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Invited review

Coupled atmosphere-ice-ocean dynamics in Dansgaard-Oeschger events

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A R T I C L E I N F O

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A B S T R A C T

The Dansgaard-Oeschger events of the last ice age are among the best studied abrupt climate changes, yet a comprehensive explanation is still lacking. They are most pronounced in the North Atlantic, where they manifest as large temperature swings, on timescales of decades or shorter, between persistent cold (stadial) and warm (interstadial) conditions. This review examines evidence that Dansgaard-Oeschger events are an unforced or “spontaneous” oscillation of the coupled atmosphere-ice-ocean system comprising the North Atlantic, Nordic Seas and Arctic, collectively termed the Northern Seas. Insights from reanalysis data, climate model simulations, and idealized box model experiments point to the subpolar gyre as a key coupling region where vigorous wind systems encounter the southernmost extension of sea ice and the most variable currents of the North Atlantic, with connections to the deep ocean via convection. We argue that, under special conditions, these components can interact to produce Dansgaard-Oeschger events. Finding the sweet spot is a matter of understanding when the subpolar region enters a feedback loop whereby changes in wind forcing, sea ice cover, and ocean circulation amplify and sustain perturbations towards cold (ice-covered) or warm (ice-free) conditions. The resulting Dansgaard-Oeschger-like variability is seen in a handful of model simulations, including some “ugly duckling” pre-industrial simulations: these may be judged as undesirable at the outset, but ultimately show value in suggesting that current models include the necessary physics to produce abrupt climate transitions, but exhibit incorrect sensitivity to the boundary conditions. Still, glacial climates are hypothesized to favour larger, more persistent transitions due to differences in large-scale wind patterns. Simplified models and idealized experimental setups may provide a means to constrain how the critical processes act, both in isolation and in combination, to destabilize the subpolar North Atlantic.

1. Introduction

The Dansgaard-Oeschger events of the last ice age represent one of the best observed yet most intriguing examples of abrupt climate change in Earth's history. In the Greenland ice cores, where they were first discovered and are still most prominent (Bond et al., 1993; Dansgaard et al., 1993), they depict large swings in temperature and snow accumulation, of a magnitude unequalled in the 100 ky-long record. Dansgaard-Oeschger events involve a wide range of time scales, with fast (decadal or shorter) transitions between persistent stadial (cold) and interstadial (warm) states lasting many centuries. The overall millennial time scale of Dansgaard-Oeschger events is not consistent with any strong external forcing such as solar variability. This fascinating combination of characteristics has fueled a lively ongoing discussion within the paleoclimate and climate dynamics communities about the causes and mechanisms behind Dansgaard-Oeschger events.

New paleoclimate records and modelling studies of the last decade have greatly advanced efforts to piece together the hows and whys of Dansgaard-Oeschger events, but a definitive answer is still pending (section 2). Many existing hypotheses invoke changes in the meridional overturning circulation of the ocean as the prime

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underlying cause (Broecker et al., 1985). They rely on a simple link between the overturning circulation and regional climate: during periods of stronger overturning, the ocean transports more heat poleward, thereby warming the region around Greenland (Clark et al., 2002; Rahmstorf, 2002). Overturning changes could also explain the Southern Hemisphere expression of Dansgaard-Oeschger events, which are antiphased with the Northern Hemisphere signals (Stocker and Johnsen, 2003; WAIS Divide Project Members, 2015). There are, however, some aspects of Dansgaard-Oeschger events that do not fit as neatly into the overturning hypothesis. For example, freshwater perturbations are typically needed to trigger overturning changes in models (Kageyama et al., 2010), but actual freshwater inputs seem to come after the abrupt climate transitions (Barker et al., 2015). Furthermore, while abrupt shutdown and recovery of the overturning circulation in response to freshwater perturbations can be simulated in intermediate complexity models (e.g., Ganopolski and Rahmstorf, 2001; Kageyama et al., 2010), state-of-the-art climate models for the most part simulate a smoother recovery (Vellinga et al., 2002; Bitz et al., 2007; Otto-Bliesner and Brady, 2010). The fact that models can behave differently is not surprising given differences in resolution, representation of physical processes, and sensitivity to freshwater, but it complicates attempts to constrain dynamical mechanisms.

Among the alternative explanations for Dansgaard-Oeschger events is one, based primarily on some recent modelling studies, that relies on dynamics internal to the climate system. These studies suggest that Dansgaard-Oeschger events may be a class of “spontaneous” climate oscillation (e.g., Drijfhout et al., 2013; Peltier and Vettoretti, 2014; Kleppin et al., 2014; Martin et al., 2015; Sidorenko et al., 2015). In other words, as opposed to being triggered by freshwater or some external forcing, they could arise from atmosphere-ice-ocean interactions that alter poleward energy transport. As in the traditional overturning hypothesis, the alternating periods of strong and weak transport would cause swings between interstadial and stadial conditions. The key interactions are hypothesized to occur in the Northern Seas, the region comprising the North Atlantic south of the Greenland-Scotland Ridge, the Nordic Seas, and the Arctic Ocean (Fig. 2). The subpolar gyre seems to be a critical coupling location, where vigorous wind systems meet the southernmost extension of sea ice and the most variable currents of the North Atlantic. Through connections to deep ocean ventilation and overturning (Hátún et al., 2005; Böning et al., 2006; Gao and Yu, 2008; Rhein et al., 2011), variations in the subpolar gyre are the dominant source of decadal climate variability in the region today (Häkkinen and Rhines, 2004; Yoshimori et al., 2010; Yeager et al., 2012; Matei et al., 2012).

In this review, we examine the role of the Northern Seas in Dansgaard-Oeschger events. Section 2 presents a brief introduction to Dansgaard-Oeschger events, their main characteristics, and the leading hypotheses. Section 3 starts with the physical processes controlling sea ice cover, ocean heat transport and atmospheric circulation in the Northern Seas today, then uses these concepts to reverse-engineer a scenario that can produce large climate swings in Greenland. Section 4 investigates how such climate change might be generated spontaneously by the coupled dynamics of the region. Section 5 gathers insights presented in previous sections to discuss how the Northern Seas system can be “primed” to support Dansgaard-Oeschger events, both in climate models and in the real world. Finally, section 6 contains some concluding remarks.

Our approach relies on existing knowledge of modern day atmosphere-ice-ocean dynamics to help understand why Dansgaard-Oeschger events occur during glacial times but not during interglacials. The overall aim is to explore the idea that we do not need to invoke exotic mechanisms or unknown forcings to explain Dansgaard-Oeschger events, but that they can arise from known physics acting under altered background conditions.

2. A primer on Dansgaard-Oeschger events

This section provides some background on Dansgaard-Oeschger events to set the stage for the rest of the paper. It reviews their basic features and the main hypotheses for why they occur. The section is intended to orient readers who are not familiar with Dansgaard-Oeschger events; for those who are well acquainted with the topic, it may be skipped without loss of continuity.

Dansgaard-Oeschger events are large, abrupt climate swings that punctuated the last ice age. They were concentrated in the period from 57 to 29 ka (known as Marine Isotope Stage 3 or MIS3; Fig. 1), and backdropped by a glacial world where massive ice sheets covered much of North America and Eurasia as well as Greenland and Antarctica. The ice sheets reached their maximum extent at Last Glacial Maximum (LGM; 21 ka), when global ice volume was equivalent to 110–130 m of sea level, the majority of which (estimated 70–90 m) was contained in the Laurentide ice sheet over North America (Clark and Mix, 2002; Lambeck et al., 2014; Clark and Tarasov, 2014; Abe-Ouchi et al., 2015). Geochemical and paleontological evidence suggest global mean surface temperature at LGM was very likely 3°C–8°C cooler than pre-industrial times, with larger cooling over land compared to ocean (see Masson-Delmotte et al., 2013, and references therein). Conditions prior to LGM, during the time of the Dansgaard-Oeschger events, are more poorly constrained by observations; estimates based on modelling and observational studies suggest a slightly warmer world compared to LGM, with a global ice sheet volume of 60–90 m sea level equivalent (Chappell, 2002; Van Meerbeeck et al., 2009).

Dansgaard-Oeschger events are most clearly seen in Greenland ice core records (Dansgaard et al., 1993; Andersen et al., 2004), where they represent temperature swings of 8–16°C (Severinghaus et al., 1998; Wolff et al., 2010). The warm (interstadial) conditions see twice as much snow accumulation as the cold (stadial) conditions, with proportionally more of the snow falling in winter (Alley et al., 1993; Cuffey and Clow, 1997). In addition, interstadials are associated with increased levels of methane and nitrous oxide, and decreased levels of dust and sea salt (see Thomas et al., 2009, and references therein), all of which indicate that the footprint of Dansgaard-Oeschger events extends well beyond Greenland. Throughout the North Atlantic, observational evidence points to warmer ocean temperatures (Bond et al., 1993; Curry and Oppo, 1997), less sea ice (Masson-Delmotte et al., 2005; Rasmussen and Thomsen, 2004; Dokken et al., 2013), and fresher surface waters in the subtropical gyre (Schmidt et al., 2006). While the events themselves are millennial in time-scale, the transitions occur within decades or less, and the stadial-to-interstadial (cold-to-warm) transitions are more abrupt (Alley and Clark, 1999).

Proxy evidence reveals that Dansgaard-Oeschger signals extended across the Northern Hemisphere and into the Southern Hemisphere (see Völker, 2002, and references therein). Stadials are associated with a northern climate that was generally cold, dry and windy. In the Northern Hemisphere tropics, the Asian and West African monsoons were weakened, and the tropical Atlantic rain belt was shifted southward (Chiang, 2009, and references therein). Where Southern Hemisphere signals exist, they suggest an antiphased relationship with the Northern Hemisphere (Blunier and Brook, 2001; Barker et al., 2011; WAIS Divide Project Members, 2015; Pedro et al., 2018).

It is important to note that this review focuses on the 20-odd repeated events numbered in blue in Fig. 1, and thus excludes from discussion other types of abrupt climate change. These include the Bølling-Allerød warming and Younger Dryas cold

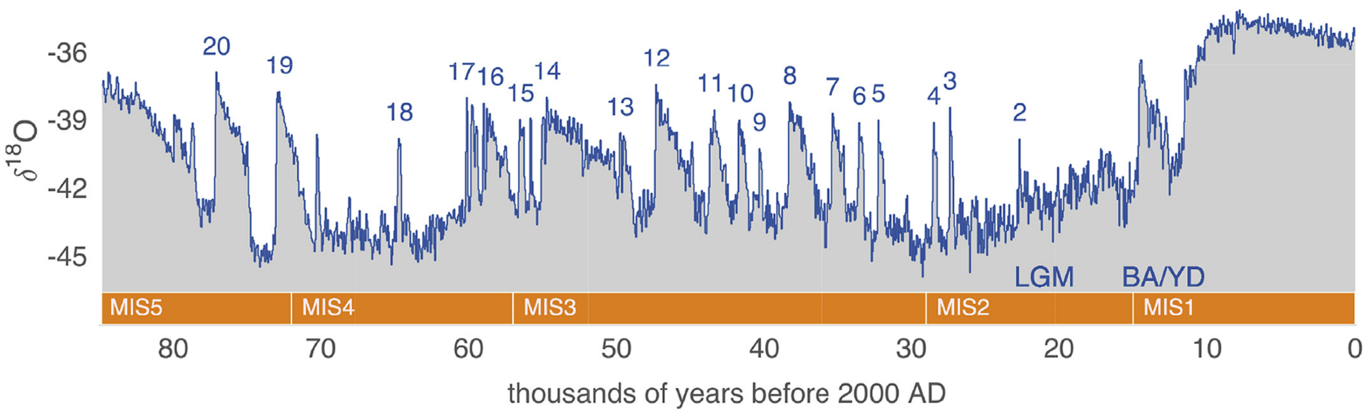

Fig. 1. Dansgaard-Oeschger events in a record of the oxygen isotopic composition (δ^{18} O) of Greenland ice in thousands of years before 2000 A.D. (ka). The ice core is from the North Greenland Ice Sheet Project site at 75.10°N, 42.32°W, elevation of 2917 m (NGRIP Members, 2004). The horizontal orange bar indicates the Marine Isotope Stages (MIS) through this time period. Dansgaard-Oeschger events are labelled with blue numbers. Also labelled are the Last Glacial Maximum (LGM) and the Bølling-Allerød/Younger-Dryas transition (BA/YD). See section 1 for more details.

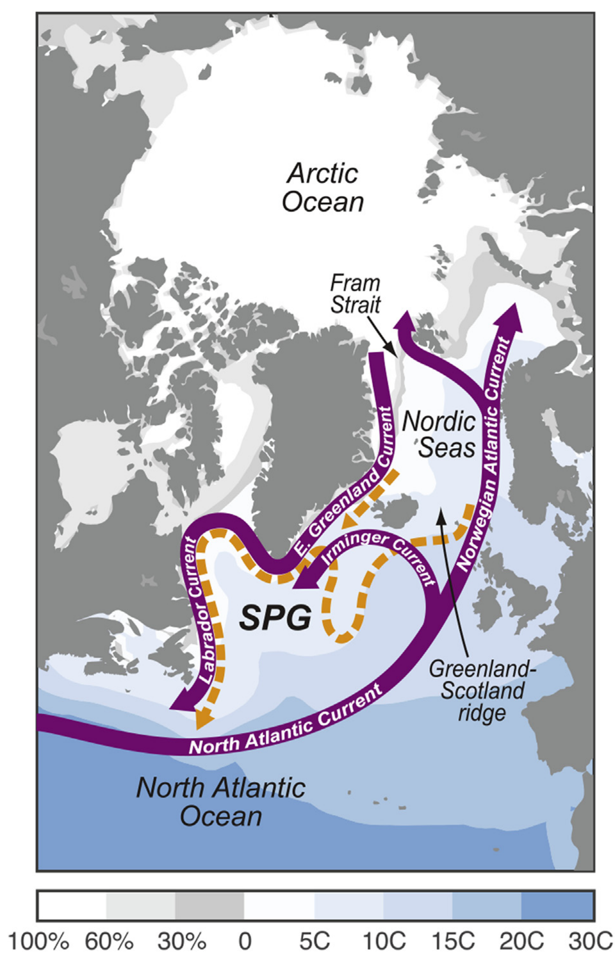

Fig. 2. Map of the Northern Seas. Purple arrows represent surface currents, dashed orange arrows represent deep currents, and SPG indicates the subpolar gyre. Shading shows annual mean sea surface temperature (blues) and sea ice concentration (greys) from the NOAA Optimum Interpolation version 2 SST reanalysis (Reynolds et al., 2007).

reversal (labelled BA/YD in Fig. 1), which occurred after LGM as the world was transitioning into the current interglacial climate, and are associated with large variations in freshwater input to the ocean from the collapsing Laurentide ice sheet (Tarasov and Peltier, 2005; Liu et al., 2009). In addition, there are a handful of Heinrich events spanning the MIS5 to MIS2 period, identified from anomalous layers of ice-rafted detritus (indicative of increased iceberg calving from ice sheets) in marine sediments throughout the North Atlantic (Hemming, 2004). The iceberg discharges occur during stadials (Barker et al., 2015), but only select ones (see Fig. 1 in Lynch-Stieglitz, 2017), and hence are not considered a general characteristic of Dansgaard-Oeschger events. Because these other phenomena involve factors not systematically associated with Dansgaard-Oeschger events, they may also arise from mechanisms distinct from those we would like to focus on.

The original proposition for why Dansgaard-Oeschger events occurred, suggested by Broecker et al. (1985), remains the prevailing view some three decades later: that they are a result of variability in the Atlantic Meridional Overturning Circulation (AMOC). The idea stems from a bistability of the thermohaline circulation found in the classic box model of Stommel (1961) as well as early climate models (Manabe and Stouffer, 1988). A strong meridional overturning circulation means more northward heat transport by the ocean, which warms northern climate, while a weak circulation means less northward heat transport, which cools northern climate (Clark et al., 2002; Rahmstorf, 2002). Other hypotheses proposed over the years (Clement and Peterson, 2008) investigate the role of alternative mechanisms for Dansgaard-Oeschger events, including reorganizations of the tropics (Clement et al., 2001; Chiang, 2009; Seager and Battisti, 2007), solar variability (Braun et al., 2008), sea ice variability (Gildor and Tziperman, 2003; Kaspi et al., 2004), shifts in Northern Hemisphere (Seager and Battisti, 2007; Wunsch, 2006) or Southern Hemisphere (Banderas et al., 2012) wind patterns, and ice-shelf growth and decay (Alley et al., 2001; Petersen et al., 2013). Each of these hypotheses can account for certain observed features of Dansgaard-Oeschger events, but the AMOC hypothesis remains central to the discussion because of its ability to capture much of the spatial pattern of the abrupt climate changes.

The observational record has offered equivocal evidence for the AMOC hypothesis to date, although there appears to be growing support for systematic changes in ocean circulation (not necessarily AMOC) with Dansgaard-Oeschger events (see Lynch-Stieglitz, 2017, for a comprehensive review of this topic). Uncertainty persists because it is no easy task to observe AMOC variability itself, much less the rapid AMOC shifts that are thought to accompany Dansgaard-Oeschger events. To do so requires both adequate spatial coverage to capture changes in AMOC structure (e.g., shoaling or deepening of the overturning cell, shoaling or deepening of the northern-southern water mass interface, shift in location of the sinking branch, etc.) as well as adequate temporal resolution to capture the fast times scale. The most direct evidence comes from proxies that reflect changes in subsurface flow strength and water-mass properties (Shackleton et al., 2000; Lynch-Stieglitz, 2017), but there is also supporting evidence from observations of Nordic Seas overflows (Kissel et al., 1999), sea ice formation (Dokken et al., 2013), spatial gradients on Greenland that reflect sea ice extent (Guillevic et al., 2013), and antiphased Southern Hemisphere signals (Barker et al., 2011). Some records show evidence for fairly systematic ocean circulation changes with most Dansgaard-Oeschger events (Burckel et al., 2015; Henry et al., 2016); others are not as conclusive (Hoogakker et al., 2007; Thornalley et al., 2013; Them et al., 2015).

The AMOC hypothesis has seen refinements over the years. Rather than a simple bistability (on/off, warm/cold or present/glacial; Stommel, 1961; Broecker et al., 1985), the hypothesis has evolved to better fit with proxy evidence suggesting three circulation modes (e.g., Sarnthein et al., 1994), referred to variously as strong/weak/off, warm/cold/off or interstadial/stadial/Heinrich (Alley and Clark, 1999; Clark et al., 2002; Rahmstorf, 2002). In the context of Dansgaard-Oeschger events, freshwater perturbations are usually invoked as a trigger for switching between AMOC states, with the continental ice sheets suggested to be responsible for the storage and release of this freshwater (Broecker et al., 1985; Birchfield et al., 1994). Modelling studies have followed up on this idea (e.g., Manabe and Stouffer, 1995; Ganopolski and Rahmstorf, 2001; Clark et al., 2002; Schmittner et al., 2002; Menviel et al., 2014), as well as offering variations on the theme (e.g., changes in precipitation or atmospheric freshwater flux alter the freshwater balance to drive changes in the overturning circulation; see Stocker and Wright, 1991; Eisenman et al., 2009). By targeting convection regions, a relatively small amount of freshwater can be very effective in shutting down the AMOC (Mignot et al., 2007; Roche et al., 2010). Otherwise, there are studies suggesting that the ocean can act as a "salt oscillator", that is, the ocean circulation can undergo self-sustained oscillations without the need for an external trigger, for example under a different mean salinity balance due to the presence of ice sheets or changing greenhouse gas concentrations (Broecker et al., 1990; Winton, 1993; Schulz et al., 2002; Timmermann et al., 2003; Eisenman et al., 2009; Zhang et al., 2017; Gent, 2018).

Despite dedicated efforts to pin down the AMOC hypothesis, there are still unresolved issues. The growing body of proxy observations undoubtedly helps, but it is not straightforward to interpret the signals in these records (Lynch-Stieglitz, 2017). From the modelling side, a number of serious challenges remain (Clark et al., 2002; Valdes, 2011), most important of which is the fact that one needs long model simulations to capture the millennial-scale nature of the events, yet high enough resolution and enough model complexity to capture the abrupt transitions. Until recently, it was thought that, without freshwater, models simulate a very smooth deglaciation compared to the spiky Greenland signal (Bethke et al., 2012), and that models have trouble reproducing the abruptness of the climate response to recoveries of the meridional overturning (Vellinga et al., 2002; Bitz et al., 2007; Kageyama et al., 2010). While obtaining the correct AMOC stability in models is an ongoing challenge (Gent, 2018), there are now some model simulations that exhibit abrupt, spontaneous events (section 4), opening promising new avenues of research (section 5).

3. An exercise in reverse engineering: how to create Dansgaard-Oeschger signals in Greenland

The launch point of section 2 is the existence of an AMOC instability during glacial times that is the root cause of Dansgaard-Oeschger events. There are multiple lines of evidence for ocean circulation changes associated with the events, some of which suggest that ocean circulation changes lead surface changes (Lynch-Stieglitz, 2017). Despite this evidence, the picture is not clear-cut. The shortest events lack an associated ocean signal (Burckel et al., 2015; Henry et al., 2016), which could be because ocean changes did not occur, or because they were too abrupt to be resolved in the proxy records. AMOC instability seems to exist during warm interglacial climates as well, but unaccompanied by abrupt climate changes (Galaasen et al., 2014). Furthermore, it is unclear mechanistically why surface climate (i.e., Greenland temperature) variability should lag the ocean heat transport variability that goes along with AMOC changes, and whether the necessary AMOC changes can occur without external forcing such as freshwater. Thus, it cannot be ruled out that the inferred AMOC changes are a response to or feedback on Northern Hemisphere climate change rather than a driver.

An alternative approach is to turn the problem around and ask: Given that we observe large climate shifts in Greenland, what is necessary to produce these shifts? In this way, one can identify the ingredients required to create the Greenland shifts, then seek a series of linking mechanisms that account for these, as well as other remote signals associated with Dansgaard-Oeschger events.

The natural geographic focus for such a study is the region surrounding Greenland, from the subpolar North Atlantic up through the Nordic Seas to the Arctic proper. We use the term "Northern Seas", as defined by Eldevik et al. (2014), to refer to this region (Figs. 2 and 3). Warm, salty Atlantic waters flow towards the Arctic with the North Atlantic Current and Norwegian Atlantic Current, losing heat to the atmosphere as they travel polewards. Cold, fresh surface waters flow back from the Arctic through Fram Strait into the East Greenland Current, continuing into the Labrador Current further south. The Labrador Current, the North Atlantic Current and the Irminger Current form a cyclonic circulation system known as the subpolar gyre (Fig. 2). The ocean circulation is set by a combination of atmospheric winds (Fig. 4), surface heat exchange, and the freshwater balance, with the presence of sea ice influencing all these factors (Fig. 3). It is understanding the functioning of this complex dynamical system in the context of Dansgaard-Oeschger-like variability that is the goal of this section.

3.1. Greenland climate shifts require sea ice changes

Sea ice changes are thought to be a key ingredient of Dansgaard-Oeschger events (Gildor and Tziperman, 2003; Li et al., 2005; Dokken et al., 2013). Sea ice is a critical component of the climate system, playing an important role in the surface heat, moisture, and momentum budgets of the polar regions (Fig. 3). Sea ice lowers surface temperatures by insulating the atmosphere from the ocean heat reservoir, and its high albedo reflects incoming solar radiation back to space. The presence or absence of an ice cover can also affect the vertical structure of the upper ocean and deep-water formation sites. In the context of glacial climates, sea ice can regulate the availability of moisture for building continental ice sheets (Hebbeln et al., 1994; Rind et al., 1995). Considering the many associated feedbacks, sea ice has the potential to strongly influence surface climate and to change rapidly in response to relatively weak forcings, which together give it great potential for creating large, abrupt climate shifts in Greenland.

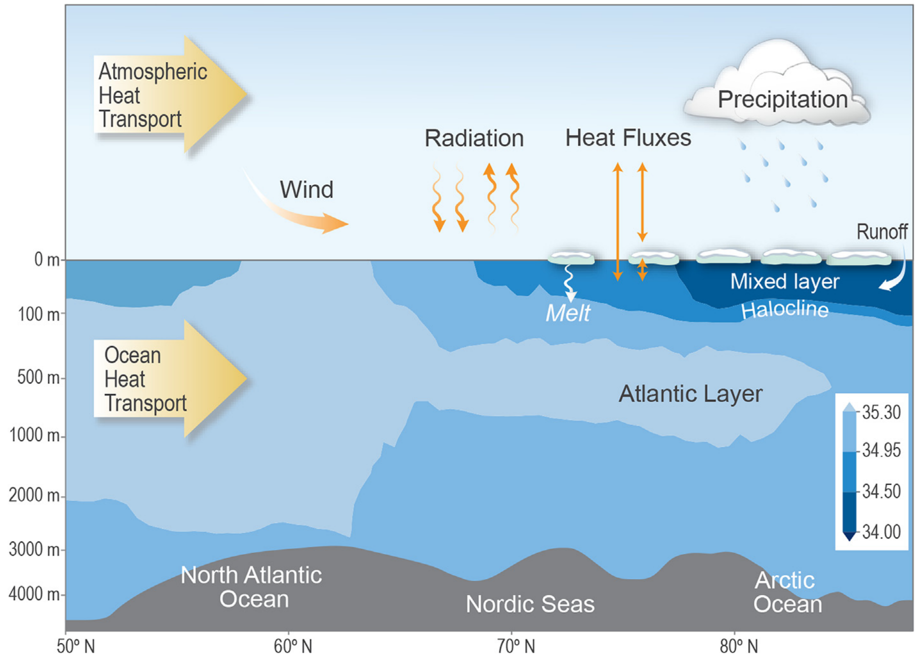

Fig. 3. Schematic of the coupled atmosphere-ice-ocean system through the Northern Seas. Shown are the main exchange processes for heat, freshwater and momentum between the atmosphere, sea ice and ocean, as well as the basic vertical structure of the ocean. Contours show the zonal-mean salinity profile from the SODA reanalysis 1958–2008. Salinity is the main control on density at low temperatures, and thus is a useful field for highlighting water masses in this region.

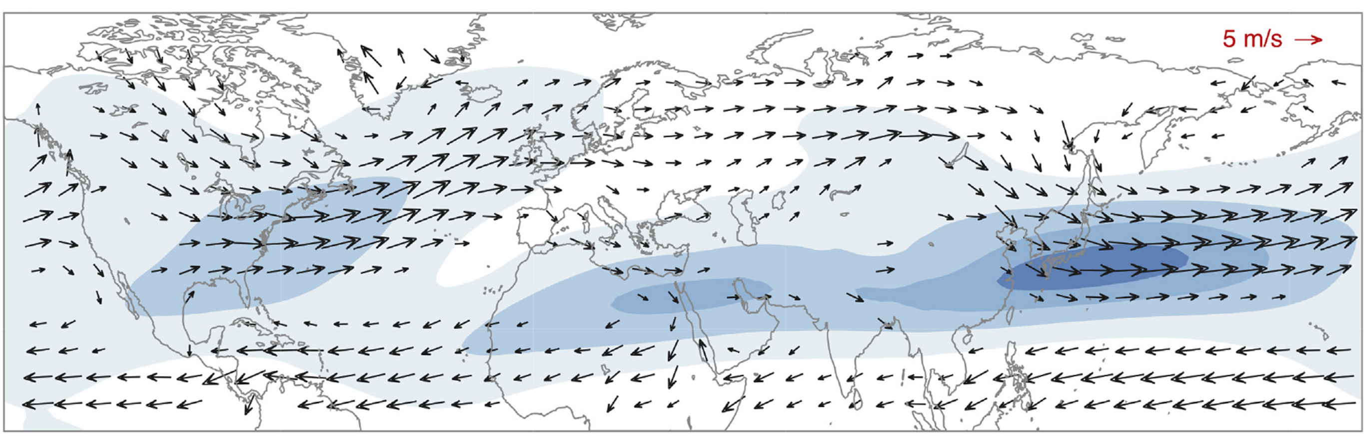

Fig. 4. Wintertime wind patterns in the Northern Hemisphere from ERA-Interim reanalysis (1979–2015). Colour shading shows upper tropospheric winds (DJFM horizontal wind speed at 250 hPa; contour interval 15 m s^{-1}). Vectors show lower tropospheric winds (DJFM horizontal wind at 850 hPa; 5 m s^{-1} reference vector at top right); vectors with magnitude less than 3 m s^{-1} have been omitted for clarity.

Although it is challenging to reconstruct past sea ice variability (de Vernal et al., 2013), there is strong observational evidence that North Atlantic sea ice cover waxed and waned during Dansgaard-Oeschger events. Deuterium excess measurements in the Greenland ice cores exhibit rapid transitions timed with stadial-interstadial transitions, and interpreted as south/north shifts of the moisture source for Greenland snow due to the advance/retreat of sea ice (Masson-Delmotte et al., 2005; Jouzel et al., 2007). Over a dozen or so sites, Greenland ice cores indicate changes in spatial gradients of temperature and accumulation that are also consistent with displacements of the sea ice edge during Dansgaard-Oeschger events (Guillevic et al., 2013). In marine cores on the Norwegian slope, there is evidence for more sea ice formation during cold stadial phases of Dansgaard-Oeschger events than during warm interstadials from benthic δ^{18} O (Rasmussen and Thomsen, 2004; Dokken et al., 2013) and biomarkers (Sadatzki et al., submitted 2018).

Sea ice is an efficient amplifier of climate signals in the context of millennial-scale variability (Gildor and Tziperman, 2003; Timmermann et al., 2003; Kaspi et al., 2004). Sensitivity experiments with climate models indicate that displacements of the winter sea ice edge are key for creating substantial (8–16°C) annual-mean temperature signals consistent with observations of Dansgaard-Oeschger events, in Greenland and across the Euro-Atlantic sector (Renssen and Isarin, 2001; Li et al., 2005). They further suggest that summer sea ice displacements were likely rather limited, otherwise the resulting accumulation signals would be several times larger than suggested by observations. Greenland surface climate appears to be most sensitive to changing ice cover in the Nordic Seas, while changes in the Labrador Sea and subpolar gyre have little additional effect (Fig. 5; see also Li et al., 2010; Merz et al., 2016). In fact, this regional sensitivity has been suggested as a possible explanation for why Heinrich stadials, which involve massive discharges of icebergs into the North Atlantic, "look like" regular Dansgaard-Oeschger stadials in Greenland ice cores (Li et al., 2010), despite having more extreme impacts elsewhere (e.g., major deep water circulation changes, strengthening of monsoon winds, colder conditions; see Hemming, 2004, and references therein).

3.2. Sea ice changes require ocean heat transport changes

Given that Greenland climate shifts require sea ice changes, the next step is to ask what must happen to rapidly and drastically alter the sea ice cover. Due to its location at the atmosphere-ocean interface, sea ice is sensitive to perturbations from above and below (Fig. 3). In addition, sea ice effectively modulates the large fluxes of energy between the atmosphere and ocean, giving rise to strong feedbacks that can cause dramatic migrations of the ice edge, consistent with the abrupt onset of Dansgaard-Oeschger interstadials and stadials.

The ice-albedo feedback produces one of the most well-known local nonlinearities associated with Arctic climate (Winton, 2013). Consider a simplified setup where meridional transports of heat are invariable, and the Arctic Ocean becomes a thermal reservoir in radiative equilibrium. Since the radiation balance strongly depends on albedo and thus on the presence of ice, early studies found that such a system rapidly transitions from an ice-covered state (high albedo, light surface, more radiation reflected away leading to more cooling) to an ice-free state (low albedo, dark surface, more radiation absorbed leading to more warming) under gradual warming, with intermediate states being unstable (North, 1984; Thorndike, 1992).

However, the ice-albedo feedback becomes much less potent once real-world considerations are introduced. One such consideration is that thin ice grows more quickly than thick ice, an effect that counteracts the ice-albedo feedback and stabilizes the ice cover against runaway change (Bitz and Roe, 2004; Eisenman and Wettlaufer, 2009; Notz, 2009). The reason for this is that to create new ice, heat must be lost from the ocean to the atmosphere (see "Heat Fluxes" in Fig. 3). This heat loss happens more readily through thin ice than through thick ice; hence, freezing – which occurs at the base of the ice – accelerates when ice thins, and slows down when ice thickens. (Melt rates do not depend on ice thickness, because melting primarily happens from the top, at the ice-atmosphere interface.) Another consideration is the response of the atmosphere itself to ice perturbations, which also stabilizes the ice cover. For example, the excess heat absorbed by an ice-free surface ocean during summer is returned to the atmosphere in early winter, and removed from the Arctic via enhanced outgoing longwave radiation and reduced atmospheric heat transport from midlatitudes (Winton, 2013; Tietsche et al., 2011).

Processes at and below the ice-ocean interface (lower portion of Fig. 3) can also give rise to rapid changes in sea ice. Today's Arctic ice cover depends on the stable stratification of the surface ocean and the presence of a cold halocline (Rudels et al., 2004), both of which are sensitive to surface freshwater input and ocean heat transport (e.g., Nummelin et al., 2016). Recent studies suggest that the interaction between the stratified ocean and sea ice may provide a mechanism for Dansgaard-Oeschger events (Singh et al., 2014; Jensen et al., 2016): starting from an ice-covered Nordic Seas, heat accumulates below the halocline until the increasing buoyancy of the subsurface layer overcomes the stable stratification above. Vertical mixing erodes the protective freshwater lid and sea ice melts. The excess ocean heat is quickly lost to the atmosphere, allowing ice to start growing again. The halocline is rebuilt by brine rejection from ice growth so that the process continues in a self-sustained oscillation.

But this mechanism alone likely cannot explain the millennial time scale of Dansgaard-Oeschger events without changes in ocean heat transport into the Nordic Seas. In the modern climate, the Nordic Seas receive about 0.4 PW of heat from the inflow of warm waters across the Greenland-Scotland ridge (Fig. 6), and exist in a relatively ice-free state. Jensen et al. (2018) found that an idealized eddy-resolving setup of the Nordic Seas can maintain strong (Arctic-like) stratification and an extensive ice cover, but only with an ocean heat transport of 0.1 PW, less than one quarter of today's value (the perpetual winter conditions used make this an upper

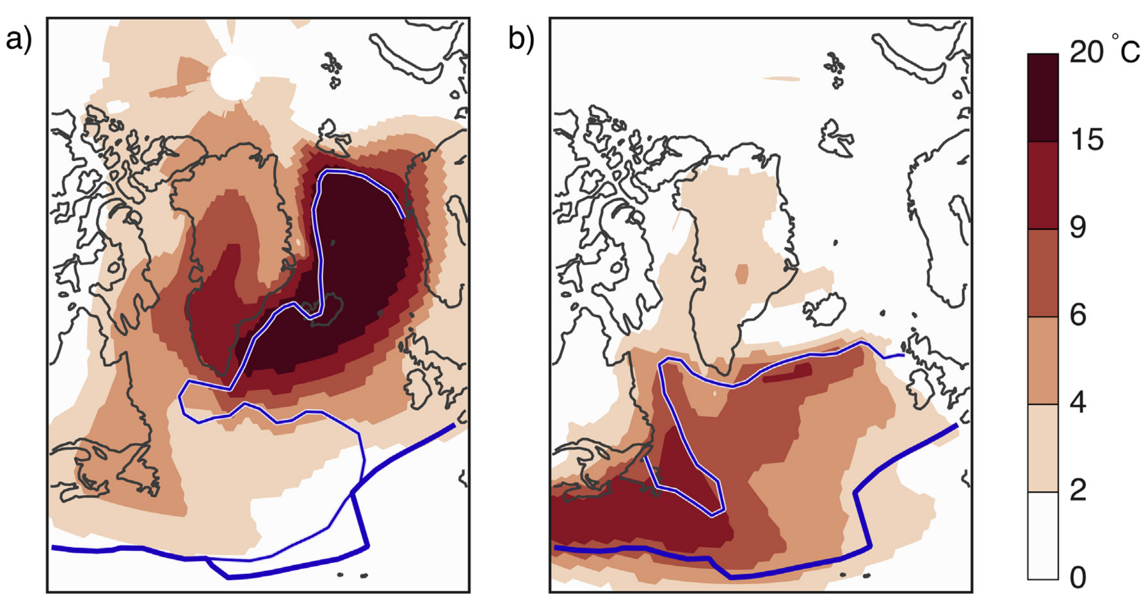

Fig. 5. Simulated warming from sea ice displacements in a glacial climate. Near-surface temperature signal (shading, °C) during the winter half year (NDJFMA) in perturbation experiments from Li et al. (2010) with removal of sea ice in (a) the Nordic Seas and (b) the North Atlantic subpolar gyre region. The dark blue line indicates March sea ice extent (15% concentration) for a cold "stadial" state, and the light blue line indicates March sea ice extent (15% concentration) for a warm "interstadial" (reduced ice) state.

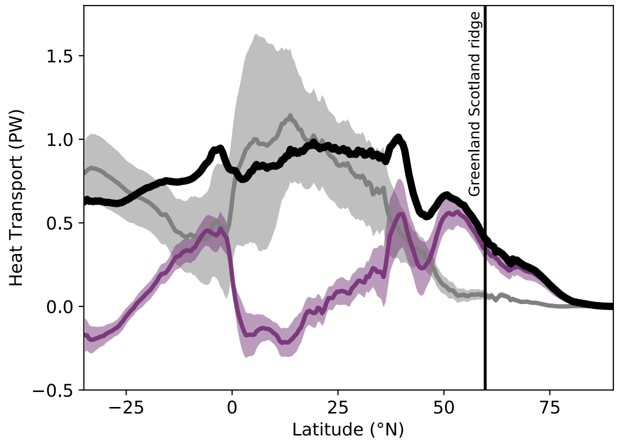

Fig. 6. Ocean heat transport (black) from the SODA reanalysis 1958–2008 separated into overturning (grey) and gyre (purple) contributions (Carton and Giese, 2008). Shading indicates one standard deviation of the annual heat transport in the period 1958–2008. The black vertical line indicates the latitude of the Greenland-Scotland Ridge separating the North Atlantic and Nordic Seas (see Fig. 2).

bound). Even lower values, around 0.005 PW, are required for the one-dimensional ocean-sea ice model of Singh et al. (2014) to produce millennial sea ice variability. The reduced ocean heat transport necessary for the existence of a Nordic Seas ice cover could be physically realizable, since ice growth is associated with decreased ocean heat loss to the atmosphere, warmer subsurface ocean temperatures, and a smaller north-south temperature contrast across the Greenland-Scotland ridge. Thus, much like atmospheric heat transport changes can reduce the impact of the one-dimensional ice-albedo effect, ocean heat transport changes put narrow limits on the role of vertical processes in the halocline.

Beyond the simplified models described above, observations suggest that ocean heat transport is an important driver of present-day sea ice changes, particularly in marginal seas where it is often the dominant source of heat (Segtnan et al., 2011). This is the case in today's Barents Sea, where warm Atlantic waters enter via the Barents throughflow (Smedsrud et al., 2013). Recent ice loss in the region is associated with increased heat transport via a stronger throughflow, a process called “Atlantification” (Årthun et al., 2012). If the marginal zone were further south in the Nordic Seas, as was likely in glacial times, one might imagine a similar Atlantification during transitions from stadials to interstadials (the reverse, in which the polar front advances into the North Atlantic, has been proposed for transitions from interstadials to stadials; see Barker et al., 2015). Today, anomalies in ocean heat transport into the Nordic Seas originate in the subpolar North Atlantic (Holliday et al., 2008; Årthun and Eldevik, 2016).

Finally, coupled climate models suggest that ocean heat transport interacting with sea ice can produce abrupt climate changes in response to weak forcing (Bengtsson et al., 2004; Semenov et al., 2009; Miller et al., 2012; Lehner et al., 2013) or even no forcing (Drijfhout et al., 2013; Sidorenko et al., 2015; Kleppin et al., 2014). A weakening of ocean heat transport stabilizes the halocline, allowing sea ice cover to expand. This mechanism is so strong that it is responsible for much of the intermodel spread in Arctic sea ice cover in climate models (Mahlstein and Knutti, 2011).

3.3. *Atmospheric effects and interactions*

The previous subsections have established that Dansgaard-Oeschger signals on Greenland require wintertime Nordic Seas ice cover to change, which in turn requires ocean heat transport to change. A final consideration is the atmosphere. As already alluded to in section 3.2, the atmosphere is coupled to the underlying ice and ocean by exchanges of heat, freshwater and momentum (Fig. 3). The large-scale atmospheric feature of interest in our focus region is the North Atlantic jet stream, a belt of strong westerly (west-to-east) winds reaching speeds over 60 $\mathrm{m\,s^{-1}}$ during the winter season (Fig. 4). Closely associated with the jet stream is the North Atlantic storm track, where midlatitude weather systems tend to travel following the prevailing westerlies.

It is well known that mean wind patterns are important in setting the circulation of the upper ocean, in particular the position and strength of the gyres. Poleward of the jet stream, in the subpolar region, there is a cyclonic wind stress curl (that is, the applied force of the winds imparts a counter-clockwise rotation to the ocean) that drives the cyclonic Atlantic subpolar gyre (Fig. 2). The strength of the subpolar gyre can be quantified via the Sverdrup transport, which shows a peak of around 30 Sv ($1\ \mathrm{Sv} = 1\times10^6\ \mathrm{m^3 s^{-1}}$; see Fig. 5.17 in Talley et al., 2011 and Fig. 9). Equatorward of the jet stream, the wind stress is anticyclonic, giving rise to the clockwise subtropical gyre.

Other air-sea interactions can further affect the ocean circulation by altering buoyancy. For example, the surface ocean becomes denser when it experiences a net heat loss to the atmosphere (via incoming radiation, more outgoing radiation, stronger turbulent fluxes) or a net loss of freshwater (via less precipitation, more evaporation, less runoff from rivers and continental ice sheets). Such processes acting in the subpolar region would spin up the gyre, thereby increasing poleward heat transport towards the Nordic Seas (see section 4.3 for more details; Born and Stocker, 2014). In the opposite direction, a net gain of heat or freshwater would make the surface ocean less dense and weaken the gyre.

Present day variability in the jet stream is important for the climate of the Northern Seas as well. In the North Atlantic sector, the jet stream exhibits frequent, large meridional displacements, fluctuating between three preferred positions – northern, central and southern (Woollings et al., 2010). A modelling study by (Czaja, 2009) suggests that the erratic nature of the jet helps maintain a strong subpolar gyre and deep water formation in the region. The mechanism in his kinematic model is that variable winds drive expansion/contraction of the subtropical and subpolar gyres, allowing for efficient export of salt from the former to the latter. A southern jet tends to have the opposite effect on the subpolar gyre because it favours the occurrence of a phenomenon called Greenland blocking (Madonna et al., 2017), during which sea-level pressure is anomalously high over the subpolar North Atlantic. The anomalous high pressure is associated with an anticyclonic (clockwise) circulation that has been shown to export more freshwater and sea ice from the Arctic, thereby freshening the subpolar gyre surface waters (Ionita et al., 2016). According to the discussion in the preceding paragraph, this should act to weaken the gyre and its associated heat transport into the Nordic Seas. We will revisit the subject of preferred jet stream positions and atmospheric blocking in the context of glacial climates in section 4.2.

Finally, atmospheric forcing drives Arctic sea ice variability over a wide range of time scales, from weekly to decadal (e.g., Fang and Wallace, 1994; Venegas and Mysak, 2000; Deser et al., 2002). The sea ice variability exhibits different dominant time scales in different regions of the Arctic depending on the mechanism, or combinations of mechanisms, that drive it: fast variability is directly due to wind forcing, while slower variability comes through changes in gyres or ocean advection (Venegas and Mysak, 2000).

In summary, the atmosphere impacts both ocean and sea ice, but the sea ice cover also modulates the effect of the atmosphere on the ocean (Fig. 3). If sea ice is present, the ocean does not feel the atmosphere directly. Thus, atmosphere-ocean exchange through a variable sea ice cover influences ocean circulation and heat transport along a continuous looped pathway through the Northern Seas (Fig. 2).

4. Abrupt events in the Northern Seas coupled system

In this section, we explore how the atmosphere, sea ice and ocean may interact to produce Dansgaard-Oeschger events. We first review existing studies documenting spontaneous abrupt climate changes in model simulations, and the coupled dynamics proposed to explain them. Then, we explore reasons why such spontaneous abrupt events might be favoured during glacial periods compared to interglacial periods. In other words, can a glacial background climate state alter the Northern Seas coupled system such that Dansgaard-Oeschger events become possible?

4.1. Mechanisms for spontaneous abrupt climate change

A growing number of simulations from coupled high-resolution climate models exhibit spontaneous abrupt climate changes that resemble Dansgaard-Oeschger events. One of the first was an unforced pre-industrial control simulation with the EC-Earth model, which exhibits an abrupt cooling event that lasts for more than one century before a similarly abrupt warming (Drijfhout et al., 2013). The mechanism starts with (1) an initial cooling due to anomalous atmospheric blocking over the eastern subpolar gyre (recall section 3.3). (2) Sea ice cover expands, creating a high sea-level pressure anomaly. (3) Continued sea ice expansion around Cape Farewell at the southern tip of Greenland introduces a freshwater perturbation that weakens deep ocean convection. (4) Surface waters that are now exposed longer to cold atmospheric temperatures freeze, reinforcing the positive sea ice anomaly and the thermally forced high-pressure anomaly. Even small changes in sea ice export into the subpolar region affect deep convection because the freshwater perturbation reaches the Labrador Sea surface in late winter (ideal location and timing), and because of strong amplifying feedbacks by the subpolar gyre circulation (Born et al., 2010; Jochum et al., 2012). This series of positive feedbacks continues for several decades, with climate impacts progressing across the Nordic Seas towards Svalbard. The subsequent abrupt warming follows the same mechanism in reverse.

A virtually identical mechanism was found to cause abrupt climate change in the Community Climate System Model version 4 (CCSM4), also in an unforced pre-industrial control simulation (Kleppin et al., 2014). Here, the coupled feedbacks include the ocean circulation as the decreased deep convection in step (3) weakens the subpolar gyre. This results in less heat and salt supplied to the western North Atlantic basin by the Irminger Current (Fig. 2), which further stabilizes and strengthens the positive sea ice anomaly (Born et al., 2016). The cold conditions last for 200 years, and the warming event is more abrupt than the cooling event.

To our knowledge, at least two more state-of-the-art coupled models exhibit similar behaviour under pre-industrial boundary conditions, ECHAM6-FESOM (Sidorenko et al., 2015) and the Kiel Climate Model KCM (Martin et al., 2015). The multiple “Labrador freezing” events in ECHAM6-FESOM are associated with an increase in Arctic freshwater export, but not necessarily caused by it, leading the authors to conclude that the dynamics of the subpolar gyre play an important role, possibly aided by changes in the deep ocean that impact the stability of the convective water column. Stability is also key for the transitions in KCM, which start in the Southern Hemisphere. Periodic decreases in deep water formation in the Weddell Sea reduce the supply of Antarctic Bottom Water, and increase the southward flow of North Atlantic Deep Water. The change in ocean hydrology favours deep convection in the Labrador Sea, which strengthens the subpolar gyre (by thermal wind; see section 4.3). In contrast to the irregular, spontaneous events in the other models, the interhemispheric teleconnection in KCM has quasi-centennial cyclicity. The ECHAM6-FESOM and KCM studies do not focus on an active role for the atmosphere, unlike the EC-Earth and CCSM4 studies.

Under full glacial conditions, the Community Earth System Model version 1 (CESM1) shows regular cycles of Dansgaard-Oeschger-like oscillations (Peltier and Vettoretti, 2014; Vettoretti and Peltier, 2015, 2016, 2018). CESM1 belongs to the same model family as CCSM4 used in Kleppin et al. (2014). As in the pre-industrial simulations, coupling between the atmosphere, sea ice and ocean is critical for the oscillation. The warm events are initiated by the opening of a large polynya in the Irminger Sea (Vettoretti and Peltier, 2016), a sensitive region for central Greenland temperatures and precipitation (Li et al., 2010; Merz et al., 2016). The transition itself is very fast (occurs in a single winter season) owing to accumulated heat at intermediate ocean depths, which melts ice and warms the atmosphere as soon as the water column becomes unstable. However, the instability depends on salt transport to the subpolar gyre through Denmark Straight and the Irminger Current in the decades prior to the abrupt warming. An additional preconditioning for the convection event is the overflow of extremely dense waters from the Nordic Seas (formed from brine rejected during sea ice formation during the cold stadial) which influences deep convection, the subpolar gyre, and heat/salt transport in the Irminger Current (Levermann and Born, 2007; Born et al., 2009; Born and Stocker, 2014), as well as the sensitivity of the subpolar gyre to wind forcing changes (Montoya et al., 2011). The simulated coolings are more gradual than the warmings, in agreement with observations (Fig. 1). One caveat with the oscillations in CESM1 is that the sea ice grows to several tens of metres in thickness. Also, the use of full glacial boundary conditions presupposes a Barents Sea covered by land ice, with sea ice export from the central Arctic Ocean restricted to the Fram Straight. For Marine Isotope Stage 3, when Dansgaard-Oeschger events were most pronounced, ocean pathways through the Barents Sea were likely open (Mangerud et al., 1998; Jakobsson et al., 2014; Hughes et al., 2015).

The unforced glacial oscillations in CESM1 are unique to date, but a handful of other models show evidence of similar dynamics under glacial boundary conditions. A simulation with the COSMOS model exhibits abrupt climate changes when forced with a gradual increase in the height of the Laurentide ice sheet (Zhang et al., 2014). Increasing the ice sheet height shifts the winds northwards, and makes the Labrador Sea more saline both by increasing salt advection into the region, and decreasing the import of freshwater/sea ice from the Arctic. At some critical ice sheet height, there is an onset of open ocean deep convection; for moderate ice volumes between 0.25 and 0.45 times the glacial maximum, there are two stable AMOC modes. The CM2Mc model, a coarse resolution GCM, simulates unforced oscillations of the AMOC under an interesting setup with 180 ppm CO_2 concentration and pre-industrial ice sheet topography (Brown and Galbraith, 2016). We speculate that the low greenhouse gas concentration sets the sea ice edge at an optimal location to engage feedbacks similar to those described above. Brown and Galbraith (2016) also cite salt advection as being important for the oscillations, in particular the exchange of salt between the subtropical and subpolar North Atlantic. Their simulations with higher CO_2 levels and different ice sheet topographies do not oscillate, and the mechanism is found to be sensitive to the axial tilt of the earth, which strongly affects the seasonality of the polar regions.

Considering this body of literature as a whole, coupled climate models demonstrate that coupled atmosphere-ice-ocean feedbacks can cause abrupt climate change in the Northern Seas. Critical features are the dynamics of the subpolar gyre, local wind feedbacks and atmospheric blocking, and the advection of freshwater through the Northern Seas, in particular by sea ice. The subpolar North Atlantic is a key meeting point for these three components, because this is where vigorous wind systems encounter the southernmost extension of sea ice and the most variable currents of the North Atlantic current, with connections to the deep ocean via convection. This suggests that climate models must correctly resolve details of this region and the interplay of processes at the atmosphere-ice-ocean boundary for a successful simulation of Dansgaard-Oeschger events or similar types of abrupt climate change. Subpolar gyre dynamics depend on comparatively small-scale details in the Labrador Sea (Spall, 2004; Straneo, 2006), which coarser resolution models fail to capture (perhaps compensating with AMOC changes instead; see section 4.2 for related discussion on AMOC biases). The ability to simulate atmospheric blocking depends somewhat on model resolution, which may explain the lack of abrupt climate change in previous generations of climate models, but also on a model's mean-state bias (Scaife et al., 2010; Dawson et al., 2012; Anstey et al., 2013).

The existence of abrupt climate changes in both pre-industrial and glacial simulations remains somewhat of a puzzle, although the explanation seems to lie in whether a model's subpolar region can enter the hypothesized feedback loop in any given experimental setup. We revisit this discussion in section 5.

4.2. Glacial winds set the background state

The previous section suggests that, instead of asking what triggers Dansgaard-Oeschger events in a glacial climate, we could ask what makes a glacial climate able to support these oscillations. A natural starting point is what is widely accepted to be the clearest difference between glacial and interglacial climates from modelling studies: winds. This section summarizes the expected wind changes and discusses the broader effects of these changes on the Northern Seas. It sets the stage for the next section, where we investigate how these wind changes influence the subpolar gyre and the stability of the atmosphere-ice-ocean coupled system. A point to bear in mind throughout this discussion is that, while climate models should not be taken at face value, there are some aspects we can trust them to simulate well. It is these aspects we should rely on in our exercise of building hypotheses for Dansgaard-Oeschger events.

Model simluations depict a different atmospheric circulation in glacial climates compared to interglacial climates. Notable differences in the Northern Hemisphere include an amplified stationary wave over North America (Manabe and Broccoli, 1985; Cook and Held, 1988); stronger winds (Li and Battisti, 2008; Pausata et al., 2011; Hofer et al., 2012; Ullman et al., 2014; Löfverström et al., 2014; Merz et al., 2015); and altered large-scale variability patterns (Justino and Peltier, 2005; Pausata et al., 2009; Rivière et al., 2010). The transient circulation (e.g., the circulation associated with baroclinic eddies that form passing weather systems) is also altered, but the differences here are more difficult to generalize. For example, consider the storm tracks, regions where midlatitude weather systems are most frequently found. Looking at the Northern Hemisphere as a whole, the differences between the glacial storm tracks compared to today's storm tracks depend on the geographic region of interest, but also somewhat on the model and boundary conditions used as well as the metric chosen to define the storm track (e.g., Kageyama and Valdes, 2000; Justino et al., 2005; Laîné et al., 2009).

Homing in on the North Atlantic, glacial-interglacial differences become more robust across the model simulations. The LGM jet stream in this sector is generally described as more zonal, equatorward-shifted and more intense (Fig. 7), although these characteristics may be more or less obvious depending on the model used (Li and Battisti, 2008; Hofer et al., 2012; Pausata et al., 2011; Ullman et al., 2014; Löfverström et al., 2014; Merz et al., 2015). The intense, zonal jet is mainly due to the presence of the Laurentide ice sheet, which has elevations reaching a maximum of approximately 3000 m (up to several hundred metres' difference depending on the reconstruction; see Abe-Ouchi et al., 2015, Fig. 1). As a result of the topographic forcing, winter stationary waves show evidence of altered propagation (Cook and Held, 1988; Held et al., 2002) and refraction (Löfverström et al., 2016) behaviour over the glacial North Atlantic. The thermal forcing of the ice sheet's white surface has little influence on winter circulation changes, but is important in determining the summer stationary waves (Ting, 1994; Roberts et al., submitted 2018).

In addition, the North Atlantic glacial jet appears less “wobbly” than it is today due to differences in wave-mean flow feedbacks (Rivière et al., 2010) that weaken the eddy-driving of the jet (Li and Wettstein, 2012; Rivière et al., 2018). The jet tends to fluctuate less north-south, and as a consequence, the North Atlantic Oscillation as we know it does not exist as the leading mode of regional variability (Pausata et al., 2009; Rivière et al., 2010). The glacial storm track is consistently narrower (meridionally confined) at the upstream end (Laîné et al., 2009; Rivière et al., 2010; Merz et al., 2015; Löfverström et al., 2016), and exhibits a slight southeastward extension towards Europe at the downstream end, consistent with a more zonal jet stream (Laîné et al., 2009; Merz et al., 2015; Löfverström et al., 2016). The narrowing manifests as weaker area-averaged storm activity over the eastern North Atlantic (Li and Battisti, 2008); elsewhere, the changes are less coherent, with localized increases and decreases in storminess due to differences in SST and sea ice (Merz et al., 2015; Löfverström et al., 2016), and upstream seeding (Donohoe and Battisti, 2009).

The altered winds are expected to have broader implications for the Northern Seas as a whole, potentially playing a critical role in priming the system to support abrupt events (Seager and Battisti, 2007; Wunsch, 2006). First, a strong and equatorward-shifted North Atlantic jet weakens atmospheric heat transport into the North Atlantic (van der Schrier et al., 2007) and favours episodes of Greenland blocking (Madonna et al., 2017), both of which could trigger the atmosphere-ice-ocean feedbacks that create abrupt climate changes in this region (see section 4.1). Second, a steadier North Atlantic jet produces steadier wind forcing at the ocean surface. The interglacial jet is highly variable (Fig. 8a and b showing 1980–2013 period from reanalysis data and pre-industrial simulation PI from Table 1), with the maximum wind location wandering freely over approximately 30° of latitude during the winter season. In the glacial climate (Fig. 8c), the maximum wind location spans a much narrower range of approximately 10° of latitude, and the wind itself is stronger than in interglacial climates (seen via the more continuous nature of the darkest filled contour at $15\,m\,s^{-1}$). Steadier wind forcing makes for stronger Sverdrup transport, as shown in pre-industrial and LGM simulations by nine PMIP3 models (Fig. 9 and Table 2), which in turn strengthens the wind-driven component of the subpolar gyre. These effects are explored quantitatively with a box model in the next subsection, and turn out to be important in setting the stability of the Northern Seas.

Although the ocean overturning circulation itself is not a focus of this review paper, it bears mention. In terms of reconstructing the overturning circulation at Last Glacial Maximum, the currently available data allow for different scenarios (Lynch-Stieglitz et al., 2007), but generally point to a shallower overturning cell of

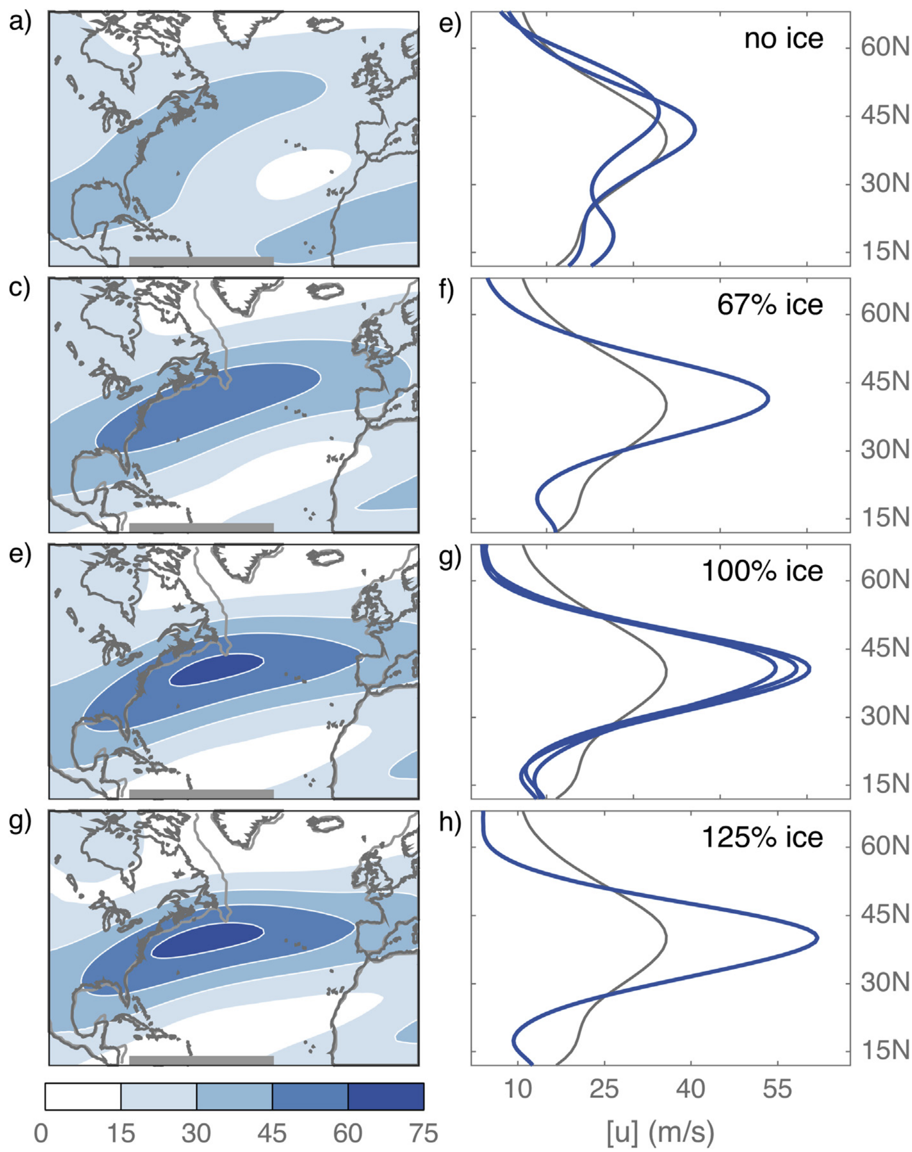

Fig. 7. North Atlantic jet stream from simulations with continental ice sheets of varying sizes. Left: Winter jet (DJFM zonal wind at 250 hPa; contour interval 15 m s^{-1}) with (a) no ice sheets; (b) ice sheets scaled to 67% of their LGM height; (c) LGM ice sheets; (d) ice sheets scaled to 125% of their LGM height. Right: Jet profiles (DJFM zonal wind at 250 hPa averaged over 35W–80W) for simulations with (e) no ice sheets; (f) ice sheets scaled to 67% of their LGM height; (g) LGM ice sheets; (h) ice sheets scaled to 125% of their LGM height. In each panel, the blue curves show jet profiles from simulations for the indicated ice sheet configuration, and the grey curve shows the profile from ERA-Interim reanalysis data. Details of the simulations appear in Table 1.

similar or slightly weaker strength compared to today (McManus et al., 2004; Hesse et al., 2011; Lippold et al., 2012; Gebbie, 2014; Boer, 2015). This does not compare well to results from climate models. The previous generation of models produced quite disparate results (Otto-Bliesner et al., 2007; Weber et al., 2007), while the newer generation of models generally agrees on a stronger and deeper glacial overturning circulation (Muglia and Schmittner, 2015). The stronger overturning in models has been linked to the presence of glacial ice sheets (Brady et al., 2013; Zhang et al., 2014; Zhu et al., 2014; Gong et al., 2015; Brown and Galbraith, 2016), specifically, their role in intensifying wind stress over the North Atlantic (Oka et al., 2012; Sherriff-Tadano et al., 2018). Interestingly, the only model of the ones shown in Fig. 9 that simulates a shallower glacial AMOC is CCSM4, which belongs to the model family that produces spontaneous abrupt changes in both interglacial and glacial climates (Kleppin et al., 2014; Peltier and Vettoretti, 2014; Vettoretti and Peltier, 2016, 2018). The CCSM4 also simulates a glacial AMOC that is only slightly (9%) stronger than its pre-industrial AMOC, compared to much (17–82%) stronger glacial AMOCs in the other models (see Muglia and Schmittner, 2015, Table 1).

Model-data discrepancies such as this can engender a general mistrust of climate models, but from a dynamical viewpoint, this particular discrepancy is not so surprising. The meridional overturning circulation results from a complicated interplay of many different processes; as such, it is understandable that models may produce different answers or “wrong” answers for a given set of boundary conditions. Possible explanations for the discrepancy are that the glacial simulations are not equilibrated (Zhang et al., 2014); that the models are overly sensitive to the stronger, more equatorward mean winds (Muglia and Schmittner, 2015); or that the models are not sensitive enough to other glacial forcings (e.g., buoyancy fluxes, reduced gyre exchange, more sea ice export into the North Atlantic) that would tend to weaken the overturning

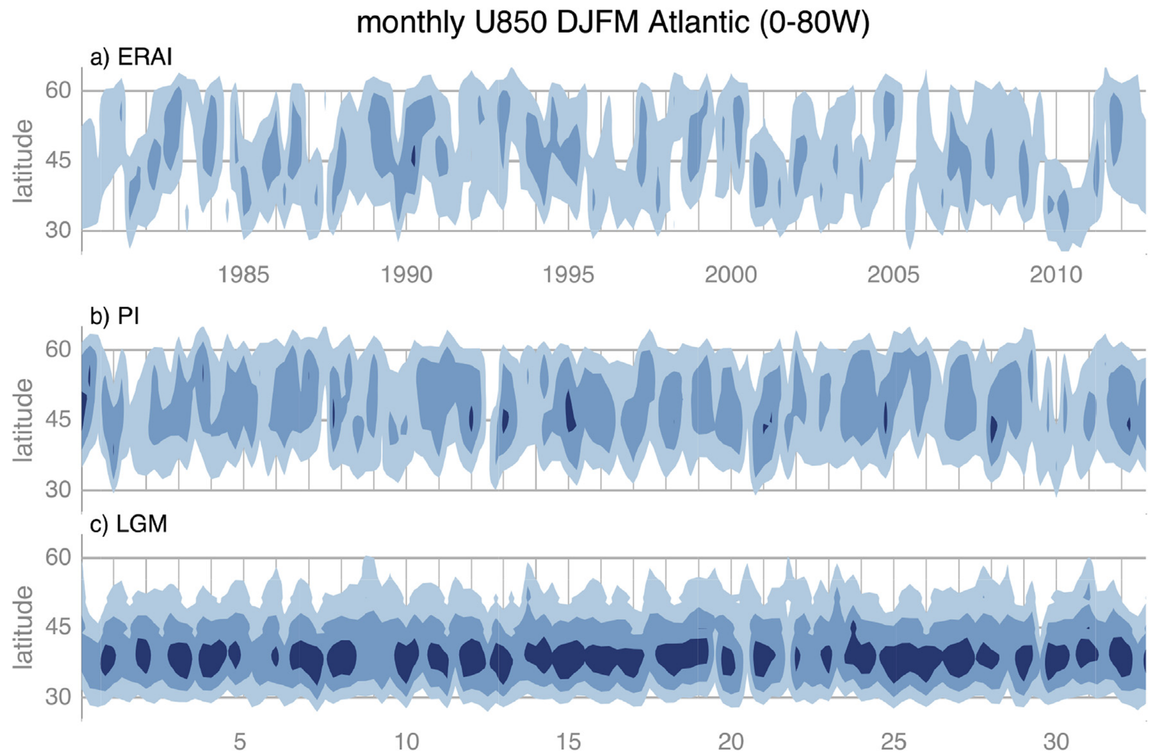

Fig. 8. North Atlantic jet variability in interglacial and glacial climates. Time-evolution of winter jet profiles (monthly DJFM zonal wind at 850 hPa averaged over 0–80W; contour interval 5 m s^{-1} with the zero contour omitted) in (a) ERA-Interim reanalysis, (b) the PI simulation from CCSM4, and (c) the LGM simulation from CCSM4. The x-axis shows actual years for panel (a) and simulation years for (b) and (c). Vertical grey lines separate each winter season.

(Czaja, 2009; Muglia and Schmittner, 2015; Sherriff-Tadano et al., 2018). For example, recall that today's erratic Atlantic jet is thought to be important for maintaining salty subpolar conditions and an overturning circulation in the Atlantic basin (section 3.3). The glacial Atlantic jet is further equatorward, stronger and steadier – similar to the wind belt in the North Pacific today, which may be responsible for fresher subpolar conditions and the lack of overturning in that basin (Czaja, 2009; see in particular Fig. 1a). If an ocean model does not respond "correctly" to the steadier glacial jet, it might maintain a healthy transport of salt into the subpolar North Atlantic; together with the enhanced surface cooling expected during glacial times, the result would be a stronger AMOC. Finally, Muglia and Schmittner (2015) suggest that some proxies may require reinterpretation; for example, a record at a certain site may not be able to distinguish between a glacial Gulf Stream that is weaker rather than simply shifted away from its modern-day position (Gong et al., 2015).

To wrap up this section, we return to the fact that differences in large-scale winds are among the clearest features of glacial simulations compared to interglacial simulations. The altered wind patterns are largely determined by the presence of ice sheets, and governed by well-studied concepts in atmospheric dynamics. Observational uncertainties in the size and geometry of ice sheets exist, but the presence of North American land ice of any appreciable size will result in a strengthening of the surface wind stress curl over the North Atlantic (Fig. 9), which produces a stronger subpolar gyre (e.g., Fig. 9 and Gregoire et al., 2018). How the altered wind forcing impacts the subpolar gyre is grounded in fundamental physics (Sverdrup theory), unlike for the AMOC, where the relationship is rather indirect (Montoya and Levermann, 2008). In the next section, we build on this established understanding of the wind-driven circulation, using clues from recent work (section 4.1) that add a non-linear twist to our textbook knowledge.

4.3. *Insights from a subpolar gyre box model*

The dynamics of the subpolar gyre have been summarized in a simplified box model (Born and Stocker, 2014) that allows us to test the impact of boundary condition changes as inferred from LGM simulations, in particular, the increased wind stress curl (section 4.2). The model consists of four boxes: two stacked cylindrical boxes forming a convective basin, and two stacked annular boxes for the boundary currents surrounding it (Fig. 10). The upper layer has a depth of 100 m to include the surface current, while the lower layer reaches down to 1500 m, a typical depth for convective mixing in this region (Yashayaev, 2007). This simplified configuration is a good approximation for the water mass distribution at the WOCE AR7 section across the Labrador Sea (Marshall and Schott, 1999; Straneo, 2006; Yashayaev, 2007). The box model is forced by wind, surface cooling and surface freshwater flux due to precipitation and sea ice melt (light blue and dark blue arrows in Fig. 10). The temperatures and salinities of the outer boundary currents are fixed, while for the inner two boxes, these properties are simulated prognostically by the model. A full description of the model set-up, its dynamics and its representation of recent decadal variability can be found in Born and Stocker (2014) and Born et al. (2015).

The box model has two stable solutions over a range of realistic forcing. Consider an initially weak gyre in today's climate. The boundary current strength has a barotropic component, set primarily by winds (Sverdrup transport), and a baroclinic component, set by the density contrast between the central basin and the boundary (thermal wind). As the freshwater flux decreases, surface waters become denser so that wintertime surface cooling promotes deep convection (i.e., mixing between the top and bottom boxes of the central basin). Convection cools the central basin to enhance the radial density gradient and strengthen the baroclinic flow

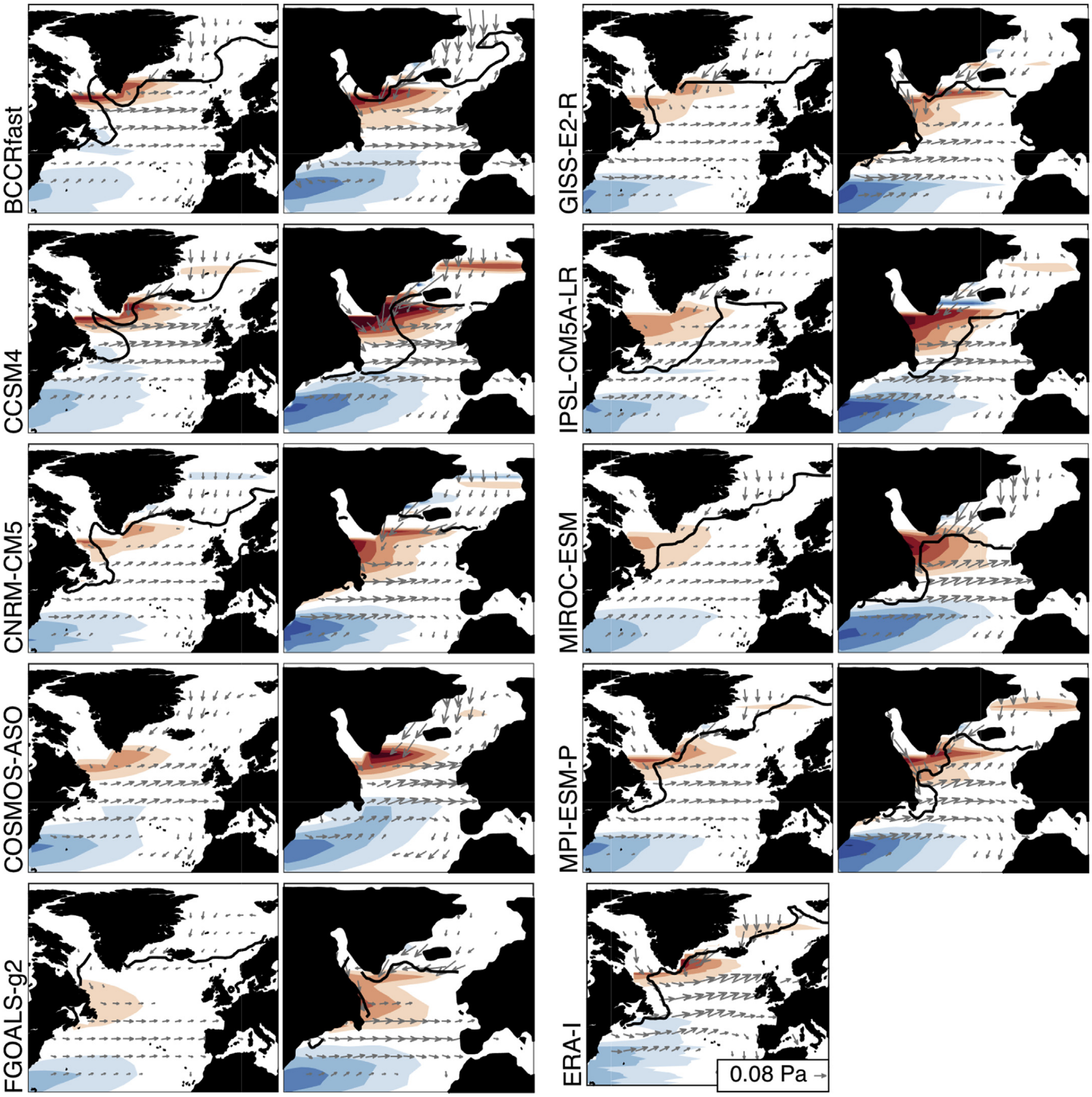

Fig. 9. Windstress (vectors) and Sverdrup transport (shading; 10 Sv intervals) for pre-industrial and glacial climates. Each two-panel block shows results from one model, with the pre-industrial simulation on the left and the LGM simulation on the right; the single panel at the bottom right is ERA-Interim. Solid black contours show March sea ice extent (15% concentration). The Sverdrup transport is calculated by integrating the wind stress curl across the ocean basin from east to west.

component of the boundary current (purple box). The now stronger current sheds more eddies from the saline boundary into the centre of the gyre (upper grey box), making it even denser and more prone to convect. This combination of convective and advective processes, once activated, is strong enough to counteract surface freshwater fluxes that would otherwise inhibit deep convection. The gyre transitions to a strong circulation mode. Shortcomings of the box model are that it only represents the local increase of salt flux by eddies and not the long-range advection of salt in the Irminger Current that could further strengthen the baroclinic flow (Born et al., 2016), and that it does not account for changes in sea ice cover. However, the same advective-convective dynamics have been found in high-resolution ocean models (Spall, 2012) and comprehensive climate models (Born and Mignot, 2012; Born et al., 2013).

Furthermore, the advective-convective feedback gives rise to a hysteresis between the weak and strong gyre modes with respect to freshwater (Fig. 11a, black curve). The dynamics behind this hysteresis are similar to those behind the classical AMOC bistability (Stommel, 1961; Marotzke, 2000), both conceptually and in terms

Table 1
List of CAM4 simulations with different ice sheet configurations. Present day levels are denoted pd, pre-industrial (1850) levels are denoted pi, LGM (21 ka) levels are denoted lgm, and MIS4 (65 ka) levels are denoted mis4. Simulations in boldface are those shown in the left column of Fig. 7. SST and sea ice fields are outputs of corresponding fully coupled CCSM3 simulations. See Merz et al. (2015) for additional details, including an illustration of the different ice sheets (their Fig. 1).

Simulation	Ice sheets	Orbital parameters	SST/ice	CO_2 (ppm)	CH_4 (ppb)	N_2O (ppb)
PI	pd	pd	pi	280	760	270
LGM_{PD}	pd	lgm	lgm	185	350	200
$MIS4_{67}$	67% lgm	mis4	mis4	205	460	210
PI_{LGM}	lgm	pd	pi	280	760	270
LGM	lgm	lgm	lgm	185	350	200
$MIS4_{LGM}$	lgm	mis4	mis4	205	460	210
$MIS4_{125}$	125% lgm	mis4	mis4	205	460	210

of the actual governing equations. Given that the subpolar gyre accounts for most of the ocean heat transport into the Nordic Seas today (Fig. 6), could variability due to this hysteresis help explain the rapid transitions and the persistent stadial/interstadial periods of Dansgaard-Oeschger events? We know that in glacial times, as the Laurentide ice sheet grows, the wind forcing increases (Fig. 9). In the box model, stronger wind forcing (light blue arrow in Fig. 10) actually weakens the hysteresis – the height and width (Fig. 11b, colour shading and horizontal extent of shaded region, respectively) of the hysteresis region shrink (also seen by comparing the black and red curves in Fig. 11a). But there are two additional effects of interest. First, stronger wind forcing increases the barotropic flow component of the gyre, which produces a much stronger gyre. Second, it makes the baroclinic flow component less important; for wind forcing corresponding to the full glacial simulation in Fig. 7, baroclinic transport in the box model contributes only 2.5–5 Sv or approximately 5–15% of the ~30 Sv Sverdrup transport, compared to 30–45% for a present-day setup (see values corresponding to positions of black dots in Fig. 11b and c). Together, these results show that the advective-convective feedback on heat and volume transports becomes weaker in glacial climates compared to today.

While gyre hysteresis alone appears to be insufficient to account for spontaneous glacial oscillations, the box model results offer an alternative explanation stemming from the effect of wind forcing on the Sverdrup transport itself. Glacial boundary conditions create a generally colder North Atlantic, with altered wind patterns that promote fresher subpolar conditions and Greenland blocking (sections 3.3 and 4.2) – all of which encourage sea ice to grow. It is difficult to quantify exactly the effect of sea ice expansion on the Sverdrup transport, but we expect a reduction, both (1) directly by reducing momentum transfer between ocean and atmosphere, and (2) indirectly by creating a cold-core surface high (section 4.1) that imparts an anticyclonic anomaly to the wind forcing on the ocean. Thus, there is potentially a much larger difference in subpolar gyre strength between an ice-free glacial state (very strong wind stress curl, very strong gyre) and ice-covered glacial state (ocean decoupled from wind forcing, very weak gyre) than between the two branches of today's hysteresis curve (moderately strong or moderately weak gyre). Because this possibility depends on atmosphere-ice-ocean processes to create and maintain transitions between the two glacial states (section 4.1), we interpret the box model results as supporting climate models in identifying coupled dynamics as key for Dansgaard-Oeschger events.

5. Discussion

The intention of this review paper was to revisit the question "What causes Dansgaard-Oeschger events?" by considering the coupled atmosphere-ice-ocean system of the Northern Seas. We see clear potential for progress. The simulated abrupt events documented in section 4.1 all invoke positive feedbacks between the atmosphere, sea ice and ocean to produce and sustain their Dansgaard-Oeschger-like climate swings. That the various studies focus on different starting points of the feedback loop is, in our opinion, secondary. Whether the events are kicked off by stochastic wind forcing (Drijfhout et al., 2013; Kleppin et al., 2014), gradual changes in ice sheets (Zhang et al., 2014), or convective instability (Sidorenko et al., 2015; Vettoretti and Peltier, 2016), the point is that they develop thanks to coupled interactions within the Northern Seas. The study of Brown and Galbraith (2016) provides additional support by examining climate responses to overturning "interruptions", both freshwater-forced (hosed) and not (unhosed), under a range of background climate states in one model. They find that the first-order response pattern is quite consistent amongst all simulations, suggesting that it is heavily shaped by atmosphere-ice-ocean dynamics inherent to the model, regardless of whether or not freshwater input is the cause. Freshwater, however, does lead to a more complete shutdown of the AMOC, with amplified global impacts evocative of Heinrich stadials.

In this coupled framework, sea ice is critical beyond its role in creating large temperature swings on Greenland; it also controls the strength of atmosphere-ocean coupling. Throughout this review (see in particular section 3), we have emphasized that sea ice responds to changes in both the atmosphere (e.g., Labrador Sea ice cover expands during prolonged periods of Greenland blocking)

Table 2
List of PMIP3 models used in the study, including basic information on the atmosphere and ocean model resolutions (longitude grid points, latitude grid points, vertical levels).

Model	Institution	Atmosphere grid	Ocean grid
BCCRfast	Bjerknes Centre for Climate Research and Norwegian Climate Centre	96×48 × L26	100×116 × L32
CCSM4	National Centre for Atmospheric Research, USA	288×192 × L26	320×384 × L60
CNRM-CM5	Centre National de Recherches Météorologiques and Centre Européen de Recherche et Formation Avancées en Calcul Scientifique, France	256×128 × L31	362×292 × L42
COSMOS-ASO	Freie Universität Berlin, Germany	96×48 × L19	120×101 × L40
FGOALS-g2	LASG, Institute of Atmospheric Physics, Chinese Academy of Sciences and CESS, Tsinghua University, China	128×60 × L26	360×180 × L30
GISS-E2-R	NASA Goddard Institute for Space Studies, USA	144×90 × L40	288×180 × L32
IPSL-CM5A-LR	Institut Pierre-Simon Laplace, France	96×95 × L39	182×149 × L31
MIROC-ESM	Japan Agency for Marine-Earth Science and Technology, University of Tokyo and National Institute for Environmental Studies, Japan	128×64 × L80	256×192 × L44
MPI-ESM-P	Max-Planck Institut für Meterologie, Germany	196×98 × L47	256×220 × L40

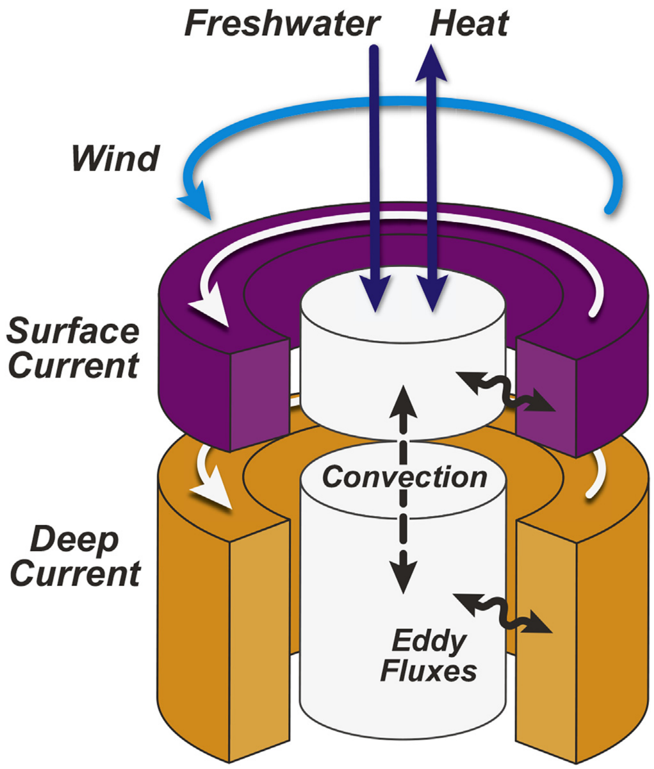

Fig. 10. Schematic of the subpolar gyre, representing the main forcings (wind, freshwater, heat), processes (convection, eddy fluxes) and components (surface current, deep current, central basin) of the box model used in this study. See section 4.3 for details.

and the ocean (e.g., Nordic Seas ice cover expands when the upper ocean is more stratified). Equally important is its ability to modulate the interaction between the atmosphere and ocean. When the subpolar region is ice-free, the ocean feels the atmospheric forcing directly — the wind-driven (barotropic) circulation is strong, as are the gyre and its associated northward heat transport. When the subpolar region is ice-covered, this direct connection is obstructed; in an extreme situation, it could be cut off entirely.

The hypothesis presented here relies on this important contrast between ice-free and ice-covered. In today's climate, relatively warm conditions and an erratic (i.e., highly variable in position, as described by Czaja (2009); see section 3.3) jet stream with southwest-northeast tilt prevent sea ice from expanding too far equatorward in the North Atlantic. As a consequence, the ocean and atmosphere are always quite tightly coupled, and the subpolar region exhibits a relatively narrow range of variability associated with hysteresis in the gyre strength (this amounts to approximately 4–5 Sv for the baseline wind forcing in our PI simulation; see Fig. 11b and c). Although the subpolar gyre hysteresis is reduced with the stronger wind forcing of the glacial climate, the variability of the subpolar region may be dramatically increased because atmosphere-ocean coupling can be turned “on” or “off” by sea ice. In the “on” state (interstadial), the maximum gyre strength is even larger than present. But as sea ice grows, the ocean is increasingly shielded from wind forcing and the gyre strength declines. If the sea ice cover becomes thick and rigid enough, it can potentially suppress most of the Sverdrup transport, resulting in a reduction on the order of tens of Sv, depending on the height of the Laurentide ice sheet (Fig. 11b and c). The concomitant decrease in poleward ocean heat transport helps sea ice spread across the region to sustain the decoupling of atmosphere and ocean.

The linchpin of the hypothesis is the ability of the subpolar region to exist in these two different states, ice-free and ice-covered. We posit that this is possible due to a combination of (a) generally colder conditions, and (b) the glacial North Atlantic jet, which is strong, steady and southward-shifted (section 4.2) compared to the jet today. The ice-free state might seem unlikely in an ice-age world, but it is realizable given the strong wind stress curl that drives a vigorous subpolar gyre and vigorous ocean heat transport towards the Nordic Seas. At the same time, the glacial system is also more likely to transition to the ice-covered state compared to today's system. The steadiness of the glacial jet reduces salt transport from the subtropics to the subpolar gyre Czaja (2009), consistent

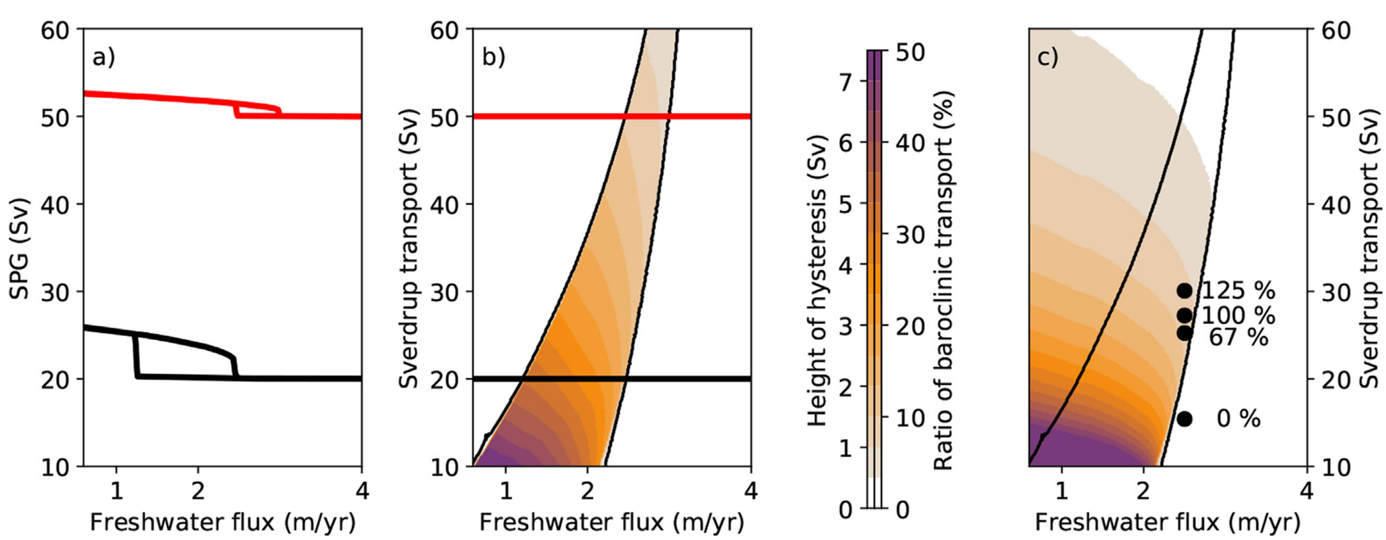

Fig. 11. Box model results showing the sensitivity of the subpolar gyre to wind and freshwater forcing. (a) Total subpolar gyre strength for two simulations with different wind forcing (i.e., Sverdrup transport). For a certain range of surface freshwater forcing, two stable solutions exist that form a hysteresis loop (black curve). Stronger wind forcing (stronger Sverdrup transport) shifts the hysteresis and reduces both its width and height (red curve). (b) Generalization of (a) for a range of wind forcing (Sverdrup transport). (c) Sensitivity of gyre dynamics to wind and freshwater forcing. The shading shows the buoyancy-driven (baroclinic) transport component expressed as a fraction of the Sverdrup transport. The edges of the hysteresis region from (b) are shown for orientation. Black dots show the Sverdrup transport averaged over the subpolar gyre region (50–63°N, 35–65°W) for the simulations in Fig. 7, showing the decreasing importance of the baroclinic transport with larger continental ice sheets.

with the idea that a fresher North Atlantic could support millennial oscillations that are stochastically excited (Timmermann et al., 2003). Additionally, the southward shift of the glacial jet cools the subpolar region (van der Schrier et al., 2007) and promotes more Greenland blocking (Madonna et al., 2017). Together, these encourage the coupled feedbacks that expand sea ice, cut off the atmosphere-ocean connection, and reduce ocean heat transport towards the Nordic Seas.

The fact that only one glacial simulation to date (CESM) exhibits spontaneous climate changes resembling Dansgaard-Oeschger events (Peltier and Vettoretti, 2014; Vettoretti and Peltier, 2015, 2016, 2018) is curious but perhaps not surprising. If the proposed mechanisms discussed in this review are indeed at play, they can only be simulated by models that have adequate spatial resolution and a good representation of the subpolar gyre, atmosphere-ice-ocean interactions, and atmospheric blocking. For example, coarser models do not resolve the comparatively small-scale details of the Labrador Sea necessary for capturing subpolar gyre dynamics (Spall, 2004; Straneo, 2006), and tend to compensate with a more reactive AMOC in response to changes that affect water mass transformations in the region. The CESM system does seem to perform better than many other models in simulating North Atlantic jet regimes and blocking (Kwon et al., 2018), and ocean stratification in the subpolar region (Sgubin et al., 2017), which suggests it is more likely to produce this type of spontaneous abrupt change. But we should also bear in mind that Dansgaard-Oeschger events occurred prior to glacial maximum, a period of milder ice age conditions than the more commonly simulated LGM time-slice. Among the important differences are smaller continental ice sheets and an open ocean pathway through the Barents Sea (Mangerud et al., 1998; Jakobsson et al., 2014; Hughes et al., 2015), both of which would alter the climate of the Northern Seas.

While it may seem worrisome that spontaneous abrupt changes are found in pre-industrial as well as glacial model runs (section 4.1), we argue that this is an extremely useful result. It suggests that the same physics can create abrupt events under a range of climate conditions, both interglacial and glacial, and that coupled climate models capture the dynamics to simulate transitions of sufficient amplitude and abruptness – but that the effect of the physics can be modified by the background climate state. These unusual pre-industrial simulations may exhibit spontaneous abrupt changes (e.g., Drijfhout et al., 2013; Kleppin et al., 2014) because, unlike more “realistic” pre-industrial simulations, their subpolar regions are somehow closer to a state allowing for both the ice-free and ice-covered states. There are a number of plausible reasons, including model biases that make the region too cold/fresh, or the jet too strong/zonal, compared to what is expected for an interglacial climate. This seems to be the case for the very cold CCSM4 pre-industrial simulation documented in Kleppin et al. (2014), and could also explain the unforced oscillations observed in the simulations of Brown and Galbraith (2016), which occur in what is essentially a pre-industrial climate with CO_2 concentrations reduced to 180 ppm. Still, the glacial abrupt events are more conspicuous, persistent and systematic (Peltier and Vettoretti, 2014), consistent with the paleoclimate record (section 2), while the pre-industrial events are typically weaker and shorter lived (Drijfhout et al., 2013; Kleppin et al., 2014).

Finally, we return to the AMOC. The idea that Dansgaard-Oeschger events are a spontaneous oscillation of the glacial Northern Seas is not inconsistent with proxy evidence linking stadial-interstadial transitions to overturning changes (Lynch-Stieglitz, 2017). The AMOC is an integrated measure of ocean circulation in the latitude-depth plane, and like the gyre, it is driven by both buoyancy and wind forcing. The processes invoked here alter Sverdrup transport, stratification, and sea ice cover in the Northern Seas; it follows that there must be an imprint on the ocean heat transport and the AMOC. This viewpoint is consistent with the possibility that, during Dansgaard-Oeschger events, the AMOC was responding to more gradual forcing (Barker et al., 2015; Barker and Knorr, 2016). If the AMOC exhibits threshold behaviour or responds with some delay to the coupled feedbacks described here, one might also expect smaller or shorter abrupt events to show little signal in the overturning, and hence, in the Southern Hemisphere. This appears to be the case for both actual Dansgaard-Oeschger events and the unexpected events in pre-industrial simulations.

6. Concluding remarks

This review has attempted to establish the idea that a mechanism for Dansgaard-Oeschger events does not require exotic physics, but rather can rely on known physics operating under altered (glacial) background climate conditions. Coupled atmosphere-ice-ocean dynamics within the Northern Seas can create Dansgaard-Oeschger-like oscillations that exhibit abrupt transitions between persistent warm interstadial and cold stadial conditions, consistent with the observational record. The background conditions dictate the character of the abrupt climate change as well as the level of climate noise, which in turn determine the likelihood of engaging critical feedbacks that alternately push the subpolar gyre towards strong (interstadial) and weak (stadial) modes. We present arguments for why glacial conditions are expected to produce larger, sustained (millennial-scale) oscillations: generally stronger winds force a vigorous subpolar gyre to allow for interstadials, while the steady, southern jet allows for periodic transitions to stadials by maintaining a cold, fresh North Atlantic and producing frequent episodes of Greenland blocking. Interactions between winds, sea ice cover, subpolar gyre strength, ocean heat transport and ocean stratification help set the millennial time scale. Similar but weaker climate changes are found in handful of pre-industrial experiments as well. In a sense, these are the ugly ducklings[1] of pre-industrial simulations: at the outset, they may be judged undesirable (i.e., unrealistic), but ultimately show value in offering important insights into the mechanisms behind abrupt climate change.

Thus, understanding Dansgaard-Oeschger events is a matter of finding the sweet spot of the Northern Seas climate system. We suggest focusing on the subpolar region, where three key ingredients meet: vigorous winds, the sea ice edge, and an oceanic gyre that connects the surface and deep ocean. In the model world, the sweet spot can be attained by various combinations of continental ice sheets, freshwater, atmospheric greenhouse gas concentrations, and other external forcings such as insolation. In reality, it seems that conditions combined optimally during Marine Isotope Stage 3, when Dansgaard-Oeschger events occurred with most regularity.

What remains to be understood is why some models have a sweet spot for producing Dangaard-Oeschger-like oscillations, while others do not. We argue that the necessary mechanisms are represented with sufficient detail in modern climate models, and that future work should concentrate on constraining these models' sensitivity to glacial boundary conditions. Which are the most important processes for hitting the sweet spot? Is it blocking, the sea ice and halocline system, wind-driven or buoyancy-driven supbolar gyre variability? Are there important processes we have neglected in our discussion, such as atmosphere-ocean interactions

[1] “The Ugly Duckling” (Danish: Den grimme ælling) is a fairy tale by poet and author Hans Christian Andersen (1805–1875).

near sharp SST gradients (Minobe et al., 2008) or those related to Arctic ice shelves (Gasson et al., 2018)? One way forward is to build intuition for how these critical processes collectively influence the stability of the climate system by better exploiting simplified models, like the box model presented in section 4.3. For example, what happens to the variability of the Northern Seas if we move the storm track and sea ice edge closer together or further apart? Are there changes in both the mean state and sensitivity of the system? A final question is whether we can understand abrupt climate change as an oscillation rooted in Northern Seas, as suggested here, or whether other components of the glacial climate system play an active role. One could consider so-called pacemaker experiments similar to those used to understand the global response to the El Niño-Southern Oscillation, where the ocean is relaxed towards some desired evolution in the pacemaker region (e.g., the subpolar North Atlantic for Dansgaard-Oeschger events), while the rest of the climate system is free to respond in a fully coupled sense.

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