GEOPHYSICAL RESEARCH LETTERS, VOL. 34, L02703, doi:10.1029/2006GL028579, 2007

Atlantic warm pool, Caribbean low-level jet, and their potential impact on Atlantic hurricanes

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Received 25 October 2006; revised 20 November 2006; accepted 13 December 2006; published 18 January 2007.

[1] The Atlantic Warm Pool (AWP) is a large body of warm water (warmer than 28.5°C) that appears in the Gulf of Mexico, the Caribbean Sea, and the western tropical North Atlantic during the summer and fall. Located to its northeastern side is the North Atlantic Subtropical High (NASH) that produces the easterly trade winds in the tropics. The trade winds carry moisture from the tropical North Atlantic into the Caribbean Sea where the flow intensifies forming the Caribbean Low-Level Jet (CLLJ). This paper finds that the easterly CLLJ is maximized in the summer and winter, whereas it is minimized in the fall and spring. The semi-annual feature of the CLLJ results from the semi-annual variation of sea level pressure in the Caribbean region associated with the east-west excursion of the NASH. The AWP's impact is to weaken the summertime NASH, especially at its southwestern edge and thus weaken the easterly CLLJ. The weakening of the easterly CLLJ, in conjunction with the AWP-induced change of upper-level wind, reduces the tropospheric vertical wind shear that favors hurricane formation and intensification during August–October. **Citation:** Wang, C., and S.-K. Lee (2007), Atlantic warm pool, Caribbean low-level jet, and their potential impact on Atlantic hurricanes, *Geophys. Res. Lett.*, *34*, L02703, doi:10.1029/2006GL028579.

1. Introduction

[2] The Atlantic Warm Pool (AWP), the region that covers the Gulf of Mexico, the Caribbean Sea, and the western tropical North Atlantic (Figure 1a), is the second largest body of very warm water on Earth. Unlike the Indo-Pacific warm pool, which straddles the equator, the AWP lies entirely north of the equator. During the winter, there is no water warmer than 28.5°C in the AWP region (Figure 1b). Water warmer than 28.5°C starts to appear in the ocean region surrounding Cuba in June (not shown). By July, water warmer than 28.5°C is well developed in the Gulf of Mexico. In August, the warm water in the Gulf of Mexico reaches its maximum with a large area covered by water warmer than 29.5°C. By September, the warm pool has expanded south into the Caribbean Sea and eastward into the western tropical North Atlantic, while the water in the Gulf of Mexico has cooled. In addition to the large seasonal cycle, the AWP also shows a large interannual fluctuation in its area [*Wang and Enfield*, 2001, 2003; *Wang et al.*, 2006]. The AWP is a climatic nexus for the North and South Americas as well as for the tropical Pacific and Atlantic Oceans, and is a moisture source for rainfall in North, Central, and South Americas [e.g., *Helfand and Schubert*, 1995; *Mo et al.*, 1997; *Wang et al.*, 2006]. It is also in the path of, and a birthplace for Atlantic hurricanes that cause loss of life and huge damage over the neighboring land areas.

[3] Located to the northeastern side of the AWP is the North Atlantic Subtropical High (NASH)—a high sea level pressure centered in the eastern portion of the subtropical Atlantic Ocean. The NASH produces the easterly trade winds in the tropical North Atlantic that carry moisture into the Caribbean Sea where the winds are intensified to form a maximum of easterly zonal winds (larger than 13 m/s) at 925-mb called the Caribbean Low-Level Jet (CLLJ) [*Amador*, 1998]. In the summer, the CLLJ splits into two branches: one turning northward into the United States, and the other one continuing westward across Central America into the eastern North Pacific. Thus, the CLLJ serves as a moisture supplier for rainfall in the United States, Central America, and the eastern North Pacific. Before Atlantic hurricanes make landfall in the North and Central Americas, they pass over the AWP region. It is thus expected that Atlantic hurricane activity would be related to the AWP and the CLLJ.

[4] The purpose of this paper is to document a newly-found feature of the CLLJ: The CLLJ varies semi-annually owing to the semi-annual feature of the Sea Level Pressure (SLP) in the Caribbean region associated with the westward extension and eastward retreat of the NASH. This study also investigates relationships among the AWP, the CLLJ, and Atlantic hurricane activity.

2. Data and Model

[5] The observational data include SLP, winds, and specific humidity from the National Centers for Environmental Prediction-National Centers for Atmospheric Research (NCEP-NCAR) reanalysis fields [*Kalnay et al.*, 1996], and Sea Surface Temperature (SST) from the Hadley Centre, United Kingdom [*Rayner et al.*, 2003]. The atmospheric general circulation model used in this study is the latest version (version 3.1) of the NCAR community atmospheric model (T42). Two sets of ensemble (with 18 members) simulations are conducted. In the control run, the monthly climatological SST is prescribed globally for forcing the model. In the second set of simulations, SST in the AWP region (from 5°N to 30°N between 40°W and the coast of the Americas) is held to its January value (Figure 1) while the twelve-monthly climatology is specified for the

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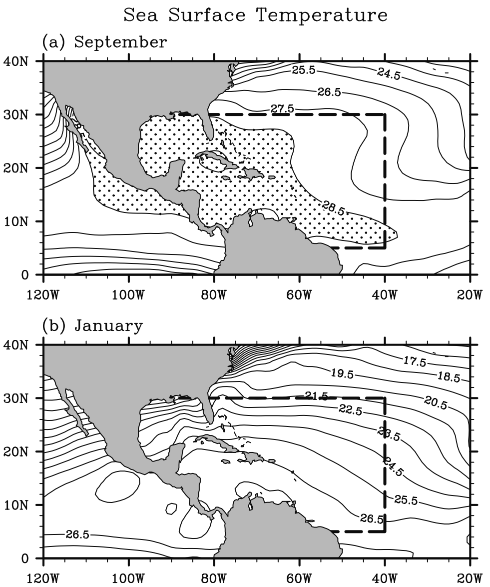


Figure 1. SST distributions near the region of the Atlantic Warm Pool (AWP) in (a) September and (b) January. The contour interval is 1.0°C and SST warmer than 28.5°C is stippled. The AWP box (from 5°N to 30°N between 40°W and the coast of the Americas) is marked. For the ensemble model run without the AWP, SST in the AWP box is held to its January value while the twelve-monthly SST climatology is specified for the rest of the global ocean. Note that the SST in the model run without the AWP is not smoothed around the edge of the AWP box since the model results are not affected by how the AWP is removed (Wang et al., submitted manuscript, 2006).

rest of the global ocean. By doing so, the warm-season AWP is removed from the model SST forcing. To isolate the AWP's effect, the difference is taken between the model runs with and without the AWP.

3. Caribbean Low-Level Jet

[6] The NASH is an important factor in determining atmospheric and oceanic circulations as well as the distribution of thermodynamical fields over the northern Atlantic. In the summer, the NASH is in a cell-type configuration with a well-defined core region located in the eastern subtropical Atlantic Ocean where the ocean temperature is relatively cool (Figures 2a and 2b). The model ensemble runs show that the AWP weakens the NASH with the largest decrease of SLP at its southwest edge, and it also strengthens the summertime continental low in the North American monsoon region (Figure 2c). This low pressure pattern results in westerly winds (the wind difference between the model runs with and without the AWP) over the Caribbean and eastern North Pacific, and northerly winds south of the Great Plains region. The physical mechanism for the response of the SLP to the AWP is explained in terms of the simple theory of *Gill* [1980]. For an off-equatorial heating anomaly, Gill's theory predicts an atmospheric response involving low pressure to the northwest of the heating, associated with a Rossby wave. To test consistency with Gill's dynamics, we perform two simple model runs with specified heating in the AWP region using: (1) the simple Gill's model, and (2) the NCAR general circulation model in an Aquaplanet setting (i.e., no land). Both model runs show a similar atmospheric response to Figure 2c (not shown). This indeed indicates that the SLP and wind distributions in Figure 2c are responses of AWP's heating, consistent with Gill's theoretical work.

[7] As shown in Figure 2, the NASH induces the easterly trade winds in the tropics that carry moisture from the tropical North Atlantic into the Caribbean Sea where the flow intensifies forming the CLLJ. The feature of the CLLJ is clearly seen by plotting the vertical-meridional section of the zonal moisture transport distribution of *qu* where *q* is specific humidity and *u* is zonal wind. Figures 3a and 3b show that the CLLJ core of maximum easterly moisture transport is located between 10°N-20°N in the lower troposphere extending to the surface. The *qu* difference between the model runs with and without the AWP in Figure 3c shows a positive zonal moisture transport difference around the CLLJ region, indicating that the AWP's impact is to weaken the easterly CLLJ. The positive zonal moisture transport difference in Figure 3c is mainly attributed to

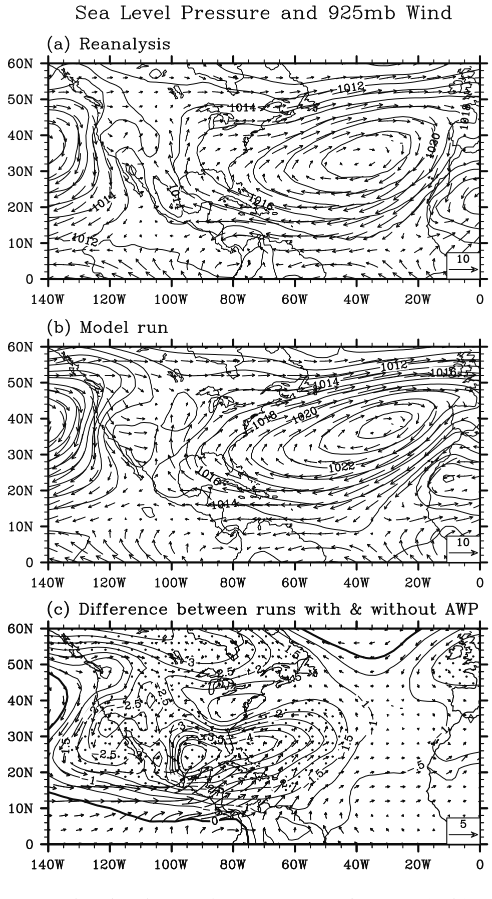
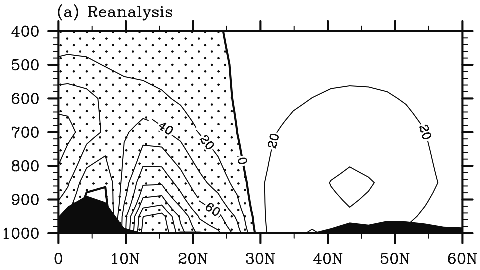


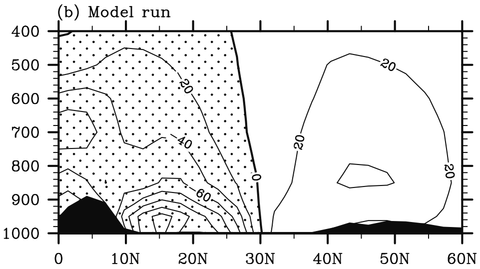
Figure 2. The horizontal structure of SLP (mb) and the 925-mb wind vector (m s^{-1}) during the summer (June to August) from (a) the reanalysis, (b) the ensemble model control run, and (c) the difference between the model runs with and without the AWP. In Figures 2a and 2b, the contour interval is 2.0 mb. In Figure 2c, the contour line of the zero SLP difference is in bold, the contour interval is 0.5 mb, and the SLP difference lower than −1.5 mb is stippled.

Zonal Moisture Transport at 75°W

(a) Reanalysis



(b) Model run



(c) Difference between runs with & without AWP

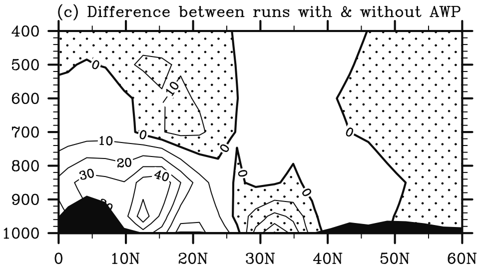


Figure 3. The meridional-vertical sections of the zonal moisture transport of qu (where q is specific humidity and u is zonal wind; in unit of g kg^{-1} m s^{-1}) at 75°W during the summer (June to August) from (a) the reanalysis, (b) the ensemble model control run, and (c) the difference between the model runs with and without the AWP. The unit in the vertical direction is mb. The negative value of the zonal moisture transport is stippled. The contour interval is 20 g kg^{-1} m s^{-1} in Figures 3a and 3b, and 10 g kg^{-1} m s^{-1} in Figure 3c.

the AWP-induced reduction of easterly wind (i.e., $q\Delta u$ where Δu represents the zonal wind difference between the model runs with and without the AWP) near the CLLJ's core, whereas the change of specific humidity (i.e., $u\Delta q$) makes a small contribution to the positive moisture transport difference (not shown).

[8] The seasonal variations of the CLLJ are examined by plotting the time-latitude section of the vertically integrated zonal moisture transport at 75°W, as shown in Figure 4. Both the observational data and the model show that the westward moisture transport between 10°N–20°N is maximized in the summer and winter, whereas it is minimized in the spring and fall (Figures 4a and 4b). That is, the CLLJ has a semi-annual feature. We perform a harmonic analysis of the zonal moisture transport between 12.5°N–17.5°N in Figure 4a. The semi-annual component explains about 56% of the seasonal variance while the annual component explains only 39%. The 925-mb zonal wind at 75°W between 10°N–20°N also displays a large semi-annual variation (not shown). Early work on the CLLJ describes it as a single peak in the summer [*Amador*, 1998; *Amador and Magaña*, 1999]. The present paper is the first to document the semi-annual feature of the CLLJ. The semi-annual feature of the CLLJ follows the semi-annual cycle of the SLP in the region of the Caribbean Sea (Figures 4c and 4d). Both the observational data and model control run clearly display a semi-annual variation: SLP is relatively high in the summer and winter, and it is relatively low in the spring and fall. The semi-annual variation of SLP (and the associated meridional SLP gradient) thus produces the semi-annual feature of the CLLJ. In other words, the high (low) pressure and meridional pressure gradient in the Caribbean during the summer and winter (spring and fall) results in the strong (weak) easterly CLLJ. July is associated with the well-known mid-summer drought in the region near Central America [e.g., *Magaña et al.*, 1999; *Mapes et al.*, 2005]. The number of tropical cyclones in the Caribbean Sea also exhibits a distinct bimodal distribution, with peaks in June and October separated by a significant minimum of tropical cyclones in July (see Figures 1 and 2a in *Inoue et al.* [2002]). These suggest that the high Caribbean SLP and the strong easterly CLLJ during the summer are related to the mid-summer drought and tropical cyclogenesis in the Caribbean Sea.

[9] The question is why the SLP in the Caribbean region varies semi-annually. The NASH is strongest in the summer with a cell-type configuration extending toward the Caribbean. As the season progresses toward the fall, the NASH weakens and its center moves eastward. In the winter, since a continental high develops over North America, the NASH's isobars extend westward and connect with the North American high. As the North American monsoon starts to develop in the spring, the NASH's isobars retreat toward the east. Thus, the yearly movement and development of the NASH result in a semi-annual feature of the SLP in the region of the Caribbean.

4. Atlantic Hurricane Activity Inferred From Vertical Wind Shear Changes

[10] The recent increase in Atlantic hurricane activity has fueled a debate on the contribution of the natural climate variability versus global warming to the increase [e.g., *Goldenberg et al.*, 2001; *Emanuel*, 2005; *Webster et al.*, 2005]. However, all of these studies agree that the increase of Atlantic hurricane activity corresponds to an increase of SST in the Atlantic Ocean. The 2005 hurricane season is the most active year on record with 28 named storms (including 15 hurricanes), associated with a very large AWP [*Virmani and Weisberg*, 2006]. We here demonstrate why and how the local SST in the AWP region may affect Atlantic hurricanes.

[11] It is well accepted that a strong vertical wind shear between the upper and lower troposphere in the hurricane main development region (from 10°N to 20°N between the west coast of Africa to Central America) inhibits the formation and intensification of tropical cyclones [e.g., *Gray*, 1968; *Pasch and Avila*, 1992], as it prevents organization of deep convection. The AWP's influence on the vertical wind shear is shown in Figure 5, which displays the vertical wind shear during August to October from the reanalysis, the model control run, and the difference between the model runs with and without the AWP. The

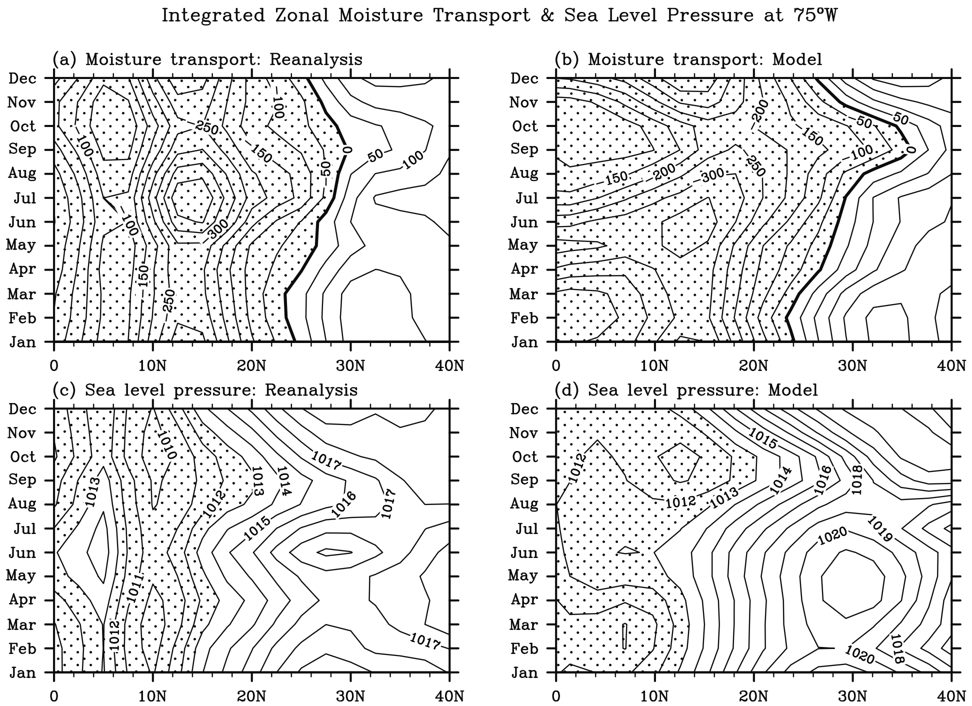


Figure 4. The time-latitude sections of the vertically integrated zonal moisture transport ($\int_{300\text{mb}}^{sfc} (qu/g)\, dp$; kg m^{-1} s^{-1}) at 75°W from (a) the reanalysis and (b) the ensemble model control run, and the SLP (mb) at 75°W from (c) the reanalysis and (d) the ensemble model control run. In Figures 4a and 4b, the negative value of the vertically integrated zonal moisture transport is stippled and the contour interval is 50 kg m^{-1} s^{-1}. In Figures 4c and 4d, the SLP lower than 1013 mb is stippled and the contour interval is 1.0 mb.

months of August to October are chosen because most of Atlantic hurricanes occur during August to October. In comparison with the mid-latitudes, the tropical and subtropical regions show a relatively low value of the vertical shear during August to October (Figures 5a and 5b). The AWP's effect on the vertical wind shear is clearly seen in Figure 5c that shows a longitudinal band of negative wind shear difference in and just north of the hurricane main development region. A calculation shows, on average, a decrease of the vertical wind shear of more than 5.0 m/s in the main development region. As shown in Section 3, the AWP largely reduces the low-level easterly trade winds in the region around the Caribbean in the summer and fall (i.e., a weakening of the CLLJ). At the same time, the AWP also induces a decrease of westerly wind in the upper troposphere (not shown), in response to the AWP's heating [*Gill*, 1980]. The combination of the AWP-induced lower and upper tropospheric wind changes shows a large reduction of the vertical wind shear in the model runs. These model experiments suggest that the AWP is important for reducing the tropospheric vertical wind shear, thus favoring hurricane intensification and formation in August to October.

5. Summary and Discussion

[12] The NASH produces the tropical easterly trade winds at its southern flank. When the easterly trade winds flow westward into the Caribbean Sea, the pressure gradient set up by the NASH becomes large and thus the easterly winds intensify forming the CLLJ. We find that the westward CLLJ's moisture transport has a semi-annual feature: two maxima in the summer and winter, and two minima in the spring and fall. This is because the SLP and meridional SLP gradient in the Caribbean region vary semi-annually in response to the east-west excursion of the NASH. Our model ensemble runs show that the AWP's effect is to weaken the summertime NASH, especially at its southwestern edge. The AWP also strengthens the summertime continental low over the North American monsoon region. In response to these pressure changes, the CLLJ's strength is weakened. In addition to the weakening of the easterly CLLJ, the AWP also weakens the upper tropospheric westerly wind. Thus, the AWP largely reduces the tropospheric vertical wind shear in the hurricane main development region that favors hurricane formation and intensification during August–October.

[13] Since the early report of the CLLJ as a single peak in the summer [*Amador*, 1998], few studies have focused on the CLLJ. We do not know any papers in the literature that document the secondary winter maximum of the CLLJ. Our model results show that the AWP weakens the CLLJ's strength. However, the AWP does not change the semi-annual feature of the CLLJ. As shown by C. Wang et al. (Impact of the Atlantic warm pool on the summer climate of the Western Hemisphere, submitted to *Journal of Climate*, 2006, hereinafter referred to as Wang et al., submitted manuscript, 2006), the semi-annual feature of the CLLJ still exists with the AWP removed. This is because a removal of the AWP does not change the east-west excursion of the NASH that is associated with the semi-annual variation of the CLLJ. Since the Caribbean Sea and the Gulf Mexico serve as a source of atmospheric moisture

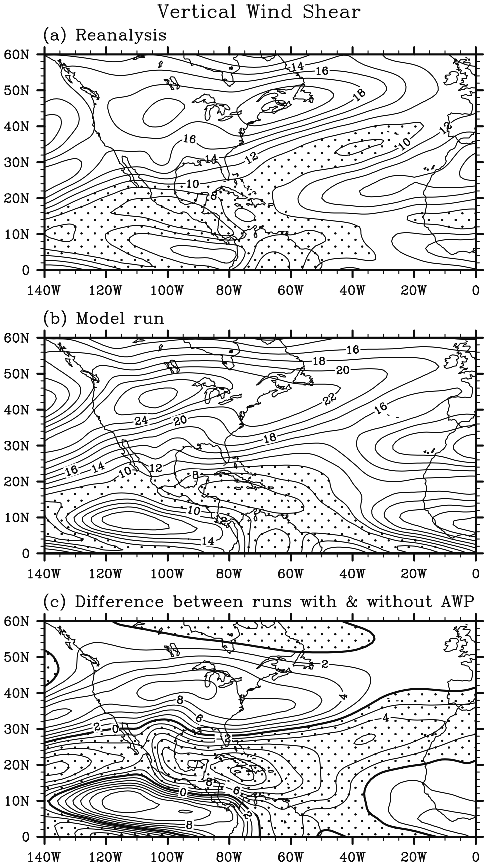


Figure 5. The tropospheric vertical wind shear ($[(U_{200} - U_{850})^2 + (V_{200} - V_{850})^2]^{1/2}$; m s^{-1}) during August to October from (a) the reanalysis, (b) the ensemble model control run, and (c) the difference between the model runs with and without the AWP. The contour interval is 2.0 m s^{-1}. In Figures 5a and 5b, the vertical wind shear smaller than 10 m s^{-1} is stippled. In figure 5c, the negative value of the wind shear difference is stippled.

for rainfall over the Americas [e.g., *Rasmusson*, 1967; *Brubaker et al.*, 2001], it is expected that the CLLJ is a potential carrier of the exported moisture. The CLLJ thus may be important for the climate over the Americas and merits further studies.

[14] This paper shows a linkage among the AWP, the CLLJ, and hurricanes through the dynamical parameter of the tropospheric vertical wind shear. In nature, there are many other (dynamical and thermodynamical) parameters that affect hurricane formation and development. For example, the moist static instability of the troposphere can affect hurricanes. The moist static instability can be measured by the convective available potential energy (CAPE) that provides the fuel for moist convection and thus is a potential indicator of hurricane intensity [*Emanuel*, 1994]. Our model ensemble runs with and without the AWP show that the AWP increases the CAPE by 72% over the main development region (not shown). Thus, the AWP is important for Atlantic hurricanes through both the dynamical parameter of the vertical wind shear and the thermodynamical parameter of the moist static instability.

[15] **Acknowledgments.** We thank two anonymous reviewers for their comments that helped improve the paper. This work was supported by a grant from the National Oceanic and Atmospheric Administration (NOAA) Climate Program Office and by the base funding of NOAA Atlantic Oceanographic and Meteorological Laboratory (AOML). The findings and conclusions in this report are those of the authors and do not necessarily represent the views of the funding agency.

References

Amador, J. A. (1998), A climatic feature of the tropical Americas: The trade wind easterly jet, *Tóp. Meteorol. Oceanogr.*, *5*(2), 1–13.

Amador, J. A., and V. Magaña (1999), Dynamics of the low level jet over the Caribbean Sea, paper presented at 23rd Conference on Hurricanes and Tropical Meteorology, Am. Meteorol. Soc., Dallas, Tex., 10–15 Jan.

Brubaker, K. L., P. A. Dirmeyer, A. Sudradjat, B. Levy, and F. Bernal (2001), A 36-yr climatological description of the evaporative sources of warm season precipitation in the Mississippi River basin, *J. Hydrometeorol.*, *2*, 537–557.

Emanuel, K. A. (1994), *Atmospheric Convection*, 580 pp., Oxford Univ. Press, New York.

Emanuel, K. A. (2005), Increasing destructiveness of tropical cyclones over the past 30 years, *Nature*, *436*, 686–688.

Gill, A. E. (1980), Some simple solutions for heat-induced tropical circulation, *Q. J. R. Meteorol. Soc.*, *106*, 447–462.

Goldenberg, S. B., C. Landsea, A. M. Mestas-Nuñez, and W. M. Gray (2001), The recent increase in Atlantic hurricane activity, *Science*, *293*, 474–479.

Gray, W. M. (1968), Global view of the origins of tropical disturbances and storms, *Mon. Weather Rev.*, *96*, 669–700.

Helfand, H. M., and S. D. Schubert (1995), Climatology of the simulated Great Plains low level jet and its contribution to the continental moisture budget of the United States, *J. Clim.*, *8*, 784–806.

Inoue, M., I. C. Handoh, and G. R. Bigg (2002), Bimodal distribution of tropical cyclogenesis in the Caribbean: Characteristics and environmental factors, *J. Clim.*, *15*, 2897–2905.

Kalnay, E., et al. (1996), The NCEP/NCAR 40-year reanalysis project, *Bull. Am. Meteorol. Soc.*, *77*, 437–471.

Magaña, V., J. A. Amador, and S. Medina (1999), The midsummer drought over Mexico and Central America, *J. Clim.*, *12*, 1577–1588.

Mapes, B. E., P. Liu, and N. Buenning (2005), Indian monsoon onset and the Americas midsummer drought: Out-of-equilibrium response to smooth seasonal forcing, *J. Clim.*, *18*, 1109–1115.

Mo, K. C., J. N. Paegle, and R. W. Higgins (1997), Atmospheric processes associated with summer floods and droughts in the central United States, *J. Clim.*, *10*, 3028–3046.

Pasch, R. J., and L. A. Avila (1992), Atlantic hurricane season of 1991, *Mon. Weather Rev.*, *120*, 2671–2687.

Rasmusson, E. M. (1967), Atmospheric water vapor transport and the water balance of North America: part I. Characteristics of the water vapor flux field, *Mon. Weather Rev.*, *95*, 403–426.

Rayner, N. A., D. E. Parker, E. B. Horton, C. K. Folland, L. V. Alexander, D. P. Rowell, E. C. Kent, and A. Kaplan (2003), Global analyses of sea surface temperature, sea ice, and night marine air temperature since the late nineteenth century, *J. Geophys. Res.*, *108*(D14), 4407, doi:10.1029/2002JD002670.

Virmani, J. I., and R. H. Weisberg (2006), The 2005 hurricane season: An echo of the past or a harbinger of the future?, *Geophys. Res. Lett.*, *33*, L05707, doi:10.1029/2005GL025517.

Wang, C., and D. B. Enfield (2001), The tropical Western Hemisphere warm pool, *Geophys. Res. Lett.*, *28*, 1635–1638.

Wang, C., and D. B. Enfield (2003), A further study of the tropical Western Hemisphere warm pool, *J. Clim.*, *16*, 1476–1493.

Wang, C., D. B. Enfield, S.-K. Lee, and C. W. Landsea (2006), Influences of the Atlantic warm pool on Western Hemisphere summer rainfall and Atlantic hurricanes, *J. Clim.*, *19*, 3011–3028.

Webster, P. J., G. J. Holland, J. A. Curry, and H.-R. Chang (2005), Changes in tropical cyclone number, duration, and intensity in a warming environment, *Science*, *309*, 1844–1846.

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