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Equatorial upper-ocean dynamics and their interaction with the West African monsoon

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Abstract

Zonal wind anomalies in the western equatorial Atlantic during late boreal winter to early summer precondition boreal summer cold/warm events in the eastern equatorial Atlantic (EEA) that manifest in a strong interannual Atlantic cold tongue (ACT) variability. Local intraseasonal wind fluctuations, linked to the St. Helena anticyclone, contribute to the variability of cold tongue onset and strength, particularly during years with preconditioned shallow thermoclines. The impact of cold tongue sea surface temperature (SST) anomalies on the wind field in the Gulf of Guinea is assessed. It contributes to the northward migration of humidity and convection and possibly the West African monsoon (WAM) jump. Copyright © 2010 Royal Meteorological Society

Keywords: **equatorial Atlantic; cold tongue; West African monsoon**

1. Introduction

Climate fluctuations in the Atlantic sector are dominated by two distinct patterns of coupled ocean/atmosphere variability, collectively referred to as tropical Atlantic variability. They are tightly phase-locked to the pronounced Atlantic seasonal cycle and vary on interannual to decadal time scales. The zonal mode, frequently viewed as the Atlantic counterpart of the Pacific El Niño-Southern Oscillation (ENSO), is most pronounced during boreal summer and coincides with the seasonal development of the eastern equatorial cold tongue (Zebiak, 1993; Kushnir *et al*., 2006). Interannual variability of sea surface temperature (SST) in the Atlantic cold tongue (ACT) during boreal summer is closely linked to rainfall variability in the countries surrounding the Gulf of Guinea and in particular to the onset of the West African monsoon (WAM). Similar to the Pacific Ocean, where the NINO3 index is defined to describe the ENSO, the ATL3 index describes the variability of the zonal mode (Figure 1).

The ocean circulation in the upper equatorial Atlantic is composed of vigorous zonal currents. At the surface, there are two branches of the South Equatorial Current (SEC) flowing westward at about 2°N and 4°S. Below the near-surface flow, the Equatorial Undercurrent (EUC) supplies the upwelling within the ACT region with saline subtropical waters. In the mean, the core depth and transport of the EUC decrease from west to east (Figure 2). The EUC seasonal cycle shows a pronounced transport maximum during late boreal summer and autumn (Brandt *et al*., 2006; Hormann and Brandt, 2007; Kolodziejczyk *et al*., 2009), when the SST in the ACT is lowest. Mean and seasonal currents in the equatorial Atlantic are modulated by intraseasonal fluctuations. Kelvin waves are found to propagate along the equator and further south- and northward along the African coast (Polo *et al*., 2008). However, tropical instability waves (TIWs) dominate the intraseasonal variability particularly during boreal summer (Düing *et al*., 1975; Weisberg *et al*., 1979; Weisberg and Weingartner, 1988; Athie and Marin, 2008). East of 10°W, where barotropic and baroclinic instabilities of the zonal currents become negligible as suggested by numerical modeling (von Schuckmann *et al*., 2008), wind-forced mixed Rossby-gravity waves are the strongest intraseasonal signal (Athie and Marin, 2008; Han *et al*., 2008; De Coëtlogon *et al*., 2010).

In this note, we will focus (1) on processes affecting the mixed layer (ML) heat budget and SST as well as (2) on the SST variability related to the WAM. We present recent progress in the understanding of the development and impact of the ACT, particularly reviewing studies that are based on observations carried out during the African Monsoon Multidisciplinary Analysis (AMMA) period.

2. ML heat budget

Extremely shallow ML depths characterize the central and eastern equatorial Atlantic (EEA). East of about

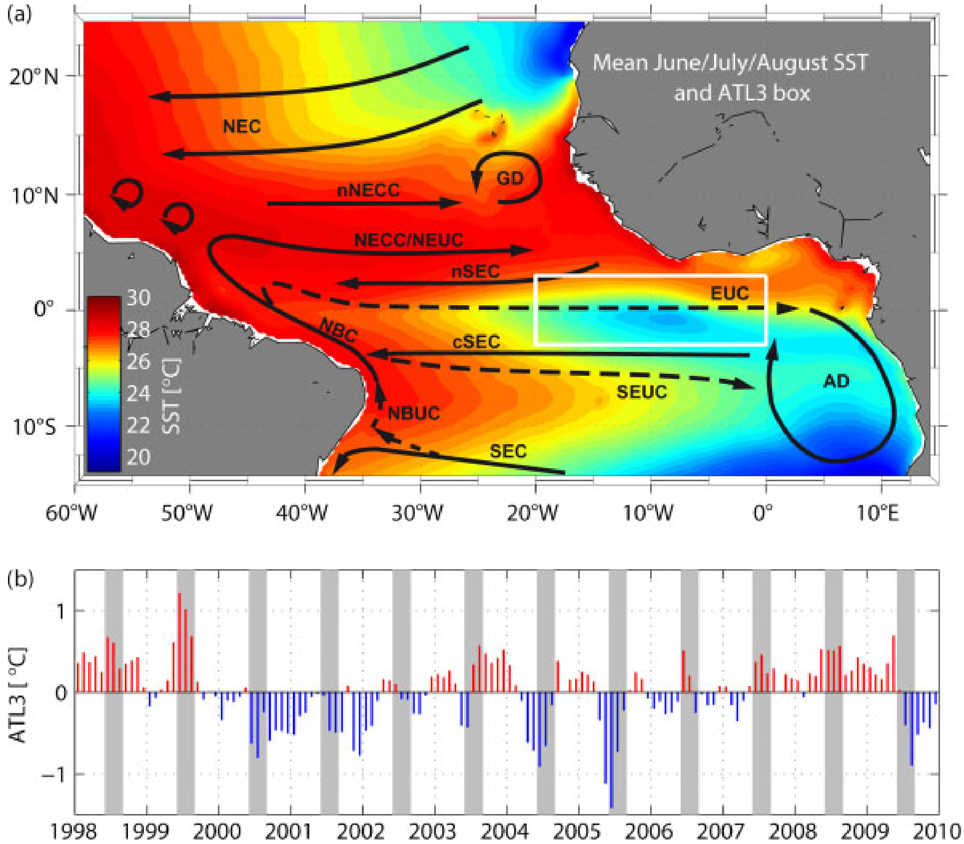

Figure 1. Mean June/July/August SSTs (averaged from 1998 to 2009) in the tropical Atlantic (a). Also included are the main surface (solid) and thermocline (dashed lines) current bands: North Equatorial Current (NEC), North Equatorial Undercurrent (NEUC), North Equatorial Countercurrent (NECC), its northern branch (nNECC), Equatorial Undercurrent (EUC), South Equatorial Current (SEC), its northern and central branches (nSEC, cSEC), South Equatorial Undercurrent (SEUC), North Brazil Current and Undercurrent (NBC, NBUC), as well as the cyclonic circulations around the Guinea and Angola Domes (GD, AD). The white box in (a) defines the region for the ATL3 SST index shown in (b). Monthly SST residuals are calculated by subtracting the mean seasonal cycle; gray bars highlight boreal summer (June/July/August). The microwave optimally interpolated (OI) SST dataset used here is available at www.remss.com.

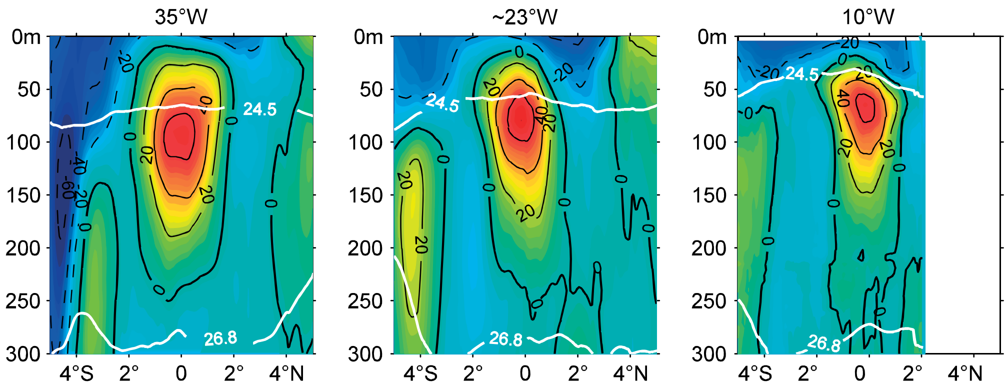

Figure 2. Mean zonal velocity [color shading and black contours (cm/s)] and potential density surfaces 24.5 and 26.8 (white solid lines) along three meridional sections: 35 °W, ∼23 °W, and 10 °W (after Brandt *et al.*, 2006; Hormann and Brandt, 2007; Kolodziejczyk *et al.* 2009).

20 °W, the ML is less than 20 m deep throughout most of the year, while it reaches a maximum depth of 30–40 m in boreal autumn (Hastenrath and Merle, 1987; de Boyer Montégut *et al.*, 2004). Thus, even moderate ML heat fluxes of O(10) Wm^{-2} may significantly impact the ML temperature. A recent modeling study indicates that the ML heat content in the central and eastern tropical Atlantic regions is mainly controlled by air–sea heat exchange, vertical processes at the base of the ML, and lateral advection of heat due to TIWs (Peter *et al.*, 2006). During the ACT development in late boreal spring to early summer, the net atmospheric heat fluxes are at a seasonal minimum due to increased latent heat flux. Heat loss due to vertical diffusion, upwelling, and entrainment of cold water into the ML (Peter *et al.*, 2006) are increased as the

strengthening of the zonal currents enhances vertical shear. Horizontal advection of heat by the mean circulation also contributes to the cooling in the central equatorial Atlantic, but these heat fluxes are lower than those by vertical processes. In late boreal summer and autumn, the net atmospheric heat fluxes increase again because of higher net solar radiation. Lateral eddy heat fluxes into the equatorial region due to TIWs play an increasing role in the ML heat budget and contribute to the disintegration of the cold tongue (Jochum *et al.*, 2004; Peter *et al.*, 2006). To date, however, models disagree on how TIW-modulated vertical mixing contributes to the ML heat budget. While simulations by Foltz *et al.* (2004) and Jochum and Murtugudde (2006) indicate a net cooling of the ML due to enhanced vertical mixing in the vortices, Menkes *et al.* (2006) found that the area-averaged diapycnal heat flux at the base of the ML decreases in the presence of TIWs.

Attempts to provide a detailed description of the seasonal cycle of the equatorial ML heat fluxes from observations have yet been unsuccessful in closing the budget for the central and eastern Atlantic. Foltz *et al.* (2003) obtained a residual heat gain in the ML heat balance of the order of 50 $\mathrm{Wm^{-2}}$ during boreal spring and early summer in the central (23 °W), and of the order of 100 $\mathrm{Wm^{-2}}$ from late boreal spring to late autumn in the EEA (10 °W). During boreal summer, the zonal advection by the mean currents was found to be the predominant cooling term in the central Atlantic while meridional advection was suggested to be the predominant mechanism cooling the ML in the EEA. Vertical advection also contributes to the cooling of the central and, to a lesser extent, the EEA ML. However, estimated heat fluxes due to vertical advection were of the order of 20–30 $\mathrm{Wm^{-2}}$ only. A recent investigation of heat fluxes due to vertical advection during boreal summer and early autumn suggested that heat fluxes of 20–30 $\mathrm{Wm^{-2}}$ are indeed frequently encountered in the central and EEA; during a period of enhanced winds in June 2006, however, elevated vertical velocities and associated enhanced heat fluxes, as high as 140 $\mathrm{Wm^{-2}}$, were determined at the equator at 10 °W (Rhein *et al.*, 2010). These strong events may be missing when using the drifter climatology (Foltz *et al.*, 2003) to determine vertical velocities; the heat flux due to vertical advection may consequently remain underestimated. Foltz *et al.* (2003) also neglected heat flux contributions due to diapycnal mixing at the base and below the ML. From recent microstructure data collected in the center of the ACT in September 2005, a diapycnal heat flux of 60 $\mathrm{Wm^{-2}}$ across the base of the ML was determined (Rhein *et al.*, 2010).

During the EGEE (Etude de la circulation océanique et du climat dans le Golfe de Guinée) cruise in June/July 2006, turbulent atmospheric heat fluxes were measured with a mast installed on the foredeck of R/V L'Atalante (Bourras *et al.*, 2009). The parameterization of fluxes deduced from these data and observed radiative fluxes were used to evaluate fluxes from different operational weather prediction models (ECMWF, NCEP2, and ARPEGE) along the ship trajectory (Table I). The net heat fluxes are similar for ECMWF and ARPEGE models but differ largely from the shipboard observations (model-data bias is around 20 $\mathrm{Wm^{-2}}$). The bias is threefold larger in magnitude for NCEP2, a model for which the variability of the fluxes is less well reconstructed. The contributors to the biases are shortwave (NCEP2 and ARPEGE) and longwave (ECMWF) radiation and latent heat flux. Biases are most likely due to model misrepresentations of clouds and aerosols and their radiative effects and to imperfect parameterizations of turbulent heat fluxes at the equator.

3. Interannual variability

The interannual tropical Atlantic SST variability has, besides in the coastal upwelling regions off Africa, strongest amplitudes in the EEA where they peak during boreal summer. The wind forcing in the western tropical Atlantic during boreal spring was found to be one of the dominant factors responsible for this SST variability; the influence of local wind forcing was found to be considerably smaller (Servain *et al.*, 1982). Since 1998, when SSTs in the tropics were consistently measured by microwave radiometers, the strongest cold event occurred during boreal summer 2005 (Figure 1). Recently, different studies compared the ACT variability of different years: while Hormann and Brandt (2009) contrasted the variability during the years 2002 and 2005, Marin *et al.* (2009) analyzed shipboard data from EGEE cruises in 2005 and 2006. Hormann and Brandt (2009) particularly studied the amplitude of equatorial Kelvin waves that were found

Table I. Comparison of the atmospheric heat fluxes (positive values mark ML warming) obtained from shipboard observations (SHIP) and different operational weather prediction models (ECMWF, NCEP2, and ARPEGE).

$\mathrm{Wm^{-2}}$ (Std)	ECMWF	NCEP2	ARPEGE	SHIP
Net longwave (N = 108)	−55.5 (10.2)	−43.0 (16.0)	−66.1 (17.9)	−75.3 (15.6)
Net shortwave (N = 54)	417.7 (74.3)	322.5 (119.2)	429.9 (97.5)	415.6 (85.3)
Latent heat flux (N = 107)	−140.4 (37.9)	−185.9 (62.4)	−135.5 (51.0)	−148.8 (54.1)
Sensible heat flux (N = 107)	−11.0 (7.1)	−13.0 (9.2)	−10.0 (6.6)	−8.2 (5.9)
Net heat flux (N = 107)	3.9 (224.3)	−78.9 (188.6)	5.4 (230.5)	−23.0 (223.9)

Comparison concerns 6-h fluxes along the ship trajectory during the EGEE cruise in June/July 2006.
Number of samples and standard deviations are given in parenthesis.

to have a minor direct impact on boreal summer SST variability. However, prior to the ACT onset in 2002 (2005), the presence of equatorial Kelvin waves was associated with a flattened (steeper) thermocline slope that represents the preconditioning for the development of the warm (cold) event. Stronger than normal easterlies during April–May 2005 in the western tropical Atlantic, which were responsible for the generation of upwelling Kelvin waves, were also identified by Marin *et al*. (2009) to force a strong thermocline slope along the equator with exceptionally shallow thermocline depths in the EEA during the succeeding boreal summer. In contrast to Servain *et al*. (1982), Marin *et al*. (2009) highlighted the role of local intraseasonal wind variability in affecting the onset date and strength of the ACT. Stronger than average intraseasonal wind fluctuations over the ACT, which were found to be linked to the St. Helena anticyclone, resulted in an early, rapid, and intense cooling of the SSTs over a broad region extending from 20 °W to the African coast and from 6 °S to the equator in mid-May 2005. Similar wind fluctuations were not observed during 2006. Local intraseasonal wind fluctuations are particularly efficient in contributing to the interannual SST variability during years of steeper thermocline slope.

SST fluctuations also retroact on the surface wind field. One mechanism is surface wind acceleration (deceleration) over warm (cold) water, likely due to a vertical readjustment of the marine atmospheric boundary layer (Wallace *et al*., 1989). De Coëtlogon *et al*. (2010) suggested that this mechanism supports a substantial part of the biweekly variability observed in the region via a negative feedback loop. It may also be responsible for the modulation of the wind bursts north of 2°N, possibly affecting the migration of humidity and convection toward the continent on short time scales (Janicot *et al*., 2011).

The interannual variability of the WAM onset (Fontaine and Louvet, 2006) is found to be correlated with the ACT onset (Figure 3). The correlation is 0.45, statistically significant at the 97% confidence level. Both late ACT and WAM onsets are associated with warm SST and convergent wind anomalies in/over the ACT during June including northerly wind anomalies over the Guinean coast (Figure 4). Using an atmospheric general circulation model, Okumura and Xie (2004) were able to show that equatorial cooling intensifies southerly winds over the Guinean coast and increases the WAM flow. The regression maps (Figure 4) also show that a late ACT onset is associated with westerly wind anomalies in the western tropical Atlantic during late boreal winter to early summer, representing the preconditioning of warm events in the east. These wind anomalies are found to be dominantly linked to variations of the St. Helena anticyclone (Lübbecke *et al*., 2010). However, they are not similarly strong correlated with the WAM onset. Instead, a late WAM onset is associated with an SST dipole from April to July, characterized by

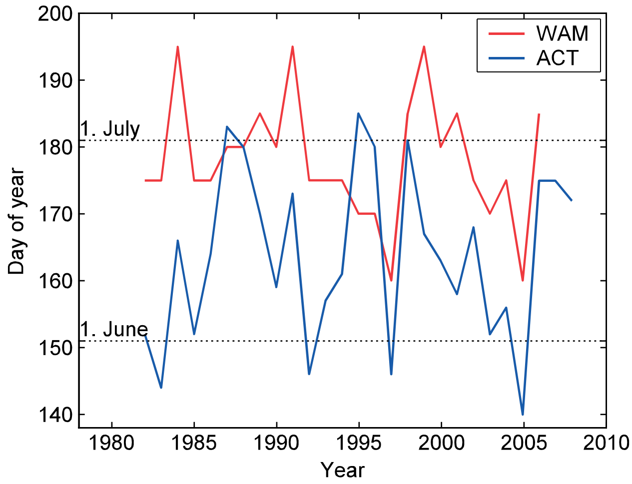

Figure 3. Onset dates for WAM (from Fontaine and Louvet, 2006; obtained using the GPCP dataset) and ACT defined as the date when the surface area, where the SST is below 25 °C inside the domain 30 °W–12 °E and 5 °S–5°N, exceeds the empirically fixed threshold of 0.4×10^6 km^2.

anomalously cool temperatures in the northeast tropical Atlantic and anomalously warm temperatures in the ACT region. This dipolar structure of the SST pattern (Rodríguez-Fonseca *et al*., 2011) can be seen as an expression of the meridional mode that peaks during boreal spring (Kushnir *et al*., 2006). A coupling between zonal and meridional modes was suggested by Servain *et al*. (1999), in which a southward interhemispheric temperature gradient weakens easterly winds on the equator in early boreal summer, thus contributing to the preconditioning of warm ACT events.

4. Discussion and conclusion

Within the AMMA program, an intense oceanic field campaign was carried out in the tropical Atlantic, particularly enhancing the observational database in the EEA. From the acquired moored and shipboard data, transport values of the main upper equatorial current branches as well as their variability on intraseasonal to seasonal time scales could be established. Ongoing mooring work in the framework of Tropical Atlantic Climate Experiment (TACE) and the Prediction and Research Moored Array in the Tropical Atlantic (PIRATA) will also allow investigating interannual transport fluctuations.

During the AMMA period, the quality of heat fluxes, finally needed to close the ML heat budget, improved considerably. For the first time, diapycnal heat fluxes at the base of the ML could be calculated systematically from microstructure measurements performed during the different cruises. The comparison of atmospheric heat fluxes obtained from observations and operational weather prediction models showed, however, large differences among the different models and systematic model-data biases.

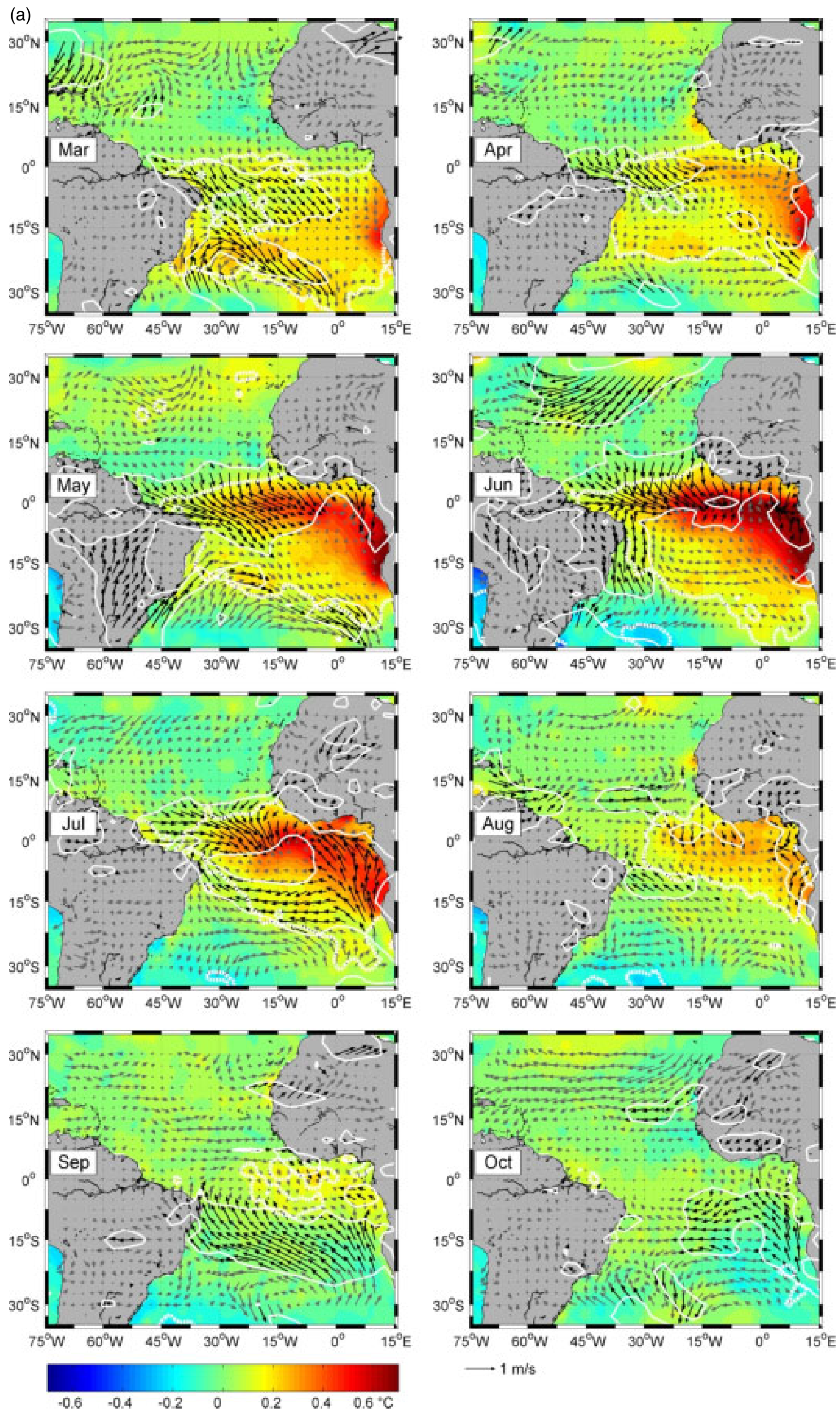

Figure 4. Regression maps of SSTs (color shading) and winds (gray and black arrows) onto the ACT onset (a) and the WAM onset (b) shown in Figure 3. Regressions are calculated separately for the different months from March to October. Significant correlations (95%) between onset dates and SSTs are marked by thick white dashed lines and between onset dates and winds by thin white solid lines and black arrows. SST and wind datasets are HadISST1.1 (Rayner *et al.*, 2003) and NCEP1 (Kalnay *et al.*, 1996), respectively.

These uncertainties make calculations of the seasonal and, even more, the interannual heat budget of the ML questionable. Besides improved atmospheric heat and freshwater fluxes, observations are required to address the impact of ML salinity, upwelling, and equatorial waves on the SST variability in the tropical Atlantic. Observations in the EEA region during the ACT preconditioning phase in boreal spring are particularly underrepresented in the observational database.

The interannual variability of the ACT, which was particularly well observed during the EGEE cruises from 2005 to 2007 as part of the AMMA program, was found to be strongly affected by wind anomalies in the western equatorial Atlantic during boreal spring and early summer. These wind anomalies set

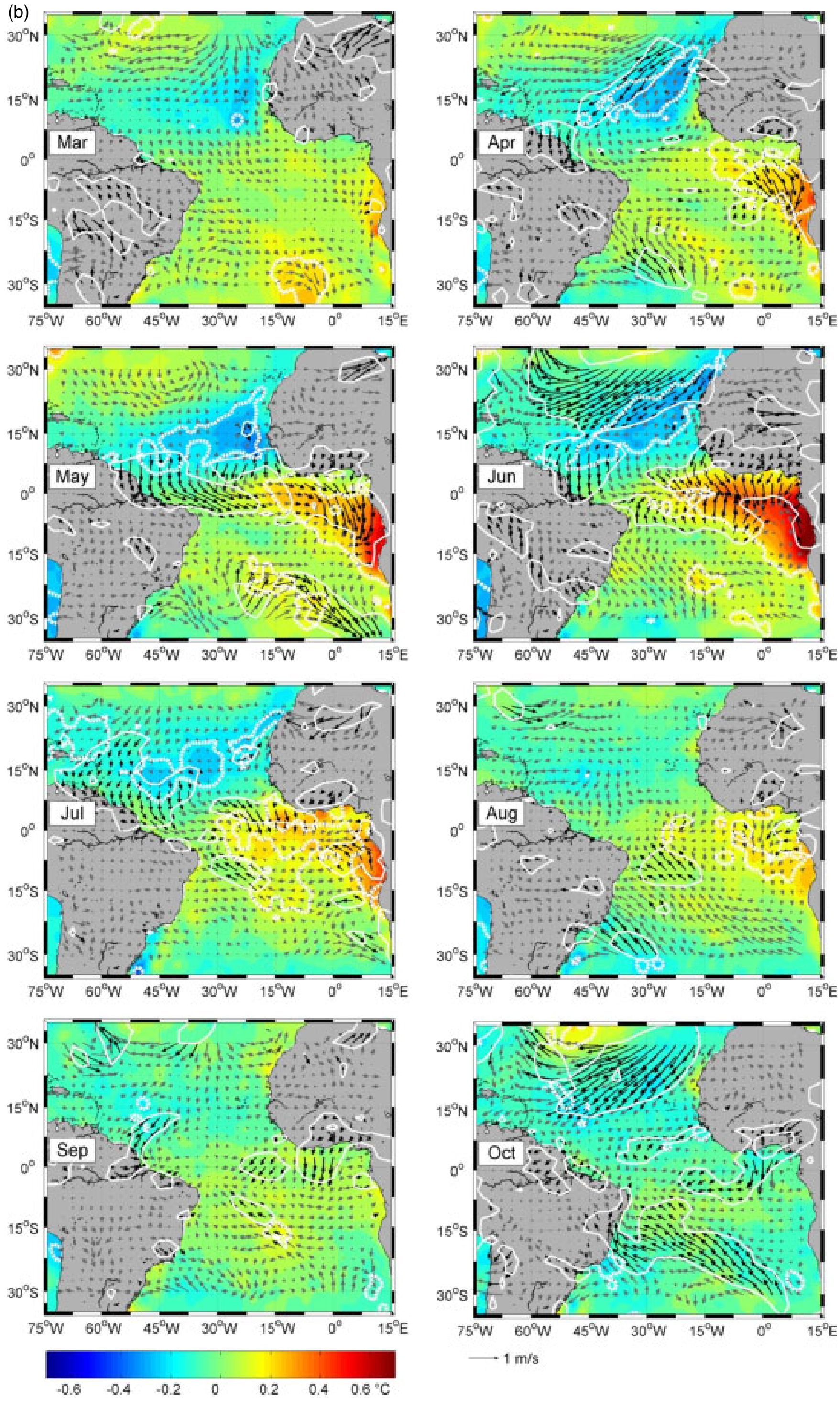

Figure 4. Continued.

the thermocline slope prior to the ACT onset and precondition warm and cold events. Stochastic processes, like local intraseasonal wind fluctuations and diapycnal mixing, are found to profoundly influence the SST evolution and possibly reduce the predictability of SST patterns.

A close link between ACT and WAM onsets could be established: late ACT and WAM onsets are associated with anomalously warm SSTs in the EEA. An SST dipole characterized, besides anomalously warm SSTs in the EEA, by anomalously cool temperatures in the northeast tropical Atlantic additionally influences the evolution of the WAM. The predictability of WAM onset and strength associated with the SST pattern remains an open question.

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References

Athie G, Marin F. 2008. Cross-equatorial structure and temporal modulation of intraseasonal variability at the surface of the Tropical Atlantic Ocean. *Journal of Geophysical Research* **113**: C08020. DOI:10.1029/2007JC004332.

Bourras D, *et al*. 2009. Turbulent air-sea fluxes in the Gulf of Guinea during the AMMA experiment. *Journal of Geophysical Research* **114**: C04014. DOI:10.1029/2008JC004951.

Brandt P, Schott FA, Provost C, Kartavtseff A, Hormann V, Bourlès B, Fischer J. 2006. Circulation in the central equatorial Atlantic: Mean and intraseasonal to seasonal variability. *Geophysical Research Letters* **33**: L07609. DOI:10.1029/2005GL025498.

de Boyer Montégut C, Madec G, Fischer AS, Lazar A, Iudicone D. 2004. Mixed layer depth over the global ocean: An examination of profile data and a profile-based climatology. *Journal of Geophysical Research* **109**: C120031. DOI:10.1029/2004JC002378.

De Coëtlogon G, Janicot S, Lazar A. 2010. Intraseasonal variability of the ocean – atmosphere coupling in the Gulf of Guinea during boreal spring and summer. *Quarterly Journal of the Royal Meteorological Society* **136**: 426–441. DOI:10.1002/qj.554.

Düing W, Hisard P, Katz E, Meincke J, Miller L, Moroshkin KV, Philander G, Ribnikov AA, Voigt K, Weisberg R. 1975. Meanders and long waves in the equatorial Atlantic. *Nature* **257**: 280–284. DOI:10.1038/257280a0.

Foltz GR, Grodsky SA, Carton JA, McPhaden MJ. 2003. Seasonal mixed layer heat budget of the tropical Atlantic Ocean. *Journal of Geophysical Research* **108**: 3146. DOI:10.1029/2002JC001584.

Foltz GR, Carton JA, Chassignet EP. 2004. Tropical instability vortices in the Atlantic Ocean. *Journal of Geophysical Research* **109**: C03029. DOI:10.1029/2003JC001942.

Fontaine B, Louvet S. 2006. Sudan-Sahel rainfall onset: Definition of an objective index, types of years, and experimental hindcasts. *Journal of Geophysical Research* **111**: D20103. DOI:10.1029/2005JD007019.

Han W, Webster PJ, Lin J-L, Liu WT, Fu R, Yuan D, Hu A. 2008. Dynamics of intraseasonal sea level and thermocline variability in the equatorial Atlantic during 2002-03. *Journal of Physical Oceanography* **38**: 945–967. DOI:10.1175/2008JPO3854.1.

Hastenrath S, Merle M. 1987. Annual cycle of subsurface thermal structure in the tropical Atlantic. *Journal of Physical Oceanography* **17**: 1518–1538.

Hormann V, Brandt P. 2007. Atlantic Equatorial Undercurrent and associated cold tongue variability. *Journal of Geophysical Research* **112**: C06017. DOI:10.1029/2006JC003931.

Hormann V, Brandt P. 2009. Upper equatorial Atlantic variability during 2002 and 2005 associated with equatorial Kelvin waves. *Journal of Geophysical Research* **114**: C03007. DOI:10.1029/2008JC005101.

Janicot S, Caniaux G, Chauvin F, de Coetlogon G, Fontaine B, Hall N, Kiladis G, Lafore JP, Lavaysse C, Lavender SL, Leroux S, Marteau R, Mounier F, Philippon N, Roehrig R, Sultan B, Taylor CM. 2011. Intra-seasonal variability of the West African monsoon. *Atmospheric Science Letters* **12**: 58–66, DOI: 10.1002/asl.280.

Jochum M, Malanotte-Rizzoli P, Busalacchi A. 2004. Tropical instability waves in the Atlantic Ocean. *Ocean Modelling* **7**: 145–163.

Jochum M, Murtugudde R. 2006. Temperature advection by tropical instability waves. *Journal of Physical Oceanography* **36**: 592–605.

Kalnay E, *et al*. 1996. The NCEP/NCAR 40-year reanalysis project. *Bulletin of the American Meteorological Society* **77**: 437–470.

Kolodziejczyk N, Bourlès B, Marin F, Grelet J, Chuchla R. 2009. Seasonal variability of the Equatorial Undercurrent at 10 W as inferred from recent in situ observations. *Journal of Geophysical Research* **114**: C06014. DOI:10.1029/2008JC004976.

Kushnir Y, Robinson WA, Chang P, Robertson AW. 2006. The physical basis for predicting Atlantic sector seasonal-to-interannual climate variability. *Journal of Climate* **19**: 5949–5970.

Lübbecke JF, Böning CW, Keenlyside NS, Xie S-P. 2010. On the connection between Benguela and Equatorial Atlantic Ninos and the role of the South Atlantic Anticyclone. *Journal of Geophysical Research* DOI:10.1029/2009JC005964, in press.

Marin F, Caniaux G, Bourlès B, Giordani H, Gouriou Y, Key E. 2009. Why were sea surface temperatures so different in the eastern equatorial Atlantic in June 2005 and 2006?. *Journal of Physical Oceanography* **39**: 1416–1431. DOI: 10.1175/2008JPO4030.1.

Menkes CER, Vialard JG, Kennan SC, Boulanger J-P, Madec GV. 2006. A modelling study of the impact of tropical instability waves on the heat budget of the eastern equatorial Pacific. *Journal of Physical Oceanography* **36**: 847–865.

Okumura Y, Xie S-P. 2004. Interaction of the Atlantic equatorial cold tongue and the African monsoon. *Journal of Climate* **17**: 3589–3602.

Peter A-C, Le Hénaff M, du Penhoat Y, Menkes CE, Marin F, Vialard J, Caniaux G, Lazar A. 2006. A model study of the seasonal mixed layer heat budget in the equatorial Atlantic. *Journal of Geophysical Research* **111**: C06014. DOI:10.1029/2005JC003157.

Polo I, Lazar A, Rodríguez-Fonseca B, Arnault S. 2008. Oceanic Kelvin waves and tropical Atlantic intraseasonal variability: 1. Kelvin wave characterization. *Journal of Geophysical Research* **113**: C07009. DOI:10.1029/2007JC004495.

Rayner NA, Parker DE, Horton EB, Folland CK, Alexander LV, Rowell DP, Kent EC, Kaplan A. 2003. Global analyses of sea surface temperature, sea ice, and night marine air temperature since the late nineteenth century. *Journal of Geophysical Research* **108**: 4407. DOI:10.1029/2002JD002670.

Rhein M, Dengler M, Sültenfuß J, Hummels R, Hüttl-Kabus S, Bourlès B. 2010. Upwelling and associated heat flux in the equatorial Atlantic inferred from helium isotope disequilibrium. *Journal of Geophysical Research* DOI:10.1029/2009JC005772, in press.

Rodríguez-Fonseca B, Janicot S, Mohino E, Losada T, Bader J, Caminade C, Chauvin F, Fontaine B, García-Serrano J, Gervois S, Joly M, Polo I, Ruti P, Roucou P, Voldoire A. 2011. Interannual and decadal SST-forced responses of the West African monsoon. *Atmospheric Science Letters* **12**: 67–74, DOI: 10.1002/asl.308.

Servain J, Picaut J, Merle J. 1982. Evidence of remote forcing in the equatorial Atlantic Ocean. *Journal of Physical Oceanography* **12**: 457–463.

Servain J, Wainer I, McCreary JP, Jr. Dessier A. 1999. Relationship between the equatorial and meridional modes of climatic variability in the tropical Atlantic. *Geophysical Research Letters* **26**: 485–488.

von Schuckmann K, Brandt P, Eden C. 2008. Generation of tropical instability waves in the Atlantic Ocean. *Journal of Geophysical Research* **113**: C08034. DOI:10.1029/2007JC004712.

Wallace JM, Mitchell TP, Deser C. 1989. The influence of sea-surface temperature on surface wind in the eastern equatorial Pacific: Seasonal and interannual variability. *Journal of Climate* **2**: 1492–1499.

Weisberg RH, Horigan A, Colin C. 1979. Equatorially trapped Rossby-gravity wave-propagation in the Gulf of Guinea. *Journal of Marine Research* **37**: 67–86.

Weisberg RH, Weingartner TJ. 1988. Instability waves in the equatorial Atlantic Ocean. *Journal of Physical Oceanography* **18**: 1641–1657.

Zebiak SE. 1993. Air-sea interaction in the equatorial region. *Journal of Climate* **6**: 1567–1586.

