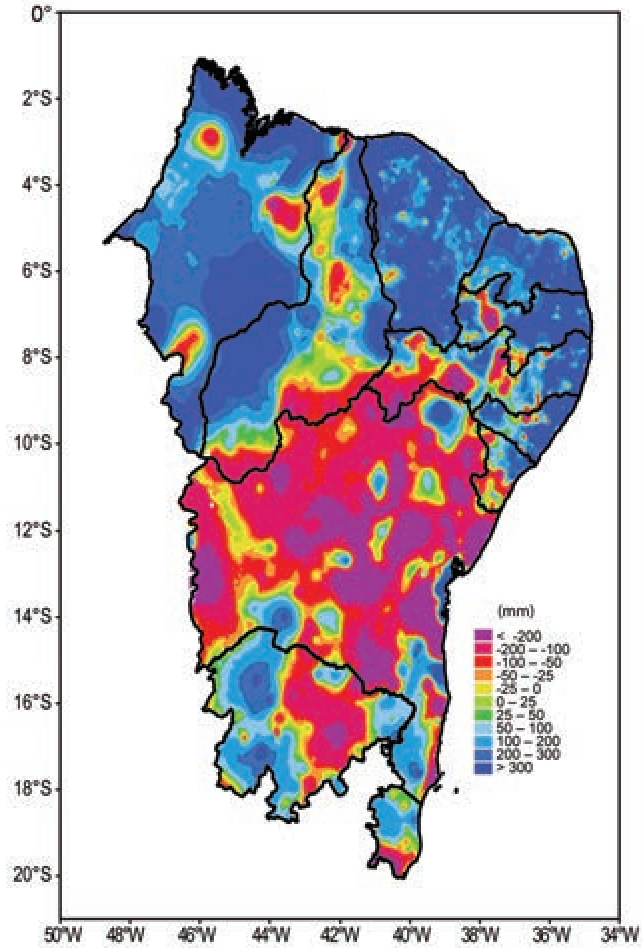

FIG. 4.34. Northeastern Brazil average 2011 precipitation anomaly (mm) with respect to 1961–90 climatology based on high-resolution station data. (Source: federal and regional networks CMCD/INPE, INMET, SUDENE, ANEEL, FUNCEME/CE, LMRS/PB, EMPARN/RN, LAMEPE/ITEP/PE, CMRH/SE, SEAAB/PI, SRH/BA, CEMIG/SIMGE/MG, and SEAG/ES)

that had previously been established in northeastern Brazil and parts of the Amazon in the previous year.

While the La Niña conditions indirectly helped enhance the ITCZ via Kelvin wave-induced upper-level divergence in the Atlantic sector, the demise of record-breaking warm SST anomalies observed over the subtropical North Atlantic (Fig. 4.33a) played a fundamental role via a favorable (enhanced) meridional SST gradient, expressed here by the Atlantic Index (see Fig. 4.33b caption for definition). A positive index in 2011, contrasting with the record-breaking negative index of 2010, implies a return to a more favorable pattern for frequent bursts of organized convection in the southern part of the basin and ultimately an overall enhancement of the ITCZ. Within this scenario, the ITCZ oscillated around its average climatological position for most of the year, with precipitation above average in parts of the eastern Amazon and northeastern Brazil (Fig. 4.34). This effect was due to the explicit dynamical response to upper-level divergence as well as the direct evaporative effect of the onshore trade winds responding to the north-south water temperature gradient, giving a typical La Niña response both in terms of positioning and strength of the ITCZ in 2011.

g. Atlantic multidecadal oscillation—C. Wang

The Atlantic Multidecadal Oscillation (AMO) is an oscillatory mode defined by the detrended North Atlantic SST anomalies over the region of 0°–60°N and from the east coast of the Americas to 0° longitude (Figs. 4.35a,b; Delworth and Mann 2000; Enfield et al. 2001; Wang et al. 2008a). A driving mechanism for the AMO may be the Atlantic meridional overturning circulation (Delworth and Mann 2000; Knight et al. 2005; Dijkstra et al. 2006; R. Zhang et al. 2007; see also section 3h of this report for detailed information on the meridional overturning circulation). A new study shows that the AMO varies with dust aerosol in the tropical North Atlantic and rain-

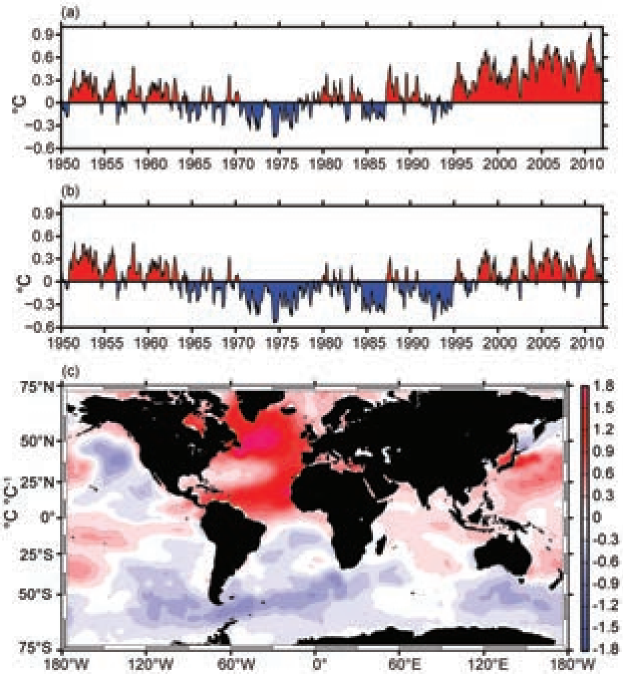

FIG. 4.35. The index of the AMO and its spatial pattern. Shown are (a) SST anomalies (°C) in the North Atlantic of 0°–60°N and from the east coast of the Americas to 0° longitude, (b) the AMO index (°C) defined by the detrended (removing the linear trend) North Atlantic SST anomalies, and (c) regression (°C per °C) of global SST anomalies onto the AMO index of (b). The monthly SST anomalies are calculated as departures from the 1971–2000 base period.

fall in the Sahel, suggesting that a positive feedback exists between North Atlantic SST, dust aerosol, and Sahel rainfall on multidecadal timescales (C. Wang et al. 2012). In addition to multidecadal variability, the AMO also demonstrates high-frequency variability such as an interannual variation via its connection to the tropical North Atlantic and the Atlantic Warm Pool (AWP; a large body of warm water comprising the Gulf of Mexico, Caribbean Sea, and the western tropical North Atlantic), and as such has exhibited a seasonal influence on the behavior of tropical cyclones (TCs) in the Atlantic and Eastern North Pacific (ENP) basins. The extended reconstructed SST (ERSST) data from 1950 to 2011 (Fig. 4.35b) shows that the AMO was in the cold phase from the late 1960s to the early 1990s and in the warm phase before the late 1960s and again after the early 1990s (Enfield et al. 2001). The AMO is related to SST anomalies over the global oceans as shown in Fig. 4.35c.

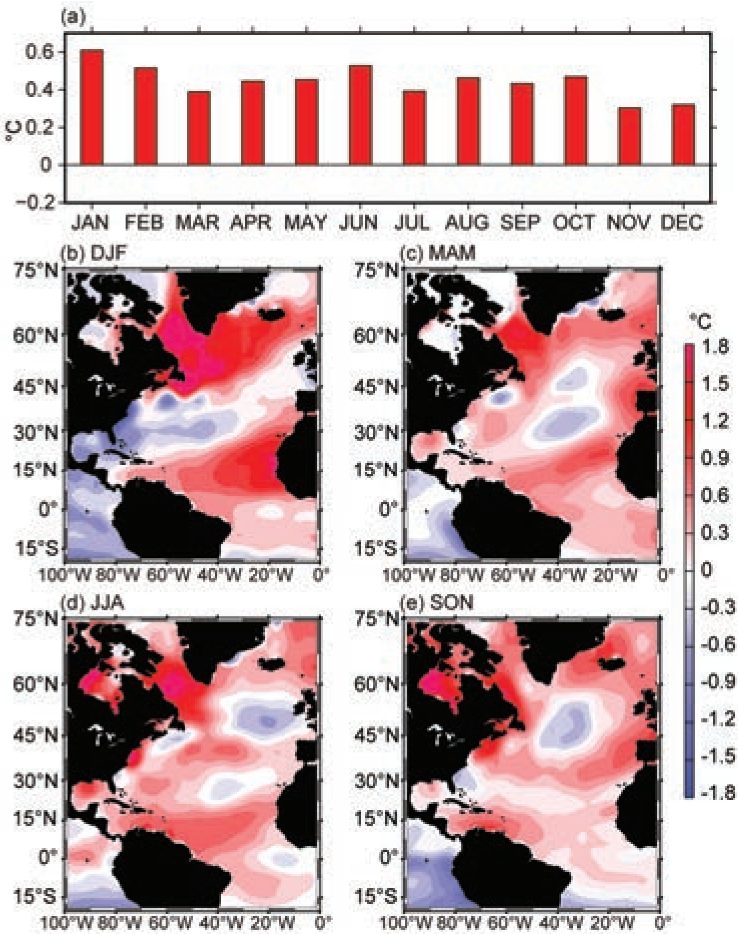

Fig. 4.36. The AMO in 2011. Shown are the (a) monthly North Atlantic SST anomalies (°C), (b) DJF 2010/11 SST anomalies (°C), (c) MAM SST anomalies, (d) JJA SST anomalies (°C), and (e) SON SST anomalies (°C). The monthly SST anomalies are calculated as departures from the 1971–2000 base period.

The AMO variability is associated with changes of climate and extreme events, such as drought and flood in North America and Europe, and hurricane activity in the North Atlantic and ENP (Enfield et al. 2001; McCabe et al. 2004; Goldenberg et al. 2001; Bell and Chelliah 2006; Wang et al. 2008a; Wang and Lee 2009). Recent studies show that the relationship of the AMO to TCs is due to its tropical component since the climate response to the North Atlantic SST anomalies is primarily forced at the low latitudes (e.g., Lu and Delworth 2005; Sutton and Hodson 2007). The AMO variability coincides with the multidecadal signal of the AWP and the AWP is in the path of, or a birthplace for, Atlantic TCs (Wang et al. 2008a). Thus, the influence of the AMO on climate and Atlantic TC activity operates through the mechanism of the AWP-induced atmospheric changes by having an effect on vertical wind shear in the hurricane main development region (MDR). A large AWP reduces such shear, while a small AWP enhances it. A large AWP also weakens the southerly Great Plains low-level jet, thus reducing the northward moisture transport from the Gulf of Mexico to the eastern United States and decreasing the boreal summer rainfall over the central US, while a small AWP has the opposite effect (Wang et al. 2006, 2008b). It has also been shown that the AWP variability can produce the observed out-of-phase relationship between TC activity in the tropical North Atlantic and ENP (Wang and Lee 2009).

As of 2011, the AMO has been in its warm phase since 1995. For all months in 2011, the North Atlantic SST anomalies were positive (Fig. 4.36a), with the maximum of the North Atlantic SST anomalies in January (+0.61°C) and the minimum in November (+0.30°C). Spatially, the North Atlantic SST anomalies during the boreal winter showed a tripole pattern with the positive SST anomalies in the subpolar North Atlantic and the tropical North Atlantic and the negative SST anomalies in the subtropical North Atlantic (Fig. 4.36b). The SST anomaly pattern divided the AWP into two parts: a colder Gulf of Mexico and a warmer Caribbean Sea/western tropical North Atlantic. The opposite SST anomaly pattern was consistent with a previous study (Muñoz et al. 2010), which showed that the air-sea fluxes as-

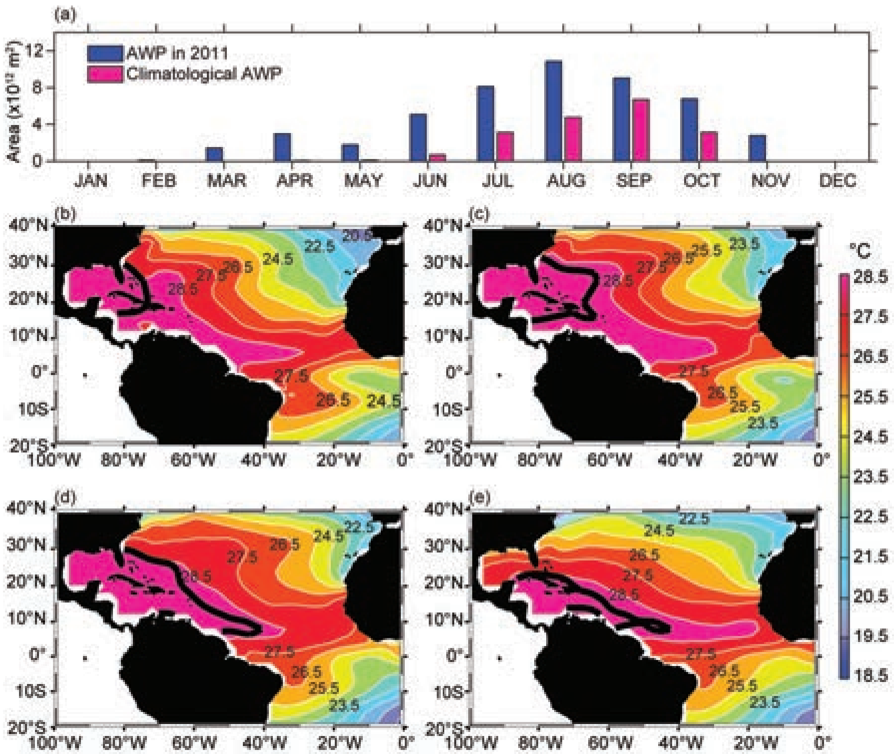

Fig. 4.37. The AWP in 2011. Shown are the (a) monthly AWP area in 2011 (10^{12} m^2; blue) and the climatological AWP area (red), spatial distributions of the 2011 AWP in (b) July, (c) August, (d) September, and (e) October. The AWP is defined by SST larger than 28.5°C. The black contours in (b)–(e) are the climatological AWP based on the data from 1971–2000.

sociated with ENSO events in the tropical Pacific and local processes were responsible for the SST anomaly distribution. During the boreal spring, summer, and fall of 2011, the cold SST anomalies in the subtropical North Atlantic weakened and the North Atlantic was mostly warm (Figs. 4.36c,d,e).

Previous studies have shown that the influence of the AMO on TC activity may operate through the atmospheric changes induced by the AWP or the Atlantic meridional mode (Wang et al. 2008a; Vimont and Kossin 2007). In particular, the AWP can play an important role in the hurricane track (Wang et al. 2011). During the 2011 Atlantic TC season of June to November, the AWP was consistently large (Fig. 4.37a) and the isotherm of 28.5°C extended eastward covering almost the entire tropical North Atlantic (Figs. 4.37b,c,d,e). An eastward expansion of the AWP shifted the hurricane genesis location eastward, decreasing the possibility for a hurricane to make landfall in the southeast United States. A large AWP also tended to shrink the North Atlantic subtropical high eastward (C. Wang et al. 2007) and produced a TC steering flow pattern along the Eastern Seaboard of the United States, which steered hurricanes away from the southeastern coast of the US (Wang et al. 2011). The large AWP in 2011 was also associated

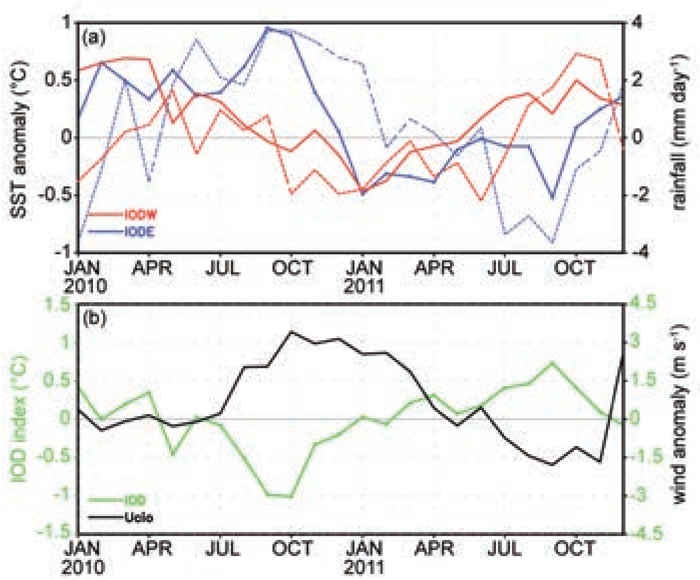

Fig. 4.38. (a) Monthly anomalies of SST (°C, solid lines) and precipitation (mm day^{-1}, dashed lines) in the eastern (IODE: 90E°–110°E, 10°S–0°; blue lines) and western pole (IODW: 50°E–70°E, 10°S–10°N; red lines) of IOD. (b) As in (a), but for the IOD index (measured by the SST difference between IODW and IODE, green line) and surface zonal wind anomaly (m s^{-1}) in the central equatorial IO (Ucio), from 70°E–90°E, 5°S–5°N, black line. The anomalies were calculated relative to the climatology over the period 1982–2010. These are based on the NCEP optimum interpolation SST (Reynolds and Chelton 2010), monthly GPCP precipitation analysis (http://precip.gsfc.nasa.gov/), and JRA-25 atmospheric reanalysis (Onogi et al. 2007).

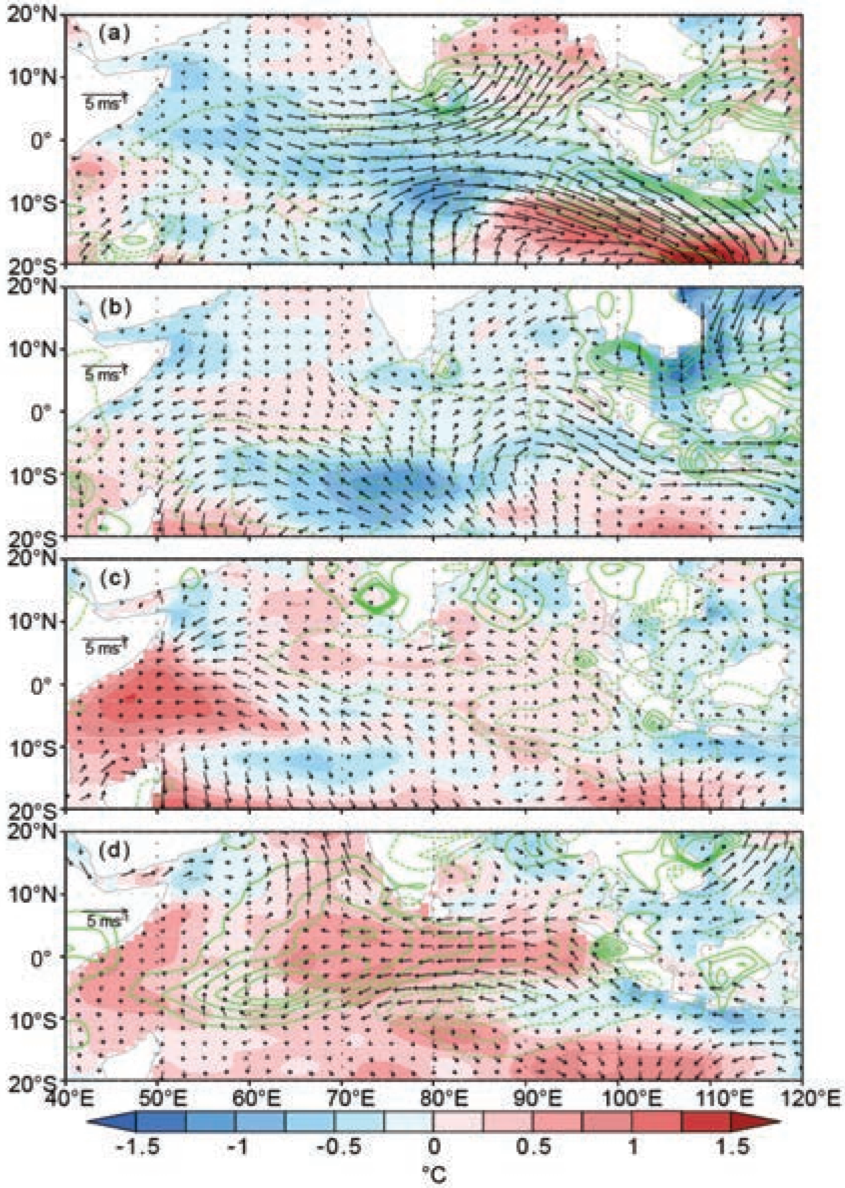

Fig. 4.39. SST (°C, colored scale), precipitation (green contours: ±1, ±2, ..., ±5 mm day^{-1}). Solid (dashed) lines denote positive (negative) values, and surface wind anomalies during (a) DJF 2010/11, (b) MAM 2011, (c) JJA 2011, and (d) SON 2011.

with the out-of-phase relationship between TCs in the North Atlantic and the ENP as described previously in this chapter.

h. *Indian Ocean dipole*—J. J. Luo

The Indian Ocean dipole (IOD)—a climate mode in the tropical Indian Ocean (IO)—usually starts in boreal summer, peaks in fall, and decays rapidly in early winter. It has been believed that El Niño (La Niña) favors the development of positive (negative) IOD. During 2011, however, a weak positive IOD event occurred in the tropical Indian Ocean (Figs. 4.38 and 4.39d) in spite of the La Niña conditions in the Pacific. In particular, the cold SST anomalies in the eastern Indian Ocean were weak,[9] and the 2011 IOD was not well developed during its peak phase in SON compared to the past three strong positive IODs in 1994, 1997, and 2006. This is similar to what happened in 2007 and 2008 (Luo et al. 2008) and reconfirms the idea that the IOD can sometimes originate from the internal air-sea interactions in the IO (Behera et al. 2006; Luo et al. 2010). The strong La Niña event in 2010 helped to induce excessive surface westerlies in the central equatorial IO by enhancing the Walker cell in the IO. The westerly anomalies tend to cool the SST in the western IO by increasing surface latent heat loss from the ocean but warm the SST in the east by deepening the oceanic thermocline there; this favors the occurrence of a negative IOD in the second half of 2010 (Fig. 4.38). After the collapse of the negative IOD near the end of 2010, the excessive westerlies caused a cool/dry condition in most places of the tropical IO and a warm/wet condition in the Bay of Bengal and northwest of Australia during early 2011 (Figs. 4.38a, 4.39a). The related anomalous cyclonic winds in the southeastern IO raised the thermocline underneath and induced strong subsurface cooling and cold SST anomalies along ~10°S. The cold signals propagated westward but disappeared before they reached the western boundary of the IO (Figs. 4.39, 4.40). The drought

[9] Like an El Niño or La Niña event, a weak or strong IOD is defined by a threshold value of one standard deviation of the IOD index.