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Abstract During the last glacial, major abrupt climate events known as Heinrich events left distinct fingerprints of ice rafted detritus, and are thus associated with iceberg armadas; the release of many icebergs into the North Atlantic Ocean. We simulated the impact of a large armada of icebergs on glacial climate in a coupled atmosphere–ocean model. In our model, dynamic-thermodynamic icebergs influence the climate through two direct effects. First, melting of the icebergs causes freshening of the upper ocean, and second, the latent heat used in the phase-transition of ice to water results in cooling of the iceberg surroundings. This cooling effect of icebergs is generally neglected in models. We investigated the role of the latent heat by performing a sensitivity experiment in which the cooling effect is switched off. At the peak of the simulated Heinrich event, icebergs lacking the latent heat flux are much less efficient in shutting down the meridional overturning circulation than icebergs that include both the freshening and the cooling effects. The cause of this intriguing result must be sought in the involvement of a secondary mechanism: facilitation of sea-ice formation, which can disturb deep water production at key convection sites, with consequences for the thermohaline circulation. We performed additional sensitivity experiments, designed to explore the effect of the more plausible distribution of the dynamic icebergs' melting fluxes compared to a classic hosing approach with homogeneous spreading of the melt fluxes over a section in the mid-latitude North Atlantic (NA) Ocean. The early response of the climate system is much stronger in the iceberg experiments than in the hosing experiments, which must be a distribution-effect: the dynamically distributed icebergs quickly affect western NADW formation, which synergizes with direct sea-ice facilitation, causing an earlier sea-ice expansion and climatic response. Furthermore, compared to dynamic-thermodynamic icebergs, a homogeneous hosing overestimates the fresh water flux in the Eastern Ruddiman belt, causing a fresh anomaly in the Eastern North Atlantic, leading to a delayed recovery of the circulation after the event.

1 Introduction

Paleoceanographic records provide ample evidence that "great armadas of icebergs" (Heinrich 1988) were a prominent feature of the last glacial climate (reviews: Hemming 2004; Andrews 1998). Six episodical discharges of icebergs into the North Atlantic Ocean, so-called Heinrich events (e.g. Broecker et al. 1993), have been registered as distinct layers of ice rafted detritus (IRD) in marine sediment cores from the North Atlantic. Sedimentological and mineralogical evidence (see references in Andrews 1998; Hemming 2004) shows that most IRD originated from the Laurentide Ice Sheet, although other ice sheets (i.e., Eurasian, Icelandic, Greenland) also contributed (e.g. Grousset et al. 2001). The icebergs were mainly discharged in the Labrador Sea area, after which they floated eastward and melted preferentially in the so-called Ruddiman belt (40°–55°N) (Ruddiman 1977), where the thickest IRD layers are recovered. Other paleoclimatic records show that Heinrich events are associated with major abrupt cooling events, and coincide with a global or

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at least Northern-Hemisphere-wide climatic footprint (Hemming 2004).

The strong cooling in the North Atlantic due to the icebergs is hypothesized to originate from a disturbed ocean circulation, which is weakened by the meltwater associated with the iceberg armadas (e.g. Broecker et al. 1993; McManus et al. 2004; Gherardi et al. 2005). This is corroborated by numerous coupled atmosphere–ocean model experiments, showing that the fresh melting water from these icebergs can trigger (partial) "shut down" of the meridional overturning circulation (MOC) (Kageyama et al. 2010 and references therein; e. g. Hewitt et al. 2006). In these so-called hosing studies, the freshening effect of the melting icebergs is typically simplified to homogeneous and instantaneous dumping of freshwater on a designated ocean area that corresponds to the Ruddiman belt.

This hosing-approach neglects the dynamic nature of icebergs, floating under influence of local forces such as the currents and the winds. Since the bulk of the icebergs will melt near the release site (with a logarithmic spread of the freshwater input with distance to the release location, cf. Jongma et al. (2009); Fig. 1), homogeneous freshwater hosing might be expected to be unrealistically efficient at inhibiting convection, when compared to a more plausible iceberg-melt distribution (Jongma et al. 2009). Indeed, a Heinrich event has been simulated recently with dynamic icebergs (Levine and Bigg 2008), suggesting that with a more realistic, localized freshwater flux of icebergs, the "MOC-shutdown may be harder to induce than previously suggested". However, neither the classical hosing approach nor the above iceberg study take into account the significant amount of latent heat that is needed to melt the icebergs. The heat involved in the phase-transition from ice to water is around 80 times higher than the heat involved in lowering water-temperature by 1 °C.

Jongma et al. (2009) separated the cooling and freshening aspects of icebergs in a series of Southern Ocean (pre-industrial) sensitivity experiments. Both the cooling and freshening effect of melting icebergs can directly affect the depth and steepness of the pycnocline, with consequences for the entrainment of saltier and warmer waters. Traditionally, cooling of the surface waters is regarded as a process that enhances deep water formation, as it results in principle in a higher density of the upper water mass. However, the surface cooling associated with melting icebergs, in conjunction with the freshening due to the melting ice, can facilitate the formation of sea-ice (Jongma et al. 2009; Wiersma and Jongma 2010). This can have
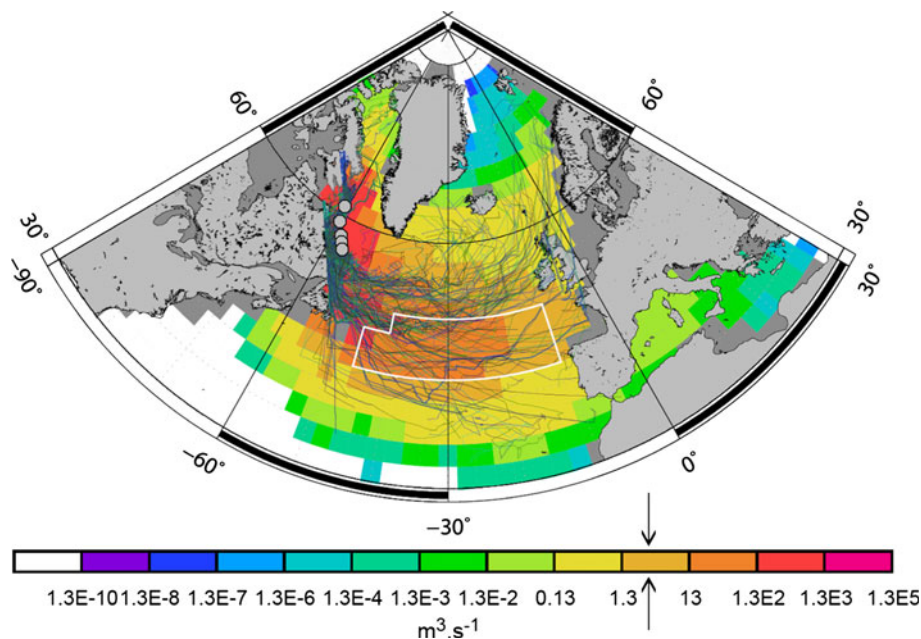

Fig. 1 Spread of the icebergs melting fluxes for the CF-icebergs (color filling, note the logarithmic scale) averaged over simulation years 151–300. Most of the melt takes place west of 30°W. Superimposed are some typical iceberg trajectories (any apparent land crossings are due to the interpolation method used to display the data from the original rotated grid). The five iceberg release sites are marked as *grey circles*. The *solid white line* marks the edges of the area of homogeneous Ruddiman belt hosing. The *solid arrows* given on the *color bar* indicates the amount used for the homogeneous hosing approach: hence any location on the map within the *solid white line* boundary having a color "colder" than the *arrows* receives less freshwater flux in the iceberg experiments during years 151–300 than the homogeneous freshwater hosing. The reverse is true for the locations with "warmer" color code. It should be noted that this comparison is indicative as the melt flux from the iceberg experiments is time transgressive by nature: the moving icebergs are releasing freshwater on their path. Thus, the same figure for an average of 50 years within years 151–300 might look different, as would as well a figure based on the early response when the thermohaline circulation is not perturbed

strong climatic implications, since sea-ice provides a couple of strong positive feedbacks to the cooling. Firstly the relatively high sea-ice albedo will increase the amount of heat reflected back into space. Secondly, sea-ice reduces the heat exchange between the ocean and the atmosphere, which can hamper deep water formation, thus reducing the advection of heat (e.g. Jongma et al. 2009; Kaspi et al. 2004). Additionally, freshwater buffered in (extra) sea-ice and intercepted precipitation can lead to a net freshening of the surface ocean (e.g. Jongma et al. 2009). Consequently, through sea-ice facilitation, the cooling aspect of melting icebergs can effectively reduce deep water formation in the Southern Ocean.

For the (glacial) North Atlantic the situation is geographically more complex. North Atlantic convection can be characterized as a three-dimensional interplay between individual convection sites and basins, influenced by the local history (e.g. Jongma et al. 2007; Schulz et al. 2007; Prange et al. 2010; Roche et al. 2010). From this perspective, the system could be severely affected by icebergs that reach local convection sites, which would not be captured in a typical hosing study. In other words the spatial distribution of the melting fluxes by dynamic icebergs might affect the character of the system's response.

To study the climatic significance of both the icebergs' freshwater and melting heat fluxes during an Heinrich event, we have coupled a dynamic thermodynamic iceberg module to a coupled climate model in LGM state. We simulate a Heinrich event under LGM climatic conditions with a large armada of dynamic-thermodynamic icebergs to mimic the conditions of Heinrich event 1. The choice of this particular Heinrich event is based on the availability of a well-described LGM background state (Roche et al. 2007) as well as the inclusion of this particular event in the third phase of the Paleoclimate Modelling Intercomparison Project (PMIP-3) that will allow further comparison with more complex models. PMIP-3 models being run with freshwater fluxes, we propose to explore the role of the spatial distribution of the iceberg melt and compare our results with the former typical homogeneous hosing approach. In a series of sensitivity experiments we unravel the cooling and the freshwater effects, paying special attention to the role of (facilitated) sea-ice and the consequences for the North Atlantic deep water (NADW) formation.

2 Model description

We use version 1.0 of the global three-dimensional coupled climate model LOVECLIM (Goosse et al. 2010), with only the oceanic (CLIO) and atmospheric (EcBilt) components activated. Despite its reasonably elaborate ocean (20-layer,

3° × 3° resolution, primitive equation free-surface ocean general circulation model with thermodynamic–dynamic sea-ice), it enables multi-millennial simulations thanks to its intermediate complexity atmosphere (T21L3, quasi-geostrophic). For a more extensive description we refer the reader to Goosse et al. 2010.

The dynamic-thermodynamic iceberg module used to simulate the Heinrich event is based on the iceberg trajectory model introduced by Mountain (1980), and further developed by Bigg et al. (1996, 1997) and Jongma et al. (2009). Summarizing [see (Jongma et al. 2009)], the iceberg model predicts the general trajectories of icebergs subject to Coriolis force, air-, water- and sea-ice drag, horizontal pressure gradient of the surrounding ocean and wave radiation. The icebergs are subject to bottom melt, lateral melt and wave erosion. The fresh water flux and latent (melting) heat flux associated with the volume-loss of the icebergs is piped to the appropriate ocean layer of the local grid cell.

We have adopted all parameter choices of Bigg et al. (Smith 1993; Bigg et al. 1996, 1997; Gladstone et al. 2001). Icebergs are assumed to remain tabular, and icebergs do not collide with each other. Increased drag of thick sea-ice is not accounted for. Direct thermodynamical interaction between the icebergs and the atmosphere is marginal (Loset 1993) and considered to be negligible, as are the water- and wind- stresses acting on the bottom respectively top surface (G.R. Bigg, personal communication). Keel shape or other turbulence related effects are not accounted for, nor is added mass due to entrained meltwater. This model has given good descriptions of the general behaviour of icebergs but cannot be expected to work well for individual icebergs.

The Jongma et al. (2009) iceberg module was refined, similarly to (Wiersma and Jongma 2010), by piping the freshwater flux and the latent heat flux associated with the icebergs' basal and lateral melt to the local ocean layer as opposed to the surface-layer. More precisely, for a given iceberg of height H, we calculate a freshwater flux for the melting that is given to all model layers between the surface and H, weighted by each model layer thickness. Adding freshwater to a series of vertical layers in the ocean tends to dampen the impact of freshwater on ocean feedbacks (as described in details below). This improvement was added by Wiersma and Jongma (2010) since the effect of freshwater induced halocline on the ocean was found to be too extreme and thus unrealistic. Furthermore, as in (Wiersma and Jongma 2010), there is a linear dependence of the wave erosion on the sea-ice fraction to mimic a first-order dampening effect of sea-ice on the (wind-dependent) waveheight In addition, icebergs that are about to get grounded are weakly repulsed (orthogonal repulsion of 0.003 m/s) instead of fixated. This almost immobilizes the grounding

icebergs, while allowing changing winds and currents the opportunity to move the icebergs to deeper waters.

3 Experimental design

All experiments in this study are simulated with last glacial maximum (LGM) boundary conditions. The LGM boundary conditions are imposed as in the PMIP-2 protocol, as described in detail in Roche et al. (2007). Summarising: Orbital parameters are taken from Berger and Loutre (1992) and set to 21 ka B.P. Greenhouse gas concentrations are set in accordance with reconstructed levels from ice-core data (Fluckiger et al. 1999; Dallenbach et al. 2000; Monnin et al. 2001), to 185 p.p.m.v., 350 and 200 p.p.b.v. for CO_2, CH_4 and N_2O respectively; The ice-sheet topography and ice-mask are based on Peltier's (2004) reconstruction, interpolated at T21 resolution. Accordingly, the sea-level drop of 120 m is taken into account for the ocean (Lambeck and Chappell 2001), with a coherent modification of the land-sea mask. We account for a change in the river routing due to these topographic changes. From the resulting quasi-equilibrium state we simulated a 1,000 year run that will be used as a control condition.

Heinrich event I is characterized by massive ice release from the Laurentide Ice Sheet. In separate experiments, the associated freshwater -and latent heat- fluxes were distributed in four different ways (Table 1). In the (cooling and freshening) CF-icebergs experiment, the ice released from the Laurentide Ice Sheet was distributed by icebergs with both freshwater and latent heat melting fluxes. For comparison, a classic hosing experiment (F-hosing) was performed where the freshwater flux was distributed homogeneously over a designated area in the Ruddiman belt (see Fig. 1), defined as in Roche et al. (2010). The differences between these two experiments can be attributed to a cooling effect (the presence of a latent heat flux) and a distribution effect (dynamic icebergs vs. homogeneous hosing). To separate the distribution effect from the cooling effect, two additional experiments were performed: an F-icebergs experiment, consisting of icebergs with the latent heat flux switched off; and a CF-hosing experiment were the freshwater hosing was accompanied by a latent heat flux to account for the phase-transition of ice to water.

Estimating the duration and equivalent freshwater flux release for the Heinrich event 1 is a complex issue. The duration of the forcing itself has to be determined by the duration of the IRD input in the North Atlantic, as this is the only definition of a Heinrich event. Different authors arrive at a different result from separate methodologies. One Heinrich event could be different from another one as well (cf. Hemming 2004 for a review). Duration evaluations range from one to several millennia (Bond and Lotti 1995; Dowdeswell et al. 1995; Elliot et al. 1998; Vidal et al. 1997; Grousset et al. 2001), between 500 and 1,000 years (François and Bacon 1994; Thomson et al. 1995; Dowdeswell et al. 1995) or less than 500 years (Grousset Pers. Comm.; Roche et al. 2004). The amount of freshwater released is even more uncertain, but is constrained by the fact that Heinrich event 1 is not well marked in the sea-level deglacial record, implying a freshwater flux of at most a few meters (e.g. Stanford et al. 2011a). The fact that a Heinrich event such as Heinrich event 1 is probably multiphased with European precursor events (Grousset et al. 2000, 2001; Stanford et al. 2011b) adds to the complexity of the period. In the following, we consider a simple experiment of a single phased Laurentide Heinrich event under LGM conditions, with characteristics as discussed in Roche et al. 2004. This decision is somewhat arbitrary but has the advantage of not being too computationally costly. Indeed, the total freshwater volume released from the Laurentide ice sheet during the event was estimated at 2.2×10^{15} m^3 (Roche et al. 2004), which was released with a constant rate over a period of 300 years, amounting to a flux of 0.235 Sv.

In the hosing experiments, this flux was added directly to the surface ocean layer in the Ruddiman belt (Fig. 1). In the iceberg experiments, the flux was converted to an equivalent iceberg volume. Heinrich event I was then simulated by releasing icebergs of 10 different size classes

Table 1 Description of experiments: We use a sensitivity approach where we compare Cool & Fresh (CF-) icebergs with three different experimental set-ups. F-icebergs are experiments with the cooling aspect switched off; CF- and F- hosing are experiments where meltwater is distributed directly in the Ruddiman Belt. Control is without additional forcing

Exp.	Icebergs	Hosing
Fresh-(freshwater flux only)	F-icebergs	F-hosing
	▬▬▬ (blue)	▬▬▬ (red)
Cool & Fresh-(freshwater and latent heat flux)	CF-icebergs	CF-hosing
	▬▬▬ (light blue)	▬▬▬ (orange)
Control-experiment	EQ (equilibrium run with LGM boundary conditions)	

"Hosing" stands for homogeneous distribution of the melting fluxes over an area corresponding to the Ruddiman belt (see Fig. 1); this is the classic way to simulate Heinrich events. The colored lines used in the boxes are those used in Fig. 3

Table 2 Size distribution of the daily iceberg release

Fraction	Length	Width	Draught	Freeboard
0.15	100.5	67	67	10.0
0.15	199.5	133	133	19.9
0.2	300	200	200	29.9
0.15	400.5	267	267	39.9
0.08	499.5	333	300	44.8
0.07	600	400	300	44.8
0.05	750	500	300	44.8
0.05	900	600	300	44.8
0.05	1,200	800	300	44.8
0.05	1,500	1,000	300	44.8

Left The fraction of the total ice volume budget that is assigned to each size class. This roughly follows a log-normal distribution. *Column 2–5* Length, width, depth and height above water. Icebergs are assumed to remain tabular with length–width ratio 1.5

(Table 2) on a daily basis from 5 different release sites near the South West of the Labrador Sea, just above New-foundland (see markers in Fig. 1).

Since reliable estimates of iceberg sizes during LGM are not available, the iceberg discharge consists of 10 classes of icebergs identical to Bigg et al. (1997), which was based on present day observations of icebergs in the Arctic (Dowdeswell et al. 1992), and roughly follows a log-normal distribution (Table 2). The total ice volume budget is divided according to this size-distribution over 300 years times 360 days and over the five release sites (Fig. 1), to determine the "amount" of daily released icebergs of a certain size class. Each iceberg trajectory thus represents a group of icebergs, proportional to this "amount", with a common source and size-class. It should be noted that using such iceberg classes to represent the distribution of icebergs possibly originating from a breakout of a large ice-shelf is a strong assumption. Indeed, it is likely that the log-normal distribution would be biased towards larger iceberg sizes, though retaining some smaller ones. Given the lack of data available to estimate such an "abnormal" iceberg distribution, we preferred to keep the present-day one. Evaluating the effect of differently shaped iceberg distributions is beyond the scope of our present study. It is likely that giant bergs would be even more efficient at affecting the AMOC, since their longer possible travel time may allow them to reach the Nordic Seas convection shortly after the beginning of the event.

4 Results and discussion

The base LGM state is characterized by a slightly stronger MOC and deeper NADW formation when compared to the pre-industrial state. Notably, there are two zones of deep water formation in the north Atlantic: one main region south of Iceland and a secondary one (producing the deepest waters) in the Nordic Seas. This aspect has been shown (Roche et al. 2007) to be consistent with proxy data (Labeyrie et al. 1992; Oppo and Lehman 1993; Dokken and Jansen 1999; Meland et al. 2008). In terms of sea-ice, also of importance here, the Nordic Seas are only covered in winter.

## 4.1 Distribution of iceberg melt

Consistent with observations, most of the dynamic icebergs melt (Fig. 1) takes place in the Ruddiman belt (Ruddiman 1977; Bond et al. 1992, 1993; Bond and Lotti 1995), while some icebergs reach the coast of Portugal [e.g. (Abrantes et al. 1998)] and others reach the northern GIN seas (consistent with the "loop" scenario, Fig. 5.24 and 5.25 in Bischof (2000)]. We note that for technical reasons the Straight of Gibraltar is relatively large in our model so the simulated (weak) iceberg melt in the Mediterranean Sea should not be taken at face value. Please note the logarithmic scale of the iceberg melt-fluxes. The melt-fluxes from the dynamic icebergs exhibit a strong West-East gradient contrary to the hosing experiments with homogeneous hosing in the Ruddiman belt area. We note that icebergs need a few months to a year to cross the North Atlantic, depending on their paths. The total lifetime of specific icebergs may extend to a few years when trapped in a cold area like the Arctic Ocean.

4.2 Climatic response to the cool and fresh icebergs

After a few hundred years, the CF-icebergs melt-fluxes have caused a drop in the sea surface salinity (SSS) of at least 1 psu between 40° and 60°N, and fanning out between 30° and 70°N in the Eastern North Atlantic (Fig. 2: SSS). In the western area near the coast and the Ruddiman belt, the salinity drops 2–3 psu. This is in agreement with reconstructions, based on the oxygen isotope composition of fossil planktonic foraminifera, by Cortijo et al. (1997, 2005). They report a 1.5–3.5 p.p.t. salinity drop between 30°N and 50°N during Heinrich event 4. Although they did not find significant salinity variations North of 50°N for this older Heinrich event 4, for Heinrich event 1 they report a more northern hydrographic pattern with a West-East decrease ("larger isotopic amplitude near the melting source") (Cortijo et al. 2005). Where they reported a cooling of about 2 °C in that area, the sea surface cooling by CF-icebergs lies in the range of 1–3.5 °C (Fig. 2: SST).

In order to unravel the impact of the freshening, the cooling and the distribution aspect of the dynamic CF-icebergs, we use a sensitivity approach where we compare CF-icebergs with three different experimental set-ups with

Fig. 2 Peak responses (years 151–300) North Atlantic: Anomalous sea surface salinity (SSS *left column*), Convection layer depth (CLD *middle column*) and Sea-ice fraction (*right column*) for: fresh-water-hosing (Fh *2nd row*); cooling and freshening -hosing (CFh *3rd row*); freshening-icebergs (Fi *4th row*); cooling and freshening -icebergs (CFi *5th row*). These bottom 4 rows display the differences ("anomalies") e.g. CFi–ctl between the mentioned event experiments and the control run (ctl *top-row*), both averaged over years 150–300. February values for SSS (psu) and sea-ice fraction. Please note CLD (m) shows the average of yearly maximum depths of the convection layer in each gridcell, implying that actual depths reached are greater in individual years
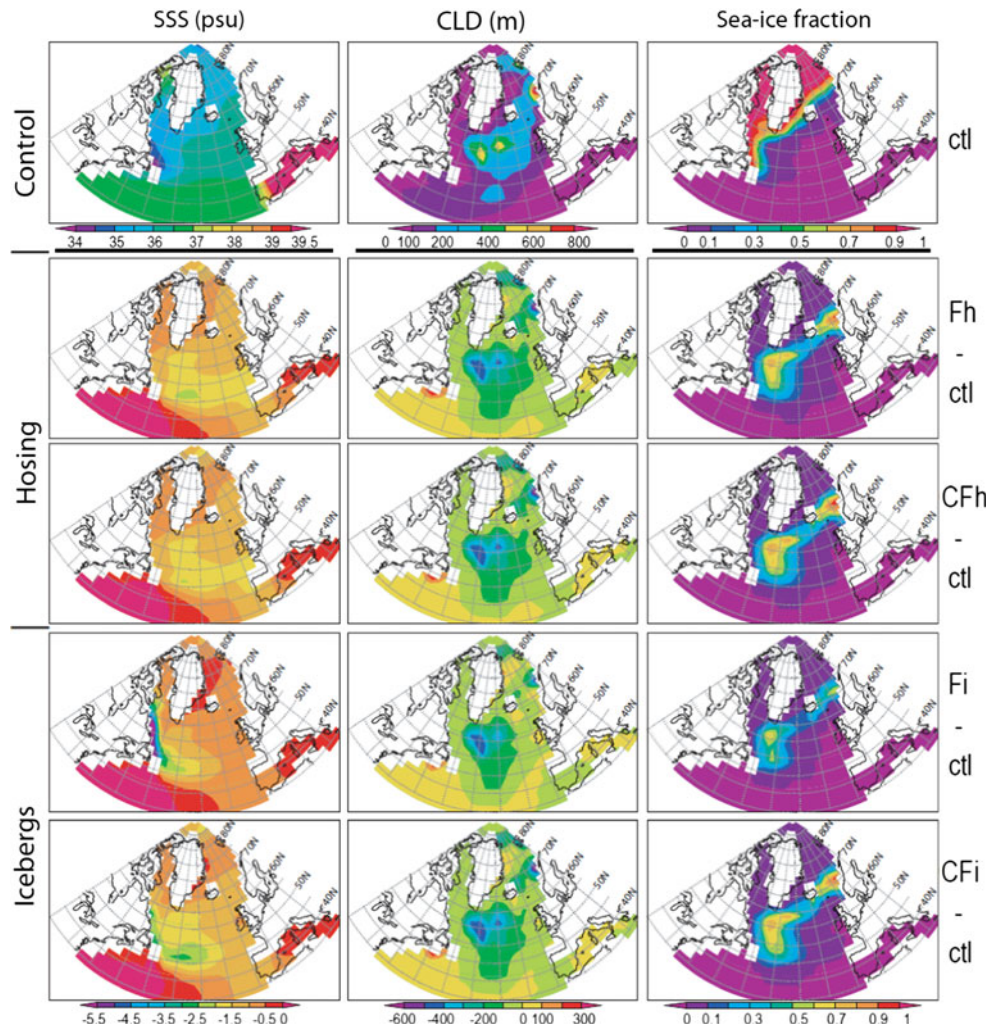
some of these aspects switched off (Table 1). We will make a distinction between the early response (first 50 years after the start of the perturbation) and the peak response (years 150–300, which is the end of the perturbation). As we are considering the response to a single peaked, single-phased event, our experiment should be compared to the main phase of the Heinrich event 1 as seen in proxy data and not to the early stages involving European ice-sheet instabilities only. In the perspective of Stanford et al. (2011b), this would be phase 2 ("main HE-ss1 phase") and phase 3 ("H1 cleanup and AMOC resumption"). The peak-response considered here would be the maximum of the phase 2 of Stanford et al. (2011b).

### 4.3 Peak-response: F-Hosing more efficient at shutting down MOC than F-icebergs

In the four experiments, the maximum Atlantic meridional overturning circulation (AMOC), which is a measure of deep water production, is affected differently by the cooling and freshening effects (Fig. 3a). At the end of the

discharge (year 300) classic freshwater hosing (F-hosing or Fh, see Table 1) has reduced the MOC from 34 Sv to about 8 Sv. This amounts to a much greater suppression of NADW formation than the F-icebergs, which still allow for about 18 Sv of circulation at that time. Specifically, the F-icebergs are not nearly as efficient at shutting down convection in the Greenland-Iceland-Norwegian (GIN) Seas (Figs. 2, 3b; CLD anomalies). We attribute this to the relatively large amount of freshwater that reaches—and accumulates in—the Eastern North Atlantic in the hosing experiments, leading to reduced sea surface salinities (SSS) (Fig. 2; SSS, Fig. 5; SSS icebergs vs. classic hosing).

4.4 Peak response: CF-icebergs are surprisingly efficient at shutting down MOC

Remarkably, the CF-icebergs are much more efficient at shutting down convection than the F-icebergs (Fig. 3a). In fact, reducing the MOC to about 10 Sv, they are nearly as efficient as the hosing-approach, even though the CF-icebergs do not manage to freshen the Eastern

Fig. 3 30-year running mean time-series of: **a** The yearly maximum meridional overturning circulation (in Sverdrup) in the North Atlantic, a measure of *NADW production*; **b** similarly defined NADW production in the Greenland-Iceland-Norwegian Seas; **c** *Sea-ice volume* in the Northern Hemisphere; **d** *Sea-ice cover* in the Northern Hemisphere. *Light-blue* CF-icebergs ("Cooling and Freshening"); *Dark-blue* F-icebergs; *Orange* CF-hosing; *Red* F-hosing; *Black* Equilibrium ("control")
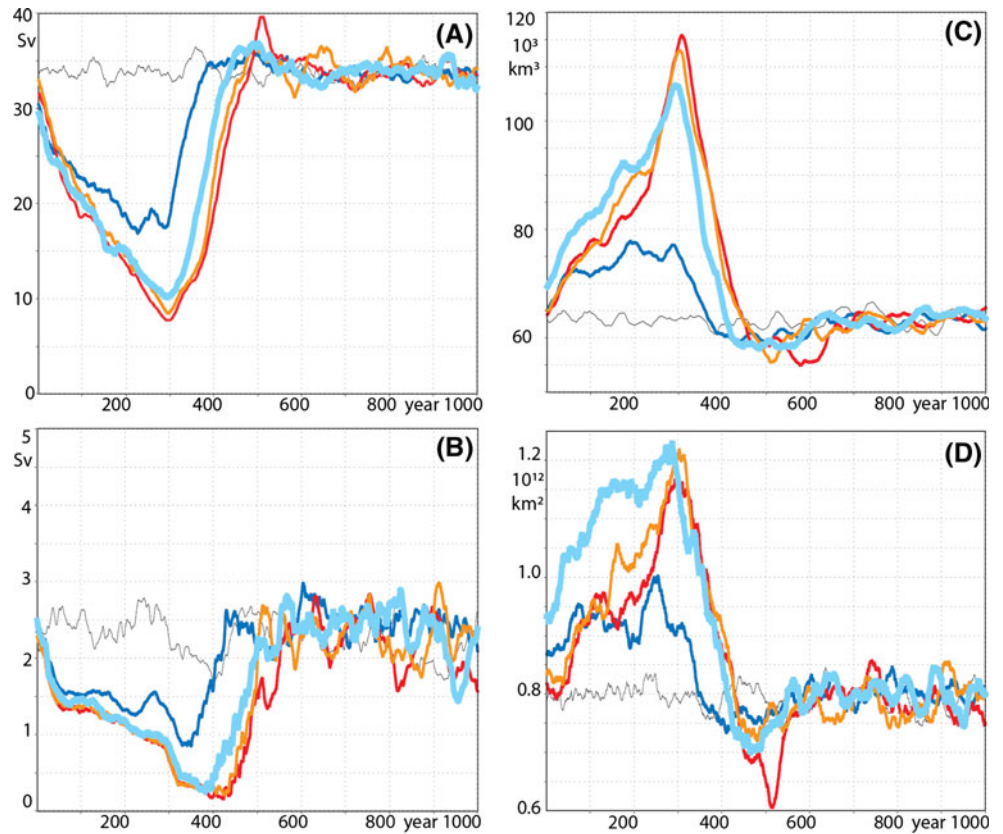
North-Atlantic and GIN-seas as much as the hosing approach does (Figs. 2, 5 SSS). Apparently, the cooling aspect of the CF-icebergs has a strong inhibitive effect on the MOC. In contrast, in the CF-hosing experiment, the cooling latent heat flux has a (marginally) positive effect on the MOC, compared to F-hosing. Likewise, "cool"- icebergs (LHF only) lead to stronger convection (not shown). Thus this somewhat counterintuitive result of CF-icebergs being more efficient at blocking the MOC than F-icebergs is clearly an indirect, non-linear, mechanism that involves the spatial distribution of the CF-icebergs. We suggest that sea-ice facilitation (Jongma et al. 2009) is the key to this mechanism (note the sea-ice differences between CF-icebergs and F-icebergs in Figs. 3c, d and in Fig. 2).

4.5 Peak-response: Delayed MOC recovery

The hosing experiments exhibit a delayed MOC recovery compared to the iceberg experiments (Fig. 3a). After the iceberg/freshwater release ends in year 300, MOC recovery shows a delay of up to 5 decades when comparing hosing with the CF-icebergs. This is quite consistent with the idea (Jongma et al. 2009; Levine and Bigg 2008) that hosing is too efficient at shutting down convection. The less convection remains active, the larger the build up of fresh surface waters in the Eastern North Atlantic Ocean

(Figs. 2, 5; SSS anomalies), which will act as a buffer against MOC recovery. This mechanism is illustrated by the fact that deep water formation in the GIN-seas re-starts 40 years later in the hosing experiments (at year 440) than in the CF iceberg experiments (at year 400, Fig. 3b).

4.6 Peak-response: The role of sea-ice

The weakening of deep convection can be studied spatially by examining CLD; the (average) yearly maximum depth of the convection layer (Fig. 2: CLD). Obviously, the anomalies in convection depth are largest at key convection sites (Fig. 2: CLD; *top row*). The CLD-anomaly patterns also relate to the SSS anomalies (Fig. 2: SSS), which directly affect the stratification. Furthermore, the spatial pattern of the negative anomalies in convection layer depth (Fig. 2: CLD) resembles the anomalous positive sea-ice pattern (Fig. 2: Sea-ice fraction), illustrating the intricate relationship between sea-ice and NADW formation.

Sea-ice can limit deep water formation by insulating the surface ocean from the atmosphere; less interaction with the atmosphere means less cooling, resulting in less deep water formation and weakening of the thermohaline circulation. In turn, a weaker THC means less heat is advected northward, which makes it easier for the sea-ice to expand.

The time series of the NADW-formation (Fig. 3a) and the Northern Hemisphere sea-ice volume (Fig. 3c) highlight the relationship between sea-ice and AMOC. At year 300, the sea-ice expansion correlates qualitatively with the pattern of reduced NADW-formation: expansion is largest in the F-hosing experiment (which exhibited the greatest weakening of deep convection); followed closely by the CF-hosing; and the CF-icebergs; while the F-icebergs stay far behind (Fig. 3c, d).

We note that sea-ice expansion (Fig. 3c, d) is about twice as strong in the CF-icebergs experiment as in the F-icebergs experiment. During the first 200 years of the event, the CF-icebergs experiment also exhibits more sea-ice than the hosing experiments, with CF-hosing starting to catch up some 100 years earlier than F-hosing. We attribute this earlier sea-ice expansion to the cooling effect of the latent heat flux, which can facilitate the formation of sea-ice in a direct manner. We make a distinction between such direct sea-ice facilitation and the indirect sea-ice expansion that can be associated with a weakened NADW-production. To separate the direct sea-ice-facilitation by the cool and fresh fluxes from such indirect sea-ice facilitation, we will now limit our analysis to the first 50 years of the event. The latter are interesting since it enables to see the direct effect of sea ice as opposed to freshwater before drastic differences in global ocean circulation, as measured by the overturning strength in the north Atlantic take place (cf. Fig 3a).

4.7 Early response: The first 50 years of Heinrich event I

Both the icebergs' fresh meltwater and the latent heat needed to melt the ice can facilitate the formation of sea-ice (Jongma et al. 2009). Compared to homogeneous hosing, dynamic icebergs have the potential to influence the North Atlantic system more directly by melting and facilitating sea-ice formation near key deep-convection sites. We analyse in detail the early response in the first 50 years. This will allow us to focus on the effects of direct sea-ice facilitation by the melting fluxes and to investigate the NADW-suppressing capability of (Cool & Fresh) CF-icebergs. Furthermore, the dynamically distributed iceberg-fluxes can be expected to lead to a more plausible simulation of the start of Heinrich events.

The feedbacks in the coupled ocean—atmosphere—sea-ice—icebergs system make it difficult to delineate all responses. In our following discussion we will follow the logical order of the basic aspects of the dynamic-thermodynamic icebergs: the freshwater flux lowers the SSS and the latent heat flux cools the upper ocean, leading to sea-ice facilitation in the local grid-cell, which in turn affects the convection, which is also directly affected by the SSS and

SST. Changes in convection then affect SSS, SST and sea-ice in a positive feedback loop.

4.8 Early response: SSS

As expected, the North Atlantic becomes fresher in all four experiments, especially in the Ruddiman belt. In this early stage, the hosing approach clearly leads to a greater freshening of the Eastern surface ocean from the Portuguese coast up to Great Britain (Fig. 4: Early response; SSS). The F-icebergs, on the other hand, cause a distinct freshening of the Western Labrador Sea, where a lot of icebergs tend to gather (not shown). In contrast, the CF-icebergs exhibit a more Eastward penetration of the freshening between 40 to 60°N, and a much weaker fresh anomaly at the Western Labrador Sea coast. This could be partly due to the fact that the iceberg erosion ("calving") is inversely proportional to the sea-ice surface fraction, but below we provide an additional explanation in terms of fresh water being buffered in sea-ice and escorted Eastward by icebergs.

4.9 Early response: SST

The CF-hosing causes a cooling of up to 2 °C in the Ruddiman belt (Fig. 4; SST) while, lacking the latent heat flux, the F-hosing only leads to slight cooling in the Western part. The CF-icebergs cause a much stronger surface cooling of 1–3.5 °C in the Ruddiman belt (Fig. 4; SST), and along the sea-ice edge (Fig. 4; HIC), corresponding to a 2–5 °C lowering of the yearly averaged atmospheric surface temperature in that area (not shown). In the Western North Atlantic one might attribute this additional cooling to the greater iceberg-melt-fluxes (Fig. 1). We note nonetheless that in the Eastern North Atlantic and Southern GIN-seas, the CF-icebergs trigger a more wide-spread cooling than the CF-hosing (Fig. 4). Furthermore, there is a distribution effect regardless of the cooling latent heat flux: the fresh (non-cooling) icebergs also cause stronger cooling in the Eastern North Atlantic than the F-hosing (Fig. 4). We attribute this to direct sea-ice facilitation by freshening the surface ocean in a susceptible region.

4.10 Early response: Sea-ice facilitation

Both the freshening and the cooling fluxes facilitate the formation of sea-ice. Compared to the control experiment, there is sea-ice facilitation in all four experiments, both in thickness (Fig. 4: HIC) and in total volume (Fig. 3c). In contrast to the peak response (Fig. 2: Sea-ice fraction), at this early stage, the F-icebergs cause a stronger sea-ice facilitation than the homogeneous hosing (Fig. 3d), indicating a more direct impact on the sea-ice (Fig. 4: note sea
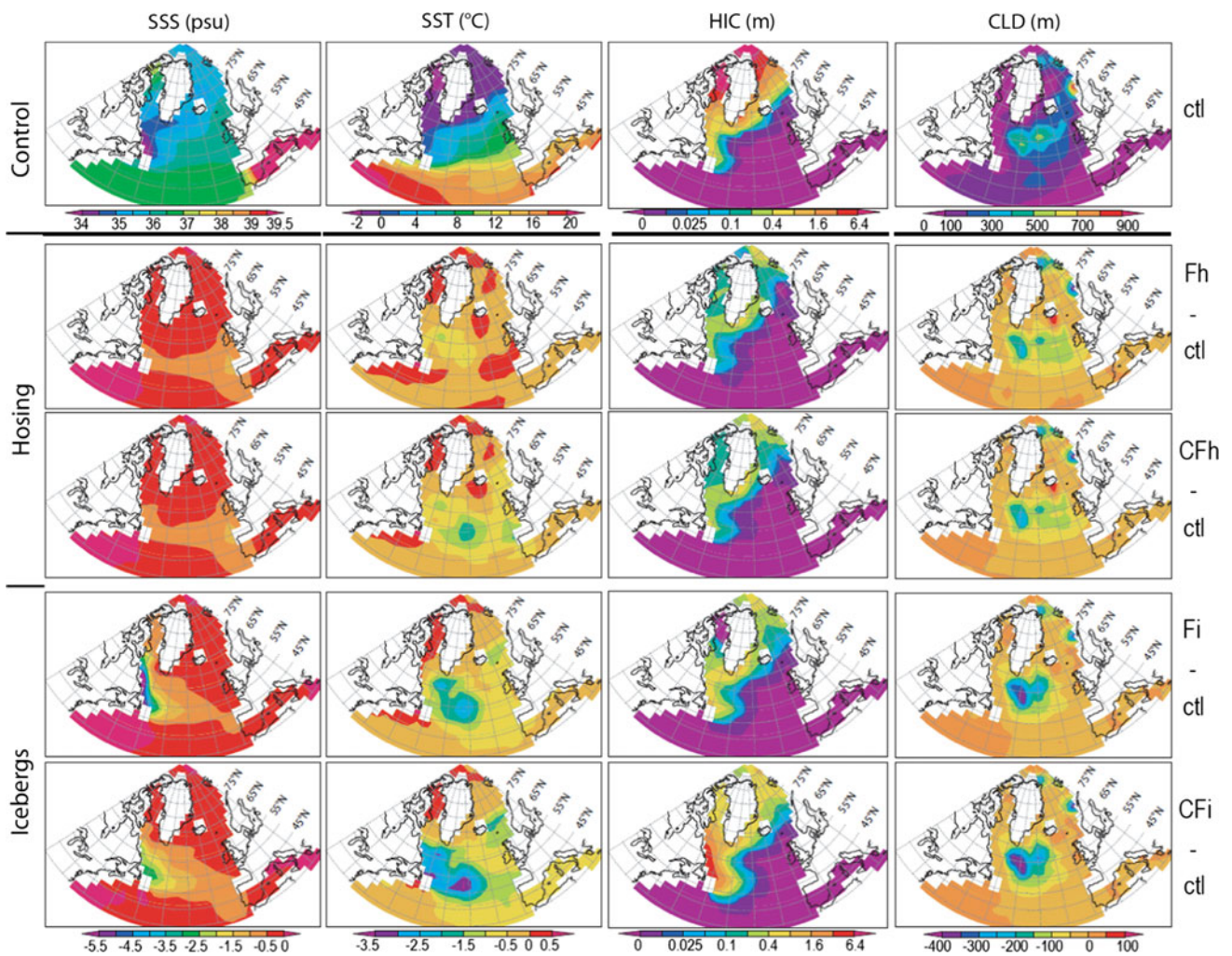
Fig. 4 Early responses (years 1–50) North Atlantic: Anomalous sea surface salinity (SSS *left column*); sea surface temperature (SST *2nd column*); Sea-ice height or thickness (HIC *3rd column*) and Convection layer depth (CLD *right column*) for: fresh-water-hosing (Fh *2nd row*); cooling and freshening hosing (CFh *3rd row*); freshening icebergs (Fi *4th row*); cooling and freshening icebergs (CFi *5th row*). These are differences between experiments and control run (ctl *top-row*), both averaged over years 1–50. Please note CLD (*right column*) shows the average of yearly maximum depths of the convection layer in each gridcell, implying that actual depths reached can be greater in individual years. Also note the logarithmic scale for sea-ice thickness HIC

ice cover below Iceland and West of Norway). The homogeneous hosing is characterized by comparatively large fluxes in the Eastern Ruddiman belt, where there is no sea-ice at this early stage. In contrast, the more plausible spatial distribution of the dynamic icebergs allows for greater melting fluxes closer to the original sea-ice edge, where the direct sea-ice facilitation can synergize with the positive feedback of sea-ice on itself (through increased albedo and isolation of the ocean from the atmosphere). This also explains the counterintuitive greater SST cooling in the Eastern North Atlantic mentioned in the previous paragraph.

Along the Western Labrador Sea coast, Near the icebergs release sites, the CF-icebergs cause a strong thickening of the sea-ice (Fig. 4: HIC), eventually leading to a thick anomalous sea-ice pack that lasts year-round (not shown). This location is characterized by strong freshening in the F-icebergs experiment and weaker freshening by CF-icebergs (Fig. 4: SSS). Here the CF-icebergs cooling-effect leads to easy freezing of the freshened upper ocean. We suggest that this extra sea-ice acts as a mobile freshwater buffer that can be transported further East than the Western SSS-anomaly in the F-icebergs experiment. Furthermore, the CF-icebergs that escort (float near) this sea-ice can be expected to provide a cool and fresh "envelope" for the sea-ice, which indeed reaches some 20° further East. This would explain the greater Eastward penetration of the Ruddiman-belt freshening in the CF-icebergs experiment, compared to the F-icebergs (Fig. 4: SSS). However, please bear in mind that even at this early stage the direct sea-ice facilitation by the iceberg-fluxes cannot be separated

completely from the secondary sea-ice facilitation by a reduced NADW formation.

4.11 Early response: Convection layer depth

At this early stage, both the fresh and CF-icebergs cause a stronger suppression of convection than the homogeneous hosing, especially at the Western-most convection site (Fig. 4: CLD; control). This is quite consistent with the fact that the dynamic icebergs lead to a more Westward release of the bulk of freshwater fluxes, so that the released fresh anomaly can affect these convection sites more directly. When comparing the CLD-anomaly pattern in the Western North Atlantic with the patterns of sea-ice and SST anomalies (Fig. 4), we note a spatial correlation between the three.

The SST anomalies between Iceland and Norway implicate sea-ice (HIC) differences as an explanation for the greater cooling by F-icebergs (which can easily reach the sea-ice edge in the GIN seas within a couple of years) than by hosing (the previously mentioned distribution effect). Indeed, there is no significant CLD or SSS changes in that area for the early response, but significant sea-ice thickening (HIC).

Given the relatively large differences in CLD between F-hosing and F-icebergs south of Iceland we must consider the possibility that at least part of the eastern cooling by the icebergs is due to secondary effects following convection anomalies. However, sea-ice facilitation must also play a key role here, since the CF-icebergs lead to slightly stronger suppression of NADW-formation than F-icebergs, while the cooling by itself would (traditionally) be expected to enhance convection.

We conclude that sea-ice facilitation and suppression of deep convection is less direct in the hosing experiments. Initially, a large part of the hosing-area in these experiments does not involve any sea-ice. It takes around 50 (Fig. 3d) to 250 (Fig. 3c) years before the fresh anomaly has spread enough to the North and the MOC has been weakened sufficiently for the sea-ice expansion to pick up in the hosing experiments. Together with the observed delayed recovery in hosing experiments, this has significant implications for understanding, interpreting and/or simulating Heinrich events.

5 Synthesis and concluding remarks

Dynamic thermodynamic icebergs can be expected to lead to a more plausible distribution of melting-fluxes than homogeneous hosing. Fluxes from icebergs increase roughly exponentially towards the release site (Fig. 1). The resulting salinity anomalies are in agreement with hydrographic reconstructions (Cortijo et al. 1997, 2005). Using a sensitivity approach we address the question whether this more realistic distribution leads to significant differences in the response of the North Atlantic Ocean and climate system. Furthermore, this is the first study of a Heinrich event that takes into account the latent heat associated with the phase-transition between ice and water. Up to now this cooling effect was considered negligible or of an opposite sign as the fresh water effect (which was confirmed by our hosing experiments). Both the cooling and the freshening effect can facilitate the formation of sea-ice. To separate direct sea-ice facilitation by the iceberg's melting fluxes from secondary sea-ice expansion due to inhibited deep water formation, we have made a distinction between the early response and the peak response of the overturning circulation.

The early response shows that, compared to hosing, icebergs affect the sea-ice and consequently temperature and convection more directly in the western as well as the eastern Atlantic, although melt fluxes from icebergs are relatively low in the East. Since this is even the case for fresh (non-cooling) icebergs, we attribute this distribution effect (at least partly) to direct sea-ice facilitation caused by a freshening of the surface ocean in a susceptible region (near the sea-ice edge). Other results indicate that sea-ice facilitation by the cooling effect also plays a crucial role. As a consequence of these qualitative differences, the hosing approach is at least 50 years too late in reacting to the start of the perturbation and takes about 200 years to catch up with the response to the CF-icebergs (Fig. 3).

At the peak of the event (Figs. 2, 5), the sea-ice anomaly is largely a function of the remaining NADW formation. Compared to the cool & fresh-icebergs, by this time hosing has caused exaggerated freshening of the Eastern North Atlantic surface ocean (Fig. 5), which has lead to a too-efficient shut-down of the AMOC (Fig. 3) resulting in a too-cold Eastern North Atlantic (Fig. 5). The implausible build-up of an Eastern fresh SSS anomaly, subsequently leads to a ~50-years delayed recovery of the circulation.

Cool & fresh-icebergs are almost as efficient as hosing at weakening the AMOC, despite the more westward distribution of icebergs' melting fluxes and despite the generally too-efficient capability of homogeneous hosing to inhibit convection. Since the CF-icebergs lead to (much) stronger suppression of deep water formation than F-icebergs, while the cooling by itself would (traditionally) be expected to enhance convection, sea-ice facilitation must play a key role here.

Early in the Heinrich event (Fig. 4), icebergs have a more direct (earlier and stronger) impact than hosing. They release relatively large cooling and freshening fluxes near the western-most convection sites, which synergizes with direct (melting-flux related) sea-ice facilitation, leading to an earlier sea-ice expansion and climatic response.
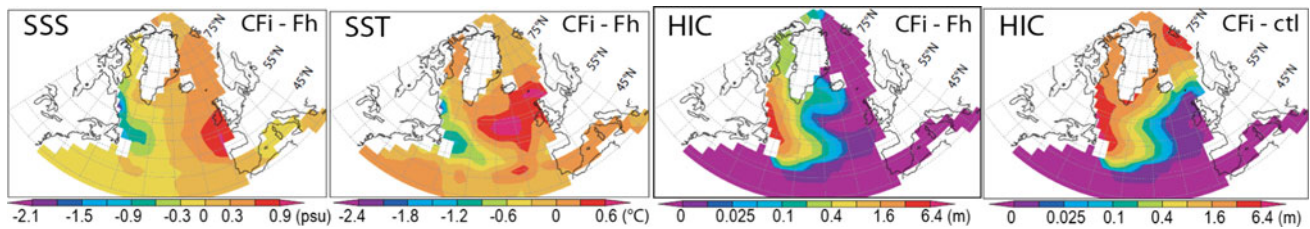
Fig. 5 Icebergs versus classic fresh water hosing at event peak. Anomalous sea surface salinity (SSS *left*), temperature (SST 2nd) and sea-ice thickness (HIC *3*rd) for Cool & Fresh icebergs (CFi) minus classic fresh water hosing (Fh). On the *right* sea-ice anomaly (HIC) for Cool & Fresh icebergs minus Control (ctl). Note logarithmic scale for HIC. All averaged over years 251–300

In fact, the CF-icebergs quickly cause the build-up of a thick pack of sea-ice (up to 6 m thick averaged over the first 50 years) near Newfoundland (Fig. 4: HIC). Given the fact that non-cooling icebergs do not have such a strong impact on the thickness of the sea-ice, and the fact that a closed sea-ice cover minimises the (sea-surface) erosion aspect of iceberg deterioration, this is probably due to the cooling of deeper, sub-surface, layers by the CF icebergs. In this context we would like to point out that Hulbe et al. (2004) have postulated, on the basis of climate controlled meltwater infilling of surface crevices, that "peripheral ice shelves, formed along the eastern Canadian seaboard during extreme cold conditions, would be vulnerable to sudden climate-driven disintegration during any climate amelioration. Ice shelf disintegration then would be the source of Heinrich event icebergs." From this point of view (and with our CF-icebergs apparently initiating the growth of an ice-shelve), icebergs originating (possibly surging) from the Laurentide ice sheet, could seed and feed the build-up of an ice-shelf, which would then be vulnerable to rapid ("catastrophic") disintegration. Similar scenarios could apply to other (e.g. Greenland, Norwegian) ice sheets. An interactive picture emerges of surging ice-sheets not only affecting the ocean and climate directly, but also initiating the growth of ice-shelves. The secondary thickening of sea-ice near Spitsbergen/northern Norway (Fig. 5 *HIC*) illustrates that the climatological feedbacks could also play an important role in the growth of ice-shelves. In turn, catastrophically disintegrating ice-shelves can become a source of iceberg armadas (Hulbe et al. 2004; Moros et al. 2002; Polyak et al. 2001).

The mechanism sketched here is not only relevant for Heinrich events, but could also help to improve our understanding of the more frequent Dansgaard/Oeschger events, which are generally characterized by a less specific and possibly sea-ice rafted IRD signature, implicating multiple sources (Bond and Lotti 1995), which hints at (synchronously disintegrating) ice-shelves as a likely source. Of course, further research encompassing dynamic ice shelves and a plausible interaction with CF-icebergs would be required to confirm the here postulated "ice-shelve facilitation" by (Cool & Fresh) icebergs. Indeed, other theories have been formulated to try explaining the likely origin of the

Heinrich events. Particularly noteworthy are theories relying on sea-level changes (Fluckiger et al. 2006) and on sub-surface warming of the ocean bathing the ice-shelves (Fluckiger et al. 2006; Alvarez-Solas et al. 2010, 2011; Marcott et al. 2011). Not all mechanisms apply to ice-sheets on both sides of the Atlantic Ocean but there might be multiple triggers that may cause the ice-shelf disintegration, hence lifting the requirement for a global mechanism affecting all northern hemisphere ice-sheets. The mechanism presented here may as well play a significant role, though assessing its significance among the other possible mechanisms is clearly beyond the scope of this study.

We conclude:

- Dynamic thermodynamic icebergs are much more efficient at suppressing deep-water formation when the latent heat flux associated with the phase transition from ice to water (the cooling effect) is not neglected. Sea-ice facilitation is a dominant mechanism in explaining this convection-suppressing efficiency of Cool & Fresh icebergs. It appears that sea-ice facilitation by CF-icebergs could also stimulate the growth of ice-shelves.
- Compared to traditional homogeneous fresh-water hosing, Cool & Fresh-icebergs are almost as efficient as hosing at weakening the meridional overturning circulation, even though hosing can be regarded as an in principle too-efficient manner to suppress convection (Jongma et al. 2009; Levine and Bigg 2008).
- Compared to Cool & Fresh icebergs, despite the general similarity of nearly complete AMOC-shut-down at the peak of the event, Ruddiman belt hosing is not a good representation for Laurentide ice-sheet disintegration (a Heinrich event) because of a number of qualitative differences:

  1. Hosing is less efficient at directly facilitating the formation of sea-ice, causing a delayed (50–200 years) climatic response of the system to the onset of a Heinrich event.
  2. Hosing causes exaggerated freshening (and cooling) of the Eastern North Atlantic, resulting in a delayed (∼50 years) recovery of the circulation after the event.

3. The more realistic dynamic CF-icebergs distribution leads to a fresher and also several degrees colder western North Atlantic at the start, as well as at the peak of the event.

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