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Variability of the Caribbean Low-Level Jet and its relations to climate

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Abstract A maximum of easterly zonal wind at 925 hPa in the Caribbean region is called the Caribbean Low-Level Jet (CLLJ). Observations show that the easterly CLLJ varies semi-annually, with two maxima in the summer and winter and two minima in the fall and spring. Associated with the summertime strong CLLJ are a maximum of sea level pressure (SLP), a relative minimum of rainfall (the mid-summer drought), and a minimum of tropical cyclogenesis in July in the Caribbean Sea. It is found that both the meridional gradients of sea surface temperature (SST) and SLP show a semi-annual feature, consistent with the semi-annual variation of the CLLJ. The CLLJ anomalies vary with the Caribbean SLP anomalies that are connected to the variation of the North Atlantic Subtropical High (NASH). In association with the cold (warm) Caribbean SST anomalies, the atmosphere shows the high (low) SLP anomalies near the Caribbean region that are consistent with the anomalously strong (weak) easterly CLLJ. The CLLJ is also remotely related to the SST anomalies in the Pacific and Atlantic, reflecting that these SST variations affect the NASH. During the winter, warm (cold) SST anomalies in the tropical Pacific correspond to a weak (strong) easterly CLLJ. However, this relationship is reversed during the summer. This is because the effects of ENSO on the NASH are opposite during the winter and summer. The CLLJ varies in phase with the North Atlantic Oscillation (NAO) since a strong (weak) NASH is associated with a strengthening (weakening) of both the CLLJ and the NAO. The CLLJ is positively correlated with the 925-hPa meridional wind anomalies from the ocean to the United States via the Gulf of Mexico. Thus, the CLLJ and the meridional wind carry moisture from the ocean to the central United States, usually resulting in an opposite (or dipole) rainfall pattern in the tropical North Atlantic Ocean and Atlantic warm pool versus the central United States.

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1 Introduction

The Caribbean Sea is bounded to the south by South America, to the west by Central America, and the north by the Greater Antilles (Cuba, Haiti, Dominican Republic, and Puerto Rico). It is connected to the Gulf Mexico by the Yucatan Channel at its northwest side and is linked with the tropical North Atlantic Ocean in the east. As part of the Western Hemisphere warm pool (Wang and Enfield 2001, 2003), the Caribbean Sea features a body of very warm water (warmer than 28.5°C) during the summer and fall (season always refers to the boreal season). Variability of the warm pool can affect the summer climate of the Western Hemisphere (Wang et al. 2006, 2007). Additionally, tropical cyclones can be formed in the Caribbean Sea or be intensified when they pass over Caribbean warm water (e.g., Shay et al. 2000). Thus, the Caribbean Sea is an important region for both weather and climate.

A maximum of easterly zonal wind (larger than 13 m/s) is observed in the lower troposphere of the Caribbean (about 925 hPa) during the summer, called the Caribbean Low-Level Jet (CLLJ) (Amador 1998; Amador and Magana 1999; Poveda and Mesa 1999; Mo et al. 2005; Poveda et al. 2006; also see Fig. 1). [Note that the CLLJ is referred to as ''San Andres Low-Level Jet'' in Poveda and Mesa

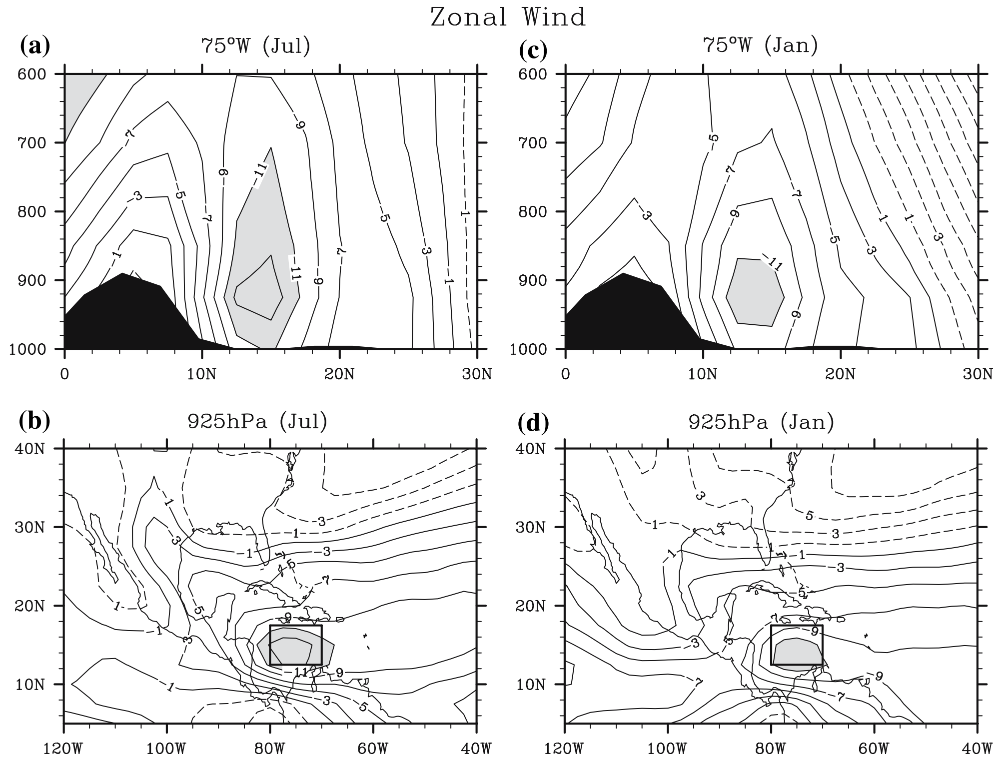

Fig. 1 The CLLJ. Shown are **a** meridional–vertical section of zonal wind at 75°W in July, **b** zonal wind at 925 hPa in July, **c** meridional–vertical section of zonal wind at 75°W in January, and **d** zonal wind at 925 hPa in January. The solid (*dash*) contours represent easterly (westerly) winds, with a contour interval of 2 m/s. The easterly winds larger than 11 m/s are *shaded*. The *box* in (**b**) and (**d**) delineates the area to calculate the CLLJ index. The unit on the vertical axis in (**a**) and (**c**) is *hPa*

(1999).] Recently, Wang and Lee (2007) and Wang et al. (2007) use the National Centers for Atmospheric Research (NCAR) atmospheric general circulation model to simulate and examine the influence of the Atlantic warm pool on the seasonal variations of the CLLJ and its moisture transport. Despite these studies, our knowledge and understanding of the CLLJ are very poor. The present paper provides an observational study of CLLJ's variability on seasonal, interannual, and longer timescales, most of which have not been documented in the literature. Since the Caribbean Sea and the Gulf Mexico serve as a source of atmospheric moisture for rainfall over the Americas (e.g., Rasmusson 1967; Brubaker et al. 2001; Bosilovich and Schubert 2002; Mestas-Nunez et al. 2005), it is expected that the CLLJ is a potential carrier of the exported moisture. Thus, it is not surprising that the CLLJ is related to the climate in the Caribbean region and in other (remote) areas such as North America, Central America, and the eastern North Pacific. However, the influence of the CLLJ on the climate has not been documented and investigated previously. This paper reports the relationship of the CLLJ with climate of the Western Hemisphere.

The present paper analyzes some of available data and has two major purposes. First, it shows in detail the seasonal and anomalous variations of the CLLJ, and discusses CLLJ's variability from the perspective of ocean–atmosphere interaction. Second, the paper shows and documents the CLLJ's relationships with the climate of the northern Western Hemisphere. The paper is organized as follows. Section 2 describes the data sets and methods that are used in this study. Section 3 shows the seasonal variation of the CLLJ. Section 4 documents variability of the CLLJ on interannual and longer timescales and shows CLLJ's relations to the climate. Finally, Sect. 5 provides a summary.

2 Data sets and methods

Many data sets are used in this study. The first one is the National Centers for Environmental Prediction–National Centers for Atmospheric Research (NCEP–NCAR) reanalysis field on a 2.5° latitude by 2.5° longitude grid (Kalnay et al. 1996). Variables that we analyze in this study include monthly sea level pressure (SLP) and horizontal wind velocity at 925, 850, and 200 hPa from January 1950 to August 2006. The second data set is an improved extended reconstructed monthly sea surface temperature (SST) data set on a 2° latitude by 2° longitude grid beginning January 1854 (Smith and Reynolds 2004), but here we only analyze monthly SST from January 1950 to August 2006 for consistency with the data record length of the NCEP–NCAR reanalysis. Another data set is monthly precipitation product of the Global Precipitation Climatology Project (GPCP) (Adler et al. 2003) that is similar to the CPC (Climate Prediction Center) Merged Analysis of Precipitation (CMAP) (Xie and Arkin 1997). The GPCP data set blends satellite estimates and rain gauge data on a 2.5° latitude by 2.5° longitude grid from January 1979 to

August 2006. With the data sets, we first calculate monthly climatologies based on the full record period and then anomalies are obtained by subtracting the monthly climatologies for each data set from the data. Our analyses include the calculations of indices and linear correlations.

3 Seasonal variability

Previous studies have shown that the CLLJ has a maximum of easterly zonal wind in the lower troposphere around 925 hPa (Amador 1998; Amador and Magana 1999; Poveda and Mesa 1999; Mo et al. 2005; Poveda et al. 2006; Wang and Lee 2007; also see Fig. 1a, c). The 925-hPa zonal winds from the NCEP–NCAR reanalysis during July and January are shown in Fig. 1b, d. A longitudinal band of strong easterly zonal wind [induced by the North Atlantic Subtropical High (NASH)] is located in the tropical North Atlantic along 15°N. As the easterly trade wind continues to flow westward to the Caribbean Sea, it intensifies forming the CLLJ with the easterly wind larger than 13 m/s in the summer. Figure 1b, d also show that after the easterly wind passes the Caribbean Sea, the easterly wind contours have two relative maxima: one turning toward the western Gulf of Mexico and the United States, and the other one continuing westward across Central America. The former corresponds to the Great Plains Low-Level Jet (GPLLJ) that transports moisture northward for rainfall over the central United States, whereas the latter carries the moisture westward to Central America and the eastern North Pacific (e.g., Wang et al. 2007).

The CLLJ resides over the Caribbean Sea throughout the year. We use the 925-hPa zonal wind in the region of 12.5°N–17.5°N, 80°W–70°W (represented by the box in Fig. 1b, d) to measure the CLLJ. Figure 2a shows the climatological zonal wind multiplying by -1 in the CLLJ region. The easterly CLLJ displays a semi-annual feature, with two maxima in the summer and winter and two minima in the fall and spring. Statistical significance of differences of the semi-annual variation is assessed by applying a Student's *t*-test. The maxima are different from the minima with the 99% significance level. Two largest values of the CLLJ for the summer and winter peaks occur in July and January, respectively although the January value of the winter peak is not significantly different from that of February. As shown in Fig. 2b, the SLP in the CLLJ region also has double peaks of high pressure in the summer and winter. Associated with the strong easterly CLLJ and high SLP in the summer is a relative minimum of rainfall in July in the region of the CLLJ (Fig. 2c). The minimum rainfall in July is the well-known phenomenon of the mid-summer drought (MSD) which is more obvious in the regions of Central America and South Mexico (e.g., Magaña et al. 1999; Mapes et al. 2005). Unlike the zonal wind, SLP, and rainfall, the SST in the CLLJ region shows a single peak around October (Fig. 2d) that corresponds to the maximum rainfall in October (Fig. 2c). Coincident with the summer strong CLLJ and high SLP, and the MSD of rainfall, 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 in July (Figs. 1, 2a of Inoue et al. 2002).

These observed relationships seem to suggest that the CLLJ is related to the MSD and tropical cyclogenesis in the Caribbean Sea. A possible mechanism for this relationship may be that the easterly CLLJ increases the moisture flux divergence in the Caribbean and thus suppresses the con-

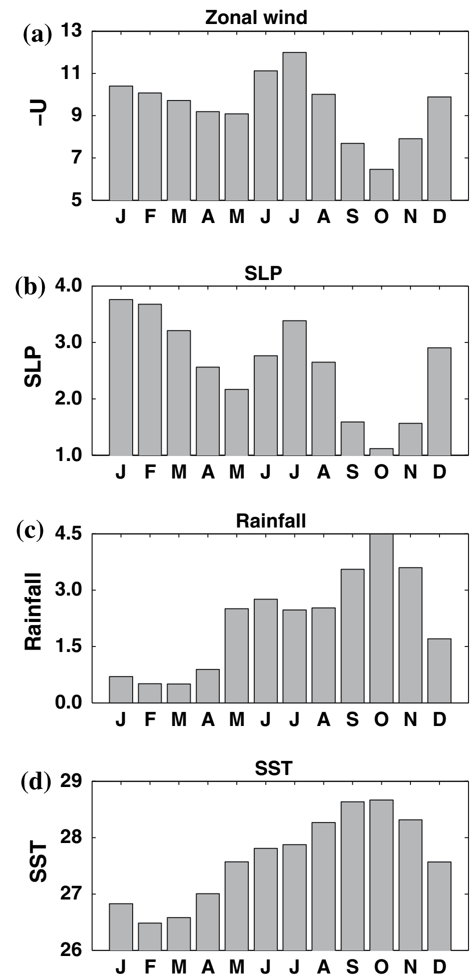

Fig. 2 Seasonal variations of **a** zonal wind (m/s) at 925 hPa, **b** SLP (total SLP – 1,010; hPa), **c** rainfall (mm/day), and **d** SST (°C) near the CLLJ region. Zonal wind and SLP from the NCEP–NCAR reanalysis are calculated over the area of 12.5°N–17.5°N, 80°W–70°W. Rainfall from the GPCP product is over the area of 11.25°N–16.25°N, 81.25°W–71.25°W and SST from NCEP is over the area of 12°N–16°N, 80°W–70°W

vection, decreasing rainfall, and suppressing the formation of tropical cyclones. Another possible explanation may be that high vertical wind shear in July prevents organization of deep convection and thus decreases rainfall and tropical cyclones in July. Inoue et al. (2002) show that the MSD and minimum tropical cyclones in July are associated with a high vertical wind shear in the southwestern Caribbean (10°N–15°N, 80°W–75°W). Using the climatological winds at 850 and 200 hPa and calculating the vertical wind shear is calculated as

$$\left[(U_{200}-U_{850})^2+(V_{200}-V_{850})^2\right]^{1/2},$$

we confirm the conclusion of Inoue et al. (2002). However, the vertical wind shear does not display a peak in July in the CLLJ's region (12.5°N–17.5°N, 80°W–70°W) (Note that the shear decreases from the spring to the fall with the shear in July being much larger than August). This suggests that the mechanism of the vertical wind shear may work in the southwestern Caribbean, but not in the CLLJ's region of the Caribbean.

Why does the CLLJ vary semi-annually? Figure 3 compares the CLLJ's variation with the meridional SLP and SST gradients in the CLLJ region. Both the meridional SST and SLP gradients display two peaks in the summer and winter that correspond to the double peaks of the CLLJ. The observed relationship between the CLLJ and meridional SLP gradient is expected from the geostrophic balance. The result of Fig. 3 is consistent with the simple model of Lindzen and Nigam (1987), which relates atmospheric wind to oceanic SST gradient in the tropics. In the model, they assume that the atmospheric boundary layer temperatures are closely related to oceanic SST by turbulent vertical mixing. These horizontal temperature variations induce horizontal pressure gradients, which directly force the low-level atmospheric flows. The observed relationship between the CLLJ and the meridional SST gradient suggests that the ocean is coupled with the atmosphere in the Caribbean region and that a positive ocean–atmosphere feedback may be involved. On one hand, the meridional SST gradient associated with the SLP gradient produces the easterly CLLJ. On the other hand, the easterly CLLJ results in negative and positive wind stress curls to the north and south of the CLLJ core, respectively (Inoue et al. 2002). The negative wind curl warms the northern Caribbean and the positive curl cools the southern Caribbean through oceanic Ekman dynamics, thus resulting in a further increase of the meridional SST gradient. This positive ocean–atmosphere feedback may help maintain the CLLJ in the Caribbean.

As pointed out by Wang and Lee (2007) and Wang et al. (2007), the SLP in the Caribbean region is related to the

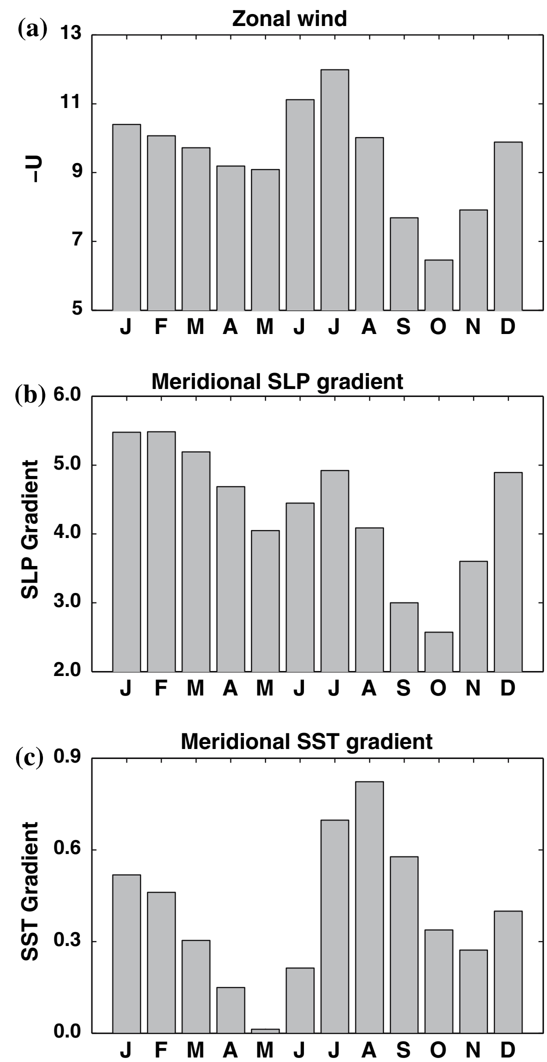

Fig. 3 Seasonal variations of **a** zonal wind (m/s) at 925 hPa, **b** meridional SLP gradient (10^{-6} hPa/m), and **c** meridional SST gradient (10^{-6}°C/m) near the CLLJ region. Zonal wind and meridional SLP gradient are calculated over the area of 12.5°N–17.5°N, 80°W–70°W, and meridional SST gradient is over the area of 12°N–16°N, 80°W–70°W

east–west excursion of the NASH. Figure 4 shows the SLP in the summer (July), fall (October), winter (January), and spring (April). The NASH is strongest in the summer with a cell-type configuration extending toward the Caribbean (Fig. 4a). As the season progresses toward the fall, the NASH weakens and its center moves eastward (Fig. 4b). In the winter, since a continental high develops over North America, the NASH's isobars extend westward and connect with the North American high (Fig. 4c). As the North American monsoon starts to develop in the spring, the NASH's isobars retreat toward the east (Fig. 4d). Thus, the yearly movement and development of the NASH result in a semi-annual feature of the SLP in the region of the Caribbean (Fig. 2b).

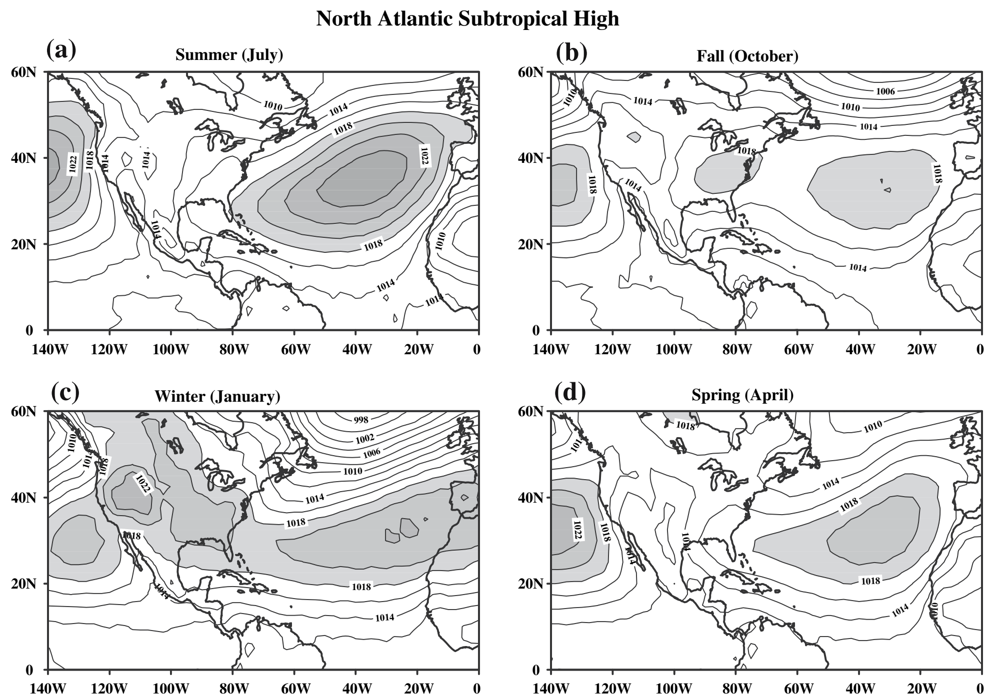

Fig. 4 Sea level pressure (hPa) in the **a** summer (July), **b** fall (October), **c** winter (January), and **d** spring (April) showing the seasonal variations of the North Atlantic Subtropical High (NASH). Sea level pressure larger than 1,018 hPa is *shaded*

4 Interannual and longer timescale variability

In this section, we first define an index for measuring the anomalous CLLJ and show CLLJ's anomalous variability. We then examine CLLJ's relations to the climate in the northern Western Hemisphere.

4.1 CLLJ Index

The 925-hPa zonal wind anomalies in the region of 12.5°N–17.5°N, 80°W–70°W are used to measure CLLJ's anomalous variability, consistent with CLLJ's index for seasonal variability in Sect. 3. For the sake of a convenient explanation of its relationship with climate variability, the CLLJ index is defined by taking the negative of the 925-hPa zonal wind anomalies in the region of 12.5°N–17.5°N, 80°W–70°W (Fig. 5). Since the lower tropospheric winds in the CLLJ region are easterly, the definition indicates that when the index is positive (negative), the CLLJ is anomalously strong (weak). We perform a spectral analysis for the CLLJ index. The CLLJ's autospectrum shows three significant peaks around the frequency bands of about 0.0978, 0.4355, and 0.8 year^{-1} (Fig. 6). These three frequencies correspond to the periods of 10.2, 2.3, and 1.25 years, respectively. The CLLJ anomalies show both the decadal (10.2 year) and interannual (2.3 and 1.25 years) variability. In comparison with the ENSO's interannual variability of 2–7 years, the CLLJ varies with relatively high interannual frequencies. This suggests that the remote ENSO forcing is not the only factor for determining the CLLJ's interannual variability. In fact, the monthly data from January 1950 to August 2006 show that maximum correlations of 0.21, -0.34, and -0.49 occur when the CLLJ zonal wind anomalies lead the Nino3 (5°S–5°N, 150°W–90°W), tropical North Atlantic (6°N–22°N, 60°W–15°W), and Caribbean (12°N–16°N, 80°W–70°W) SST anomalies by 4, 1, and 1 months, respectively. We will show later that the CLLJ correlation with ENSO largely

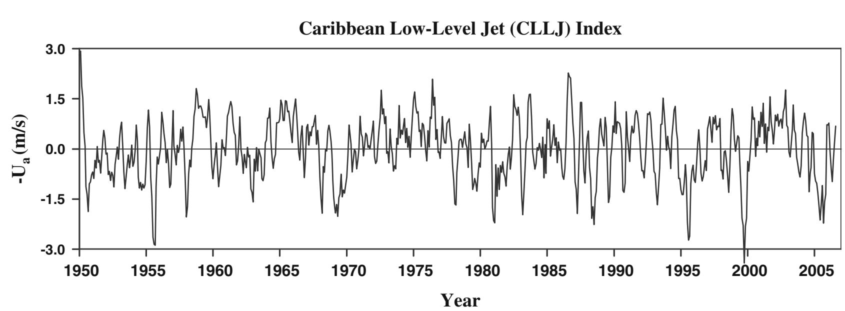

Fig. 5 The CLLJ index. The CLLJ index is calculated by averaging the 925-hPa zonal wind anomalies multiplying by -1 over the region of 12.5°N–17.5°N, 80°W–70°W. Three-month running mean is applied to the index

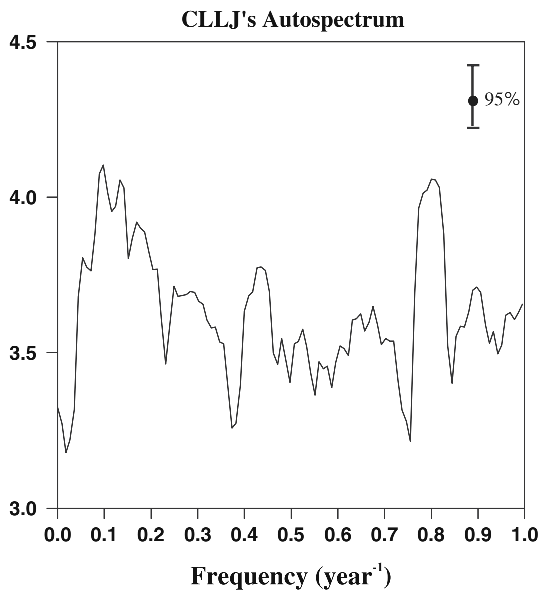

Fig. 6 Logarithm of autospectrum for the CLLJ anomaly index. Vertical line is the 95% confidence interval (Emery and Thomson 1997). All three peaks (around the frequency bands of about 0.0978, 0.4355, and 0.8 year^{-1}) are significant with the 95% confidence level

depends on the season since ENSO's teleconnections are different in the winter and summer.

The standard deviation of the CLLJ anomaly index as a function of month is shown in Fig. 7. The largest variance occurs around September and the secondary large variance is around February. August, September, and October are months when Atlantic hurricanes are busy and active (e.g., Gray 1984). The large variations of the CLLJ around September can change the vertical wind shear between the lower and upper troposphere, which can then affect hurricane activity (Wang and Lee 2007). The February peak of the CLLJ's standard deviation may be more related to ENSO since ENSO is in its mature phase in the wintertime

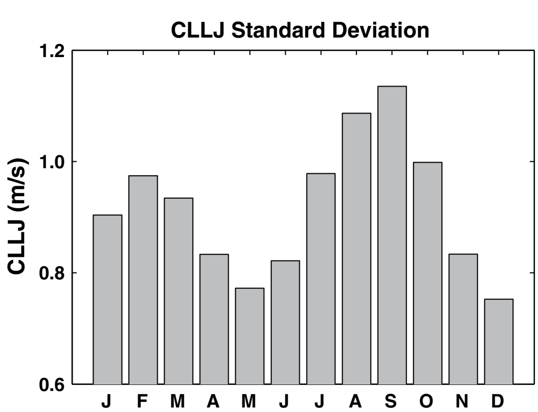

Fig. 7 Standard deviation of the CLLJ anomaly index as a function of month

and the ENSO's influence on the tropical North Atlantic is strong in the subsequent spring (e.g., Enfield and Mayer 1997).

4.2 Relation to SLP anomalies

The atmospheric pressure field and the wind distribution are in approximate geostrophic balance, so we first examine the relationship between SLP anomalies and the CLLJ. As shown in Fig. 7, the CLLJ has two peaks of variability around February and September. In this paper, we will thus focus on variability during both January-February (JF) and August-September (AS) that represent the winter and summer, respectively. Figure 8a, b shows the correlation maps of SLP anomalies with the CLLJ index in JF and AS, respectively. During both the winter and summer, a significant positive correlation is located to north and northeast side of the CLLJ region where the NASH resides. The

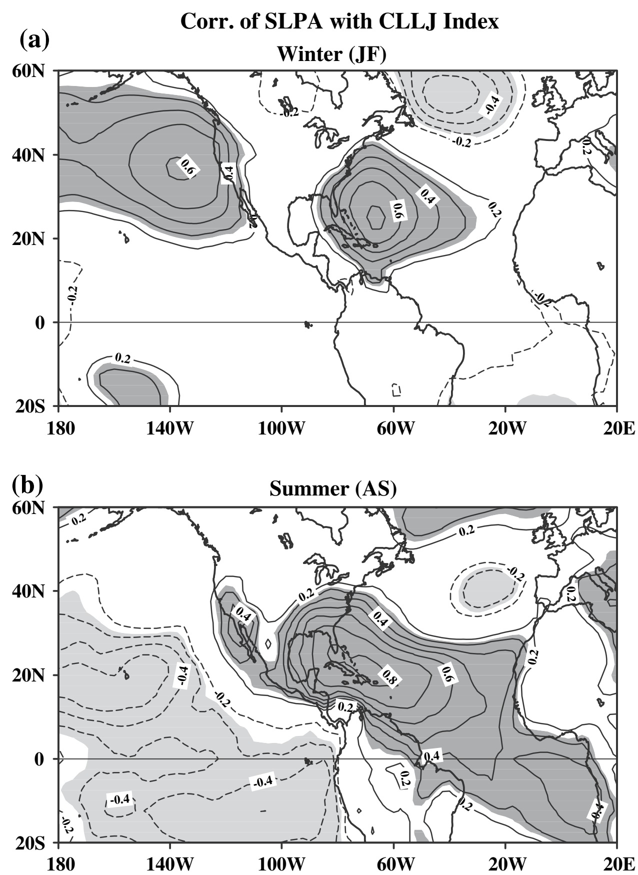

Fig. 8 Correlation maps of SLP anomalies with the CLLJ index during the **a** winter (JF) and *b* summer (AS). The calculations are based on data from 1950 to 2006. The contour interval is 0.1 and the correlations below ±0.2 are not plotted. The 95 and 99% significant levels are 0.26 and 0.34, respectively. The *shadings* represent correlation larger than the 95% significant level (*dark* for positive and *light dark* for negative)

positive SLP correlation indicates that a strengthening (weakening) of the NASH corresponds to a strong (weak) easterly CLLJ. This is consistent with the modeling result of Wang et al. (2007) who use the NCAR atmospheric model to demonstrate that a removal of the Atlantic warm pool strengthens the NASH and then increases the CLLJ's strength.

In the winter, the extratropical North Atlantic shows a negative SLP correlation and the North Pacific has a positive correlation (Fig. 8a). We keep in mind that in the winter the North Atlantic and Pacific are under the influence of the Icelandic low and the Aleutian low, respectively. The North Atlantic Oscillation (NAO), represented by the meridional SLP seesaw between the NASH and the Icelandic low, links the SLP in the subtropical Atlantic with that in the North Atlantic. A strengthening of the NASH, usually associated with strengthening of the easterly CLLJ, thus tends to correspond to a strengthening of the Icelandic low (with negative SLP anomalies). This explains the negative correlation between the CLLJ and the SLP anomalies in the North Atlantic. The positive SLP correlation in the North Pacific shows that the easterly CLLJ is associated with a weakening of the wintertime Aleutian low (with positive SLP anomalies) in the North Pacific.

During the summer (Fig. 8b), an additional positive SLP correlation is centered in the southwest coast of North America–the North American monsoon region. This shows that a strong (weak) summertime easterly CLLJ corresponds to positive (negative) SLP anomalies near the North American monsoon region and thus a weak (strong) summer monsoon. This is again consistent with the modeling result of Wang et al. (2007) who show that the effect of the Atlantic warm pool is to weaken the CLLJ and to decrease SLP northwest of the warm pool. The physical basis for this response is Gill's (1980) theory that shows a Rossby wave (with low SLP) to the northwest of atmospheric heating. Figure 8b also shows a negative SLP correlation in the subtropical North Pacific, indicating that the summertime strength of the North Pacific subtropical high is inversely related to the CLLJ.

4.3 Relation to SST anomalies

The correlation maps of SST anomalies with the CLLJ index in the winter and summer are shown in Fig. 9a, b, respectively. During both the winter and summer, a significant negative SST correlation is located near the Caribbean region. When the SST in the Caribbean is anomalously warm (cold), the easterly CLLJ is anomalously weak (strong). This is consistent with the modeling result of Wang and Lee (2007) and Wang et al. (2007) who show that a removal of the Atlantic warm pool strengthens the NASH and thus increases the easterly CLLJ. The physical mechanism for the SST–SLP–CLLJ relationship may be explained in terms of Gill's (1980) simple theory. For an off-equatorial heating anomaly (warm SST anomalies), Gill's theory predicts an atmospheric response involving low pressure to the northwest of the heating, associated with a Rossby wave. The low pressure then decreases the easterly zonal wind anomalies (the CLLJ) in the heating region. The negative SST correlation in the Caribbean is consistent with Gill's theoretical work.

The difference between the winter and summer Atlantic SST correlation is that in the winter it displays an alternating tripole pattern of zonally oriented negative-positive SST correlation, whereas in the summer it does not (Fig. 9a, b). The wintertime tripole pattern features a negative SST correlation in the regions of the Caribbean to the western tropical North Atlantic and of the North Atlantic, and a positive correlation from the Gulf of Mexico to the southeastern coast of the United States (Fig. 9a). This difference may reflect that the NAO links the Icelandic low and the NASH and that the NAO is strong in the winter. The CLLJ is largely affected by variability of the NASH. When the NASH is strong and extends westward, the easterly CLLJ becomes strong. The strong easterly CLLJ results in a decrease in SST through evaporation and/or ocean dynamics in the CLLJ region and in the western tropical North Atlantic. In the region from the Gulf of Mexico to the southeastern coast of the United States, the CLLJ-related SLP pattern (Fig. 8a) suggests a westerly wind anomaly that can warm SST (Fig. 9a). The negative SST correlation at the high latitudes of around 45°N–50°N may involve the NAO, which represents a meridional SLP seesaw between the Azores high (or the NASH) and the Icelandic low (Hurrel 1995). When the NASH is strengthened (the easterly CLLJ is strong), the Icelandic low is strengthened (Fig. 8a). This corresponds to strong westerly wind in the high latitudes, and thus decreasing SST there (Fig. 9a).

On the Pacific side, Figure 9a, b show different signs of SST correlation in the tropical Pacific: negative in the winter and positive in the summer. This indicates that during the winter, a weak (strong) easterly CLLJ corresponds to warm (cold) SST anomalies in the tropical eastern and central Pacific. During the summer, a strong (weak) easterly CLLJ is associated with warm (cold) SST anomalies in the tropical Pacific. That is, ENSO has a different relationship with the CLLJ in the winter and summer. To investigate why, we focus on the difference of ENSO's teleconnections between the winter and summer. Figure 10 shows the correlation maps of SLP anomalies with the Nino3 SST anomalies in the winter and summer. In the winter, a large negative correlation is located in the Caribbean and in the subtropical western North Atlantic (Fig. 10a). However, in the summer a positive SLP corre-

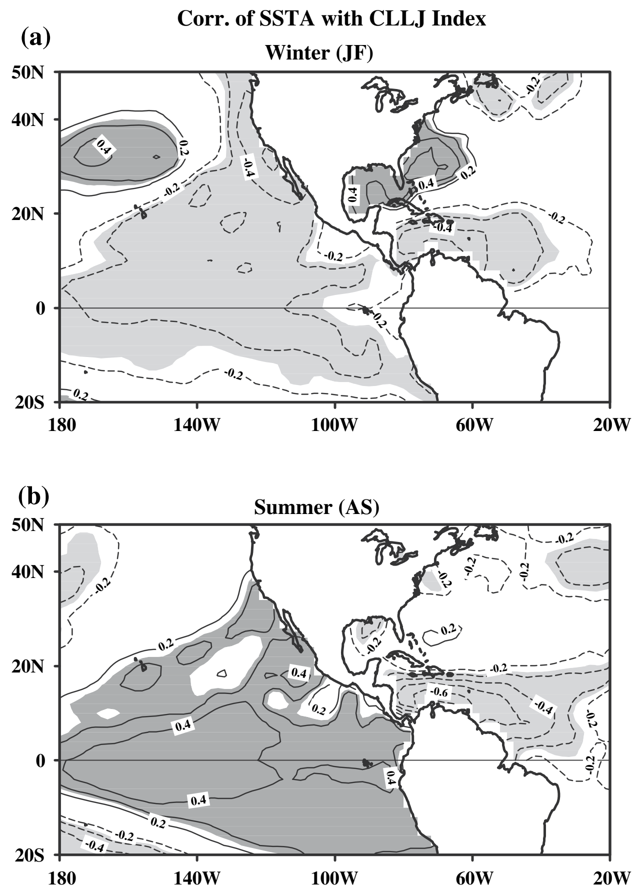

Fig. 9 Correlation maps of SST anomalies with the CLLJ index during the **a** winter (JF) and **b** summer (AS). The calculations are based on data from 1950 to 2006. The contour interval is 0.1 and the correlations below ±0.2 are not plotted. The 95 and 99% significant levels are 0.26 and 0.34, respectively. The *shadings* represent correlation larger than the 95% significant level (*dark* for positive and *light dark* for negative)

lation in the tropical Atlantic extends westward to Central America and then northward to the western central United States, with a significant correlation in the western Gulf of Mexico and the western central United States (Fig. 10b). This means that ENSO's teleconnection is to decrease (increase) the SLP near the Caribbean region during the winter (summer). As shown in Figs. 2 and 3, high (low) SLP in the Caribbean region is associated with large (small) meridional SLP gradient. Thus, the decrease (increase) of the SLP in the winter (summer) can result in the weak (strong) easterly CLLJ and thus a wintertime negative (summertime positive) correlation between the CLLJ index and tropical Pacific SST anomalies.

4.4 Relation to anomalous meridional flows to the United States

The NASH produces the easterly trade winds at its southern flank that carry moisture from the tropical North Atlantic into the Caribbean Sea where the flow intensifies forming the CLLJ. The CLLJ then splits into two branches: one turning northward, and the other one continuing westward across Central America into the eastern North Pacific. The northward one flows into the United States via the Gulf of Mexico. It is thus expected that the meridional flows to the United States fluctuate with the CLLJ. Figure 11 shows the correlation maps of the 925-hPa meridional wind anomalies with the CLLJ index. Both the winter and summer show a significant positive correlation in the Gulf of Mexico and the central/eastern United States. This indicates that when the CLLJ is anomalously strong (weak), the meridional wind anomalies in the Gulf of Mexico and the central/eastern United States are southerly (northerly). If we define the 925-hPa meridional wind anomalies in the region of 25°N–35°N, 100°W–95°W as an index for measuring the GPLLJ, the correlations between the CLLJ and GPLLJ during the winter and summer are 0.40 and 0.54, respectively. A strong (weak) CLLJ corresponds to a strong (weak) GPLLJ. Additionally, Fig. 11 also shows significant correlation patterns in the subtropical North Atlantic and in the North Pacific west coast of North America, reflecting the relationship between the CLLJ and the subtropical highs.

The mechanism of the meridional wind correlations with the CLLJ can be explained in terms of atmospheric response to the oceanic SST forcing. A cold SST in the Caribbean region strengthens the NASH that in turn increases the easterly CLLJ. The pressure response in Fig. 8 shows a high SLP located north and northeast of the Caribbean region. This high SLP pattern produces southerly (northerly) wind anomalies at its western (eastern) side, thus resulting in a positive (negative) meridional wind correlation in the Gulf of Mexico and the central/eastern United States (the subtropical North Atlantic). The negative (positive) meridional wind correlation in the North Pacific in the winter (summer) in Fig. 11 represents northerly (southerly) wind anomalies that are associated with the wintertime Aleutian low (the summertime North Pacific subtropical high).

4.5 Relation to rainfall anomalies

The associations of the CLLJ with rainfall anomalies during the winter and summer are shown in Fig. 12a, b, respectively. The central United States and the Atlantic Ocean show an opposite rainfall correlation pattern in both the winter and summer although significant correlations are located differently during the winter and summer. In the winter, the significant positive correlation is over the southern central United States and the negative correlation is in the subtropical North Atlantic (Fig. 12a). In the summer, the significant positive correlation is located over

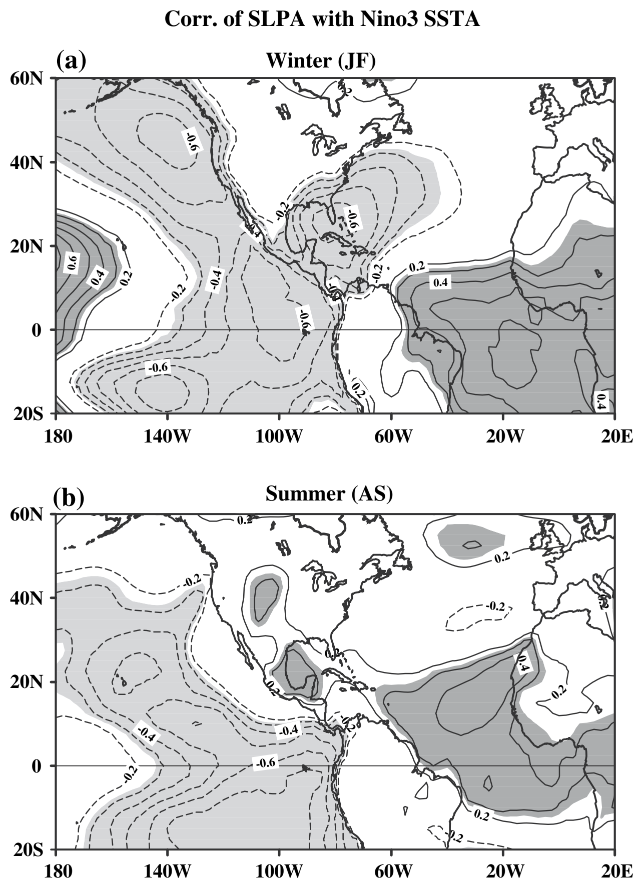

Fig. 10 Correlation maps of SLP anomalies with the Nino3 (5°S–5°N, 150°W–90°W) SST anomalies during the **a** winter (JF) and **b** summer (AS). The calculations are based on data of all winter and summer from 1950 to 2006. The contour interval is 0.1 and the correlations below ±0.2 are not plotted. The 95 and 99% significant levels are 0.26 and 0.34, respectively. The *shadings* represent correlation larger than the 95% significant level (*dark* for positive and *light dark* for negative)

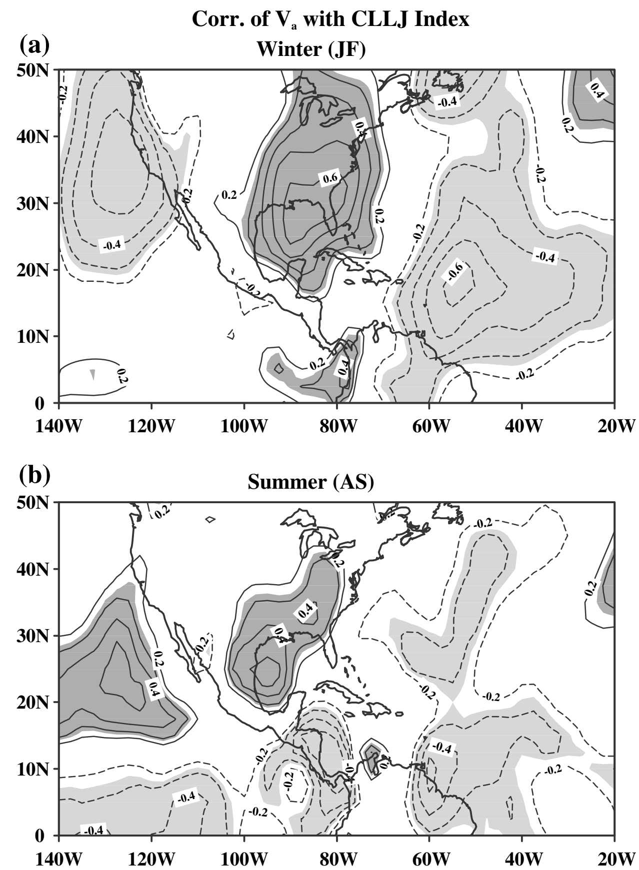

Fig. 11 Correlation maps of 925 hPa meridional wind anomalies with the CLLJ index during the **a** winter (JF) and **b** summer (AS). The calculations are based on data from 1950 to 2006. The contour interval is 0.1 and the correlations below ±0.2 are not plotted. The 95 and 99% significant levels are 0.26 and 0.34, respectively. The *shadings* represent correlation larger than the 95% significant level (*dark* for positive and *light dark* for negative)

the northern central United States, and the negative is in the tropical North Atlantic, the eastern Caribbean Sea, and Central America (Fig. 12b). The opposite rainfall pattern between the ocean and the central United States may be explained from a moisture transport perspective. The ocean is a source of water vapor for rainfall in North, Central, and South America (e.g., Mo and Higgins 1996; Hu and Feng 2001). Our analysis (Fig. 11) shows that the CLLJ is positively correlated with the 925-hPa meridional wind anomalies from the Gulf of Mexico to the United States. This suggests that a strong (weak) easterly CLLJ is associated with southerly (northerly) wind anomalies to the United State that transport more (less) moisture for rainfall over the United States. At the same time, if more (less) moisture is exported from the ocean to the United States, it seems plausible that less (more) moisture would be available for local rainfall over the ocean region. In addition to the enhanced (reduced) moisture transport into the United States, a strong (weak) CLLJ is associated with less (greater) moist static instability over the Caribbean, thus decreasing (increasing) convection and rainfall there and increasing (decreasing) the moisture over the ocean available for export. All of these may explain why the CLLJ is associated with an opposite rainfall pattern between the ocean and the continental United States in Fig. 12.

In the summer, the CLLJ index is positively correlated with rainfall anomalies in the eastern North Pacific in a longitudinal band along 9°N (Fig. 12b). This positive correlation may also be associated with the CLLJ's moisture transport. As the CLLJ passes the Caribbean Sea, one of the CLLJ's branches is to continue westward across Central America into the eastern North Pacific. It is possible that a strong CLLJ enhances the moisture convergence in the eastern North Pacific and thus increases rainfall there. Figure 12 also shows that the eastern subtropical North Pacific displays significant rainfall

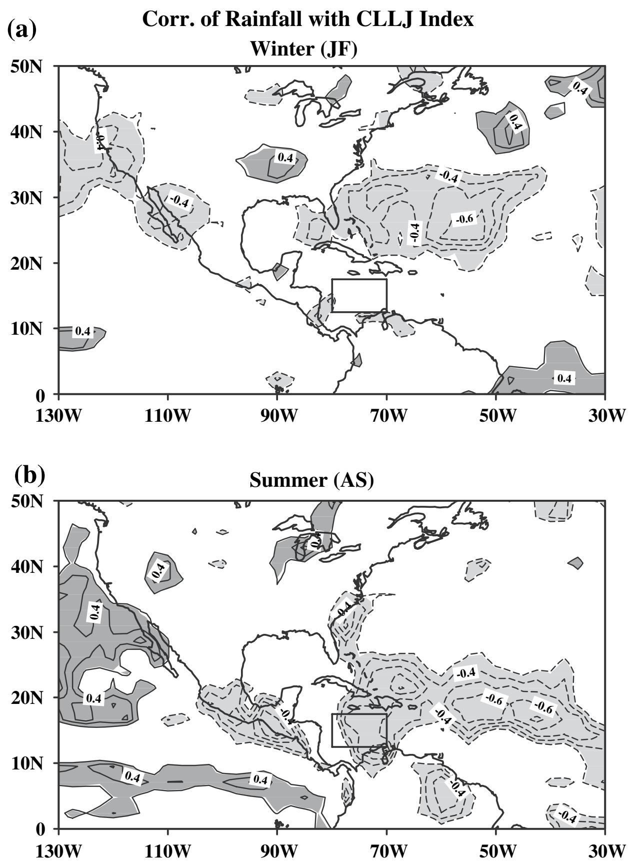

Fig. 12 Correlation maps of rainfall anomalies with the CLLJ index during the **a** winter (JF) and **b** summer (AS). The calculations are based on data from 1979 to 2006. The contour interval is 0.1 and the correlations below ±0.3 are not plotted. The 90 and 95% significant levels are 0.31 and 0.37, respectively. The *shadings* represent correlation larger than 0.31 (*dark* for positive and *light dark* for negative). The *box* delineates the area (12.5°N–17.5°N, 80°W–70°W) to calculate the CLLJ index of the 925-hPa zonal wind anomalies

correlations during both the winter and summer. This is consistent with the distributions of SLP anomalies and 925 hPa meridional wind anomalies in Figs. 8 and 11. In the winter, the North Pacific is under the influence of the Aleutian low. The positive SLP anomalies in the North Pacific (Fig. 8a) means a weakening of the Aleutian low which is associated with northerly wind anomalies (Fig. 11a). The northerly wind anomalies are not favorable for rainfall, resulting in negative rainfall correlation in the eastern subtropical North Pacific and the west coast of the United States (Fig. 12a). In the summer, the North Pacific is controlled by the North Pacific subtropical high. The negative SLP anomalies in the subtropical North Pacific (Fig. 8b) indicate a weakening of the subtropical high which is associated with southerly wind anomalies (Fig. 11b). The southerly wind anomalies carry moisture to the eastern subtropical North Pacific for more rainfall there (Fig. 12b). However, how the CLLJ is related to the Aleutian low and the North Pacific subtropical high is unknown.

As shown in Wang et al. (2007), one branch of the CLLJ turns northward for merging with the GPLLJ after the CLLJ passes the Caribbean. It is thus interesting to see rainfall correlation with the GPLLJ. The correlation maps of rainfall anomalies with the GPLLJ index during the winter and summer are shown in Fig. 13a and b, respectively. The opposite rainfall correlation pattern (or the dipole pattern) in the central United States versus the tropical North Atlantic Ocean and Atlantic warm pool region is more obvious in Fig. 13. This reflects that the GPLLJ is positively correlated with the CLLJ (0.40 and 0.54 in the winter and summer) and that the GPLLJ is directly involved the northward moisture transport to the central United States.

5 Summary

Using observational data, the present paper shows variability of the CLLJ and its relationships with the climate of the northern Western Hemisphere. The key for CLLJ's variability and influences on the climate is through the NASH. NASH's strength and its east–west excursion change the SLP gradient in the Caribbean and thus the CLLJ that are associated with variations of SST and rainfall in the Caribbean region. The NASH also serves as a bridge to link the CLLJ with other important climate phenomena such as ENSO, the NAO, and tropical Atlantic variability. In this paper, we examine the CLLJ's seasonal and anomalous variability. On seasonal timescale, the major findings are:

- The easterly CLLJ is observed to show a semi-annual variation, with two maxima in the summer and winter and two minima in the fall and spring.
- The summertime strong easterly CLLJ is associated with a maximum of SLP, the MSD of rainfall, and a minimum of tropical cyclones in the Caribbean region. A possible mechanism for this relationship may be that the easterly CLLJ increases the moisture flux divergence in the Caribbean and thus suppresses the convection, decreasing rainfall, and suppressing the formation of tropical cyclones.
- The semi-annual strengthening of the easterly CLLJ results from the semi-annual variation of the meridional SST and SLP gradients. A positive ocean–atmosphere feedback may be operating for maintaining the easterly CLLJ. A meridional SST gradient in the Caribbean induces a meridional SLP gradient that produces the easterly CLLJ. The easterly CLLJ in turn results in negative and positive wind stress curls to the north and

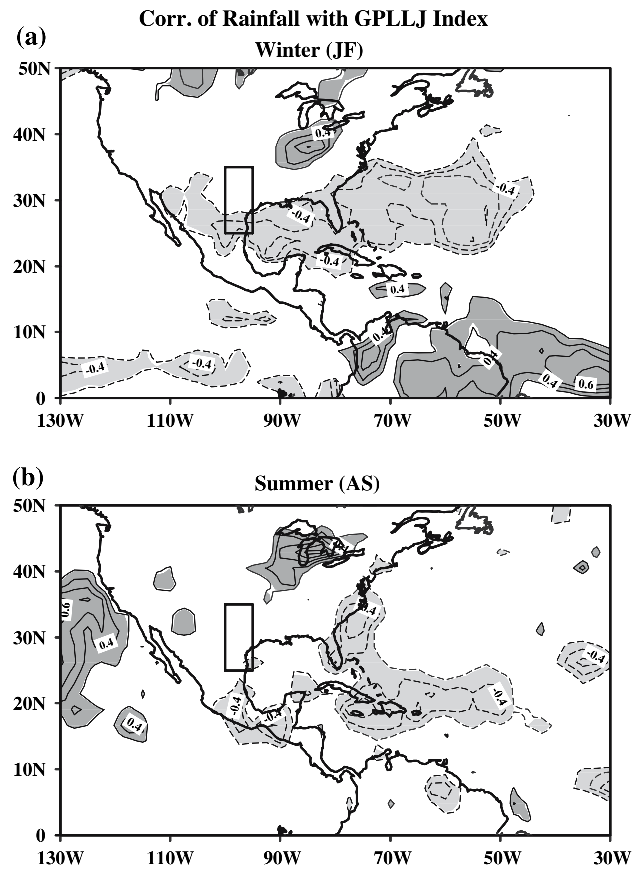

Fig. 13 Correlation maps of rainfall anomalies with the Great Plains Low-Level Jet (GPLLJ) index during the **a** winter (JF) and **b** summer (AS). The calculations are based on data from 1979 to 2006. The contour interval is 0.1 and the correlations below ±0.3 are not plotted. The 90 and 95% significant levels are 0.31 and 0.37, respectively. The *shadings* represent correlation larger than 0.31 (*dark* for positive and *light dark* for negative). The *box* delineates the area (25°N–35°N, 100°W–95°W) to calculate the GPLLJ index of the 925-hPa meridional wind anomalies

south of the CLLJ core, respectively. The negative wind curl warms the northern Caribbean and the positive curl cools the southern Caribbean through oceanic Ekman dynamics, thus resulting in a further increase of the meridional SST gradient.

The main findings in the CLLJ's anomalous variability and its relations to the climate are as follows:

- The CLLJ index of zonal wind anomalies at 925 hPa shows peaks of variability on timescales of 1.25, 2.3, and 10.2 years.
- Two maxima of the standard deviation of CLLJ anomalies are found around September and February. The largest variance near September coincides with the busy and active months of Atlantic hurricanes, suggesting that the large variation of the easterly CLLJ changes the vertical wind shear between the lower and upper troposphere that then affects hurricane activity.
- The CLLJ anomalies are inversely related to the Caribbean SST anomalies. Based on Gill's theory, warm (cold) SST anomalies in the Caribbean are associated with low (high) SLP anomalies that weaken (strengthen) the easterly CLLJ.
- The CLLJ varies in phase with the NAO. This reflects the fact that both the CLLJ and NAO are related to the NASH. A strong (weak) NASH is associated with strengthening (weakening) of the easterly CLLJ and also corresponds to the positive (negative) phase of the NAO.
- The CLLJ has an opposite relationship with ENSO in the winter and summer. During the winter a weak (strong) easterly CLLJ corresponds to warm (cold) SST anomalies in the tropical Pacific, whereas during the summer a strong (weak) easterly CLLJ is associated with warm (cold) SST anomalies in the tropical Pacific. This is because ENSO's teleconnections are different: ENSO induces negative and positive SLP anomalies in the subtropical North Atlantic during in the winter and summer, respectively.
- The CLLJ is positively correlated with the surface meridional wind anomalies over the Gulf of Mexico and the central/eastern United States or the GPLLJ index. When the CLLJ is anomalously strong (weak), the meridional wind anomalies in the Gulf of Mexico and the central/eastern United States are southerly (northerly).
- A strong (weak) easterly CLLJ is associated with southerly (northerly) wind anomalies to the United State that transport more (less) moisture for rainfall over the United States. At the same time, if more (less) moisture is exported from the ocean to the United States, it seems plausible that less (more) moisture would be available for local rainfall over the ocean region. Thus, the CLLJ and the GPLLJ are associated with an opposite rainfall pattern (i.e., a dipole rainfall pattern) in the tropical North Atlantic Ocean and Atlantic warm pool region versus the central United States.

All results and conclusions in this paper are inferred from the diagnostic analyses of observational data and reanalysis field. Although observations are invaluable for showing what happens in the real world, they are limited in determining causality in light of competing influences. Numerical modeling (from both uncoupled and coupled models) studies are needed to diagnose cause and effect for the results presented in this paper.

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