC (Bingham et al., 2007; Zhang, 2010), anomalies propagate

at advective speeds through the subpolar gyre to the subpolar-subtropical interface. There, they excite a rapid adjustment of the MOC that spreads meridionally across the subtropical gyre within months (Johnson and Marshall, 2002). While the mechanisms tying anomalies in the subpolar regions to subtropical latitudes are unclear, the interannual variability of the MOC and property anomalies in the subtropical regions lag the overturning in the subpolar regions by some years (ranging from three to nine years in the model studies cited above).

Using Argo float data (Roemmich and Gilson, 2009), we update the density anomaly in the Labrador Sea, finding a continued decline through 2013 (Figure 5b). Since 2014, densities increase due to deeper convection (de Jong and de Steur, 2016). Shifting the density anomaly time series later by 10 years aligns some of the peaks in density anomaly with peaks in subtropical overturning anomalies. The transport estimates from 26°N and 41°N show a local peak in overturning during the 2004–2006 period, about 10 years after the increasingly deep convection from 1987 to 1994/95 (Yashayaev, 2007). Even with the return of deep convection in the past couple of years, the deep density anomaly is still below the average over

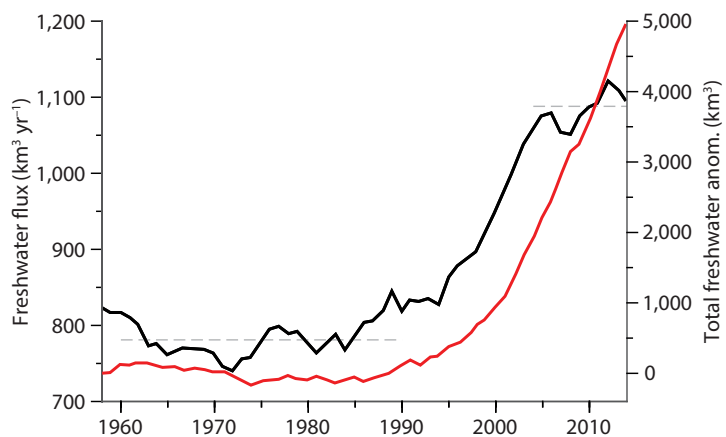


FIGURE 6. Absolute freshwater fluxes from Greenland (black line, left-hand axis) and cumulative freshwater flux anomalies with respect to the reference period 1960–1990 (red line, right-hand axis). The data are updated from Bamber et al. (2012), with CryoSat-2 estimates of mass trends for 2011–2014 inclusive. The gray dashed lines show the means for 1960–1990 and 2004–2014. The difference between these two periods is slightly more than 300 km³ yr⁻¹.

the Argo period (Figure 7).

While convection in the Labrador Sea is only one source of deep water to the southward MOC flow, the peak response of the MOC at 26°N lags high-latitude buoyancy forcing by nine years (Pillar et al., 2016). Using this response time scale, a strong Atlantic MOC would be observed at 26°N in 2003. From the RAPID observations, MOC strength was at its highest when observations began in 2004 and has been declining since. Using extended estimates from 41°N transports from sea surface height altimetry suggests that the Atlantic MOC was not anomalous prior to 2005, while extended estimates of the overturning at 26°N have larger variability, perhaps due to wind forcing, but also peak in 2004.

In this section, we considered density anomalies as a measure of convective strength as well as the important influence convection and deepwater formation have on the MOC. During periods with several years of convection, the density of deep waters formed may increase, while during periods without convection (as during the Great Salinity Anomaly in the 1960s), deep densities gradually decline due to restratification by lighter,

surrounding waters. The asymmetry between convection and absence of convection has been difficult to tie to changes in meridional transports (see section on Relationship Between Deepwater Formation and Atlantic MOC) because years without convection still showed southward flow. With density as the measure of convection, years without convection mean that MOC strength should gradually decline due to restratification, not abruptly halt, as would be implied by using mixed layer depth as a measure of convection. Here, we have shown a tentative connection between density anomalies in the Labrador Sea and overturning strength, using a model-derived estimate for the time lag between them. We have not identified a robust link between deep density anomalies and freshwater fluxes. Yang et al. (2016) investigated this link and found that recent freshening resulted in reduced LSW formation since the mid-1990s; however, the return of deep convection in 2013/14 and 2014/15 would seem to defy their conclusions. While freshwater fluxes have been accelerating since the 1990s (Bamber et al., 2012), Argo float data in the Labrador Sea indicate only marginal

freshening, and the most recent three winters have been accompanied by deep convection (Figure 7).

SUMMARY AND OUTLOOK

In high-latitude regions, air-sea fluxes cool surface waters, working against stratification to form deep, homogeneous waters through the process of deep convection. These newly ventilated deep waters spread southward away from the convection regions through the lower limb of the Atlantic MOC, as seen in chemical tracer and Lagrangian studies. However, the speed at which these waters move southward (i.e., the strength of the overturning circulation) is not set by the mixed layer depth during convection nor by the volume of deep water formed. Rather, pressure gradients (typically calculated from profiles of seawater density) drive circulation. Since 1994, the density of deep waters in the Labrador Sea has been decreasing while in situ observations of Atlantic MOC strength have shown an accompanying weakening tendency since 2004. Numerical models (forced and adjoint) show that the MOC response to high-latitude changes should be lagged (with lags identified in models ranging from three to nine years).

In the last two decades, satellite and in situ technology have enabled continuous estimates of the variability of ocean transports and ice mass. There has been a marked increase in GrIS melting since the 1990s (Figure 6 and Bamber et al., 2012), while deep densities in the Labrador Sea (possibly linked to the strength of the overturning) have been declining (Figure 5 and Robson et al., 2014). At the same time, the overturning circulation shows a tendency to slow (Figure 5), with a possible lag of O(10 years) following density anomalies in the Labrador Sea. With the advent of observations of oceanic transports by the RAPID array at 26°N (and other estimates, for example, at 41°N and 16°N, and in the sub-polar gyre), and Argo float profiles of hydrography in the top 2,000 m, we can look forward to better investigating

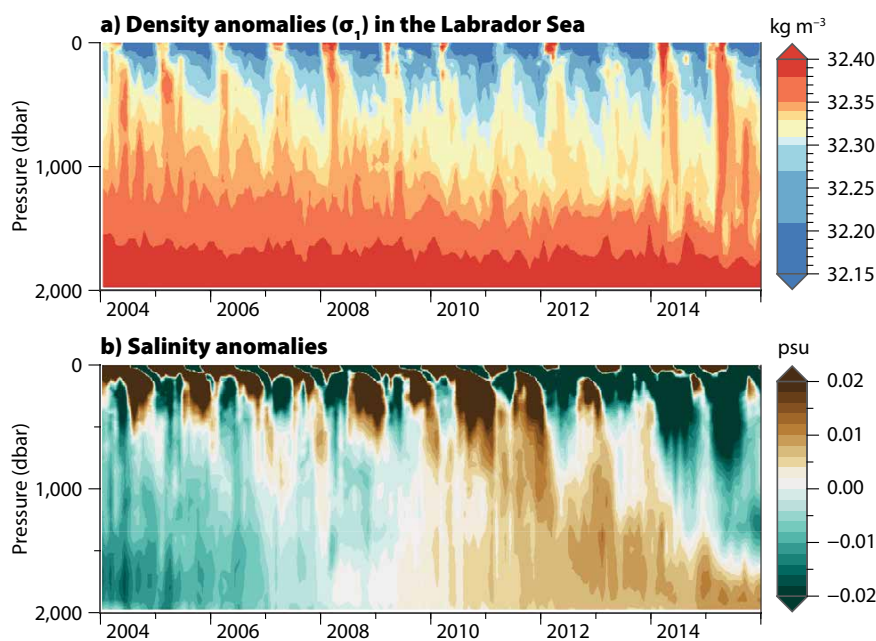


FIGURE 7. (a) Density time-depth sections (σ_1) are shown for the region 50°N–65°N, 38°W–65°W from Argo climatology (Roemmich and Gilson, 2009). (b) Salinity anomalies for the same region.

pressing questions of Greenland's influence on the Atlantic MOC. New satellite data sets provide more detailed estimates of freshwater melt from the Greenland Ice Sheet. As these records extend in time, we can answer: How has (or will) melt from the GrIS influence convection? While convection responds to a range of forcings—surface fluxes, preconditioning, and lateral fluxes/restratification—the anticipated continued and accelerating melt of the GrIS (Bamber et al., 2012) may lead to freshwater forcing becoming a dominant influence on convection in the North Atlantic (Böning et al, 2016).

Several key gaps in knowledge limit investigations of the influence of Greenland melt on ocean circulation. While freshwater hosing experiments can shut down the MOC, pathways of freshwater transport from source areas to regions of convection is not well understood. Numerical simulations of the processes are challenging to represent at adequate resolution and with appropriate parameters for resolving the eddy and boundary processes, and they tend to be model and/or resolution dependent (Myers, 2005; Dukhovskoy et al., 2016). Satellite altimetry does not resolve small-scale eddies in the Labrador Sea, and Argo floats do not profile at the shelf edge. While surface drifters can be used to track freshwater at the surface, their numbers—particularly on the shelves—are sparse. Drifters deployed in the open ocean seldom reach the Greenland shelves, and they would be difficult to deploy at sufficient temporal frequency to resolve seasonal variations. Tracer methods using noble gases have been used within Greenland's fjords to distinguish the sources of freshwater in the fjords (surface or submarine melt; Beaird et al., 2015) and could be applied to trace the spread of Greenland meltwater into the Labrador Sea.


Due to numerical challenges in simulating cross-shelf exchange, observations are likely to be needed to identify pathways and processes of freshwater transport. To elucidate the link between

convection and the strength of the Atlantic MOC, we must determine what metric of convection the overturning responds to, whether it is the depth of wintertime mixing, the volume of deep water formed, or the density of deep water formed. While the observations of overturning at 26°N provide a continuous time series of MOC strength since 2004, a new program of observations of ocean circulation in the subpolar gyre (Overturning in the Subpolar North Atlantic Program, OSNAP) may show an ocean response to high-latitude forcing sooner. Although we focused on open-ocean convection and the MOC in this article, dense water formation in the Nordic Seas (north of Iceland) and freshwater influences or intrusions of salty North Atlantic water are both likely to be important to the evolution of ocean circulation, dense water formation, and overflows in the coming decades. The Nordic Seas present their own sets of modeling challenges in regions of limited observations (under Arctic ice, ice-ocean influences on circulation, and dense water flowing over sills).

The MOC has garnered much attention in the literature, both as a key component of the present climate system and because it has been implicated in dramatic global shifts in paleoclimate records. Observations of the MOC over the past 10 years show a marked reduction in its strength (Frajka-Williams et al., 2016), but this slowing down is indistinguishable from natural variability (Roberts et al., 2014). Numerical models suggest that dramatic changes in the MOC may result from small perturbations (Lenton et al., 2008), possibility due to the sensitivity of the MOC to freshwater feedbacks (Rahmstorf, 1996). In particular, the sign of freshwater transport relative to the volume transport may be an indicator of MOC stability. If the sign is positive, then a weakening of the MOC brings less freshwater to the subpolar region, leading to recovery of the MOC through enhanced convection (negative feedback). If the sign is negative, then a weakening of the

MOC exports less freshwater from the subpolar region, leading to an increase in surface buoyancy and further reduction of the MOC (positive feedback). At 26°N, observations show that the MOC exports freshwater southward (McDonagh et al., 2015). In the absence of other changes, the recent reduction of overturning at 26°N should result in an accumulation of freshwater north of 26°N, possibly contributing to a further slowdown of the overturning. At present, the combination of reduced MOC and increased GrIS inputs is resulting in freshwater accumulation in the subpolar regions. But, freshwater content there is not yet anomalous because the increase in freshwater is reducing the anomalously salty state of the 1990s (Kelly et al., 2016). However, meltwater from Greenland could initiate a slowdown of the MOC, which would then be compounded by the freshwater feedback (reducing MOC exports less freshwater to the south), leading to further slowing of the MOC (Hawkins et al., 2011).

In this article, we explored evidence to support the hypothesis that melting of the Greenland Ice Sheet influences ocean circulation. We discussed how the ocean's response to varied forcings is complicated, occurring on a range of time scales and mediated by interactions with the cryosphere and atmosphere. Several possible responses (and time lags) have been identified through modeling or have been explained by theory, suggesting that observed circulation changes may result from several competing influences and from a range of past events. Identifying the driver of MOC variations is further complicated by the fact that the dominant influence may change over time. Evidence is building that there is a time lag between the formation of deep density anomalies in the Labrador Sea and their expression in Atlantic MOC strength (Robson et al., 2014; Pillar et al., 2016, and figures therein). Yang et al. (2016) argue that LSW thickness and density anomalies have been decreasing due to recent freshwater inputs. While the 2013/14 and 2014/15 winters showed strong convection in the

Irminger and Labrador Seas (apparently contradicting the findings of Yang et al., 2016), the continued intensifying freshwater fluxes from the GrIS (Bamber et al., 2012) may soon bring us past a threshold where freshwater influences outweigh other forcings of convection to bring convection to a halt. The resulting reduction in deep density anomalies should then reduce the strength of the MOC and, through freshwater feedbacks, result in compounded freshening and reduction of density in the subpolar region (Kelly et al., 2016). As observational records lengthen and freshwater fluxes increase, the influence of freshwater on convection may soon become dominant over other factors responsible for MOC variability (Böning et al., 2016). 

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