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Modelling abrupt glacial North Atlantic freshening: rates of change and their implications for Heinrich events

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Abstract

The abrupt delivery of large amounts of freshwater to the North Atlantic in the form of water or icebergs has been thought to lead to significant climate change, including abrupt slowing of the Atlantic Ocean meridional overturning circulation. In this paper we examine intermediate complexity coupled modelling evidence to estimate the rates of change, and recovery, in oceanic climate that would be expected for such events occurring during glacial times from likely sources around the North Atlantic and Arctic periphery. We show that rates of climate change are slower for events with a European or Arctic origin. Palaeoceanographic data are presented to consider, through the model results, the origin and likely strength of major ice-rafting, or Heinrich, events during the last glacial period. We suggest that Heinrich events H1-H3 are likely to have had a significant contribution from an Arctic source as well as Hudson Strait, leading to the observed climate change. In the case of H1 and H2, we hypothesise that this secondary input is from a Laurentide Arctic source, but the dominant iceberg release for H3 is hypothesised to derive from the northern Fennoscandian Ice Sheet, rather than Hudson Strait. Earlier Heinrich events are suggested to be predominantly Hudson Strait in origin, with H6 having the lowest climate impact, and hence iceberg flux, but H4 having a climate signal of geographically variable length. We hypothesise that this is linked to a combination of climate-affecting events occurring around the globe at this time, and not just of Laurentide origin.

Key-words: Heinrich events; modelling; Quaternary; icebergs

1. Introduction

The sedimentary record in the glacial (Marine Isotope Stage 2-3) North Atlantic domain is rich in complexity. Marine, ice and terrestrial records show evidence for long term climate decline interspersed by periods of more pronounced temperature decline and sharp, but short-lived, temperature rises (Fig. 1). Marine records show clear evidence for a number of times of sedimentary fall-out from enhanced iceberg rafting in the North Atlantic – Heinrich events – during this period (Hemming, 2004). There is also evidence for catastrophic freshwater outbursts from under ice sheets or ice-dammed lakes (e.g. Fisher, 2003; Leverington and Teller, 2003; Lekens et al., 2006; Murton et al., 2010). The Younger Dryas (YD) is the canonical event of this type, but while having a St. Lawrence origin, there is evidence that other, later contributions from the Arctic (Leverington and Teller, 2003) and Northern Europe (Nesje et al., 2004) may have prolonged the freshwater supply.

This tendency for abrupt change is centred in the Northern Hemisphere, and particularly the North Atlantic, having little signature in the Antarctic (Voelker, 2002). The classical portrayal of these signals is seen in the oxygen isotope, and hence temperature, record of the Greenland ice sheet, where the abrupt warmings became known as Dansgaard-Oeschger (D-O) events (Dansgaard et al., 1984). The linkages between these semi-periodic events and the ice-rafting peaks of Heinrich events became hypothesised as part of the Bond cycle (Bond et al., 1999), with a series of D-O events caused by stochastic freshwater forcing leading to ice accumulation on North American ice sheets that then became unstable and purged ice (Timmermann et al., 2003). These massive releases of icebergs are hypothesised to lead to major climate change and cessation of the Atlantic thermohaline circulation (Broecker, 1994).

However, the origin of these abrupt climate changes is not well established, even though there is good evidence for their existence in various forms of palaeoclimatic archives. A recent review by Clement and Peterson (2008) surveyed the many records around the world, demonstrating the existence of the abrupt climate change during the last glacial period and discussed the three main mechanisms proposed for their cause: ocean thermohaline circulation change, sea-ice feedbacks and tropical processes. Their well argued conclusion was that none of these fitted the observations when compared with models of the different processes and that more work considering other, or combined, feedbacks using coupled climate models are required to understand abrupt change.

A subset of environmental change during the last glacial period consists of Heinrich events. These are periods, of 500±250 years duration (Hemming, 2004) occurring roughly every 10 000 years, with extensive deposits of ice-rafted debris (IRD) in the North Atlantic marine record. These peaks of IRD are normally ascribed to episodic iceberg releases from the Hudson Strait Ice Stream of the Laurentide Ice Sheet, set off by binge-purge oscillations within the ice sheet (MacAyeal, 1993). There is good lithological evidence linking the IRD to North America (e.g. Grousset et al., 1993; Gwiazda et al., 1996a). However, there are other theories for their generation, and alternative possible sources, including the Fennoscandian Ice Sheet, particularly for Heinrich events H3 (~ 30 000 cal. yr B.P.) and H6 (~60 000 cal. yr B.P.) (Gwiazda et al., 1996b). These two events appear to be smaller in magnitude and may have multiple sources, or have an insufficiently large primary source to overwrite the lithic signature of more normal glacial IRD levels in the eastern Atlantic. It can sometimes be difficult to distinguish lithic signatures from the two sides of the North Atlantic (Farmer et al., 2003), however. Hemming (2004) gives an excellent review of the state of knowledge concerning Heinrich events, their causes, origins and the spread of IRD.

Whatever their origin, Heinrich events provide an unequivocal signal of disturbance to the marine environment through enhanced iceberg fluxes. There is also evidence of disturbance to the atmosphere on a hemispheric scale, through enhanced dust deposits from Asia in the Greenland ice cores (Biscaye et al., 1997), and, since Bond et al. (1993), a link has frequently been made between climate cooling, followed by abrupt warming, and Heinrich events. The classic picture is that the release of icebergs into the North Atlantic, and their subsequent melting, stabilises the surface ocean, preventing deep convection, and so shutting off the Atlantic meridional overturning circulation (Broecker, 1994). A wide range of climate modelling experiments have demonstrated that this scenario is consistent with climate physics (e.g. Rind et al., 2001; Ganopolski and Rahmstorf, 2001; Vellinga and Wood, 2002; Stouffer et al., 2006; Levine and Bigg, 2008; see Clement and Peterson (2008) for a full review).

In this paper we take an intermediate complexity climate model, spun-up for glacial climates and with iceberg-ocean coupling embedded within it (Levine and Bigg, 2008), to examine the rates of climate change in the glacial world consistent with a range of release rates of icebergs and freshwater into the North Atlantic and Arctic, from a range of possible source regions. The modelled rates of change are then compared with observed rates of change during the Younger Dryas and Heinrich events H1-H6 in a range of climate-related indices. This allows us to comment on the origin and magnitude of these unequivocal abrupt freshwater events, and their true, individual, climate impact.

We first discuss the climate model and the range of experiments used to simulate freshwater or Heinrich event-led change to the Atlantic overturning. The results of the modelling experiments are then discussed. We next present the high temporal resolution glacial ocean, terrestrial and cryosphere proxy climate indices, followed by a comparison of the modelled rates of change with rates of change found in these climate indices around the time of Heinrich events. We conclude by discussing the implications of these comparisons for the strengths and origins of the Younger Dryas and H1-H6, and the consequences for our understanding of abrupt change during glacial times.

2. Model

The climate model used is an intermediate complexity coupled ocean-atmosphere model, with an energy balance atmosphere derived from that of Fanning and Weaver (1996) and a curvilinear coordinate ocean model, whose North Pole has been displaced to central Greenland (Wadley and Bigg, 1999). There is a free surface to the ocean (Webb, 1996) and a dynamic and thermodynamic sea-ice model at the interface of the ocean and atmosphere (Wadley and Bigg, 2002). In addition, icebergs are allowed to move, and melt, within the model in a way that is coupled to the ocean model processes (Levine and Bigg, 2008). The model's curvilinear grid enhances model resolution in the North Atlantic and Arctic, and in particular in the Greenland and Labrador Seas. Typically, in the Nordic Seas the horizontal resolution is 1–2^{o}, whereas in the Southern Hemisphere it is 6–8^{o}. Time step length is a function of grid spacing, to allow efficient integration of the variable resolution grid (Wadley and Bigg, 1999). Full details of the model can be found in Levine and Bigg (2008).

Here we are interested in abrupt change during glacial times, so the model experiments all start from a glacial control state. This state, and the equivalent representation of the climate for a present day control, is described in Levine and Bigg (2008). The background iceberg flux for the glacial Northern Hemisphere uses that calculated by Bigg and Wadley (2001), based on a steepest gradient algorithm draining atmospheric precipitation fields, from the atmospheric general circulation model runs for the Last Glacial Maximum (LGM) by Dong and Valdes (1998), off the Peltier (1994) ice sheet in a state of mass balance. For the Southern Hemisphere we use climatological iceberg fluxes from Gladstone et al. (2001) that are based on Present Day mass balance calculations for the Antarctic ice sheet. The Antarctic ice sheet has decreased in volume since the LGM, however, there is evidence from cores taken at various latitudes in the South Atlantic that IRD delivery was at a minimum at periods surrounding the LGM and during the Holocene (e.g., Kanfoush et al. (2000)). We assume the Antarctic ice sheet to be in a steady state for both PD and LGM simulations and thus use the PD iceberg fluxes for both PD and LGM simulations.

2.1. Control Run

The performance of the PD model provides guidance for interpreting the reliability of the glacial simulations. Thus, the PD coupled model (see Levine and Bigg, 2008) has a rather high peak North Atlantic overturning of 28.1±0.3 Sv although the amount of North Atlantic Deep Water (NADW) that flows southward across the equator contributing to the global thermohaline circulation is only ~13 Sv, which is reasonably consistent with observations (Gordon, 1986; Schmitz, 1995). The PD sea surface properties compare reasonably well with the climatological values, although gradients are not fully resolved. This is particularly true for the modelled meridional temperature gradients across the Southern Ocean, because of the relatively coarse resolution of the grid in this region, leading to a weak Antarctic Circumpolar Current (63.5±4.2 Sv) compared to observations (130-140 Sv, Nowlin and Klinck (1986)). However, the temperature and salinity distribution in the northern North Atlantic and Nordic Seas corresponds quite well with the climatology and leads to realistic North Atlantic Deep Water formation, in terms of convection location and depth penetration. The tropical sea surface temperature (SST) and air temperature are a little low, leading to reduced evaporation and fresher tropical sea surface conditions than observed, with salinity anomalies over 1 psu. There is also less sea-ice in both hemispheres. In the Northern Hemisphere this anomaly is mainly on the continental shelves of the Arctic Ocean, and so does not directly influence the Atlantic convection areas. The Southern Hemisphere sea-ice areas of 1.7±0.1 million km^2 compares with observations of around 11 million km^2 for the annual mean (Cavalieri et al., 1997). However, poor reproduction of PD Southern Ocean sea-ice is a common climate model problem (Hansen et al., 2007).

The strength of northern Atlantic currents is an important factor in the speed with which salinity anomalies are moved around in the ocean, so it is relevant to examine the PD simulation's performance here. The strength of the North Atlantic sub-polar gyre (20.5±0.2 Sv) compares reasonably with the range in the literature (13-16 Sv according to Tomczak and Godfrey (2003)). The Denmark Strait Overflow in the model is 5-6 Sv, while observations suggest it varies between 3-4 Sv (Macrander et al., 2005). Through the other route into the Nordic Seas the Faroe Shetland Channel has ~ 9 Sv of Atlantic inflow in the model, but observations suggest ~ 7 Sv (Østerhus et al., 2005). The model therefore produces reasonable, but slightly too large, fluxes into and out of the Arctic.

In the glacial control run the main North Atlantic convection occurs to intermediate depths, with a maximum penetration to 1800 m. This is centred at 45^{o}N in the central and eastern Atlantic, which is consistent with other studies suggesting the shallower convection of the last glacial period occurred south of Iceland (Seidov and Maslin, 1999). The strength of this is only a third (9.6±0.3 Sv) of the PD peak overturning. This is within the uncertainty ranges provided by palaeo-observations (see Levine and Bigg (2008) for a fuller discussion). Estimates of coupled models of the LGM vary widely for this quantity (Weber et al., 2007), and the present model falls within this range of model uncertainty. Our modelled sea surface temperatures in the tropics and subtropics are lower than CLIMAP, which is consistent with the MARGO glacial analysis (Kucera et al., 2005). However, the winter limits of near freezing ocean surface temperatures in the model between 40–50^{o}N are further south than in the MARGO reconstructions. Thus, our model produces sea-ice all year round in the Nordic Seas, while this area is thought to have been seasonally ice-free (Pflaumann et al., 2003; Kucera et al., 2005). This problem is similar to the experience of other LGM coupled models (Kageyama et al., 2006). In the SH the Drake Passage flux (88.6±2.3 Sv) is a third higher than for the PD simulation. There is a five-fold increase in SH sea-ice area, to 9.3±0.9 million km^2. This amount of Southern Ocean sea-ice is still approximately 20% lower than the PD observed annual mean but covering a much more realistic extent of ocean.

2.2. Experiments

In this paper we are examining the signature of abrupt change resulting from known catastrophic adjustments to last glacial ice sheets, either through fresh water release (e.g. the Younger Dryas) or iceberg melting (the Heinrich events). Consequently, we performed a number of experiments where the basic glacial control run of the model was perturbed by freshwater or iceberg additions from a number of possible release locations around the North Atlantic and Arctic periphery (Fig. 2). The length of time during which the perturbation was imposed was determined from estimates in the literature. The extensive review of Heinrich events in the palaeoceanographic literature by Hemming (2004) suggested a period of 500

±250 years. We have therefore imposed our perturbations for 500 years, after 5500 years of a glacial control simulation, and run on the experiments for at least an additional 500 years to study the rate of return of the climate towards an unperturbed state. Green (2009) showed that while changing the duration of the pulse affects the detail of the response, the general character is the same. We have examined a range of possible freshwater-equivalent release rates, ranging over 0.1-0.4 Sv. These match with estimates from the palaeoceanographic (Hemming, 2004; Roche et al., 2004) and modelling (Calov et al., 2002) communities; they also cover a range over which our model response varies from a circulation perturbation to a complete collapse (Levine and Bigg, 2008).

The release locations have been chosen through field evidence of known or suspected catastrophic events. The ice stream feeding Hudson Strait has long been acknowledged as a likely source for Heinrich events (Broecker et al., 1992) because of the lithology of North Atlantic IRD, the latter's strong carbonate content, which is characteristic of sediments underlying Hudson Bay, and the geographic pattern of IRD deposition. The ice stream draining through the Gulf of St. Lawrence overlay areas of similar geology to the Hudson Strait ice stream, and would have produced icebergs feeding into the same North Atlantic IRD pattern. There is evidence that IRD in Heinrich deposits has a signature consistent with at least a contribution from the Gulf of St. Lawrence (Piper and Skene, 1998; Piper and DeWolfe, 2003). We have therefore used this as a second release location. Further afield, there has been debate over a European origin for H3 and H6. While this now seems less likely (see Hemming (2004) for a review) there is increasing evidence for an ice-bridge across the northern North Sea (Sejrup et al., 2009) and a major ice stream in the Norwegian Channel (Nygård et al., 2007), either of which are possible candidates for major ice release from the southern arm of the Fennoscandian Ice Sheet, at least for H3 (Lekens et al., 2009).

The Arctic also provides potential release sites for either iceberg or freshwater releases. Iceberg scour marks and erosion on the deep Lomonosov Ridge in the central Arctic and the Yermak Plateau northwest of Svalbard are consistent with a catastrophic release of deep draft icebergs from the northern Barents Sea section of the Fennoscandian ice sheet (Kristoffersen et al., 2004; Green et al., 2010). There is evidence that the very deep St. Anna Trough in the eastern Barents Sea continental margin had ice grounded to its base during the last glacial period (Polyak et al., 1997). This, or the nearby Franz Victoria Trough (Green et al., 2010), therefore represents a possible source for a catastrophic release of icebergs. Finally, the Mackenzie basin and M'Clure Strait in western Arctic Canada drained the Keewatin Dome of the Laurentide Ice Sheet. The Mackenzie is likely to have been the route down which there was a freshwater release during the Younger Dryas, through a partial collapse of Lake Agassiz (Teller et al., 2002, Murton et al., 2010). There were also major ice-rafting events during the Last Glacial from the M'Clure Strait in the western Canadian Archipelago (Stokes et al, 2005; Darby and Zimmerman, 2008). This therefore forms the fifth source region for our iceberg and freshwater release experiments (Fig. 2).

3. Modelling Results

In the modelling results presented below there is an intrinsic assumption that sufficient ice was present in the catchment of each release site during the Last Glacial for our range of fluxes to be possible. This may, or may not, be true for any particular time during this long time period but allows comparison of the impact between different release sites. Later sections will address the probability of such releases within the palaeoclimate data assessment.

3.1. Iceberg experiments

The clearest variable to show the response of the climate to the iceberg forcing is the strength of the peak Meridional Overturning Circulation (MOC) of the North Atlantic. This is shown for the 1000 years following the start of release of icebergs at fluxes of 0.1 Sv, 0.2 Sv and 0.4 Sv from the five release points around the North Atlantic and Arctic in Fig. 3. The control run's MOC is also shown for comparison.

In all cases the 0.1 Sv release causes a decline in the strength of the MOC by 2-3 Sv, or 20%, with an eventual recovery of some extent. However, the speed of the decline, and the recovery, depend on the location of the iceberg release. The majority of the decline occurs within a decade from eastern North American releases, while from other locations it is slower, up to 1-200 years. There is also a difference in the rate of recovery once the iceberg release ceases in Year 6000 of the model run. The eastern North American release experiments show a rapid return of the MOC to values near those of the control. However, the recovery of eastern Atlantic and Arctic release experiments is significantly more gradual, taking at least a century, but up to 300 years from Mackenzie releases. In the case of the St. Lawrence release, the MOC does not recover to its original level but equilibrates at a new, slightly lower, MOC stength.

This tendency to equilibrate at a new MOC level on recovery is seen in several of the experiments with higher releases (Fig. 3). This new equilibrium state after recovery is not invariably one with a lower MOC, and hence colder North Atlantic, but can lead to a higher MOC (for example, the 0.2 Sv St. Anna Trough release). The equilibrium recovery behaviour is very dependent on the specific release location, but the rates of change respond to a broader geographical imperative. Thus, rates of initial decline are always very rapid from eastern North American releases whatever the release strength, while the behaviour of experiments from other release sites varies with the release magnitude. For these sites the 0.2 Sv and 0.1 Sv releases result in similar behaviour, however, the rate of decline is faster when collapse of the MOC occurs for 0.4 Sv releases, if still up to a few decades slower than for collapse generated from eastern North America.

Recovery rates are generally independent of the release magnitude, but dependent on the release site. Thus, for eastern North American release experiments recovery is extremely rapid, Mackenzie experiments recover over about a century, while NCIS and St. Anna Trough release experiments require several hundred years for recovery. Note that for the stronger releases only Hudson Strait and Mackenzie experiments show consistent recovery to pre-release MOC strengths.

The explanation for the sometimes striking differences between the oceanic responses to release location lies in where the fresh water from the iceberg melting enters the ocean and the consequent response of the ocean density field, currents and ocean and atmospheric temperature fields. The icebergs are released in a range of sizes (Levine and Bigg, 2008) and allowed to move, and melt, through the interaction of the icebergs with the ocean and atmosphere (Bigg et al., 1997). The bergs will melt rather slowly in cold conditions, and those originating in the Arctic may take some years to decades (Bigg et al., 1996; Green, 2009) to leave this Ocean. Rapid melting, and so addition of freshwater to the ocean, only occurs once the icebergs enter warmer and windier climates (Bigg et al., 1997). Note that this may not relate closely to where IRD is deposited (Death et al., 2006).

Due to all these factors, the model salinity fields reveal the path along which icebergs travel, and melt, once they are released, rather than the result of freshwater advection and diffusion from a point source. The sea surface salinity fields for the control and 0.4 Sv experiments are shown in Fig. 4 300 years after the catastrophic iceberg releases began. The release points on the eastern coast of North America show the movement of icebergs into the glacial Gulf Stream and North Atlantic Drift with freshwater entering this system and then moving south into the sub-tropical gyre re-circulation. Rather little impact is seen in the sub-polar gyre and Arctic, as few icebergs from these sources penetrate into such regions, and relatively little surface water advects unaltered from the modelled glacial sub-tropical gyre northwards. The freshwater does, however, enter the region of intermediate water formation in the central North Atlantic rather quickly, hence leading to rapid decline of the MOC, and almost as rapid a return once this source is cut-off and the fresh anomaly of the NW Atlantic has been advected past the convection region.

In the case of a Mackenzie release, Fig. 4 shows that much of the freshwater, as both ice and freshwater, leaves the Arctic in the East Greenland Current and then freshens the Labrador Sea and northwestern Atlantic. The Arctic is also freshened generally. Thus there is a delay in the MOC decline, as shown by Fig. 3, relative to eastern North American releases, as it takes longer for sufficient freshwater to enter the northern Atlantic, both from the delay due to the water transit time and the slow melting of icebergs in a cold Arctic and Greenland Sea. The cessation of iceberg input also leads to a slower recovery because it takes some decades to centuries, depending on the run, for the excess salinity built-up in the Arctic to be flushed out into the ocean more generally.

The European releases show this delay in the onset of MOC decline (Fig. 3), but respond rather differently to North American releases thereafter. Many of the NCIS icebergs go north and melt in the Nordic Seas, or Arctic, while relatively few of the St. Anna Trough icebergs get entrained into the East Greenland Current and exit into the NW Atlantic. In both cases, this leads to large-scale freshening in the Northeastern Atlantic and eastern Arctic, which creates a pool of low salinity that gradually leaks out into the Atlantic, and results in a continuation in MOC decline beyond the time of initial response, unless the circulation was shut down (Fig. 3). Similarly, once the iceberg input ceases this slow leaking of freshwater significantly delays the return to a strong MOC. It must be remembered that there is a net decrease in global salinity due to the Heinrich events, but while North American inputs tend to get mixed globally to minimise the net impact of this on the MOC, the eastern Arctic input leads to a long-term decrease of the North Atlantic salinity field. In this case, areas between ~50-60^{o}N retain upper ocean salinity values some 1 ‰ lower even 500 years after the iceberg input has ceased.

The SST patterns tend to be similar for the different events because they are strongly tied to the strength of the MOC. Thus, there is significant cooling over the central Atlantic, as the North Atlantic Drift adjusts southwards, and some weak warming further north and south (Levine and Bigg, 2008). This is similar to the results of Vellinga and Wood (2002) for a present day freshwater release. For atmospheric temperature anomalies, again there is cooling over the North Atlantic, whose centre and magnitude varies depending on the release site (see Levine and Bigg, 2008). The Mackenzie and NCIS releases produce the maximum cooling, and so biggest overall climatic effect. Note that all releases lead to some, at least localised, warming; for the Hudson Strait release circum-Arctic warming is a strong characteristic, with the maximum warming centred over northern Greenland (Fig. 5). Releases from the St. Anna Trough also lead to localised warming over northeastern Greenland, and slight warming over the eastern Arctic and much of northern Eurasia. Releases from the St. Lawrence, NCIS and to a lesser extent, the Mackenzie, result in significant western European cooling.

3.2. Freshwater experiments

The MOC values for the freshwater release experiments are shown in Fig. 6. These are similar to the iceberg releases in character, both in terms of the relative rates of change and the recovery. In the case of releases from eastern North America, the main difference is that the MOC reacts more to weaker inflows, but the rates are very similar as the fresh water is effectively injected into the same current systems in both iceberg and freshwater releases. In contrast, the Mackenzie release has less impact per unit freshwater equivalent release because the freshwater enters the western Arctic directly, rather than being carried further towards the North Atlantic as icebergs before release. More of the freshwater therefore remains in the Arctic for longer, reducing the impact, although also slowing the recovery somewhat.

Freshwater releases from the European sites show substantially larger impacts on the MOC than do similar iceberg releases. The freshwater in both cases reaches the central North Atlantic convection zone in a few decades and caps the ocean. In fact, the releases from both the NCIS and St. Anna Trough cause such rapid change that the model becomes numerically unstable for larger releases.

3.3. Experimental summary

Several key differences between sites of release and the type of freshwater release are apparent from these numerical experiments. Firstly, whatever form the release takes, inputs from eastern North America cause substantial and rapid change in North Atlantic climate, with equally rapid recovery to states similar to the original climate. Secondly, Arctic and European releases of icebergs show a slower response of several decades to centuries for climate cooling, and similar, slower, timescales of recovery. Thirdly, experiments with freshwater releases from these sites show a rapid onset of climate cooling but a slower recovery than is the case for experiments with eastern North American inputs.

The extent of the climate effect also depends on the release location and type. Iceberg inputs show a more linear variation with MOC decline, whatever the release site, until effective collapse occurs, than is the case for freshwater releases. For the latter, the MOC is most sensitive to releases from eastern North America (although the numerical problems caused by rates of change may distort this result).

We have here considered idealised experiments for a particular time, and hence orbital parameter, atmospheric carbon dioxide concentration and ice sheet configuration. We have also considered single release experiments, rather than the impact of multiple release sites on glacial ocean circulation and climate, in order to disentangle the basic signature deriving from each release site. It is possible, indeed likely, that any observed palaeoclimatic signal will not have been due to such a pure event as we have modelled. Nevertheless, sensitivity experiments that we have performed using different glacial forcings and combinations of releases suggest that there is sufficient signal produced in the idealised experiments for us to usefully proceed to explore the palaeoclimatic record. For example, mixing equal strength Arctic and Hudson Strait releases produces a response dominated by the Hudson Strait release, as the latter affects the convection site first, because of its proximity.

These idealised modelled differences therefore mean that it should be possible to attempt to infer rates, types and locations of releases from the rates, and absolute magnitudes, of change in the palaeoclimate record. In the following section we will examine such records with high temporal resolution, concentrating on the known freshwater release of the Younger Dryas (c. 11-12 000 cal. yr B.P.) and the known iceberg releases of Heinrich events H1-H6 during the Weichselian.

4. Palaeoclimate Analysis

4.1. Representative data sets

To compare the numerical experimental conclusions with palaeodata it is necessary to look at a representative set of high resolution (sub-centennial) but long-term records covering the North Atlantic region where the climatic impact was seen to be strongest in the numerical experiments (Fig. 5). It has not been possible to find appropriate datasets to cover the whole of the period back to 70 000 cal. yr B. P. for all areas, but a representative sample of different geographical regions and data types has been selected. Fig. 7 shows the location of the datasets chosen and Fig. 8 shows the various timeseries on a common timescale. These timeseries address different aspects of climate variability in regions where the model experiments showed most clearly defined differences between the different release experiments. No exactly comparable proxies with the necessary temporal resolution and scientific basis were discovered across the whole North Atlantic, where model difference was greatest. However, coverage is available in some manner in most crucial areas; we discuss drawbacks as well as advantages to using potentially problematic datasets in what follows.

Ice cores from Greenland provide an anchor against which many studies compare their local results; in addition Greenland is an area where we expect significant atmospheric climate change to be seen for the abrupt changes to be studied (Fig. 5). The GISP2 ice core record from central Greenland provides a temperature reconstruction from oxygen isotope and ice accumulation records extending back to 50 000 cal. yr B. P.. The original reconstruction derives from Cuffey and Clow (1997), with smoothing by Alley (2000). Another important indicator of change is the sea level record. A high resolution record of sea level should show the rate of transfer of freshwater, as either ice or water, from land to sea, and hence be a proxy for the temporal length and magnitude of an iceberg or freshwater release respectively. Siddall et al. (2003) used a combination of an oxygen isotope record from a Red Sea core with a hydraulic model of exchange between the Red Sea and the Arabian Sea to reconstruct sea level change, down to centennial scale for much of the last 70,000 years.

The main indicator used for comparing the numerical experiments was the MOC strength (Figs. 3 and 6). There are few modern day records of this, let alone palaeo-records. However, one proxy is the sortable silt grain size of bottom sediments (McCave et al., 1995) under the deep return flow of the MOC in the North Atlantic. We use here such a timeseries, extending over 26-62 000 cal. yr B. P., that has been sampled at centennial to sub-centennial scale from ODP Site 1060 on the Blake Outer Ridge of the western Atlantic by Hoogakker et al. (2007). The mean grain size of the 10-63 μm sediment fraction was used. Larger sizes imply stronger currents, and hence a stronger MOC. Previous work using this parameter, and a discussion of its advantages and drawbacks, can be found in McCave and Hall (2006).

The biggest climatic impact of abrupt North Atlantic freshwater injections is found in the North central Atlantic (e.g. Fig. 5). Hence SST indicators from either side of the Atlantic are also used. From the Gulf Stream dominated western region, a Marine Isotope Stage 3 record

(24 -64 000 cal. yr B. P.) of faunal and alkenone reconstructed SSTs is available (Vautravers et al., 2004) from the same site, ODP 1060, as is used for the MOC proxy. This will tell us about both the local atmospheric and upper ocean climate variability. In the eastern Atlantic, core MD01-2444 underlies a seasonally active upwelling zone off the Portuguese coast. The northern limit of this upwelling will move latitudinally as climate fluctuates, leading to SSTs from this site being a sensitive indicator of climate change. We use a faunal SST reconstruction covering a similar time period to that of the western Atlantic SST site to represent this area of North Atlantic climate (Vautravers and Shackleton, 2006). Another sensitive oceanic indicator of climate change is the exchange of water between the Mediterranean and the Atlantic through the Strait of Gibraltar, as this tells us something about the relative densities of the eastern Atlantic and Mediterranean, and therefore acts as a regional climate proxy (Rogerson et al., 2010) . We use a high resolution record of alkenone-derived SST in the Alboran Sea, at site MD952043, as an indicator of conditions near the exchange; this dataset is available back to 52 000 cal. yr B.P. (Cacho et al., 1999).

The numerical experiments indicate that the climate anomaly caused by some Heinrich events is likely to have spread over Europe (Fig. 5). Thus, from Lago Grande di Monticchio in southern Italy a high resolution 100 000 year record of carbon content in the lake sediments, measured through their loss fraction on ignition and the biogenic silica content (Allen et al., 1999) is used. These two indicators are linked to the proportion of the sediment entering the lake from erosion of bare or forested environments, thus high weight percentages for both are characteristic of high organic fractions in runoff, while low values suggest barer soils with rather low organic content. Both can approach zero in particularly cold climates. Such records may respond strongly to local topographic influences as much as the wider, regional climate. However, this particular record correlates strongly with changes in the Greenland ice core (Allen et al., 1999), suggesting that it is largely responding to large-scale, rather than local, influences.

A second terrestrial site is chosen from Brazil, as the numerical experiments suggest the cooling during Heinrich events caused by Hudson Strait releases may have led to a western Atlantic-centred cooling extending to South America, with potential impact on tropical climate teleconnections (Fig. 5). High resolution stalagmite oxygen isotope records, extending back to 116 000 cal. yr B. P., from the sub-tropical Botuverá Cave (Cruz et al., 2005) are used to examine climate change in this region. These data also act as an indicator of climate change in the Southern Hemisphere, to gauge the cross-equatorial spread of any abrupt change. The oxygen isotope record appears to mostly be a reflection of the local precipitation record (Cruz et al., 2005), with less negative values reflecting more regional winter than summer rainfall because of a weakening of the strength of the local summer monsoon, which imports moisture from afar and is characterised by intense convection, during such times. Thus, the long-term signal is dominated by the precessional (21 000 year) orbital cycle. Nevertheless, over the short-term we can ignore this trend and examine evidence for abrupt change in the isotopic rate of change.

4.2. Evidence for abrupt change

To examine our selected palaeoclimate records for abrupt change we first need to specify the time periods to examine. These are shown in Table 1. We follow Hemming (2004) for estimates of the timing of Heinrich events, and Alley (2000) for the Younger Dryas. The durations of these events are estimated as 500±250 years (Hemming, 2004) for Heinrich events and 1500 years for the Younger Dryas, with a number of freshwater releases maintaining the oceanic freshening (Teller et al., 2002) in the case of the latter. In intercomparing records the question of the relative accuracy of their chronologies arises. Records with very well established calendar year chronologies have been chosen to minimise this problem, and in general, as we will see, there is very good agreement on the relative timings of events. However, here we are interested in rates of change in what are normally significant events. Thus, slight off-sets in the absolute time between the different records are not a major problem.

In the case of the Younger Dryas (YD) and Heinrich Event 1 (H1) a number of the chosen high resolution datasets do not cover this period so an additional high resolution, but shorter, SST dataset has been added from the Caribbean basin (Lea et al., 2003). Variations in these data are thought to indicate changes in the movement of the ITCZ (Lea et al., 2003), but they may also be linked to changes in the input of warm water to the sub-tropical gyre.

The palaeohydrological reconstructions of Leverington and Teller (2003) and Nesje et al. (2004) offer the current view of the YD being largely flood-induced, but with successive flooding from a range of locations prolonging the cooling event. Figure 9 shows a comparison of the various datasets through the YD. The Greenland temperature record shows very abrupt onset c. 12.9k cal. yr B.P. and abrupt recovery c. 11.6k cal. yr B.P.. These dates correspond quite well with change in the SST records across the Atlantic, particularly with an abrupt onset of cooler conditions. Recovery is also abrupt in the ice core record, but less so in both SST records around the same time. The Brazilian ITCZ record also suggests a change to weaker ITCZ convection during the YD, indicating cooler conditions reaching into the Southern Hemisphere. The southern European terrestrial record shows little sign of the abrupt return to glacial conditions; the sharp spike around 12.2k cal. yr B.P. is seen in just one point and may be a data problem. The abrupt onset and slower recovery in SST, combined with a weak southern European climate response (see Fig. 5), is consistent with the modelling response of a St. Lawrence flood initiating the YD, but lake drainage from other sources and directions, such as the Mackenzie and Baltic, prolonging it. This is also consistent with the palaeohydrological reconstructions.

We now turn to Heinrich events, starting with H1. This was the main deglaciation iceberg release, which has a distinct and abrupt signature in the palaeo-records (Fig. 10), with the rise in sea level beginning c. 17.8k cal. yr B.P., approximately coinciding with abrupt fall in the Alboran Sea SST (f in Fig. 10), a decrease in the carbon content in runoff in Italy (g and h in Fig. 10), a fall in rainfall in the ITCZ proxy in Brazil and short-lived dips in Greenland temperature and Caribbean SST (* in Fig. 10). These correspondences are consistent with a Hudson Strait iceberg release, particularly seen through the mid-latitude abrupt change but limited Greenland response (e.g. as seen in Fig. 5). However, the Alboran Sea SST shows a gradual, rather than abrupt, return over several hundred years to pre-event temperatures, with short-term coinciding temperature drops in Greenland (a in Fig. 10), the Caribbean and southern Europe during this period (g & h in Fig. 10). This contrast in rates of change between the beginning and end of H1, through comparison with the modelling results, suggest that H1 may have consisted of two events that affected climate: an initial Hudson Strait release followed by an Arctic or European release. The magnitude of the European signal suggests a possible Mackenzie source (Fig. 5). Several authors have found evidence for European IRD events preceding H1 (Grousset et al., 2000; Peck et al., 2006; Peck et al., 2007); there is also evidence for a Laurentide iceberg release into the Arctic about the same time as H1 (AL2; Darby et al., 2002). On the basis of the comparison of modelling and palaeo-records in Fig. 10, we hypothesise that any British-Irish Ice Sheet (BIIS) precursor to a Hudson Strait release was not large enough to have a significant climate impact, but that part of the North American response of the Laurentide ice sheet saw a later iceberg release enter the Arctic that continued the climate event associated with H1, leading to a slower recovery than there would otherwise have been.

The palaeoclimate data for the previous Heinrich event, H2, are shown in Fig. 11. The Atlantic SST data only begin during the event, so just the start of H2 is shown in these two fields (d & e in Fig. 11). However, across a wide number of variables onset of a cooling event is seen between 24.1-24.4k cal. yr B.P.. In most fields this onset is abrupt, with the majority of change occurring in less than 200 years. The exception is the Greenland temperature (a in Fig. 10), where there is relatively little impact. The bottom-sediment grain-size variable (c) is low throughout much of the interval; as will be seen repeatedly there is only a loose temporal association between this variable and what is occurring in the atmosphere and the surface ocean. The rising sea level (b in Fig. 11) suggests that iceberg loss continued until around 23.3k cal. yr B.P.. The Greenland temperature, Mediterranean SST and Italian biogenic silica proxy all return to pre-event levels around this time, suggesting the recovery follows the cessation of enhanced iceberg flux quite quickly. This style of response is compatible with a predominantly Hudson Strait release, as H2 is normally considered to be (Hemming, 2004). However, the recovery in SST is slower than the initiation of change, which could be due to a climatic impact from the coinciding Arctic IRD event AL3 (Darby et al., 2002). Once again, the European H2 pre-cursor event identified by Scourse (2000), Grousset et al. (2001) and Peck et al. (2006) does not appear to have had a significant climatic impact.

H3, the selected palaeoclimate data for which are shown in Fig. 12, has long been seen as a problematic Heinrich event, with evidence of European source material in the eastern Atlantic, but of low concentration, and North American-sourced IRD in the west, but more abundant. Hemming (2004) summarises this evidence and concludes that H3 was of Hudson Strait origin, but of smaller size than other events, so that its IRD did not cover the North Atlantic, as in more characteristic events. One has a very different impression, however, from examining the palaeoclimate record, as there is a very strong and prolonged climate signal associated with this event. In all the temperature records in Fig. 12 there is a gradual decline of up to 5^{o}C in SST (d, e & f in Fig. 12) and 10^{o}C in Greenland air temperature (a in Fig. 12) during the interval 32-31k cal. yr B.P.; the MOC proxy (c in Fig. 12) is also lower during this interval. The end of this period is normally taken as the indicative time for H3 (Table 1), when the North Atlantic IRD signature is at its peak. A number of the records show some recovery around 30-30.5k cal. yr B.P., although in most cases it is centuries-long rather than abrupt. A full recovery of the Greenland air temperature and the SST fields, however, does not occur until c. 29k cal. yr B.P..

H3 therefore poses a problem – why is there so little IRD but such a strong climate anomaly? The nature of the anomaly suggests that despite the strong evidence for a regional IRD event originating from Hudson Strait there must have been some coincident or preceding cause of the major climate change. If this were due to an iceberg release then the slowness of the change suggests a European or Arctic origin. As Hemming (2004) shows, there is little evidence for a strong European source. However, Figures 2 and 7 of Darby et al. (2002) show that the maximum IRD signature during the last 35 000 years in core PS1230 from the Fram Strait occurred prior to 30k cal. yr B.P.. Their analysis was unable to link this peak with any of the source regions that they examined, and, in particular, it did not seem to be linked to Arctic North America. Our hypothesis to reconcile these various facts is that there was a major loss of ice from the Barents Sea ice shelf at this time, followed by a later, and smaller, Hudson Strait event. Lekens et al. (2006) show evidence in contemporary planktonic foraminiferal oxygen isotope anomalies of extensive meltwater across the surface of the whole Nordic Seas during H3, but little enhanced IRD flux at a core in the southern Norwegian Sea. This meltwater could have come from local sources as suggested by Lekens et al. (2006), but could also partially originate from melted Arctic icebergs (e.g., as seen in Fig. 4). Additional evidence for a significant IRD event originating from the eastern Arctic during H3 comes from core GC070 on the Yermak Plateau, NW of Svalbard and to the east of Fram Strait (Howe et al., 2008), where by far the largest IRD event recorded in this core during the main glacial period dates to around this time.

Fig. 13 shows the range of palaeodata around the time of H4. Just prior to 40k cal. yr B.P. there are abrupt changes in a number of indices, particularly eastern Atlantic SST (e in Fig. 13), the ITCZ proxy in Brazil (i), the biogenic silica in Italy (h) and Greenland air temperature (a). At the same time, the sea level (b) begins to rise. These abrupt changes, and the relatively lesser response over Greenland, is consistent with the modelling results for a Hudson Strait iceberg release, which is normally considered the cause of H4 (Hemming, 2004). The MOC proxy (c) also suggests weak return flow at this time, although the onset of this pre-dates the start of change elsewhere. The end of H4 is difficult to determine. The sea level ceases to rise around 39k cal. yr B.P. (b), shortly before Atlantic SST (e) and the MOC (c) return to more normal glacial conditions. However, the Mediterranean SST (f) and Greenland temperature (a) persist in an anomalous state for another 500 years before abrupt rises. This abruptness is consistent with the modelling results for a Hudson Strait recovery, however, the variable end point of the signal points to a more complex event. It is noteworthy that at this same time there is evidence for changes in Antarctica, leading to the major sea level rise (Rohling et al., 2004), a Heinrich-like event and ice sheet collapse in the North Pacific (Bigg et al., 2008) and localised IRD events in the Nordic Seas (Dowdeswell et al., 1999). H4 may be part of a larger perturbation to the climate system.

In contrast to H3 and H4, the climate signal for H5 is hard to discern (Fig. 14). There are substantial amounts of IRD in the North Atlantic (Hemming, 2004) associated with this event, implying a Hudson Strait origin, but the Atlantic climate anomaly is small in the eastern Atlantic (e and b) to missing in the western Atlantic (d). There is a slow decline in the Alboran Sea SST (f) that correlates well with a decline in air temperature over Greenland (a), and the return to normal glacial conditions occurs around the same time in these two parameters, although much more abruptly over Greenland. However, these changes look more likely to be due to some other mechanism; H5 itself appears to have a weak climate impact, although one consistent in terms of rates of change with a Hudson Strait origin.

The final Heinrich event we consider is H6, which occurred around 60k cal. yr B.P. (Table 1). Hemming (2004) shows a number of IRD records associated with this event, although many suggest a much reduced flux (e.g. her Fig. 9). It is again difficult to see a significant climate signal in the proxy records chosen here (Fig. 15, although some (a & f in Fig. 15) do not extend this far back). There is a rise in sea level (b) around 60.7k cal. yr B.P. that may be associated with slight temperature falls in the Atlantic (d) and a reduced MOC (c), but the correspondence is weak.

5. Discussion

Comparison between a number of long term palaeoclimate records and modelling results for idealised releases of freshwater or icebergs into the Atlantic or Arctic Oceans has confirmed the important role of the Laurentide Ice Sheet in affecting glacial climate through ice and freshwater releases into the western Atlantic. Marine core evidence firmly points towards Hudson Strait as the primary origin of these releases (Hemming, 2004). However, the comparison has highlighted the variable climate impact of different Heinrich events, and hence likely differences between events in terms of iceberg release magnitude and/or duration. In addition, we have seen strong suggestions that other release areas have also played a role in producing the climate change observed in a number of events. In particular, past workers' concentration on the IRD record of the North Atlantic has tended to downplay the possibility of significant fluxes from Arctic sources. Few icebergs from the glacial Arctic will have reached the main Atlantic basin to leave a lithic signature even from very large releases, because of the restriction of last glacial Arctic ice export to the narrow Fram Strait, and a long subsequent ocean passage through the Greenland Sea.

The rate of climate recovery from H1, H2 and H3 all suggested an Arctic release contribution to the climate impact. In the case of H1 and H2 we have seen that the IRD evidence from the Arctic and Fram Strait points towards a North American origin for a release coinciding with (or, in the case of H1, slightly later than) the Hudson Strait release. Darby et al. (2002)'s analysis of the FeO grain sizes in IRD suggests that Arctic event AL2 (cf. H1) was largely Laurentide in origin, while AL3 (cf. H2) has a mix of Laurentide, Innuitian and North Greenland peaks. Stokes et al. (2005; 2009) suggest that the only likely source for a major Arctic Laurentide ice stream during this time interval was M'Clure Strait, roughly corresponding to one of our model release sites. In these two Heinrich events, even if there were a European pre-cursor as some authors have suggested, we hypothesise that the major climatic influence was due to a North American ice sheet collapse with both an eastern and northern signature.

H3 has a rather different character, however. While the model-data comparison suggests an Arctic component, the palaeoceanographic evidence suggests any release from the North American ice sheet c. 31k cal. yr B.P. was relatively minor, whether to the east (Hemming, 2004) or the north (Darby et al., 2002). Nevertheless, there was a strong meltwater signal in the Nordic Seas (Lekens et al., 2006), a strong IRD signal in Fram Strait at PS1230 (Darby et al., 2002) and a strong IRD signal off northern Svalbard at GC070 (Howe et al., 2008). The combination of palaeoclimate and model evidence therefore points towards a release from the St. Anna Trough region being the dominant climate-altering cause of H3.

Earlier in the Last Glacial the Hudson Strait origin of Heinrich events H4, H5 and H6 is clearer, with the IRD record in the Atlantic being consistent with the rates of change in the simulations. The Arctic also may have had a more constrained level of glaciation during this time (Svendsen et al., 2004). H6 had quite a small climate signal, consistent with the often weaker, and less widespread, IRD signal in the North Atlantic (Hemming, 2004). H4, however, had an unusual signature, with variable length of the climatic signal depending on location (Fig. 13). With evidence for widespread environmental change around this time elsewhere in the globe (Dowdeswell et al., 1999; Rohling et al., 2004; Bigg et al., 2008) we suggest that this period merits further study to unravel the abrupt climate change occurring at this time.

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Table 1

Approximate ages (calendar years BP) of the start of the abrupt events considered here. Hemming (2004) gives estimates of the uncertainty in H3 of ±1000 years and H6 of ±5000 years. Our estimate comes from the timing of disturbance in the Northern Atlantic and represents the mid-point of the event (±250 years).

Event	Hemming (2004) Age (BP)	Estimated Age (BP) here
Younger Dryas	12 900	12 600
H1	16 800	17 500
H2	24 000	24 000
H3	31 000	31 000[a]
H4	38 000	39 200
H5	45 000	46 100
H6	60 000	60 000

[a]H3 appears to be unusually prolonged from a number of records.

Figure Legends

Fig. 1. High temporal resolution records spanning the last 50 000 years of GISP2 Greenland ice core temperature (bottom panel; data from Alley (2000)), Alboran Sea surface temperature (centre panel; data from Cacho et al. (1999)) and Italian lake sediment biogenic silica (top panel; data from Allen et al. (1999)). Note the fast rates of change and the variations in amplitude and frequency in the records.

Fig. 2. Iceberg and fresh water release sites for a glacial Arctic. A schematic of model annual mean glacial ocean currents is also shown.

Fig. 3. MOC strength (in Sv) for 500 year iceberg releases of 0.1 (dashed), 0.2 (dot-dashed), 0.4 (solid) Sv from the 5 sites, clockwise around the Atlantic and Arctic from bottom to top. The control MOC variation over this time is shown by the dotted line in each segment.

Fig. 4. Sea surface salinity at model Year 5800 (ie 300 years into an iceberg release) for the following experiments: Control (bottom right), and 0.4 Sv iceberg releases for the St. Lawrence (bottom left), Hudson Strait (centre left), MacKenzie (top left), St. Anna Trough (top right) and NCIS (centre right). Absolute values are shown for the Control but anomalies relative to the Control for all other experiments. Contours are every 0.5, with labels at integer values. The data have been transformed onto a 1 degree conventional latitude-longitude grid, so there is a slight discrepancy between the model ocean data (land shown in white) and modern day land boundaries relative to a zero height at the 123m bathymetric contour (shown in black).

Fig. 5. Lower atmospheric temperatures at model Year 5800 (ie 300 years into an iceberg release) for the following experiments: Control (bottom right), and 0.4 Sv iceberg releases for the St. Lawrence (bottom left), Hudson Strait (centre left), MacKenzie (top left), St. Anna Trough (top right) and NCIS (centre right). Absolute values are shown for the Control but anomalies relative to the Control for all other experiments. Contours are every 5^oC for the Control and 0.5^oC for the anomalies, with labels at integer values for the latter. The darker shading shows the more negative contours for the anomaly plots. The modern day land boundaries, relative to a zero height at the 123m bathymetric contour, are shown hatched.

Fig. 6. MOC strength (in Sv) for 500 year freshwater releases of 0.1 (dashed), 0.2 (dot-dashed), 0.4 (solid) Sv from the 5 sites, clockwise around the Atlantic and Arctic from bottom to top. The control MOC variation over this time is shown by the dotted line in each segment. Note that the 0.4 Sv releases from the two European sites led to numerical instabilities, while the 0.2 Sv release from the NCIS, while complete, was also affected by numerical problems.

Fig. 7. Map of sites for palaeoclimate data, mostly labelled as in Fig. 8 (a-b, e-f, and i) or Fig. 9 (*). Note that data sets c and d from Fig. 8 are from the same marine core in the sub-tropical west Atlantic, so this site is labelled “W”, and sets g & h are from the same lake in southern Italy, hence labelled “L”. Dataset * is only used for the Younger Dryas comparison (see Fig. 9).

Fig. 8. Plot of palaeoclimate datasets on a common timescale. From the bottom up there is a) a Greenland ice core temperature (Alley, 2000), b) a sea level record (Siddall et al., 2003), c) an MOC sortable silt index (Hoogakker et al., 2007), d) a western Atlantic SST (Vautravers et al., 2004), e) an eastern Atlantic SST (Vautravers and Shackleton, 2006), f) a western Mediterranean SST (Cacho et al., 1999), g) a terrestrial organic carbon record (Allen et al., 1999), h) a terrestrial biogenica silica record (Allen et al., 1999), and i) a speleotherm $\delta^{18}O$ record (Cruz et al., 2005). See Fig. 7 for a key as to the location of the different datasets.

Fig. 9. Comparison of palaeoclimate records during a time interval focused on the Younger Dryas (c. 12.5ka). The panels are labelled to correspond to the locations shown on Fig. 7. See the longer sets of timeseries, of which this is an excerpt, in Fig. 8, except for the Caribbean SST record (*, Lea et al., 2003).

Fig. 10. Comparison of palaeoclimate records during a time interval focused on H1 (c. 18ka). The panels are labelled to correspond to the locations shown on Fig. 7. See the longer sets of timeseries, of which this is an excerpt, in Fig. 8, except for the Caribbean SST record (*).

Fig. 11. Comparison of palaeoclimate records during a time interval focused on H2 (c. 24ka). The panels are labelled to correspond to the locations shown on Fig. 7. See the longer sets of timeseries, of which this is an excerpt, in Fig. 8.

Fig. 12. Comparison of palaeoclimate records during a time interval focused on H3 (c. 31ka). The panels are labelled to correspond to the locations shown on Fig. 7. See the longer sets of timeseries, of which this is an excerpt, in Fig. 8.

Fig. 13. Comparison of palaeoclimate records during a time interval focused on H4 (c. 40ka). The panels are labelled to correspond to the locations shown on Fig. 7. See the longer sets of timeseries, of which this is an excerpt, in Fig. 8.

Fig. 14. Comparison of palaeoclimate records during a time interval focused on H5 (c. 46ka). The panels are labelled to correspond to the locations shown on Fig. 7. See the longer sets of timeseries, of which this is an excerpt, in Fig. 8.

Fig. 15. Comparison of palaeoclimate records during a time interval focused on H6 (c. 61ka). The panels are labelled to correspond to the locations shown on Fig. 7. See the longer sets of timeseries, of which this is an excerpt, in Fig. 8.

Figure 1
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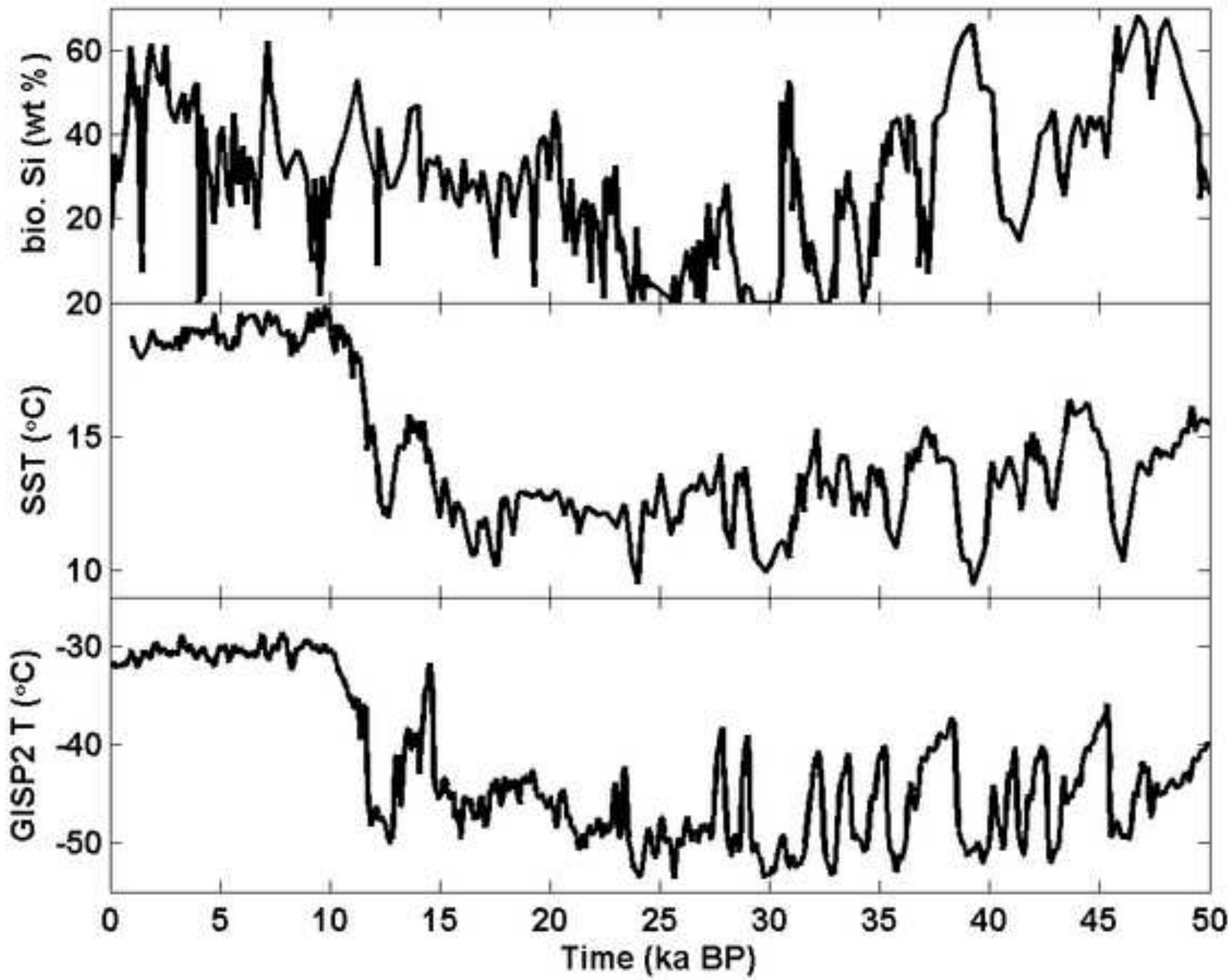

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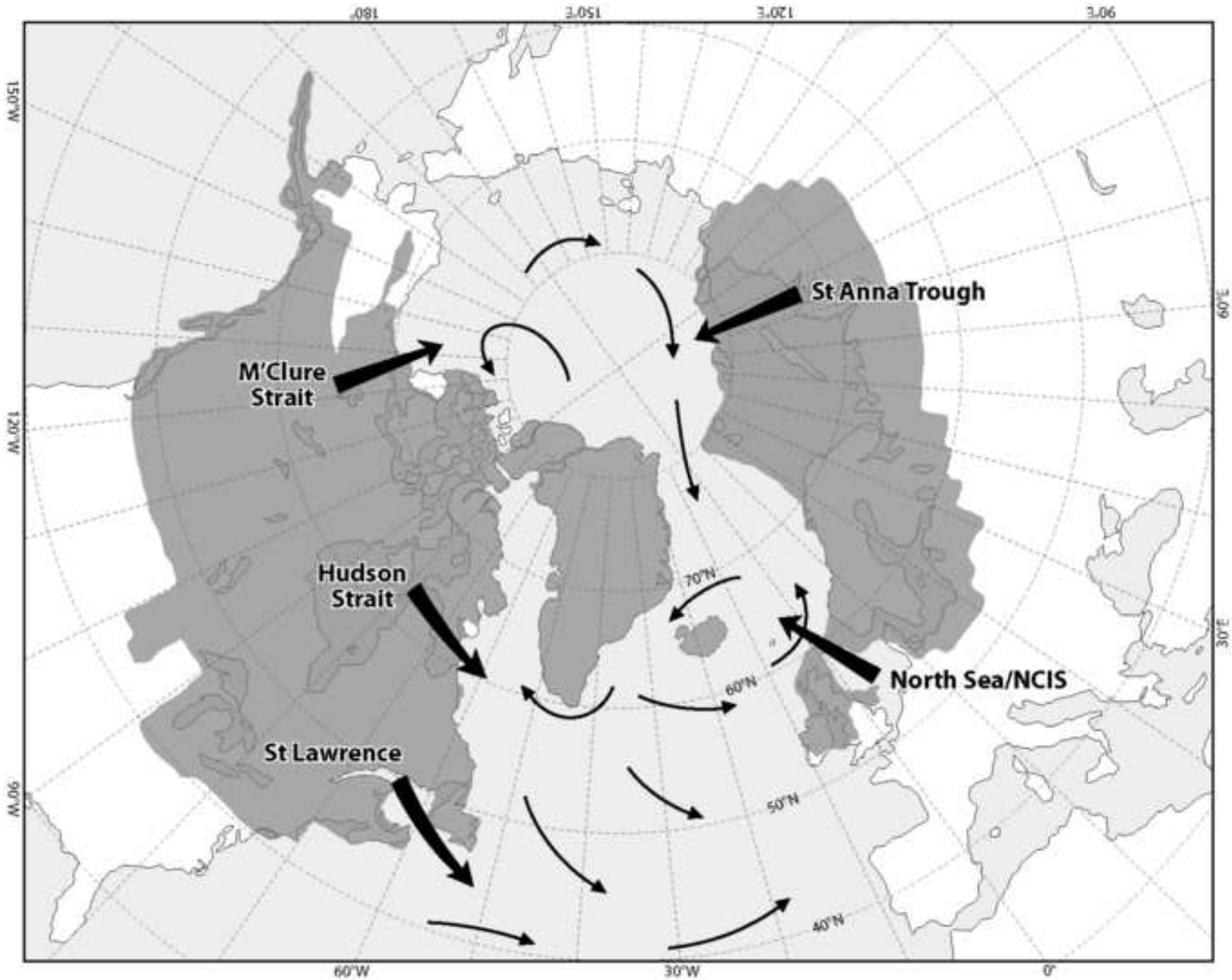

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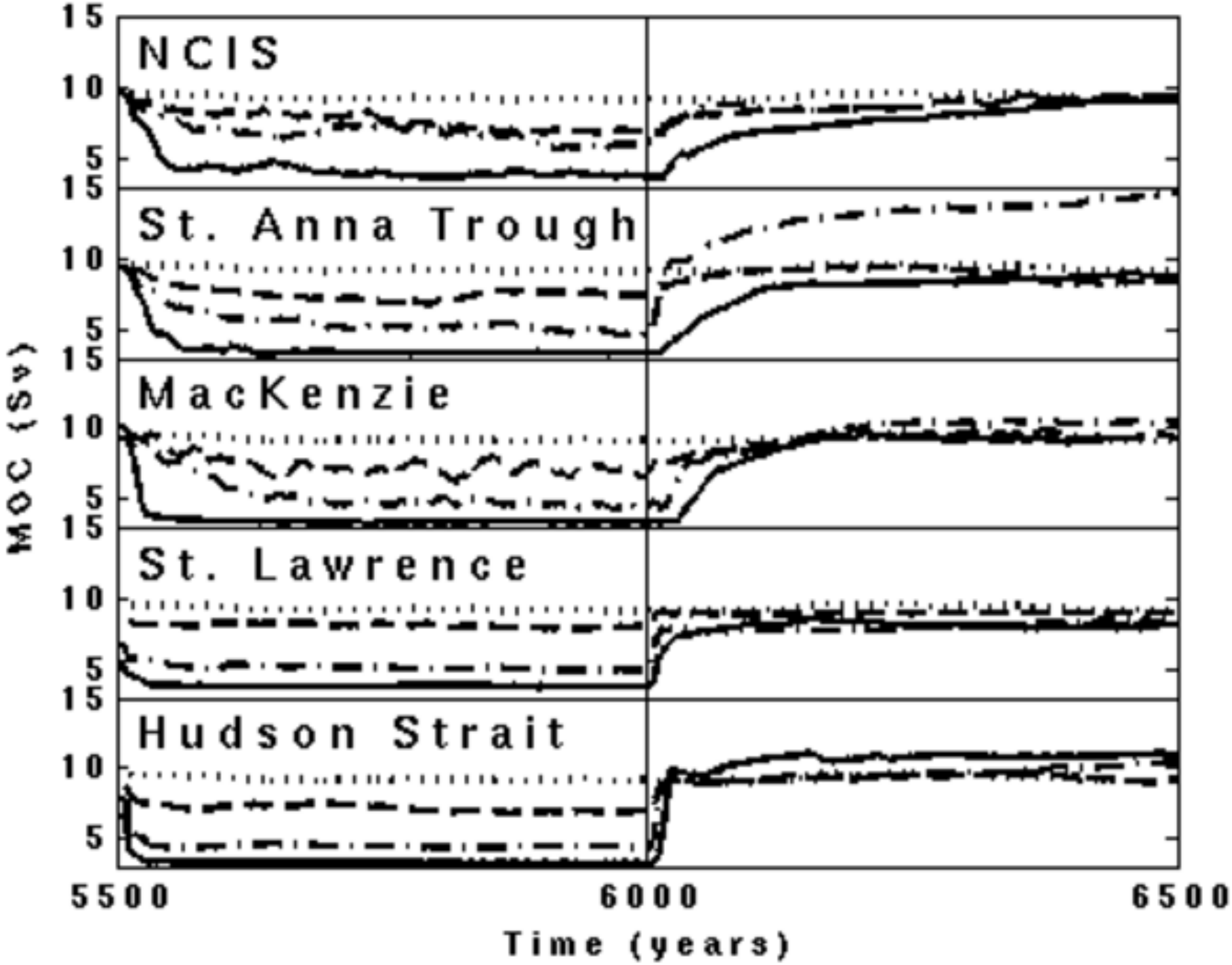

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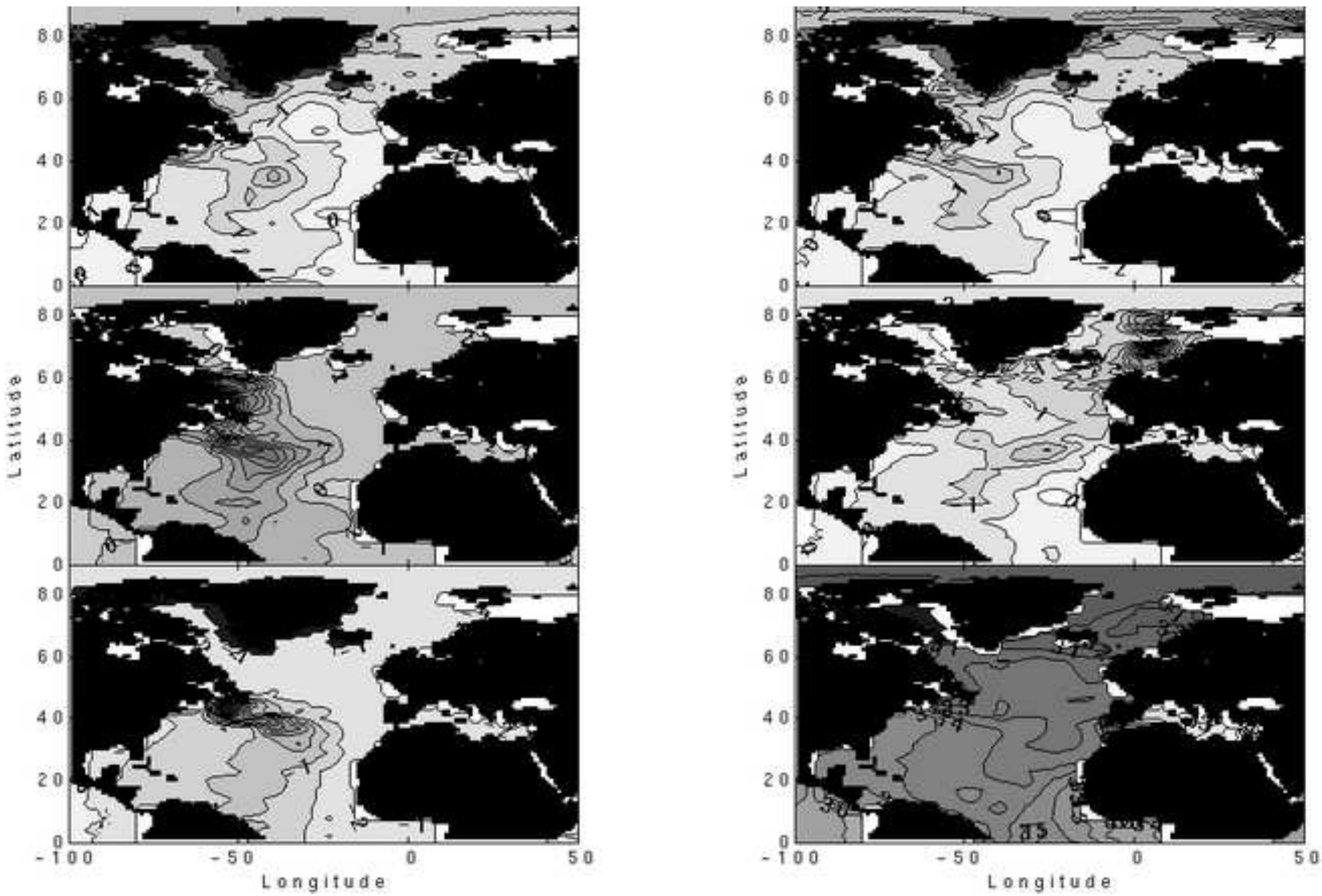

Figure 5

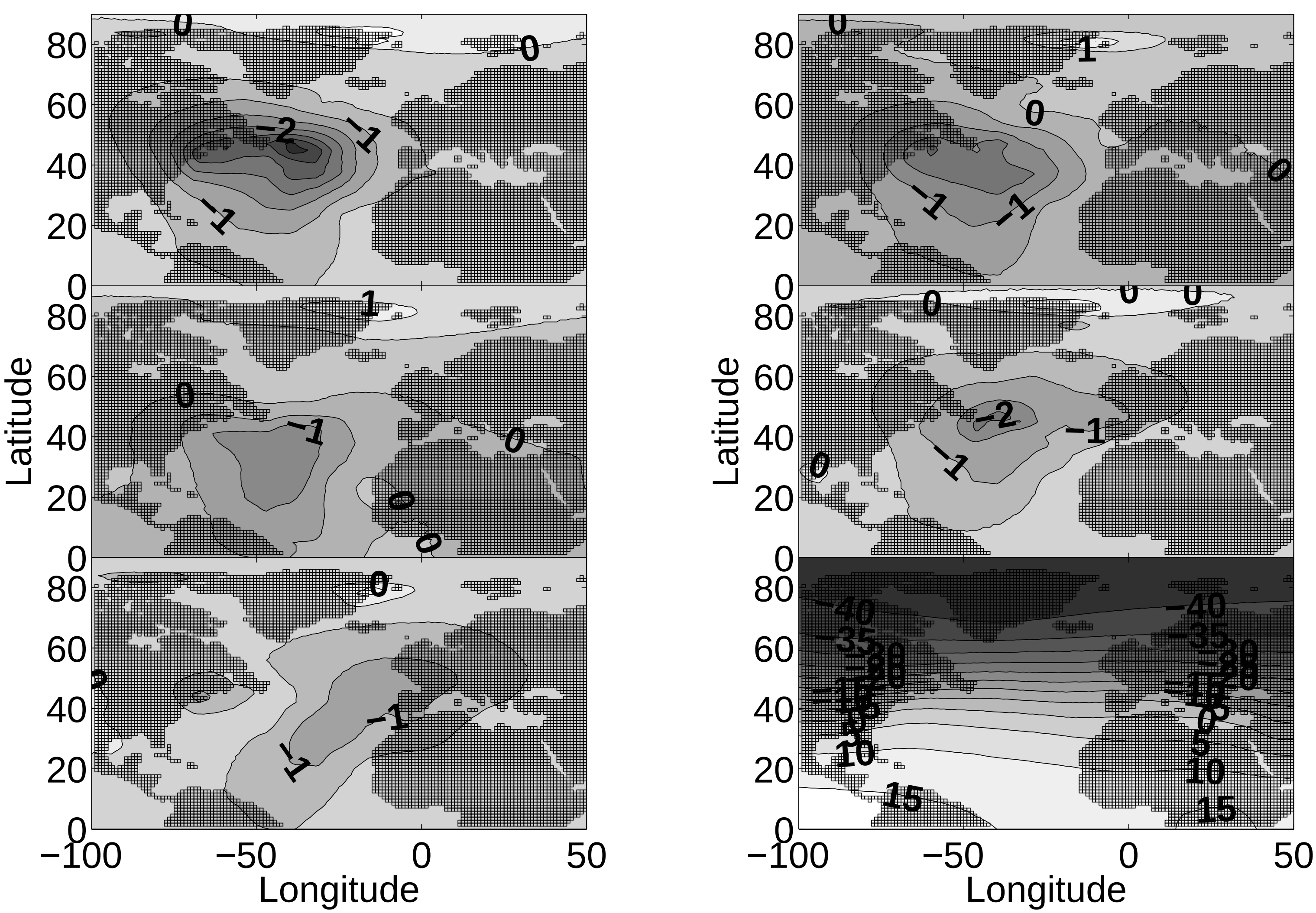
Latitude
Latitude
Longitude
Longitude
0
20
40
60
80
−100
−50
0
50

Figure 6

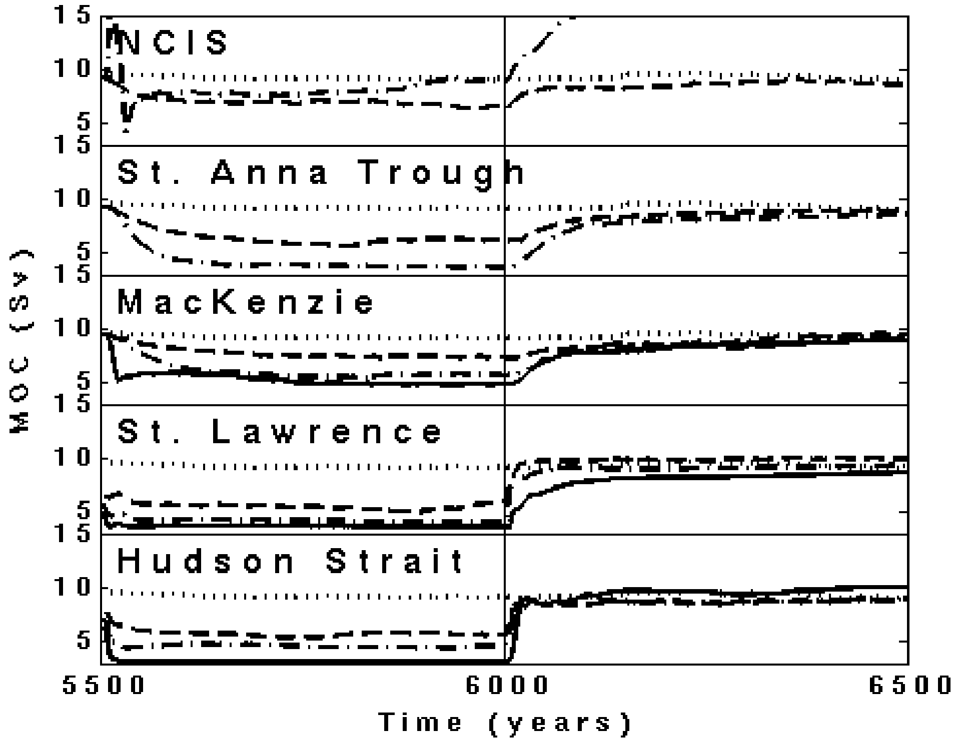
NCIS
St. Anna Trough
MacKenzie
St. Lawrence
Hudson Strait
MOC (Sv)
Time (years)
5500
6000
6500
15
10
5

Figure 7

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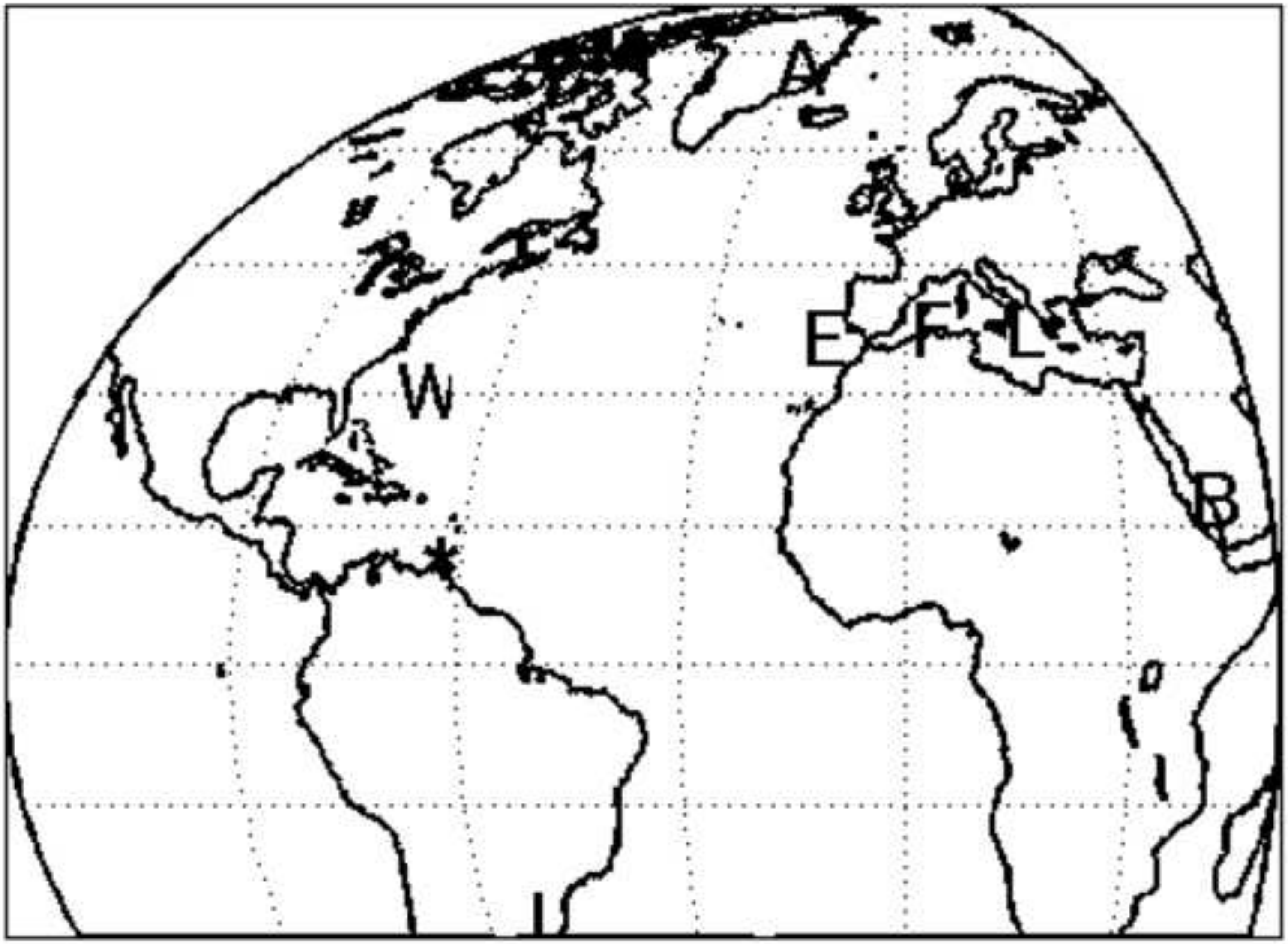

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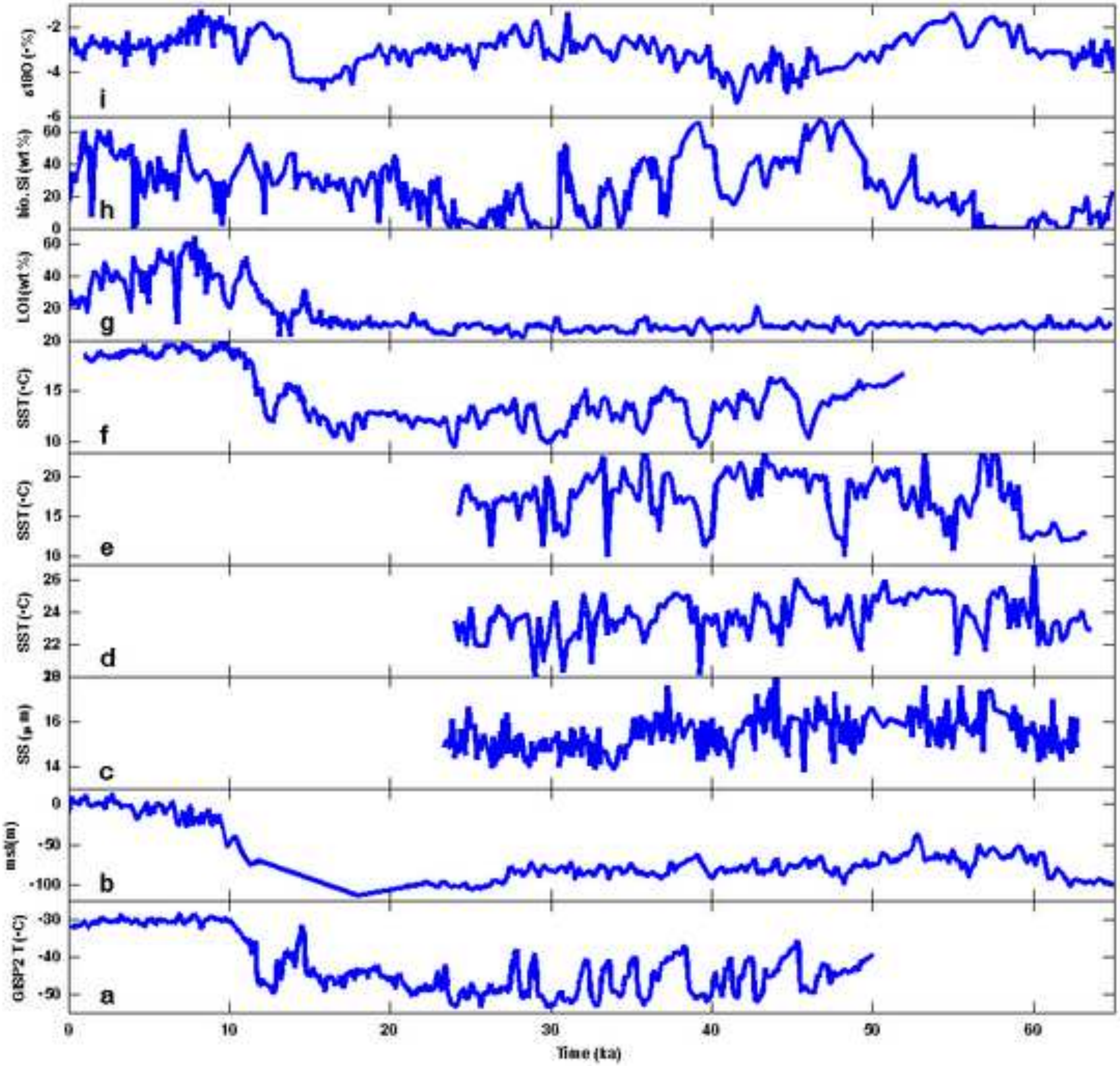

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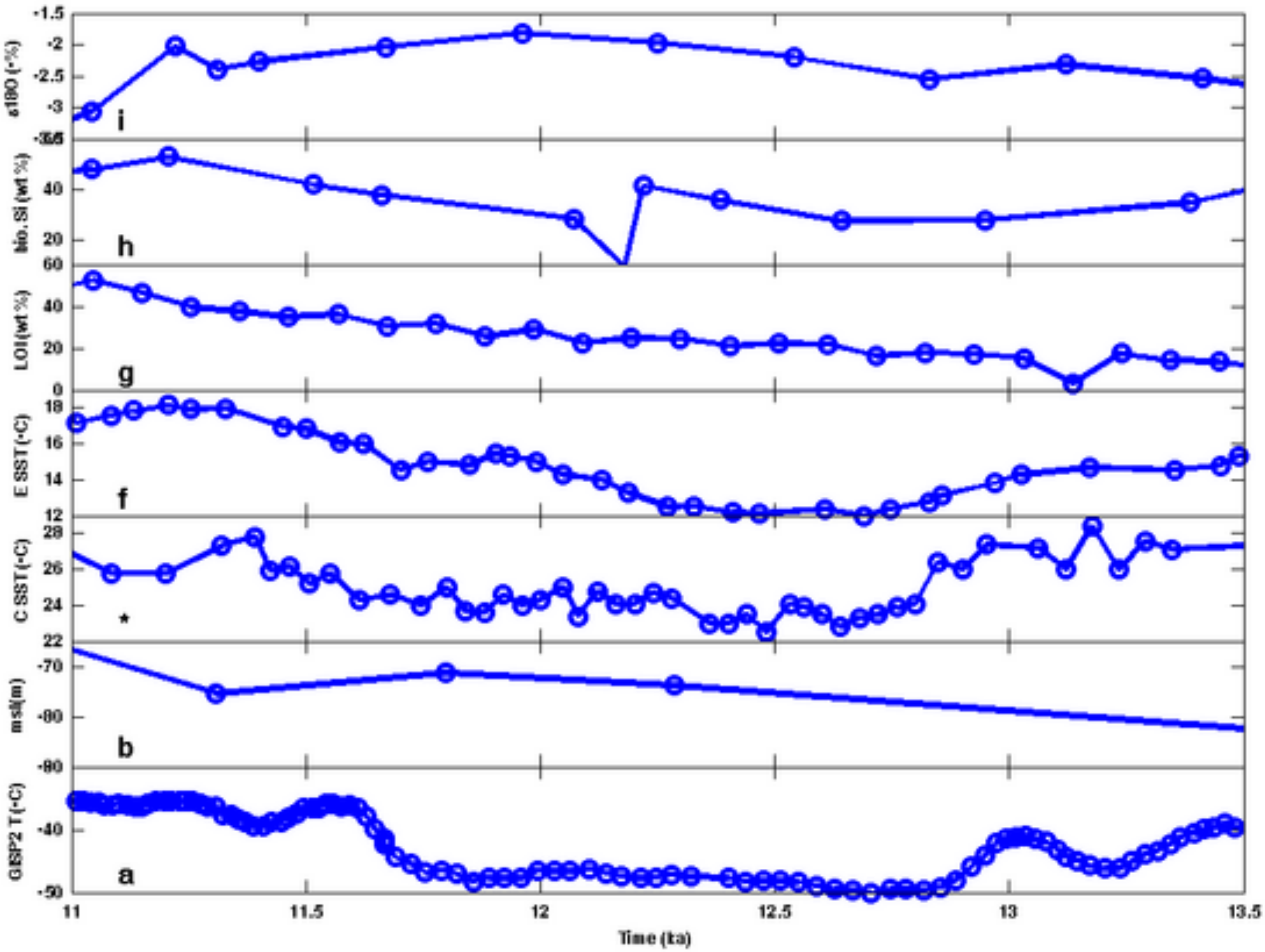

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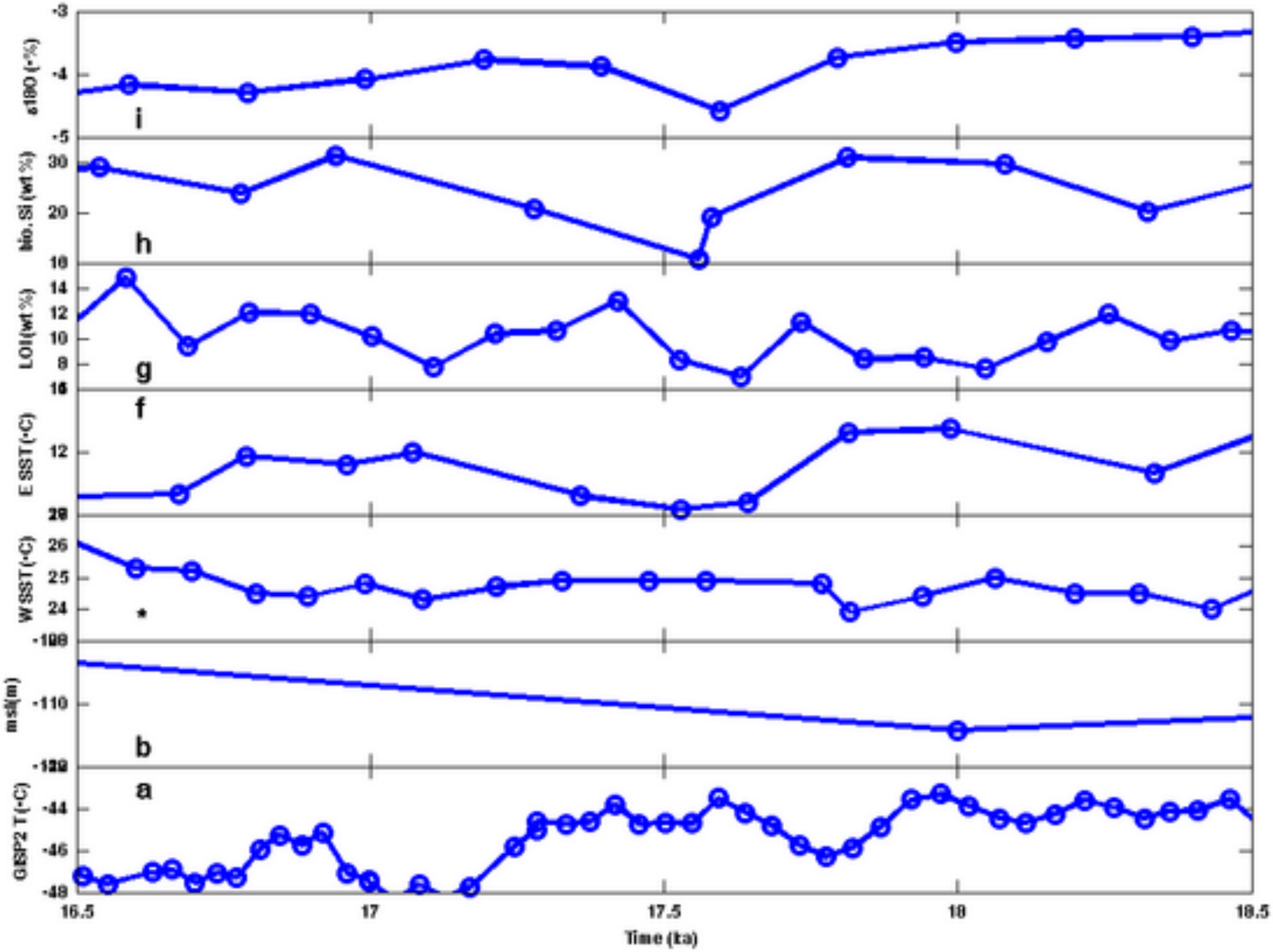

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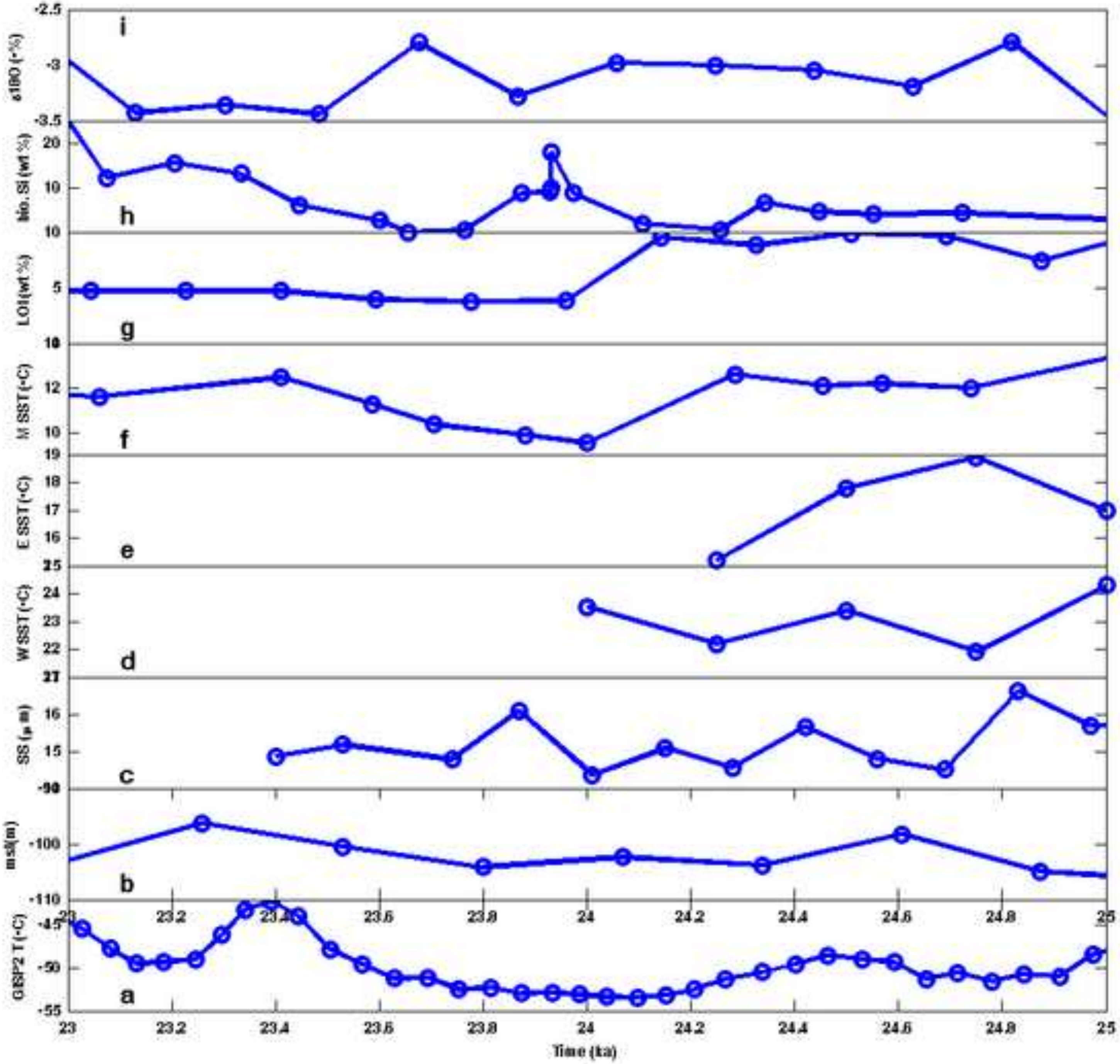

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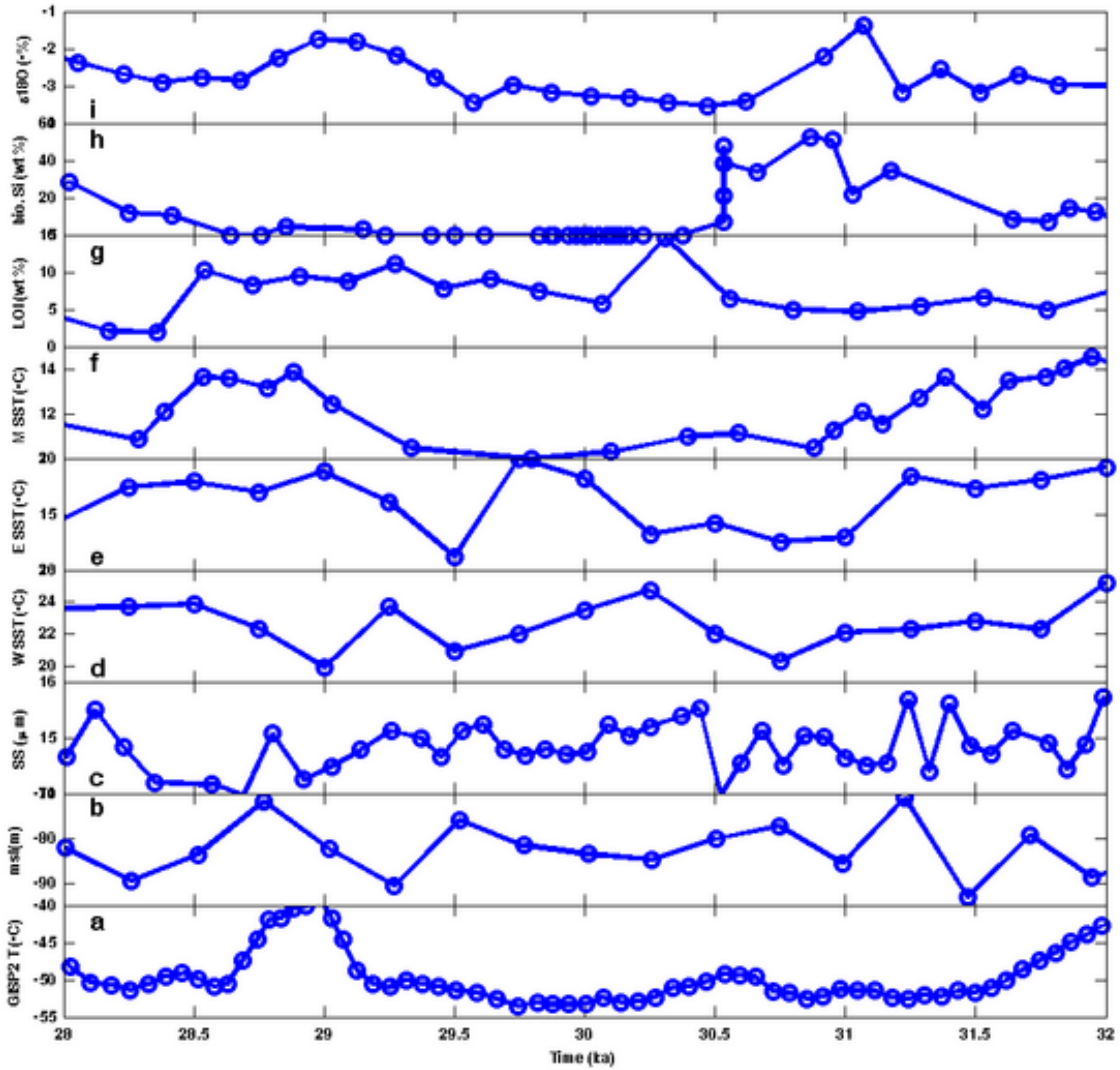

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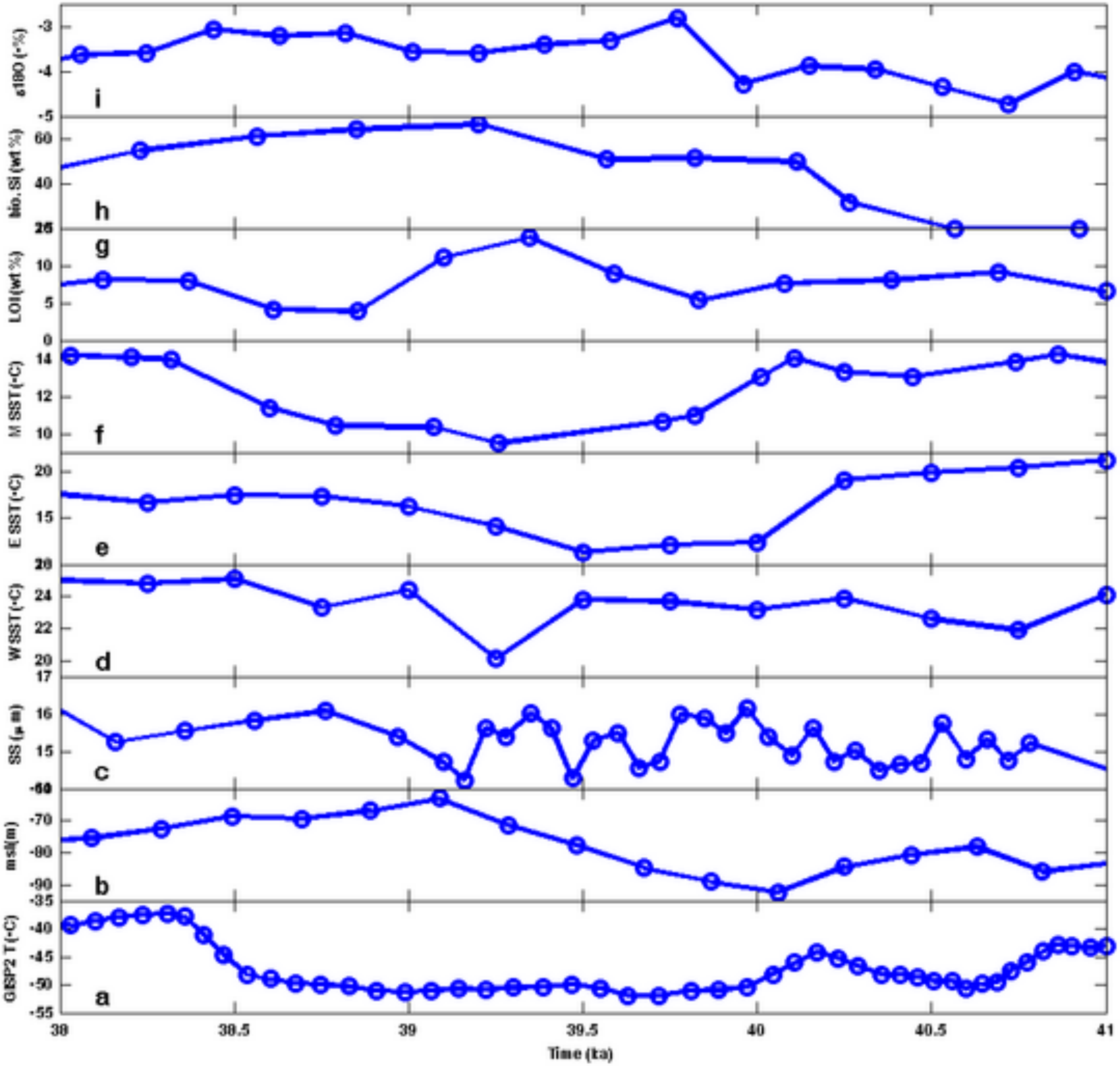

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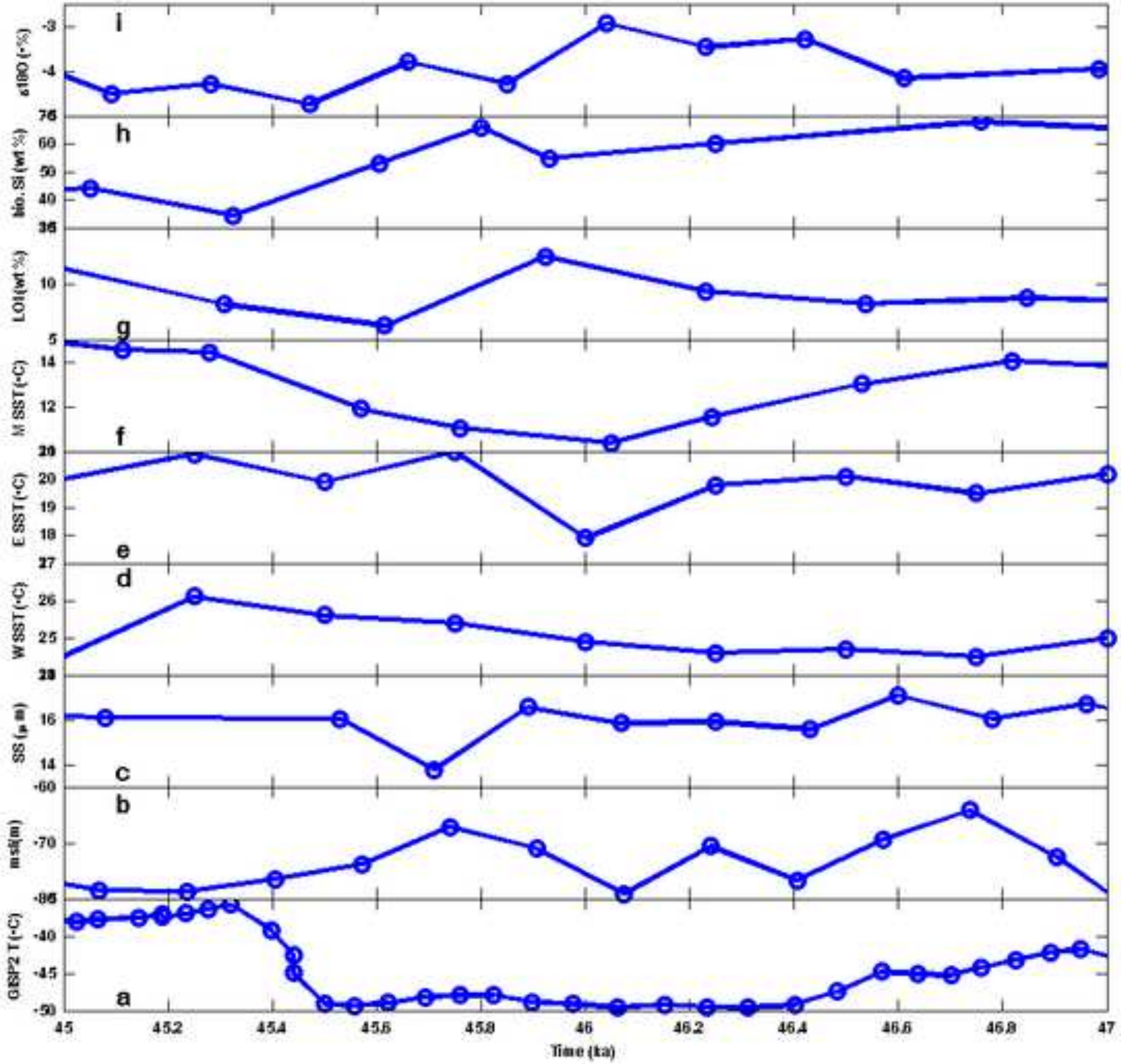

Figure 15

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