A decadal climate variation in the tropical Atlantic Ocean from thermodynamic air–sea interactions

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Rainfall variability in northeast South America[1] and the Sahel region of Africa[2–4] is profoundly influenced by the sea surface temperature (SST) of the tropical Atlantic Ocean. Of particular importance are relative changes in SST between the hemispheres on decadal timescales, a phenomenon often called the Atlantic SST dipole[1,5]. Here we propose that the decadal variation in the tropical SST dipole may be attributed to an unstable thermodynamic ocean–atmosphere interaction between wind-induced heat fluxes and SST. Using coupled ocean–atmosphere models, we show that the coupled dipole mode has a typical oscillation period of about a decade. The notion that the Atlantic dipole-like SST variability may be related to an oscillatory coupled mode might assist attempts to predict decadal climate variability in the tropical Atlantic region.

In some respects, the tropical Atlantic Ocean is similar to the tropical Pacific, having, for example, an equatorial cold tongue of sea surface temperature just south of the Equator, and the northeasterly and southeasterly trade winds converging in the Intertropical Convergence Zone (ITCZ) north of the Equator. The ITCZ and cold tongue complex in the Atlantic has a pronounced annual cycle, more prominent than that in the Pacific[6]. A phenomenon similar to but much weaker than the Pacific Ocean El Niño also occurs in the Atlantic[7]. A mode of variability seemingly unique to the tropical Atlantic is the SST dipole mode—the decadal variation of the interhemispheric SST gradient across the Equator. Studies have suggested that rainfall variabilities in the northeast Brazil and Sahel regions are closely related to the cross-equatorial SST gradient variation[1–4]. In particular, the Brazilian rainfall tends to be more strongly correlated with the Atlantic SST dipole than with the Pacific El Niño, despite a stronger SST anomaly in the Pacific.

Dipole-like SST variabilities in the tropical Atlantic have been documented in several studies[5,8,9]. We have applied a joint singular-value decomposition (SVD) analysis to a 40-year (1950–89) monthly mean surface data set that includes SST, surface wind-stress and heat-flux anomalies. This data set is derived from the reprocessed Comprehensive Oceans Atmosphere Data Set (COADS)[10]. The SVD analysis provides an efficient way to identify coupled patterns in the ocean–atmosphere system[11]. Our analysis reveals that the first leading SVD mode, which explains about 45% of the total squared covariance, has a dipole-like SST pattern with the maximum amplitudes at about 15° S and 15° N, and opposite signs of response on either side of the Equator (Fig. 1a). A spectral analysis on the SST time coefficient of the first SVD (Fig. 1b) indicates a dominant spectral peak at about 13 years (Fig. 1c), consistent with previous studies[8,9].

Associated with the variability of the cross-equatorial SST gradient is a southeasterly wind anomaly in the Southern Hemisphere and a southwesterly wind anomaly in the Northern Hemisphere. These winds act to enhance the southeasterly mean trade winds in the Southern Hemisphere and reduce the northeasterly trades in the Northern Hemisphere. As a result, the surface heat flux from the ocean is decreased in the Northern Hemisphere the converse applies in the Southern Hemisphere. The anomalous heat flux enhances the SST dipole. The high correlation between the surface heat flux and SST anomalies (SSTA) is illustrated in Fig. 1a and suggests a positive ocean–atmosphere feedback through thermodynamic processes. This positive feedback differs considerably from the El Niño/Southern Oscillation (ENSO) mechanism, where ocean dynamics (changes in equatorial upwelling and thermocline depth in response to anomalous winds) play a central role in regulating tropical SSTs. Carton *et al.*[12] and Zhou[13] demonstrated that wind-induced latent heat flux is a major contributing factor to the variability of the SST dipole, whereas wind forcing is the main source of the interannual SST variability associated with the 'Atlantic El Niño'.

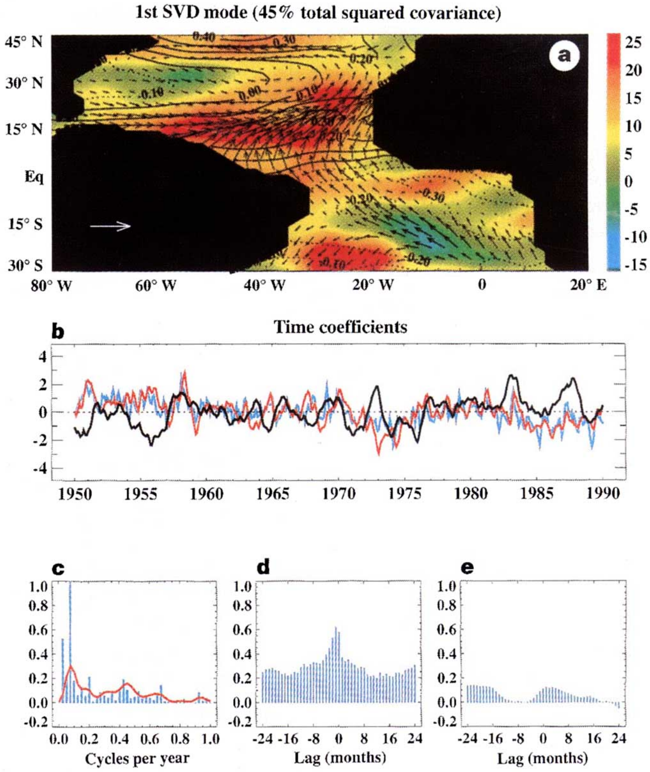

Figure 1 The Atlantic dipole-like variability revealed by a joint SVD analysis. **a**, Spatial structure of the first SVD mode. Contours represent SST anomaly in °C. Vectors depict wind-stress anomalies. The unit vector on the lower-left corner represents an easterly wind-stress anomaly of 1.0 dyn cm^{-2}. Colours indicate the strength of surface heat-flux anomalies in W m^{2}. **b**, The two associated time coefficients (red line, 1st SVD of Atlantic SSTA; blue line, 1st SVD of Atlantic atmospheric variation). The time coefficients have been normalized by their own standard deviations. The black line in **b** is the normalized time coefficient of the first EOF of the global SST anomaly. **c**, Periodogram of the SST time coefficient, along with an estimated spectrum (red line) using the Hanning window with a bandwidth $M = 216$ months. **d**, Cross-correlation between the two time coefficients of the SVD, as a function of lag. The maximum correlation of 0.7 occurs at a lag of −1, indicating that the atmosphere leads the ocean by one month. **e**, Cross-correlation between the time coefficients of the first SVD of the tropical Atlantic SST and the first EOF of the global SST. Low correlation values in **e** suggest that the coupled dipole mode is not strongly affected by changes in the global SST.

Information on causal linkages of air–sea feedbacks may be contained in the cross-correlation between the time coefficients of the atmospheric forcing and SST[14]. A positive feedback implies that the ocean and atmosphere are mutually reinforcing each other. The cross-correlation in this case will be more or less symmetrical and of the same sign for both positive and negative lags, although it usually peaks when the atmosphere leads the ocean. In contrast, when the atmosphere forcing has negative or no feedbacks, the cross-correlation will have either an anti-symmetric appearance or a very asymmetric form, peaking when the atmosphere leads but dropping

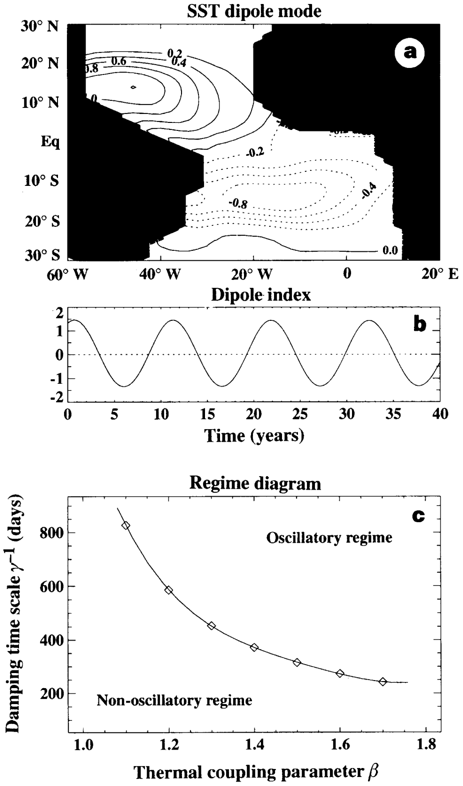

Figure 2 Dipole modes of the intermediate coupled model. **a**, Spatial pattern obtained by regressing the simulated SST upon a dipole index (shown in **b**) generated by differencing area-averaged SSTs in the Northern and Southern hemispheres. The area-averages were taken over longitudinal bands 18° wide centred at 15° S and 15° N across the basin. The results shown in **a** and **b** were based on simulation using coupling parameters $\alpha = 0$ and $\beta = 1.3$. The standard deviation of the dipole index is about 1.0 °C. Contour intervals are 0.2 °C per standard deviation of the dipole index. The regime diagram at the bottom was obtained by varying the thermodynamic coupling parameter β and damping parameter γ. The damping coefficient γ^{-1} determines the time over which the model SSTs are relaxed back to the annual mean SST by means of a newtonian damping term $-\gamma(T - T_{\mathrm{mean}})$ in the model temperature equations. A boundary in parameter space separates the non-oscillatory regime from the oscillatory regime.

rapidly to zero when the atmosphere lags. Figure 1d shows the cross-correlation for the first SVD mode. The cross-correlation shows a peak when the atmosphere is leading the ocean by about a month and remains positive for all the lags, suggesting a positive feedback. Figure 1e illustrates the correlation between the dipole variation and global SST fluctuation represented by the time coefficient of the first leading empirical-orthogonal-function (EOF) mode of the global SST anomaly (black line in Fig. 1b). There is a low correlation (less than 0.2) throughout ±24-month lags, suggesting that the Atlantic dipole mode involves mainly local dynamics and is not strongly influenced by global processes.

To explore further the role of air–sea interactions in the tropical Atlantic, we have developed two coupled ocean–atmosphere models. The simpler model is an intermediate coupled model (ICM), which consists of an empirical atmospheric feedback model and a reduced-gravity ocean model[15,16]. The more sophisticated model—a hybrid coupled general circulation model (HGCM)—combines the same empirical atmosphere with a 3-D primitive-equation Ocean General Circulation Model (OGCM; R. C. Pacanowski, K. Dixon and A. Rosati, personal communication). The model configurations are similar to those developed for the Pacific ENSO studies described in ref. 17. The atmospheric feedback is constructed using the joint SVD analysis described earlier, which establishes an empirical relation between the SST anomaly and the corresponding surface wind-stress and heat-flux anomalies from observations of low-frequency variability[18]. Two scaling factors α and β are introduced for, respectively, surface wind stress and heat flux. These coupling parameters control the strength of dynamic and thermodynamic feedbacks. In the coupled models, only the spatial patterns associated with the singular eigenvectors are taken into consideration—that is, the SSTA pattern determines the patterns of wind stress and heat flux. Temporal information in the observations is not used in the coupled models.

A large number of numerical experiments were conducted with the coupled ocean-atmosphere models. For the ICM, each experiment consists of a 5-year spin-up forced with the annual mean wind stress and surface heat flux followed by a 60-year coupled integration. The HGCM experiments use a 120-year spin-up and a 120-year coupled run, as a fully stratified ocean model needs a much longer spin-up time. In all the experiments, the coupled models are forced externally with the COADS annual mean wind stress and heat flux. Figure 2 summarizes the results from a set of ICM experiments where the atmosphere and ocean are coupled only through thermodynamic processes; feedbacks between winds and SST are disallowed by setting α to zero. the spatial pattern of the coupled mode bears a striking resemblance to the observed SST dipole, with an oscillation period varying from 8 to 14 years depending on the thermodynamic coupling strength β. The larger β, the shorter is the oscillation period. When only the dynamic coupling is allowed ($\beta = 0$), the model produces an ENSO-like variability with oscillation period ranging from 2.5 to 4 years (not shown). This result is consistent with previous findings[7]. Figure 2c illustrates the dependence of oscillatory behaviour of the coupled modes on coupling and damping parameters. The HGCM experiments gave a similar result. A self-sustained dipole oscillation was obtained, when $\alpha = 0$, $\beta = 1.05$ and damping coefficient $\gamma^{-1} = 150$ days, with a period of about 13 years, indicating that the existence of the coupled dipole mode is model-independent.

To gain further understanding of the physical processes that control the dipole-like oscillations, we have analysed the evolution of model SSTs by diagnosing the upper-ocean heat budget. For a dipole-like oscillation, the dominant physical processes include a positive feedback between the heat-flux and SST anomalies, and a negative feedback due to the advection of heat by steady ocean currents. The negative feedback can be briefly described as follows: if the Northern Hemisphere SST is warmer than the Southern Hemisphere, the steady ocean currents, particularly the intensified western boundary currents such as the North and South Brazil currents, will transport anomalous cold (warm) water to the Northern (Southern) Hemisphere. This horizontal heat advection acts against the atmospheric warming (cooling) in the Northern (Southern) Hemisphere by transporting the anomalously cold (warm) water to the dipole region. Atmospheric radiative and turbulent energy fluxes, together with ocean mixing processes, damp SST anomalies, acting as additional negative feedbacks. When these positive and negative feedbacks properly balance each other, the system gives rise to a self-sustained oscillation. For the dipole oscillation, we have found that thermocline variability does not contribute significantly to balancing the heat budget, indicating that ocean wave dynamics may be of secondary importance. Thus, this coupled mode can be regarded as a generalization of the SST mode studied by Neelin[19].

Although simple coupled models suggest that a self-sustained oscillatory dipole mode is possible in a coupled system, in reality stochastic processes generated by internal dynamics of the atmosphere and oceans can have a large impact on the response of

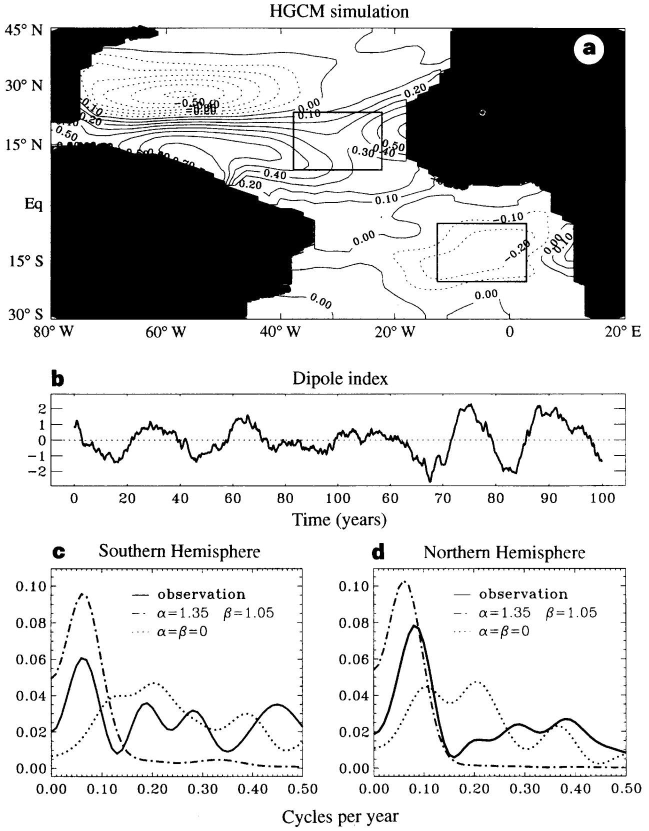

Figure 3 SST fields simulated by the hybrid coupled general circulation model forced with wind noise with both dynamic and thermodynamic coupling ($\alpha = 1.35$, $\beta = 1.05$). **a**, Dipole pattern generated using a regression analysis as in Fig. 2, with a dipole index (shown in **b**) derived by taking the difference of the model SSTs averaged over a 15° × 15° area in each hemisphere indicated by two rectangles. The dipole pattern is insensitive to the choice of the index. The two areas were chosen because they had the best availability of 100-year SST observations. The 100-year (1890–1990) observed monthly mean SST time series were derived from the UK Meteorological Office Main Marine Data Bank[20] by area-averaging over the two regions. The observed SSTs were compared with similar SST time series derived from model simulations via a spectral analysis. **c**, **d**, SST spectra in the Southern and Northern hemispheres, respectively, where solid lines are for observations, dash-dotted and dotted lines are for simulations with $\alpha = 1.35$, $\beta = 1.05$ and $\alpha = \beta = 0$, respectively. The SST spectra were estimated using the Hanning window with a bandwidth $M = 240$ months and were normalized by their own variance. The standard deviations of the observed SST in the Northern and Southern hemispheres are about 0.5 °C. Similar values for the simulated SSTs are about 0.3 °C for $\alpha = 1.35$, $\beta = 1.05$ and are less than 0.1 °C for $\alpha = \beta = 0$.

ocean–atmosphere coupled systems. Recent studies[16] suggest that random atmospheric forcing can resonantly excite coupled modes in a system that contains no self-sustained oscillatory modes in the absence of stochastic processes. To explore the effects of stochastic processes on tropical Atlantic climate variability, we have done experiments with the addition of external atmospheric noise forcing. Figure 3 depicts the SST response from an HGCM experiment using $\alpha = 1.35$ and $\beta = 1.05$ with wind-noise forcing. The wind noise was constructed using the high-frequency component of the observed wind stress from COADS. It retains a coherent structure in space that mimics short period 'weather', but has a white spectrum in time. The coupled model produces dipole-like SST variability similar to observations, although the SST variabilities in the model are weaker than observed values (Fig. 3). A comparison between the observed and simulated SST spectra indicates that the coupled model reproduces the observed decadal spectral peak at a period around 13 years in both hemispheres (Fig. 3c and d). In contrast, the model SST response in the absence of any air–sea feedbacks ($\alpha = \beta = 0$) produces neither dipole-like SST patterns nor pronounced decadal spectral peaks. Experiments using white spatiotemporal wind and heat-flux noises lead to similar results, indicating that ocean–atmosphere interactions are essential to the decadal dipole-like SST variability in the tropical Atlantic Ocean.

In light of these results, we propose that the dipole-like variability in the tropical Atlantic may be attributed to ocean–atmosphere interactions involving primarily positive thermodynamic feedbacks between surface heat flux and SST. The coupled model experiments indicate that in a realistic parameter regime the dipole mode oscillates on a timescale of 12–13 years. In reality, stochastic processes in the atmosphere may play an important role in resonantly exciting the dipole mode and causing irregularities in the dipole variability. Therefore, the decadal dipole-like SST variability are perhaps best viewed as a resonant response of the coupled system to short-period 'weather' disturbances in the atmosphere. ☐

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Triggering basaltic volcanic eruptions by bubble–melt separation

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Understanding the processes by which volcanic eruptions are triggered is crucial for volcanic hazard prediction and assessment. In intermediate and silicic magmas, where water is the dominant volatile phase[1,2], 'second boiling' provides an effective eruption trigger[1]—as the magma cools and crystallizes, volatiles are increasingly concentrated in the residual melt where they eventually become saturated and are exsolved, thereby raising the magma pressure to the point of eruption[1,2]. In contrast, in basaltic magma chambers in which carbon dioxide is the dominant volatile phase, the effect of second boiling is to decrease the chamber pressure. But basaltic magmas are also relatively inviscid, so volatile bubbles can separate from the turbulently convecting melt to produce a foam layer at the chamber roof[3,4]: decompression of the rising bubbles can increase the chamber pressure and so trigger an eruption[5,6]. Here we model the complex competition between the processes of magma cooling and bubble