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Climate Simulation and Change in the Brazilian Climate Model

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Abstract

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The response of the global climate system to atmospheric CO₂ concentration increase in time is scrutinized employing the Brazilian Climate Model (BESM-OA2.3). Through the achievement of over two thousand years of coupled model integrations in ensemble mode, it is shown that the model simulates the signal of recent changes of global climate trends, depicting a steady atmospheric and oceanic temperature increase and corresponding marine ice retreat. The model simulations encompass the time period from 1960 to 2105, following the Coupled Model Intercomparison Project 5 (CMIP5) protocol. Notwithstanding the accurate reproduction of large scale ocean-atmosphere coupled phenomena, like the ENSO phenomena over the Equatorial Pacific and the interhemispheric gradient mode over the Tropical Atlantic, the BESM-OA2.3 coupled model shows systematic errors on sea surface temperature and precipitation which resemble those of other global coupled climate models. Yet, the simulations demonstrate the model's potential to contribute to the international efforts on global climate change research, sparking interest in global climate change research within the Brazilian climate modeling community, constituting a building block of the Brazilian Framework for Global Climate Change Research.

33 1. Introduction

34 The conclusions of the Intergovernmental Panel on Climate Change (IPCC) Assessment
35 Report No. 4 (IPCC 2007) are unequivocal regarding the human interference on the
36 global climate via the buildup of Green House Gases (GHG) into the atmosphere. The
37 steady increase of GHG concentration in the atmosphere, CO₂ being the most notable of
38 them, has led the scientific community and Governments alike to debate the issue to
39 exhaustion. Therefore, a detailed scrutiny of the influence of atmospheric GHG
40 concentration on the climate system is a must and has been delved in by several research
41 groups worldwide. Such scrutiny is presently applied to matters ranging from the
42 development of state-of-the-art coupled climate models up to complex earth system
43 models (ESM) which incorporate the complexity of the many components of the earth
44 system. Examples of such complex system models are those developed by the largest
45 climate research centers in the world, such as the National Center for Atmospheric
46 Research – NCAR’s CCSM4 (Bates et al. 2012; Grodsky et al. 2012), Geophysical Fluid
47 Dynamics Laboratory – GFDL’s CM2.3 (Delworth et al. 2012), and the Hadley Centre’s
48 HadGEM-ES. Yet, other climate centers around the world are also developing their own
49 climate models, either by adapting a particular state-of-the-art global ESM, or by
50 combining ocean, atmosphere, and surface component models from various sources;
51 creating their “own” global climate model. This is the path chosen by Brazil, which has
52 decided to develop its ESM, whose initial steps are underway and are detailed in this
53 article.

54 The effort to develop Brazil’s ESM is part of a national coordinated effort to build
55 a multidisciplinary research framework geared at understanding the causes of global

56 climate change, its effects, and impacts on society. In this coordinated effort, the
57 Brazilian National Institute for Space Research (INPE) is leading the construction of a
58 global climate model capable of predicting the consequences of the buildup of GHG in
59 the atmosphere. This article describes the first version of this model, here represented by
60 the global ocean-atmosphere coupled model, named Brazilian Earth System Model –
61 Ocean-Atmosphere version 2.3 (BESM-OA2.3). For this article, we have followed the
62 criteria for participation in the Coupled Models Intercomparison Project 5 (CMIP5)
63 protocol (Taylor et al. 2009), aiming at simulating the behavior of the coupled ocean-
64 atmosphere system on decadal time scales under varying GHG concentrations in the
65 atmosphere.

66 In so doing, we have developed a model which is capable of performing seasonal
67 to decadal predictions, moving in the direction of a seamless prediction system (Palmer et
68 al. 2008; Shukla 2009), as other centers have already embraced worldwide. With this goal
69 in mind, this article presents the ability of BESM-OA2.3 to simulate global trends and
70 global climate variability related to historical variations of GHG concentrations in the
71 atmosphere. Such exercise constitutes a benchmark of the model's capability to be used
72 for studies involving more complex phenomena in the future.

73 BESM-OA2.3 is an evolution of previous versions of the Brazilian Center for
74 Weather Forecasting and Climate Studies (CPTEC) coupled ocean-atmosphere model,
75 which have been used to study the impacts of Amazon deforestation on climate (Nobre et
76 al. 2009), the dynamics of the South Atlantic Convergence Zone (SACZ) over cold
77 waters (Chaves and Nobre 2004; Nobre et al. 2012), the importance of ocean-atmosphere
78 coupling to represent the Atlantic thermocline depth (Siqueira and Nobre 2006), as well

79 as for regional seasonal predictions with the Eta regional model over South America
80 (Pilotto et al. 2012). All these studies highlighted the importance of ocean–atmosphere
81 coupling to better predict the phenomena of study.

82 The challenge in this article is to demonstrate that the coupled ocean-atmosphere
83 model developed at INPE is suitable to simulate climate phenomena along the last few
84 decades and into the near future, for the period from 1960 to 2035. This article shows that
85 the model is capable of reproducing global trends of ocean heat content and top- of-
86 atmosphere radiation balance, as well as important modes of climate variability. The
87 effects of observed variations of atmospheric CO₂ during the period of study and its
88 projection into the near future are also investigated. Regardless of the simplicity of
89 BESM-OA2.3 compared to full ESMs (*e.g.*, Muñoz et al. 2012; Delworth et al. 2012), its
90 results are comparable to other global climate models.

91

92 2. Model description

93 The BESM-OA2.3 coupled climate model used in this research is constituted by the
94 CPTEC/INPE Atmospheric General Circulation Model (AGCM) coupled to GFDL’s
95 MOM4p1 Oceanic General Circulation Model (OGCM) via GFDL’s Flexible Modular
96 System. The component models and coupling are described as follows.

97

98 *Atmosphere model*

99 The CPTEC/INPE AGCM has been constantly modified during the last few years. The
100 implementation of two new dynamic cores (Eulerian and Semi-Lagrangian) and the

101 algorithm for the use of a reduced grid are now employed operationally at CPTEC. For
102 the present work, the Eulerian core and reduced grid have been used. The options for
103 physical parameterizations include the Simplified Simple Biosphere (SSiB) module
104 surface (Xue et al. 1991), turbulence level 2 module (Mellor and Yamada 1982), gravity
105 wave module (Anthes 1977), deep convection module (Arakawa and Shubert 1974; Grell
106 and Devenyi 2002), shallow convection module (Tiedtke 1984), short-wave radiation
107 CLIRAD-SW-M (Tarasova et al. 2006, Chou and Suarez 1999), longwave module
108 (Hashvardhan and Corsetti 1984; Hashvardhan et al 1987), and the model for interaction
109 of radiation with clouds (Slingo, 1987; Hou, 1990; Kinter et al. 1997). In the shortwave
110 and longwave radiation schemes the carbon dioxide is globally specified, assuming either
111 a 370 ppm fixed value (for the control integrations, labeled “physics2” for the CMIP5
112 protocol) or a time dependent function (transient integrations, “physics1”).

113 The new deep convection scheme implemented in the CPTEC–AGCM is based
114 on the convective parameterization developed by Grell and Devenyi (2002), hereafter
115 referred as Grell-ensemble (GRE) scheme. This scheme includes several alternative
116 closure assumptions to estimate the cloud base mass flux. Both the first type of closure
117 (Grell closure) and the second type (Arakawa-Schubert closure) assume a quasi-
118 equilibrium state between large-scale forcing and convection. Other closures used are the
119 Kain-Fritsch closure, in which the atmospheric stability is removed by the convection
120 within the specified convective time, moisture convergence closure, and low level
121 vertical velocity mass flux closure. The set of convection scheme ensembles used in the
122 CPTEC-AGCM includes 144 different versions of the cumulus parameterization, 16 from
123 dynamic control and 9 from static control. The details of implementation of this scheme

124 in the CPTEC-AGCM and its impact on tropical precipitation can be found in Figueroa et
125 al. (2012).

126 The exchanges of heat, moisture, and momentum between the surface and
127 atmosphere in the CPTEC-AGCM over the ocean and continents are computed
128 differently by various physical processes that define the surface fluxes. The model uses
129 an empirical relationship that approximates the theory of Monin-Obukhov similarity to
130 determine the transfer and the drag coefficient. The relationships over the ocean differ
131 from those over land. Over the oceans, the momentum, heat, and moisture fluxes are
132 parameterized considering the global aerodynamic scheme (bulk aerodynamic schemes)
133 where the fluxes are assumed to be proportional to wind speed and the surface potential
134 (difference in temperature or moisture between the ocean surface and the air near the
135 surface). The proportionality or drag coefficient is calculated by an empirically
136 determined functional fit. On the surface of the ocean the functional fit is similar to the
137 one used by Sato et al. (1989). With regard to momentum, heat, and moisture fluxes over
138 the continent, the parameterization used is more sophisticated. The surface scheme on the
139 continent considers a variety of different types of vegetation, soil, and geographical
140 formation. The Simplified Simple Biosphere (SSiB) (Xue et al. 1991) scheme used in the
141 CPTEC-AGCM, despite being a simplified model, has many physical processes that
142 represent the interactions between land surface and atmosphere.

143 The AGCM has spectral horizontal resolution truncated at triangular wave
144 number 62, which gives an equivalent grid size of 1.875 degrees, and 28 sigma levels
145 unevenly spaced in the vertical (i.e. T062L28).

146

147 *Ocean model*

148 The ocean model used in this work is the “Modular Ocean Model - MOM4p1 (Griffies
149 2009) from GFDL, which includes the Sea Ice Simulator – SIS built-in ice model
150 (Winton 2000). The horizontal grid resolution is set to 1° in the longitudinal direction,
151 and in the latitudinal direction the grid spacing is $1/4^\circ$ in the tropical region (10°S – 10°N)
152 decreasing uniformly to 1° at 45° and to 2° at 90° of both hemispheres. For the vertical
153 axis, 50 levels are adopted with a 10-m resolution in the upper 220 m, increasing
154 gradually to about 370 m of grid spacing at deeper regions. The variant of the ocean
155 model not coupled to an atmospheric model, here referred to as “OGCM” or “forced”
156 model, is forced by winds, temperature, specific humidity at 10 m, sea level pressure,
157 radiative fluxes, sea ice cover and thickness, and river discharges. The OGCM spinup
158 was done in a manner similar to that used by Nobre et al. (2012): starting from zero
159 currents velocities and Levitus (1982) T-S structure of the world ocean, running the
160 forced model for 13 years with climatological atmospheric forcing fields, followed by an
161 additional 58 years (1950–2007) forced by interannual fields from Large and Yeager
162 (2009); except for river discharges and the sea ice variables, which were kept
163 climatological in both cases. During the forced model run, the dynamic and
164 thermodynamic structure of the ocean was saved at regular periods of time in order to be
165 used as initial states of the ocean for the coupled model experiments.

166 The model settings are similar to the COREs experiment settings (Griffies and
167 coauthors 2009) apart from modifications made to obtain better results for our forced and
168 coupled experiments at the chosen ocean grid, such as changing the smoothing of the sea
169 surface height (“laplacian” instead of “biharmonic”), disabling the “mixing downslope”

170 module (which treats the mixing of tracer between dense shallow parcels and deeper
171 parcels downslope), disabling the tidal forcing module, enabling the “convect” module
172 (which vertically adjusts gravitationally unstable columns of ocean fluid), changing
173 parameters of the frazil module (which computes the heating of seawater due to frazil ice
174 formation), disabling the restoring masks, as well as other restoring options, with the
175 exception of the salinity restoring which is kept on with a time scale of 60 days. The
176 turbulent flux calculation described by Large and Yeager (2009) is used in the forced
177 model in order to compute the surface drag coefficients.

178 The coupled version of the ocean model has additional changes in relation to the
179 forced model. The background vertical diffusivity was defined by the Bryan Lewis
180 scheme (Bryan and Lewis, 1979), zonally constant, but varying both meridionally and
181 vertically. Between 10°N and 10°S, the Bryan Lewis diffusion coefficient span from
182 $2 \times 10^{-5} \text{ m}^2/\text{s}$ near the surface to $7.8 \times 10^{-5} \text{ m}^2/\text{s}$ at the bottom. Such values increase linearly
183 poleward to compensate for the coarser grid, measuring $2.3 \times 10^{-5} \text{ m}^2/\text{s}$ at the surface and
184 $8.7 \times 10^{-5} \text{ m}^2/\text{s}$ at the bottom near the poles. The consequent near surface eddy kinetic
185 energy is coherent with the observations (not shown), considering the limitations of the
186 simulations setup. The BESM-OA2.3 simulations develop up to $1.2 \text{ cm}^2/\text{s}^2$ average eddy
187 kinetic energy over a few degrees around the equator. This value rapidly decays poleward
188 due to the gradually coarser grid resolution near the poles. Moreover, in the coupled
189 model all restoring options including the salinity restoring are disabled.

190

191 *Coupling strategy*

192 The coupled model uses GFDL's Flexible Modeling System (FMS) coupler to couple the
193 CPTEC-AGCM to GFDL MOM 4p1 (Griffies 2009) and SIS (Winton 2000). The main
194 characteristics of the CPTEC-AGCM are described in the previous subsection. The
195 original configuration of MOM4p1 is similar to that used in the COREs release of the
196 model for the forced experiments using Large and Yeager (2009) fields. For the coupled
197 model, all restore flags were turned off. Wind stress fields are computed, using Monin-
198 Obukhov scheme within MOM4p1, from the wind field 10 meters above the ocean
199 surface. Adjustments were done to the Monin-Obukhov boundary layer scheme, whose
200 parameters were tuned according to the wind fields output by the CPTEC-AGCM. The
201 hydrological cycle is balanced by an indirect method, as suggested in Griffies (2010).
202 The AGCM receives the following two fields from the coupler: sea surface temperature
203 (SST) and ocean albedo from ocean and sea ice models at an hourly rate (coupling time
204 step). Adjustments were also made to ocean shortwave penetration parameters due to the
205 CPTEC-AGCM supply of visible and infrared short wave radiation. The coupling
206 variables supplied by the AGCM are: fresh water (liquid and solid precipitation), specific
207 humidity, heat, vertical diffusion of velocity components, momentum fluxes, and surface
208 pressure.

209

210 *Experiment design*

211 The numerical experiments design followed the CMIP5 protocols (Taylor et al. 2009) for
212 decadal, short term simulations, and near future predictions using two types of physics
213 for the coupled model differing on the concentration of atmospheric CO₂. Integrations

214 labeled "physics1" used atmospheric CO₂ concentrations derived from *in situ* air samples
215 collected at Mauna Loa Observatory, Hawaii. Historical data from Bacastow et al. (1985)
216 were regressed onto a parabolic equation defined by:

217

$$218 \text{ CO}_2 = a t^2 + b t + c$$

219

220 where the CO₂ concentration is in ppm, t is the number of fractional years since 2000,
221 $a=0.0116696$, $b=1.79984$, and $c=369$. Integrations labeled "physics2" used atmospheric
222 CO₂ concentrations fixed at 370 ppm and are referred to as control run throughout the
223 text.

224 Ensemble members were integrated for 10 years each with Initial Conditions (IC)
225 on 1–10 December of years: 1960, 1965, 1970, 1975, 1980, 1985, 1990, 1995, 2000, and
226 2005, for each physics1 and physics2. Three of these ensembles, namely, 1960, 1980, and
227 2005, were extended for an extra 20 years for each of the ten members, completing 30
228 years long integrations each, referred in the text as decadal runs, and an extra 70 years for
229 only the IC conditions of 1st December, completing 100 years long single simulations,
230 referred in the text as centennial runs, respectively. In addition, one member CMIP5
231 "1pct-co2" run was computed, assuming a 1% yearly increase in the atmospheric CO₂
232 concentration, starting at the 20th year of the 1960, physics2 run and completing 80 years.

233 The CPTEC-AGCM initial conditions for each ensemble member used the
234 National Centers for Environmental Prediction/National Center for Atmospheric
235 Research (NCEP/NCAR) reanalysis fields (Kalnay et al. 1996) for the 00:00 GMT of
236 each day from 1st to 10th of December of the chosen years. The ocean initial states were

237 chosen for the same dates from a spinup run of MOM4p1 that used prescribed
238 atmospheric fields of momentum, solar radiation, air temperature, and fresh water
239 described earlier in this section.

240 The outputs from the ocean model were generated directly at the proper
241 frequencies requested according to the CMIP5 experiment specification. The atmospheric
242 data were output at a three hourly frequency and later processed using the CMOR2
243 software (Doutriaux and Tayler 2011) to satisfy all CMIP5 output requirements.

244 3. Results

245 *Atmospheric CO₂ concentration, Ocean Heat content and TOA fluxes*

246 The residual SST between the physics1 experiment and the physics2 control run show a
247 uniform global SST warming for the 100-years long experiments (Figure 1a) showing a
248 trend of 0.054 K/decade ($p < 0.01$). This corresponds to approximately 50% of the
249 observed global SST warming in the 1980-2011 period, as estimated from averaging
250 Hadley Centre HadSST2 anomalies globally (figure not shown).

251 The propagation of the CO₂-induced extra heat into the surface layers of the
252 ocean can be illustrated in Figure 1b, which shows the global average physics1 minus
253 physics2 mean temperature as a function of depth and time for the centennial2005 run. It
254 is noteworthy in Figure 1b the surface warming penetrating characteristic with warmer
255 physics1 ocean during the 2005–2105 period. The residual temperature propagation due
256 to atmospheric CO₂ concentration changes reach the upper 400m of the oceans in
257 approximately 50 years. Such ocean warming is consistent with the global ice reduction
258 shown in Figure 1c in terms of physics1–physics2, acting to reduce the surface albedo.

259 Yet, the analysis of decadal trends in SST are only weakly indicative of changes in the
260 top- of- atmosphere radiation balance (TOA). The importance of ocean heat content
261 (OHC) as a diagnostic for changes in the climate system has been emphasized in many
262 recent papers (Domingues et al. 2008; Levitus 2012; Palmer et al. 2011). Trends in total
263 ocean heat content strongly constrain TOA since the ocean is the primary heat store in the
264 Earth System. The essential balance in the Earth System is between total OHC (the
265 primary heat storage term) and F_{TOA} (the total radiation flux entering or leaving the
266 system). Therefore, we use conservation of energy to write the following equation

267

$$268 \quad F_{\text{TOA}} t = \frac{1}{A_{\text{earth}}} \frac{dH_o}{dt} (\text{W m}^{-2}), \quad (1)$$

269

270 where $F_{\text{TOA}} t$ is the net inward radiative flux through the top of the atmosphere,
271 A_{earth} is the area of the Earth, and H_o is the ocean heat content.

272 Numerous assumptions underlie this equation, particularly the neglect of the
273 thermal energy of the other components of the climate system (atmosphere, land,
274 cryosphere, and geothermal fluxes). Therefore, changes in OHC should approximately
275 balance the time integrated TOA. Climate modeling and observations indicate that to
276 understand the global energy budget fully, one needs to include measurements of the
277 deep oceans in the OHC analysis. The surface layers, even down to 700 meters, are not
278 robust indicators of total OHC, although the OHC-700m is strongly correlated (though
279 not perfectly) to the TOA imbalance.

280 The key questions here is whether this CO_2 -induced extra heat into the oceans is
281 indicative of changes in the Earth's top- of- atmosphere radiation balance, and thus an

282 increase of the accumulation of energy in the Earth System. Figure 2a shows the full
283 depth ocean heat content anomaly (OHCA) of the centennial2005 (physics1 minus
284 physics2) runs. This global average show a steady rise, to the end of the 100yr, of the full
285 depth ocean heat content for the centennial2005 (physics1) relative to the control run
286 with fixed CO₂ (physics2).

287 In order to assess the energy flux change on the top of the atmosphere due to
288 different CO₂ forcing conditions, we compute the top-of-atmosphere radiation flux
289 imbalance (TOA) implied by the ocean heat content difference in terms of its trend over
290 the 100 years period shown in Figure 2b. The estimated trend in TOA for the 2005-2105
291 period is of +0.3W/m⁻² per year represented by the red dashed line in Figure 2b. In a
292 recent determination of OHC for the 0-2000 m layer, Levitus et al. 2012 report an
293 implied rate in TOA of +0.27 W/m⁻² per unit area of earth's surface for the 1955-2010
294 period.

295 The radiative flux change on the top of the atmosphere due to different CO₂
296 forcing conditions can be affected by changes in clouds radiative forcing (e.g., Senior
297 1999; Stephens 2005). Radiative feedbacks involving low level clouds are a primary
298 cause of uncertainty in global climate model projections (IPCC 2007). In order to
299 investigate the low cloud responses to climate perturbations, we computed the time series
300 of global low cloud cover physics1–physics2 runs during the 2005-2105 period. The
301 increase in atmospheric CO₂ concentration in BESM-OA2.3 model produces a small
302 change in low cloud cover, around 0.1 % increase over 100 years (figure not shown). The
303 result is in accordance with other cloud responses to climate perturbations studies (Wyant
304 et al. 2006; Williams et al. 2006).

305 The results in this subsection indicate that there are mechanisms operating in the
306 BESM-OA2.3 climate model that are capable of re- distributing substantial quantities of
307 heat at all depths of the ocean on decadal time scale due to changes in atmospheric CO₂
308 concentration.

309

310 *The Atlantic Meridional Overturning Circulation*

311 The BESM-OA2.3 is able to reproduce the Atlantic Meridional Overturning Circulation
312 (AMOC), a fundamental process in the Atlantic Ocean. North Atlantic Deep Water
313 (NADW) is formed on the north Atlantic and exported southward below 1000m, while a
314 warmer near surface flow closes the system determining a northward net heat transport
315 (Talley 2003). Figure 3 shows the meridional spatial pattern of the AMOC as represented
316 by the BESM-OA2.3 and obtained from the average along 25 years (from year 95 to 120
317 of physics2 centennial run) of the vertical cumulative sum of the zonal total transport for
318 each depth and longitude. In other words, each point shows the average meridional total
319 transport through a section along that longitude, from that depth up to the surface. The
320 upper 1000m in the Atlantic basin is dominated by a northward transport up to 35°N.
321 Between 1000 and 3500m the cumulative transport reduces as a consequence of the
322 southward NADW, while below 3500 the Antarctic Bottom Water flows northward.

323 The BESM-OA2.3 not only reproduces the 3 layer meridional overturning
324 circulation system, but also well determines its intensity. Probably the best estimate of
325 the AMOC has been done by the RAPID-MOC, an observing array system along the
326 26.5°N cross section operating since 2004, suggesting a transport of 18.5 ± 4.9 Sv (Johns
327 et al. 2011). At the same section, the BESM-OA2.3 reproduces a maximum transport of

328 approximately 20 Sv, slightly higher than, but still within the uncertainty of the
329 observations.

330 The most impressive point is that this is the preferred stable state of the model.
331 The small frame in Figure 4 shows the maximum AMOC transport throughout the 122
332 years elapsed after the spinup, *i.e.*, once the coupled model runs free with physics2
333 configuration (constant atmospheric CO₂ concentration). The transport is initially just
334 below 10 Sv, nearly half of the observed value, but it continually increases until
335 stabilizing at approximately 20 Sv, after 80 years. There were no prescribed forcing
336 driving to this result, except for the model internal dynamics. The adequate adjustment of
337 the AMOC reproduced on the BESM-OA2.3, despite the bias from the spinup, is a strong
338 argument that the model is capturing the main Earth system processes, and hence should
339 be a good approximation of reality.

340

341 *Model climatology and systematic errors*

342 *SST AND PRECIPITATION ANNUAL BIASES*

343 In order to evaluate the ability of BESM-OA2.3 to reproduce the observed climate, the
344 annual biases of SST and precipitation were computed and contrasted with the biases for
345 other three CMIP5 global models, namely; Hadley Centre Coupled Model version 3
346 (HadCM3), NCAR's CCSM4 coupled model, and the Canadian CanCM4 coupled model.
347 The Optimal Interpolation SST (OISST) dataset, combining *in situ* and satellite SST
348 interpolated onto a one-degree grid with weekly resolution (Reynolds et al. 2002), was
349 used to assess the SST biases; while the CPC Merged Analysis of Precipitation (CMAP),
350 constructed from gauge observations, satellite estimates, and numerical model outputs

351 (Xie and Arkin 1998) was used to evaluate the precipitation biases. Both datasets were
352 averaged in time, resulting in global maps of climatological values.

353 The annual mean SST biases relative to OISST are presented in Figure 5. The
354 results indicate that the BESM-OA2.3 model has an extensive negative SST bias over the
355 tropical and subtropical oceans (Figure 5a), much like both the HadCM3 model (Figure
356 5b) and the CanCM4 model (Figure 5d). The exception is on the eastern boundaries of
357 the Pacific and Atlantic Ocean, where the models are unable to properly reproduce the tilt
358 of the equatorial thermocline (not shown), resulting in a warm bias of approximately 2 K.
359 The pattern of a colder tropical North Atlantic and a hotter tropical South Atlantic is
360 related to wind stress biases and has also been observed in other models (Muñoz et al.,
361 2012). Bottino and Nobre (2012) have studied the impact of cloud cover
362 parameterizations on BESM-OA2.3, obtaining a reduction of the North Atlantic cold bias
363 and South Atlantic warm bias on this model.

364 The most outstanding cold bias occurs in the simulation of the North Atlantic in
365 all four CMIP5 models in Figure 5, which is also similar to GFDL CM2.5 (Delworth et
366 al. 2012), since both CM2.5 and BESM-OA2.3 coupled models share the same ocean
367 component; and on the Southern Ocean, where warm biases are predominant much like
368 the NCAR CCSM4 model (Figure 5a,c). The model overestimates SSTs by as much as 3
369 K in eastern boundary currents (e.g. Humboldt, California, and Benguela currents). One
370 possible explanation for the spatial extent and magnitude of the warm bias in the eastern
371 currents is that it results from positive feedback between the processes involved in the
372 formation of marine stratocumulus clouds over cold waters and their cloud shading effect

373 reducing the net surface radiative forcing (Ma et al. 1996, Siqueira and Nobre 2006,
374 Zheng et al. 2011).

375 The SST biases in BESM-OA2.3 show no significant variation when biases from
376 the ensemble mean of individual experiments decadal1960, decadal1980, and
377 decadal2005 are intercompared. The biases are also indifferent to CO₂ values, displaying
378 no significant difference between the physics1 and physics2 integrations (figures not
379 shown). These results suggest that the observed biases in SST are essentially determined
380 by the model formulation.

381 Annual mean precipitation biases are shown in Figure 6, following the same
382 approach used for SST. While the BESM-OA2.3 model is able to reproduce global
383 observed values of precipitation, it displays biases that arise from the inability to position
384 the ITCZ (Intertropical Convergence Zone) correctly, a common feature of all the four
385 CMIP5 models in Figure 6. The Tropical Pacific is characterized in BESM-OA2.3 by a
386 double ITCZ and a missing SPCZ (South Pacific Convergence Zone) (Figure 6a). The
387 Tropical Atlantic is better reproduced and although the ITCZ is shifted towards the south,
388 the SACZ (South Atlantic Convergence Zone) has an improved representation compared
389 to its Pacific counterpart. These deficiencies, of the double ITCZ in the Pacific and the
390 southward shift of the ITCZ in the Atlantic, are known problems of current global models
391 (Lin 2007; Richter and Xie 2008).

392 The biases originated from not representing the ITCZ properly include an excess
393 of precipitation over the South Atlantic Ocean, Indian Ocean, and most of the South
394 Pacific Ocean, with the exception of the SPCZ region. The Equatorial Pacific and
395 northern Tropical Atlantic are characterized by negative biases in precipitation, while

396 most of the North Pacific and North Atlantic Oceans show positive biases. Over land, the
397 strongest signal comes from the Amazon basin, where the model is characterized by
398 negative biases, similar to the HadCM3 (Figure 6b) and CanCM4 (Figure 6d) models.
399 These biases are caused by the reduced convection over the Amazon forest and are
400 possibly connected to the SST biases via the Walker circulation over the Atlantic Ocean,
401 through trade winds, and the tilt of the equatorial thermocline (Braconnot et al. 2007;
402 Breugem et al. 2008; Richter and Xie 2008; Tozuka et al. 2011). In earlier versions of the
403 coupled ocean-atmosphere model (e.g. Nobre et al. 2009), base for the BESM-OA
404 coupled model, the trade winds over the Atlantic Ocean had their direction reversed
405 during part of the year, producing strong positive SST biases when the equatorial
406 thermocline collapsed, reducing convection over land. This has been improved in the
407 current version of the model, mostly due to the Grell-ensemble cumulus
408 parameterizations used (Grell and Devenyi 2002), although some bias still remain.
409 Similar to the SST, precipitation biases are time and physics invariant, showing no
410 significant differences between periods or CO₂ physics.

411 Figure 7 shows the seasonal cycle in the deviations of SST at the equator about
412 the annual mean for OISST and four current CMIP5 models (top panel). The variation in
413 observed SST is dominated by the annual harmonic that peaks at about 100°W, with the
414 warm phase in April slightly stronger than the cold phase in September-October (Figure
415 7a). All four models fail to capture the correct start and duration of the warm phase
416 (Fig7b-e). The main part of seasonal variation in equatorial SST is realistically located
417 over the eastern Pacific in all four models, although deviations are concentrated too close
418 to the coast for BESM-OA2.3 (Figure 7b) or extend too far west as in CanCM4 (Figure

419 7e). The seasonal cycle is quite weak in HadCM3 and CCSM4 (Fig7.c-d) and the
420 simulated SSTs are correctly dominated by the annual harmonic, except in BESM-OA2.3
421 for which the semi-annual harmonic is the dominant feature (Figure 7b).

422 In order to investigate to what extent convection is taking place south of the
423 equator in the eastern Pacific and its seasonality, Hovmöller diagrams of precipitation for
424 CMAP and four current CMIP5 models averaged from 100°W to 150°W are also plotted
425 in Figure 7 (bottom panel). The rainfall from CMAP observations is largely confined to
426 an ITCZ located between 5°N and 15°N (Figure 7f). The seasonal variation in latitude for
427 the observations is small with the ITCZ reaching 15°N around September and its
428 southern-most position in March-April, almost reaching the equator. The rainfall
429 maximum simulated by the BESM-OA2.3 model has realistic magnitude, but the
430 precipitation is severely overestimated south of the equator (much like CSSM4 and
431 CanCM4 in Figure 7i-j) and produces a rather persistent double ITCZ throughout the year
432 (Figure 7g).

433

434 *Interannual variability*

435 EL NIÑO–SOUTHERN OSCILLATION

436 The spatial patterns of tropical interannual SST variability for the BESM-OA2.3 model
437 and observations are shown in Figure 8. The BESM-OA2.3 model exhibits weaker
438 equatorial Pacific variability than that observed in recent decades, correctly placing the
439 center of the Pacific SST variability relative to the observed pattern, although with too
440 little variability along the coast of Peru. It is worth noting that the tropical Atlantic SST
441 simulations also show weaker interannual SSTA fluctuations than the observations.

442 The ENSO variability is frequently characterized by indexes of SST variability
443 for specific regions over the equatorial Pacific, however the SST and wind patterns may
444 differ among models and observation. Therefore, performing comparisons can be
445 difficult. Nobre and Shukla (1996), for example, discussed the tropical Atlantic ocean–
446 atmosphere variability through empirical orthogonal function (EOF) analysis (Lorenz
447 1956), joining the fields of SST and wind stress anomalies normalized by the local
448 standard deviation prior to the EOF calculations, a technique referred as “joint EOF” (or
449 “jEOF”) analysis. Hence it is more convenient to characterize ENSO variability in terms
450 of joint EOF of the SST and wind stress fields so that the model’s ENSO can be studied.
451 In this case, ENSO variability is considered to be represented by the leading jEOFs of
452 anomalous monthly SST and wind stress over a region that nearly spans the equatorial
453 Pacific, and by the time series of the corresponding principal components. In this
454 analysis, the jEOFs are normalized to have unit spatial variance, and hence the variances
455 of each principal component (PC) reflect spatiotemporal variances of corresponding SST
456 and wind stress anomalies, with variances of all the PCs summing to the total variance.
457 The jEOF analysis is presented in Figure 9, where (a) shows the first jEOF of observed
458 SSTs (Reynolds et al. 2002) and wind stress fields estimated with observed winds from
459 NCEP/NCAR reanalysis (Kalnay et al. 1996), from January 1950 to December 2010 (60
460 years).

461 The main pattern of variability associated with ENSO on the outputs of BESM-
462 OA2.3 physics2 centennial run with fixed atmospheric CO₂ concentration can be
463 determined through the first jEOF of monthly SSTs and wind stress, and is plotted in
464 Figure 9b for a 100 years period. The first jEOF (jEOF1) reflects the mature phases of the

465 ENSO life cycle, i.e., the mature phases of El-Niños and La-Niñas. Despite broadly
466 resembling the observed variability (Figure 9a), we detect a strong meridional
467 confinement of the SST anomalies around the equator and the center of action of this
468 mode is displaced eastward when compared to the observations. The meridional
469 confinement can be caused by a narrow meridional span of zonal wind stress anomaly
470 and has implications in the interannual variability (Kirtman 1997, Deser et al. 2006). The
471 first jEOF explains only 19% of variance, versus over 38% in the observations.

472 The first jEOF pattern of the BESM-OA2.3 model in the equatorial Pacific does
473 not change significantly over the 100yr period for a twin physics1 simulation (not
474 shown), suggesting that the ENSO variability is only affected in its characteristic period
475 by weak global temperature increase.

476 The power spectra of the first principal component (PC1) for observed SST (solid
477 black) and for the twin 100yr simulations starting at 2005 physics1 (solid red) and
478 physics2 (solid blue) are shown in Figure 9c. The seasonal cycle and linear trends were
479 removed when computing the jEOFs and the respective PCs to better isolate the modes of
480 variability. The observed ENSOs have a relatively broad spectrum spanning 3–5 years.
481 The model captures this feature qualitatively with peak ENSO period in BESM-OA2.3 at
482 approximately 5 years, which is on the higher end of the 3-5 year band (towards lower
483 frequencies). Changes in ENSO frequency under CO₂ increase can be noticed in Figure
484 9c (solid red) and an overall tendency for ENSO period to decrease is evident in
485 accordance with Merryfield (2006) multimodel ensemble analysis.

486 Overall the model has a robust ENSO signal with multi-decadal fluctuations in
487 amplitude (not shown), and an irregular period between 3 and 5 years with peak ENSO

488 period at approximately 5 years. The main ENSO pattern of SST variability simulated by
489 the model is shifted about 30° east of the observed patterns, and the transition pattern
490 matches the observations well with some deficiencies mainly over the equator (not
491 shown). Such problems may be linked to the model mean state biases namely the
492 equatorial cold bias, the double ITCZ, and the overstratified surface waters near the
493 South American coast.

494

495 EXTRATROPICAL TELECONNECTIONS

496 The extratropical teleconnections are recurrent atmospheric circulation patterns
497 associated with variations in the strength and location of the polar jet stream, the
498 subtropical jet stream, or mid-latitude storm tracks (Chang and Fu 2002).

499 The Pacific-North American teleconnection pattern (PNA) is one of the most
500 prominent and influential modes of low-frequency variability in the Northern Hemisphere
501 mid-latitudes beyond the tropics (Wallace and Gutzler 1981). The PNA pattern features
502 large-scale above-average surface pressure anomalies that arc from the Hawaiian Islands
503 and over the intermountain region of North America, and below-average surface pressure
504 located south of the Aleutian Islands and over the southeastern United States (Figure
505 10a).

506 The PNA pattern is associated with noticeable fluctuations in the strength and
507 location of the East Asian jet stream. The positive phase is associated with an intensified
508 East Asian jet stream, and with an eastward shift in the jet exit region toward the western
509 United States. The negative phase is associated with a westward shrink of the jet stream
510 toward eastern Asia, and blocking activity over the high latitudes of the North Pacific.

511 Although the PNA and ENSO are distinct natural internal modes of climate
512 variability, the PNA has been found to be strongly influenced by the ENSO phase (Straus
513 and Shukla 2002). The positive phase of the PNA pattern tends to be associated to El-
514 Niños, and during La-Niñas, the PNA tends to be in the negative phase.

515 The BESM-OA2.3 model matches well the PNA pattern and its amplitude (Figure
516 10b) with the center of action over the intermountain region of North America shifted
517 about 20° east of the observed pattern. This mode of climate variability explains 12% of
518 variance in the model versus over 17% in the observations.

519 In the Southern Hemisphere (SH), there is also one prominent low-frequency
520 mode in the extratropics that resembles the PNA pattern, which is characterized by a
521 well-defined wave train extending from the central Pacific to Argentina, and is referred to
522 as the Pacific–South American (PSA) pattern (Ghil and Mo 1991; Mo and Higgins 1998;
523 Lau et al. 1994).

524 The center of action of Pacific-South American (PSA) teleconnection pattern is
525 located in the west of the Antarctic Peninsula in the observations and comprises a wave-
526 train of alternate positive and negative geopotential height (or pressure) anomalies across
527 the South Pacific emanating from the tropics (Figure 11a).

528 It has been suggested that this teleconnection pattern that acts over South America
529 impacts the South Atlantic Convergence Zone, which results in a regional seesaw pattern
530 of alternating dry and wet conditions (Nogués-Paegle and Mo, 1997). The PSA has also
531 been found to be influenced by ENSO conditions (Mo and Nogués-Paegle 2001).

532 The model simulations show two main centers of action instead of one, a negative
533 center located in the west of the Antarctic Peninsula similar to observations, and another

534 with similar amplitude over the Antarctic Peninsula. This mode of climate variability in
535 the SH explains 18% of variance in the model, versus over 14% in the observations.

536

537 TROPICAL ATLANTIC VARIABILITY

538 The Tropical Atlantic Variability (TAV) is characterized by two main modes of
539 variability, an ENSO-like equatorial mode (Zebiak 1993) and an interhemispheric
540 gradient mode (e.g. Nobre and Shukla 1996). This last mode, also referred as the
541 “Atlantic dipole mode” or the “Atlantic meridional mode,” is related to the cross-
542 equatorial gradient of the SST anomalies and resultant migrations of the intertropical
543 convergence zone. Consequently, it has an impact on rainfall over the tropical Atlantic,
544 including the continental areas of the northeast Brazil and African’s Sahel region, at
545 interannual and decadal climate scales (Doi et al. 2010 and references therein).

546 Several studies have employed the technique of EOF analysis to study covariant
547 patterns of variability over the Atlantic Ocean (Venegas et al. 1997; Weare 1977).
548 According to Nobre and Shukla (1996), from 1963 to 1987 the first jEOF mode spatial
549 pattern of SST and wind stress anomalies showed a zonal, dipole-like structure in both
550 SST and wind stress anomalies over the Tropical Atlantic, suggesting that warmer SSTs
551 are connected to weaker trade winds, and colder SSTs to stronger trade winds
552 (Dommenget and Latif 2000). A similar analysis is presented in Figure 12a, considering
553 observed SSTs (Reynolds et al. 2002) and wind stress fields of the NCEP/NCAR
554 reanalysis, from January 1960 to December 2010 (51 years), where all linear trends and
555 seasonal cycles were removed prior to the analysis. The spatial pattern of the first jEOF,
556 accounting for 12.5% of the total variance (Figure 12a), shows that the structure of the

557 Tropical Atlantic variability described by Nobre and Shukla (1996) was sustained until
558 2010, with the dipole-like feature in the SST pattern, and analogous connections SST–
559 trade winds. The power spectrum analysis of the amplitude time series of this mode
560 (Figure 12c) reveals a low frequency oscillation, which is represented as the main peak at
561 11.75 years, confirming the decadal timescale behavior of this pattern as noticed by
562 Nobre and Shukla (1996).

563 The occurrence of SST biases in the tropical Atlantic is a common issue of
564 coupled models and, as discussed by Richter and Xie (2008) and Doi et al (2010), almost
565 all CMIP3 models show the zonal SST gradient along the equator opposite to what is
566 observed. This deviation is possibly attributed to the lack of the cold tongue in the
567 equatorial Atlantic in many coupled model simulations. Even recent coupled models,
568 with higher horizontal resolution and updated parameterizations present SST biases in the
569 eastern equatorial Atlantic and the Angola–Benguela area (Doi et al. 2012). The BESM-
570 OA2.3 model also has the SST bias in the tropical Atlantic as discussed before (Figure 5).
571 Notwithstanding, it is able to simulate the Atlantic dipole mode quite well. Figure 12b
572 shows the jEOF analysis results of the physics2 centennial run simulation, starting from
573 December 2005 and with fixed CO₂ concentration. The spatial patterns of SST and wind
574 stress anomalies are very similar to the ones for observed data (Figure 12a), except for
575 the region near the Gulf of Guinea, where the negative SST patterns are stronger and the
576 wind stress patterns are preferably from SE instead of from S–SW, reflecting the SST
577 biases and the consequent discrepancies in the trade wind anomalies over this region. The
578 main period of this simulated Atlantic dipole mode is 15 years (Figure 12c). For the twin
579 centennial simulation with increasing CO₂ of physics1, jEOF analysis of the resultant

580 simulation also shows a dipole structure (figure not shown) similar to Figure 12b, but
581 with the period of maximum spectral energy density at 6.43 years (Figure 12c).
582 Apparently, the variations of the atmospheric radiative balance caused by an increasing
583 atmospheric CO₂ concentration act to increase the Atlantic dipole frequency. Yet, the
584 mechanisms by which it occurs still have to be further investigated.

585 4. Summary and Conclusions

586 This article has explored the global features and regional aspects of the BESM-OA2.3
587 coupled model for climate change research. The numerical experiment design was set
588 accordingly to the CMIP5 protocol and was aimed at explicitly revealing the model
589 sensitivity to atmospheric CO₂ concentration increase, as a representation of the buildup
590 of Green House Gases in the atmosphere. Extended runs with over 2,000 years of
591 ensemble members were generated to scrutinize the model behavior for decadal climate
592 variability and change. It was demonstrated that the BESM-OA2.3 model responds to
593 increasing atmospheric CO₂ concentrations in a consistent manner. The model
594 simulations showed changes in Earth's top-of-atmosphere radiation balance, an
595 increase of the accumulation of energy in the Earth System, and associated global sea
596 surface temperature increase and marine ice reduction.

597 In terms of annual biases, the simulations show some of the same characteristics
598 of other global coupled ocean-atmosphere models, namely the double ITCZ and an
599 almost absent SPCZ over the Pacific Ocean, in addition to the southward displacement of
600 the Atlantic ITCZ. Concurrent with most global coupled ocean-atmosphere models, a
601 warm SST bias is present along the western coasts of South America and Africa.

602 Regarding precipitation, the model shows excess rainfall over the oceans and deficiencies
603 over the continents, particularly over the Amazon basin.

604 The model biases showed no significant sensitivity to atmospheric CO₂
605 concentrations, suggesting that such biases are a consequence of more basic model
606 physics deficiencies. The excess rainfall over the oceans was diagnosed after the fact, and
607 a new formulation for water vapor surface fluxes over the ocean has been developed at
608 CPTEC, and will be tested in a future experiment. Yet, the BESM-OA2.3 simulations
609 presented an improved representation of the SACZ relative to previous versions of its
610 base coupled ocean-atmosphere model, which we interpret as a direct consequence of the
611 improvements on deep cumulus parameterizations used for this research.

612 On the interannual spectrum of climate variability, the simulations have shown a
613 robust ENSO signal over the equatorial Pacific Ocean, with peak energy in the 3 to 5 year
614 range, and a dipole-like pattern of SST and surface winds with peak energy on the
615 decadal scale in the Atlantic Ocean, in agreement with observations. However, for the
616 centennial model runs under increasing atmospheric CO₂ concentrations the coupled
617 ocean-atmosphere modes of variability over both the Pacific (e.g. ENSO) and the tropical
618 Atlantic (e.g. the meridional mode) presented signals of enhanced periodicity, which may
619 be an indication that under growing atmospheric CO₂ concentrations, interannual climate
620 variability may grow. The model also represented atmospheric teleconnections in close
621 agreement with observations, which is an important benchmark for its use for climate
622 variability and change research.

623 The BESM-OA2.3 model has been used for operational seasonal predictions at
624 CPTEC and constitutes a building block on the Brazilian framework for climate change
625 research – Rede CLIMA.

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837 **Figures Captions**

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839 centennial2005 runs for (a) Sea Surface Temperature (SST), the red line is the linear fit
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841 global ocean temperature (K), depicting the propagation of the temperature signal
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874 control run.

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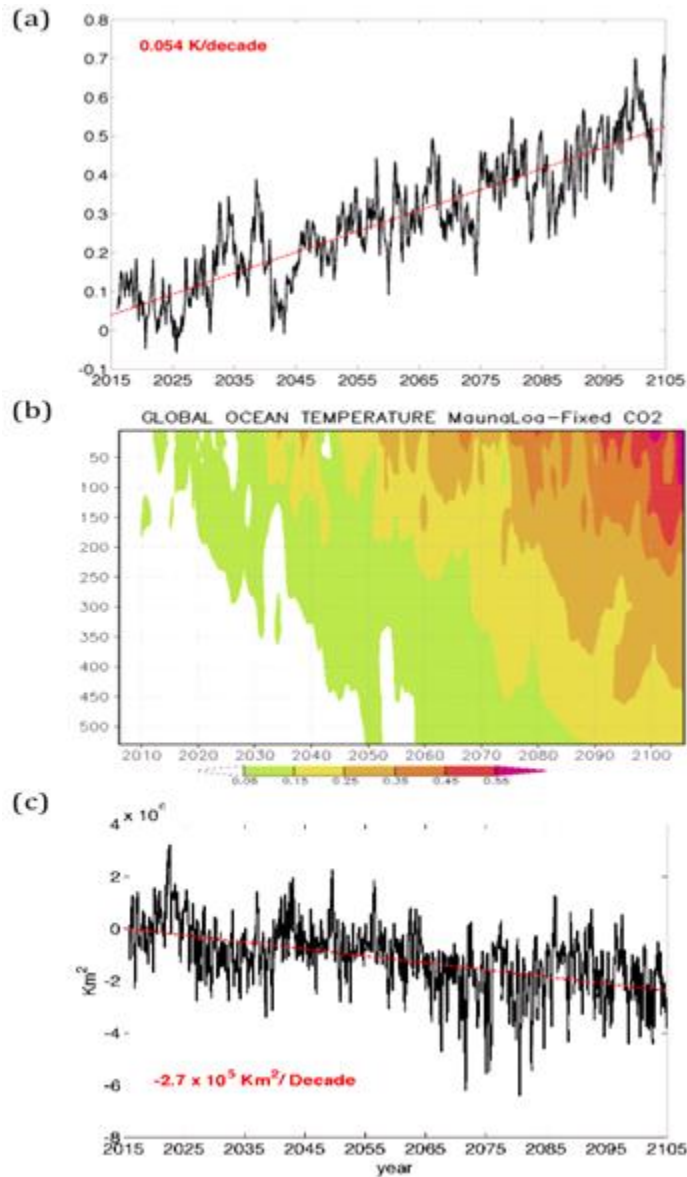
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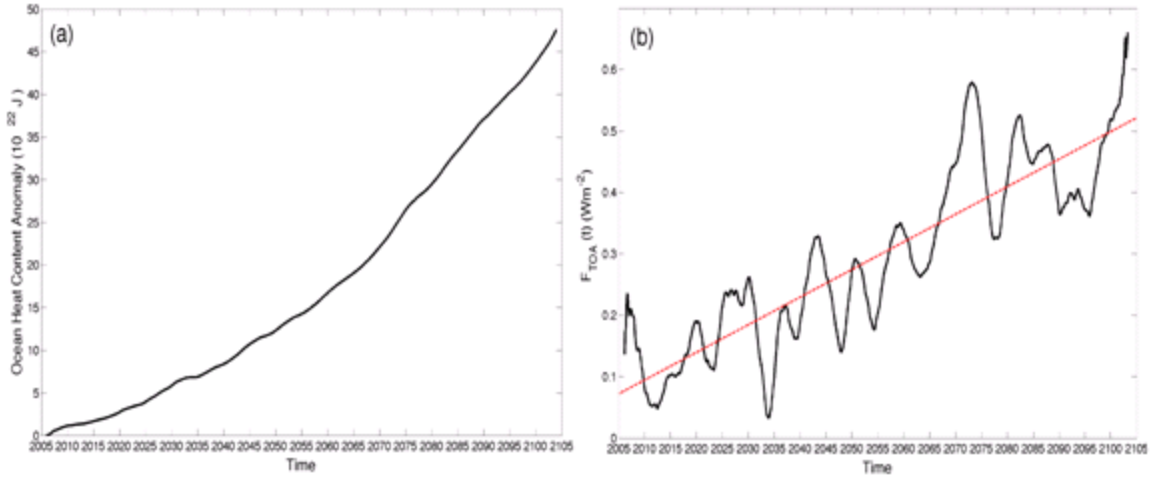
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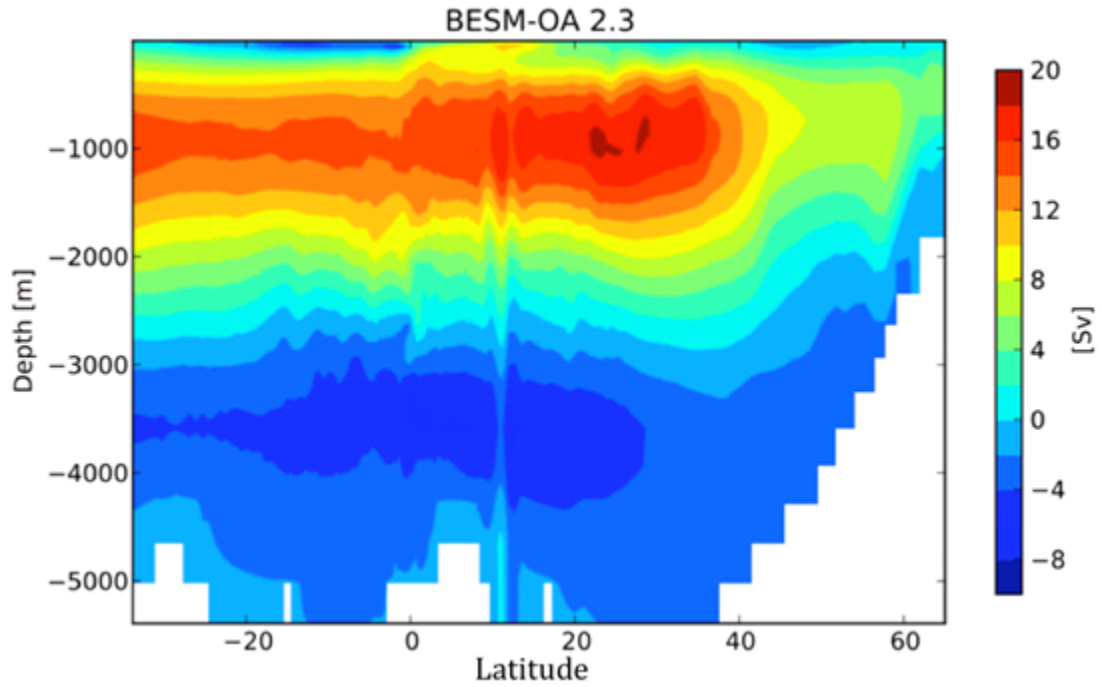
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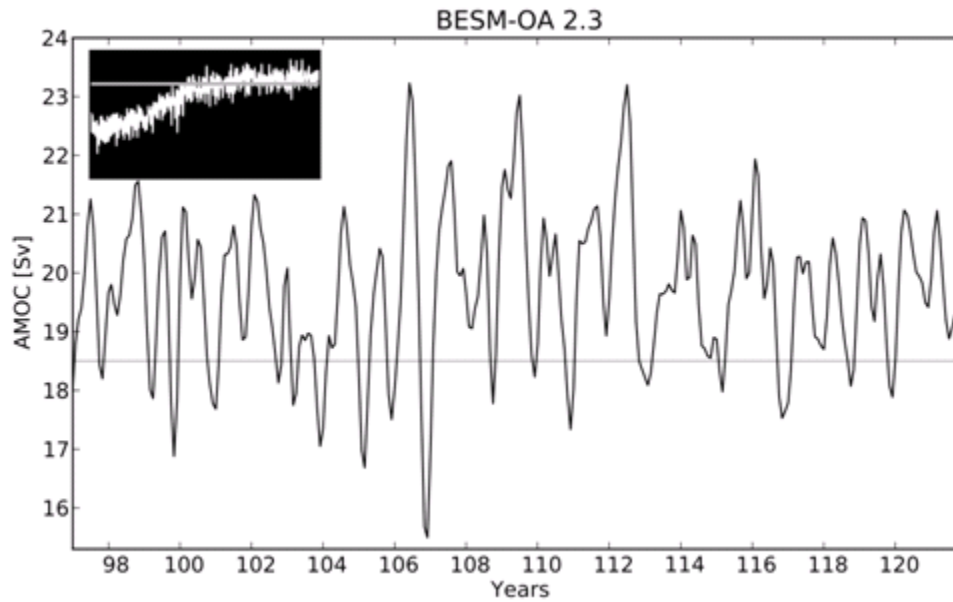
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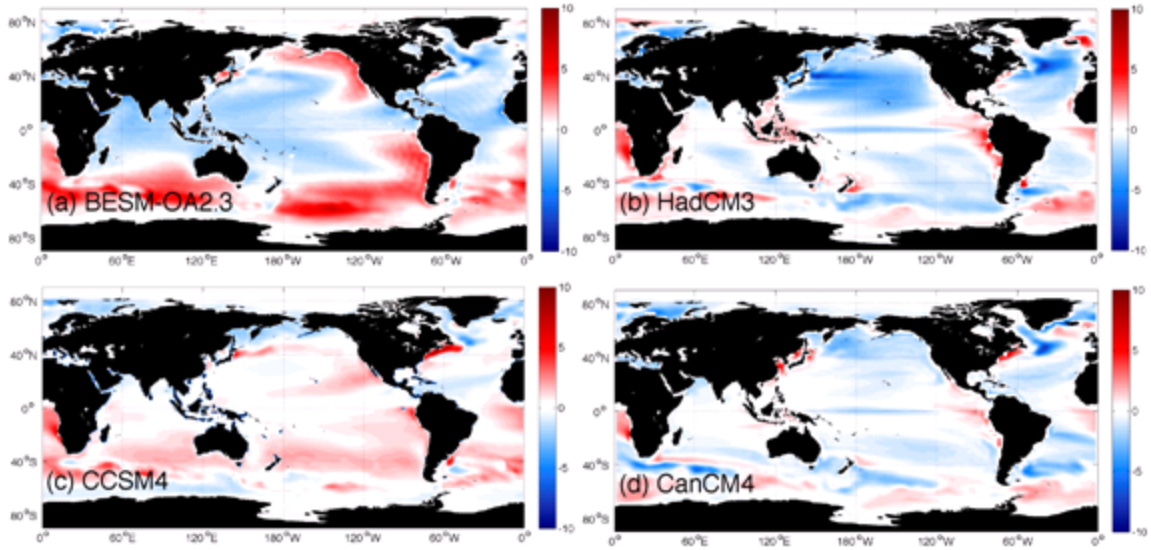
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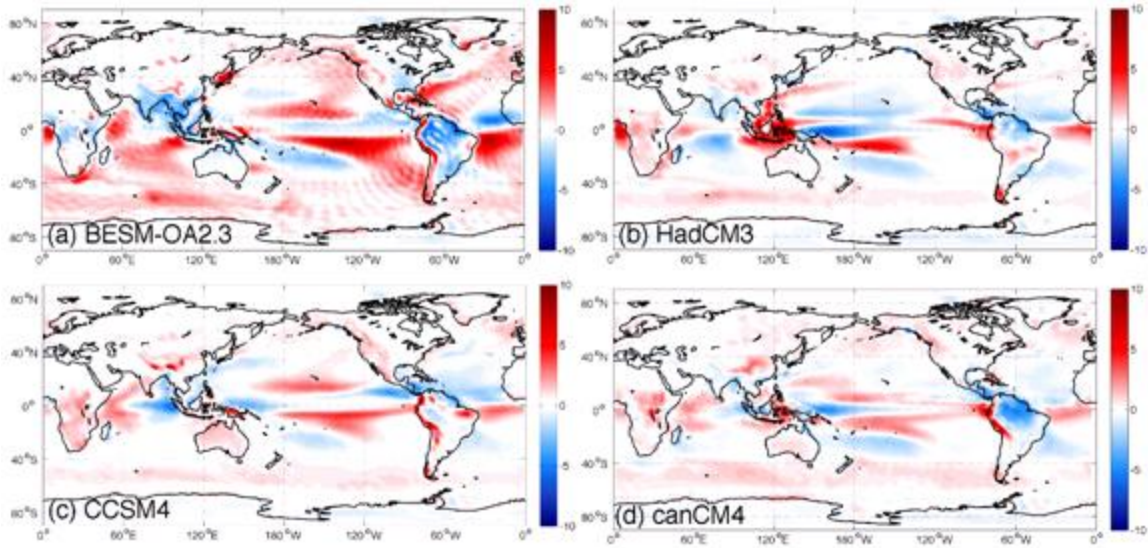
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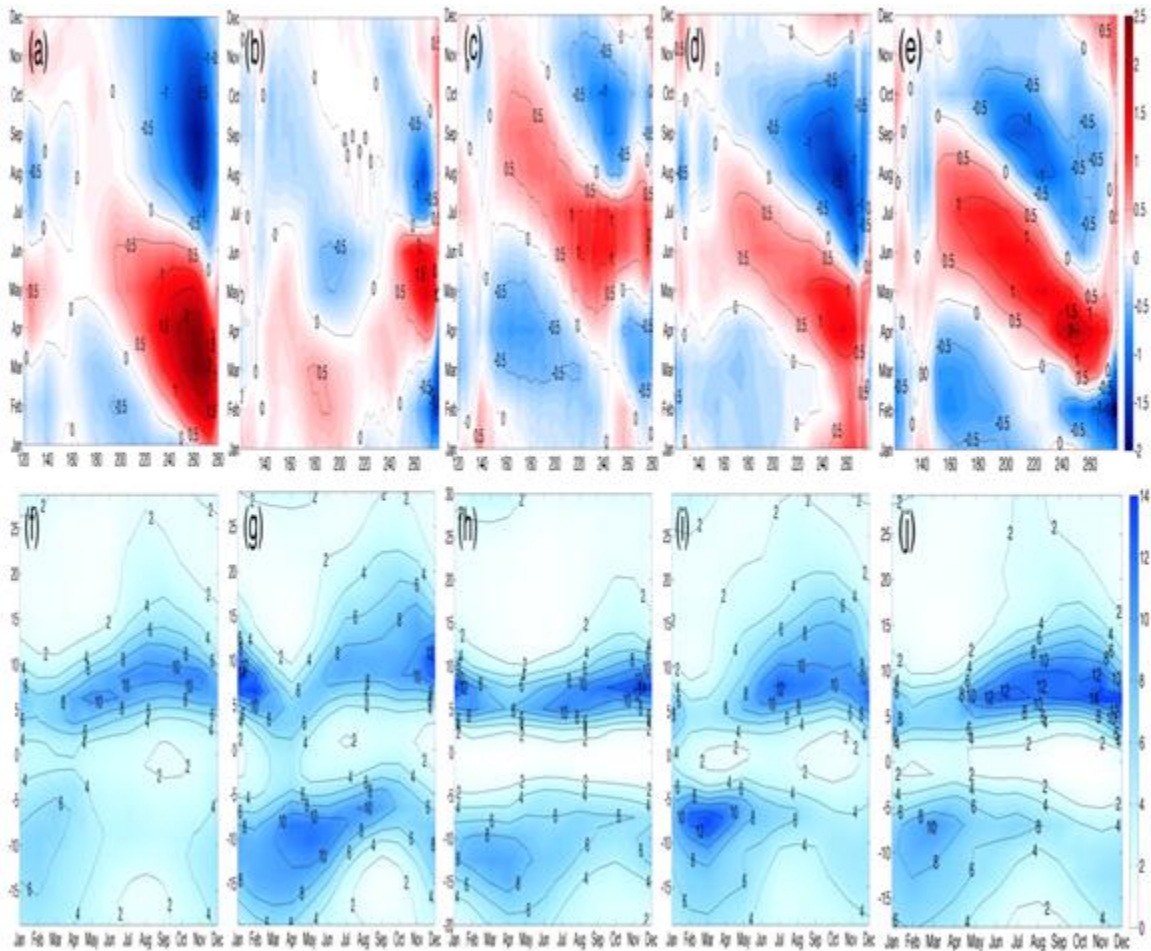
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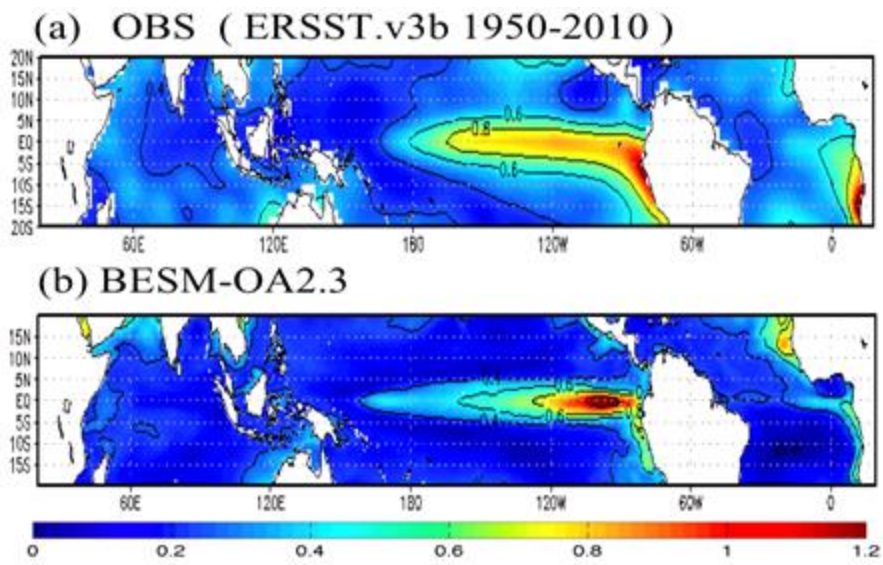


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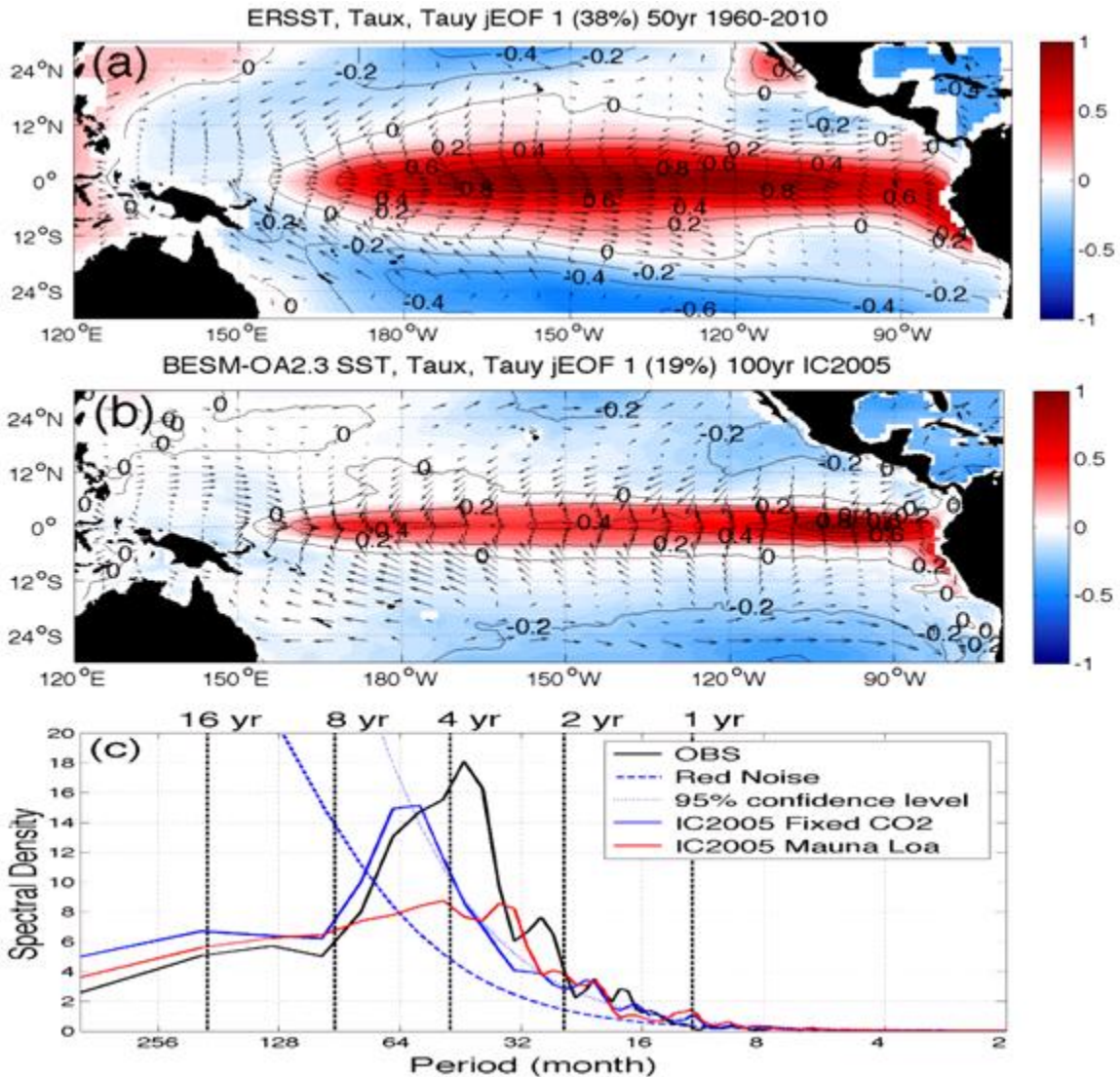
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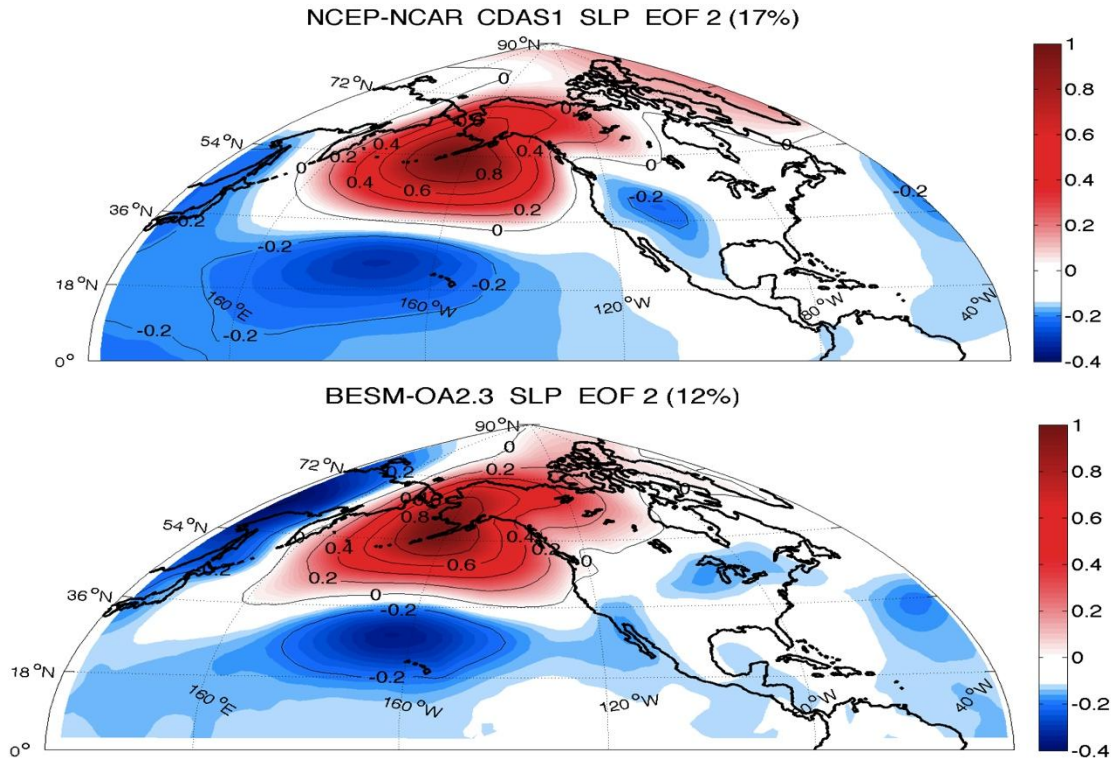
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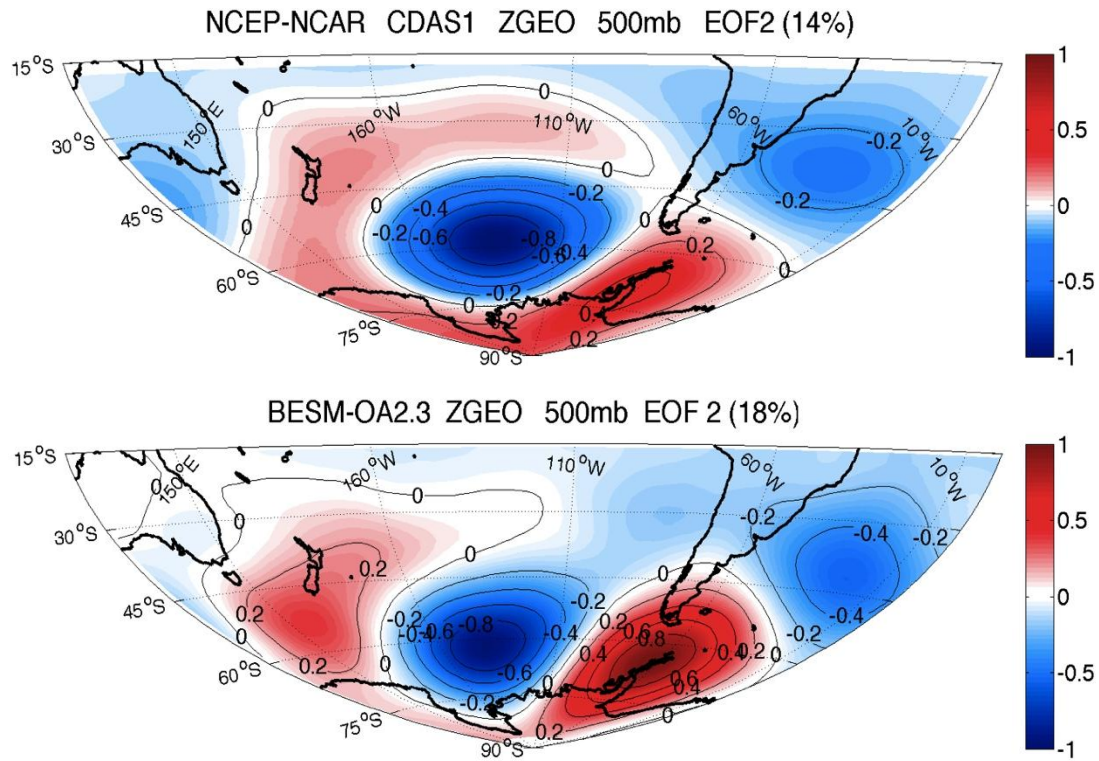
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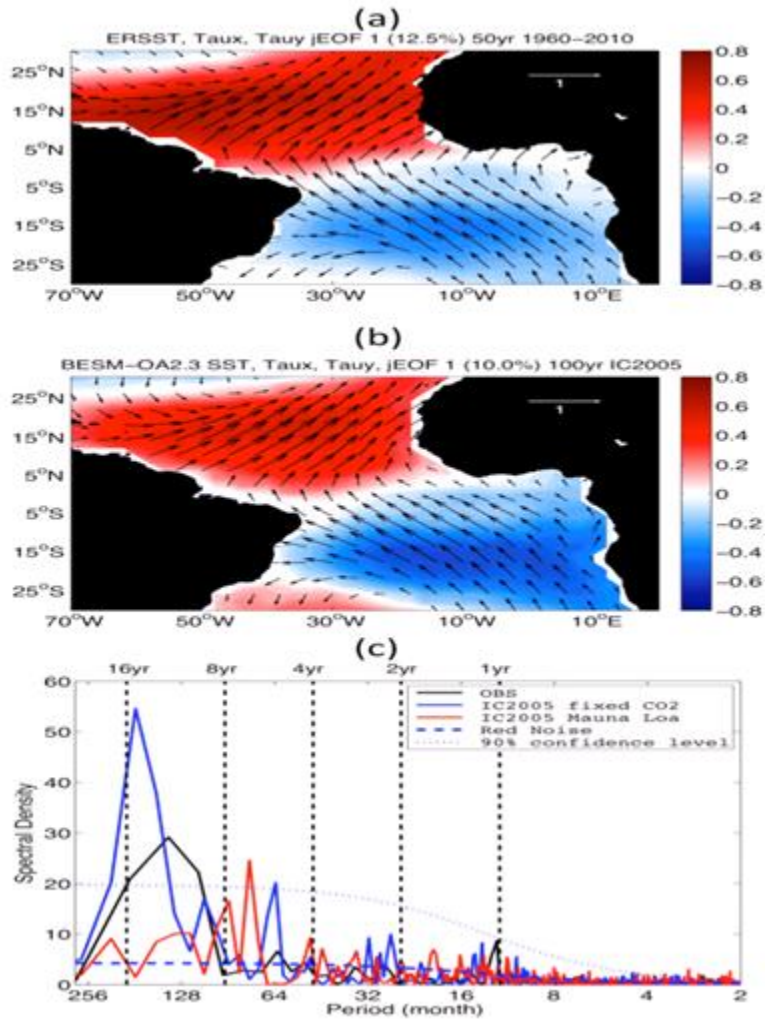
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