Remote influences on freshwater flux variability in the Atlantic warm pool region

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Received 9 August 2012; revised 6 September 2012; accepted 11 September 2012; published 11 October 2012.

[1] The understanding of freshwater flux variability is both scientifically and socially important. Local freshwater flux response to a large Atlantic warm pool (AWP) is excessive freshwater or negative Evaporation minus Precipitation (EmP) anomalies, whereas the response is deficient to a small AWP. However, the EmP anomalies in the AWP region are also influenced by the SST anomalies in the tropical eastern Pacific and in the tropical South Atlantic. These remote influences operate through the inter-basin mode represented by the SST gradient between the tropical North Atlantic and eastern Pacific and the Atlantic meridional mode (AMM) defined as the SST gradient between the tropical North and South Atlantic. When either of these two modes is in the negative phase, the EmP and sea surface salinity anomalies in the AWP region can be positive although the AWP is large. This indicates that the remote influences of the inter-basin mode and/or the AMM can overwhelm the local effect and induce an opposite freshwater response. Additionally, although ENSO and the AMM sometimes coincide with AWP variability, an El Niño in the preceding winter or a positive AMM in the spring does not necessarily follow a large AWP in the summer. **Citation:** Zhang, L., and C. Wang (2012), Remote influences on freshwater flux variability in the Atlantic warm pool region, *Geophys. Res. Lett.*, *39*, L19714, doi:10.1029/2012GL053530.

1. Introduction

[2] Freshwater flux of evaporation minus precipitation (EmP) in the Atlantic warm pool (AWP) region (i.e., the Gulf of Mexico, the Caribbean, and the western tropical North Atlantic) plays a significant role in the agriculture, economy, and environment of surrounding continents and islands. If EmP in the AWP is negative, there is less moisture available for transport into the surrounding land areas. EmP variability can also affect North America by inducing diabatic heating and atmospheric circulation change such as the Great Plains low-level jet [e.g., *Hu and Feng*, 2001]. Additionally, EmP in the AWP or the tropical North Atlantic (TNA) can induce ocean salinity variability in the TNA. This salinity anomaly can be advected toward the deep-water formation regions of the North Atlantic, changing density there and thus influencing the Atlantic meridional overturning circulation [e.g., *Thorpe et al.*, 2001; *Krebs and Timmermann*, 2007]. Therefore, understanding the mechanisms for EmP variability in the AWP region has important applications.

[3] The AWP shows a large variation from year to year and can induce local rainfall change, and affect moisture export across Central America and moisture transport to the central United States [*Wang and Enfield*, 2003; *Wang et al.*, 2008b]. Using various data sets, *Wang et al.* [2012] examine the local EmP response to AWP variability. All of the data sets used show a consistent response: A large (small) AWP is associated with a local freshwater gain (loss) to the ocean, less (more) moisture transport across Central America and a local low (high) sea surface salinity (SSS). A moisture budget analysis further demonstrates that the freshwater change is dominated by the atmospheric mean circulation dynamics, while the effect of thermodynamics (i.e., the change due to humidity) is of secondary importance. Further decomposition points out that the contribution of the mean circulation dynamics primarily arises from its divergent part which mainly reflects the wind divergent change in the low level as a result of SST change. In association with a large (small) AWP, warmer (colder) than normal SST over the TNA can induce anomalous convergence (divergence), which favors anomalous ascent (decent) and thus generates more (less) precipitation.

[4] However, one question is not addressed by these previous studies: Will a large AWP inevitably lead to a local freshwater gain? If not, why do some large AWP years correspond to a freshwater loss in the AWP region? The obvious candidate is the remote forcing such as ENSO [e.g., *Enfield and Alfaro*, 1999; *Giannini et al.*, 2000; *Wu and Kirtman*, 2011] and tropical Atlantic SST variability [e.g., *Enfield*, 1996; *Taylor et al.*, 2002; *Spence et al.*, 2004]. The purpose of this paper is to investigate the remote influences on the EmP and SSS anomalies in the AWP region. The data sets used in this study are described in Section 2. Section 3 briefly summarizes the local response of the EmP and SSS anomalies to AWP variability. Section 4 documents the remote forcing of the EmP and SSS anomalies and the associated mechanisms. Section 5 provides a summary and discussion.

2. Data Sets

[5] Two atmospheric reanalysis data sets are used in this study. The first one is the NCEP-NCAR reanalysis field on a 2.5° × 2.5° horizontal grid. The data consist of daily fields from 1948 to 2010. The other data set is the 20th Century

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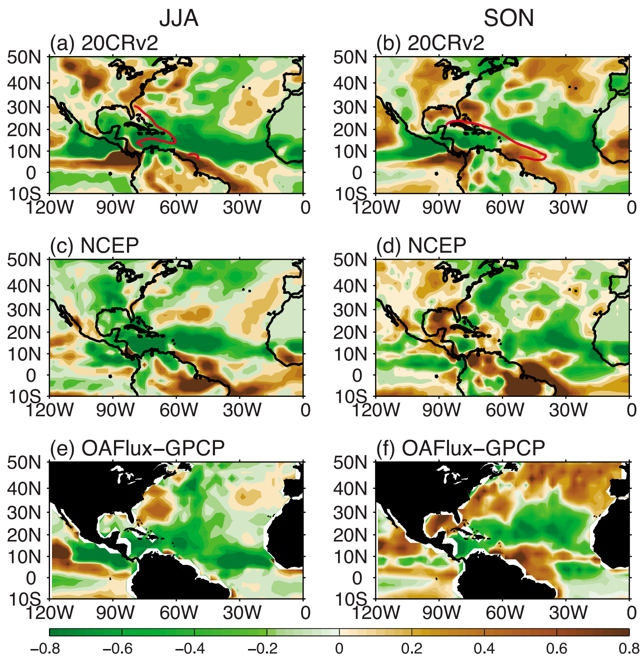

Figure 1. Regression of EmP anomalies onto the AWP index in the (left) summer (JJA) and (right) fall (SON) based on various data sets of (a, b) 20CRv2, (c, d) NCEP, (e, f) OAFlux and GPCP precipitation. Unit is mm/day per 100%. The red contours in Figures 1a and 1b represent the eastern boundary of the climatological AWP (SST warmer than 28.5°C) in JJA and SON, respectively, which comprises the Gulf of Mexico, the Caribbean Sea and the western TNA.

Reanalysis version 2 (20CRv2), which contains the estimate of global tropospheric variability spanning from 1871 to 2010 with a 6-hourly temporal resolution and a spatial resolution of 2° × 2° [*Compo et al.*, 2011]. In addition, the global ocean–atmosphere flux (OAFlux) product [*Yu and Weller*, 2007] is also used to examine the evaporation change associated with the AWP. We also use the Global Precipitation Climatology Project (GPCP), which blends satellite estimates and rain gauge data on a 2.5° × 2.5° grid from January 1979 to 2010 [*Adler et al.*, 2003].

[6] SST is from the NOAA extended reconstructed SST version 3. The monthly objectively analyzed ocean temperature and salinity version 6.7 [*Ishii et al.*, 2006] at 24 levels in the upper 1500 m from 1945–2010 is also used to study the salinity variability associated with the AWP. The analysis is based on the World Ocean Database/WOA05, the global temperature-salinity in the tropical Pacific from IRD/France, and the Centennial in situ Observation Based Estimates (COBE) SST. The *Ishii et al.* analysis also includes the Argo profiling data in the final several years and the XBT depth bias correction.

[7] In this study, no filtering or detrending is applied to the data. This allows us to better quantify the real amount of freshwater flux in the AWP region that is directly related to, for instance, the damage to local economies. Anomalies are calculated by subtracting the climatology (based on the data from 1950–2010) from the monthly fields. We focus on the period of 1950 to 2010 except for GPCP which starts from 1979. If the 20CRv2 and SST data are extended back to the entire 20th century, the conclusions in this paper still hold.

3. Local Response of EmP and SSS Anomalies

[8] The detailed local response of freshwater and SSS to the AWP variations has been shown in *Wang et al.* [2012] who calculated the local moisture budget and the freshwater flux composites for the large and small AWPs. For the sake of the continuity and understanding, here we briefly discuss the relationship of EmP with the AWP. As done in previous studies [e.g., *Wang and Enfield*, 2001; *Wang et al.*, 2006], the AWP index is calculated as the anomalies of the area of SST warmer than 28.5°C divided by the climatological AWP area. Based on this definition, the AWP almost does not exist during the winter and spring. Thus, we only address the EmP anomalies associated with the AWP in the summer (June-August, JJA) and fall (September-November, SON). Regression maps show that during JJA the entire TNA features the negative EmP anomalies (Figure 1, left), indicating that a large (small) AWP is associated with excessive (deficient) freshwater flux to the ocean. A similar response also exists in the fall (Figure 1, right).

[9] A further moisture budget analysis shows that a large portion of the EmP changes in the AWP region can be accounted for by the moisture flux divergence (on monthly to longer timescales), whereas the role of the transient eddies is much smaller (Figure S1 in the auxiliary material).[1] When we decompose the moisture flux divergence anomaly into the mean circulation dynamics contribution (the change due to wind) and the thermodynamics contribution (the change due to humidity), it is shown that the mean circulation dynamics provides a nearly full accounting of the EmP change in the tropics, while the thermodynamics contribution is small (Figure S1). This indicates that the EmP variation is mainly determined by the atmospheric circulation change and the change in humidity plays a small role.

4. Remote Influence on EmP and SSS Anomalies

[10] We first define the AWP index as the anomalies of the area of SST warmer than 28.5°C divided by the climatological AWP area. Large AWP years are identified when the AWP index is larger than 25% at least for three consecutive months from June to November. Based on this definition and the data from 1950–2010, 17 large AWPs are identified (Figure 2). Since the AWP index used here is a total index (undetrended and unfiltered), the large AWPs in Figure 2a include the low-frequency multidecadal variations and the secular warming trend as well as the interannual variability [*Wang et al.*, 2008a]. For most cases, large AWPs occur in the fall (SON) because the AWP normally peaks in September. However, for 5 large AWPs (1958, 1969, 1998, 2005, and 2010), the AWP peaks in the summer (JJA).

[11] Consistent with previous studies [e.g., *Enfield et al.*, 2006; *Wang et al.*, 2006], a large AWP in the summer and fall is sometimes preceded by an El Niño event in the Pacific during the preceding winter (Figures 2a and 2b). However, this is not always true. Of the 17 large AWPs in Figure 2a, 7 large AWP years (1999, 2000, 2001, 2002, 2006, 2008, and 2009) are not preceded by an El Niño event in the preceding

[1] Auxiliary materials are available in the HTML. doi:10.1029/2012GL053530.

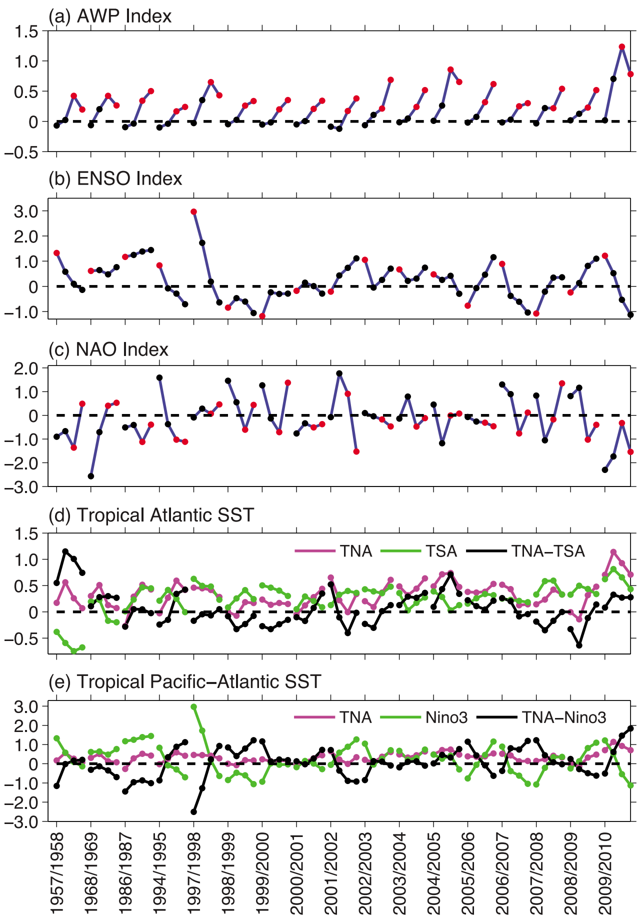

Figure 2. Quarterly (DJF, MAM, JJA, and SON) (a) AWP index (100%), (b) Niño3 SST anomalies (°C), (c) NAO index (normalized), (d) the TNA and TSA SST anomalies (°C) and the index of the tropical AMM (TNA minus TSA), and (e) the index of the inter-basin mode (TNA minus Nino3) for the 17 large AWP years selected during the period 1950–2010. In Figures 2a and 2c, red dot represents the summer (JJA) and fall (SON). In Figure 2b, red dot represents the winter (DJF).

winter. These 7 large AWP years (except 2008) are associated with the negative phase of the NAO in the simultaneous summer or fall (Figure 2c), consistent with *Enfield et al.* (2006) (the NAO index during the summer and fall of 2008 is neutral and positive, respectively). All of the 17 large AWPs in Figure 2a seem to correspond to a warming of the TNA as shown in Figure 2d. This is not surprising because the AWP is in the west of the TNA and they overlap in the portion of the western TNA [*Wang et al.*, 2006]. However, a large AWP does not necessarily follow a positive tropical Atlantic meridional mode (AMM) defined as the SST anomaly difference between the TNA and TSA (tropical South Atlantic). Of the 17 large AWPs, 8 large AWPs (1987, 1998, 1999, 2000, 2002, 2007, 2008, and 2009) are not associated with the positive AMM (Figure 2). These 8 large AWP years correspond to either a nearly neutral AMM (1987, 1998, 2002, 2007, 2008, and 2009) or a negative AMM (1999 and 2000). In summary, although ENSO and the AMM are related to AWP variability, an El Niño in the winter or a positive AMM in the spring does not necessarily follow a large AWP in the subsequent summer and fall. The AWP variations depend on both the remote forcing and local oceanic/atmospheric processes.

[12] The EmP and SSS anomalies in the AWP region for the 17 large AWPs (Figure 2a) are shown in Figure 3 based on various data sets. All data sets show a consistent result that for the majority of large AWP years, the EmP and SSS anomalies are negative during the summer and fall, consistent with the regression analysis in Figure 1. However, the positive EmP anomalies are also seen in both the summer and fall for 5 large AWPs (1987, 2000, 2002, 2008, and 2009) and the SSS anomalies are positive for these 5 large AWPs (Figure 3). Note that the summer of 2001 has a positive EmP anomaly, but the SSS anomalies are nearly neutral. Thus, 2001 is not counted as a large AWP with deficient freshwater. The result in Figure 3a shows that a large AWP does not definitively feature excessive freshwater in the AWP region. *Wang et al.* [2012] have shown the negative EmP anomaly response: A large AWP is associated with a lower (upper) tropospheric anomalous convergence (divergence) and an anomalous upward motion and thus more rainfall and freshwater gain. The question is: What are the mechanisms for the response of the positive EmP anomalies in the AWP region? That is, why do some large AWP years correspond to deficient freshwater?

[13] Previous studies have revealed that remote forcing such as ENSO plays a significant role in rainfall variability

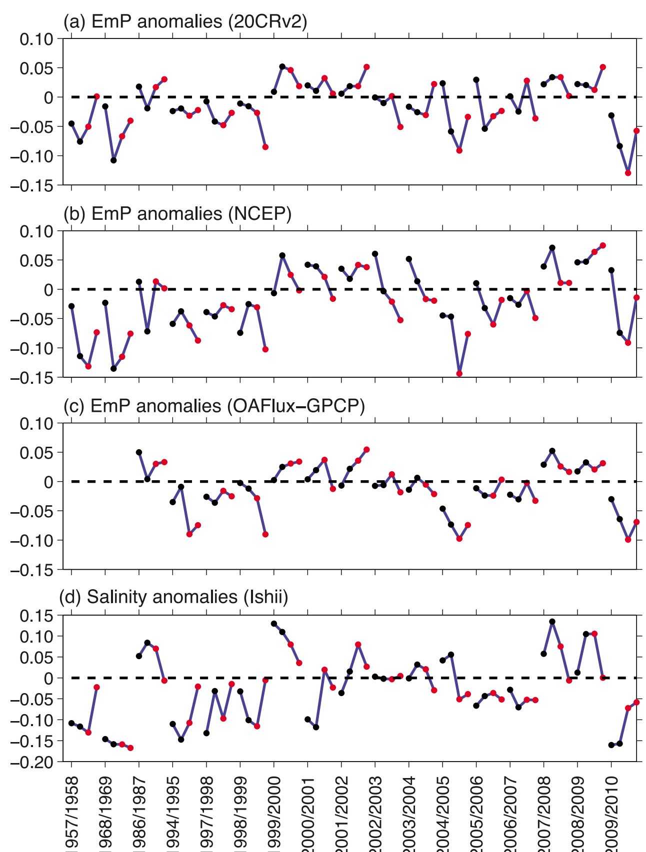

Figure 3. Quarterly (DJF, MAM, JJA, and SON) EmP anomalies (Sv) and sea surface salinity anomalies (psu) in the AWP region (5°N–30°N from the American coast to 40°W) based on various data sets. Red dot represents the summer (JJA) and fall (SON).

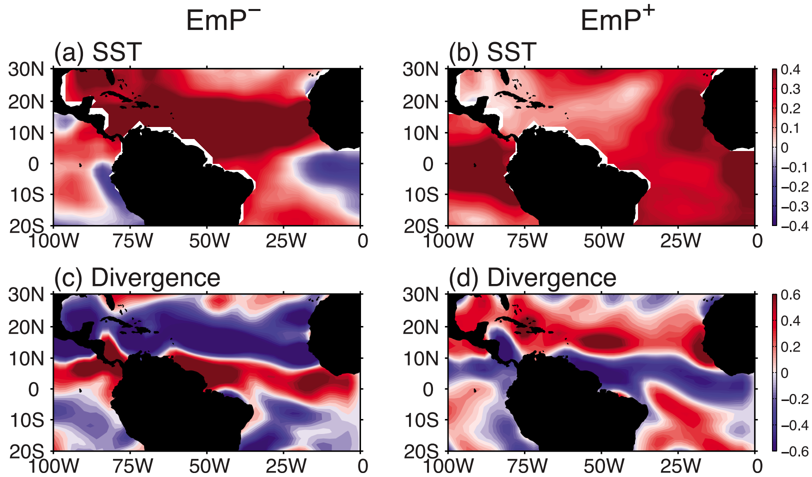

Figure 4. Composites of (a and b) the SST anomalies (°C) and (c and d) the 925-hPa wind divergence anomalies (10^{-6} s^{-1}) for 5 large AWPs with the (left) negative (1958, 1969, 1999, 2005, and 2010) and (right) positive (1987, 2000, 2002, 2008, and 2009) EmP anomalies.

in the Intra-Americas Sea [e.g., *Enfield and Alfaro*, 1999; *Giannini et al.*, 2000; *Taylor et al.*, 2002]. They argued that the strength of the rainfall response seems to depend on how the SST anomalies in the tropical Atlantic and eastern Pacific combine. The strongest response occurs when the tropical Atlantic is in the configuration of a meridional dipole (antisymmetric across the ITCZ), and the tropical eastern Pacific SST anomalies are of opposite sign to those of the TNA which is called the inter-basin mode. This is because the low level winds tend to converge (diverge) from the TSA to the TNA as the AMM index is positive (negative). Similarly, when SST is cold over the TNA and warm in the tropical eastern Pacific, the surface anomalous atmospheric flow is divergent to the southwest, toward the tropical eastern Pacific, and vice versa. Both the inter-basin mode and the AMM can induce an anomalous convergent (divergent) circulation over the TNA, which in turn is in favor of generating a positive (negative) rainfall anomaly. Additionally, the TNA SST is influenced by the NAO through the trade wind variations associated with the North Atlantic subtropical high and thus rainfall in the AWP region [*Enfield et al.*, 2006]. Figures 2 and 3 show these remote influences on the EmP anomalies in the AWP region.

[14] Among the 5 large AWPs with the positive EmP anomalies, 3 years (1987, 2002, and 2009) are associated with negative values of the inter-basin mode index and a small AMM index. The other 2 years (2000 and 2008) are with negative values of the AMM index and a small inter-basin mode index. This indicates that for the 5 large AWPs with the positive EmP anomalies, either the inter-basin mode or the AMM can contribute to the deficient freshwater change in the AWP region. It can thus be concluded that the effects of remote forcing on the EmP anomalies are larger than the local AWP effect for the 5 large AWP years of 1987, 2000, 2002, 2008 and 2009.

[15] To further demonstrate the mechanisms of local and remote response, we compute composites of the SST and wind divergence anomalies for the 5 large AWPs with the negative (1958, 1969, 1999, 2005, and 2010) and positive (1987, 2000, 2002, 2008, and 2009) EmP anomalies in the AWP region (Figure 4). For the case of the negative EmP anomalies, the SST anomalies in the TNA are much larger than those in the TSA and tropical eastern Pacific (Figure 4a), which is in favor of generating an anomalous low-level convergent circulation in the TNA (Figure 4c) and therefore the local freshwater is excessive to the ocean. In contrast, for the case of the positive EmP anomalies, the SST anomalies in the TSA and tropical eastern Pacific are larger (Figure 4b), suggesting that the AMM and the inter-basin mode are important (the local effect on EmP is smaller because the AWP SST anomalies are only weakly positive). These results indicate that the local response of EmP in the AWP region is to increase (decrease) freshwater when the AWP is large (small). However, remote forcing such as the inter-basin mode and the AMM can overwhelm the local effect and thus the freshwater is decreased (increased) although the AWP is large (small).

5. Summary and Discussion

[16] The freshwater and SSS variations in the AWP region are dependent on the local and remote responses. Locally, large (small) AWPs are associated with the warm (cold) SST anomalies in the TNA which induce an anomalous low-level moisture convergence (divergence) and an anomalous ascent (decent) motion. This situation generates more (less) precipitation and thus leads to the negative (positive) EmP and SSS anomalies. However, the EmP anomalies in the AWP are also influenced remotely by the SST anomalies in the tropical eastern Pacific and in the TSA. These remote influences operate through the inter-basin mode and the AMM. The negative phase of either mode can produce an atmospheric low-level anomalous divergence away from the AWP region, which in turn induces positive EmP anomalies. We found that for most of 17 large AWPs, the local response can explain the negative EmP and SSS anomalies in the AWP. Of the 17 large AWPs, 5 large AWPs are associated with the positive EmP and SSS anomalies in the AWP region. This is because the warm SST anomalies in the tropical eastern Pacific and/or TSA can remotely force an anomalous divergence in the AWP region and then decrease rainfall there. That is, the effect of the remote forcing (either the inter-basin

mode or the AMM) is larger than the local effect during these 5 large AWP years.

[17] Both ENSO and the NAO can influence the AWP variations [*Enfield et al.*, 2006; *Wang et al.*, 2006]. Of the 17 large AWPs, 10 large AWPs in the summer and fall are preceded by an El Niño event in the Pacific during the preceding winter, whereas 7 large AWPs are not preceded by an El Niño event in the preceding winter. These 7 large AWP years except 2008 are associated with the negative phase of the NAO in the simultaneous summer or fall. As in the case of ENSO, a positive AMM does not necessarily follow a large AWP either. Of the 17 large AWPs, about half of large AWPs are not associated with the positive AMM. It seems that a combination of ENSO and the NAO can explain the AWP variations. Both ENSO and the NAO can influence the SST anomalies in the TNA which indirectly affect the freshwater response in the AWP region. Numerical model experiments are needed to investigate the local and remote freshwater response.

[18] **Acknowledgments.** We thank Dave Enfield and Marlos Goes for serving as AOML's internal reviewers. This work was supported by grants from NOAA Climate Program Office and the base funding of NOAA AOML.

[19] The Editor thanks one anonymous reviewer for assisting in the evaluation of this paper.

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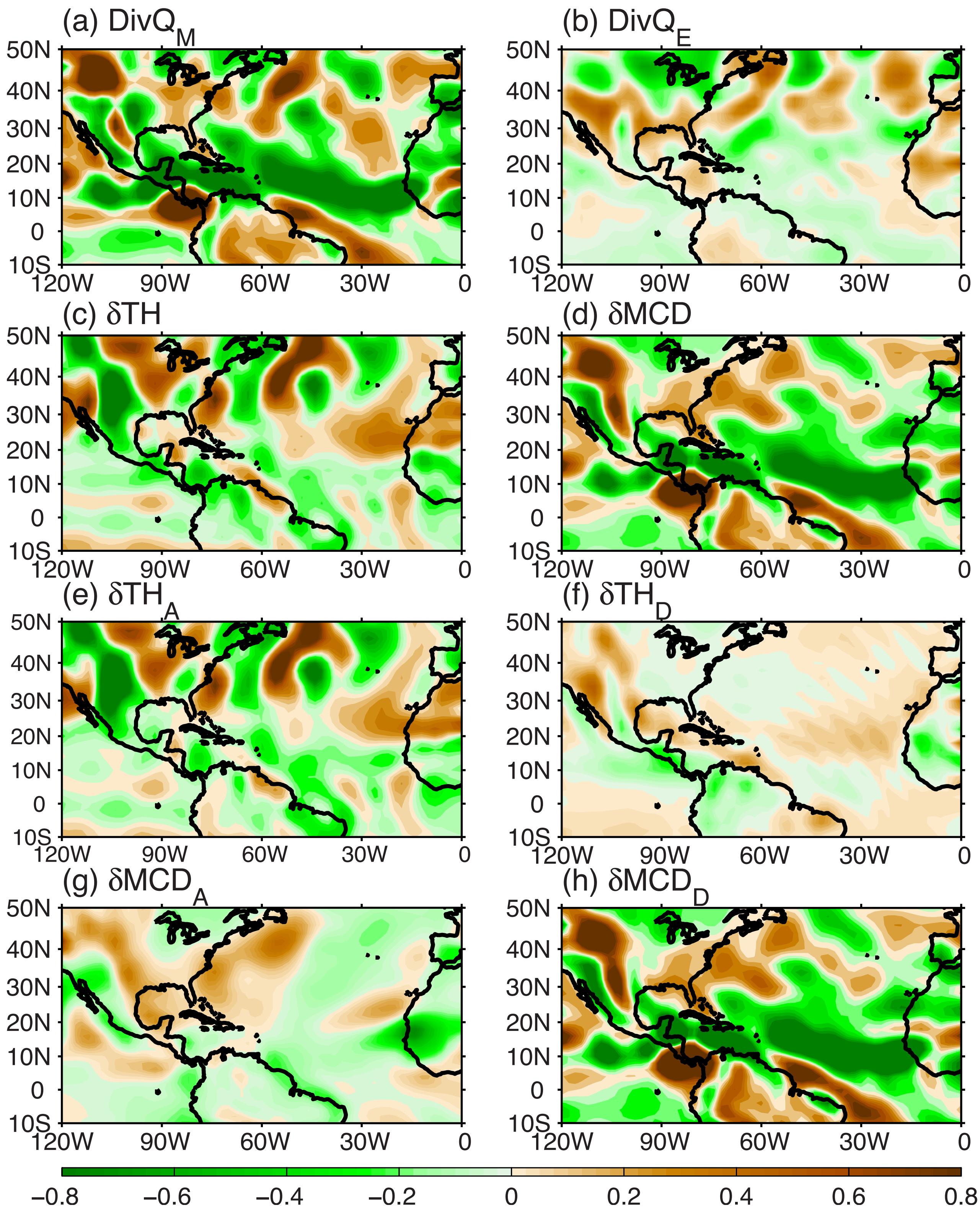

Figure S1. Regression of moisture flux divergence with contributions (a) from monthly to longer time scales (DivQM), (b) from the transient eddies (DivQE), (c) thermodynamics contribution (δTH) by humidity change, and (d) mean circulation dynamics contribution (δMCD) by wind change. δTH is further decomposed into the terms due to (e) the wind advection of humidity (δTHA) and (f) the wind divergence (δTHD). δMCD is decomposed into the terms due to (g) the advection of wind change (δMCDA) and (h) the wind divergence change (δMCDD). The calculations are based on summer (JJA) AWP index and the 20CRv2 data set.
Unit is mm/day per 100%.