A Northern Hemispheric Wave Train Associated with Interannual Variations in the Bermuda High during Boreal Summer

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ABSTRACT: The processes that lead to the spatial and temporal evolution of the Bermuda high (BH) during July and August (JA) are investigated on the basis of linear regression analysis. The analysis is based on a Bermuda high index (BHI): the difference in standardized, deseasonalized, and detrended sea level pressure (SLP) between northeast of Bermuda (40°N, 60°W) and New Orleans (30°N, 90°W). Negative values of BHI indicate a westward expansion of the Bermuda high relative to its climatological-mean location and reduced precipitation in the southeastern United States (SEUS), whereas positive values correspond to BH contraction and enhanced precipitation in the SEUS. Linear regression of the 200-hPa geopotential height based on the BHI reveals the existence of a Rossby wave train that extends zonally from the eastern North Pacific to the eastern North Atlantic. The troughs and ridges associated with this wave train are spatially collocated with the climatological-mean jet stream, indicating that the jet serves as their waveguide. Anomalous troughing in the SEUS associated with this wave train is linked to the contraction of the Bermuda high during JA. The enhanced precipitation is associated with anomalous ascent to the east and south of this trough where anomalous warm advection is observed. Based on these results, it is hypothesized that this Rossby wave train may partially explain the occurrence of suppressed precipitation tied to midsummer drought in the SEUS during July and August. It is found that the BHI has trended from negative to positive in recent decades, suggesting that it may be influenced by low-frequency variability.

KEYWORDS: Rossby waves; Teleconnections; Interannual variability; Northern Hemisphere; Precipitation

1. Introduction

The North Atlantic subtropical high (NASH; also known as the Bermuda–Azores high) is a region of climatological-mean high pressure centered over the North Atlantic near 32.5°N, 45°W. When the NASH is located on the western Atlantic basin during boreal summer, its lower-tropospheric center is located near Bermuda and it is often regarded as the Bermuda high (BH) (Zishka and Smith 1980; Davis et al. 1997; Li et al. 2011). We will use the term “Bermuda high” from here on to denote the western phase of NASH. The BH plays an important role in the climate of southeastern North America. The shape, location, and intensity of the BH have been associated with variations in precipitation over the southeastern United States (SEUS) (Stahle and Cleaveland 1992; Henderson and Vega 1996; Keim 1997; Katz et al. 2003; Li et al. 2011; Ortegren et al. 2011) and over the U.S. Great Plains (Zhu and Liang 2013). The BH also plays a central role in the occurrence and strength of the midsummer drought (Hastenrath 1976, 1978), a local minimum in precipitation that occurs in July and August (JA) over the SEUS, the Greater Antilles, and part of Central

America (Henderson and Vega 1996). An intensified and westward expanded BH brings higher sea level pressure and incites stronger trade winds and lower sea surface temperature (Giannini et al. 2000), along with stronger subsidence (Knaff 1997), resulting in a decrease in precipitation over the Caribbean region (Small et al. 2007). Gamble et al. (2008) verified these conclusions but pointed out differences in the spatial influence of Bermuda high over the Caribbean region. Instead of initiating the midsummer drought over the whole Caribbean region, the midsummer drought initiates over the eastern Caribbean and proceeds westward, having a different duration over different locations. Kelly and Mapes (2011) argued that a teleconnection pattern that arises from the South Asian monsoon strengthened the BH, causing the midsummer drought in the western North Atlantic. Li et al. (2012) analyzed the relationship between moisture transport and vertical motion associated with displacements of the BH over the SEUS. They found that a southwest displacement of the BH causes increased moisture transport and rainfall whereas water vapor and precipitation decrease when the BH shifts toward the northwest.

The origin of the BH was also investigated in several studies. Using both observational data and a linear quasigeostrophic model, Chen et al. (2001) revealed that Asian monsoonal heating is the main energy source of the subtropical high. The same conclusions were drawn by Rodwell and Hoskins (2001) through a primitive equation model. Their hypotheses were also supported by Seager et al. (2003), who concluded that summer monsoonal heating over land brings about the development of the subtropical anticyclone. However, these studies found that the subtropical high forced by the Asian monsoon

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was weak in their models. Miyasaka and Nakamura (2005) proposed that, in addition to the teleconnections with the Asian summer monsoon, localized zonal contrasts in near-surface heating between the Atlantic Ocean and cooling near the continental west coast of Africa is the primary driver for the BH.

In this study, we will investigate the mechanisms that modulate the intensity and position of the Bermuda high and the mechanisms in which it modifies precipitation in its adjacent regions. We propose to answer the following questions: 1) Are variations in the BH linked to the rainfall variations during July and August in the SEUS? 2) If so, what is the mechanism that leads to variations in the BH and the occurrence of the precipitation variations in the midsummer?

This study is structured as follows: section 2 discusses the data and methods employed in this study. Section 3 discusses the variability of Bermuda high as well as its relations with precipitation over the SEUS. Section 4 investigates the structure of a teleconnection pattern that modulates the Bermuda high. A concluding discussion is presented in section 5.

2. Data and methods

We make use of monthly-mean data from the fifth reanalysis from the European Centre for Medium-Range Weather Forecasts (ECMWF) (ERA5; Hersbach et al. 2019). The data covers the time period starting in 1979 and ending in 2018. The data have a horizontal resolution of $0.25^\circ \times 0.25^\circ$ and 32 vertical levels ranging from 1000 to 10 hPa. The variables used in this study are geopotential height (z), temperature (T), horizontal and vertical winds (u, v, ω), specific humidity (q), sea level pressure (SLP), and mean total precipitation rate (P). For comparison, we use monthly precipitation rate data from the Global Precipitation Climatology Project (GPCP) (Adler et al. 2003, 2018), starting from 1979 to 2018, with a spatial resolution of $2.5^\circ \times 2.5^\circ$.

Many of the fields analyzed in this study are anomalies obtained by removing the seasonal cycle and linear trends. The seasonal cycle is removed by subtracting the monthly climatology from each respective month. Linear trends are also removed to avoid contamination with any long-term trends that may not be related to the interannual variations that are of interest to this study.

The position and shape of Bermuda high can be quantified through the use of the Bermuda high index (BHI) (Stahle and Cleveland 1992). The BHI was initially defined as the standardized SLP northeast of Bermuda (blue dot in Fig. 2) minus the standardized SLP over New Orleans (red dot in Fig. 2):

$$\text{BHI} = \text{SLP}'(40^\circ\text{N}, 60^\circ\text{W}) - \text{SLP}'(30^\circ\text{N}, 90^\circ\text{W}), \quad (1)$$

where SLP' denotes a standardized SLP anomaly field:

$$\text{SLP}' = \frac{\text{SLP} - \text{avg}(\text{SLP})}{\text{std}(\text{SLP})}, \quad (2)$$

where “std” is the standard deviation and “avg” is the average. The seasonal cycle and linear trends are also removed from the SLP anomalies.

Because we are interested in using the BHI as a time series for linear regression, we will standardize the time series so that

an amplitude of 1 corresponds to one standard deviation in the BHI:

$$\text{BHI}' = \frac{\text{BHI}}{\text{std}(\text{BHI})}. \quad (3)$$

While additional definitions of the BHI exist (Katz et al. 2003; Ortegren et al. 2011; Zhu and Liang 2013), we use the definition described by Eq. (1). We have verified that our results are reproducible with alternative definitions of the BHI (not shown).

Most of the results presented in this study are obtained from linear regression analysis based on the BHI. The statistical significance of the regression fields is tested by examining whether the correlation coefficient of each grid point exceeds a critical value. A critical coefficient value is obtained from a Fisher transformed correlation, following the method employed by Chen (1982) and Adames and Wallace (2014). Values that exceed this critical correlation coefficient are significant at the 95% confidence interval. To obtain the critical coefficient, we first calculate the effective number of degrees of freedom discussed by Davis (1976) and Chen (1982). The effective number of degrees of freedom is calculated through the decorrelation time scale (Adames and Wallace 2014), which is the sum of multiplied autocorrelations over all time lags for each individual grid point. One of the autocorrelations comes from the BHI and the other one is from the time series of a variable field (e.g., precipitation anomalies). With the effective number of degrees of freedom on each grid, the critical coefficient can be calculated through the method presented by Adames and Wallace (2014). A grid point will be deemed significant if its cross-correlation's absolute value is higher than the largest absolute value of critical coefficient in the domain.

We also use empirical orthogonal functions (EOFs, or principal component analysis) on the regional precipitation field. The EOFs' time series, also called principal components (PCs), are adopted to make regressions and comparisons with the BHI. Similar to EOF analysis, maximum covariance analysis (MCA) (Bretherton et al. 1992; Newman and Sardeshmukh 1995; Cherry 1996; Hu 1997) is also implemented in a joint field of precipitation and geopotential height at the 1000-hPa level. The results of MCA are also shown in appendix B (Fig. B1). To test if the EOFs are statistically distinguishable from one another, we use North et al.'s (1982) method to determine the confidence interval of the eigenvalue of the EOF/MCA analysis. The EOF and MCA patterns shown here were found to be statistically significant at the 95% confidence interval. As shown in the appendixes, the second leading pattern of variability in both analyses is highly correlated with the BHI (Figs. A1a and B1, respectively).

Some additional plots are obtained from compositing. These composites represent strong Bermuda high contraction (BHI+) and expansion (BHI−) events, defined as the times when the BHI exceeds +1 standard deviation or −1 standard deviation, respectively.

3. Variability in the Bermuda high during July and August

Figure 1a shows the time series of the BHI from 1979 to 2018. Figure 2 shows composites based on the months in which

FIG. 1. (a) Time series of the standardized BHI in July and August (JA) from 1979 to 2018, defined following Eq. (1). The red spikes denote a BHI amplitude greater than +1 std and the blue ones an amplitude less than -1 std. (b) Deseasonalized and detrended precipitation (mm day^{-1}) from ERA5 averaged over the SEUS (23° – 35°N , 75° – 95°W ; red box in Fig. 2). (c) As in (b), but for GPCP precipitation data. The dots in (b) and (c) are as in the spikes in (a), showing the months of strong BHI. (d) Scatterplot of standardized BHI and ERA5 precipitation anomalies. July is marked by blue dots and August is marked by red.

the BHI exhibited an amplitude greater than unity. As seen in Fig. 2, in years in which the BHI is positive, the Bermuda high remains over the ocean and contracts relative to its climatological position (Fig. 2a). In years which it is negative, the BHI expands westward into North America, relative to its climatological position (contours in Fig. 2b). The index exhibits an amplitude greater than 1 in July 1979, July 1984, July 1985, August 1988, July 1994, July 2004, August 2005, July 2007, July 2008, August 2012, July 2013, July 2014, and July 2018. It exhibits an amplitude of less than -1 during August 1980, July 1983, July 1986, July 1987, July 1992, July 1993, July 1999, July 2002, July 2012, July 2015, July 2016, and August 2018. The BHI is skewed: both strong positive and negative Julys outnumber the Augusts. This result indicates that the BH is more variable during July than during August. In spite of this variability, the large-scale patterns associated with fluctuations in the BH are similar during both months (not shown). Also, even the BHI is calculated from detrended data, there is a multidecadal trend. Before 2003, the BHI had more strong negatives but has tended toward strong positives since then. This trend indicates a shift of BH from mainly expansion to contraction in the recent two decades.

Figures 1b and 1c show JA precipitation from ERA5 and GPCP, respectively, averaged over the SEUS (23° – 35°N ,

75° – 95°W ; marked by red box in Fig. 2). The months when the BHI is positive are dotted in red and the negatives are in blue. Dry conditions are more likely to happen over the SEUS when the BHI is negative, which is consistent with previous studies (Henderson and Vega 1996; Zhu and Liang 2013). The correlation coefficient of ERA5 precipitation time series with BHI is 0.68 and 0.66 for GPCP. Figure 1d shows a scatterplot of the BHI and rainfall anomalies over the SEUS. The clustering of the cloud of points in Fig. 1d suggests a roughly linear relationship between the BHI and rainfall in the SEUS.

The shading in Figs. 2a–d shows the interannual variation of precipitation in strong positive (Figs. 2a,c) and strong negative (Figs. 2b,d) BHI phases. The contour lines show the area of the BH. During the positive years, a region of mean rainfall greater than 6 mm day^{-1} is observed over the SEUS following the northwestern edge of the Bermuda high. Consistent with Fig. 1b, during negative years the precipitation in SEUS decreases. In agreement with previous findings, the westward extension of the Bermuda high is correlated with suppressed rainfall (Giannini et al. 2000; Small et al. 2007; Gamble et al. 2008). When the BH is in the negative phase, rainfall is suppressed along with the western edge of BH (Kelly et al. 2018). The horizontal distribution and magnitude of precipitation are similar in ERA5 and GPCP data. Minor differences show up in

FIG. 2. (a),(b) JA composite maps of 850-hPa geopotential heights (contours; units: gpm), corresponding horizontal wind (vectors; units: m s^{-1}), and GPCP precipitation (shading) for BHI exceeding (left) 1 std and (right) -1 std. The contour for 1560 gpm, which is considered the edge of the Bermuda high (Li et al. 2011) is shown as a thick line. (c),(d) and (e),(f) As in (a) and (b), but showing ERA5 precipitation and column-integrated specific humidity ($\langle q \rangle$; units: kg m^{-2}), respectively. The SEUS region is shown as a red box in (a)–(d). The locations used to calculate BHI are marked by red (30°N , 90°W) and blue dots (40°N , 60°W).

the SEUS, however. For example, the ERA5 data exhibit less rainfall over Florida in the positive phase.

The distribution of column-integrated specific humidity ($\langle q \rangle$) is shown in Figs. 2e and 2f. The overall patterns are similar, except that during positive BHI months (Fig. 2e) higher concentrations of humidity are observed over Florida, whereas during negative BHI months higher values of $\langle q \rangle$ are observed in the U.S. Great Plains.

Figure 3a shows linear regressions of 850-hPa geopotential height and precipitation onto BHI. As described in section 2, the fields are anomalies obtained by removing the mean and seasonal cycle from the data. An anomalous anticyclone is seen over the northwestern Atlantic and an anomalous cyclone is seen over the SEUS and Gulf of Mexico. Collocated with the anomalous cyclone is a region of enhanced precipitation. Suppressed precipitation is seen near the northwestern Atlantic anticyclone. Apart from the two strong centers, another cyclone is seen in the northeastern Atlantic. No significant precipitation anomalies are seen near this cyclone.

Figure 3b shows the distribution of anomalous precipitation along with the anomalies in column-integrated specific humidity. The moisture anomalies are located at the edges of the anomalous cyclone and anticyclone shown in Fig. 3a. Anomalous humid conditions are seen at the equatorward component of the cyclone, while the reverse conditions are seen in the poleward side. To first order, the regions of enhanced precipitation correspond to the regions of increased water vapor and the suppressed rainfall corresponds to reduced water vapor. However, the precipitation anomalies are shifted to the north of the water vapor anomalies over the SEUS.

4. Northern Hemisphere wave train

The regression maps of anomalous geopotential height (contours) and precipitation (shadings) in Fig. 3a show a horizontal structure suggestive of a large-scale wave train in the Northern Hemisphere. To investigate this idea further, we

FIG. 3. (a) Linear regression of 850-hPa geopotential height anomalies (contours; interval: 2 gpm) and ERA5 precipitation rate anomalies (shading; units: mm day^{-1}) onto the BHI. (b) As in (a), but for column-integrated specific humidity anomalies (contours; interval: 0.2 kg m^{-2}). The black contours and blue and red shading are significant at the 95% confidence level.

FIG. 4. Vertical cross section of 40°–50°N averaged geopotential height anomalies (contours; interval: ± 4 gpm, beginning at 2 and -2 gpm, respectively) and vertical velocity anomalies (shaded; units: Pa s^{-1}) regressed onto the BHI. The black contours and blue and red shading are significant at the 95% confidence level.

examine longitude–height cross sections of geopotential height and vertical velocity averaged over the 40°–50°N latitude band (Fig. 4). We chose this band because most of the troughs and ridges of the wave, especially the anticyclone that is closely related to the BH over the North Atlantic, are located within this band (refer to Fig. 5). This cross section reveals a wave train, with alternating regions of high and low geopotential heights that encompass most of the Western Hemisphere. The vertical wind anomalies are located in the regions where the horizontal gradients in geopotential height are strongest, suggestive of quasigeostrophic forcing as indicated by the omega equation (Holton and Hakim 2013). We can qualitatively express the omega equation for this cross section as

$$\omega \propto \frac{\partial \mathbf{v}}{\partial p} \cdot \nabla_h \Phi, \quad (4)$$

where ω is the vertical velocity in isobaric coordinates, $\Phi = gz$ is geopotential, and p is pressure. Equation (4) indicates that when $\partial u / \partial p < 0$, which is typical for the midlatitudes, ascent occurs in regions where the geopotential is increasing with longitude and descent occurs in regions where the geopotential decreases with longitude. This type of pattern is clearly seen throughout the Western Hemisphere (180°–0°) in Fig. 4.

The results from Fig. 4 reveal the existence of a wave train that may affect the position and amplitude of the Bermuda high. Its strongest variations are located near 200 hPa. Such wave trains are usually guided by the midlatitude jet stream (Wirth et al. 2018). To verify this idea, Fig. 5 shows the geopotential height anomalies regressed onto the BHI and the JA-mean jet stream at 200 hPa. From 120°W to 0°, near the 40°–50°N band where the jet is strong, four clearly defined, statistically significant ridges and troughs are observed following the region of maximum westerly winds over the Atlantic and North America region. Such a pattern indicates that the westerly jet acts as a waveguide for this pattern (Branstator 1983). In addition to these four centers, there is a hint of a statistically significant signal over the Pacific Ocean. Most of the

FIG. 5. 200-hPa geopotential height anomalies (contours; interval: 3 gpm) regressed onto the BHI. The black contours are significant at the 95% level. The shading depicts the 200-hPa JA-mean zonal wind (units: m s^{-1}).

centers are nearly collocated with the lower tropospheric geopotential height signatures seen in Fig. 3a, indicative of an equivalent barotropic structure. For the centers over SEUS and North Atlantic, a small northwestward tilt in height is seen in the geopotential, as hinted at by Figs. 4 and 5.

Spectral analysis of the 200-hPa height anomalies reveals that the wave train exhibits a horizontal scale between zonal wavenumbers 6 and 7 (Fig. 6), corresponding to a wavelength of 3000–3500 km. Previous research has shown that Rossby wave trains with a zonal wavenumber of ~ 6 tend to be trapped around the latitude of the jet stream core (Hoskins and Karoly 1981; Branstator 2002). The alternating troughs and ridges in Fig. 5 are suggestive of such trapping. For a Rossby wave that propagates in the westerly flow, and has little variations in the latitudinal direction compared to its zonal scale, the stationary Rossby wavenumber is calculated to support the result from spectral analysis. Based on the method proposed

FIG. 6. Power spectral density (units: gpm^2) as a function of zonal wavenumber for 200-hPa geopotential height anomalies regressed onto the BHI and meridionally averaged over the 40°–50°N latitude belt. The spectral power at each zonal wavenumber is shown as a red dot.

FIG. 7. Geopotential height anomalies at 200 hPa (gpm; contours; interval: 6 gpm), Takaya–Nakamura (T–N) wave activity flux ($\text{m}^2 \text{s}^{-2}$; vectors), and vertical velocity anomalies (m s^{-1} ; shading) composited for negative BHI years. The vertical velocity is averaged from 300 to 500 hPa and smoothed in longitude with a running window of 2.5° . The black contours and blue and red shading are significant at the 95% confidence level. The vectors are masked out where the wave train is not significant at the 95% confidence level.

by Hoskins and Karoly (1981), Hoskins and Ambrizzi (1993), and Wills et al. (2019), the stationary Rossby wavenumber is calculated as follows:

$$K_s = a \cos \varphi \left(\frac{\beta - \frac{\partial^2 [U]}{\partial y^2}}{[U]} \right)^{1/2}, \quad (5)$$

where K_s is the stationary wavenumber, a is the radius of Earth, $\cos \varphi$ is cosine of latitude, $\beta = 2 \times 10^{-11} \text{ m}^{-1} \text{ s}^{-1}$ is the gradient

of planetary vorticity (i.e., the Rossby parameter), and $[U]$ is the zonal-averaged zonal wind. Based on the values of the fields shown in Fig. 5, we obtain a value of K_s of 6. Thus, the value of the stationary wavenumber is consistent with the observed scale of the wave train shown in Figs. 4 and 5.

Figure 7 shows the Takaya–Nakamura (T–N) wave activity flux (Takaya and Nakamura 1997, 2001) of the 200-hPa geopotential anomalies, overlaid with vertical velocity anomalies averaged over the 300–500-hPa layer, the layer in which vertical motions are strongest in Fig. 4. The wave activity fluxes are used to depict the propagation of energy in a wave train. Unlike the previous figures, these are composites based on the negative phases of the BHI. We chose to show composites due to nonlinearities in the wave activity flux calculation, which project poorly onto a linear regression. Wave activity flux arrows indicate eastward propagation of wave activity from the North Pacific toward the Atlantic. The regions of ascent and descent over the Atlantic sector are collocated with the enhanced and suppressed precipitation, respectively, suggesting that vertical motions associated with this wave train may at least partly explain the fluctuations in precipitation over this region. Statistically significant regions of vertical motion are also observed over the northeast Pacific and Great Lakes region.

To see if anomalous precipitation associated with this wave train occurs away from the BH region, Fig. 8 shows the precipitation anomalies over the Northern Hemisphere extratropics. Although significant vertical motions and associated geopotential height anomalies are seen throughout the extratropics, statistically significant precipitation anomalies are observed only over the northwestern United States, SEUS, and

FIG. 8. (a) 200- and (b) 500-hPa geopotential height anomalies (contours; interval: 3 gpm) and ERA5 precipitation anomalies (shading; units: mm day^{-1}) regressed onto the BHI. The black contours and the blue and red shading are statistically significant at the 95% confidence level.

Shadings: anomalies ($\text{kg}^2 \text{m}^{-3} \text{s}^{-3}$) regressed onto BHI, Contours: 850 hPa z' (gpm) regressed onto BHI

FIG. 9. 850-hPa geopotential height anomalies (gpm; contours; interval: 1 gpm) and column-integrated vertical velocity contributions ($\text{kg}^2 \text{m}^{-3} \text{s}^{-3}$; shading) regressed onto the BHI. The vertical velocities are the (a) ERA5 column-integrated vertical velocity, (b) column-integrated vertical velocity estimated from Eq. (8), (c) contribution from latent heating, (d) contribution from radiative heating, (e) zonal temperature advection, (f) meridional temperature advection, and (g) total horizontal temperature advection. Black contours and blue and red shading are statistically significant at the 95% confidence level.

northwestern Atlantic. A region of statistically significant rainfall is seen in the tropical western Pacific, although from Fig. 8 it is unclear if this region of rainfall is connected to the rest of the wave train. The latter two exhibit the strongest precipitation anomalies. This is likely because the SEUS and northwestern Atlantic exhibit high concentrations of water vapor during JA (Figs. 2 and 3; see also Källberg et al. 2005). The consistency in the structure of the wave in the 200- and 500-hPa levels supports the idea that this wave train exhibits an equivalent barotropic structure.

To quantify whether large-scale motions are the cause of the precipitation anomalies in the SEUS, we will now analyze the thermodynamic energy equation, which can be written following Nie and Sobel (2016) as

$$C_p \frac{\partial T}{\partial t} + C_p \mathbf{V} \cdot \nabla_h T + \omega \frac{\partial S}{\partial p} = J, \quad (6)$$

where $C_p = 1004 \text{ J K}^{-1} \text{ kg}^{-1}$ is the specific heat of dry air under constant pressure, $\partial/\partial t$ is the local time derivative, T is the temperature, $\mathbf{V} \cdot \nabla_h T$ is the horizontal advection of temperature, ω is the vertical pressure velocity, $\partial/\partial p$ is the partial derivative with respect to pressure, $S = C_p T + \Phi$ is the dry static energy (Φ is geopotential), and J is the diabatic heating rate. We assume that $\partial T/\partial t$ is negligibly small due to the low frequency of the pattern that we are analyzing, and $\partial S/\partial p$ is nearly constant (Adames and Ming 2018). The vertical integration of J describes the sum of the surface precipitation rate and the column-integrated radiative heating rate (Yanai et al. 1973):

$$\langle J \rangle = \frac{1}{g} \int_{100 \text{ hPa}}^{1000 \text{ hPa}} J dp \simeq L_v P + \langle Q_r \rangle, \quad (7)$$

where $\langle \cdot \rangle$ denotes the vertical integration, $g = 9.8 \text{ m s}^{-2}$ is gravitational acceleration, $L_v = 2.5 \times 10^6 \text{ J kg}^{-1}$ is the latent heat of vaporization, P is precipitation rate, and Q_r is the

radiative heating rate. Applying Eq. (7) to Eq. (6), and rearranging the terms, it becomes a diagnostic equation for ω :

$$\langle \omega \rangle \simeq \left(\frac{\partial \bar{S}}{\partial p} \right)^{-1} \left[L_v P + \langle Q_r \rangle - \left\langle C_p u \frac{\partial T}{\partial x} \right\rangle - \left\langle C_p v \frac{\partial T}{\partial y} \right\rangle \right], \quad (8)$$

where u is the zonal wind and v is the meridional wind. The overline in the dry static energy is used to denote that we are using a tropospheric mean value. The first term on the right-hand side is latent heat from phase changing, the second term is the column radiative heating anomaly, the third term is zonal temperature advection, and the fourth term is meridional temperature advection.

Equation (8) allows us to quantify the vertical velocity in the column as a result of diabatic heating or anomalous temperature advection, which is related to the QG omega equation in Eq. (4). The results, shown in Fig. 9, reveal that the pattern of vertical velocity in the SEUS is predominantly a result of latent heat release in precipitating clouds. However, ascent associated with radiative heating and warm advection is also seen in this region, although these are of a much smaller magnitude. Radiative heating contributes to the anomalous ascent near the northern part of the Gulf of Mexico and to the east of Florida. Both zonal (Fig. 9e) and meridional (Fig. 9f) temperature advection contribute to the anomalous ascent seen in this region. Zonal temperature advection predominantly contributes to ascent over the northeastern Gulf, while meridional temperature advection is more important over the western Gulf and to the east of Florida. That the precipitation in the Gulf of Mexico is located in a region of warm air advection suggests a potential relationship between the two.

5. Discussion and conclusions

Our study of the Bermuda high (BH) is based on the Bermuda high index (BHI) during July and August. The BHI shows that the BH varies on an interannual scale, expanding westward during some seasons and contracting during others. The strongest contraction and expansion tend to take place in July. A multidecadal variation is also found in the BHI, indicating a shift from majorly expansion to contraction in the last two decades. We hypothesize that this trend may be a result of low-frequency multidecadal oscillations. More work is needed to elucidate the case of this multidecadal shift.

To answer question 1—Are variations in the BH linked to the rainfall variations during July and August in the southeast United States (SEUS)?—we created a scatterplot comparing the BHI to rainfall in the (Fig. 1d). A strong correspondence is seen between the two, with a correlation of 0.68. Suppressed rainfall over the SEUS is related to the westward expansion of the BH.

Composites of mean precipitation illustrate the rainfall distribution in the two phases. During the contraction phase, precipitation shifts toward the southern, southeastern, and eastern coasts of the United States. This region also corresponds to the northwestern edge of the Bermuda high. During the expansion phase, the coastal areas are drier while the U.S. Midwest area is wetter. These findings are consistent with

previous studies that showed a correlation between the westward expansion of the BH and the occurrence of suppressed rainfall (Giannini et al. 2000; Small et al. 2007). Linear regression analysis of rainfall and geopotential on the BHI clearly show this relationship (Fig. 3). The same results are obtained on the basis of EOF and MCA analysis (see appendix B).

Fluctuations in tropospheric-averaged temperature across the United States are also observed in association with the fluctuations in the BH. The displacement of cold and warm centers with respect to anomalous cyclones and anticyclones, as well as the existence of temperature anomalies in the western United States, indicates that fluctuations in the BHI are the result of an extratropical wave train. The potential existence of this wave train would lead to a hypothetical answer to question 2: What is the mechanism that leads to variations in the BH and the occurrence of the precipitation variations in the midsummer? We performed BHI-based linear regression analysis of the geopotential height anomalies on the 200-hPa level, revealing a wave train that encompasses much of the northern extratropics. It exhibits a zonal wavenumber near 6 or 7 and appears to be guided by the midlatitude westerly jets. The wave train exhibits an equivalent barotropic structure with its strongest amplitude occurring at the 200-hPa level.

While the wave train encompasses a large region of the Northern Hemisphere extratropics, its modulation of precipitation is largely confined to the BH region. The anomalous vertical motion over the SEUS is mainly attributed to anomalous diabatic heating (Fig. 9). Radiative heating and horizontal temperature advection are much smaller in amplitude, but nonetheless contribute nonnegligibly to the total vertical motion. That the anomalous precipitation is collocated with warm advection suggests a potential physical connection, at least over the Gulf of Mexico. It is possible that ascent associated with warm air advection moistens the troposphere through isentropic lifting, leading to a more favorable environment for precipitation. Upper-tropospheric clouds then reinforce the lifting by reducing the outgoing longwave radiation. Previous studies also suggested that horizontal advection of moisture also plays a role in precipitation anomalies associated with BH fluctuations (Henderson and Vega 1996; Diem 2006). While we did not examine the role of horizontal moisture advection in this study, we do not discard the possibility that both temperature and moisture advection are creating a thermodynamic environment that is more favorable for precipitation in this region. Additionally, because of the way the precipitation anomalies are oriented with respect to the land, it is possible that interactions between the land, the sea, and the atmosphere also play a role in determining the distribution of rainfall. More work is needed to establish the mechanism that leads to rainfall in the SEUS in association with fluctuations in the BH.

Our results suggests that fluctuations in the BH modulates rainfall in the SEUS, and that this modulation is associated to a Rossby wave train in the midlatitudes. We sought to understand the energetics of this wave trough by calculating barotropic or baroclinic energy conversions but the results were inconclusive. It is possible that the Rossby wave train is the result of mechanisms that vary from year to year. Future work may be able to elucidate possible causes of the wave train, such

FIG. A1. (a) 850-hPa geopotential height anomalies (contours; interval: 2 gpm) and ERA5 precipitation rate anomalies (shading; units: mm day⁻¹) regressed onto the second principal component of the plotted domain, which explains 7.2% of the variance and has a correlation with BHI of 0.7. Black contours and blue and red shading are statistically significant values at 95% confidence level. (b) Standardized principal component 2.

as changes in Asian monsoon, through model simulation of varying levels of complexity.

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APPENDIX A

EOF Analysis

Regression patterns similar to those shown in the main text can be obtained through EOF analysis of rainfall data over the Atlantic–North America sector (17°–50°N, 10°–125°W). The first and second EOFs are statistically significantly separated according to the method presented by North et al. (1982). The leading EOF of rainfall variability in this region (not shown) is reminiscent of teleconnection patterns associated with ENSO variability during boreal summer (Ropelewski and Halpert 1987), and exhibits a correlation of 0.53 with the Niño-3.4 index. The second EOF (Fig. A1a) explains 7.2% of the total JA precipitation variance in the Atlantic–North America domain. Regression maps of rainfall and 850-hPa geopotential height

FIG. B1. (a) Homogeneous and (b) heterogeneous MCA2 of precipitation anomalies (shading) and 1000-hPa geopotential height anomalies (contours). It explains 12.4% of the total variance. The left expansion coefficient (precipitation) has a correlation with BHI of 0.78, and the right expansion coefficient (geopotential height) has a correlation with BHI of 0.79. The dark contours and blue and red shading are significant values at the 95% level.

are nearly identical to those obtained using the BHI. The principal component time series associated with this pattern exhibits a correlation with BHI of 0.7.

APPENDIX B

Maximal Covariance Analysis

Maximal covariance analysis is applied to precipitation and geopotential height anomalies. As in the EOF analysis discussed previously, the leading MCA pattern is correlated with ENSO. The MCA is also tested using North et al.'s (1982) method and proved to be significantly separated. The second MCA is more closely related to the BHI (shown in Fig. B1, with Fig. B1a showing the homogeneous map and Fig. B1b showing the heterogeneous map) and explains 12.4% of total variance. The map exhibits some similarity to the patterns shown in Fig. 3a, except the MCA emphasizes rainfall anomalies occurring over the tropical Atlantic more. The time series associated with the second-leading pattern exhibit a correlation with the BHI of 0.78. The principal component 2 shown in Fig. A1b has a trend similar to that of the BHI, of BH shifting from negative to positive during the recent two decades, even it is calculated from detrended precipitation.

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