

Decadal prediction skill using a high-resolution climate model

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1 Decadal prediction skill using a high-resolution 2 climate model.

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6

7 Abstract

8

9 The ability of a high-resolution coupled atmosphere-ocean general circulation model (with a
10 horizontal resolution of a quarter of a degree in the ocean and of about 0.5° in the atmosphere) to
11 predict the annual means of temperature, precipitation, sea-ice volume and extent is assessed based on
12 initialized hindcasts over the 1993-2009 period. Significant skill in predicting sea surface temperatures
13 is obtained, especially over the North Atlantic, the tropical Atlantic and the Indian Ocean. The Sea Ice
14 Extent and volume are also reasonably predicted in winter (March) and summer (September). The
15 model skill is mainly due to the external forcing associated with well-mixed greenhouse gases. A
16 decrease in the global warming rate associated with a negative phase of the Pacific Decadal Oscillation
17 is simulated by the model over a suite of ten-year periods when initialized from starting dates between
18 1999 and 2003. The model ability to predict regional change is investigated by focusing on the mid-
19 90's Atlantic Ocean subpolar gyre warming. The model simulates the North Atlantic warming
20 associated with a meridional heat transport increase, a strengthening of the North Atlantic current and
21 a deepening of the mixed layer over the Labrador Sea. The atmosphere plays a role in the warming
22 through a modulation of the North Atlantic Oscillation: a negative sea level pressure anomaly, located
23 south of the subpolar gyre is associated with a wind speed decrease over the subpolar gyre. This leads
24 to a reduced oceanic heat-loss and favors a northward displacement of anomalously warm and salty
25 subtropical water that both concur to the subpolar gyre warming. We finally conclude that the subpolar

26 gyre warming is mainly triggered by ocean dynamics with a possible contribution of atmospheric
27 circulation favoring its persistence.

28

29

1. Introduction

Climate prediction at interannual-to-decadal timescales can help decision makers to calibrate plans and actions related to climatic events due to climate variability and change (Hibbard et al., 2007; Cox and Stephenson, 2007). Model error and forcing uncertainties are however a source of near-term global prediction biases, as underlined by the global temperature fluctuation of the 2000's, i.e. the so-called "hiatus", that has not been predicted by the CMIP5 climate models (Kosaka and Xie., 2013; Watanabe et al., 2013, Meehl et al., 2014; among others). Decadal prediction experiments are an opportunity to analyze the physical mechanisms associated with climate predictability and variability in a near-future (2010-2035). They can thereby allow model physics and projection improvements. Meehl et al. (2011, 2014), Watanabe et al., (2013), Guemas et al. (2013b) and England et al. (2014) have thus used CMIP5 models to evaluate this 2000's temperature hiatus in climate models. This is a relevant question and is also one goal of this study. In this study we explore the ability of a newly developed high resolution climate model to predict the decadal evolution of temperature, sea-ice extent and volume. In addition, we assess conditional skill by focusing on two specific events, the 2000's temperature hiatus and the 1990's warming of the subpolar gyre.

Motivation for initialized decadal predictions comes from various studies that have successfully predicted regional climate on decadal time scales (Smith et al., 2007; Keenlyside et al., 2008; Pohlmann et al., 2009; Mochizuki et al., 2012; Chikamoto et al., 2013; Bellucci et al., 2013, 2014; Doblus-Reyes et al., 2013a; Garcia-Serrano et al., 2015; Karspeck et al., 2015). The strongest skill is found in retrospectively predicting surface air temperature over large areas. This is primarily determined by the response to well-mixed greenhouse gases (GHGs), especially for lead-time greater than 2 years (Oldenborgh et al., 2012). Additional skill is given by the major volcanic events that induce an abrupt cooling through the ejected volcanic aerosols (Guemas et al., 2013a; Mehta et al., 2013; Timmreck et al., 2016). Another source of predictability is the initialization of the ocean that can lead to additional skill for the first years of the hindcasts, improving the near-term predictability

57 (Yeager et al., 2012, Robson et al., 2012b; Matei et al., 2012; Chikamoto et al., 2013; Doblas-Reyes et
58 al., 2013a). This is especially relevant for the North Atlantic Ocean where an accurate initialization of
59 the Atlantic meridional overturning circulation (AMOC) allows Atlantic multidecadal variability to be
60 predicted a decade or more in advance (Griffies and Bryan, 1997; Boer, 2000; Collins et al., 2006;
61 Pohlmann et al., 2013). Model studies have then shown that it is possible to produce an accurate
62 simulation of the mid 90's subpolar gyre (SPG) warming thanks to a correct initialization of the ocean
63 mean state (Robson et al., 2012ab; Yeager et al., 2012; Msadek et al., 2014a). The SPG warming is
64 therefore an excellent case study to test the ability of a prediction system, due to its large magnitude
65 and potential predictability, and is thereby one of the main goal of this study.

66

67 The impact of the land surface initialization has also been assessed and is associated with a more
68 limited but significant improvement of the precipitation and temperature predictions (Koster and
69 Suarez, 2003; Paolino et al., 2012). Over the Arctic the sea ice thickness initialization plays a role in
70 the seasonal forecast skill of sea ice extent, has shown in Day et al. (2014) and suggested in Msadek et
71 al., (2014b). Important efforts are thus currently being undertaken to improve sea-ice initialization
72 (Lindsay et al., 2012). The role of other drivers has been summarized in the review of Bellucci et al.
73 (2015).

74

75 It has been suggested that models with larger systematic errors tend to produce lower prediction
76 performance (DelSole and Shukla, 2010). The improvement of the decadal prediction systems by
77 increasing spatial resolution and improving physical parametrization is one of the main challenges
78 proposed in Doblas-Reyes et al. (2013b). There is indeed some evidence that increasing model
79 resolution can improve some aspects of the mean climate. It has been shown that high resolution
80 ocean-atmosphere coupled models produce smaller errors than low resolution models in sea surface
81 temperature (SST) over coastal upwelling regions (Shaffrey et al., 2009; Gent et al., 2010; McClean et
82 al., 2011; Delworth et al., 2012; Sakamoto et al., 2012; Small et al., 2014), due to a better simulation
83 of the wind and its effect on the ocean (Gent et al., 2010; Sakamoto et al., 2012). It reduces the double
84 ITCZ in the tropics (Delworth et al., 2012), El-Niño Southern Oscillation (ENSO) (Shaffrey et al.,

85 2009; Sakamoto et al., 2012; Small et al., 2014) and the north Atlantic SSTs (Gent et al., 2010) are
86 better simulated. Gent et al. (2010) with an ocean-atmosphere coupled model and Jung et al (2012)
87 with a atmosphere model have shown a more accurate atmospheric circulation at higher resolution: in
88 Gent et al. (2010) the maximum sea-surface temperature biases in the major upwelling regions are
89 reduced by more than 60%, the precipitation patterns are improved in the summer Asian monsoon, and
90 the atmospheric circulation in the Arctic is also improved. In Jung et al. (2012) the tropical
91 precipitation and atmospheric circulation, the occurrence of Euro-Atlantic blocking and the
92 representation of extratropical cyclones are better represented in increasing horizontal resolution. Over
93 land, the improvement of the precipitation is partly due to a better resolved orography (Gent et al.,
94 2010; Delworth et al., 2012; Sakamoto et al., 2012) as shown with regional downscaling.

95

96 With the high resolution Community Atmosphere Model (CAM) Bacmeister et al. (2014) have found
97 however that their simulations are not dramatically better than low resolution simulations.
98 Furthermore some problems remain or are exacerbated at these resolutions: some simulations exhibit
99 too much warming and cooling, especially over polar regions where sea ice are not enough accurately
100 represented (McClean et al., 2011; Kirtman et al., 2012), and Bacmeister et al. (2014) have found an
101 exacerbated Pacific double ITCZ bias. The improvement due to a higher resolution is therefore model-
102 dependent. Increasing the resolution does not necessary lead to more accurate simulations.

103

104 Increasing the resolution can also lead to improved seasonal predictions: with the high-resolution
105 GFDL climate model Jia et al. (2015) have shown an improvement of the seasonal prediction of 2-m
106 air temperature and precipitation over land and of the Nino-3.4 index.

107 The question of the skill improvement with a high-resolution climate model remains a wide-opened
108 question. We address it here by using a set of decadal hindcasts performed with the CERFACS-HR
109 high-resolution model. Unfortunately, we cannot directly assess the potential improvements due to
110 resolution only as the model physics version is slightly different from our low resolution version that
111 was used for CMIP5 (Sanchez-Gomez et al., 2015). Furthermore, the initialization dates are different

112 making impossible a fair comparison of the two prediction systems. We thereby simply assess the skill
113 of our new modeling system and qualitatively compare it to the range obtained with low-resolution
114 models. Note also that we do not have historical simulations (uninitialized hindcasts) that would allow
115 us to clearly assess the additional influence of ocean initialization in addition to that of the external
116 forcing.

117

118 The aim of this paper is thus to address three main questions:

119 - What is the skill of our high-resolution coupled climate model in predicting global and regional
120 surface quantities (temperature and sea-ice)?

121 - Is the model able to represent the 2000's global mean temperature fluctuation when initialized
122 with observed ocean initial conditions?

123 - Is the model able to predict the mid-90's warming of the Atlantic subpolar gyre? What are the
124 dominant mechanisms?

125 The paper is structured as follows. The model, data and methodology are described in section 2.
126 Section 3 focuses on the model ability to predict the annual means of SST, air-surface temperature,
127 sea-ice volume and extent. We analyze the simulated 2000's global mean temperature (GMT) changes
128 in section 4. The case-study of the mid-90's north Atlantic warming is investigated in section 5. A
129 summary and discussion are given in section 6.

130

131 **2. Data and Methodology**

132 **2.1 The CERFACS-HR model**

133 CERFACS-HR is a high resolution Atmosphere-Ocean General Circulation Model (AOGCM)
134 developed at Centre Européen de Recherche Avancée en Calcul Scientifique (CERFACS). The
135 atmospheric model is the ARPEGE-climate v5.3 with a horizontal resolution of about 50 kilometers
136 and 31 vertical levels (T359L31), developed at Météo-France/CNRM (Centre National de Recherches
137 Météorologiques). This version of ARPEGE shares the same physic as ARPEGE v2, used in the
138 CMIP5 exercice (CNRM-CM5.1, Voldoire et al., 2013), but runs at a higher resolution (truncature
139 T359 instead of T127) and does not use the SURFEX (SURFface EXternalisé) modeling system. The
140 land-surface Interaction between Soil Biosphere and Atmosphere (ISBA) model (Noilhan and Planton
141 1989; Noilhan and Mahfouf 1996) is used in order to represent continental surfaces (soil, vegetation,
142 snow).

143 The ocean model is the Nucleus for European Models of the Ocean (NEMO) v3.4 (Madec, 2008) from
144 the Institut Pierre-Simon Laplace (IPSL). NEMO is discretized on a ORCA025L75 grid (horizontal
145 resolution of about 0.25° and on 75 vertical levels) at global scale decreasing poleward (an isotropic
146 Mercator grid in the Southern Hemisphere, a quasi-isotropic bipolar grid in the Northern Hemisphere
147 with poles over land at 107°W and 73°E). Vertical grid spacing is finer near the surface and increases
148 with depth. Further information on the DRAKKAR ORCA-R025 grid is given in Barnier et al. (2006).
149 The Louvain-la-Neuve Sea Ice model LIM2 3.3 (Vancoppenolle et al., 2009a, 2009b) is embedded
150 into the ocean modeling system NEMO, on a C-grid dynamic-thermodynamic model and includes the
151 representation of the subgrid-scale distributions of ice thickness, enthalpy, salinity and age. The
152 atmospheric and oceanic components are coupled with OASIS3-MCT2.0 (Valcke et al., 2013).

153 **2.2 Initialization and external forcing**

154 The hindcasts are full-field initialized every year from 1993 to 2009 (initialized November 1st every
155 year). Ocean and Sea-ice initial conditions are given by the GLORYS2V1 ocean reanalysis product

156 (Ferry et al., 2012). This reanalysis is based on an ocean and sea-ice general circulation model at 1/4°
157 horizontal resolution assimilating in situ profiles of temperature, salinity and sea surface temperature.

158 The atmosphere is initialized from a SST-forced ARPEGE simulation over the same period. The
159 different members are obtained by lagged perturbation of the atmospheric initial conditions (initialized
160 from a different day of the same month). Finally, ten ensembles (1993-2002) of five members
161 integrated over a ten-year period and seven additional ensembles (2003-2009) of five members
162 integrated over a five-year period are available. Each member shares the same radiative forcing and
163 differs due to a different atmospheric initialization and to the model internal variability.

164 The model is constrained by the observed external forcings such as past and current greenhouse gases
165 concentration, solar irradiance and various types of aerosols (black carbon, particulate organic matter,
166 dust, sea salt and sulfate). The optical depths of the tropospheric aerosols are similar as that of the
167 CMIP5 prescribed emissions (Szopa et al., 2013).

168 **2.3 Model evaluation**

169 The magnitude of the bias and drift of CERFACS-HR is assessed by comparing the raw hindcast data
170 with the ocean reanalysis. CERFACS-HR exhibits positive biases of SST over the Austral Ocean (up
171 to 6°C) and over the eastern equatorial Pacific and Atlantic Oceans (up to 4°C) (Fig. 1a). These strong
172 biases, in particular over the equatorial and subtropical Oceans, are commonly obtained with climate
173 model simulations. The bias in monthly GMST increases during the first two years, reaching 1.2°C,
174 and decreases then after 10 years (Fig. 1b).

175 The Arctic experiences a cold bias of 5°C, at the east of Greenland (Fig. 1a). The sea ice extent is
176 therefore anomalously large ($+0.5 \cdot 10^6 \text{ km}^2$) during the first months of the simulations. The bias
177 increases with time and can reach up to $2.5 \cdot 10^6 \text{ km}^2$ after 9 years (Fig. 1c).

178 The AMOC spatial pattern is reasonably simulated (Fig. 1d, e) despite a negative bias at the location
179 of the core of North Atlantic deep water (the AMOC is weaker in CERFACS-HR than in GLORYS2V3

180 north of 30°N). Note that the ocean reanalysis AMOC data may also suffer from large biases and does
181 not necessarily reflect the observed AMOC.

182

183 **2.4 Observed data sets**

184 The new Mercator product (GLORYS2V3; Lellouche et al., 2013) covers a longer period (1993-2012)
185 than GLORYS2V1 (1993-2009) and allows us to take into account all the hindcasts integrated over a
186 ten-year period. A comparison between the GLORYS2V1 and GLORYS2V3 products shows a similar
187 mean state of the SST (not shown). Furthermore these products were produced with the same ocean
188 NEMO global configuration, i.e. the ORCA025 grid (Lellouche et al., 2013), than CERFACS-HR.

189 The model ability to predict SSTs is investigated by comparing with the ERSST3b analysis. This
190 product covers the 1854-present period on a monthly frequency and on a 2° x 2° horizontal resolution.
191 The ERSST3b analysis includes the Extended Reconstruction SST version 2 (ERSST) improvements
192 due to analysis methods tuning and to the inclusion of bias-adjusted satellite data (Smith et al., 2008).

193 The observed sea level pressure (SLP) and surface fluxes are given by the ERA-interim reanalysis
194 (Simmons et al., 2007; Uppala et al., 2008; Dee et al., 2011) over a global domain and available from
195 1979 to present. ERA-interim corrects some of the ERA-40 errors: an improved representation of the
196 hydrological cycle, a more realistic stratospheric circulation and a better temporal consistency of the
197 reanalysis fields (Dee et al., 2011).

198 For the surface-air temperature, three sets of observations are used. The Hadley Centre/Climate
199 Research Unit Temperature version 4 (HadCRUT4) provides near-surface air temperature from 1850
200 to present on a 5° x 5° horizontal resolution (Morice et al., 2012). HadCRUT4 covers about 84% of
201 the globe (with a weaker concentration of observations at the poles and over Africa). The coverage is
202 varying in time and missing values are associated to grid points receiving a weak coverage of
203 observations. The incomplete global coverage is a potential source of bias in global temperature

204 reconstructions. HadCRUT4 has been improved in Cowtan and Way (2014) with a better infilling of
205 poorly sampled regions. This additional dataset is hereafter named COWT.

206 The NASA's Goddard Institute for Space Studies Surface Temperature Analysis (GISTEMP) consists
207 of an update of the Goddard Institute for Space Studies (GISS) analysis of surface temperature. The
208 GISTEMP record attempts to address the coverage issue by extrapolating temperatures into
209 unmeasured regions by means of kernel smoothing using a canonical kernel of radius 250 and 1200
210 km (Hansen et al., 2010). We used the data smoothed at 250 km. GISTEMP provides temperature
211 anomalies (the base period is 1960-1990) over land and ocean from 1880 to present on a $2^\circ \times 2^\circ$
212 horizontal resolution.

213 We also used the NOAA Merged Land-Ocean Surface Temperature (MLOST) version 3b of the
214 National Climatic Data Center (NCDC). MLOST is a merged land air and sea surface temperature
215 anomaly analysis, based on data from the Global Historical Climatology Network (GHCN) of land
216 temperatures and the International Comprehensive Ocean-Atmosphere Data Set (ICOADS) of Sea
217 Surface Temperature (SST) data. The temperature anomalies, with respect to 1961-1990 are analyzed
218 separately and then merged to form the global analysis. More details are given in Smith et al. (2008)
219 and Vose et al. (2012). MLOST is a spatially gridded ($5^\circ \times 5^\circ$) global surface temperature dataset, with
220 monthly resolution from January 1880 to present.

221 The Sea-Ice Extent (SIE) is given by the National Snow Ice Data Center (NSIDC) (Fetterer et al.,
222 2009). As for the NSIDC sea ice index, the CERFACS-HR SIE is computed as the surface of the grid
223 cells covered by ice when the ice concentration is above 0.15. The Sea-Ice Volume (SIV) time series is
224 reconstructed from the Pan-Arctic-Ice-Ocean-Modeling-System (PIOMAS) (Zhang et al., 2008;
225 Schweiger et al., 2011).

226 CERFACS-HR results were interpolated on different grids so to compare them with different as the
227 reanalysis, i.e. a 1.5° resolution for SLP and surface fluxes, a 2.5° resolution for precipitation and a 5°
228 resolution for surface air temperature. SSTs have been interpolated at the ERSST3b 2° resolution. The
229 oceanic variables used in the section 4 (temperature, salt, zonal and meridional velocities) are

230 produced on the same ORCA025 grid as GLORYS2V1 and GLORYS2V3, and have therefore not
231 been interpolated.

232 In section 4 the observed heat content is computed with the Hadley Center reanalysis (EN4 data; Good
233 et al., 2013) that has used all the available data sets including Argo floats of recent years. It is a global
234 quality controlled ocean temperature data set, available from 1900 to present at a monthly frequency at
235 1° resolution with 42 vertical levels.

236 **2.5 Methodology**

237 The hindcasts cover a short period starting every year over a seventeen-year period (1993-2009).
238 Hindcasts starting in 1993-2002 (2003-2009) were run for ten (five) years. Such a short period can be
239 strongly impacted by decadal variability and the removal of a parametric trend associated with the
240 increase in well-mixed GHGs may lead to spurious results. Removing the trend (linear or quadratic)
241 could therefore lead to remove part of the decadal signal. We thus chose to not detrend the temporal
242 data fields.

243 **2.5.1 Bias adjustment**

244 Hindcasts start from the initial conditions and drift away as time progresses. This drift is due to model
245 errors and reflects the model adjustment from the initial observed state back to its equilibrium state.
246 We used a standard procedure following the World Climate Research Program recommendations
247 (ICPO report on mean bias adjustment, 2011) to remove the drift, a posteriori and in a linear way.

248 Let $Y_j^i(\tau)$ the hindcasts and $X_j(\tau)$ the corresponding verifying observations for each member i of the m
249 simulations; where $j=1$ to n indicates the initial year of the prediction (start-date), and τ is the lead
250 time. For each lead-time τ we can thus define the mean model predictions $\bar{Y}(\tau)$ and observations $\bar{X}(\tau)$
251 as

$$252 \quad \bar{Y}(\tau) = \frac{1}{nm} \sum_{j=1}^n \sum_{i=1}^m Y_j^i(\tau)$$

253
$$\bar{X}(\tau) = \frac{1}{n} \sum_{j=1}^n X_j(\tau)$$

254 The drift $dr(\tau)$ is thus estimated for each lead-time as the difference between the mean model
 255 prediction $\bar{Y}(\tau)$ and the observation $\bar{X}(\tau)$, as

256
$$dr(\tau) = \bar{Y}(\tau) - \bar{X}(\tau)$$

257 The corrected model prediction, for each member i , initial year j and lead-time τ is then defined as

258
$$\hat{Y}_j^i(\tau) = Y_j^i(\tau) - dr(\tau)$$

259 This method assumes that the drift is independent of the initial conditions. The drift is however not
 260 stationary and the results may be sensitive to this methodology (in the case where there is strong non-
 261 stationary drifts it could be preferable to use the method developed in Kruschke et al., 2015). Here we
 262 assume that the ICPO method is a reliable bias adjustment method.

263 In section 3 we choose to remove the drift of SST based on the observed datasets used to assess skill
 264 (ERSST3b). The drift in air-surface temperature is removed with ERA-interim since HadCRUT4 is
 265 given in anomaly. We have test the sensitivity to the selected reanalysis used to remove the drift: we
 266 have found weak differences among the used reanalysis (not shown). In section 4 we analyzed the
 267 subpolar gyre warming and used the GLORYS2V3 reanalysis (that are close to the GLORYS2V1
 268 reanalysis used to initialize the ocean model while spanning a longer time period) to remove the drift
 269 on the original ORCA grid of the model. We used ERA-interim for heat fluxes, surface pressure and
 270 surface winds.

271 **2.5.2 Evaluation of hindcast skill**

272 The skill is computed with an Anomaly Correlation Coefficient (ACC) r , given by

273

$$r(\tau) = \frac{\sum_{j=1}^n [X_j(\tau)Y_j(\tau)]}{\sqrt{\sum_{j=1}^n [X_j(\tau)]^2 \sum_{j=1}^n [Y_j(\tau)]^2}}$$

274 Where Y_j is the ensemble mean anomaly for the j^{th} hindcast starting in year j ; X_j is the observation
275 anomaly for the corresponding starting date j . τ is the lead-time and n the number of starting date.
276 Anomalies for X_j and Y_j are calculated independently, both having a zero mean over the hindcast
277 period.

278 We estimate the significance of the ACC through re-sampling and 1000 permutations in a Monte-
279 Carlo framework. The values are judged significant at the 5% level if the correlation values are
280 stronger than 97.5% of the randomly obtained correlation values. We also use the root mean squared
281 error (RMSE). RMSE measures the magnitude of the error between the hindcasts and the observations
282 while ACC measures only the phase difference between observations and hindcasts. RMSE and ACC
283 thus provide complementary information on the prediction skill of CERFACS-HR.

284 The 4-year persistence is computed for the SST, SIE and SIV based on the observed values of SST,
285 SIE and SIV in the 4 years prior to the start date. The skill of this empirical model to predict the
286 climate depends on the initial condition given the climate system memory.

287

288 **3. Model skill**

289 **3.1 Sea Surface Temperature**

290 The ACC is computed for the first year of the hindcasts (lead-time 1) and averaged over two different
291 lead-times: 2-5 and 6-9 years (based on the hindcasts for which the time-period is long enough). The
292 skill of the first year is strongly dependent on the initial conditions and represents the seasonal-to-
293 interannual predictions; the 2-5 years lead-time represents the interannual timescale (Goddard et al.,
294 2013). The comparison of the 2-5 and 6-9 year means allows analyzing the dependence of the skill on
295 lead time. The 4-year average reduces higher frequency noise. Another reason of the choice of the
296 lead-times 2-5 and 6-9 is that they are frequently used (Bellucci et al., 2013, 2014; Goddard et al.,
297 2013; Karspeck et al., 2015, among others) and results are thus comparable with other modeling
298 studies.

299 The skill in predicting annual mean sea-surface temperature (SSTs) is given by the ACC between the
300 hindcasts and the ERSST3b analysis. The first year exhibits significant correlations over the Atlantic
301 Ocean (mostly over the tropical Atlantic and the Caribbean), the Indian Ocean and the Pacific Ocean,
302 west of 150°W (Fig. 2a). The 2-5 year lead-time still exhibits skill over tropical and North Atlantic
303 (but with a lesser extent), Indian and West of the Pacific Ocean (Fig. 2b). The 6-9 year lead-time
304 exhibits negative correlations over the Pacific Ocean and the skill over the Caribbean vanishes,
305 underlying the loss of skill in some areas (Fig. 2c).

306 It has been shown in Sanchez-Gomez et al. (2015) that the CNRM-CM5 model drift is due to quasi-
307 systematic excitation of an ENSO event at the first year of the hindcasts. Since the drift is not totally
308 removed by the ICPO adjustment method a similar mechanism could occur in CERFACS-HR (which
309 also used the ARPEGE and NEMO model but at higher resolution) and can partly explain the weak
310 skill obtained in the Pacific Ocean.

311 For the 6-9 year lead-time, the most prominent result is the skill obtained over the tropical Atlantic
312 that is unexpected since biases, due to the poor representation of the mean intensity and dynamic of

313 the cold tongue and seasonal cycle of the SSTs (Okumura and Xie, 2004; Richter and Xie., 2008)
314 generally lead to low skill in these regions (Stockdale et al., 2006). Fig. 2c shows annual mean
315 correlation but significant correlations remain in March-April-May over the tropical Atlantic (not
316 shown), the period of the cold-tongue set-up and the season where the bias is at its maximum
317 (Zermeño-Díaz and Zhang, 2013).

318 The RMSE is strong but no significant over the equatorial Pacific and the subtropical north Atlantic,
319 for the first year (Fig. 3a). The 2-5 and 6-9 year lead-time areas with large and significant RMSE value
320 correspond to the areas of large and negative ACC value, i.e, the Pacific Ocean for 2-5 and 6-9 year
321 lead-time and the Austral Ocean and subtropical North Atlantic for the 6-9 year lead-time (Fig. 3b, c).
322 The loss of skill is thus due to inabilities of CERFACS-HR to reproduce both the magnitude and phase
323 of the SST.

324 Different boxes are defined in order to show the ability of the model to predict regional indexes: the
325 Atlantic multidecadal oscillation (AMO) index ($[0^{\circ}\text{N}-60^{\circ}\text{N}; 80^{\circ}\text{W}-0^{\circ}\text{W}]$), the nino3.4 index ($[5^{\circ}\text{S}-$
326 $5^{\circ}\text{N}; 170^{\circ}\text{W}-120^{\circ}\text{W}]$), the tropical Atlantic ($[3^{\circ}\text{S}-3^{\circ}\text{N}; 20^{\circ}\text{W}-10^{\circ}\text{E}]$) and Indian SSTs ($[30^{\circ}\text{S}-10^{\circ}\text{S};$
327 $60^{\circ}\text{E}-100^{\circ}\text{E}]$)(see Fig. 2a). The global mean SST domain is defined as the average of all the observed
328 grid points containing non-missing values.

329 Fig. 4a shows the correlation score for these five indexes and for 4-year averaged periods. Gray
330 shading represents the non-significant correlations (according to the Monte-Carlo test) and the 4-year
331 persistence is represented within the purple line.

332 When we consider the global domain, results are significant for the first lead-time, decrease as time
333 progresses, and become non-significant for the lead-time 3-6 year (Fig. 4a). It is almost zero at a lead-
334 time 6-9 year. The 4-year mean persistence is not beaten by CERFACS-HR leading to the conclusion
335 that the model ability to predict the mean global SSTs is mostly due to the external forcing. The
336 decrease of the skill is partly due to the negative correlation over the Pacific Ocean from the lead-time
337 2-5 to 6-9 year (Fig. 4e). This is consistent with the Fig. 2c which exhibits negative correlations over
338 the Pacific Ocean.

339 Correlations are significant for the AMO, Tropical Atlantic and Indian Ocean. The skill of the AMO is
340 positive and significant for the lead-times 1-4 year to 4-7 years and becomes non-significant for the
341 lead-times 5-8 and 6-9 year. A positive and significant skill for the AMO was obtained by numerous
342 authors (Kim et al., 2012; Chikamoto et al., 2013; Bellucci et al., 2014; Garcia-Serrano et al. 2015).
343 This high predictability was attributed to multidecadal variability of the ocean dynamics (Pohlmann et
344 al., 2004) and to the initialization of the ocean (Garcia-Serrano et al., 2015).

345 Over the tropical Atlantic, correlations using CERFACS-HR are significant and stronger than
346 persistence for lead-times up to 3-6 years. The skill in predicting tropical Atlantic SSTs is associated
347 with variables linked to the tropical Atlantic SSTs variability: the westerly winds (Richter et al., 2014a,
348 2014b), precipitation over the adjacent areas (Chang et al., 2008; Richter and Xie, 2008; Zermelo-
349 Diaz and Zhang, 2013) and deep convection (Richter et al., 2014b). Since no skill is found on the
350 pressure field (involving the subtropical gyre), the heat content (involving the memory of the ocean)
351 and the meridional wind and precipitations (Supplementary material S1) we hypothesize that this
352 result could be due to the sampling (we only used ten start-dates). The question of the relationship
353 between these variables and the tropical Atlantic SSTs in CERFACS-HR is however not under the
354 scope of this study.

355 The skill over the Indian Ocean is positive and marginally significant for several lead times (1-4 years,
356 2-5 years and 6-9 years). Guemas et al. (2013a) used a set of decadal predictions and found that the
357 Indian Ocean stands out as the region where the predictions of SST perform the best worldwide. The
358 models ability to predict the SSTs is mainly due to the long-term warming trend (as in Guemas et al.,
359 2013a).

360 The CERFACS-HR model is thus able to predict the decadal trends of several indexes of SST (AMO,
361 Tropical Atlantic and Indian) and with at least similar skill as that shown in studies based on low-
362 resolution models (Bellucci et al., 2013; Chikamoto et al., 2013; Guemas et al., 2013a).

363 **3.2 Sea ice extent and volume**

364 The observed shrinking of summer Arctic sea ice extent (SIE) (Serreze et al., 2007; Stroeve et al.,
365 2012) has increased accessibility to marine waters and has raised the possibility of an ice-free Arctic in
366 the near-future (Wang and Overland, 2009). This rapid decline also contributes to the polar
367 temperature amplification (Screen and Simmonds, 2010) and is thereby an indicator of climate change.
368 Accurate prediction of SIE and the sea ice volume (SIV) can thus lead to a better understanding of the
369 incoming climate change. There are plenty of prediction studies at a monthly to seasonal timescale
370 (Sigmond et al., 2013; Day et al., 2014; Msadek et al., 2014b; among others) about sea-ice but only a
371 few studies at decadal time-scales with for instance Blanchard-Wrigglesworth et al., (2011) and Germe
372 et al., (2014).

373 The skill in simulating the SIE and SIV 10-year trends is investigated with the Fig. 5 for ten start dates
374 (from the 1994-2003 to 2003-2012 trend). The observed SIE and SIV trends are included in the
375 CERFACS-HR spread for a majority of start dates, highlighting the importance of both the external
376 forcing and the initial conditions in predicting these quantities. The mean simulated and observed
377 trends are negative (except observed SIE in winter), indicating a continuous shrinking of SIE and SIV.
378 CERFACS-HR well simulates the sign of the long-mean trend with a better result in summer (Fig. 5bd)
379 than in winter (Fig. 5ac) (the SIE and SIV shrinking is not apparent in the observations in winter and is
380 over-estimated with CERFACS-HR). The internal variability of SIE and SIV is strong (the ensemble-
381 spread of summer SIE is of 2 million km² for the 2002-2011).

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4. The global mean temperature fluctuation of the 2000's

386

387 Despite a continuous increase in atmospheric GHGs, the GMT has shown a quasi-stabilization since
388 1998. This recent temperature evolution has been named "the hiatus" and has been associated with
389 several possible factors: a negative phase of the Pacific decadal oscillation (PDO) with anomalously
390 cold SSTs over the eastern Pacific ocean and more heat uptake (Meehl et al., 2011; Kosaka and Xie,
391 2013; Watanabe et al., 2013; Trenberth and Fasullo, 2013; Guemas et al., 2013b; England et al., 2014;
392 Meehl et al., 2014; Douville et al., 2015), the recent stratospheric volcanic aerosol trend (Fyfe et al.,
393 2013; Santer et al. 2014; 2015; Haywood et al., 2014; Ridley et al., 2014; Schmidt et al., 2014; Brühl
394 et al., 2015; Mills et al., 2016), the decrease in tropospheric water vapor concentration (Solomon et al.,
395 2010) and the solar minimum around 2009 (Kopp and Lean., 2011).

396 We analyzed the ability of CERFACS-HR to simulate the so-called hiatus over the recent decades.
397 Figure 6 represents the decadal linear trends for the hindcasts initialized from November 1997 to
398 November 2002. The hindcast surface air temperatures are interpolated on the same horizontal grids
399 and with the same missing values as in COWT, MLOST and in GISTEMP and the respective trends
400 are calculated over the ten-year period starting in January following the initialization (e.g. over the
401 1998-2007 period for the hindcasts starting in November 1997).

402 The uncertainty of the annual-mean GMT evolution is addressed using these three observed datasets,
403 which trends are shown on Figure 6, i.e. black crosses for COWT, blue squares for MLOST and
404 orange triangles for GISTEMP. The trends computed over the last decades are weaker as time
405 progresses and highlight the decrease of the global warming rate. The observations are globally
406 consistent but with COWT exhibiting values about $.1^{\circ}\text{C}$ per 10 years higher. This is mainly due to the
407 differences in spatial coverage among the various observations and shows the impact of high-latitude
408 regions since regions north of 60°N present a trend of about 0.5°C per 10 years higher (Fig. 7b).
409 Indeed Cowtan and Way (2014) have shown that the coverage bias causes a cool bias in recent
410 temperature trend relative to the late 1990s, that increases from around 1998 to present. Hansen et al.

411 (2010) have also shown that the global temperature change is sensitive to estimated temperature in
412 polar regions where observations are limited.

413 The large spread among the hindcasts stands out when different grid points are replaced by missing
414 values as in COWT (black circles), MLOST (blue circles) and in GISTEMP (orange circle). This is
415 due to strong positive trends over the North Pole in the model (Fig. 7a), which are more or less
416 masked depending on the various observed masks. For each start date, the gray area represent the
417 spread (+/- one standard deviation) computed with all the hindcasts, and the white area is the average
418 over all the 3 interpolated hindcasts, for each start-date. The spread is strong with mean trends ranging
419 from -0.08 to +0.19 °C per ten years. From 1999 and on, the warming rate decreases with time in the
420 hindcasts as in the observations with the weakest trends occurring over the last decade (2003-2012).
421 The model is thus able to simulate the hiatus of the 2000's. Interestingly, the hindcasts considered with
422 the spatial coverage of MLOST (CERFACS-HR/MLOST, blue circles) exhibit 10-year trends closer to
423 the raw MLOST observations (blue squares) than CERFACS-HR/GISTEMP or CERFACS-
424 HR/COWT with their respective observations.

425 Note that for the hindcasts starting in November 1997, the model is initialized with an anomalously
426 warm SSTs over the eastern Pacific ocean (the 1998 El-Nino) and the 1998-2007 trends are largely
427 weaker than in the observations (this is even more evident when we only consider the tropical latitudes,
428 not shown). With the standard bias-correction method, it is assumed that the drift is stationary and
429 independent of the initial conditions. The strong discrepancy between the hindcasts and the
430 observations for these strong events suggests that the bias-adjustment method may not be able to
431 correctly remove the drift.

432 The last decade (2003-2012) exhibits the weakest 10-year trend and is reasonably simulated since the
433 difference between the observations and the model does not exceed the model spread (Fig. 6). Figure 7
434 shows the 10-year trend in air surface temperature for CERFACS-HR and COWT (Fig. 7a, b), and
435 ocean heat content [0-2000m] for CERFACS-HR and EN4 (Fig. 7c, d). We used the 0-2000m depth to
436 take into account the upper and mid ocean layers, which have undergone an increase in heat uptake

437 during the hiatus period (Meehl et al., 2011). The horizontal resolution of EN4 is weaker than for
438 CERFACS-HR and each grid cell point contains thus, by construction, more heat. The heat content
439 comparison was made available by scaling the observations and CERFACS-HR by the surface of each
440 grid cell. The heat content values represent thus the heat contained in a 1m^2 area down to a depth of
441 2000 meters. The trend in air surface temperature is positive over the Arctic and the Indian Ocean (Fig.
442 7a). The positive trend over the western Pacific and the negative trend over the eastern Pacific Ocean
443 indicate that CERFACS-HR simulates a negative phase of the PDO. This is consistent with the
444 observed trend (Fig. 7b). Over the Pacific Ocean stippling indicates that the trend is more negative in
445 the observations than in CERFACS-HR. This pattern is however consistent with the literature and is
446 associated with an increase (decrease) in the ocean heat content over the Western Pacific and Indian
447 (eastern Pacific) Ocean (Fig. 7c, d). The positive and significant heat content trend over the Indian
448 Ocean is consistent with Lee et al., (2015) who show that the heat has been transferred from the
449 Pacific Ocean to the Indian Ocean through the Indonesian straits.

450 During this decade the volcanic activity has led to an increase in stratospheric aerosol optical depth
451 that is able to explain a part of the hiatus (Fyfe et al., 2013; Santer et al. 2014; Haywood et al., 2014).
452 In CERFACS-HR the stratospheric aerosol optical depth is prescribed based on the data of Vernier et
453 al. (2011). Unfortunately a technical problem led to an inaccurate representation of these volcanic
454 aerosols in the model leading to a very small effect equivalent to that of a small and constant volcanic
455 aerosol background. Hence, the ability of CERFACS-HR to simulate the hiatus cannot be attributed to
456 the impact of the stratospheric aerosols and can therefore mainly be explained by the ocean
457 initialization and heat uptake. Additional simulations are needed to conclude on a possible impact of
458 the stratospheric aerosols on the modeled hiatus. This is the main topic of a current work with
459 CERFACS-HR.

460

461 The ability of CERFACS-HR to simulate the recent 2000's temperature fluctuation could
462 seem contradictory to inability of CERFACS-HR to simulate the GMST at decadal scale (6-9

463 years). These two results are however not directly linked since the former is associated with
464 the hindcasts of only a few start-dates and the latter is based on all the hindcasts.

465

466 **5. North Atlantic warming of the mid-90's**

467 In this section we address the question of the model ability to predict the North Atlantic warming that
468 occurred during the mid-90's, when the subpolar gyre warmed significantly by 1°C in just 2 years
469 (1995-1996) (Robson et al., 2012a). At the same time the North Atlantic Oscillation (NAO) went from
470 a long positive phase (1989-1995) to a negative phase (1996). Different hypotheses, linking the
471 subpolar gyre and the North Atlantic Ocean heat content (OHC) with the phase of the NAO, have been
472 proposed to explain this sudden warming (Hurrell, 1995; Bersch, 2002; Hurrell et al., 2003; Bersch et
473 al., 2007; Sarafanov et al., 2008; Ortega et al., 2011, 2012; Barrier et al., 2015). The first hypothesis is
474 based on the long positive phase of the NAO (NAO+) and the second on the short and strong negative
475 NAO (NAO-) phase that occurs in 1996 (Lohmann et al., 2009):

476 - A NAO+ phase is associated with heat loss over the subpolar North Atlantic (Barrier et al.,
477 2014) through increased westerlies across the subpolar gyre (Hurrell et al., 2003; Visbeck et
478 al., 2003; Yashayaev, 2007; Barrier et al., 2014). Temperatures cool down, leading to an
479 increase in deep-water formation and to a strengthening of the AMOC (Lohmann et al., 2009,
480 Ortega et al., 2012). This favors an increase in northward oceanic heat transport leading to the
481 warming of the subpolar North Atlantic Ocean several years later (Johns et al, 2011; Msadek
482 et al, 2013).

483 - The NAO- phase has been associated with an increase in OHC over the subpolar gyre through
484 a change of the oceanic circulation and air-surface heat fluxes (Bersch, 2002; Bersch et al.,
485 2007; Sarafanov et al., 2008; Lohmann et al., 2009; Barrier et al., 2015). The NAO- phase can
486 favor a shrinking and a weakening of the subpolar gyre and thus a northward motion of warm
487 subtropical waters into the subpolar gyre (Hatun et al., 2005; Sarafanov et al., 2008). The north
488 Atlantic warming could thus be linked to the abrupt change from a positive to a negative phase

489 of the NAO (as shown by Lohmann et al., 2009 maintaining NAO- conditions in numerical
490 experiments).

491

492 The first hypothesis involves the NAO- phase and therefore implies low-predictability (Collins, 2002;
493 Muller et al., 2005) while the second one implies higher predictability through a lagged AMOC
494 response (Griffied and Bryan, 1997; Pohlmann et al., 2004; Collins et al., 2006, Pohlmann et al.
495 (2013)).

496 5.1 Oceanic changes

497 The subpolar gyre simulated in our hindcasts is represented with the annual-mean barotropic stream
498 function (BSF) in Fig. 8. Negative values indicate a cyclonic gyre that extents from the Labrador Sea
499 to the Iceland Basin with a maximum south-east of Greenland. The subpolar gyre (hereafter noted
500 SPG) area is defined from 60°W to 10°W and from 50°N to 65°N as in Robson et al. (2012ab) and
501 Msadek et al. (2014a) (see the box in Fig. 8).

502 Fig. 9 exhibits the OHC anomalies integrated over the SPG area and from the surface to a 500-meter
503 depth for CERFACS-HR and GLORYS2V3. We used the 0-500-meter integrated anomalies but
504 checked that the warming is consistent over the 0-1000-meter depth (Supplementary material S2).
505 Since the shift occurs in 1995-1996, we focus on hindcasts initialized few years before (i.e. in 1993
506 and 1994; hereafter DH93 and DH94) as well as those initialized in 1995 and 1996 (hereafter DH95
507 and DH96). GLORYS2V3 exhibits a strong warming from 1993 to 1998 (Fig. 9a; black line). DH93
508 and DH94 do not simulate a strong and abrupt warming (with a positive anomaly in heat content) over
509 the SPG (Fig. 9a, b). However the simulations initialized in 1995 and 1996 exhibit a shift (Fig. 9c, d),
510 as shown by the large and sudden increase of OHC during the first four years. DH95 simulates a
511 strong warming (year 1 to 6), with the SPG remaining anomalously warm during 5 years (year 4 to 8),
512 and finally cools down from year 6 to 10 (Fig. 9c). DH96 exhibits a warming that maintains itself
513 during 8 years (Fig. 9d) and thus simulates the observed persistent SPG warming. The following
514 analysis focuses on DH95, which is the first hindcast of the series to simulate the SPG warming.

515 A comparison of the vertically integrated temperature [0-300m] between the GLORYS2V1,
516 GLORYS2V3, ORA-S4 (Balmaseda et al., 2013), and Nemovar1 (Balmaseda et al., 2010) shows that
517 GLORYS2V1 and GLORYS2V3 are close to but warmer than ORA-S4 and Nemovar1 of about 0.5
518 degrees Celsius over the SPG area (Supplementary material S3). ORA-S4 and Nemovar1 cover a
519 longer period than GLORYS2V1 and GLORYS2V3; however, their lower temperature may lead to an
520 overestimation of the warming and therefore these reanalysis were not used throughout this study.

521 The evolution of the volume weighted annual-mean temperature anomalies [0-500m] are shown on
522 Fig. 10 over the 1996-2005 period. The first two years exhibit a positive temperature anomaly
523 centered at 45°N and 40°W (Fig. 10a, b). This anomaly then propagates northward following the
524 eastern side of the SPG (Fig. 10c, d) and reaches the northern SPG boundary during the fifth year (the
525 maximum of warming is located at the south-east of Greenland on Fig. 10e). As in Fig. 9c the SPG
526 then starts to cool down and the positive temperature anomaly slowly vanishes (Fig. 10h-j). The
527 timing of DH95 is not exactly synchronized with the reanalysis since the maximum warming occurs in
528 2001 (instead of 1998 in GLORYS2V3) but its amplitude is well reproduced (about 1°C).

529 Some studies have suggested that the SPG warming is due to the northward migration of subtropical
530 anomalously warm and salty waters. The salt anomalies [0-500m] simulated by DH95 are shown
531 within the figure 11. A positive salt anomaly propagates from the subtropical gyre (Fig. 11a) to the
532 SPG (Fig. 11e-h). At the same time the subtropical gyre exhibits a negative salt anomaly suggesting a
533 transfer from the subtropical to the SPG. As for temperature the maximum is located at the south-
534 eastern tip of Greenland. The positive anomaly located on the eastern side of the subtropical gyre
535 highlight the importance of the gyre as a source of salt and temperature for the SPG warming. This
536 anomaly is also found in Msadek et al. (2014a) in the simulation that reproduces the SPG warming. It
537 also suggests a possible role of the Mediterranean Sea in exporting anomalously warm and salty
538 waters into the subtropical gyre.

539 The subtropical gyre strengthens along with the North Atlantic current during the first five years of
540 DH95 (Fig. 12a-e). In 1998 a robust and positive anomaly extends from the north-west (55°W-40°N)

541 of the subtropical gyre to the east (40°W-55°N) of the SPG (Fig. 12c). It reinforces in 1999 (Fig. 12d)
542 denoting a northward shift of the Gulf Stream (for more clarity the mean climate is not shown but the
543 reader can refer to the Fig. 8). The BSF weakens from 1998 to 2000 over the eastern boundary of the
544 SPG. The SPG continues then to weaken (Fig. 12d-f), especially in 2003 (Fig. 12f).

545 In summary the strengthening of the north Atlantic current favors the northward move of a salty and
546 warm anomaly that reaches the Iceland Basin, south of Greenland and the Labrador Sea several years
547 later (see Fig 11e for the salt and Fig 10e for the temperature).

548 The evolution over the 1996-2002 period of the Meridional Heat Transport (MHT) integrated over the
549 whole oceanic column is analyzed in Fig. 13. MHT increases from 1996 to 2000 (with significant
550 anomalies for the first 2 years) north of 40°N with a maximum at 55°N (Fig. 13a). MHT decreases
551 after 2000. The MHT change is thus consistent with the temperature evolution over the SPG, from
552 40°N to 60°N (Fig. 10).

553 The MHT anomaly is decomposed in the anomalous advection of mean temperature ($\overline{Tv'}$), the mean
554 advection of anomalous temperature ($\overline{vT'}$) and the cross-product of anomalies ($T'v'$) (Fig. 13).

555 This decomposition is computed at a monthly time-scale, bars denoting monthly means and prime the
556 departure from the monthly means. These products, averaged in annual means are shown in figure 13.

557 The anomalous advection $\overline{Tv'}$, denoting an enhanced northward heat transport over the SPG
558 associated with a strengthening of the oceanic circulation, dominates the MHT change (the anomalies

559 of $\overline{Tv'}$ are significant from 1996 to 2002) between 50°N and 55°N (Fig. 13b). The anomaly of MHT
560 due to the mean advection of anomalous temperature ($\overline{vT'}$) is positive and significant north of 55°N

561 during the first five years of DH95 (Fig. 13c). This ($\overline{vT'}$) anomaly is located at 55°N during the first
562 3 years, then moves northward and reaches 60°N from 1998 to 2001; it is generally weaker than the

563 $\overline{Tv'}$ anomaly but dominates the MHT anomalies at 60°N during the 1998-2000 period. This mean
564 advection of temperature anomalies is thus sustaining the warming first created by the anomalous

565 advection of mean temperature. The ($T'v'$) component is weak and can be neglected over the SPG

566 (Fig. 13d). During the last two years of DH95, the decrease of temperature is associated with a weaker
567 MHT linked with a decrease of both \overline{vT}' and \overline{Tv}' .

568 Fig. 14 shows the mixed-layer depth anomalies averaged over the years 1996-2000. The mixed layer is
569 deeper, consistent with more deep water formation over the Labrador Sea. It denotes an increase of
570 convective activity (Ortega et al., 2011), leading to an AMOC strengthening and an increase of
571 meridional heat transport (Johns et al., 2011; Msadek et al., 2013). The deepening of the mixed layer is
572 also consistent with the \overline{Tv}' increase and confirms (as in Msadek et al., 2014a) the prominent role of
573 the ocean dynamics on the SPG warming. The AMOC is particularly strong in November 1995 (and
574 stronger than for the other start date) in GLORYSV21 (not shown). The DH95 dynamical shift is
575 probably, at least partly, built-in in the initial condition. The mixed-layer depth anomaly weakens
576 after 2002 and vanishes during the second half of the simulation (2001-2005).

577 5.2 Atmospheric changes

578 Some studies have also suggested a possible role of heat fluxes in relation with the atmospheric
579 variability (Bersch, 2002; Bersch et al., 2007; Sarafanov et al., 2008; Lohmann et al., 2009). The
580 surface fluxes are shown in the Fig. 15. By convention positive (negative) values denote downward
581 (upward) anomalies. The net surface heat flux (the sum of net shortwave, net longwave, sensible and
582 latent heat fluxes) averaged over the SPG exhibits positive values over the 1998-2003 period (Fig.
583 15a), which means that the atmosphere is warming the ocean in DH95. The surface flux anomalies are
584 stronger over the Labrador Sea, the Irminger Sea and the Iceland Basin (Fig. 15b). However, the
585 strongest positive anomaly is located south of 50°N (southern boundary of the SPG). The net
586 shortwave and longwave fluxes show weak and non-significant anomalies (Fig. 15c, d). Sensible and
587 latent heat fluxes exhibit stronger positive anomalies (Fig. 15e, f), denoting a decrease in oceanic heat
588 loss.

589 We now discuss changes in the surface wind speed and sea-level pressure (SLP) average over the
590 1998-2003 period (Fig. 16). The SLP denote a negative phase of the NAO with positive (negative)

591 anomalies of SLP over Iceland (the subtropical latitudes). The SLP exhibits a cyclonic anomaly
592 between 30°N and 50°N and an anticyclonic anomaly north of 50°N. The wind decreases over the
593 south-eastern boundary of the SPG and acts to weaken the SPG, favoring the northward move of a
594 salty and warm anomaly from the subtropical to the SPG, as described in Hatùn et al. (2005). The
595 decrease of the wind speed over the SPG, associated with the pressure anomalies, is able to weaken
596 the oceanic heat loss through the weakening of the latent and sensible heat fluxes (Fig. 15e, f). These
597 changes are strong in winter (DJF) but not consistent in summer (JJA) (not shown). CERFACS-HR is
598 able to reproduce the negative phase of the NAO. However this does not indicate a skill in predicting
599 the NAO since only concerning one start-date: the NAO is not skillful on interannual time-scales in
600 CERFACS-HR (not shown). We have computed the fraction of heat increase due to the net oceanic
601 heat transport (difference between the heat entering the southern boundary of the SPG and the heat
602 leaving its northern boundary) and the part coming from the atmosphere (the net heat surface fluxes).
603 One third of the OHC change is due to atmospheric changes over the 1998-2003 period. The
604 anomalies of net surface heat fluxes are thus not negligible.

605 The pressure anomaly may be due to a remote effect of the Atlantic SST (Czaja and Frankignoul, 2002)
606 and/or East Pacific SST (Huang et al., 2002; Brönnimann, 2007) on the NAO. With only five members
607 we cannot assess the robustness of such a relationship. This will need to be confirmed when larger
608 ensemble sizes will be routinely used with high-resolution models.

609

610 **6. Conclusion**

611 CERFACS-HR was first evaluated in term of its ability to predict temperature and precipitation.
612 CERFACS-HR shows some skill at predicting the AMO, Indian and tropical Atlantic SST as shown
613 by related indexes that exhibit reasonable correlations for 1-4 to 3-6 year lead-time. However
614 CERFACS-HR presents no skill in predicting the eastern pacific SSTs. Skill in predicting SSTs was
615 found in other studies (Chikamoto et al., 2013; Bellucci et al., 2013, 2014; García-Serrano et al., 2015;
616 Karspeck et al., 2015; among others) and are mainly due to the warming trend. The skill improvement
617 due to a spatial resolution increase is not assessed in this study. We nevertheless observe that the
618 CERFACS-HR skill in predicting temperature is slightly weaker than in the literature. This can be
619 explained by several factors: (i) the model tuning aims to improve the ability of a climate model to
620 simulate the mean climate. High resolution coupled climate models are still computationally very
621 expensive. Less sensitivity tests are thus performed with high-resolution models meaning that they are
622 usually less optimized than low-resolution AOGCMs. The bias reduction associated with the time
623 devoted to preliminary tests is important as it is related with prediction performance (DeIsole and
624 Shukla, 2010). (ii) We only ran a few start dates covering a period with no major volcanic events. In
625 contrast, the studies based on the CMIP5 models used longer time periods (1960-2009), including
626 major volcanic events (El-Chichón, Pinatubo) that improve model skill (Guemas et al., 2013a). (iii)
627 Another difference comes from the start-date frequency, every 5 years in the CMIP5 protocol and
628 every year in the present study. García-Serrano et al. (2015) have noticed different results using 5-year
629 rather than 1-year date frequency.

630 Interestingly, the model is able to simulate the observed slow-down of the global warming rate over
631 the recent period. This hiatus is associated with a negative phase of the PDO and an increase of the
632 ocean heat uptake in both CERFACS-HR and observations. This is consistent with Meehl et al. (2011),
633 Kosaka and Xie (2013), and Watanabe et al. (2013) among others. Since the observed PDO has
634 experienced a negative phase during the 2000's (Trenberth and Fasullo, 2013), the question arises as
635 to the main driver of the PDO variations, external forcing (e.g. the volcanic forcing due to the small to

636 medium size eruptions after 2005) or internal variability. In order to confirm the role of ocean
637 initialization versus the external forcing, we have performed two other sets (6 members each) of 10-
638 year simulations starting from November 2002 and differing by the applied volcanic forcing (a
639 constant weak background is used in one set while the observed forcing is used in the second set). The
640 two sets simulate a negative phase of the PDO (Monerie et al, to be submitted) with the main
641 contribution coming from the ocean initialization and an additional cooling due to volcanic forcing
642 (Fyfe et al. 2013; Santer et al. 2014; Haywood et al. 2014; Ridley et al., 2014; Schmidt et al., 2014;
643 Brühl et al., 2015; Mills et al., 2016).

644 The last part of the paper focuses on the case study of the North Atlantic warming of the mid 1990's.
645 The ability of the model to reproduce such an event is not directly related to the overall skill over the
646 North Atlantic. The simulation initialized in November 1995 simulates a quick warming, of about 1°C,
647 in just five years. The analysis of temperature, salt and circulation anomalies reveals that this warming
648 is associated with a strengthening of ocean dynamics. The subtropical gyre strengthens and its north-
649 western boundary transports the anomalously warm water to the SPG. The SPG then warms and
650 weakens. The increase of the North Atlantic temperature through a strengthening of the oceanic
651 circulation is consistent with Robson et al. (2012ab), Yeager et al. (2012) and Msadek et al. (2014a).

652 DH95 exhibits a robust strengthening of the heat transport by the anomalous advection of mean
653 temperature ($\overline{Tv'}$). The SPG warming predictability comes mainly from the initialization of the ocean.
654 Note that DH96 simulates an increase of the mean SPG temperature with a MHT change largely
655 dominated by the mean advection of temperature anomalies ($\overline{vT'}$) north of 50°N (Supplementary
656 material S4). In DH96 the shift has already occurred and the temperature anomaly is incorporated in
657 the initial conditions. The warm anomaly is located inside the SPG at the beginning of DH96 and stay
658 inside the SPG during the simulation period (Fig. 9d). The reproducibility of the shift mechanisms is
659 thereby dependent on the initial conditions. Interestingly analysis of DH96 confirms that the SPG
660 warming may be sustained by the $\overline{vT'}$ anomaly.

661 There is also a contribution from the net surface heat flux indicating a warming of the ocean by the
662 atmosphere over the 1998-2003 period. A negative phase of the NAO is simulated during this period
663 and favors a decrease of the wind speed over most of the SPG. This has multiple consequences: (i) the
664 latent and sensible heat fluxes weaken and reduce heat loss from the ocean (ii) the decrease of the
665 wind speed over the south-eastern boundary of the SPG acts on the SPG circulation, favoring the
666 northward move of the warm anomaly along the gyre eastern boundary. The warming of the SPG in
667 CERFACS-HR is thus due to both an increase of the oceanic heat transport and of the local change
668 due to atmospheric forcing, which sustains the positive anomaly of temperature. This is consistent
669 with the relation found in Bersh (2002), Bersh et al. (2007), Sarafanov et al. (2008), Lohmann et al.
670 (2009) and Barrier et al. (2014) between a NAO- phase and anomalously warm SSTs over the North
671 Atlantic Ocean. We hypothesize that this negative phase of the NAO acts to reduce the meridional heat
672 transport and the anomaly of mixed-layer over the Labrador Sea, in consistency with Ortega et al.
673 (2012).

674 In this study we show that the north Atlantic warming of the mid-1990's is predictable due to the
675 ocean initialization (ocean memory): DH95 is initialized with a strong AMOC and thus with an
676 anomalously strong oceanic heat convergence.

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682

683 Captions

684 Figure 1: (a) CERFACS-HR minus ERAI difference in surface air-temperature ($^{\circ}\text{C}$), averaged over all the start-dates and
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687 computed with GLORYS2V3 (contour) and CERFACS-HR (color), and its (e) CERFACS-HR minus GLORYS2V3
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690 Figure 2: Anomaly coefficient correlation of the SST hindcasts (with respect to ERSST3b) for a) year 1, b) years 2-5 and c)
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700 Figure 4: Anomaly coefficient correlation of the SST hindcasts (with respect to ERSST3b) for a) the global ocean, b) the
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705 Figure 5: Ten-year linear trends for 1994-2003 to the 2003-2012 periods for a) winter SIE, b) summer SIE, c) winter SIV and
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711 Figure 6: Ten-years linear trends ($^{\circ}\text{C}$ by ten years) for 1998-2009 to the 2003-2012 time-period for COWT (black cross),
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713 with the same spatial coverage of the observations. Each hindcast is represented by a circle and the ensemble mean by a
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717 Figure 7: Ten-year linear trend ($^{\circ}\text{C}$ by decade) for the 2003-2012 period for the annual mean air-surface temperature for a)
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724 (clockwise) circulation. The subpolar gyre area (SPG) is defined as the box represented in black: $[60^{\circ}\text{W} - 10^{\circ}\text{W}; 50^{\circ}\text{N} -$
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728 spread is computed on the 5 members as more or less 1 standard deviation (red shading). The zero line is in gray.

729 Figure 10: Temperature anomalies ($^{\circ}\text{C}$) [0-500m] for DH95 and from 1996 to 2005. The anomalies are computed with
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732 Figure 11: Same as Fig. 10 but for salt anomalies (psu).

733 Figure 12: Same as Fig. 10 but for barotropic streamfunction anomalies (Sv).

734 Figure 13: a) Meridional heat transport anomalies tv (PW) and its decomposition into b) $\overline{T}v'$, c) $\overline{v}T'$ and d) $v'T'$ for
735 DH95. The anomalies are computed in regard to the average over all the 10-years hindcasts. Bars denote the monthly means
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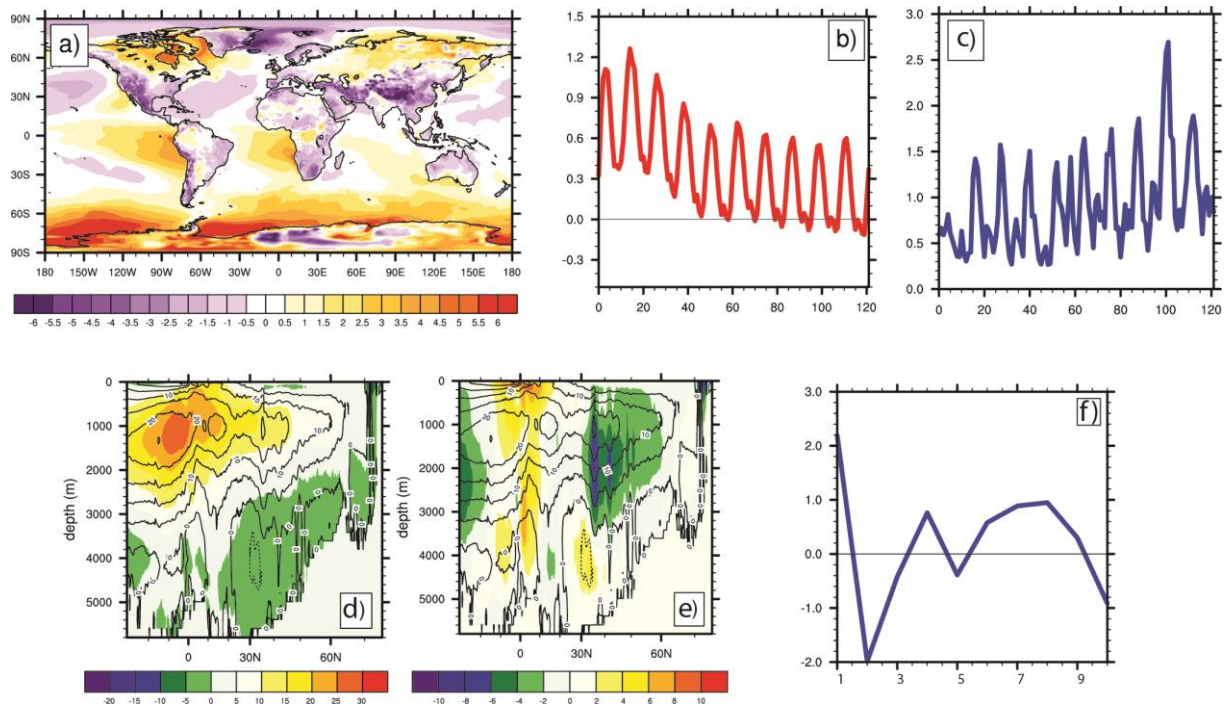
738 Figure 14: Mixed layer depth anomalies (m) for the years 1-5 of DH95. The anomalies are computed with respect to the
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741 Figure 15: a) Surface net fluxes (net shortwave + net longwave + sensible + latent) anomalies ($W m^{-2}$) averaged over the
742 SPG, b) 3-8 year average of net surface flux anomalies and its decomposition into the c) net shortwave anomaly, d) net
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746 Figure 16: 3-8 year average of sea-level pressure (hPa, color) and surface wind speed anomaly ($m s^{-1}$, vector). The anomalies
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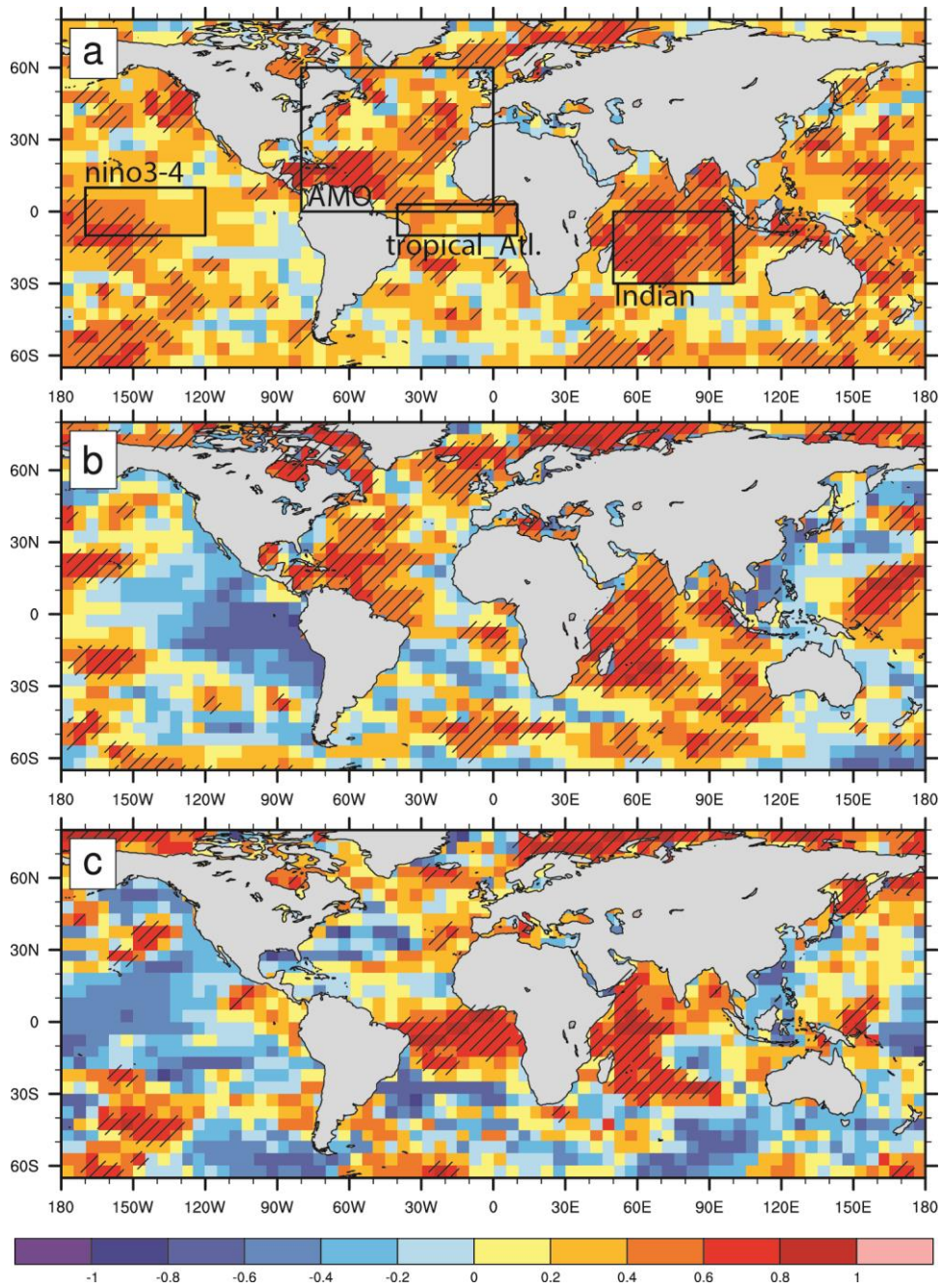
751 **Figures**



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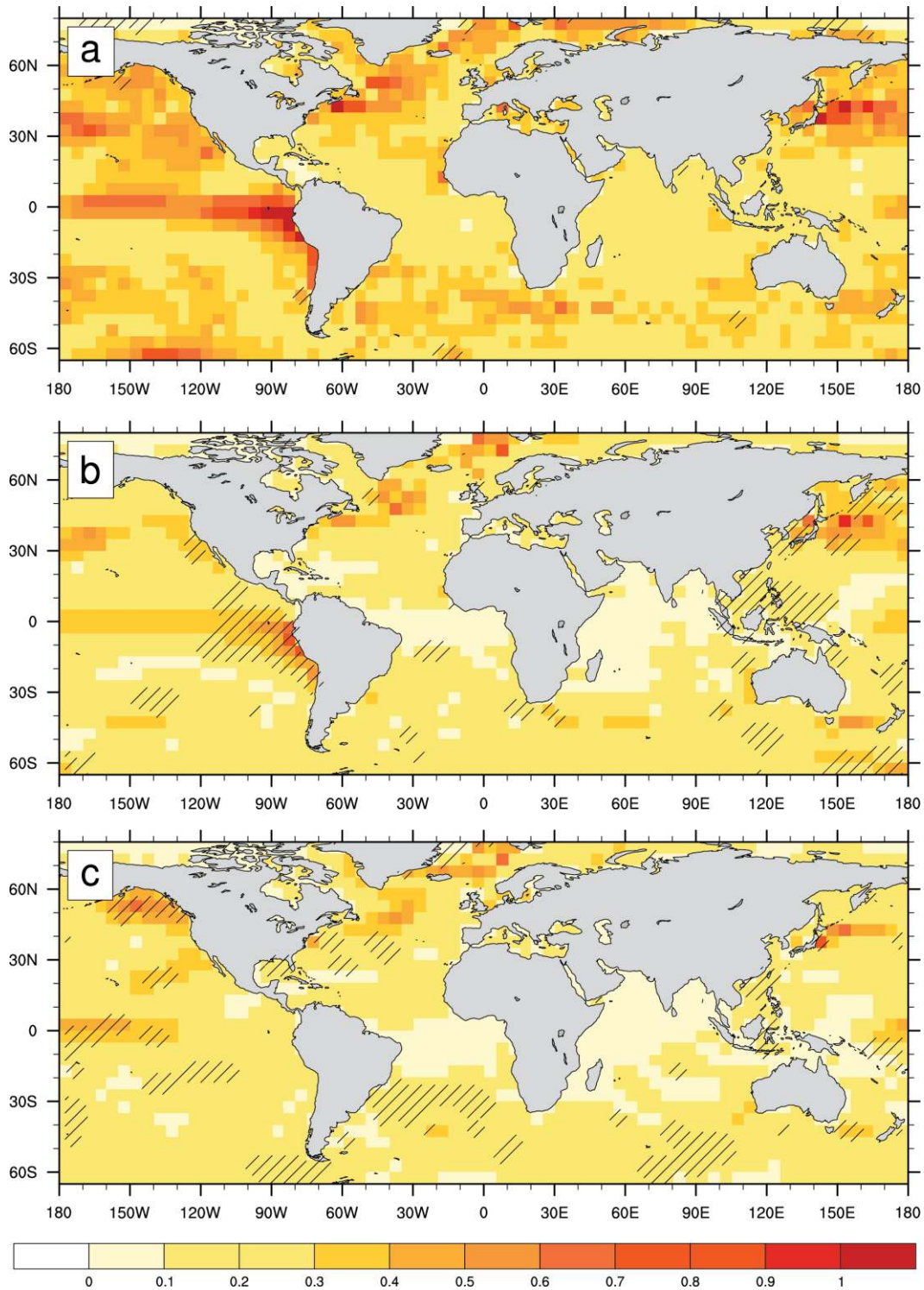
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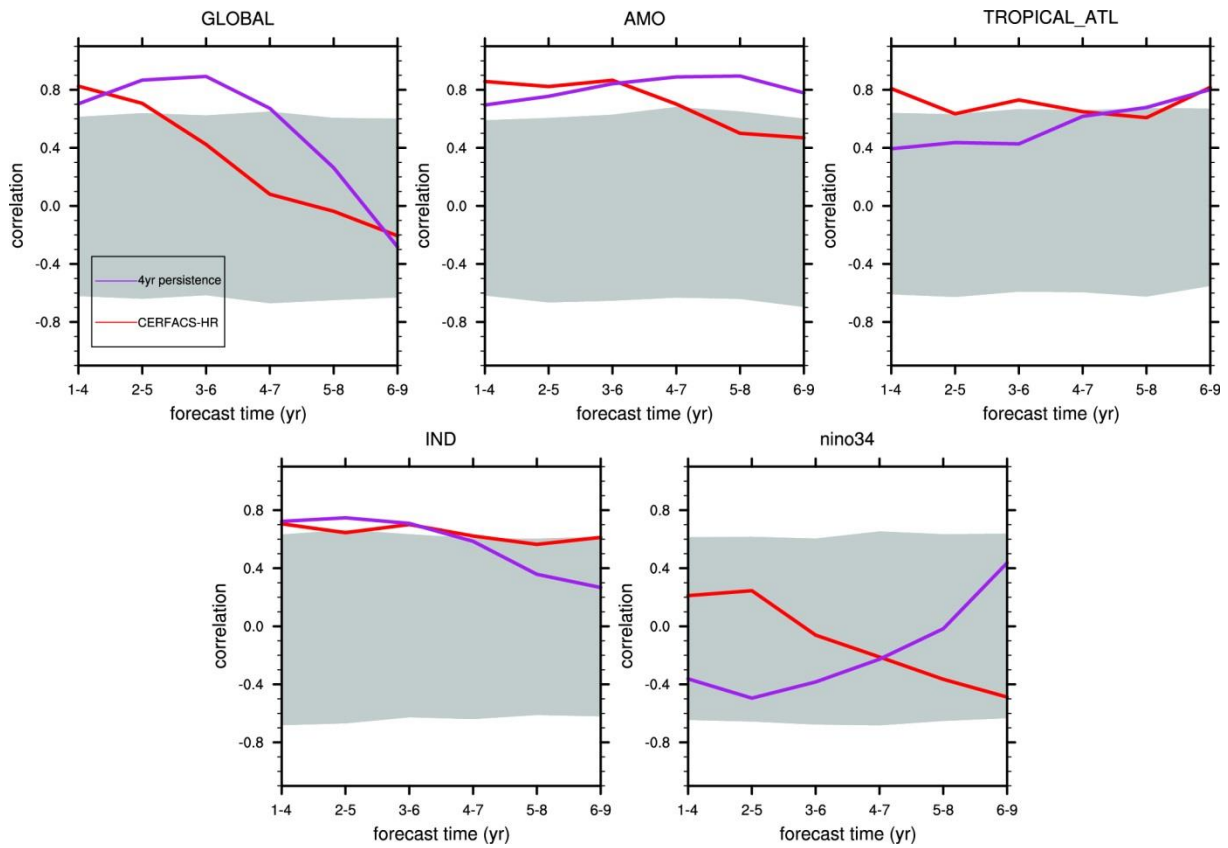


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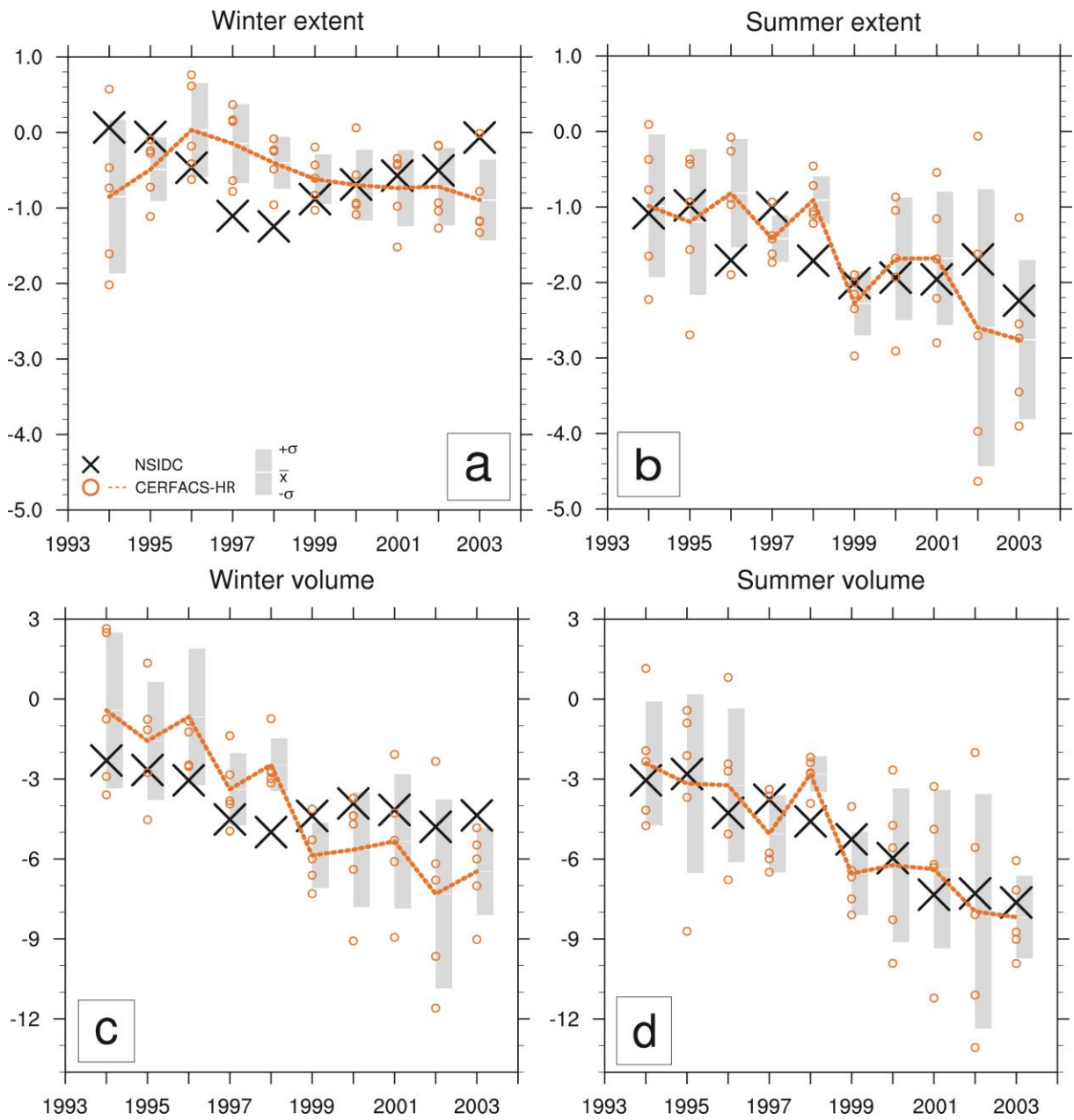
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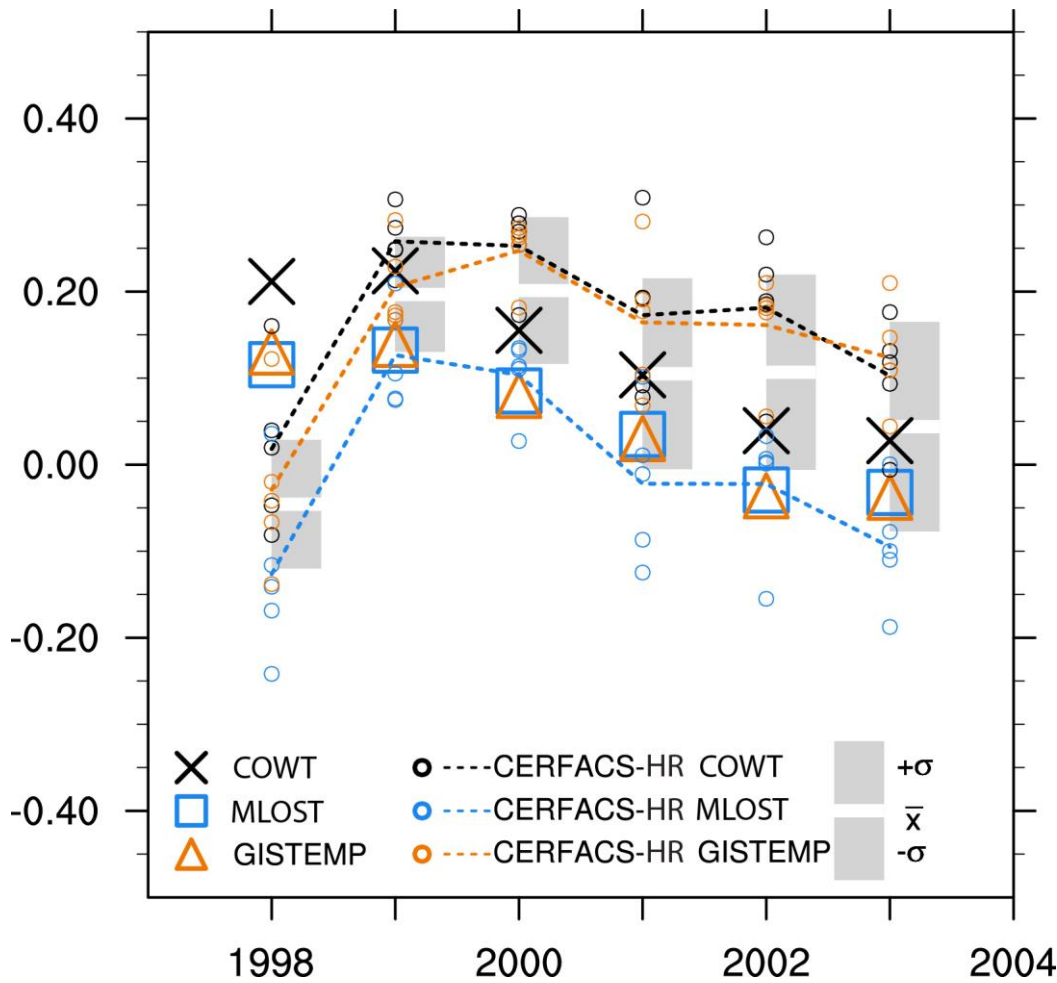
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798 Figure 6: Ten-year linear trends (expressed in °C by ten years) for 1998-2007 up to the 2003-2012 time-period for COWT
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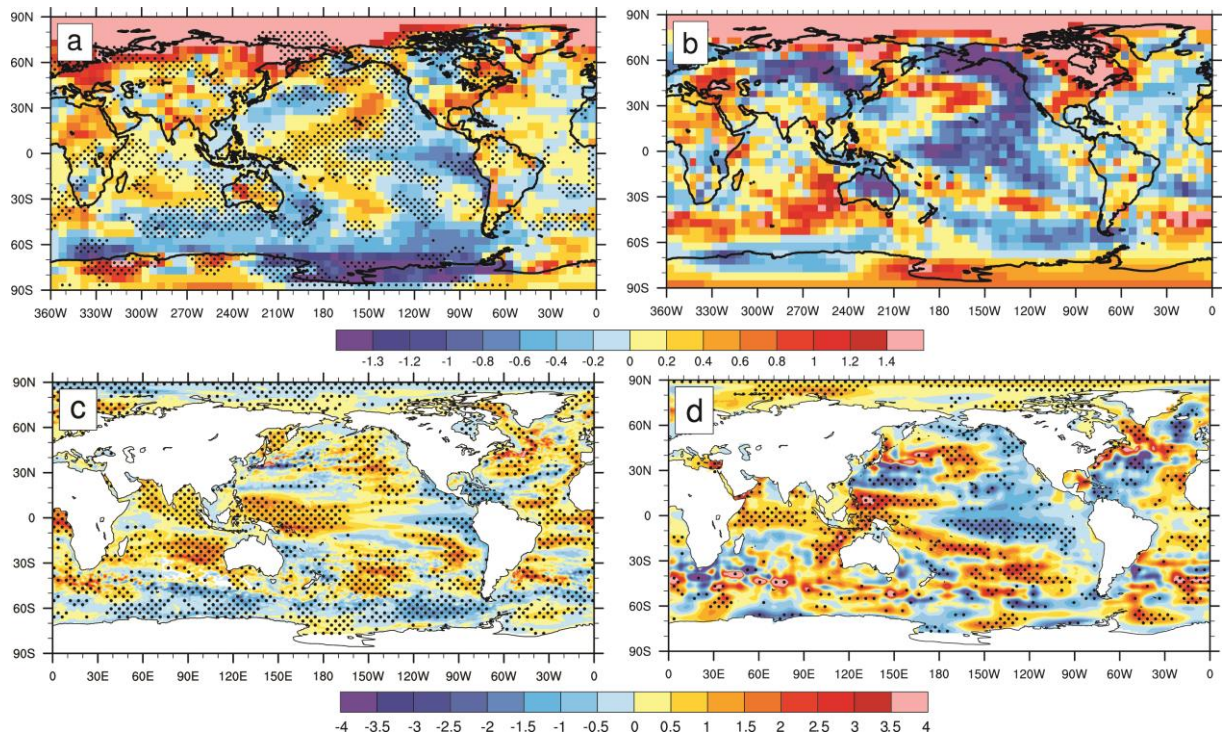
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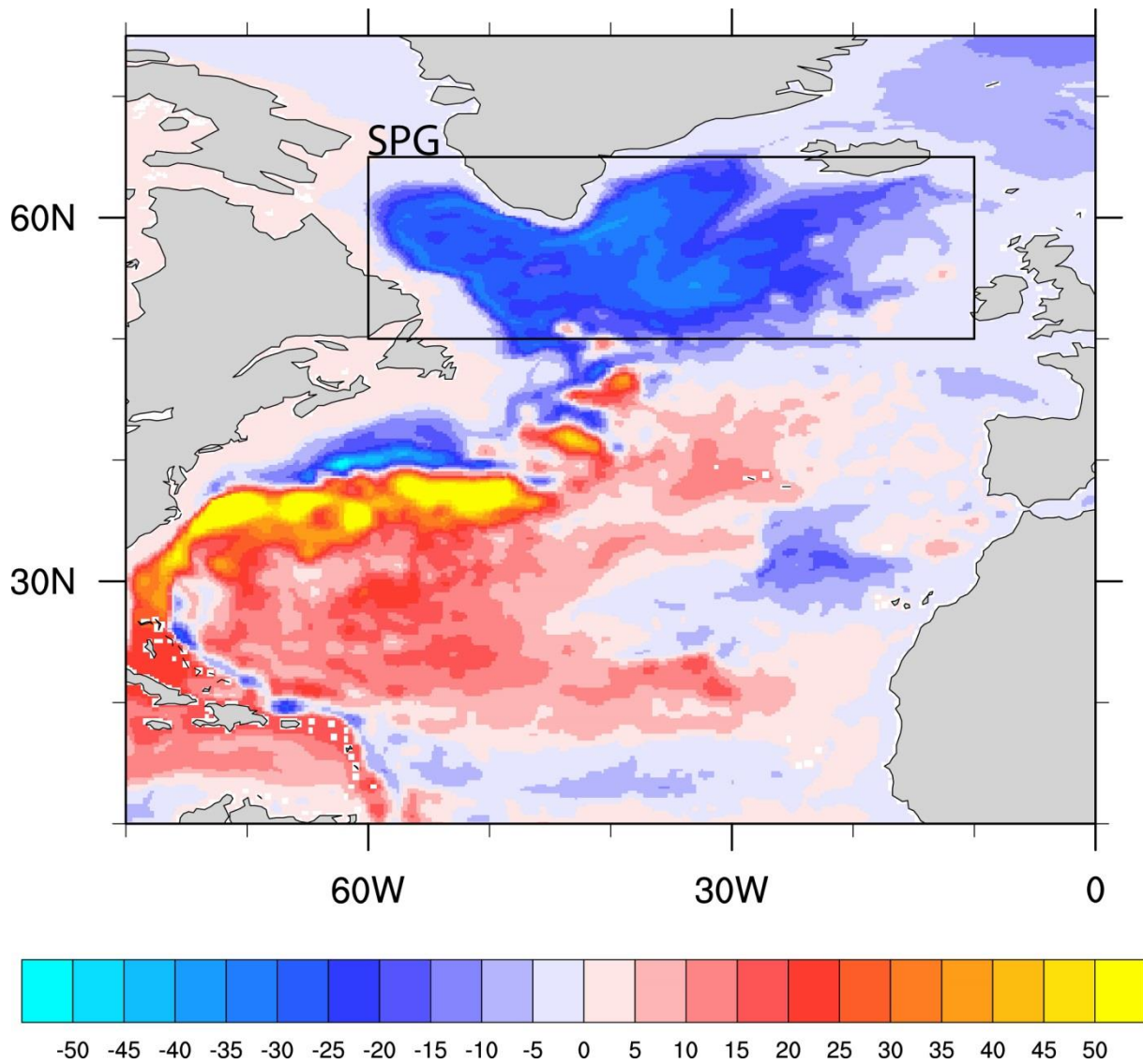
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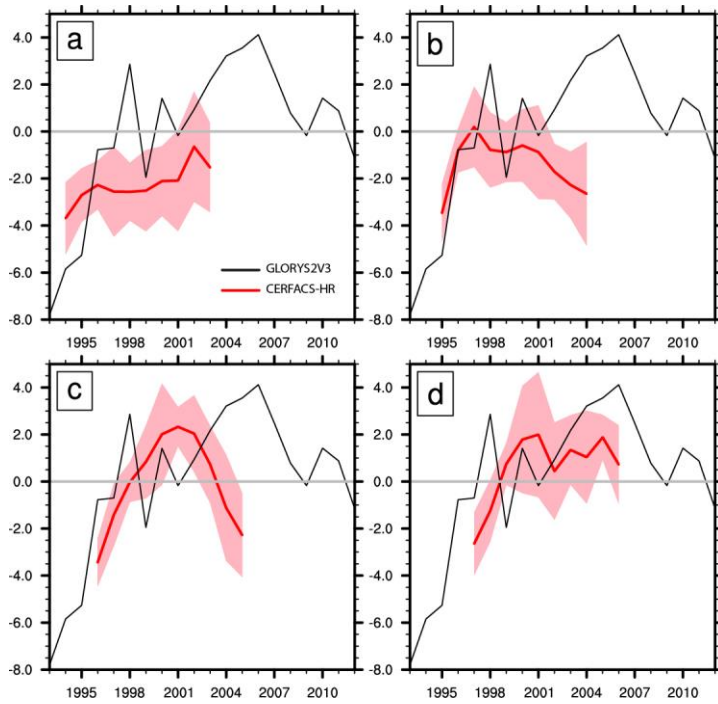


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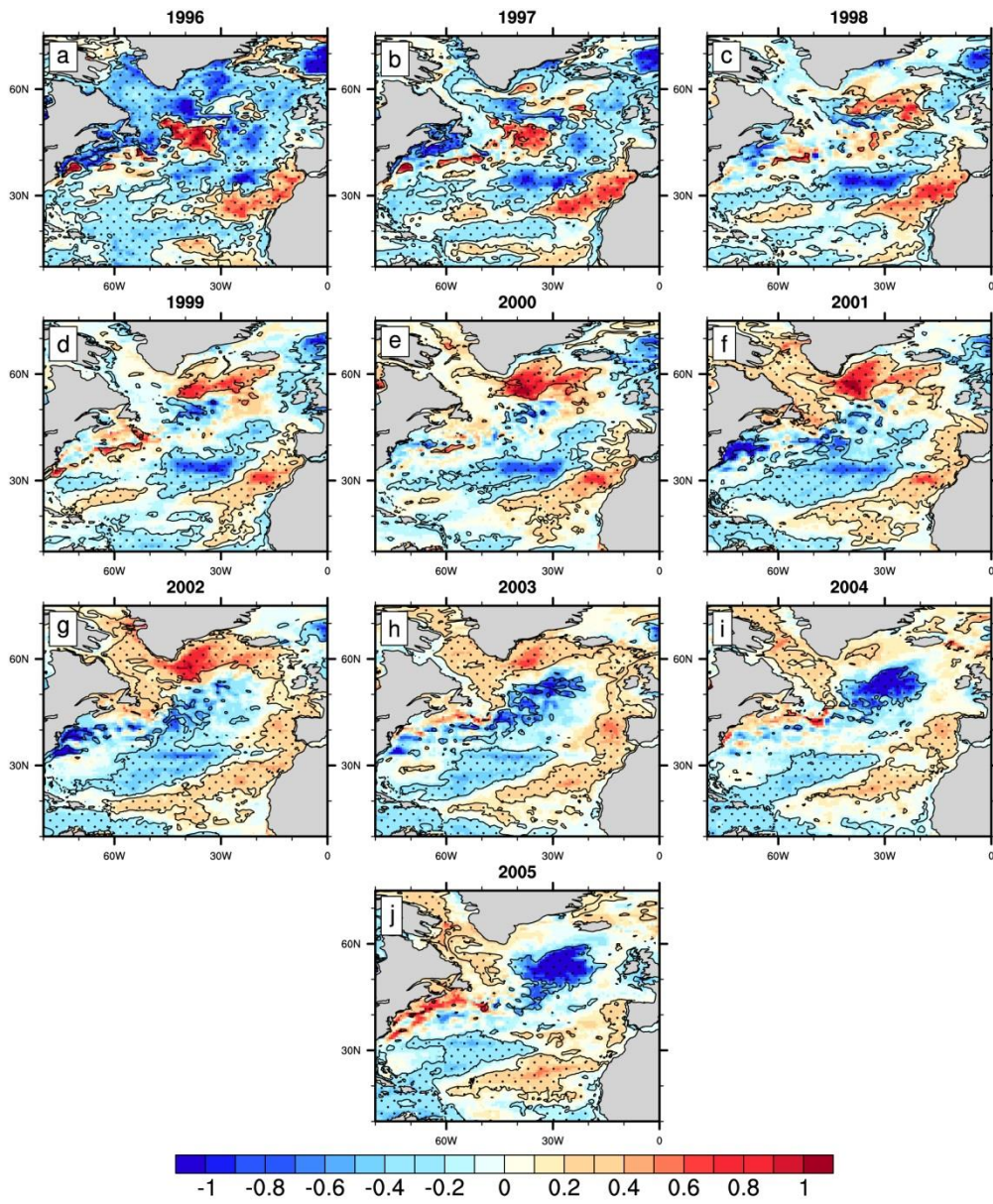
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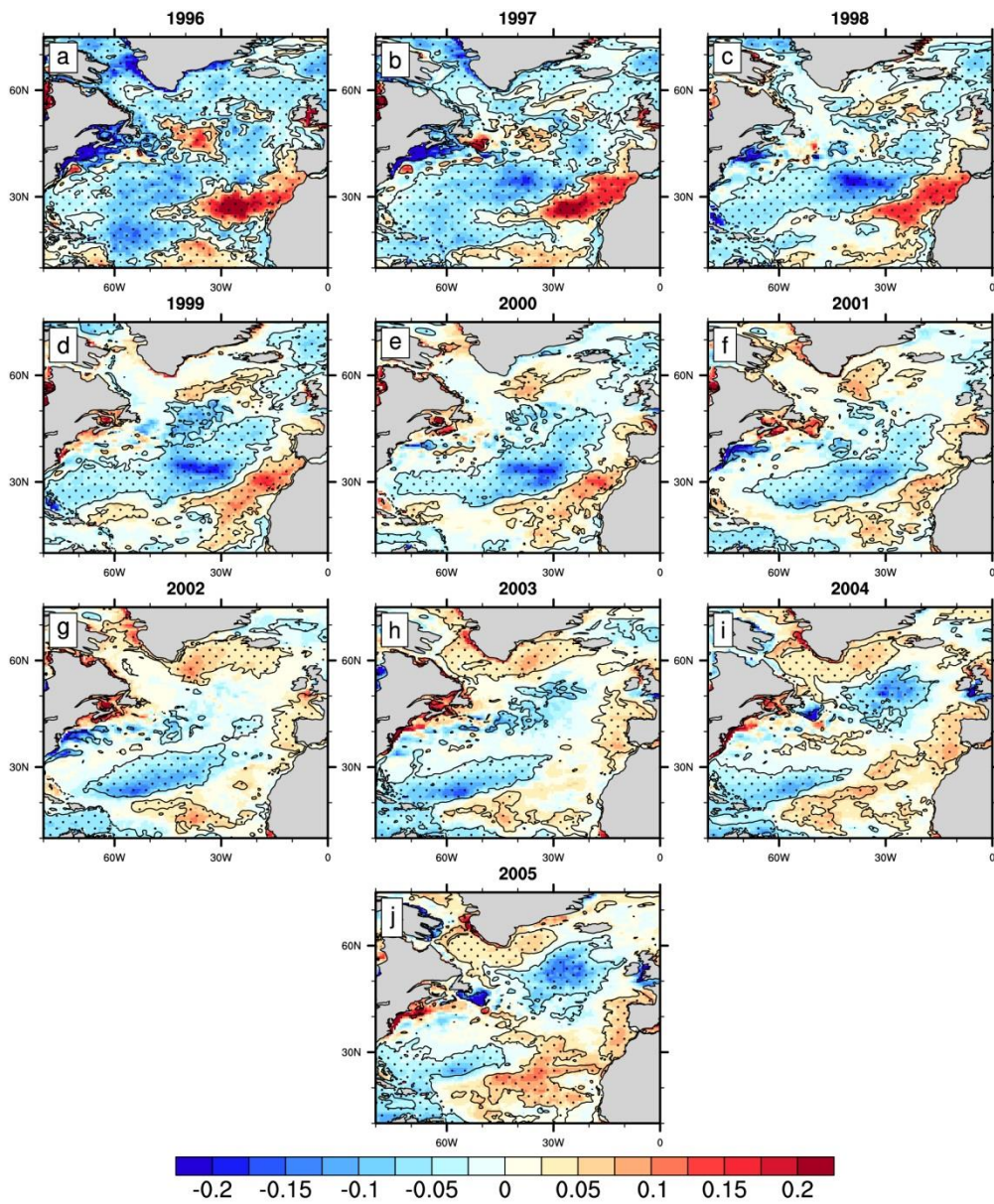
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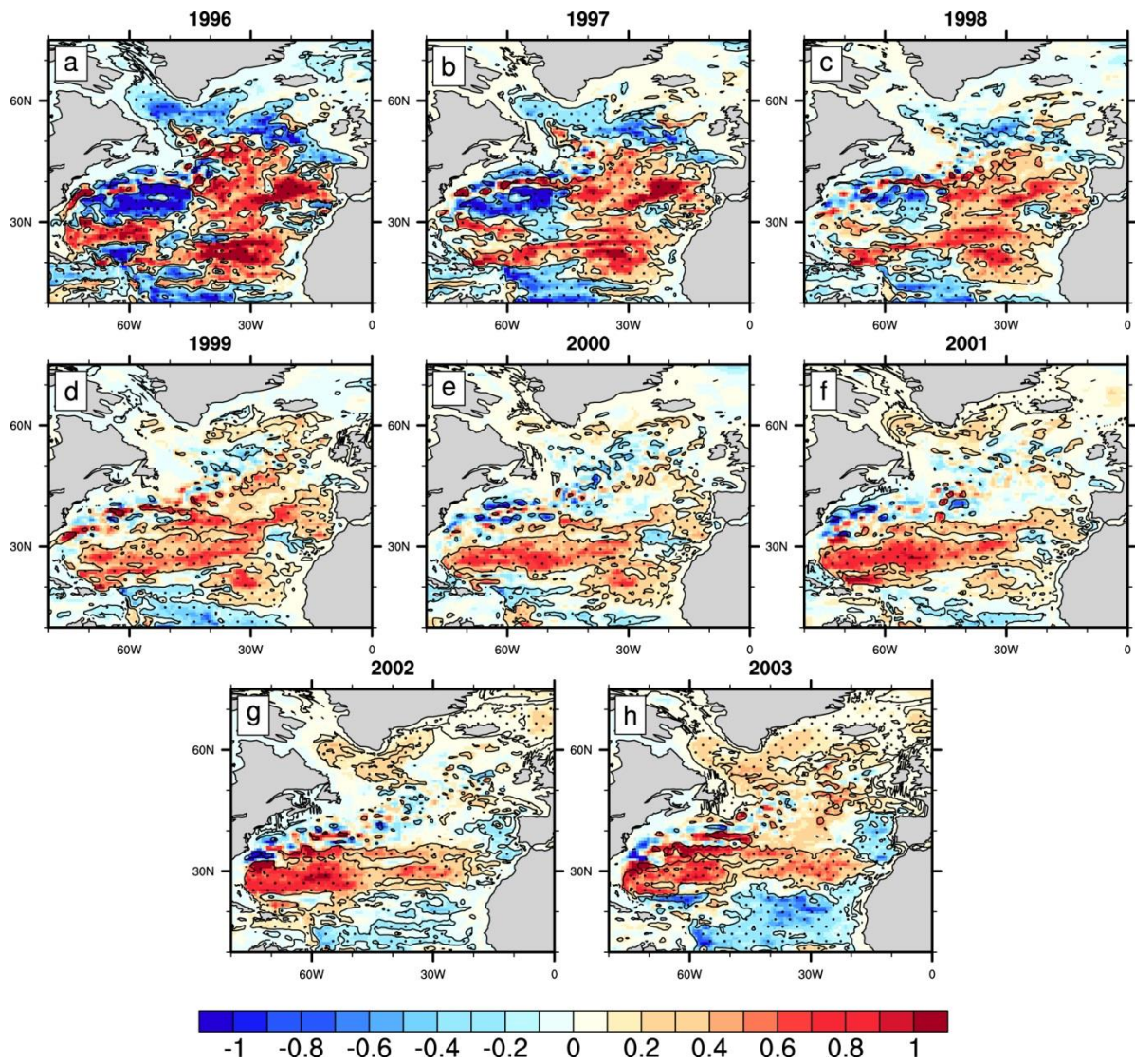
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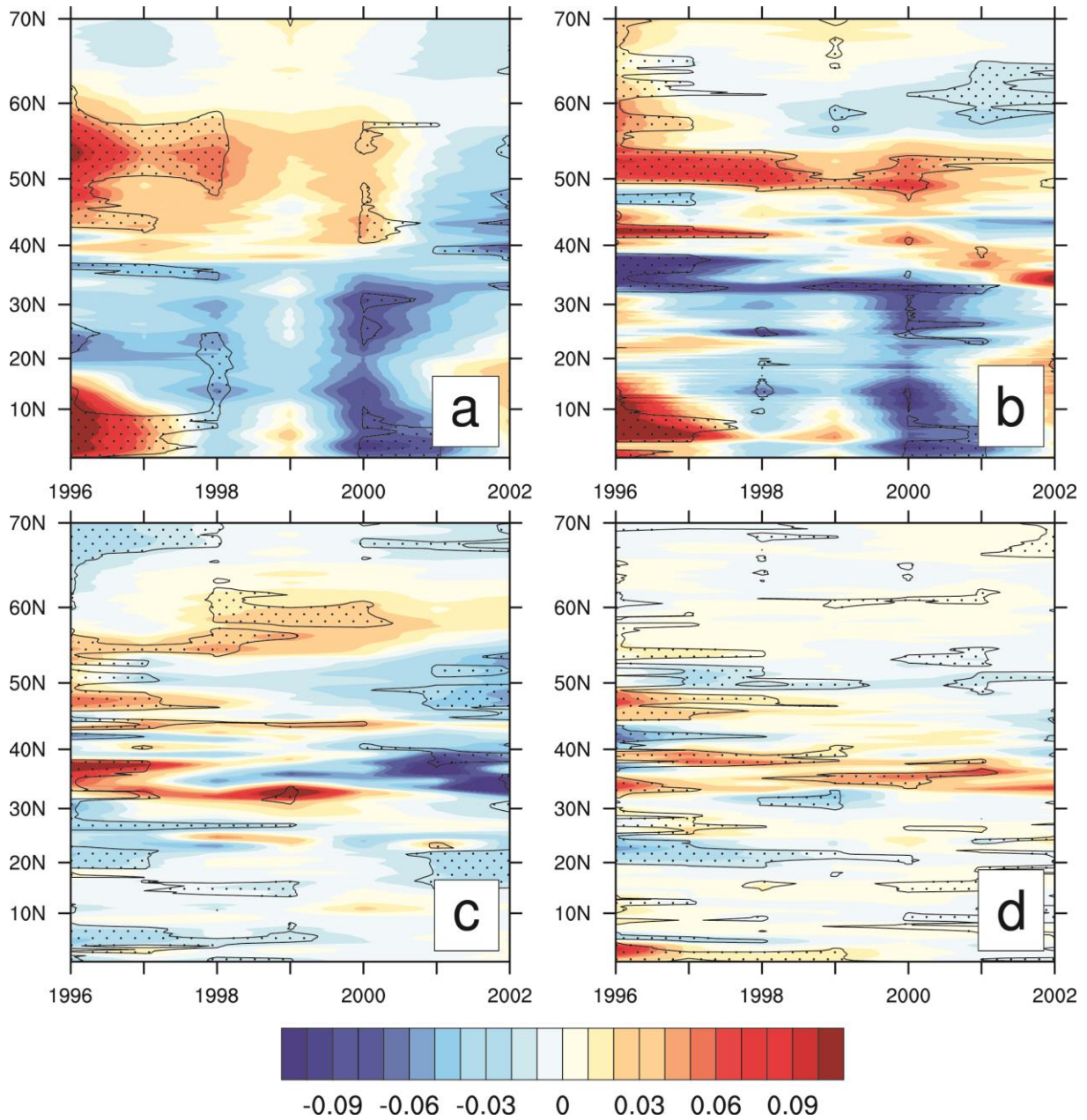
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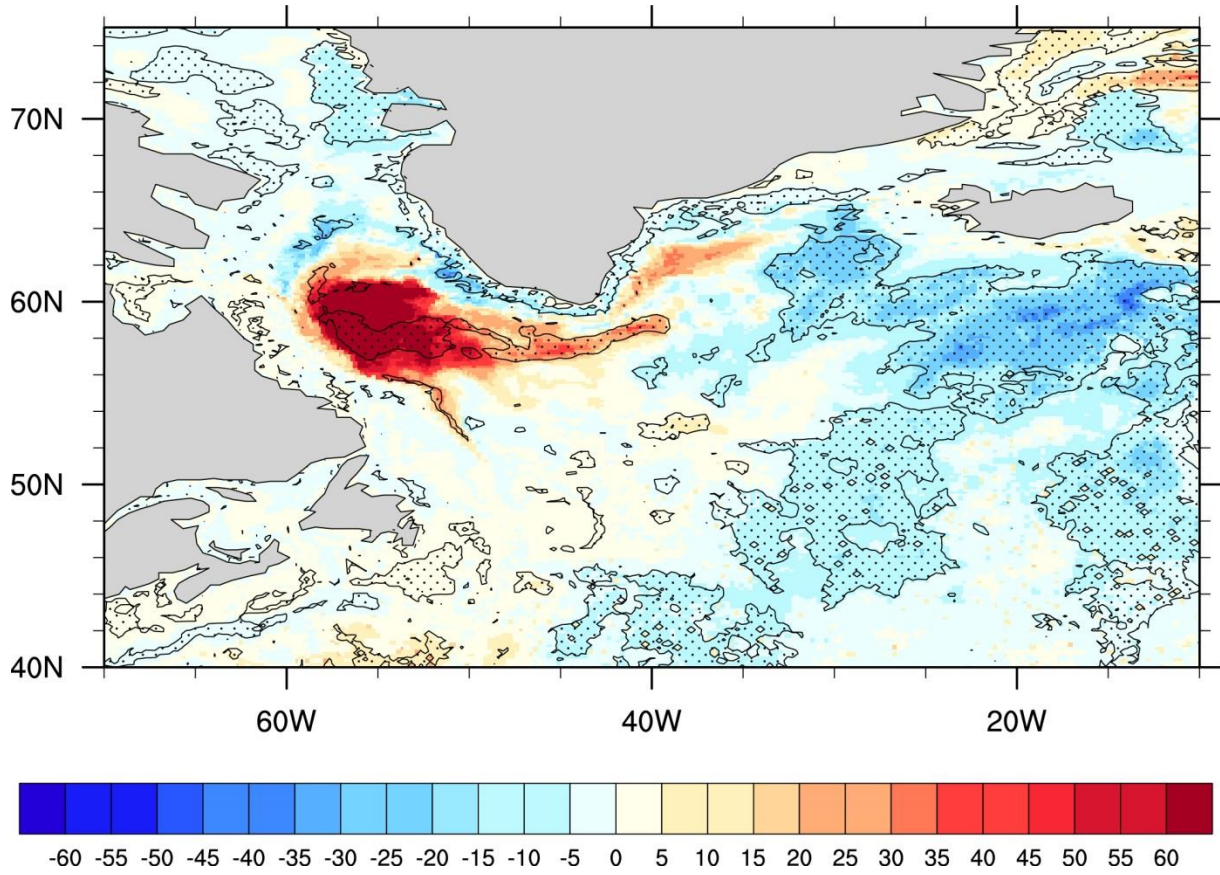
869 Figure 13: Evolution of vertically integrated (in PW) a) meridional heat transport anomalies vT and its decomposition into
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Figure 14: Mixed layer depth anomalies (m) for the years 1996-2000 of DH95. The anomalies are computed with respect to the average over all the 10-year hindcasts. Dots indicate that anomalies are significant at the 95% confidence level according to a Student t-test.

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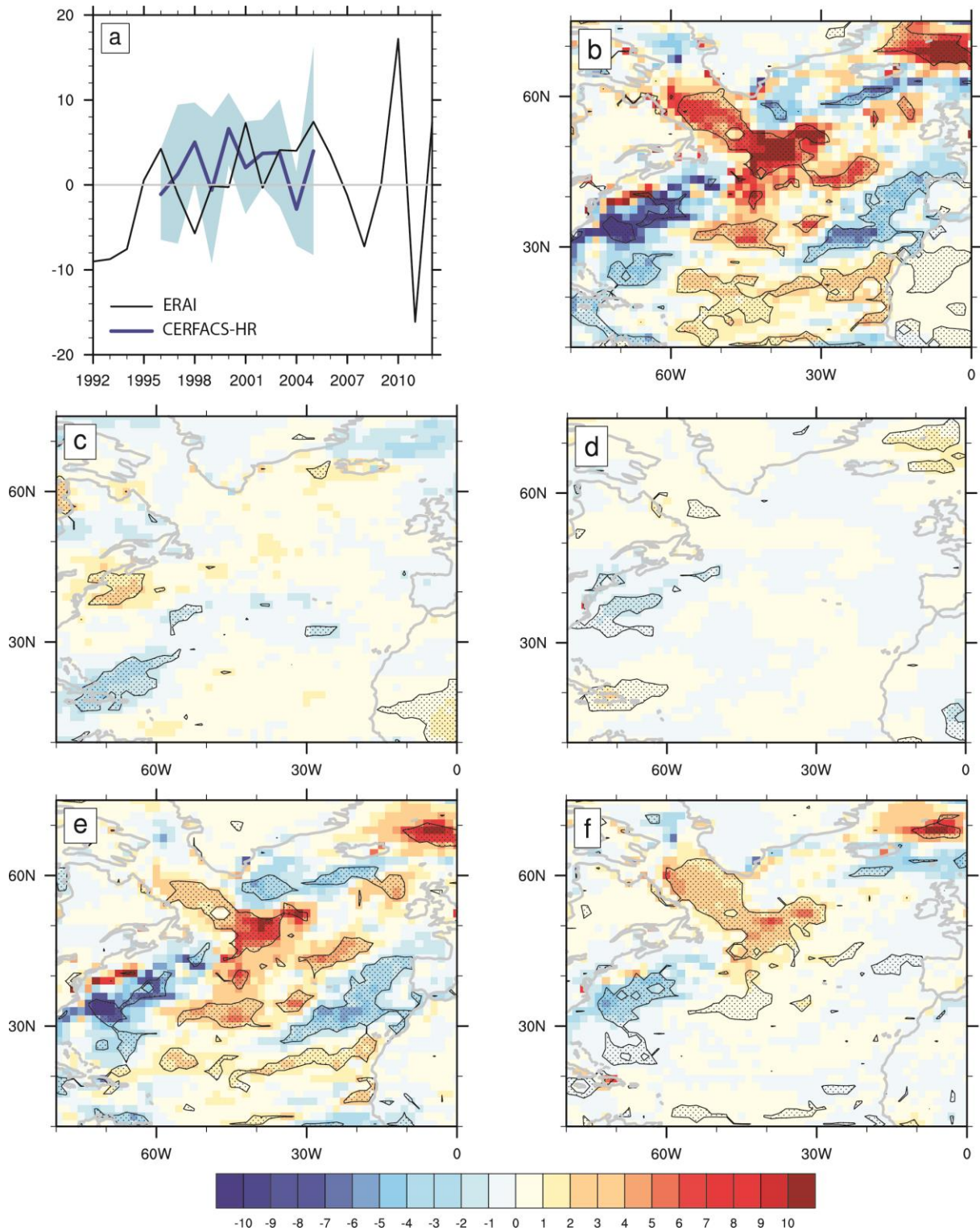
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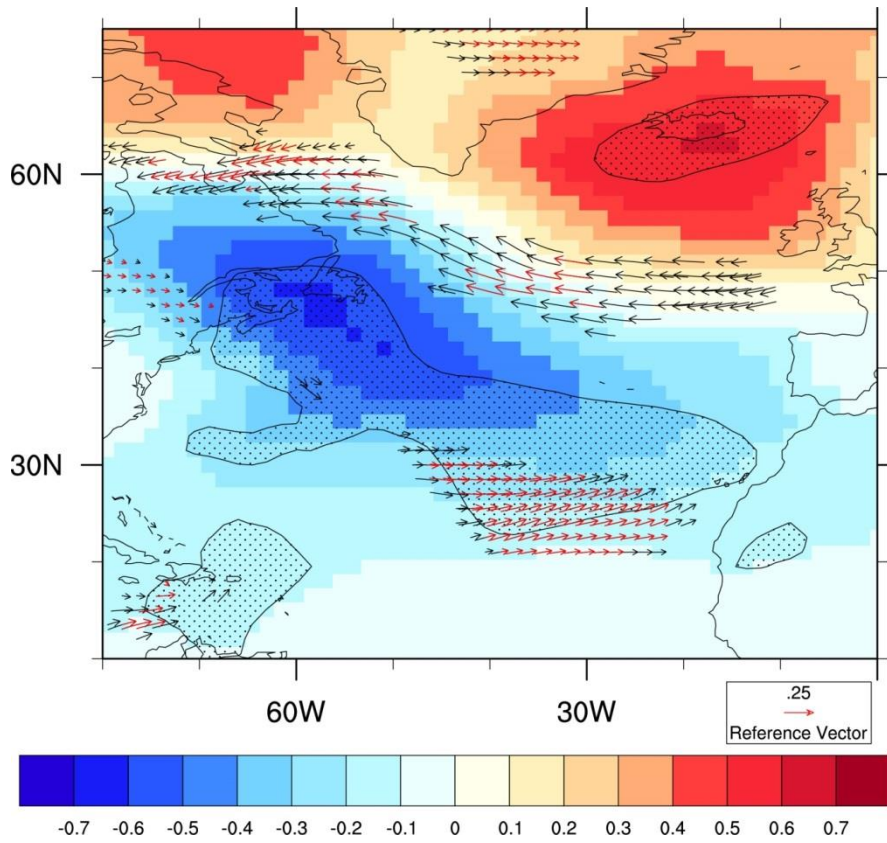
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897 Figure 16: 1998-2003 (3-8 year lead time) average of sea-level pressure (hPa, color) and surface wind speed anomaly (m s⁻¹,
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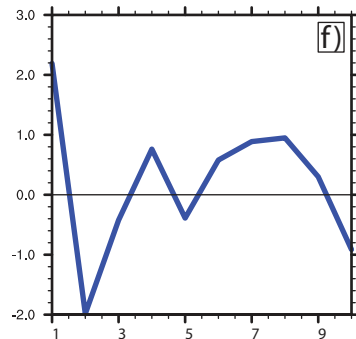
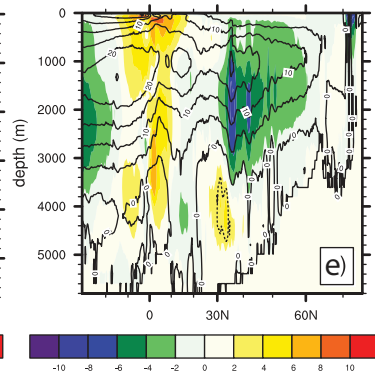
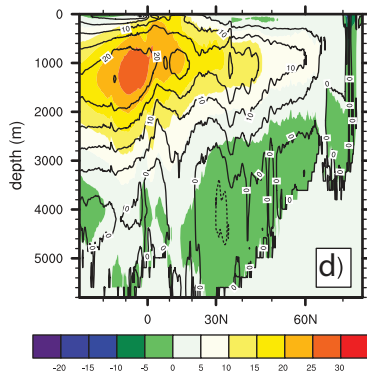
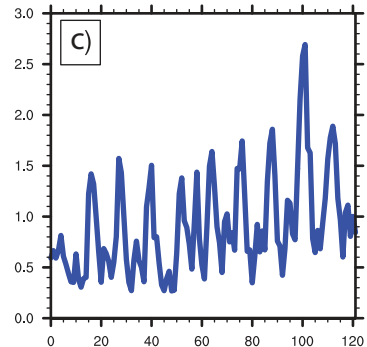
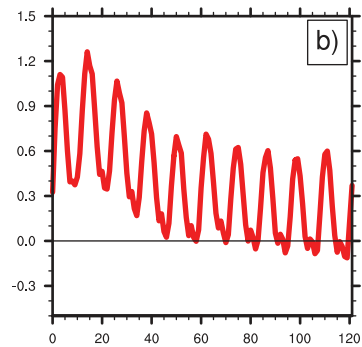
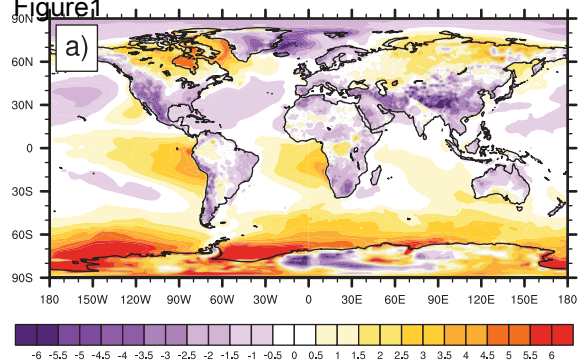
Figure 1

Figure 2

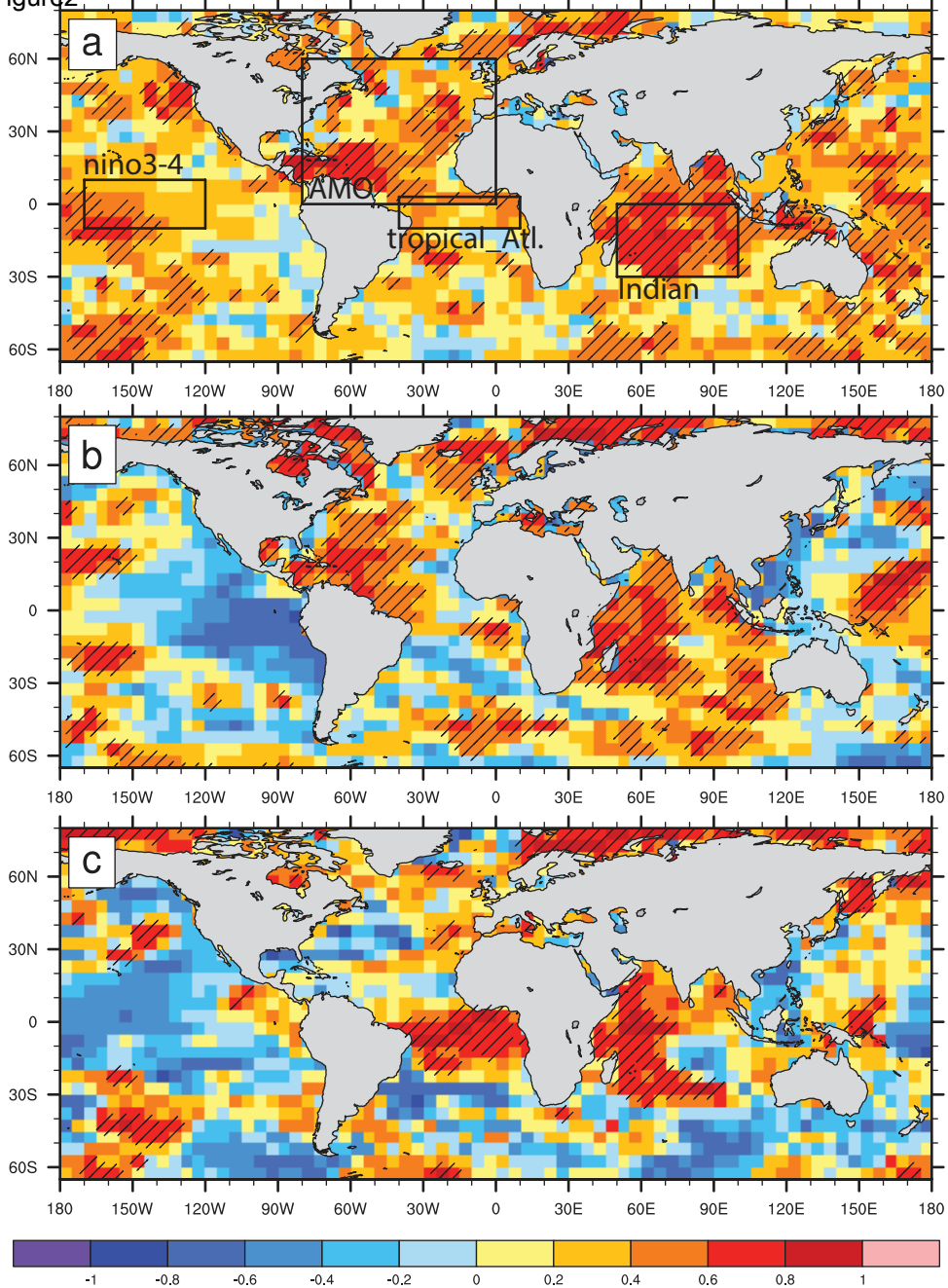


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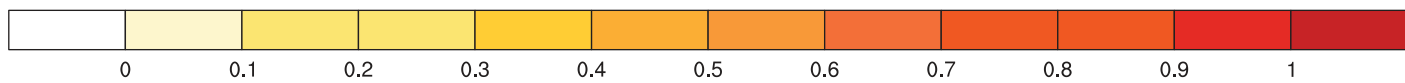
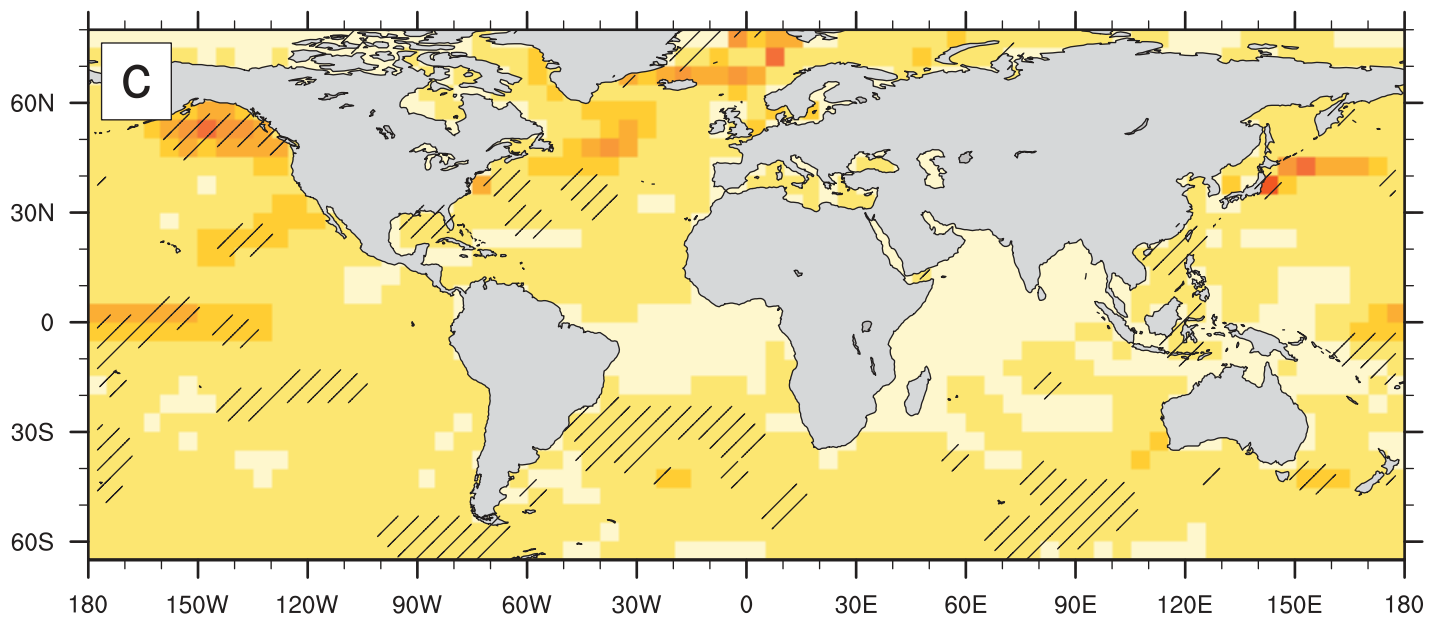
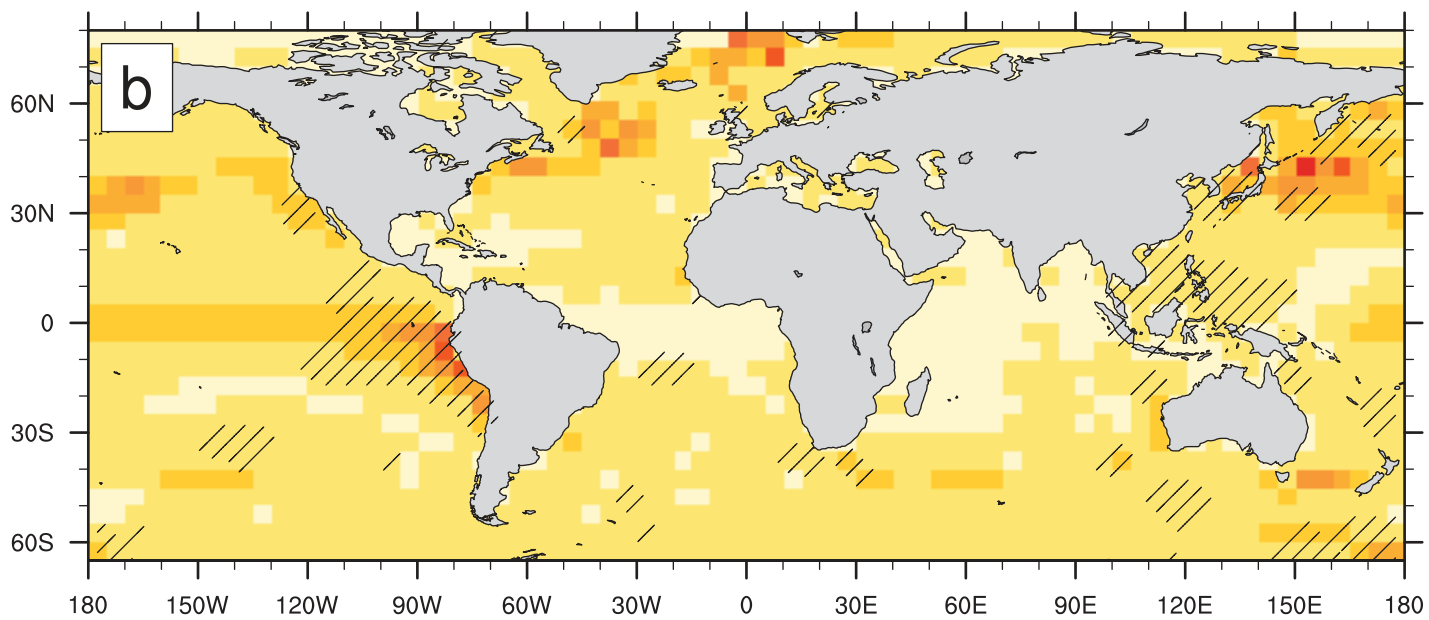
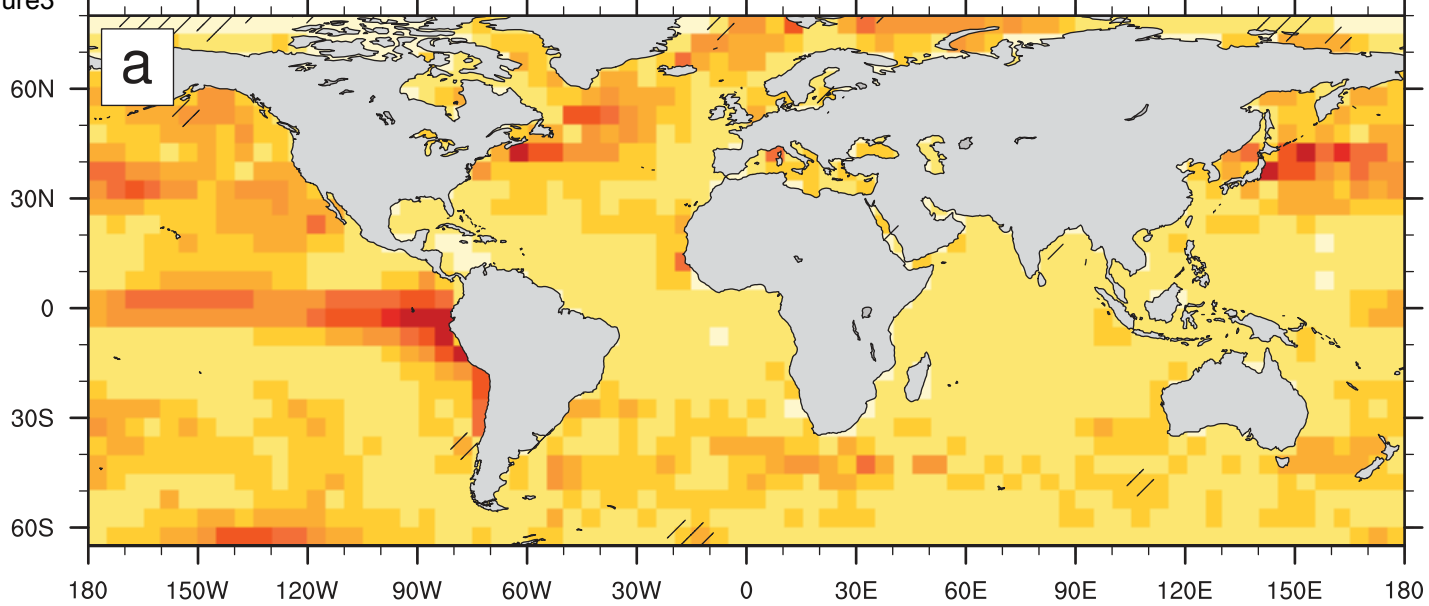
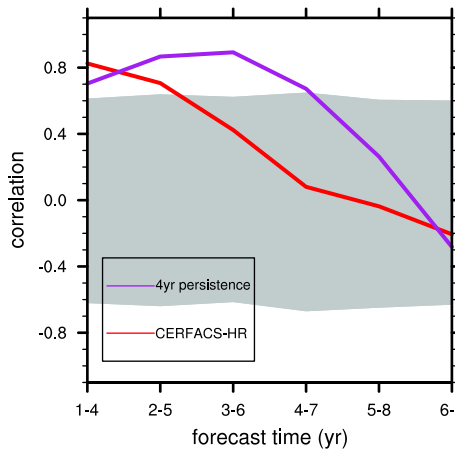
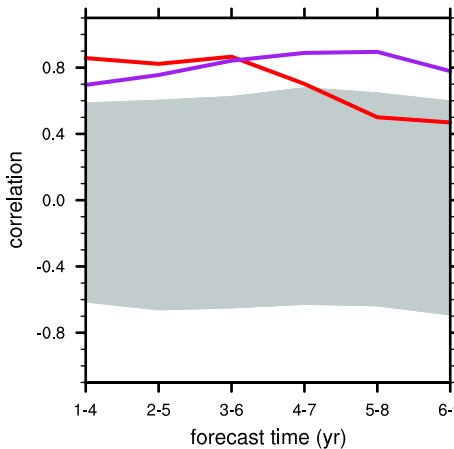


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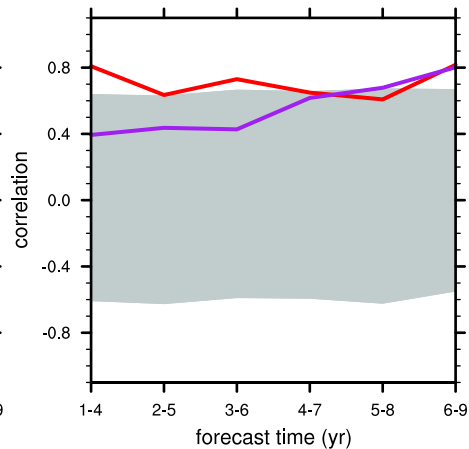
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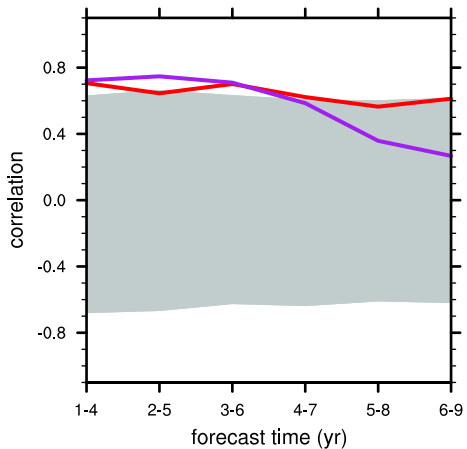
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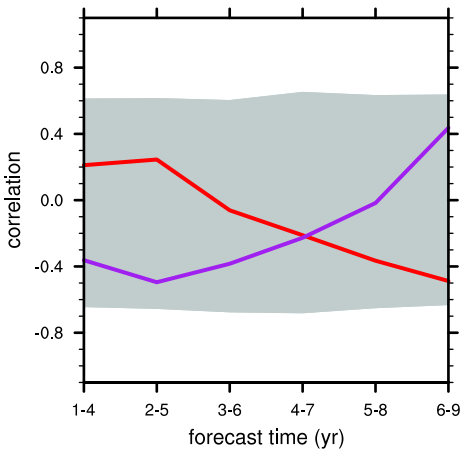
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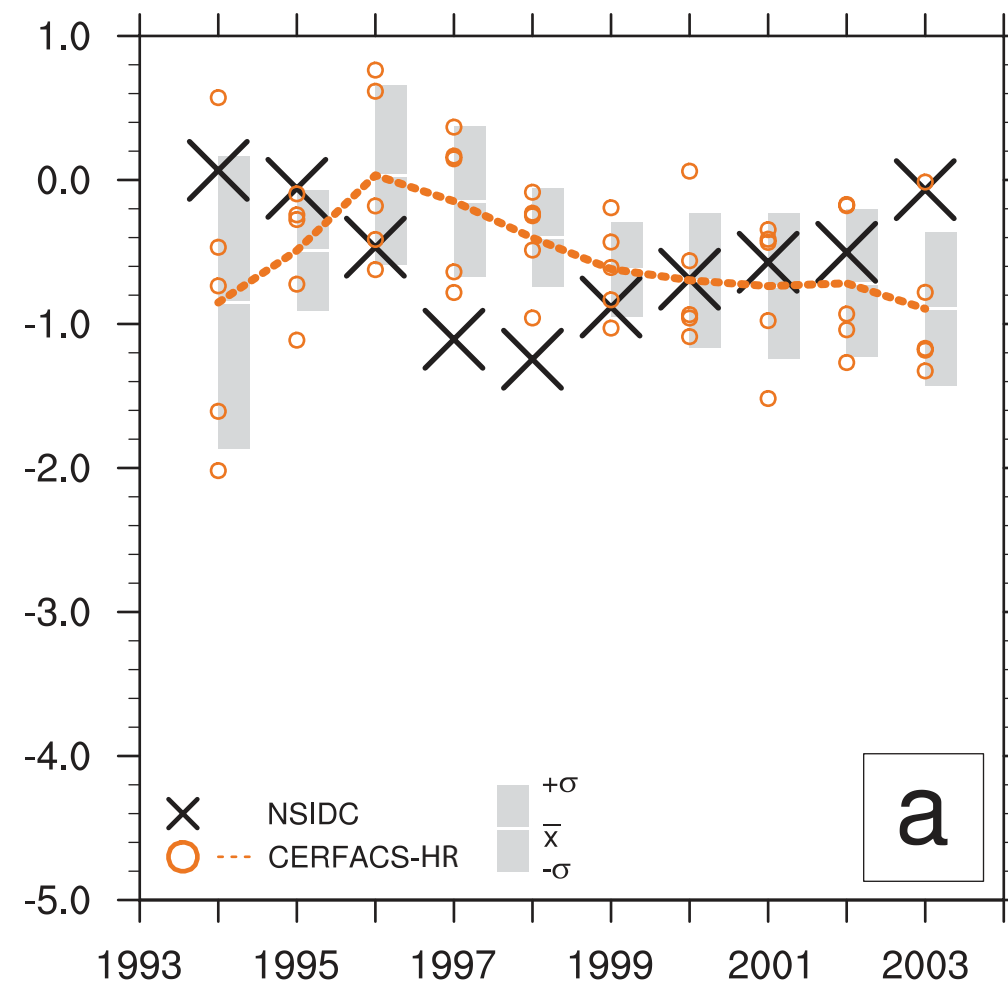
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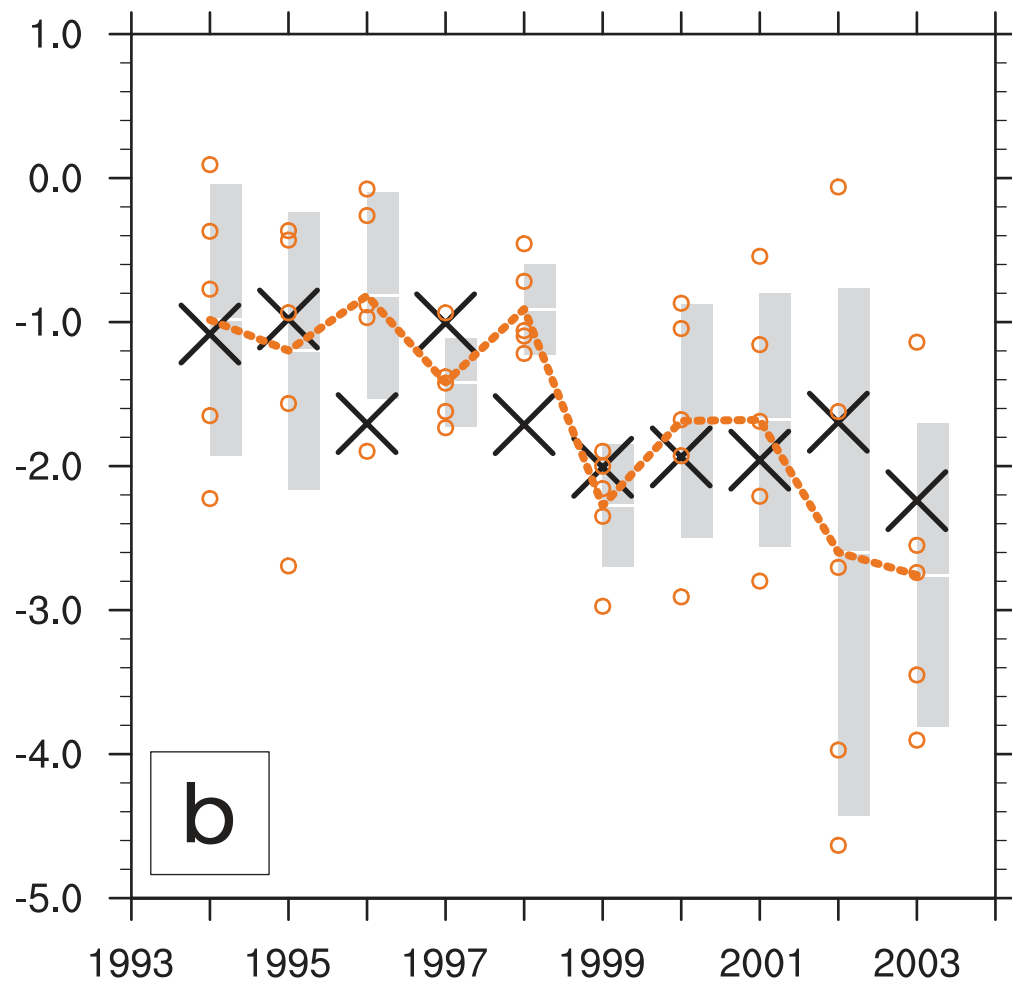
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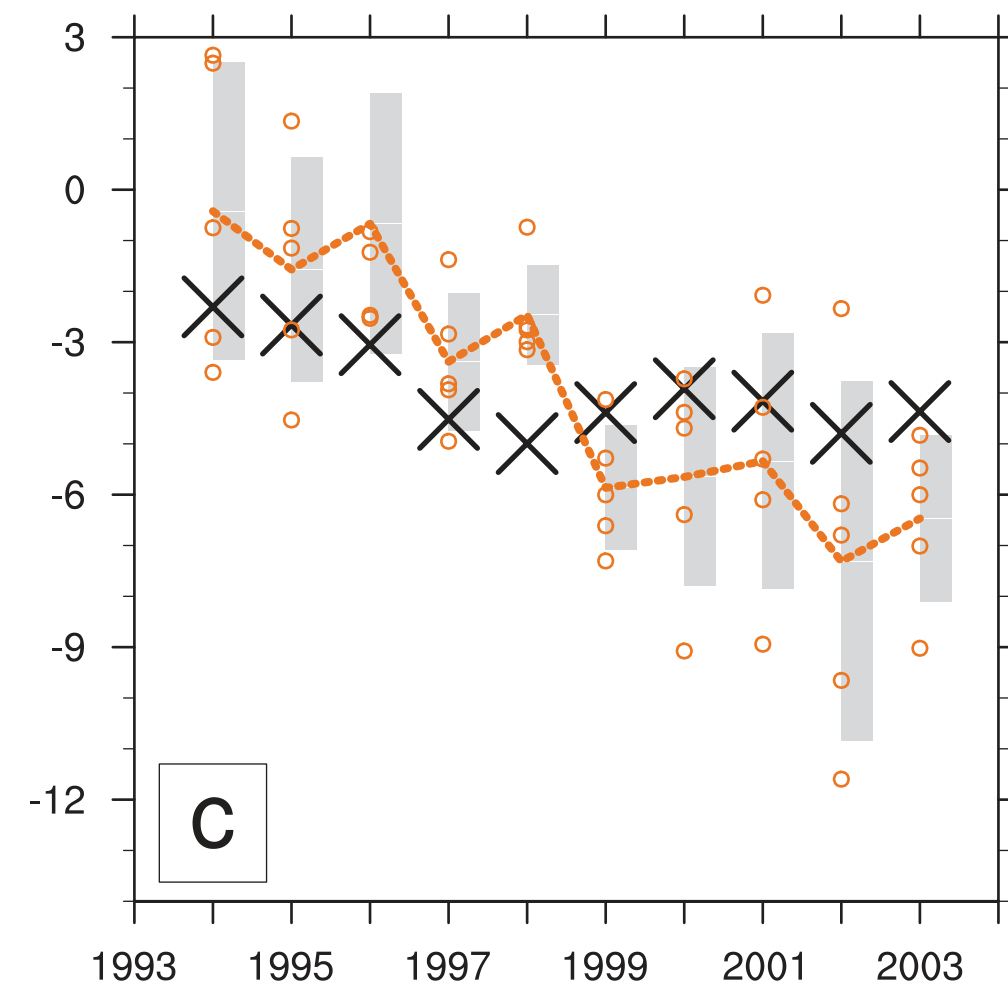
Winter extent



Summer extent



Winter volume



Summer volume

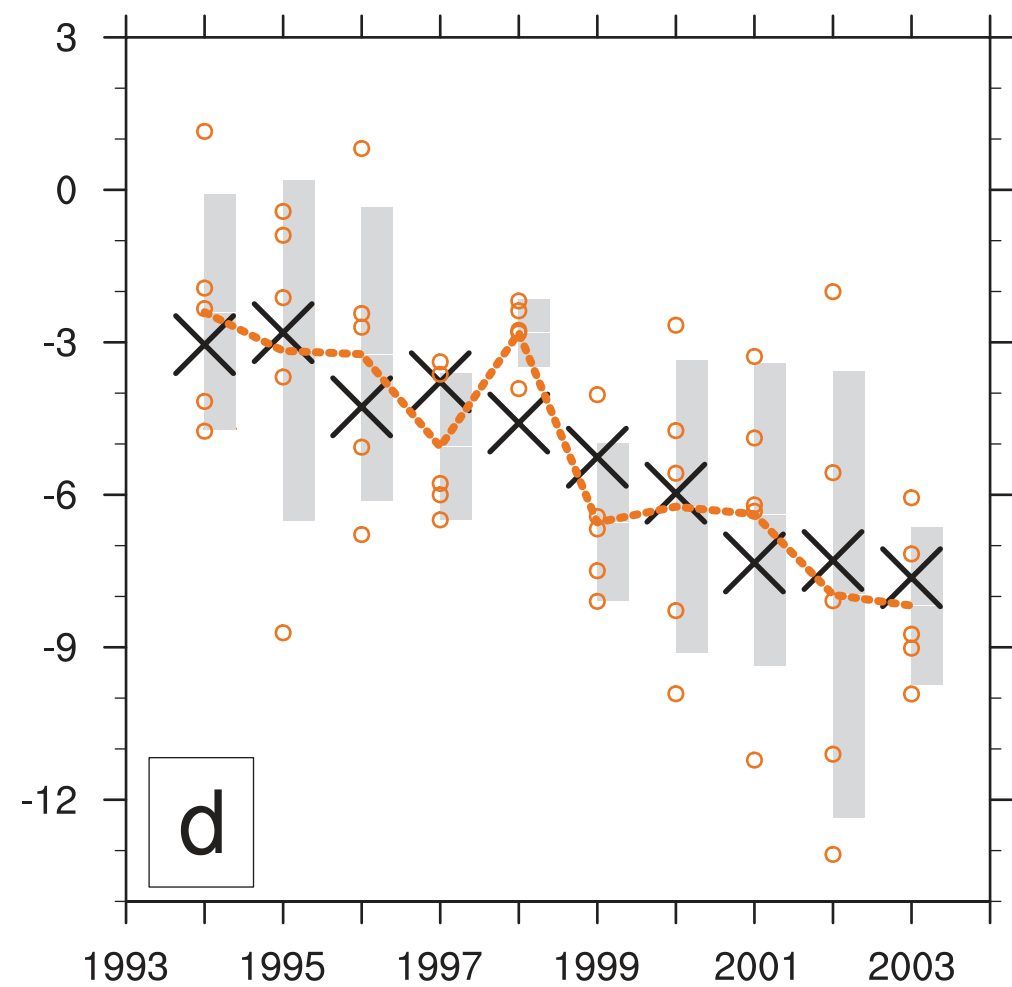


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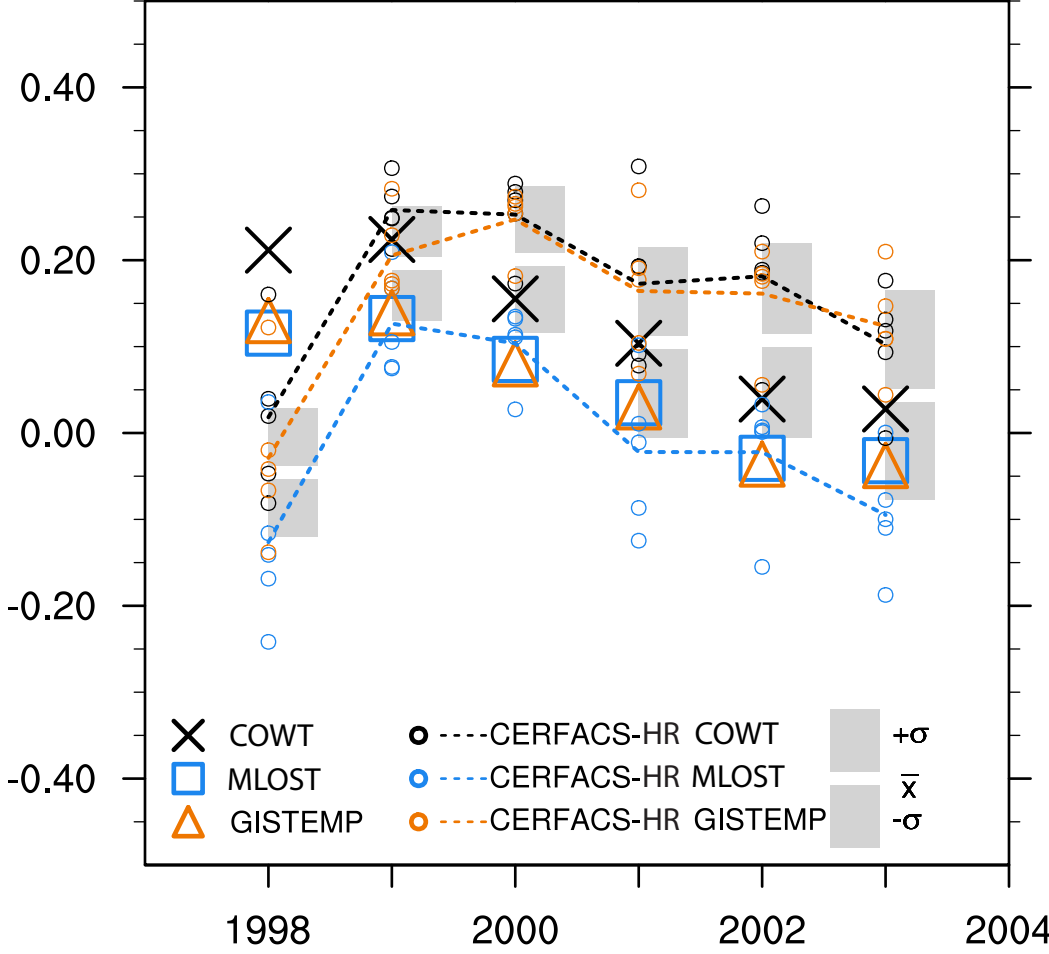


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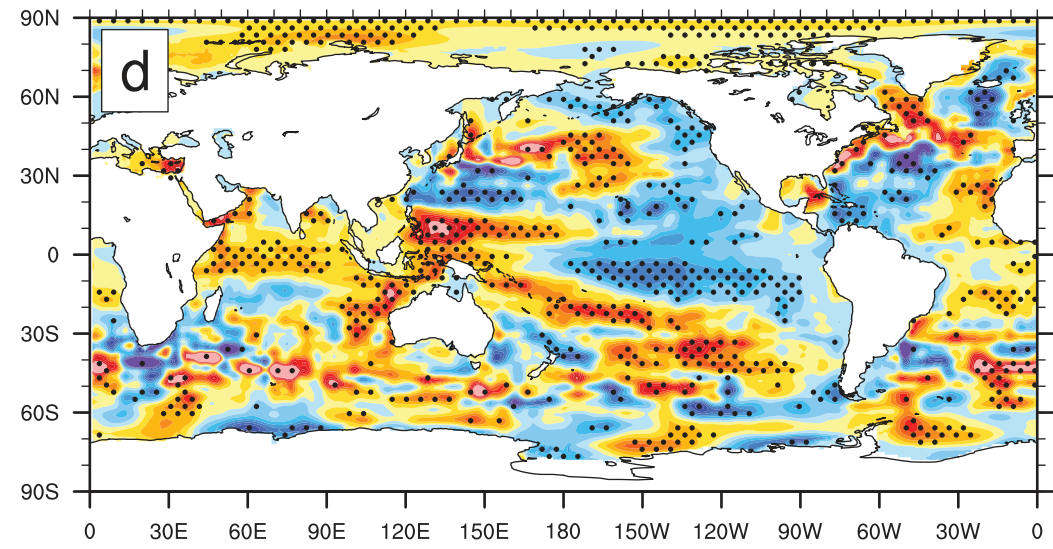
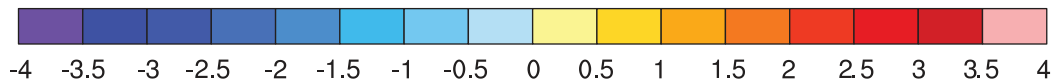
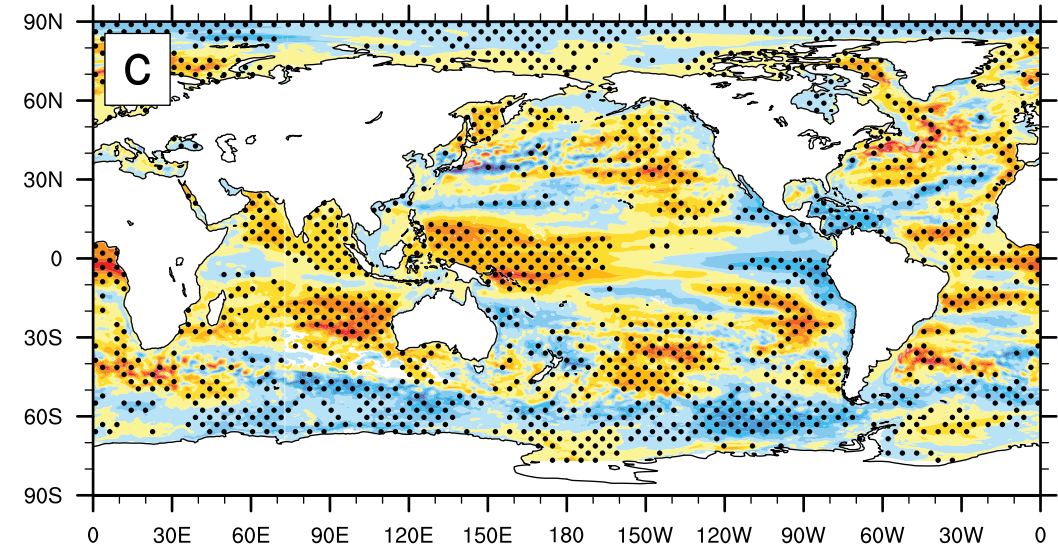
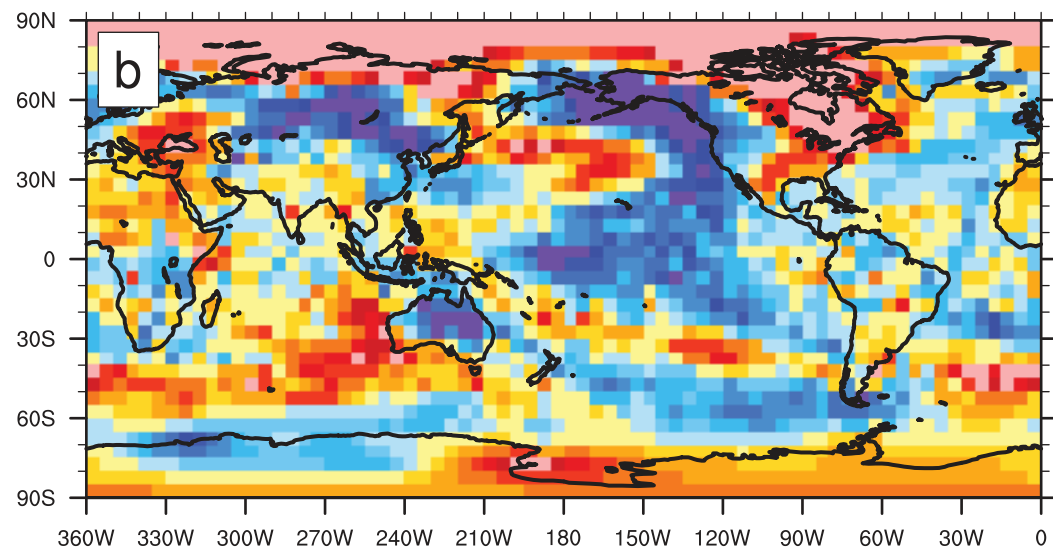
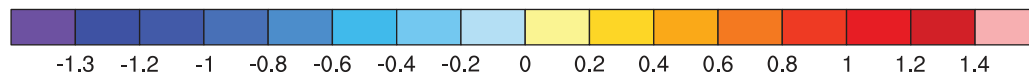
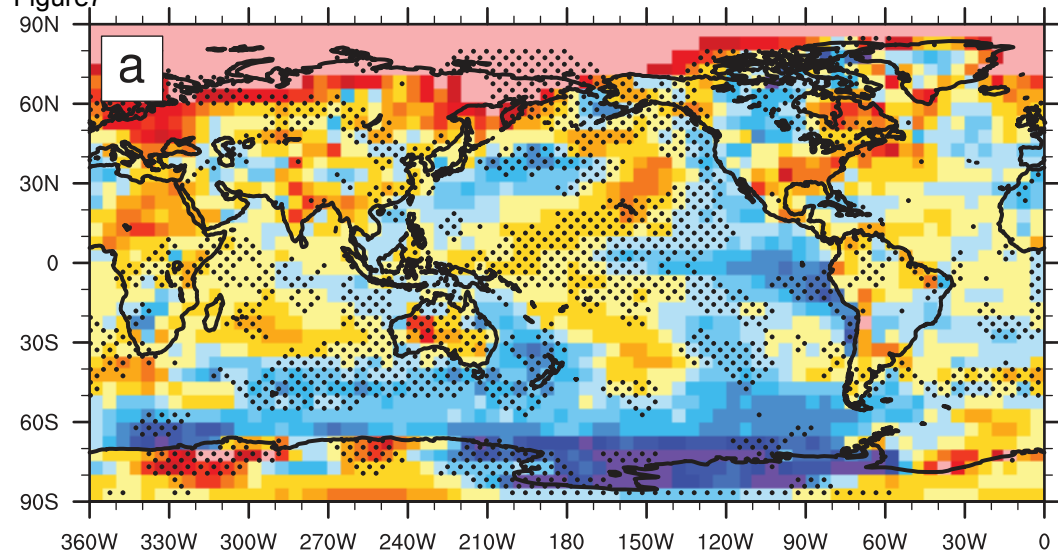


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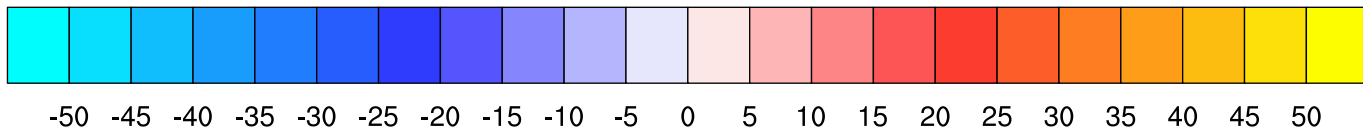
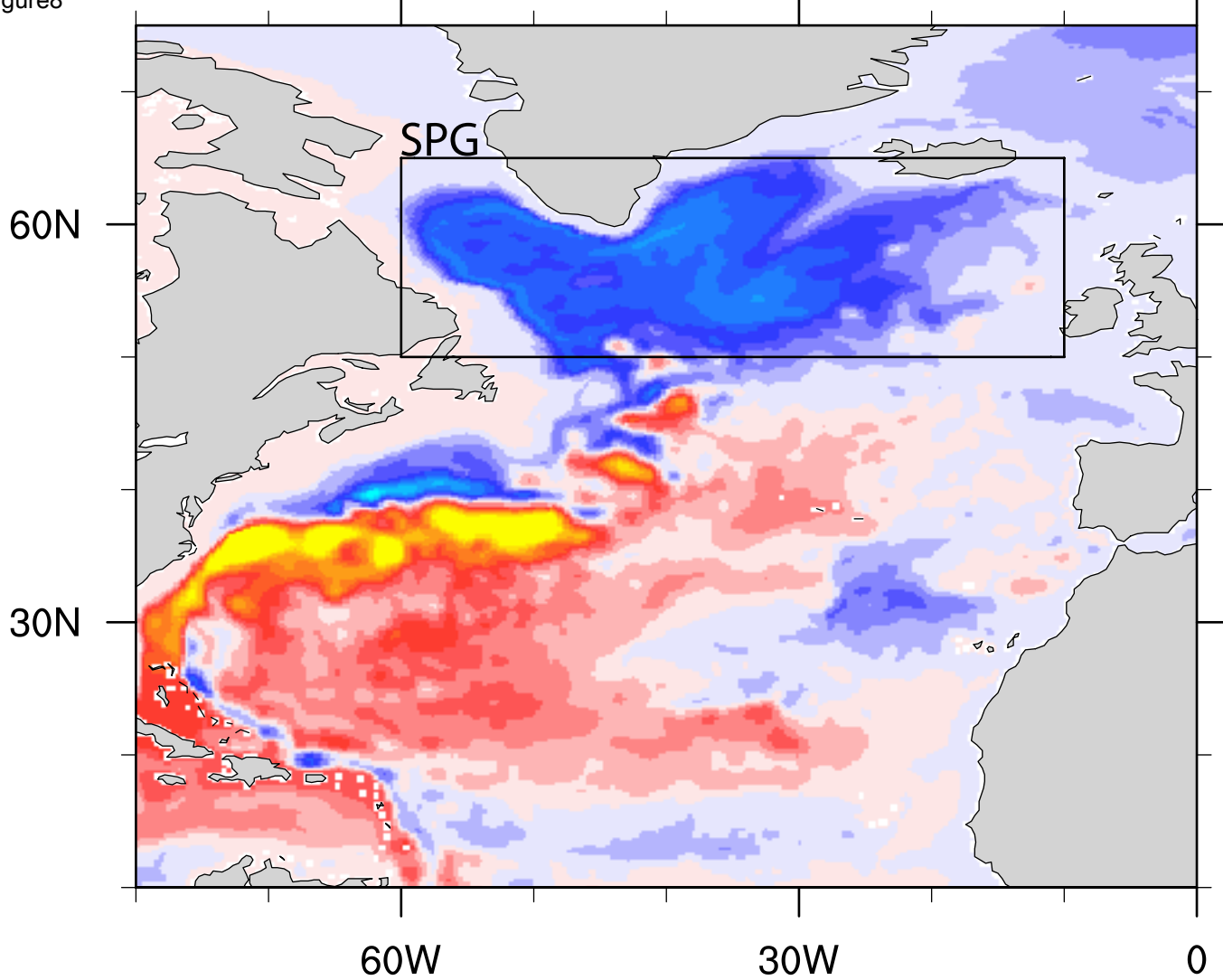


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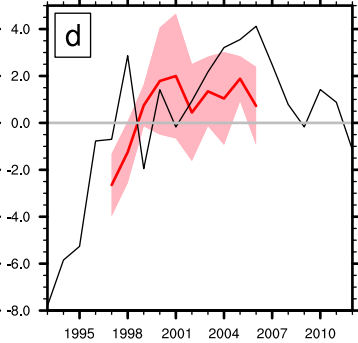
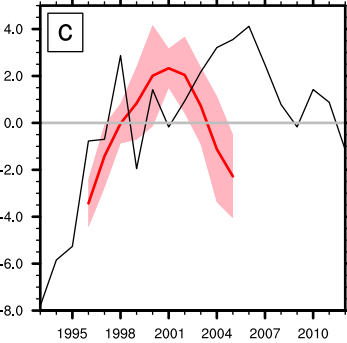
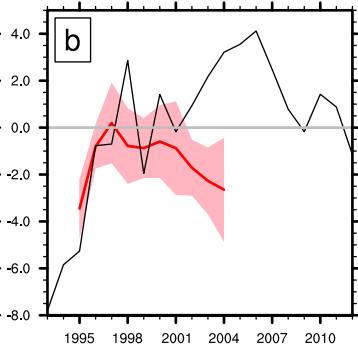
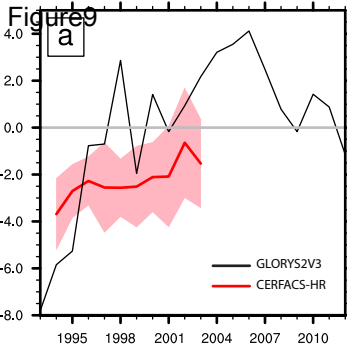
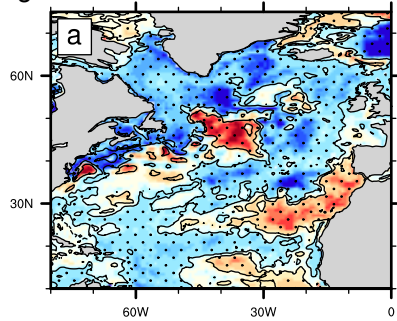
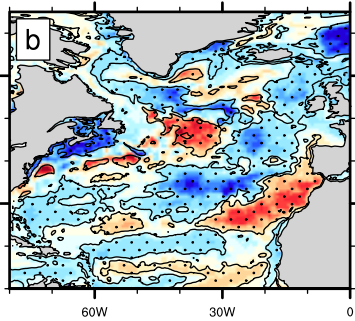


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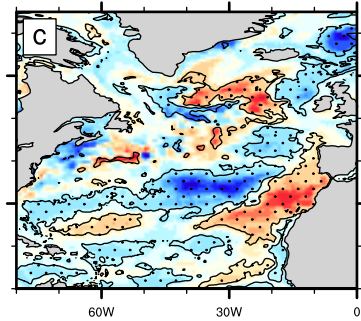
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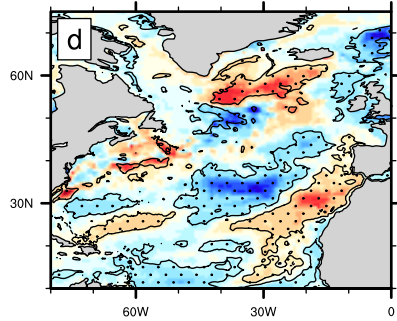
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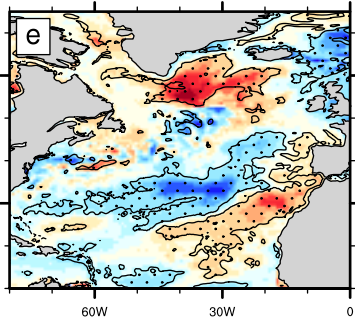
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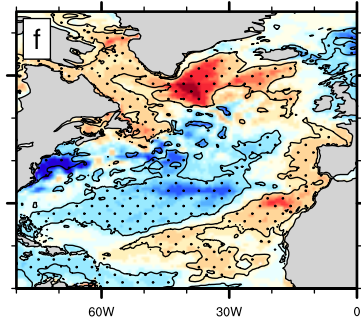
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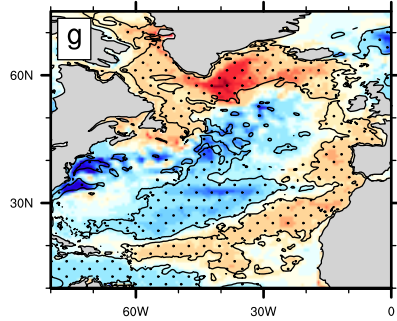
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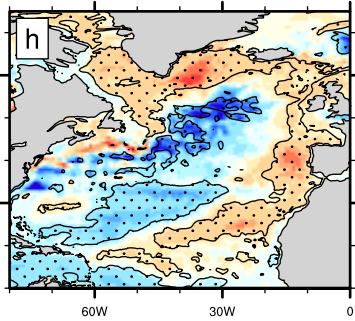
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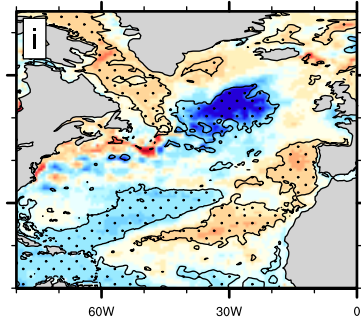
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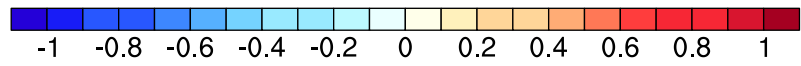
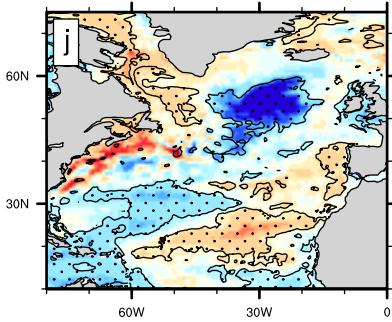
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2004



2005



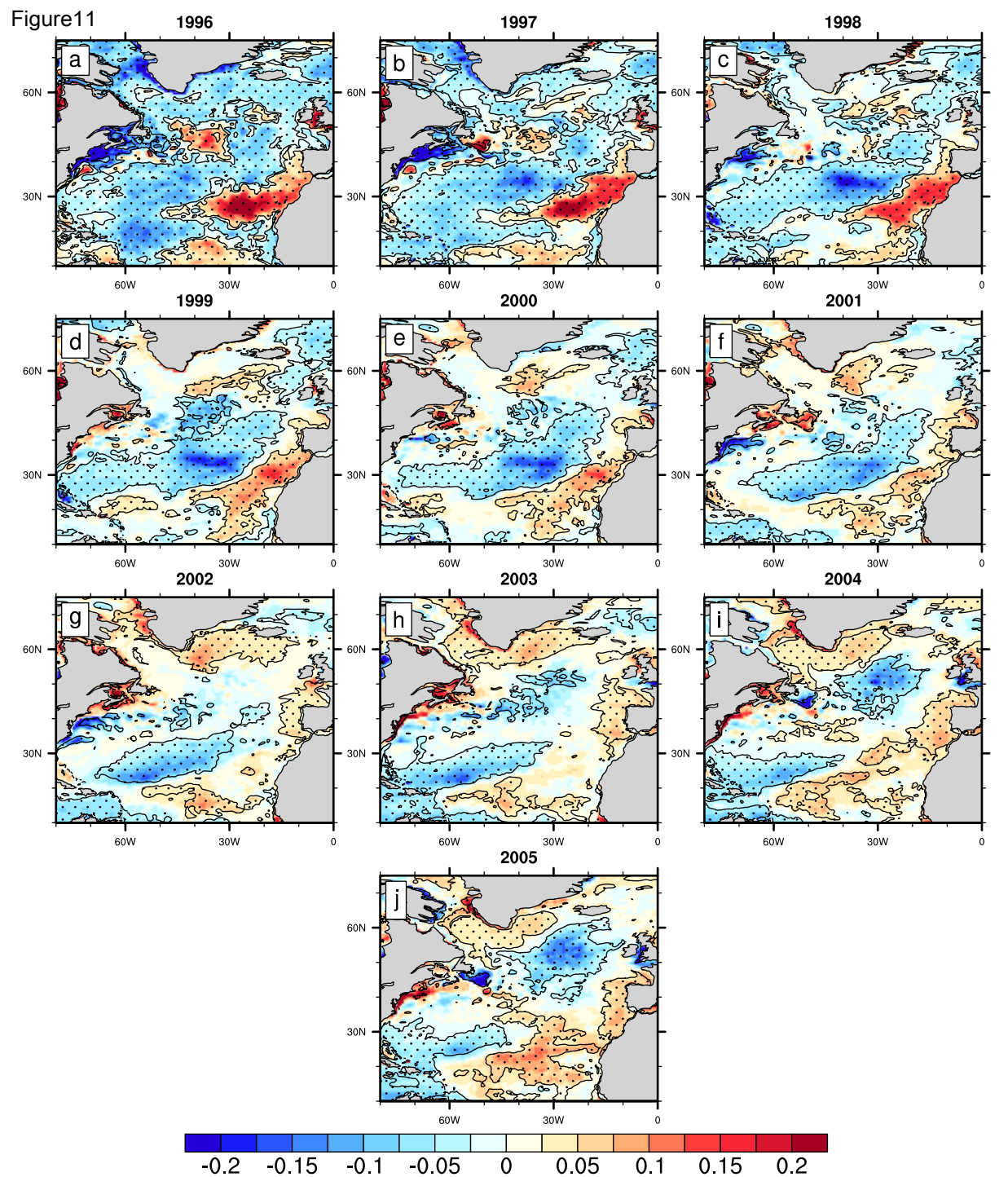
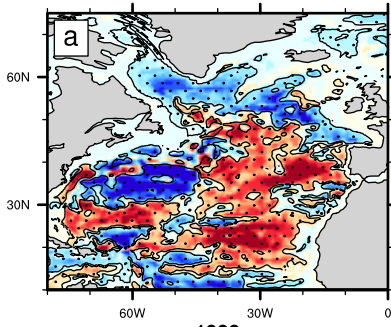
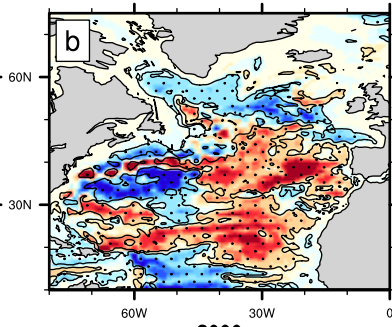


Figure 12

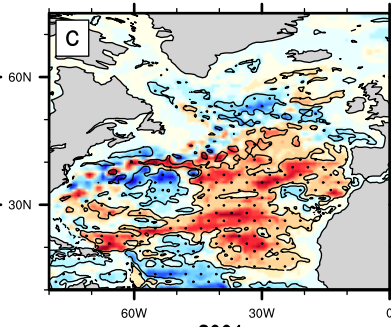
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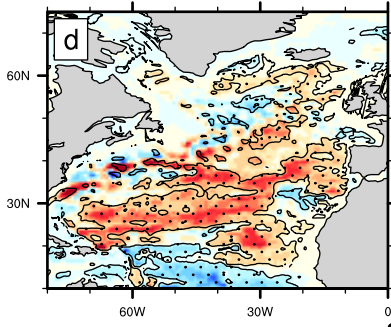
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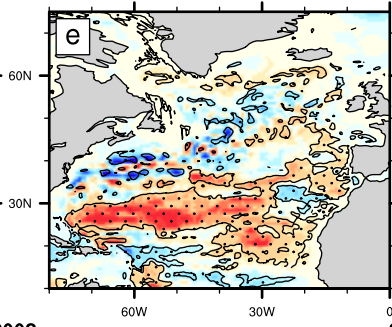
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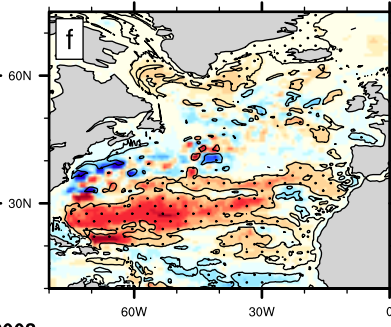
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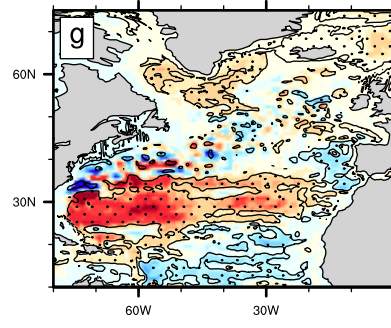
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2001



2002



2003

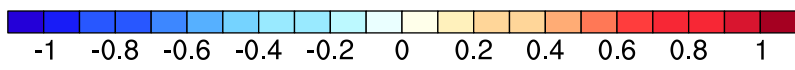
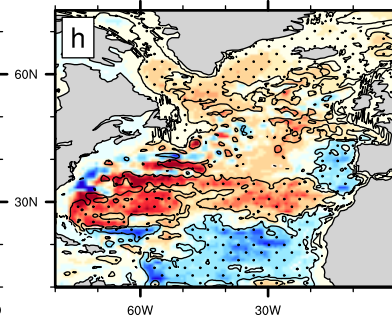


Figure 13

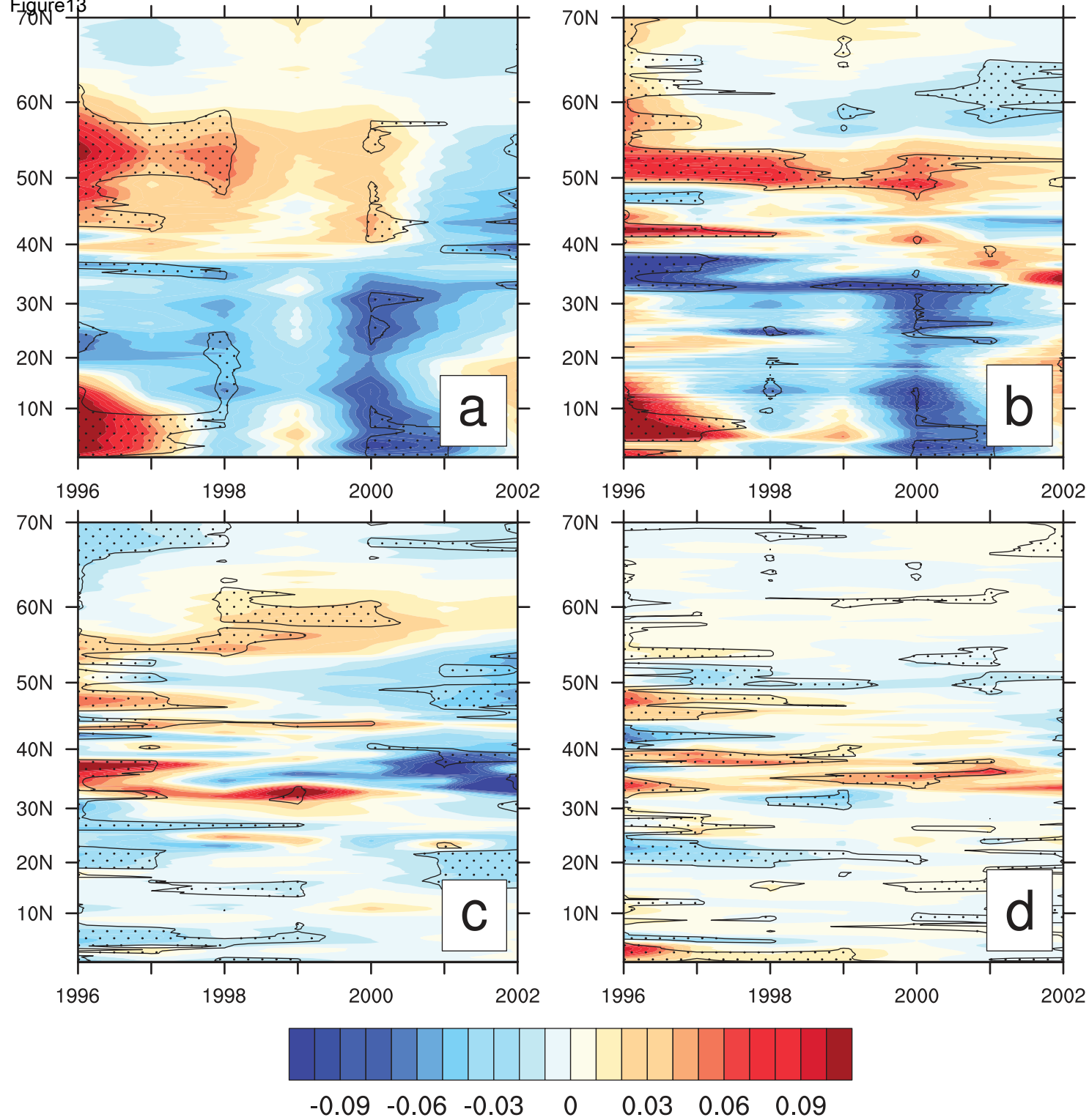
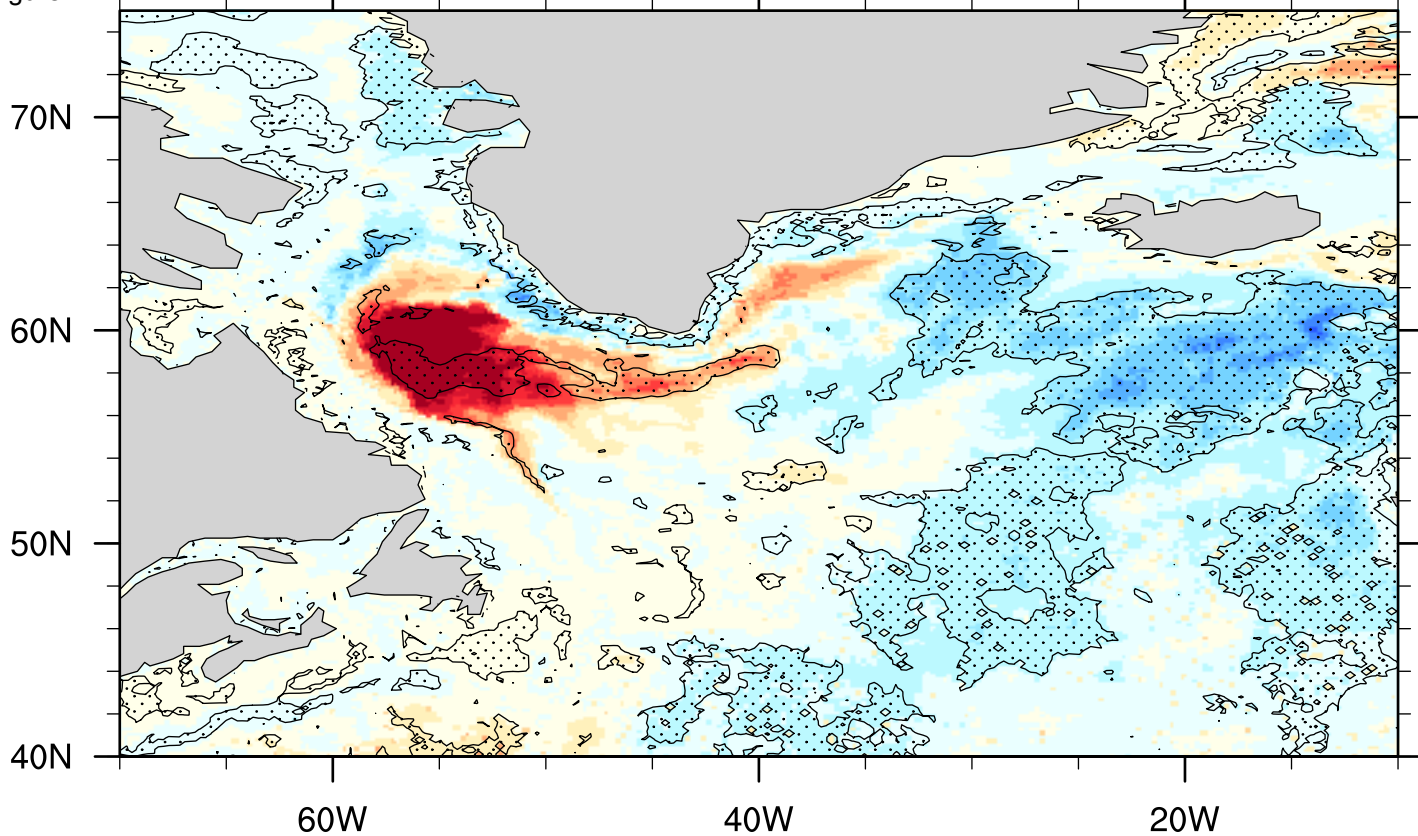


Figure14



60W

40W

20W

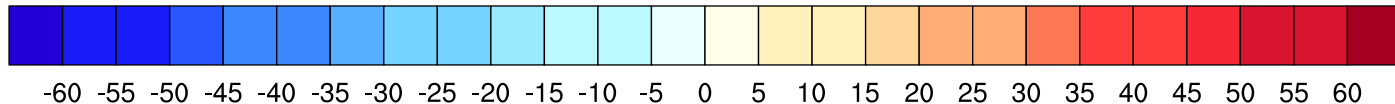


Figure 15

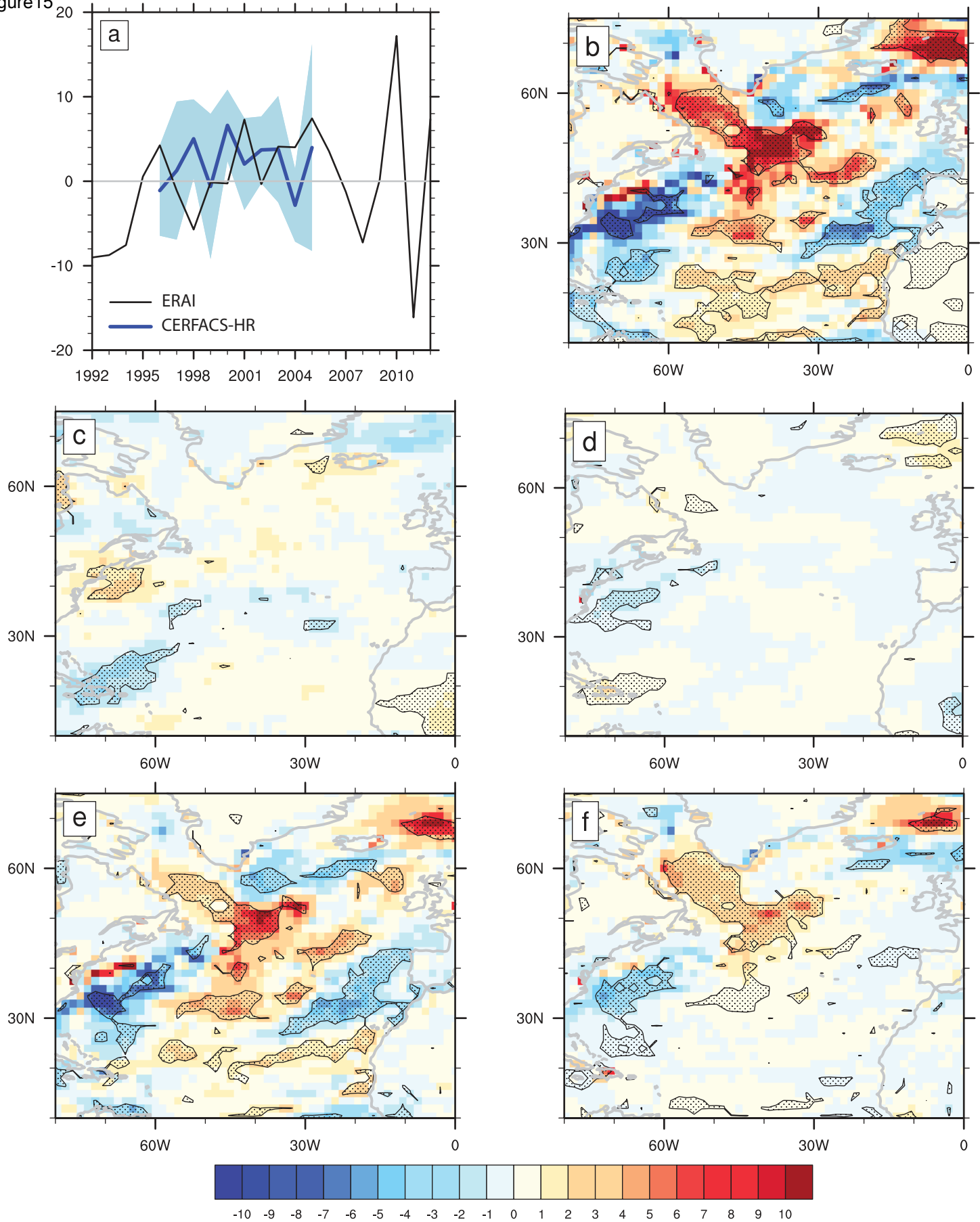
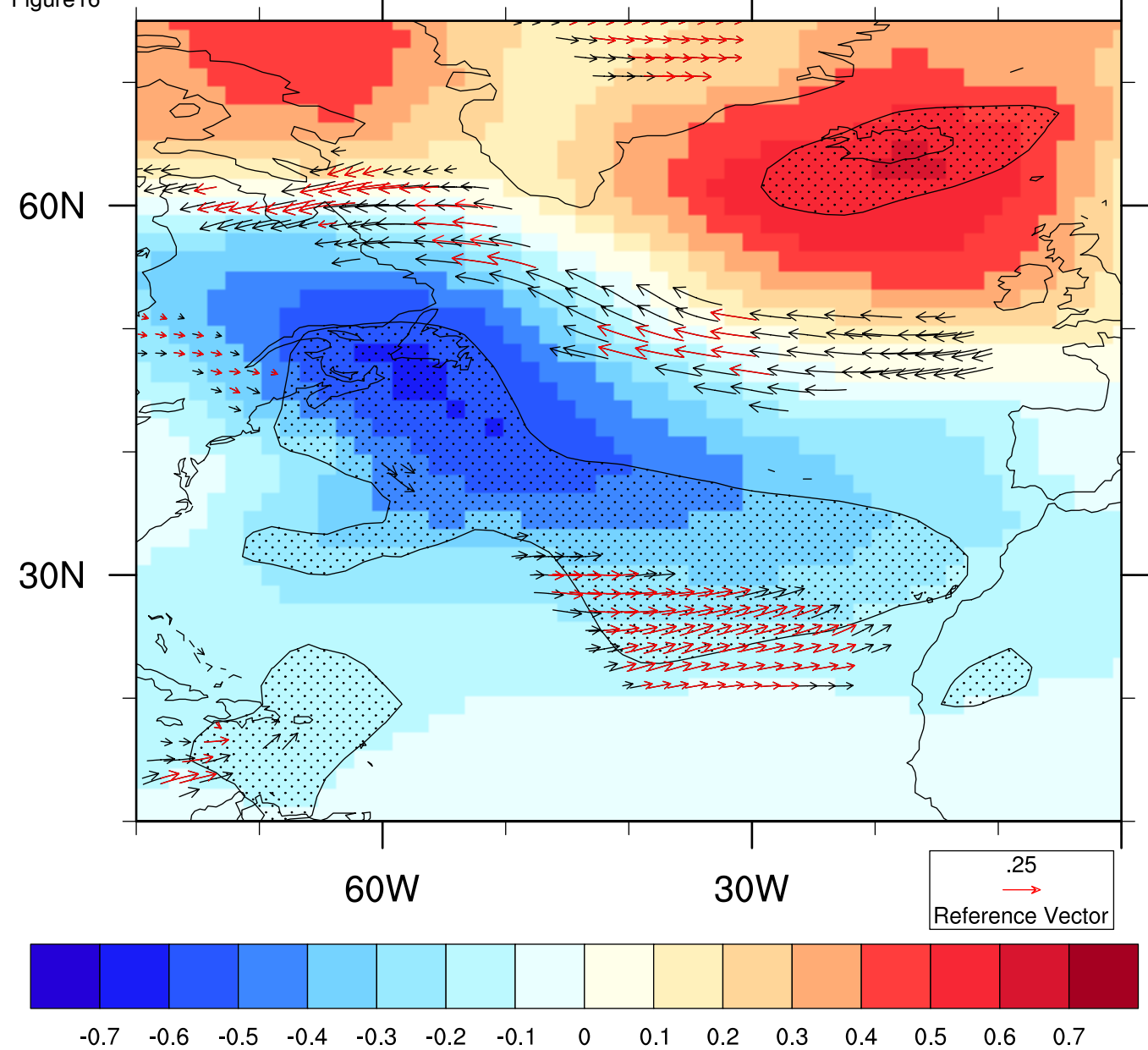
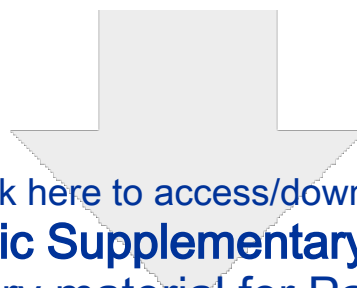


Figure16





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