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Climate of the Past Discussions

Can oceanic paleothermometers reconstruct the Atlantic Multidecadal Oscillation?

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Abstract

Instrumental records of North Atlantic sea surface temperature reveal a large-scale low frequency mode of variability that has become known as the Atlantic Multidecadal Oscillation (AMO). Proxy and modelling studies have demonstrated the important consequences of the AMO on other components of the climate system both within and outside the Atlantic region. Over longer time scales the past behavior of the AMO is predominantly constrained by terrestrial proxies and only a limited number of records are available from the marine realm itself. Here we use an Earth System-Climate Model of intermediate complexity to simulate AMO-type behavior in the Atlantic with a specific focus placed on the ability of ocean paleothermometers to capture the associated surface and subsurface temperature variability. Given their lower prediction errors and annual resolution, coral-based proxies of sea surface temperature appear to be capable of reconstructing the temperature variations associated with the past AMO with an adequate signal-to-noise ratio. Contrastingly, the relatively high prediction error and low temporal resolution of sediment-based proxies, such as the composition of foraminiferal calcite, limits their ability to produce interpretable records of past temperature anomalies corresponding to AMO activity.

1 Introduction

The Atlantic Multidecadal Oscillation (AMO) is a leading mode of sea surface temperature (SST) variability with a period of 30–80 years (Delworth and Mann, 2000). The SST pattern attributed to the AMO plays an important role in modulating key components of the climate system, for example, the African and Indian summer monsoons, North American and European summer climate, and the activity of hurricanes in the Atlantic (Sutton and Hodson, 2005; Zhang and Delworth, 2006; Shanahan et al., 2009). Therefore understanding the evolution of the AMO and its variability through time is of key importance.

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The AMO was first identified and characterized on the basis of instrumental temperature records spanning the last ∼150 years (Mann and Park, 1994; Schlesinger and Ramankutty, 1994). Subsequently, a focus has been placed on exploring the spectrum of AMO variability more fully via the quantification of past spatial and temporal evolution of temperature anomalies using calibrated proxies (Delworth and Mann, 2000; Gray et al., 2004; Saenger et al., 2009).

Proxy reconstruction efforts have been complemented by a variety of modelling experiments, spanning atmosphere-only models forced to mimic the AMO (Sutton and Hodson, 2005), through to fully coupled models that exhibit spontaneous AMO-type variability (Knight et al., 2005). The specific forcing mechanism of the AMO is still a matter of debate, however, modelling studies provide strong support for an active role of the Atlantic Meridional Overturning Circulation (AMOC), at least partially, driving the development of the SST patterns that characterize the AMO (for a review see Delworth et al., 2007).

The currently available proxy reconstructions of the AMO have mainly relied on dendroclimatological networks, where high resolution (annual) records with precise chronologies are available (Delworth and Mann, 2000; Gray et al., 2004). However, given the hypothesized link between the AMO and AMOC it is desirable to attempt to reconstruct past AMO activity directly from marine archives (Saenger et al., 2009). Corals have the potential to provide absolutely dated and annual (or sub-annual) resolution reconstructions of SST, but their availability is limited and often the duration of the records is not sufficient to isolate the AMO component (Kuhnert et al., 2002; Kilbourne et al., 2008; Saenger et al., 2009). In contrast, temperature reconstructions derived from sediment archives, for example based on the composition of foraminiferal calcite (Lea, 1999), have a lower temporal resolution but are capable of spanning much longer periods than coral records. The limiting factors for sedimentary records are their accumulation rate and processes such as bioturbation, both of which act to smooth the recorded temperature signal. Recent sediment-based studies have achieved resolutions that have allowed the examination of interannual and subdecadal modes (Black et al., 2007; Sicre et al., 2008), however, such records are generally only available in special settings such as marginal basins and not the open ocean.

Here we perform a feasibility study to determine with what fidelity past AMO-type behavior can be reconstructed based on calibrated proxies of seawater temperature. We employ a climate model that is forced in such a manner as to produce AMO-type fluctuations similar to those in the observational record and determine how reliable a temperature proxy would need to be in order to record such fluctuations with an acceptable signal-to-noise ratio. It is shown that many of the currently employed (surface and subsurface) temperature proxies should be capable of capturing AMO variability providing that appropriate sampling locations are chosen and an annual resolution can be achieved. At lower resolutions, for example decadal, the sediment-based proxies carry too large an error to provide reliable reconstructions of the temperature anomalies associated with the AMO.

2 Model and experimental design

The University of Victoria Earth System Climate Model (UVic ESCM version 2.8, Weaver et al., 2001) consists of a two-dimensional atmospheric energy-moisture balance model (Fanning and Weaver, 1996) coupled to a dynamic-thermodynamic sea ice model (Weaver et al., 2001) and a three-dimensional ocean model (MOM2, Pacanowski, 1995). The model is forced with seasonally-varying insolation at the top of the atmosphere and prescribed monthly wind stresses (Kalnay et al., 1996). The horizontal resolution of the model is 1.8° in latitude and 3.6° in longitude, whilst the ocean component is constructed with 19 vertical levels, varying in thickness from 50 m at the surface to 500 m at the base.

Unlike more complex models, the UVic ESCM does not generate AMO-type multidecadal behavior as part of its internal variability (Stouffer et al., 2006). Instead, we adopt a forcing concept similar to that of Zhang and Delworth (2006) and apply a pattern of anomalous heat fluxes to the surface of the Atlantic to mimic the observed

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structure of the AMO over the 20th century. The advantages of such an approach are twofold, first, the spatial pattern, amplitude and frequency content of the AMO can be controlled. Second, any anomalies in the model with respect to a control run can be unequivocally attributed to the AMO-type forcing.

We adopt the following definition of the AMO index: the low-pass filtered (10 year moving average, Enfield et al., 2001) spatial average of detrended mean SST anomalies within the North Atlantic. The annual AMO index was first calculated using the monthly version of the HadISST sea surface temperature component (Rayner et al., 2003). A spatial pattern of the sea surface temperature anomalies associated with the AMO was then obtained by calculating the difference, δSST(K), in detrended SST between 1941–1960 CE (observed positive AMO phase) and 1971–1990 CE (observed negative AMO phase) as a function of location over the Atlantic. The map of δSST(K) was then converted in anomalous heat fluxes, δQ (W/m^2), by multiplication with a coupling coefficient of 30 W/m^2/K (Haney, 1971). Finally, a small constant was subtracted from δQ to ensure that its spatial integral was zero (Zhang and Delworth, 2006).

The anomalous heat fluxes form a basin scale dipole about the equator and thus are suitable to induce northward heat transport in the Atlantic, mimicking changes in the AMOC (Zhang and Delworth, 2006). At a given time, t, when the AMO index is AMO(t), the anomalous heat flux, $Q(t)$, applied to the surface of the Atlantic in the UVic ESCM was defined as $Q(t) = \delta Q \times (\mathrm{AMO}(t)/\delta\mathrm{AMO})$. Here δAMO was a scaling factor selected in order that the amplitude of the AMO index generated in the model was similar to that in the instrumental record. On the basis of a number of trial runs, δAMO was determined to be ~0.18 K.

A 2000-year integration was employed to bring the UVic ESCM into equilibrium with boundary conditions and insolation forcing appropriate for 1850 CE. The anomalous heat flux was then introduced with a repeating pattern of the AMO index following the time series obtained from the data of Rayner et al. (2003) before low-pass filtering. After the run was completed, the temperature values for the final sequence of index values were extracted and anomalies were calculated with respect to a control run. The modelled AMO index was then obtained from the temperature anomalies calculated for the upper layer of the ocean.

3 Results and discussion

Figure 1 shows the results of the modelled AMO index compared to the observed index values obtained from the SST data of Rayner et al. (2003). The unfiltered indices obtained from the observations are more variable than those produced in the model. The same phenomena appeared in the unfiltered index values of Zhang and Delworth (2006), who modelled the AMO using a similar forcing approach. It is assumed that the smoother model values are a product of the redistribution of the heat introduced by the forcing. After filtering (10 year moving average) to isolate the multidecadal variability, the observed and modelled indices display consistent phases and amplitudes.

The influence of the anomalous heat flux on the strength of the AMOC was quantified by performing a complex (time domain) empirical orthogonal function analysis (CEOF, von Storch and Zwiers, 1999) of the Atlantic annual mean meridional overturning streamfunction over time. The advantage of CEOF analysis is that individual components can contain contributions with different phases. Thus it is possible to capture the AMO-type variability throughout the streamfunction even though different locations may not be in phase. The AMOC in the control run has a maximum value of ~22 Sv, which shows very little variability over time (<0.04 Sv), Fig. 2. In contrast, when the AMO-type forcing is introduced, the AMOC displays variation over a large area, with the leading CEOF (explaining ~97% of the total variance) having a maximum amplitude of ~1.5 Sv, Fig. 2. Previous modelling studies have proposed that the AMOC plays a controlling role in modulating the strength of the AMO (Vellinga and Wu, 2004; Knight et al., 2005), whilst in our simulations the introduced anomalous heat flux drives changes in the AMOC. This role reversal does not allow us to comment on the nature of AMO forcing in reality, but does help to demonstrate the important relationship between overturning and heat in the upper ocean.

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As expected the UVic ESCM does redistribute the anomalous heat introduced by Q. Thus the spatial distribution of the SST anomalies resulting from the AMO-type forcing in the UVic ESCM is different to the instrumental record (which was the basis of the δSST forcing pattern). The map of δSST from the instrumental record and an equivalent determination (positive phase minus negative phase) of the AMO δSST pattern in the model run are shown in Fig. 3. In comparison to the instrumental record, the SST pattern in the model run shows an enhanced bipolar structure about the equator. It would appear most likely that this phenomenon is a product of the variations induced in the AMOC by the anomalous heat flux (Broecker, 1998; Stocker, 1998; Stocker and Johnsen, 2003). The key aspect for this study, however, is that although the spatial distributions differ to some extent, the magnitude of the SST anomalies are similar in both the instrumental record and the AMO-type forced UVic ESCM. It is also important to bear in mind that the spatial SST pattern of the AMO obtained from the instrument record is based on a single oscillation and may therefore not be representative of the long-term behavior.

Of specific interest to this study is the amplitude of the temperature anomalies generated in the surface and deeper layers of the ocean as a result of the anomalous heat flux forcing. Such information will allow us to address the question if proxy studies based on, for example, the composition of foraminiferal tests or the growth-rate/composition of corals are able to reconstruct the kind of temperature oscillations expected as part of AMO variability? More specifically, are such proxies sufficiently accurate and of a high enough temporal resolution to reconstruct the temperature variations associated with past AMO variability?

3.1 Statistical framework

It is common for the quality of temperature proxy calibration studies to be assessed using the *standard error of estimation* (for examples see Saenger et al., 2009; Lear et al., 2002; Marchitto et al., 2007; Bamberg et al., 2010). If T is the observed temperature, $\hat{T}$ is the temperature predicted by the estimated calibration relationship and N is the

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number of samples employed in the calibration, the standard error of estimation, s, is given by (Taylor, 1997):

$$s = \sqrt{\frac{\sum_{i=1}^{N}(T_i - \hat{T}_i)^2}{N-2}} \qquad \text{where} \qquad i = 1, \ldots, N. \tag{1}$$

The minimum achievable standard error, S_{E}, on a temperature prediction obtained from a new observation, i.e. applying the derived calibration to a new sample, is given by (Devore, 2008):

$$S_{\mathrm{E}} = s\sqrt{1 + \frac{1}{N}}. \tag{2}$$

A single realization of the possible error on the calibrated value can then be found by multiplying S_{E} by a random number drawn from Student's t distribution with N-2 degrees of freedom (Davies and Goldsmith, 1984).

To assess the ability of a given proxy in a given location to produce a reliable reconstruction of temperature changes resulting from the AMO, we performed the following analysis, which requires some simplifying assumptions to make the results generally applicable. If the calibration of a proxy is based on a sufficiently high number of data points, then $S_{\mathrm{E}} \approx s$ and Student's t distribution approximates a standardized normal distribution. Therefore the temperature uncertainty (effectively noise) in new predictions obtained from the estimated calibration equation can be assumed to be normally distributed with a mean of 0 and standard deviation equal to s. Thus, under an assumption of independent and identically distributed error values it is possible to simulate the noise that can be expected to be included in the estimated temperature values obtained from a calibrated proxy.

For the case of a reconstructed temperature record with a temporal resolution of δt and low-pass filtered with a τ year moving average in the same manner as the AMO index (thus δt must be $\leqslant \tau$ years), the variance of the noise component, σ_n^2, will be:

$$\sigma_n^2 = \frac{\delta t \times s^2}{\tau} \tag{3}$$

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If the variance, σ_T^2, of the noise free low-pass filtered temperature anomalies obtained from the model for a given location is compared to the expected variance of the smoothed noise inherent in a given proxy-calibration, a simple expression for the signal-to-noise ratio, SNR, of the potential proxy reconstruction can be determined:

$$\mathrm{SNR} = \frac{\sigma_T^2}{\sigma_n^2} \tag{4}$$

This means if s is known, the SNR of the low-pass filtered temperature anomaly record can be determined, or alternatively the magnitude of s required to achieve a given SNR can be assessed. We will adopt the latter approach and work under the assumption that the minimum possible value of SNR indicative of an interpretable reconstruction is SNR = 1, i.e. the noise and the true temperature variation make equal contributions to the final reconstruction.

Figure 4 shows the maximum value of the standard error of estimation, $s^{\max}$, which can be tolerated in a proxy calibration in order that the 10 year moving average filtered reconstructed temperature variability resulting from the AMO-type forcing has a SNR$\geqslant$1. Given the assumptions discussed above, this example represents the best case scenario. It is based on the smallest possible error distribution for the predictions (i.e. high N) and assumes an annual resolution. Records with a lower resolution or which exhibit disturbance due to processes such as bioturbation can be expected to have a reduced σ_T^2, whilst s is independent of such factors. Thus, such records would be expected to be characterized by lower SNRs and the values of the tolerable $s^{\max}$ in Fig. 4 would be lower (this will be discussed in more detail below).

A clear pattern emerges in Fig. 4 with surface proxies requiring a low standard error between 0–30° N, whilst higher standard errors are tolerable north of 30° N. This pattern follows the general spatial structure of SST variations associated with the AMO extracted from the instrumental record by Sutton and Hodson (2005) but, as mentioned above, contains an enhanced bipolar structure. Low values of $s^{\max}$ between 0–30° N do not imply that the reconstruction of multidecadal temperature variations possibly attributable to the AMO are unfeasible in the low-latitude North Atlantic, but simply the uncertainty associated with the proxy calibration must be sufficiently low. Saenger et al. (2009) reconstructed multidecadal SST variations for the region around the Bahamas using a coral-based SST proxy. The reconstruction had an annual resolution with an estimated $s \approx 0.3$ K, in comparison, the corresponding s^{max} in Fig. 4 for the study location is ∼0.4 K, indicating that the SNR of the Saenger et al. (2009) record could be >1 providing an adequate number of samples were included in the calibration to fulfill the assumptions given above.

North of 30° N the value of s^{max} increases, particularly within the northward and southward flowing branches of the subtropical gyre. In particular, maximum values of s^{max} are found in the southern portion of the subpolar gyre (SPG), supporting studies which have demonstrated a link between the AMO index and the strength of the SPG (Zhang, 2008). The elevated values of s^{max} in these regions suggest that a number of existing paleothermometers should be capable of recording the temperature variations associated with the AMO in the mid-latitude North Atlantic.

The Mg/Ca ratio of the tests of planktonic foraminifers has been shown to be (at least partially) a function of calcification temperature (Lea, 1999). A wide array of species-specific and species-nonspecific calibrations indicate that s for Mg/Ca based estimates of SST is ∼1 K (Lea et al., 1999; Dekens et al., 2002; Anand et al., 2003). Alternative ocean paleothermometers such as TEX_{86} (thought to reflect the annual mean temperatures of the upper mixed layer) and $U_{37}^{K'}$ (a SST proxy) have quoted s values of 1.7 K (Kim et al., 2008) and 1.2 K (Conte et al., 2006), respectively. The magnitude of these errors suggest that providing sufficiently high-resolution sediment archives exist (a point discussed below), such temperature proxies could be applicable at appropriately chosen locations, but not throughout the entire Atlantic.

Of particular interest in this investigation is the subsurface fingerprint of the AMO and the values of s^{max} that will allow its reconstruction in proxy studies, Figure 5. The temperature signature of the AMO reaches a depth of ∼1400 m but the values of s^{max} are low. Determined standard errors for candidate proxies suggest that subsurface

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temperature variations resulting from the AMO could only be reconstructed with relatively low SNR values. For example, benthic foraminiferal Mg/Ca based temperature estimates have quoted s values of 1.7 K (Lear et al., 2002), 2.4 K (Marchitto et al., 2007) and 0.7 K (Bamberg et al., 2010), suggesting that only the latter case would be capable of producing reconstructions with SNR > 1.

3.2 Limiting factors: sedimentation rate and bioturbation

As discussed above, a number of the considered seawater temperature proxies appear to be sufficiently accurate to record the signature of the AMO providing they can achieve an annual resolution. In the open ocean such resolutions are highly unlikely and it is necessary to consider the effects of low sedimentation rates and processes such as bioturbation, which both act to smooth the recorded temperature signal. A demonstration of this problem can be given using the results presented in Figs. 4 and 5.

The mean annual temperature anomaly records produced by the UVic ESCM were smoothed with a 10 year moving average in order to make them compatible with the definition of the AMO index. These filtered data can also be used to represent a record with a reduced sedimentation rate (and/or increased bioturbation) acting to smooth out the high frequency variation, leaving only the decadal and multidecadal components. Whilst such processes smooth the recorded signal they have no influence on the noise resulting from the uncertainty in the proxy calibration.

Figures 4 and 5 were produced by considered a noise component smoothed according to Eq. 3, where $\delta t = 1$ yr and $\tau = 10$ yrs. For the case of a temperature signal smoothed by the accumulation rate / bioturbation and a non-smoothed noise component, the values of $s^{\max}$ presented in Figs. 4 and 5 need to be divided by a factor of $\sqrt{10}$. For example, the maximum value of $s^{\max}$ in Fig. 4 is ~2.5 K in the subpolar gyre, when considering a non-filtered sediment record with decadal resolution this value of $s^{\max}$ becomes $2.5/\sqrt{10} = 0.79$ K. Similarly, the maximum value of $s^{\max}$ in Fig. 5 is ~1.7 K, which becomes $1.7/\sqrt{10} = 0.54$ K. This is a particularly demonstrative result showing that the majority of the considered sediment-based proxies would be capable of recording the AMO if annual resolution could be achieved, however, with only decadal resolution (which is still optimistic for the open ocean), none of the proxies would yield a SNR$\geqslant$1.

4 Conclusions

Our model and subsequent statistical analysis reveal that, under a best-case scenario, existing ocean paleothermometers may be capable of recovering interpretable records of the temperature variations in the North Atlantic associated with the past activity of the AMO. Whilst such a scenario appears to be appropriate for coral-based reconstructions, sediment records have a lower resolution and may have been disrupted by processes such as bioturbation. For the case where sediment records have a decadal resolution, the quoted accuracy of proxies such as Mg/Ca, TEX_{86} and $U_{37}^{K'}$ suggests that they would yield reconstructions of the temperature anomalies resulting from the AMO with SNRs less than 1.

A simple strategy when the characteristics of a proxy record are sub-optimal would be to filter the reconstructed temperatures at longer periods, thus decreasing the noise contribution. Such an approach will of course reduce comparability with the AMO characteristics obtained from the observational record, but may represent the only opportunity to recover some of the useful information buried within a relatively high noise component. It is, however, important to remember that the form of the final signal will be controlled by the specific filter design and thus must be taken into consideration when interpreting and comparing signals.

Finally, it would appear that climate models have an invaluable role to play in estimating the potential SNR with which different proxies could be expected to reconstruct certain modes of variability. Such information is essential if the quality of proxy records is to be assessed objectively, but it should always be kept in mind that the results will be model dependent.

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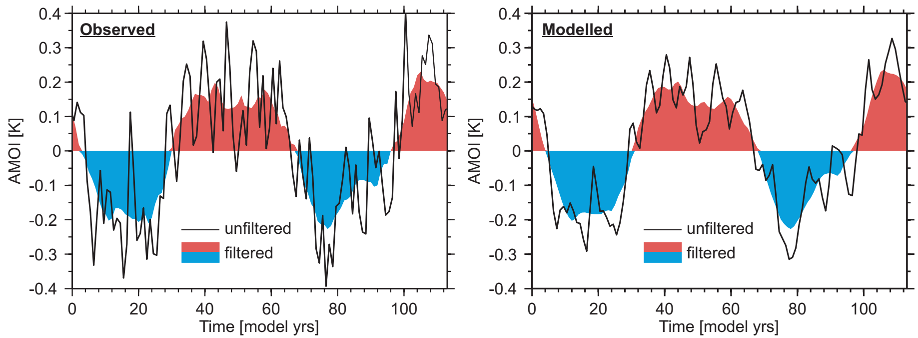

Fig. 1. Observed and modelled AMO index, pre- (lines) and post- (shaded) low-pass filtering (10 year moving average). Positive and negative phases of the AMO are shown by red and blue shading, respectively.

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Reconstructing the Atlantic Multidecadal Oscillation

D. Heslop and A. Paul

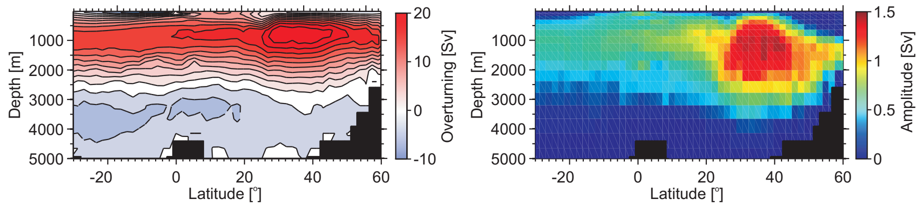

Fig. 2. (Left) Annual mean meridional overturning streamfunction for the Atlantic obtained from the control run with no AMO-like forcing. (Right) Amplitude of the leading complex empirical orthogonal function, showing the amplitude of the variability induced in the AMOC by the AMO-like forcing.

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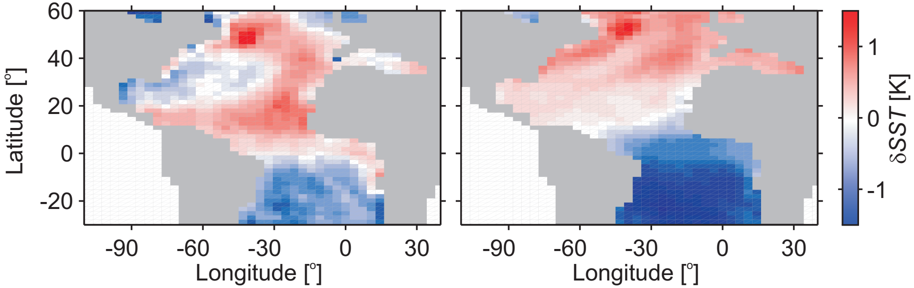

Fig. 3. (Left) Anomalous SST pattern associated with the AMO component of the instrumental record. The temperature anomalies during a negative phase of the AMO were subtracted from those of a positive phase to yield the difference map, δSST. (Right) Anomalous SST pattern obtained from the difference of positive and negative AMO phases in the UVic ESCM.

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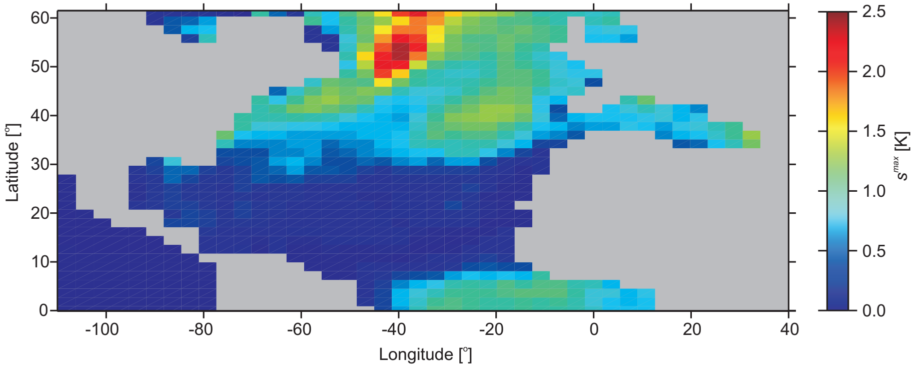

Fig. 4. The maximum allowable standard error of estimation, s^{max}, in a calibrated SST proxy if the temperature variability attributable to the AMO is to be reconstructed with a SNR$\geqslant$1. It is assumed that the temporal resolution of the proxy data is 1 year and the estimated values of s^{max} represent the best case scenario. If the temporal resolution was decreased then s^{max} would also decrease.

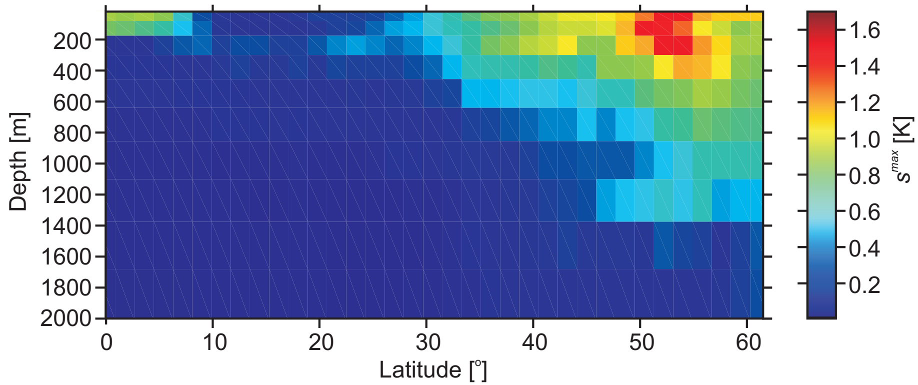

Fig. 5. Meridionally averaged s^{max} in a calibrated temperature proxy if the temperature variability attributable to the AMO is to be reconstructed for annually resolved data with a SNR⩾1.

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