Mean and Variability of the Tropical Atlantic Ocean in the CCSM4

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Abstract:

In this study we analyze important aspects of the tropical Atlantic Ocean from simulations of the 4th version of the Community Climate System Model (CCSM4): the mean sea surface temperature (SST) and wind stress; the Atlantic warm pools; the principal modes of SST variability; and the heat budget in the Benguela region. The main goal was to assess the similarities and differences between the CCSM4 simulations and observations. The results indicate that the tropical Atlantic is more realistic in CCSM4 compared to the previous version of the CCSM. Yet, there are still significant biases in the CCSM4 SSTs, with a colder tropical North Atlantic and a hotter tropical South Atlantic, that are related to biases in the wind stress. These are also reflected in the Atlantic warm pools in April and September with its volume greater than in observations in April, and smaller than in observations in September. The variability of SSTs in the tropical Atlantic is well represented in CCSM4. However, in the equatorial and tropical South Atlantic regions CCSM4 has two distinct modes of variability. A heat budget analysis of the Benguela region indicates that the variability of the upper ocean temperature is dominated by vertical advection, followed by meridional advection.
1. Introduction

It is important to understand the climate dynamics in the tropical Atlantic basin. The tropical Atlantic impacts the climate of Northeast Brazil, Northwestern Africa, Central America and the Caribbean regions. The tropical Atlantic is also important because of its remote impacts on the tropical Pacific (Ding et al. 2011; Losada et al. 2009; Saravanan and Chang, 2000) and on the Indian Ocean (Kucharski et al. 2008). Even though the tropical Atlantic has been recognized as an important region in the Earth’s coupled climate system, it has been challenging to model adequately by coupled climate models. A few recent reports summarize the advances in the understanding of the tropical Atlantic climate and its variability (e.g., Hurrell et al. 2006; Xie and Carton, 2004; Garzoli and Servain, 2003; Visbeck et al. 2001). Some of the main aspects of the tropical Atlantic Ocean discussed in the current study include: the seasonal cycles of wind stress and sea surface temperature (SST), the Atlantic warm pools and the interannual variability of SST, including that of the Benguela region.

The seasonal cycle of SST in the tropical Atlantic is related to the seasonal cycles of wind stress and of the Intertropical Convergence Zone (ITCZ). The seasonal cycle is the largest ocean-atmosphere signal in the region. As discussed in Servain et al (1998), the timing and characteristics of the seasonal evolution of SST, winds and the ITCZ depend on the coupled air-sea-land dynamics. For example, when the precipitation is greatest over the equatorial Amazon in boreal spring, the equatorial easterlies are weak and Atlantic SSTs are quasi-uniform in the near-equatorial belt. As the seasons progress the ITCZ shifts northward and stronger southeasterly winds amplify the eastward tilt of the thermocline leading (through upwelling and Kelvin waves) to a colder eastern equatorial Atlantic in late spring and early summer. Farther
north in the Caribbean Sea, the low-level (~925 hPa) zonal winds have a semi-annual cycle with peaks in February and July (Muñoz et al. 2008) given the seasonal changes in pressure and thermal wind. The Caribbean low-level jet affects precipitation in the Caribbean and Central America, and also affects regional SST (Small et al. 2007; Muñoz et al. 2008; Wang et al. 2008) through coastal upwelling.

A significant aspect of the tropical Atlantic seasonal cycle is its warm pool. Warm pools (WP) have been defined as those regions of the ocean with temperatures greater than 28.5°C (Tian et al. 2001; Wang and Enfield, 2003). Beyond its surface manifestation and extent the WP in the Tropical Atlantic has vertical and horizontal profiles that are important with respect to the heat content of the upper layer of the ocean. The Atlantic WP has a component in the northwestern tropical Atlantic spanning the Gulf of Mexico and the Caribbean Sea (i.e., the Intra-Americas Sea) that peaks during boreal summer and early fall (Wang et al. 2003). The heat content in the tropical North Atlantic (TNA) WP is an important regulator of the air-sea interactions in tropical storms and hurricanes passing through that region (Wang et al. 2006; Landsea et al. 1999; Gray, 1990). The impact of the TNA-WP SSTs extends to the tropical Pacific through an upper-level Gill-type circulation (Saravanan and Chang (2000); Wang et al. (2008)). In contrast to the TNA-WP, much less attention has been given to the tropical South Atlantic warm pool where SSTs also exceed 28.5°C during boreal spring.

A majority of coupled GCMs have biases in the mean state of both the tropical Pacific and Atlantic including notorious warm biases in the southeastern tropical basins (Large and Danabasoglu, 2006; Zuidema et al., 2011). In the tropical Atlantic, the most severe warm SST bias (in excess of 5°C) occurs along the Benguela coast of southwestern Africa (e.g. Chang et al., 2007; Richter and Xie 2008), i.e., the TSA warm pool region. This spurious pool of abnormally
warm water simulated by most coupled climate models alters the large scale meridional SST
gradient across the tropical Atlantic, and thus may project on the natural mode of the tropical
Atlantic variability known as the meridional or inter-hemispheric SST mode (e.g. Xie and
Carton, 2004). Furthermore, some coupled models also have relaxed (weaker) zonal winds along
the equator related to weaker precipitation over the Amazon region, that is in turn related to the
warm bias in the eastern equatorial Atlantic (Doi et al. 2012; Tozuka et al. 2011; Richter et al.
2011; Wahl et al. 2011; Richter and Xie, 2008; Chang et al. 2007). Chang et al. (2007) and
Richter and Xie (2008) have shown that abnormally weak equatorial easterly wind is responsible
in part for the time mean warm bias of the Benguela SST. Tozuka et al. (2011) showed that the
magnitude of the SST bias is related to differences in the convection parameterization in
atmospheric models.

In addition to remote mechanisms, the impact of local meridional winds and upwelling on
the Benguela SST has been discussed by Large and Danabasoglu (2006). They show that the
warm bias in eastern boundary upwelling regions in the CCSM are due to a combination of weak
ocean currents, weak upwelling, weak along-shore wind, too little stratus cloud, and neighboring
mountainous regions. The impact of local upwelling is emphasized by Grodsky et al. (2011) who
argue that adequate representation of the magnitude of the southerly Benguela low-level wind jet
(Nicholson, 2010) is crucial for maintaining the zonal sea level gradient in the coastal ocean, and
thus cold water transport by the coastal jet of Benguela Current. Independent of its origin, any
warm SST bias in the Benguela region may grow and expand via the positive feedbacks from
marine stratocumulus clouds (e.g. Mechoso et al., 1995).

Observed changes in Atlantic SST occur on a wide range of time scales. On multidecadal
timescales, the Atlantic Multidecadal Oscillation (Enfield et al., 2001) is mostly confined to the
North Atlantic, and possibly reflects changes in the strength of the Atlantic Meridional Overturning Circulation. At decadal timescales, the out-of-phase variations of SST in the northern and southern tropical Atlantic are driven by the wind-evaporation-SST feedback in the trade winds (Carton et al., 1996; Chang et al., 1997). At interannual timescales, remote impacts of the El Niño-Southern Oscillation (ENSO; Enfield and Mayer, 1997; Deser et al., 2006) and the North Atlantic Oscillation (Czaja et al., 2002) produce different SST responses in the northern and southern tropical sectors and thus contribute to the observed lack of coherence between SST variations in the two regions. Deser et al. (2006) found that in CCSM3 the correlation between the tropical Pacific Nino3 SST index in December-February (DJF) and tropical North Atlantic SST anomalies in March-May was similar to the correlation from observations. However, in CCSM3 the tropical Pacific Nino3 index had a high correlation with SSTs in the deep tropical Atlantic (between 3°N and 10°S) something not observed in the same analysis from observations (Deser et al., 2006).

In addition to a tropical Atlantic lagged response to ENSO, there is increasing evidence (based on observations and modeling studies) that the equatorial Atlantic SST anomalies are anti-correlated with tropical Pacific SSTs as an intrinsic phenomenon. This tropical Pacific-Atlantic inter-basin anti-correlation is also evident (and more widespread) in the sea level pressure (SLP) anomalies. Some studies that make this evident are: Giannini et al. 2000; Wang, 2006; Munoz et al. 2008; Garcia-Serrano et al. 2008; Polo et al. 2008; Losada et al. 2010; and Ding et al. 2011. For example, it is observed from Figure 8 from Polo et al (2008) that the equatorial Atlantic SST anomalies lead the Pacific Nino-3 index with an anti-correlation greatest at 8 months. Also, it is observed from Figure 14 from Munoz et al (2008) that the inter-basin gradients of SST and SLP are highly correlated to the variations in the strength of the Caribbean
low-level jet. Explanations of this inter-basin gradient mode of variability has been provided by Wang (2006), Garcia-Serrano et al. (2008), Polo et al. (2008), Losada et al. (2010), and Ding et al. (2011).

The variability of tropical Atlantic SSTs has been observed to have a few main modes that are predominant at different times of the year (Servain et al. 1990; Servain et al. 2003). One of the modes of variability is the so-called meridional mode or interhemispheric mode and is the dominant mode in the boreal spring (Servain et al. 1998; Mahajan et al. 2010). The meridional mode is characterized by a north-south gradient of SST anomalies from one subtropical region to its counterpart in the other hemisphere, and a pattern of surface wind anomalies from the colder sub tropics to the warmer sub tropics (Nobre and Shukla, 1996). Another mode of variability is the so-called zonal mode or Atlantic Niño and is predominant in the boreal summer (Tokinaga and Xie, 2011; Carton and Huang, 1994; Zebiak, 1993; Shannon et al. 1986). Yet, the identification of Tropical Atlantic modes of SST variability has also benefited from the use of statistical techniques such as rotated Empirical Orthogonal Functions (rEOFs). The resulting modes using rEOFs have been referred to as: the southern tropical Atlantic (STA) pattern; the northern tropical Atlantic (NTA) pattern; and the southern subtropical Atlantic (SSA) pattern (Huang and Shukla, 2005; Bates, 2008). Previous studies have analyzed these modes and the dynamics and thermodynamics that explain their variability (Bates, 2008, 2010; Huang and Shukla, 2005; Florenchie et al. 2004; Chang et al. 1997; Carton et al. 1996; Shannon et al. 1987).

These periodic changes of SST in the northern and southern tropical Atlantic displace the ITCZ and affect rainfall over surrounding continents in the northeastern Brazil and African Sahel (see e.g. Xie and Carton, 2004 and references therein).

Observation-based analyses of SST variability in the tropical Atlantic indicate that the
standard deviation of anomalous SST is strongest in areas adjacent to the western coast of Africa (Doi et al., 2010) and reaches a maximum in the Angola-Benguela frontal zone (referred to as the Benguela region in this paper; see e.g. Florenchie et al., 2003). SSTs in the Benguela region are affected by local and remote impacts according to observations and model simulations (Lubbecke et al. 2011; Richter et al. 2010; Rouault, 2010; Florenchie et al. 2003). Florenchie et al. (2003) have suggested a link between the Benguela warm events and weakening of the zonal equatorial winds 1 to 2 months in advance, which remotely impact the Benguela region via Kelvin waves propagating eastward along the Equator and further south along the coast. Richter et al. (2010) have demonstrated based on observations and model simulations that impact of local upwelling on Benguela SST is comparable to the remote impact of the Equatorial winds.

The recent availability of the 4th version of the Community Climate System Model (CCSM4) provides an opportunity to re-assess the status of the simulation of the tropical Atlantic by one of the leading coupled climate models. In this study the main aspects of the tropical Atlantic Ocean are analyzed, expanding upon the CCSM3 analysis of the tropical Atlantic by Deser et al. (2006) and other CCSM3 studies. To evaluate how well the CCSM4 is able to simulate the tropical Atlantic seasonal cycle we provide some standard diagnostics for comparison to the CCSM3 (Section 3) and to observations. We then discuss how the CCSM4 compares to observations relative to the well-known and important features described above. These include analysis of the Atlantic Warm Pools (Section 4), the modes of SST variability (Section 5), and a detailed analysis of the upper ocean heat budget in the Benguela region (Section 6). A summary and discussion are presented as the final section of the manuscript.

2. Data and Methods
a. Model data

The main CCSM4 data set analyzed is an ensemble of five 20th Century (20C) CCSM4 simulations. (These were archived as b40.20th.track1.1deg.[005-009], and are referred here as runs R005-R009). These CCSM4 ensemble simulations were run in the same manner with the exception of their initial state. Each of the CCSM4 members was initialized from the 1850 control simulation, with the five initializations chosen to represent different states of the Atlantic meridional overturning circulation. The period used for the analyses in this study span 1950 to 2005 of the CCSM4 20C simulations. Complete descriptions of these simulations are provided by Gent et al. (2011). The CCSM4 20C ensemble simulations were compared to observations, to a CCSM3 20C ensemble mean, and to an ocean-sea ice hindcast experiment forced by observed winds.

The ocean-sea ice hindcast simulation (from here on “POP-CORE”) was conducted using the 1-degree horizontal resolution version of the CCSM4 ocean model coupled to an active freely evolving dynamic-thermodynamic sea ice model (CICE). Monthly climatological river runoff in the POP-CORE is based on Dai et al. (2003) discharge estimates. The model was forced by four cycles of the CORE v2 Inter-Annual Forcing data (Large and Yeager, 2009), which span the 60-year period from 1948 to 2007. The POP-CORE data analyzed is from the fourth (last) forcing cycle.

There were eight ensemble members in the CCSM3 simulations, and when possible, all eight members are used. One of the ensemble members did not contain monthly output and, therefore, is not used in the annual cycle comparisons. The CCSM3 ensemble members were initialized from the 1870 control simulation (the control was switched to 1850 for CCSM4) at
20-year intervals with no tie to a physical feature (see Gent et al., 2006, Table 1 for details on CCSM3 simulations). The CCSM3 simulations ended in December of 1999; therefore, the 20-year mean used in this study is from 1980-1999.

Differences between CCSM3 and CCSM4 are expected due to various changes in the model physics and the spin-up and tuning procedures, which are all intended to produce a more realistic model. Both sets of simulations are the nominal 1-degree resolution in all components; however, this resolution has increased in the atmospheric and land components from approximately 1.4 degrees in CCSM3 to approximately 1 degree in CCSM4. The depth resolution of the ocean model has also increased from 40 levels in CCSM3 to 60 levels in CCSM4, with the majority of the additional layers in the upper ocean. The dynamical core is different in the atmospheric component (Neale et al. 2011, this issue); the ice model component, with a new radiation scheme and different albedo values, produces different sea ice cover (Holland et al. 2011, this issue), and; the addition of new ocean parameterizations include more ocean physics than were present in CCSM3 (Danabasoglu et al. 2011, this issue). Due to differences in model tuning and spin-up (Gent et al., 2011, this issue), the ocean in the CCSM3 20C simulations loses heat over the length of the run, while in the CCSM4 simulations it more realistically gain heat.

A control CCSM4 run was also used to evaluate the variability of the heat budget in the Benguela region. The monthly averaged fields were from the 1-degree 1850 control run with 1300 years of simulation forced by fixed pre-industrial levels of ozone, solar, volcanic, greenhouse gases, carbon, and sulfur dioxide/trioxide (archived as b40.1850.track1.1deg.006). The Benguela heat budget analysis focuses on data from a 97-year period (model years 863-959). A sensitivity examination has been carried out to ensure that the climatology of this particular
period is similar to that of later periods.

250 \textit{b. Observational data}

For the comparisons of the mean and seasonal cycle of SST and wind stress, the observations used include: SST from Hurrell et al. (2008), and wind stress were derived from the Coordinated Ocean-ice Reference Experiments (CORE; Griffies et al. (2009), Large and Yeager (2009)) data set. The time periods used from these observations were 1980-1999 when comparing with CCSM3 ensemble means, and 1986-2005 when comparing with CCSM4 ensemble means. These datasets and time periods are chosen since they are used in the literature to assess the CCSM, and also used in publications within this issue to assess ocean behavior.

For analyses of the Atlantic warm pools, two observational data sets with subsurface temperatures were used: the World Ocean Atlas 2009 (WOA09; Locarnini et al. 2010; Levitus et al. 1998; hereafter “Levitus”), and the Ishii et al (2003, 2006; hereafter “Ishi”) temperature data sets. The Levitus 12-month climatology has been used in many other studies of the mean state of the ocean, including the Intra-Americas Sea warm pool study of Wang and Enfield (2003). The Levitus climatology provides climatological temperature data for the twelve calendar months based on available observations during the period from the year 1773 to the year 2008 (as indicated in the metadata). Also used is the observational data set developed by Ishii et al (2003, 2006) with monthly temperature data interpolated to a 1x1 degree grid. For the warm pool estimates, the averages from the Ishii et al product were computed based on the period 1950-2005. Both of these observational data sets have the same horizontal 1x1 degree grid, and the same vertical levels, with data at the surface, 10m, 20m, 30m, 50m, 75m, 100m, 125m, 150m, 200m, 250m, and at 100m intervals between 300m and 700m.
The Extended Reconstructed SST version 3b (ERSSTv3b; Smith et al., 2008) observational data set is used to compute rotated Empirical Orthogonal Functions (rEOFs) of SSTs.

c. Methods

For the warm pool analysis, since the vertical resolution of the model data is finer than that of the observational data sets, the model data were interpolated to the vertical levels of the observations before calculating the depth of the 28.5°C isotherm \((Z_{28.5})\) as a metric of the warm pool. Once \(Z_{28.5}\) was calculated, the result was interpolated to the horizontal grid of the observational products at each month. These monthly values of Atlantic warm pool volume were used to calculate the warm pool mean, seasonal cycle and other statistics, such as rank histograms.

Rank histograms from the time series of the Atlantic warm pool (WP) were computed according to Hamill (2001). To create the rank histogram, the five CCSM4 ensemble simulations are considered in addition to an observational estimate (either Ishii et al., or the POP-CORE simulation), thereby having 6 bins in the histogram. For each time step in the WP time series, the ensemble member simulations are ranked from lowest to highest after including the estimate from observations. As explained by Hamill (2001) if there is equal probability that the observation will fall in each bin, then the histogram should be uniformly distributed or flat, and one can conclude that on the average, the ensemble spread represents correctly the uncertainty. However, if the histogram is distributed non-uniformly one can refer to either underdispersion or overdispersion of the ensemble (Hamill, 2001).

To compare the main modes of tropical Atlantic SST variability, rotated Empirical
Orthogonal Functions (Rotated EOF, or rEOF) were computed from the CCSM4 20C simulations, the POP-CORE hindcast and the ERSSTv3b observational product. The area of interest is the Atlantic Ocean between 30°S and 30°N, and the analysis is performed for the period 1948-2005, the era common to these data sets. In contrast to many previous studies, the Caribbean and Gulf of Mexico SSTs are included in these EOF analyses of the tropical Atlantic. First, an EOF analysis is performed on the area-weighted, detrended, monthly anomaly time series of SST, to focus on internal variability and reduce the impact of secular warming trends. The EOFs and their Principal Components (PCs) are renormalized so that the PCs have unit variance and the EOFs carry the standard deviation. Then a varimax rotation is applied to the dominant 10 EOFs. The rotation technique removes the orthogonality constraint on the EOFs and leads to more localized spatial patterns that might be easier to interpret in terms of dynamical processes (Richman, 1986; Dommenget and Latif, 2002).

A heat budget analysis of the upper ocean layer in the Benguela region was performed in order to quantify relative contributions of the air-sea heat fluxes versus heat advection terms. To evaluate relative impacts of heat advection and the surface fluxes on anomalous heat content we focus on the relationship between the terms of the vertically integrated heat balance equation (1) spatially averaged over the Benguela region. In common notations the vertically-integrated heat balance equation is:

\[
\begin{align*}
C_p \rho \int_{z=0}^{z=H} \frac{\partial T}{\partial t} dz = & C_p \rho \int_{z=0}^{z=H} \left( -u \frac{\partial T}{\partial x} - v \frac{\partial T}{\partial y} - w \frac{\partial T}{\partial z} \right) dz + NSHF - VDIFF(z = H) + R, \\
\end{align*}
\]

where \( NSHF \) is the net surface heat flux, and vertical integration is taken down to \( H=80m \). In this region \( H=80m \) is below the mixed layer year-round. We focus only on the terms available in the history files of CCSM4 output, and so the vertical diffusion \( VDIFF(z = H) \) is combined in \( R \) with other unresolved terms that include lateral diffusion, diffusion introduced by
the mixed layer model, and errors due to the use of monthly (instead of model) sampling to
calculate the time derivative in (1). Heat advection terms in the right hand side of (1) are
computed on the original model grid using the POP numerics.

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3. Long-term mean and seasonal cycle

a. Time mean of wind stress and SST

In this section we perform standard diagnostics used to assess previous versions of the
CCSM in order to quantify progress in the CCSM4. Figure 1 shows the mean state of zonal and
meridional wind stress from observations, and the difference with CCSM3 and CCSM4. The
overall pattern of easterlies throughout the tropics with centers near 15°S (southeasterlies) and
15°N (northeasterlies) is captured by both CCSM4 and CCSM3 (Fig. 1). The direction of the
wind stress, for the most part, seems to be correct in the model. Yet the magnitude of the wind
stress is the main difference between the model and the observations as observed in Fig. 1b-c.

As in the CCSM3, the CCSM4 exhibits wind stress smaller in magnitude than the CORE
wind stress throughout the equatorial (deep tropics) region (Fig. 1b-c). In the western basin this
is mostly due to weakened easterlies, and in the eastern basin weakened southerlies are
responsible for the decrease in magnitude. Both of these differences are related a reduced
upwelling in the east. From the model-observations differences in SST (Fig. 2) it is observed that
there are warm SST biases along the Equator and in the tropical southeastern Atlantic. Along the
Equator, observations indicate a west-to-east warm-to-cold SST gradient. CCSM3 exhibited a
zonal SST gradient opposite to that from observations. On the other hand, the CCSM4 shows
significant improvement in the SST gradient along the Equator with SSTs in the western basin
closer to those from observations (Fig. 2c).

In the regions of strongest easterlies (centered around 15°N/S), the wind stress in CCSM4 is enhanced over CORE with the largest differences in the eastern basin near the African continent and in the southern Caribbean region. In general, these differences have been reduced in CCSM4 compared to CCSM3 (Fig. 1); yet they still contribute to biases in the SSTs (Fig. 2). For example, the largest SST differences within the 30°S to 30°N domain are found in the southern Caribbean Sea and in the southeastern basin along the coast of Africa (Fig. 2a-b). Nonetheless, as with wind stress, these SST differences have improved in CCSM4. In the southern Caribbean Sea the SST model-observations difference has improved from -4.0°C (in CCSM3) to -2.5°C (in CCSM4). In the southeastern basin near the African coast the SST difference has increased from 6.5°C (in CCSM3) to 7.5°C (in CCSM4) approximately.

In the tropical Atlantic (as in the global ocean), a shift to warmer surface ocean temperatures is noted in the CCSM4 versus CCSM3 (Danabasoglu et al, 2011). (This shift is mostly due to the spinup procedure effects described in the data and methods section). The mean values from 30°S to 30°N of the tropical Atlantic SST difference from observations (excluding the Mediterranean Sea) are -0.50°C for CCSM3 and 0.41°C for CCSM4, reflecting the warming shift. The root mean square (RMS) error of the SST differences over the same region does show improvement with a value of 1.65°C for CCSM3 and 1.20°C for CCSM4. Therefore, the improvements in TNA SSTs described above are influenced by both the overall warming in the transition from CCSM3 to CCSM4, and by the improvements in the mean wind stress field.

b. Seasonal cycle of wind stress and equatorial SST

In Figure 3, we present the seasonal cycle of wind stress from CORE observations (left
panels) along with the difference from CCSM4 (right panels). In all seasons, wind stress in CCSM4 is too weak in the deep tropics and too strong at higher tropical latitudes. The largest and most widespread of these differences occurs during the December-February (DJF) and March-May (MAM) seasons, corresponding to the seasons when the equatorial easterlies are the weakest and monsoonal flow in the Gulf of Guinea the strongest. The weakened easterlies are related to the incorrect SST gradient along the equator, and perhaps the model is under-representing the air-sea temperature difference driving the monsoonal flow. In MAM, a large difference in the southerly wind stress occurs in the central basin just north of the equator. This difference is due to the displacement of largest cross equatorial flow westward in CCSM4; in CORE this maximum occurs in the central basin.

In all of the areas described above, the mean bias of wind stress influences the seasonal bias quite strongly. Therefore, we also compare the wind stress seasonal departure from the annual mean (figures not shown). Compared to observations the amplitude of the zonal wind stress in the Caribbean Sea is weaker in June-August (JJA) and stronger in September-November (SON). In the TNA centered at 25°N and next to the African coast, where northeasterly winds dominate, we find the zonal wind stress difference to be the largest in DJF and also present in MAM, and the meridional wind stress weakened in DJF and strengthened in JJA. In the TSA, we find the same pattern of southerlies (too weak in the Gulf of Guinea and too strong to the south) to be present in DJF, with opposite sign differences occurring in JJA (southerlies too strong in the Gulf of Guinea and too weak to the south).

Figure 4 shows the SST monthly departure from the annual mean, which helps in assessing the phasing and magnitude of the seasonal cycle. The seasonal cycle of SST along the equator in the CCSM4 has improved greatly over the CCSM3 (Fig. 4), most likely due to
improvements in the wind stress forcing from the atmosphere. Most notable of these improvements is found in the warm phase in late boreal winter and spring. In CCSM3, the warm phase begins in the western basin and propagates eastward, neither of which occur in the observations, and are corrected in the CCSM4. Though the phasing in the CCSM4 lags observations by approximately half a month, the character and magnitude are quite similar (Fig. 4).

c. Seasonal cycle of the Atlantic Warm Pool

Figure 5 (a-d) shows the month of the calendar year when the 28.5°C isotherm is deepest in the long-term mean. When the CCSM4 ensemble mean is compared to the Levitus and the Ishii data sets, a better agreement in the WP is obtained between CCSM4 and Ishii. For example, in both the CCSM4 and Ishii, the Z28.5 is present throughout the TSA; in the TNA the Z28.5 crosses the basin at about 10°N in both products. Yet, the warm pool extent and timing in the POP-CORE hindcast seems to match better the Ishii observational estimate, indicating that some of the major differences between CCSM4 and the observations are in the coupled framework (not strictly in the ocean component of CCSM4: POP).

The seasonal cycle of the warm pool volume in the tropical Atlantic is shown in Fig. 5e. In all products there are relative maxima in boreal spring and in boreal summer indicating greater extent in April and in September, respectively. The peak in April corresponds to the TSA warm pool (TSA-WP), whereas the peak in September corresponds to the TNA warm pool (TNA-WP) (Figs. 5a-d). The TSA-WP in observations is smaller than its northern counterpart, the TNA-WP. The major differences in the TNA-WP and the TSA-WP between model and observations are described below.
In the TNA, the main differences in the CCSM4 and the POP-CORE hindcast are in the Caribbean Sea. Both the CCSM4 and the POP-CORE show a Z28.5 in the northern Caribbean Sea deepest in November. However, in the Levitus and Ishii observational products the Caribbean Z28.5 is deepest in October. Figure 6 shows the mean warm pool in September, when the TNA warm pool has its greatest volume (Fig. 5e). Among the similarities in all products is the presence of a warm pool deepest between Cuba and Central America with Z28.5 greater than 60 meters. The POP-CORE simulation and the Ishii observational estimate have September warm pools very similar in spatial extent; that is to say, extending to the southern Caribbean Sea, to the northeast of the Antilles, and across the basin approximately between 5°N and 15°N. Yet, in CCSM4 as in Levitus there is a lack of subsurface temperatures greater than 28.5°C in the southern Caribbean Sea. In the CCSM4 this is related to the strong zonal wind stress along the southern Caribbean Sea (as observed in Fig. 3). This stronger Caribbean low-level jet would induce stronger upwelling in the southern Caribbean Sea thereby simulating temperatures colder than observations, and reducing the volume of the warm pool.

In the TSA, a main difference between the Z28.5 from observations and from CCSM4 is the timing of the deepest Z28.5 in the TSA (Fig. 5). In the Gulf of Guinea and Benguela regions the observations have the deepest Z28.5 in March and April respectively, while the CCSM4 has its deepest Z28.5 in May. This indicates that the CCSM4 is staying warmer than observations during the decay of the TSA warm pool in the eastern TSA. April is the month of greatest volume of Z28.5 in the TSA (Figs. 5, 7). Yet, as observed from the seasonal cycle in Fig. 5e, in the CCSM4 the WP volume in April surpasses that of September (whereas in observations is the WP is largest in boreal summer-fall peaking in September). This greater TSA warm pool during boreal spring seems to be associated with the pattern of wind stress biases in CCSM4 from
December to March. As observed from Fig. 4, the wind stress magnitude in CCSM4 is weaker than in observations in an area collocated with the TSA warm pool in the boreal springtime (as in Fig. 7a). The weaker wind stress in the TSA reduces the evaporative cooling in the region. Furthermore, in the tropical southeastern Atlantic the meridional wind stress in CCSM4 is weaker than in observations thereby reducing the upwelling along the eastern boundary, and reducing the intensity of the Benguela Current, Angola Current and the Atlantic South Equatorial Current systems which normally advect cold water from the south.

4. Tropical Atlantic variability

a. Atlantic Warm Pools

The September time series of TNA volume (km3) encompassed by the 28.5°C isotherm (i.e., the TNA-WP) are shown in Fig. 6e. The volume was calculated from the 5°N latitude to the north and across the basin. The Ishii observational product shows anomalously large volumes in the late 1980s and after 1994. Between 1950 and 2005 there are years of decreased observed volume in the mid-1970s (minimum in 1974) and the mid-1980s (minimum in 1984). The POP-CORE hindcast has a similar evolution than Ishii with relative minima in the mid-1970s and in the mid-1980s. After the 1970s the CCSM4 ensemble spread encompasses the observational estimate or the POP-CORE hindcast.

The April time series of TSA volume (km3) encompassed by the 28.5°C isotherm (i.e., the TSA-WP) are shown in Fig. 7e. The volume was calculated from the 5°N latitude to the south and across the basin. The TSA-WP volume has periods of relative minima in the early 1950s and the late 1970s. The POP-CORE hindcast has a similar evolution than Ishii. However,
because of the warm bias, the CCSM4 ensemble simulations have a TSA-WP much larger than observations and their spread does not encompass the observational estimates.

Even though the time series of the CCSM4 ensemble simulations do not correspond to the time series from observations, we can compare basic statistics such as the trend, the standard deviation, and the auto-correlation of the WP time series. Furthermore, a rank histogram was calculated to determine if the CCSM4 ensemble has undervariability with respect to observations and with respect to the POP-CORE simulation.

The trend of the WP indices of each product (including each ensemble simulation) is shown in Table 1. The trend was calculated from the period 1950 to 2005. The September TNA-WP in the CCSM4 20C simulations has a greater trend than that of the observational estimate, and than that of the POP-CORE simulation. However, the April TSA-WP trends in CCSM4 are about the same as those from observations and the POP-CORE simulation. To analyze other statistics the long-term mean and the trends in Table 1 were removed from the corresponding WP time series. Then, the standard deviation, auto-correlation and rank histograms were computed from these detrended time series.

The standard deviations of the WP indices are shown in Table 2. The WP indices from the POP-CORE simulation have the greatest standard deviation of all products. The TNA-WP in September has lower standard deviation in CCSM4 than in Ishii, but the TSA-WP in April has greater standard deviation in CCSM4 than in Ishii. From a rank histogram analysis (Fig. 8) it can be observed that the CCSM4 TNA-WP has underdispersion with respect to both the Ishii WP and the POP-CORE WP. Yet, the POP-CORE may have excessive interannual variability as indicated by the low auto-correlation values in Table 3. For the TSA-WP, even though the CCSM4 has undervariability with respect to the POP-CORE, the variability with respect to Ishii
is unclear (Fig. 8).

**b. Modes of SST variability**

The dominant rEOFs of SSTs from observations, the CCSM4 ensemble mean, and the POP-CORE experiment are shown in Fig. 9, while the spectra of the associated rotated Principal Components (rPCs) are shown in Fig. ES-1. The associated levels of variance accounted for by each of these modes (both in relative and absolute sense) are tabulated in Table 4. The dominant modes in observations are the so-called patterns of the Southern Tropical Atlantic (STA), Northern Tropical Atlantic (NTA) and Subtropical South Atlantic (SSA) modes (e.g., Huang et al. 2004; Bates 2008, 2010). These modes are represented by the dominant rEOFs of the CCSM4 ensemble members and the POP-CORE experiment, although they account for different levels of variance (Table 4). Note that the rEOFs are not very well separated in terms of explained variance, so the relative order of the modes is not of critical importance.

The NTA and SSA modes are well represented in CCSM4 (Fig. 9, lower panels) with centers of action located off West Africa and in the central South Atlantic, respectively, as in observations. The domain-averaged variance of the NTA is underestimated by the ensemble members (0.014 vs. 0.022°C²), making it the weakest mode in all but one of the ensemble members (R008). Nonetheless, the variance accounted for by the SSA mode is well represented (0.019 vs. 0.018°C²). The spectral content of the rPCs of the NTA and SSA modes is consistent with an AR-1 process, as no significant spectral peaks are present in the ensemble mean, nor in the observations (Fig. ES-1). Only the NTA mode in the POP-CORE run displays some enhanced energy at the annual frequency. Lagged correlations between the rPCs and the wind stress (Fig. 10) shows that both the NTA and SSA modes (lower panels) are associated with a
weakening of the trade winds, in agreement with observations (e.g., Barreiro et al 2004; Bates 2008). The correlation peaks when wind stress leads by one month. The NTA lead-lag correlations are consistent with the so-called wind-evaporation-SST (WES) feedback that has been proposed as the dominant mechanism for these modes (Chang et al., 1997; Sterl and Hazeleger, 2003); a negative wind stress perturbation reduces evaporation and evaporative cooling, inducing a positive SST anomaly that amplifies the wind stress anomaly (Moura and Shukla, 1981). The NTA mode is also associated with a strengthening of the southeasterly trade winds in the tropical South Atlantic. No such inter-hemispheric relationship is seen in the SSA mode. Instead, the SSA mode appears to be related to in-phase variability in the tropical South Pacific (Fig. 11).

The largest variability in the observational STA mode (a standard deviation of close to 1ºC) is found off Angola (Fig. 9). The signal attenuates northward, and achieves amplitudes below 0.3ºC along the equator. In the POP-CORE run the emphasis of the STA mode is on the equatorial region, lacking energetic SST variability in the Benguela upwelling region. The POP-CORE equatorial emphasis is probably due to an underestimation of wind stress variability resulting from the CORE forcing (not shown; Grodsky et al. 2011). Yet the CCSM4 model has two separate modes in the tropical South Atlantic. The dominant rEOF in all but one (R007) of the ensemble members (here called the STA-BG mode) is characterized by strong (>0.6ºC) SST variability in the southern segment of the Benguela upwelling zone, off the coast of Namibia. The variability extends northward, but does not have an equatorial tongue. In contrast, variability in equatorial SSTs (>0.4 ºC) in the CCSM4 ensemble runs is captured by a different mode, indicated here as STA-EQ.

Figure 10 (top left) shows that a warm phase of the STA-BG mode is related to a
northerly wind stress anomaly that peaks one month in advance, in agreement with the model study of Richter et al. (2010). The spectrum of the corresponding rPC cannot be distinguished from a red-noise process (Fig. ES-1). In the STA-BG mode, the area of highest SST variability also corresponds to the region of maximum mean meridional wind stress and a weak wind stress bias (from December to May) in the model, compared to observations (Fig. 4). From the global perspective (Fig. 11), the STA-BG mode seems to be related to the tropical southeastern Pacific climate.

The STA-EQ mode is highly correlated to wind stress fluctuations in the central equatorial Atlantic that peak about one month earlier (Fig. 10). Also, the wind stress in the central tropical South Atlantic responds to the SST anomaly with a one-month lag. The spectrum of the ensemble-mean rPCs display some enhanced energy at a period of about 9 months, a feature that does not seem to be present in the rPC of the observational STA mode (Fig. ES1). The CCSM4 STA-EQ mode is collocated with a decrease in SLP anomalies. The SLP anomalies in the tropical Atlantic are anti-correlated with SLP anomalies in the tropical Pacific (Fig. 11). Also, the anti-correlation with the Nino-3 index peaks nine months after the peak of the STA-EQ mode.

c. **Heat budget of the Benguela region**

For the purposes of this section the model Benguela index is defined based on the variability of the heat content rate of change (HCR). In the east, where the thermocline is shallower, the regions of high HCR variability are roughly collocated with regions of high SST variability. This is in contrast to the western equatorial Atlantic where SST variability is relatively weak regardless of rather strong HCR variability. The model Benguela region extends
from 20°S to the northern edge of the time mean SST front at 13°S (where standard deviation of HCR ≥ 250 Wm-2), and from 9°E to the coast of southwestern Africa (Fig. 12). Although the observed SST front is mostly confined to the meridional extent of the Benguela region, the model SST front is stretched further south (Fig. 12). As a result, the region of high SST variability is also stretched southward. The model Benguela region, as defined above, covers only the northern part of the high model SST variability zone, but is influenced by both the STA-BG and STA-EQ modes (Fig. 9). This selected northern part is close to the observed region of high SST variability in the Angola-Benguela front (see e.g. Florenchie et al., 2003).

In the CCSM4, anomalous SST events in the Benguela region last for approximately 4 months (Fig. 13). Heat balance analysis based on equation (1) suggest that anomalous HCR varies in phase with instantaneous anomalous heat advection and surface flux. In general this is confirmed by correlation analysis of the region-averaged time series, which identifies ocean heat advection as the dominant contribution to the heat budget in Benguela region. The largest influence is provided by vertical heat advection (upwelling). The magnitude of its correlation with anomalous HCR at zero-lag exceeds 0.7; i.e., vertical advection accounts for 51% of the anomalous HCR variance. The second strongest contribution is from anomalous meridional heat advection (correlation of 0.5), which accounts for about 26% of the anomalous HCR variance. Anomalous meridional heat advection is dominated by an anomalous meridional current acting on the mean meridional gradient of temperature (as explained by Colberg and Reason, 2007b). The impact of zonal advection is weak in CCSM4 due to predominantly zonal orientation of the isotherms in the Angola-Benguela front. Local surface flux accounts for only 12% of the anomalous HCR variance. Surface flux also provides a weak negative feedback on anomalous SST in two months after the peak of HCR via latent heat flux.
To explore possible atmospheric forcing of anomalous heat advection in the Benguela region, the area-averaged time series of anomalous vertical and meridional advection have been lag-correlated with wind stress anomalies elsewhere (Fig. 14). This analysis illustrates that anomalously warm vertical advection in the Benguela region (reduced upwelling) occurs in phase with weakening of southeasterly trade winds. The maximum correlation is at zero lag, suggesting that impact of local upwelling dominates. This is evident in an anomalous cyclonic wind pattern driven by an anomalously weak South Atlantic subtropical high. The anomalous wind pattern includes a northerly (downwelling) component along the coast. Although the upwelling is attenuated along a major portion of the South African coast its impact on SST is stronger in the Benguela region where the thermocline shoals. An interesting (but not yet well understood) feature of the air pressure pattern is the area of anomalously high pressure over South Africa. A zonal gradient between the high over land and the low over the ocean further accelerates anomalous downwelling winds along the coast. An increase in air pressure over the land during warm Benguela events may be linked to cooling of the land due to above-average rainfall along the coast of Angola and Namibia as observed by Rouault et al. (2003). But in CCSM4 the Benguela SST does not correlate significantly with either land temperature or rainfall.

The wind pattern corresponding to anomalous meridional advection (Fig. 14b) is different from that for anomalous vertical advection (Fig. 14a) in many aspects. For warm Benguela events the area of weaker southeasterly trades does not cover the Benguela region itself. But, there is significant correlation with the zonal Equatorial winds that lead the meridional advection by about a month. This suggests that non-local processes translating wind impacts from the Equatorial region (such as equatorial and coastal Kelvin waves) are responsible for anomalous
meridional heat advection in the Benguela region.

7. Summary and Discussion

In this study we assess the CCSM4 with respect to some of the main aspects of the tropical Atlantic mean climate and its variability. We have performed a suite of analyses to address differences and improvements achieved by the CCSM4 model in the tropical Atlantic Ocean as compared to observations, an ocean-only POP simulation forced with the CORE data set (referred here as the POP-CORE hindcast) and in some occasions CCSM3. Various analyses are presented and discussed covering the main differences in the surface fields, the structure and variability of the tropical Atlantic warm pools, and the variability of sea surface temperatures. The variability of the heat budget in the Benguela region was analyzed from a control CCSM4 simulation. The analyses and results presented and discussed will be useful for further evaluations of CCSM4 simulations of the tropical Atlantic climate and for predictive studies of this region. In the following we present a summary of the results.

Both improvements and degradations in the simulated SST occur in the transition from CCSM3 to CCSM4. Some of these differences are related to the changes in the wind stress biases. In CCSM4 there has been a large reduction in the cold biases centered at 20°N as well as in the Caribbean Sea and Gulf of Mexico. This reduction of the cold bias in CCSM4 represents a significant improvement in the simulation of the tropical North Atlantic warm pool (TNA-WP) given that in CCSM3 the TNA-WP was non-existent (Misra et al. 2009). The CCSM3 warm bias in the southeastern tropical Atlantic upwelling regions was attributed by Chang et al (2007) to a weak bias in the equatorial easterlies. However, even though the magnitude of equatorial easterlies has improved in CCSM4 (over CCSM3) the warm bias in the southeastern tropical
Atlantic is worse in CCSM4 compared to CCSM3. Both the reduction of the TNA cold bias and the worsening of the TSA warm bias may be related to the wind stress. They could also be partly due to the overall warming in CCSM4 compared to CCSM3.

The warm pool is analyzed by its vertical structure throughout the year in both the tropical North Atlantic (TNA) including the Intra-Americas Sea (IAS), and the tropical South Atlantic (TSA). The volume of the WP in the tropical South Atlantic (TSA-WP) peaks in April and in the tropical North Atlantic (TNA-WP) peaks in September. The timing of the WPs in CCSM4 is similar to that of the observations, although the vertical structure indicates that the TSA-WP pool is deeper in the CCSM4 than in observations. This deeper TSA-WP is related to the CCSM4 warm bias in the TSA region, a common challenge to many coupled models.

Regardless of the warm bias, the ensemble spread of the TSA-WP seems to correctly represent the uncertainty or spread of the ensemble.

In the Intra-Americas Sea, the CCSM4 warm pool is smaller than in observations, as a result of the CCSM4 cold bias (and a stronger low-level jet) in the southern Caribbean Sea and to the northeast of the Caribbean Sea. The CCSM smaller warm pool limits the simulation of tropical cyclones by regional models with boundary conditions obtained from CCSM. In particular, Holland et al (2010) found that given a colder TNA, the CCSM develops a wind vertical shear that is much stronger than observations and therefore inhibits the formation of tropical cyclones. In addition to the CCSM4 simulating a TNA-WP smaller than POP-CORE, the ensemble spread of the TNA-WP is underdispersed compared to the observations. Yet, the trend in the TNA-WP is greater in CCSM4 than in observations, mostly because of a near-monotonic increase in warm pool volume in CCSM4 as opposed to the observational estimates which show large warm pools in the 1950’s and 1960’s.
To assess the CCSM4 variability of tropical Atlantic SSTs, rotated Empirical Orthogonal Functions (rEOFs) were applied to SST fields of the various CCSM4 ensemble simulations, the POP-CORE simulation and to an observational data set for the period 1948-2005. The spatial patterns of the main modes of variability in the model are similar to those from the observations. The lead-lag relationship of the rotated Principal Components (PCs) with wind stress and SST anomalies indicate that the mechanisms of the NTA and SSA modes are consistent with the findings of previous studies. However, in the tropical South Atlantic the coupled model separates the SST variability as two different modes of variability: one centered in the Benguela region (STA-BG) and another one centered along the Equator (STA-EQ).

A break-up of the TSA variability in a pattern containing equatorial SST variability and another one capturing variability off Angola seems to be characteristic of at least one other coupled model. Based on analysis of the COLA AGCM coupled to the Poseidon ocean model, Huang et al. (2004) ascribe this disconnection between equatorial and off-equatorial STA variability to: 1) an artificial warm pool related to the Intertropical Convergence Zone (ITCZ) having two preferred locations, and 2) to two different timescales of SST variability. As a parallel, during March-May, when observations indicate the Benguela variability to be most active (Florenchie et al. 2004), the seasonal wind stress bias in CCSM4 is a reduction in magnitude and variability of these southerly coastal winds.

Regardless, the STA-EQ mode in CCSM4 is characterized by tropical Atlantic SLP variability anti-correlated with tropical Pacific SLP variability and manifested as an inter-basin gradient of SLP anomalies (and to a lesser degree an inter-basin SST gradient although not statistically significant). Such an inter-basin gradient of SLP anomalies has been observed (Giannini et al. 2000; Wang, 2006; Munoz et al. 2008; Garcia-Serrano et al. 2008; Polo et al.
and is increasingly recognized as an intrinsic mode of tropical Atlantic variability that feeds back on tropical Pacific climate anomalies (Garcia-Serrano et al. 2008; Losada et al. 2010; and Ding et al. 2011).

Analysis of the model heat budget in the Benguela region suggests that anomalous vertical advection accounts for about 50% of the anomalous heat content rate (HCR) variance while the contribution by anomalous meridional heat advection is half as strong. Local surface flux accounts for only 12% of the anomalous HCR variance. The impact of zonal advection is weak. Anomalously warm vertical advection in the Benguela region (reduced upwelling) occurs in phase with the weakening of southeasterly trade winds. Correlation is maximum at zero lag suggesting that the impact of local upwelling dominates. In contrast to vertical heat advection the anomalous meridional heat advection is forced by zonal equatorial winds, which lead it by about a month. This suggests that non-local processes translating wind impacts from the Equatorial region (such as Equatorial and coastal Kelvin waves) are responsible for anomalous meridional heat advection in the Benguela region. In distinction from observations of Florenchie et al. (2003) this wave-based teleconnection is not the dominant mechanism of heat content variability in the Benguela region in CCSM4.
8. Acknowledgements

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FIGURE CAPTIONS

Figure 1: (top) Total mean CORE-based wind stress derived for the period 1980-2005. Model minus CORE-based difference for CCSM4 (middle) and CCSM3 (bottom). The period used for each model is explained in the text. Shading indicates wind stress magnitude and vectors indicate the direction. Units are N/m².

Figure 2: Model minus observations of mean sea surface temperature (°C) for (a) CCSM4 and (b) CCSM3. Panels a and b correspond to Figure 6 from Danabasoglu et al. (2011, this issue) with a focus on the tropical Atlantic. (c) Mean SST along the Equator from observations (black line, 1980-2005), CCSM4 (red line, 1986-2005), and CCSM3 (blue line, 1980-1999).

Figure 3: (left) Seasonal means of the CORE-based wind stress. (right) Seasonal differences of the CCSM4 wind stress minus CORE-based wind stress. The units are N/m², and the time period used for both CORE and CCSM4 is 1986-2005. Shading indicates magnitude, and the vectors are the direction.

Figure 4: Seasonal cycle of SST along the equator calculated as the mean of each month minus the total mean for a) observations, b) CCSM4, and (c) CCSM3. Units are °C, and the seasonal cycles are calculated over 1980-1999 for CCSM3, 1986-2005 for CCSM4, and 1980-2005 for observations.
Figure 5: (a-d) Horizontal distribution of the month of deepest 28.5°C isotherm from the long-term mean from 1950 to 2005. The numbers 1 to 12 correspond to the months from January to December. The Pacific data has been masked. Panel (a) corresponds to the CCSM4 ensemble mean. Panel (c) corresponds to the POP ocean model forced with CORE surface forcing. Panels (b) and (d) correspond to the observational products, Ishii and Levitus, respectively. (e) Seasonal cycle of the volume of the 28.5°C isotherm between 40°S-40°N.

Figure 6: The tropical North Atlantic (TNA) Warm Pool in September. (a-d) Mean depth (meters) of the 28.5°C isotherm in September. The CCSM4 ensemble mean (panel a) is the mean of five different simulations. (e) Time series of the volume (10^4 km^3) encompassed by the 28.5°C isotherm in September north of 5°N. The black line is the Ishii observational product; the blue line is the ocean POP simulation forced by CORE forcing; the red line is the CCSM4 ensemble mean with the ensemble spread in gray. The ensemble spread is the minimum and maximum value of any of the ensemble members.

Figure 7: The tropical South Atlantic (TSA) Warm Pool in April. (a-d) Mean depth (meters) of the 28.5°C isotherm in April. The CCSM4 ensemble mean (panel a) is the mean of five different simulations. (e) Time series of the volume (10^4 km^3) encompassed by the 28.5°C isotherm in April south of 5°N. The black line is the Ishii observational product; the blue line is the ocean POP simulation forced by CORE forcing; the red line is the CCSM4 ensemble mean with the ensemble spread in gray. The ensemble spread is the minimum and maximum value of any of the ensemble members.
Figure 8: Rank histograms of the CCSM4 ensemble spread against the POP ocean simulation forced by CORE (purple), and against the Ishii observational estimate (blue). The top panel corresponds to the index of the tropical North Atlantic (TNA) Warm Pool in September. The bottom panel corresponds to the index of the tropical South Atlantic (TSA) Warm Pool in April. The black line represents a uniform distribution.

Figure 9: Dominant rotated EOFs (rEOFs) of SST for the ERSSTv3b data set (left), the mean of the five 20C ensemble members of the CCSM4 (center), and the CORE-forced ocean-ice simulation (right). The rEOFs are based on a varimax rotation of the 10 dominant EOFs of detrended, area-weighted, monthly SST anomalies. The North Tropical Atlantic (NTA) and Subtropical South Atlantic (SSA) modes are found in all data sets. In CCSM4, the South Tropical Atlantic (STA) variability is represented by the STA-EQ and STA-BG modes, with SST variability in the equatorial region and the Benguela upwelling zone, respectively. The rEOFs carry the standard deviation. Negative, zero, and positive contours are thin dashed, thick solid, and thin solid, respectively, with contour interval of 0.1°C.

Figure 10: Statistical relationship between wind stress, SST, and the rPCs of the four dominant SST modes (STA-BG, STA-EQ, NTA and SSA) in the 20C ensemble member 005 (R005) of CCSM4 at lags of -1, 0 and 1 months. Negative lags indicate the wind stress and SST are leading. Correlations and regressions cover the 1948-2005 period. Shadings correspond to the correlation between the rPCs and wind stress magnitude (only values significant at the 95% level are shown). Arrows correspond to the linear regression of the rPCs on the wind stress components (legend arrows are 0.0089, 0.0069, 0.0084, 0.0087 N/m² for the STA-BG, STA-EQ,
NTA, and SSA columns, respectively). Contours correspond to the linear regression of the rPCs on the sea surface temperature (contour interval 0.1, negative, zero and positive contours are indicated with gray, thick black, and black, respectively). The rPCs are renormalized to have standard deviation of 1, so the regressions correspond to unit amplitude of the rPCs.

Figure 11: Lagged correlations between each of the rotated PCs and SST (shades) or SLP (contours) from the ensemble run R005. Rotated PCs lag by one month. The top of each panel indicates the corresponding mode as explained in the text. Shadings are correlations with SST (gray areas are not significant at the 95% level). Contours are correlations with SLP: black contours are positive and gray contours are negative, with a contour interval of 0.1 starting at ±0.1 (the zero contour is not plotted).

Figure 12: Standard deviation (STD) of anomalous heat content rate of change in the upper 80m (shading, Wm$^{-2}$), STD of anomalous SST (black contours), and time mean SST (gray contours). Box is the model Benguela region. All data are from the 1deg 1850 control run of CCSM4.

Figure 13: Lagged autocorrelation of anomalous SST and lagged correlation of anomalous heat content rate of change (HCR) with anomalous vertical (VERT), meridional (MER), zonal (ZON) heat advection, and anomalous net surface heat flux (NHF). All variables are spatially averaged over the Benguela region box and vertically integrated in the upper 80m.

Figure 14: (a) Correlation of anomalous vertical heat advection in the Benguela region with wind stress elsewhere at zero lag. Arrows show correlation with wind stress components. Correlations
significant at the 95% confidence level are shown in red. Temporal regression of anomalous vertical heat advection on anomalous mean sea level pressure elsewhere at zero lag is overlain as contours [mbar/(100Wm²)].

(b) Correlation of anomalous meridional heat advection in the Benguela region with wind stress leading by 1 month. Inlay shows the lagged correlation averaged over the equatorial box. Wind stress leads for positive lags.

Figure ES-1: Power spectra of the rotated PCs (rPCs) for the different modes featured in Figure 9. A 13-point Daniell filter is applied to smooth the spectra. For CCSM4 (black) the spectra are averaged over the five 20C ensemble members. The spectra of the ERSST (dark gray) and CORE (light gray) data sets are offset by factors 0.25 and 0.0625, respectively. The thin lines are 95% confidence limits, based on a best-fit AR-1 model to the time series, and a 2500-member ensemble of AR-1 processes with these same parameters.
TABLE CAPTIONS

Table 1. Linear trend \((10^4 \text{ km}^3/\text{year})\) of the September TNA and April TSA warm pool indices for each CCSM4 ensemble simulation, the POP ocean simulation forced by CORE, and the observational estimate of Ishii. The columns R005 through R009 correspond to the ensemble simulations.

Table 2. Standard deviation of the September TNA and April TSA warm pool indices for each CCSM4 ensemble simulation, the POP ocean simulation forced by CORE, and the observational estimate of Ishii. The columns R005 through R009 correspond to the ensemble simulations.

Table 3. Spearman \((R_s S)\) and Pearson \((R_s P)\) auto-correlations of the September TNA and April TSA warm pool indices for each CCSM4 ensemble simulation, the POP ocean simulation forced by CORE, and the observational estimate of Ishii. The columns R005 through R009 correspond to the ensemble simulations.

Table 4. Leading rotated EOFs (rEOFs) of SST for the ERSSTv3b data set, the five 20C ensemble members of CCSM4 (R005-R009), the ensemble mean (Ens), and the CORE-forced ocean-ice simulation. The rEOFs are based on a varimax rotation of the 10 dominant EOFs of the detrended, area-weighted, monthly SST anomaly time series. The North Tropical Atlantic (NTA) and Subtropical South Atlantic (SSA) modes are found in all data sets. In the CCSM4 ensemble members, the South Tropical Atlantic (STA) variability is represented by the STA-EQ and STA-
BG modes, with SST variability in the equatorial region and the Benguela upwelling zone, respectively. Lightest gray cells indicate relative ordering of the modes, while medium and dark gray cells indicate relative and absolute (domain-averaged, °C²) levels of variance accounted for by the modes.
Figure 1: (top) Total mean CORE-based wind stress derived for the period 1980-2005. Model minus CORE-based difference for CCSM4 (middle) and CCSM3 (bottom). The period used for each model is explained in the text. Shading indicates wind stress magnitude and vectors indicate the direction. Units are N/m².
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Figure 5: (a-d) Horizontal distribution of the month of deepest 28.5°C isotherm from the long-term mean from 1950 to 2005. The numbers 1 to 12 correspond to the months from January to December. The Pacific data has been masked. Panel (a) corresponds to the CCSM4 ensemble mean. Panel (c) corresponds to the POP ocean model forced with CORE surface forcing. Panels (b) and (d) correspond to the observational products, Ishii and Levitus, respectively. (e) Seasonal cycle of the volume of the 28.5°C isotherm between 40°S-40°N.
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Figure 7: The tropical South Atlantic (TSA) Warm Pool in April. (a-d) Mean depth (meters) of the 28.5°C isotherm in April. The CCSM4 ensemble mean (panel a) is the mean of five different simulations. (e) Time series of the volume ($10^4$ km$^3$) encompassed by the 28.5°C isotherm in April south of 5°N. The black line is the Ishii observational product; the blue line is the ocean POP simulation forced by CORE forcing; the red line is the CCSM4 ensemble mean with the ensemble spread in gray. The ensemble spread is the minimum and maximum value of any of the ensemble members.
Figure 8: Rank histograms of the CCSM4 ensemble spread against the POP ocean simulation forced by CORE (purple), and against the Ishii observational estimate (blue). The top panel corresponds to the index of the tropical North Atlantic (TNA) Warm Pool in September. The bottom panel corresponds to the index of the tropical South Atlantic (TSA) Warm Pool in April. The black line represents a uniform distribution.
Figure 9: Dominant rotated EOFs (rEOFs) of SST for the ERSSTv3b data set (left), the mean of the five 20C ensemble members of the CCSM4 (center), and the CORE-forced ocean-ice simulation (right). The rEOFs are based on a varimax rotation of the 10 dominant EOFs of detrended, area-weighted, monthly SST anomalies. The North Tropical Atlantic (NTA) and Subtropical South Atlantic (SSA) modes are found in all data sets. In CCSM4, the South Tropical Atlantic (STA) variability is represented by the STA-EQ and STA-BG modes, with SST variability in the equatorial region and the Benguela upwelling zone, respectively. The rEOFs carry the standard deviation. Negative, zero, and positive contours are thin dashed, thick solid, and thin solid, respectively, with contour interval of 0.1°C.
Figure 10: Statistical relationship between wind stress, SST, and the rPCs of the four dominant SST modes (STA-BG, STA-EQ, NTA and SSA) in the 20C ensemble member 005 (R005) of CCSM4 at lags of -1, 0 and 1 months. Negative lags indicate the wind stress and SST are leading. Correlations and regressions cover the 1948-2005 period. Shadings correspond to the correlation between the rPCs and wind stress magnitude (only values significant at the 95% level are shown). Arrows correspond to the linear regression of the rPCs on the wind stress components (legend arrows are 0.0089, 0.0069, 0.0084, 0.0087 N/m² for the STA-BG, STA-EQ, NTA, and SSA columns, respectively). Contours correspond to the linear regression of the rPCs on the sea surface temperature (contour interval 0.1, negative, zero and positive contours are indicated with gray, thick black, and black, respectively). The rPCs are renormalized to have standard deviation of 1, so the regressions correspond to unit amplitude of the rPCs.
Figure 11: Lagged correlations between each of the rotated PCs and SST (shades) or SLP (contours) from the ensemble run R005. Rotated PCs lag by one month. The top of each panel indicates the corresponding mode as explained in the text. Shadings are correlations with SST (gray areas are not significant at the 95% level). Contours are correlations with SLP: black contours are positive and gray contours are negative, with a contour interval of 0.1 starting at ±0.1 (the zero contour is not plotted).
Figure 12: Standard deviation (STD) of anomalous heat content rate of change in the upper 80m (shading, Wm$^{-2}$), STD of anomalous SST (black contours), and time mean SST (gray contours). Box is the model Benguela region. All data are from the 1deg 1850 control run of CCSM4.
Figure 13: Lagged autocorrelation of anomalous SST and lagged correlation of anomalous heat content rate of change (HCR) with anomalous vertical (VERT), meridional (MER), zonal (ZON) heat advection, and anomalous net surface heat flux (NHF). All variables are spatially averaged over the Benguela region box and vertically integrated in the upper 80m.
Figure 14: (a) Correlation of anomalous vertical heat advection in the Benguela region with wind stress elsewhere at zero lag. Arrows show correlation with wind stress components. Correlations significant at the 95% confidence level are shown in red. Temporal regression of anomalous vertical heat advection on anomalous mean sea level pressure elsewhere at zero lag is overlain as contours [mbar/(100Wm$^2$)].

(b) Correlation of anomalous meridional heat advection in the Benguela region with wind stress leading by 1 month. Inlay shows the lagged correlation averaged over the equatorial box. Wind stress leads for positive lags.
Figure ES-1: Power spectra of the rotated PCs (rPCs) for the different modes featured in Figure 9. A 13-point Daniell filter is applied to smooth the spectra. For CCSM4 (black) the spectra are averaged over the five 20C ensemble members. The spectra of the ERSST (dark gray) and CORE (light gray) data sets are offset by factors 0.25 and 0.0625, respectively. The thin lines are 95% confidence limits, based on a best-fit AR-1 model to the time series, and a 2500-member ensemble of AR-1 processes with these same parameters.
Table 1. Linear trend ($10^4$ km$^3$/year) of the September TNA and April TSA warm pool indices for each CCSM4 ensemble simulation, the POP ocean simulation forced by CORE, and the observational estimate of Ishii. The columns R005 through R009 correspond to the ensemble simulations.

<table>
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<tr>
<th>Linear trend</th>
<th>R005</th>
<th>R006</th>
<th>R007</th>
<th>R008</th>
<th>R009</th>
<th>POP</th>
<th>Ishii</th>
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</thead>
<tbody>
<tr>
<td>September</td>
<td>0.347</td>
<td>0.337</td>
<td>0.355</td>
<td>0.360</td>
<td>0.353</td>
<td>0.283</td>
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<tr>
<td>April</td>
<td>0.184</td>
<td>0.190</td>
<td>0.276</td>
<td>0.196</td>
<td>0.251</td>
<td>0.254</td>
<td>0.207</td>
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Table 2. Standard deviation of the September TNA and April TSA warm pool indices for each CCSM4 ensemble simulation, the POP ocean simulation forced by CORE, and the observational estimate of Ishii. The columns R005 through R009 correspond to the ensemble simulations.

<table>
<thead>
<tr>
<th>Standard deviation</th>
<th>R005</th>
<th>R006</th>
<th>R007</th>
<th>R008</th>
<th>R009</th>
<th>POP</th>
<th>Ishii</th>
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<tbody>
<tr>
<td>September</td>
<td>6.42</td>
<td>5.00</td>
<td>5.38</td>
<td>5.89</td>
<td>6.31</td>
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<td>April</td>
<td>5.56</td>
<td>4.90</td>
<td>5.35</td>
<td>4.61</td>
<td>5.85</td>
<td>8.14</td>
<td>4.76</td>
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Table 3. Spearman ($R_aS$) and Pearson ($R_aP$) auto-correlations of the September TNA and April TSA warm pool indices for each CCSM4 ensemble simulation, the POP ocean simulation forced by CORE, and the observational estimate of Ishii. The columns R005 through R009 correspond to the ensemble simulations.

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<tr>
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<th>Auto-correlation</th>
<th>R005</th>
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<th>R007</th>
<th>R008</th>
<th>R009</th>
<th>POP</th>
<th>Ishii</th>
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<tr>
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<td>0.31</td>
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<td>$R_aP$</td>
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<tr>
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<td>$R_aS$</td>
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<td>0.12</td>
<td>0.18</td>
<td>0.20</td>
<td>0.02</td>
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<td>0.39</td>
<td>0.10</td>
<td>0.03</td>
<td>0.24</td>
<td>-0.05</td>
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Table 4. Leading rotated EOFs (rEOFs) of SST for the ERSSTv3b data set, the five 20C ensemble members of CCSM4 (R005-R009), the ensemble mean (Ens), and the CORE-forced ocean-ice simulation. The rEOFs are based on a varimax rotation of the 10 dominant EOFs of the detrended, area-weighted, monthly SST anomaly time series. The North Tropical Atlantic (NTA) and Subtropical South Atlantic (SSA) modes are found in all data sets. In the CCSM4 ensemble members, the South Tropical Atlantic (STA) variability is represented by the STA-EQ and STA-BG modes, with SST variability in the equatorial region and the Benguela upwelling zone, respectively. Lightest gray cells indicate relative ordering of the modes, while medium and dark gray cells indicate relative and absolute (domain-averaged, °C²) levels of variance accounted for by the modes.

<table>
<thead>
<tr>
<th></th>
<th>ERSST</th>
<th>R005</th>
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<th>R007</th>
<th>R008</th>
<th>R009</th>
<th>Ens</th>
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<td>17.8%</td>
<td>11.7%</td>
<td>13.7%</td>
<td>15.5%</td>
<td>14.5%</td>
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<tr>
<td></td>
<td>--</td>
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<td>0.0184</td>
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<td>11.6%</td>
<td>13.6%</td>
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<td>10.7%</td>
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<tr>
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<tr>
<td>SSA</td>
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