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INTERANNUAL VARIABILITY OF HEAT CONTENT, AND ITS RELATIONSHIP WITH THE SUMMER MONSOON INTENSITY, IN THE SOUTH CHINA SEA

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Abstract: This study analyzes the Ishii 700 m heat content (HC) in the South China Sea (SCS). During the 1978-2012 period, the HC in the SCS changed dramatically on interannual timescales. Three main findings emerged from the analysis. 1) The first spatial pattern of the empirical orthogonal function (EOF1) was consistently distributed over most of the SCS, whereas that of the second empirical orthogonal function (EOF2) showed a dipole signal. 2) The HC anomalies in the SCS were closely related to the SCS summer monsoon intensity. When the HC over most of the SCS increased (decreased) in previous winter, the SCS summer monsoon was strengthened (weakened). Therefore, the HC behavior in the SCS during previous winter can well predict the intensity of the SCS summer monsoon. 3) HC anomalies in the SCS largely influence the monsoon and Walker circulations, in turn affecting the western Pacific subtropical high and finally the SCS summer monsoon.

Key words: South China Sea; heat content; summer monsoon; interannual variability; intensity
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1 INTRODUCTION

The South China Sea (SCS) is a marginal sea situated on the west side of the Pacific Ocean within the typical monsoon region. The SCS is an important part of the largest ocean heat library in the world and is sensitive to air-sea interactions (Li and Pan [1]). The SCS and its surrounding oceans are closely related to the onset and evolution of the southwest monsoon. The onset and strength of this monsoon depend on the air-sea heat exchange and early stages of the sea surface temperature (SST) (Lin and Zhang [2]; Ding et al.[3]). Because the oceans influence the atmosphere via heat exchange (Sui et al. [4]), understanding the thermal processes in the sea is important for clarifying the regional air-sea interaction mechanism.

The relationship between the SCS summer monsoon and ocean thermal conditions has been investigated in several studies (Gu et al.[5]; Wang et al.[6]; Liang et al.[7]; Chen and Hu [8]; Zheng et al.[9]). Gu et al. pointed out that a strong (weak) SCSSM induces a strong (weak) cross-equatorial flow over the Maritime Continent, which cools (warms) the local SST through anomalous surface latent heat flux and oceanic mixing[5]. In winter and spring, the HC distribution in the middle south of the SCS opposes that of late or early onset years of the SCS summer monsoon. An early (late) breakout of the SCS summer monsoon is accompanied by positive (negative) anomalies in the HC[6]. The Indian Ocean SST influences the onset of the SCS summer monsoon without removing the influence of the El Niño Southern Oscillation (ENSO). Positive and negative anomalies in the Indian Ocean SST are associated with early and late breakouts of the SCS monsoon, respectively [7]. The state of the warm pool of the West Pacific Ocean is also closely related to the SCS summer monsoon[8].

The impact of tropical ocean thermal conditions on the SCS summer monsoon has been extensively researched. As the SST is greatly affected by the air-sea heat flux, the SST reflects the ocean surface heat condition, whereas the HC and subsurface sea temperature better express the upper-level ocean dynamics. Less is known about the interannual variability of the SCS HC and its effect on the SCS summer monsoon activities because the ocean observation data are limited. Therefore, this paper analyzes the spatiotemporal distribution of the interannual variability of the HC in the SCS and its

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influence on the SCS summer monsoon intensity. It also proposes a possible mechanism for these behaviors.

2 DATA AND METHODS

The atmospheric datasets were provided by the National Centers for Environmental Prediction/National Center for Atmosphere Research (NCEP/NCAR) reanalysis (Kalnay et al.[10]). The data were resolved to 2.5° × 2.5° and extended from June 1979 to August 2012. We also employed the HadiSST monthly global SST dataset (Rayner et al. [11]) and the Ishii 700 m monthly HC data (version V6.13) (Ishii and Kimoto[12]). Both datasets were resolved to 2.5° × 2.5° and covered the January 1978 to December 2012 period.

From the NCEP/NCAR monthly data, we calculated the normalized SCS summer monsoon intensity of the southwesterly, as defined by Wu and Liang[13]:

$$I_s = \frac{V_{sw} - \bar{V}_{sw}}{\delta_{sw}}$$

In this expression, $V_{sw} = (u+v)/\sqrt{2}$, where u and v denote the 850 hPa zonal wind and the 850 hPa meridional wind in summer, respectively. $\bar{V}_{sw}$ and δ_{SW} are the average and standard deviation of V_{sw}, respectively, over the 1979–2012 period.

3 TIMESCALES OF HC ANOMALIES IN THE SCS

To determine the dominant timescales of the HC in the SCS, the monthly HC anomalies were subjected to wavelet analysis. Panels (a) and (b) of Fig.1 show the time series and wavelet analysis results of the monthly HC anomalies, respectively, averaged over the SCS (0-25° N, 100-125° E). From 1978 to 2012, the time series exhibited a clear interseasonal and interannual variability. The interannual variability of the HC anomalies in the SCS was stable from 1978 to 1998. After 1998, the HC anomalies increased until around 2002, and then declined before increasing again from 2006. According to the wavelet analysis (Fig.1b), the interannual timescale varied with periods of 1-3 and 4-6 years, consistent with the conclusions of Tong et al. [14] and He [15]. A semi-annual timescale variability was also found. Therefore, the ocean heat over the SCS varies on both interannual and semi-annual timescales (Gao et al.[16]; Tan et al.[17]). Such interannual variability is one of the main characteristics of climate change. This paper focuses on the interannual variability of the HC anomalies in the SCS and its relationship with the SCS summer monsoon intensity.

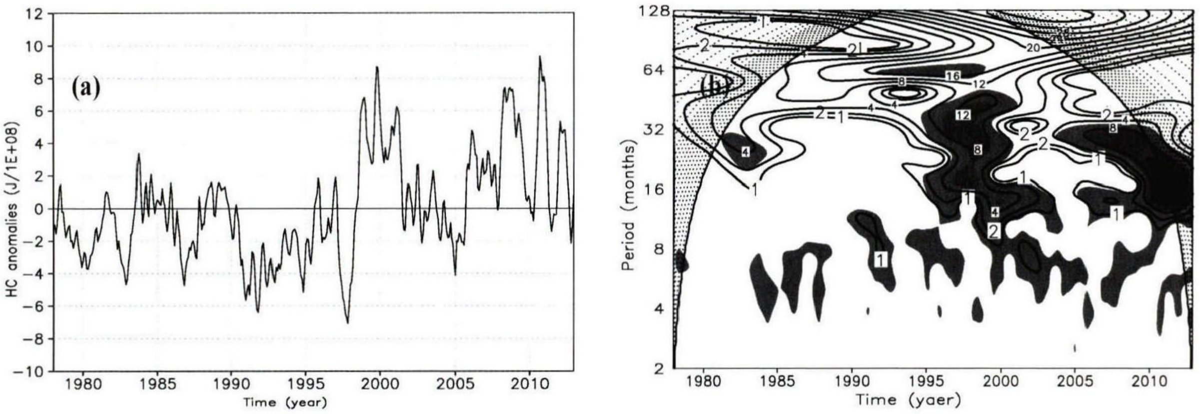

Figure 1. Time series (a) and wavelet analysis (b) of the monthly average HC anomalies. The averaging was performed over the SCS (0-25°N, 100-125°E) (shaded areas exceed the 95% confidence level).

4 CHARACTERISTICS OF INTERANNUAL VARIABILITY OF HC ANOMALIES IN THE SCS

To examine the spatial structure of the interannual variations in the HC anomalies, we performed an EOF analysis on the time series of the HC anomaly field over the SCS. After applying a 1-8 year band filter to highlight the interannual component and remove the linear trend (Huang et al. [18]), we calculated the anomalies by subtracting the seasonal cycle. The first four EOF modes accounted for 51.8%, 12%, 8.7%, and 6.2% of variance, respectively. The spatial patterns and time coefficients (Fig.2) of the first two modes pass the North criterion (North et al.[19]) for pattern distinction. The spatial pattern of EOF1 (Fig.2a) is consistent over most of the SCS signal. EOF1 is centralized in the middle east of the SCS. However, the time coefficient of EOF1 (Fig.2c) shows obvious interannual variability; in particular, it is largely negative during warm ENSO years (82/83, 86/87, 91/92, 97/98, 02/03, and 09/10) and largely positive during cold ENSO years (84/85, 88/89, 99/00, 00/01, and 10/11). The spatial pattern of EOF1 may reflect the ENSO (Yan et al.[20]; Wang et al.[21]; Sun et al. [22]), but the cold ENSO event in 07/08 did not yield a large positive time coefficient of EOF1, revealing that the HC anomalies in the SCS possess regional characteristics. The wavelet analysis shows a

2-7 year interannual periodicity in the time coefficient of EOF1, consistent with the ENSO primary period[20]. The spatial pattern of EOF2 (Fig.2b) shows a "+ –" dipole signal. The anomalies in the northeast part of the SCS HC oppose those over most of the SCS, and the center locates in the middle west of the SCS. This means that when the HC increases (decreases) in the northeast part of the SCS, it decreases (increases) over the rest of the SCS, especially in the mid-western region. The wavelet analysis reveals a 1-4 year interannual periodicity in the time coefficient of EOF2 before 2000, which increases to 1-5 years after 2000.

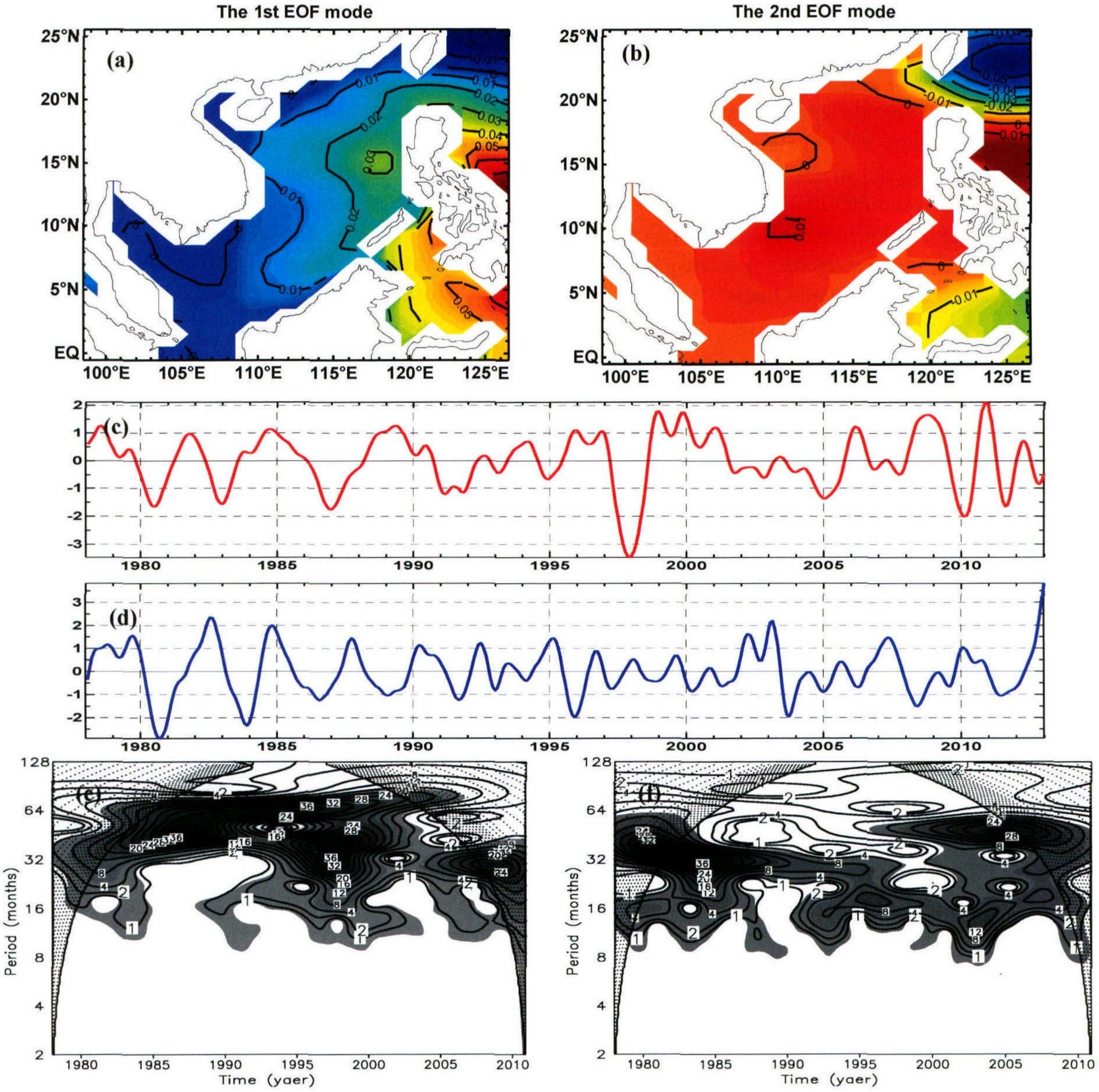

Figure 2. Spatial pattern (a), time coefficients (c) and wavelet analysis (e) of time coefficients of EOF1 for the monthly averaged HC anomalies; (b, d, and f) as for (a, c, and e) but for EOF2.

5 RELATIONSHIP BETWEEN HC IN SCS AND SCS SUMMER MONSOON INTENSITY

The influence of the oceans on the atmosphere eventually affects the heat exchange between the oceans and atmosphere. The oceanic HC is more stable than the SST and plays a greater role in the sustainable development of the weather and climate (Wu and Xu[23]). HC changes in the tropical oceans are known to be closely related to the SCS summer monsoon onset[6], but how do the HC anomalies in the SCS influence the strength of the SCS summer monsoon?

To relate the interannual variability of the HC anomalies in SCS and the SCS summer monsoon intensity, we present the lag correlation between the time coefficients of EOF1 and the SCS summer monsoon intensity in Fig.3a. The SCS HC anomalies distinctly affected the SCS summer monsoon intensity from previous summer to the following spring. The highest correlation is found in previous winter

(December-February), especially in February, when the correlation coefficient reaches 0.52 (above the 95% confidence level). Fig.3b shows the time series of SCS summer monsoon intensity and the previous winter (December-February) time coefficients of EOF1. The HC anomalies in the SCS are clearly related to the intensity of the SCS summer monsoon. In strong SCS summer monsoon years, such as 1985, 1994, 1997, and 1999, the HC in the SCS was abnormally high. In weak SCS summer monsoon years, such as 1983, 1988, 1995, 1998, and 2010, the HC anomalies in the SCS become negative. This indicates that when the HC in the SCS increases (decreases) in the previous winter, the SCS summer monsoon is strengthened (weakened). The increase or decrease of HC in the SCS well predicts the intensity of the SCS summer monsoon. In contrast, the SCS summer monsoon intensity is only weakly correlated with the time coefficients of EOF2, so this is not discussed here.

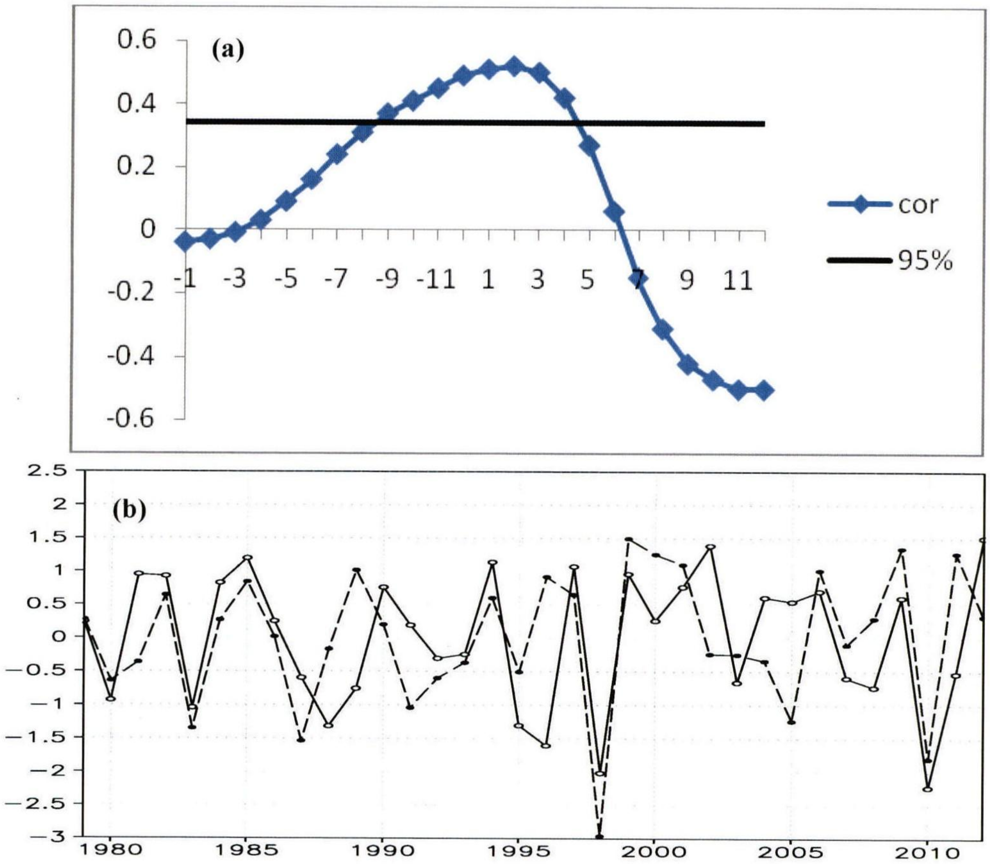

Figure 3. (a) Lag correlation between SCS summer monsoon intensity and time coefficients of EOF1 (negative and positive numbers on the abscissa refer to the previous and current month, respectively), (b) standardized time series of SCS summer monsoon intensity (solid line) and the previous winter (December-February) time coefficients of EOF1 (dashed line).

6 POSSIBLE IMPACT MECHANISMS OF HC ANOMALIES IN SCS ON THE INTENSITY OF SCS SUMMER MONSOON

To analyze how the previous winter HC anomalies in the SCS affect the SCS summer monsoon intensity, we calculated the correlations between the time coefficients of EOF1 and the atmospheric elements.

Figure 4a correlates the previous winter time coefficients of EOF1 with the 500 hPa vertical velocity in summer (June-August). The east of the SCS is negatively related to the vertical velocity in the western Pacific Ocean and positively related to those of the Arabian Sea, southeast Indian Ocean, and the middle-east Pacific Ocean in the southern hemisphere. This indicates that when the HC anomaly over the SCS is positive (negative) in previous winter, the rising (sinking) movement is strengthened in the east of the SCS and the western Pacific Ocean, while the sinking (rising) movement is strengthened in the Arabian Sea, southeast Indian Ocean, and the middle-east Pacific Ocean. There is a large region of significantly negative relatedness in the 500 hPa geopotential height field (Fig. 5b), covering South China, the eastern part of India, the Bay of Bengal, the SCS, and the western Pacific Ocean. This region corresponds to the control region of the western Pacific subtropical high (WPSH). The negative correlation of the geopotential height shows that when the HC in the SCS increases in previous winter, the WPSH weakens in the following summer. The weakening of the WPSH encourages the northward advance of the southwest summer monsoon, and vice

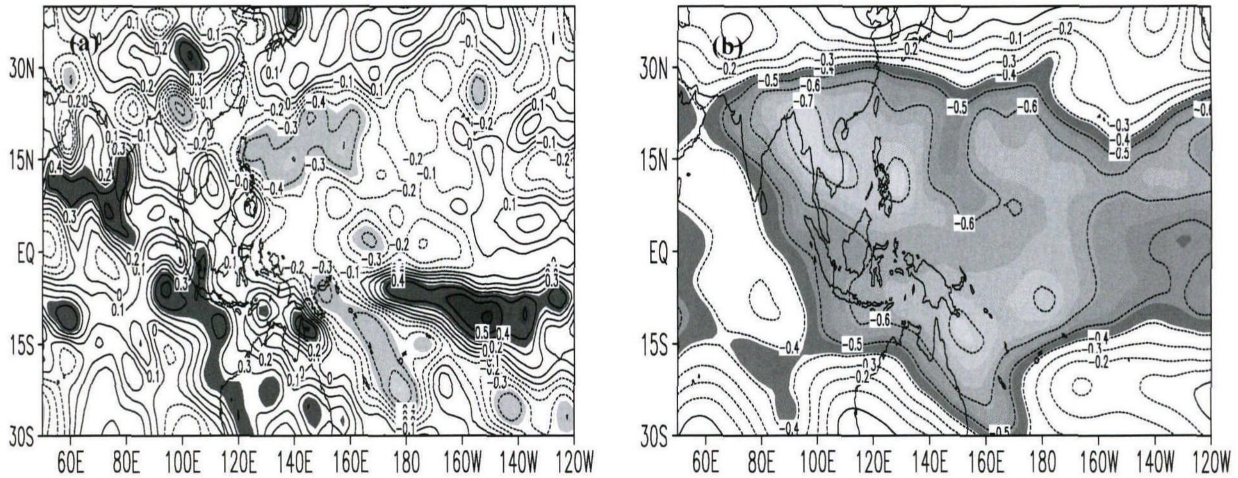

Figure 4. Correlation between the time coefficients of EOF1 in previous winter and the 500 hPa vertical velocity (a) and geopotential height field (b) in summer. (Shaded areas exceed the 90% and 95% confidence level in (a) and (b), respectively).

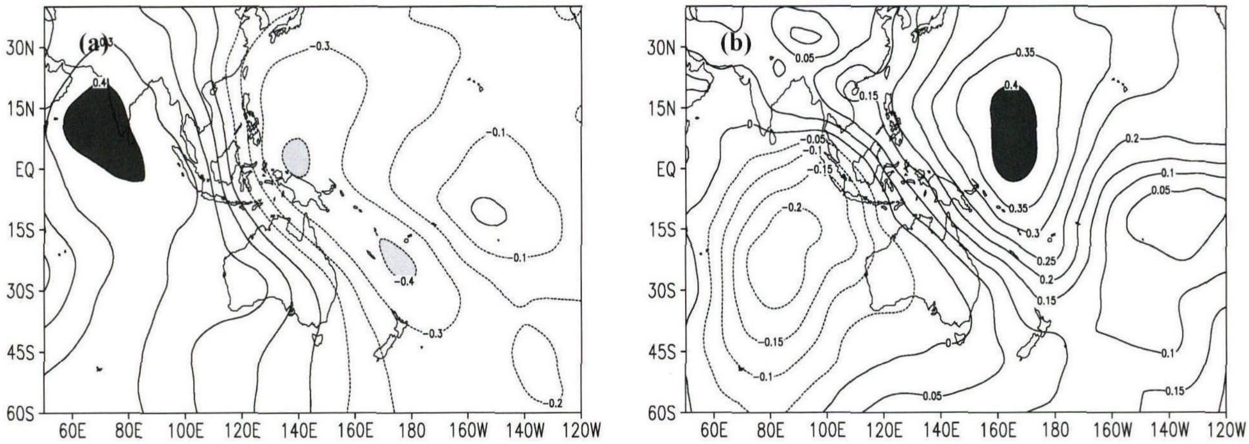

Figure 5. Correlation between the time coefficients of EOF1 in previous winter and the velocity potential at 200 hPa (a) and 850 hPa (b) in summer. (Shaded areas exceed the 95% confidence level).

versa. The previous winter HC anomalies in the SCS affect the summer monsoon similarly to the springtime HC anomalies in the Indian Ocean (Yu and Fen[24]).

The large-scale divergent circulation in tropical regions is associated with the imbalanced heating distribution. Panels (a) and (b) of Fig. 5 correlate the time coefficients of EOF1 in previous winter with the 200 hPa and 850 hPa velocity potentials in summer, respectively. Positive (negative) velocity potentials signify convergence (divergence). The 200 hPa velocity potential (Fig.6a) is significantly positively and negatively related to the HC in the Indian Ocean and western Pacific Ocean, respectively. That is, when the HC in the SCS increases (decreases) in previous winter, the movements significantly converge (diverge) in the Indian Ocean and diverge (converge) in the western Pacific Ocean. High-level convergent (divergent) movements correspond to vertical movements at the middle level (500 hPa). The enhancement of high-level convergence (divergence) strengthens the sinking (rising) movements in the middle level. The low-level (850 hPa) velocity potential correlation is oppositely distributed to high-level correlation. This correlation is significantly positive in the western Pacific Ocean and negative (but not significantly so) in the Indian Ocean. Therefore, the high-level convergence strengthens with strengthening low-level divergence, and vice versa.

The SCS summer monsoon intensity is closely related to the strength of the southwesterly winds. Fig. 6 shows the correlation distributions between the time coefficients of EOF1 in previous winter and the zonal winds at the high and low levels in summer. At the low level (850 hPa), the correlation is significantly positive around the Indian Peninsula, the Bay of Bengal, the SCS, the Philippines, and the ocean east of the Philippines, and significantly negative around the Yangtze River. This means that when the HC in the SCS increases (decreases) in the previous winter, the zonal wind increases (decreases) around the Indian Peninsula, the Bay of Bengal, the SCS, the Philippines, and the ocean east of the Philippines. In contrast, at the high level (200 hPa; see Fig.6b), the correlation is

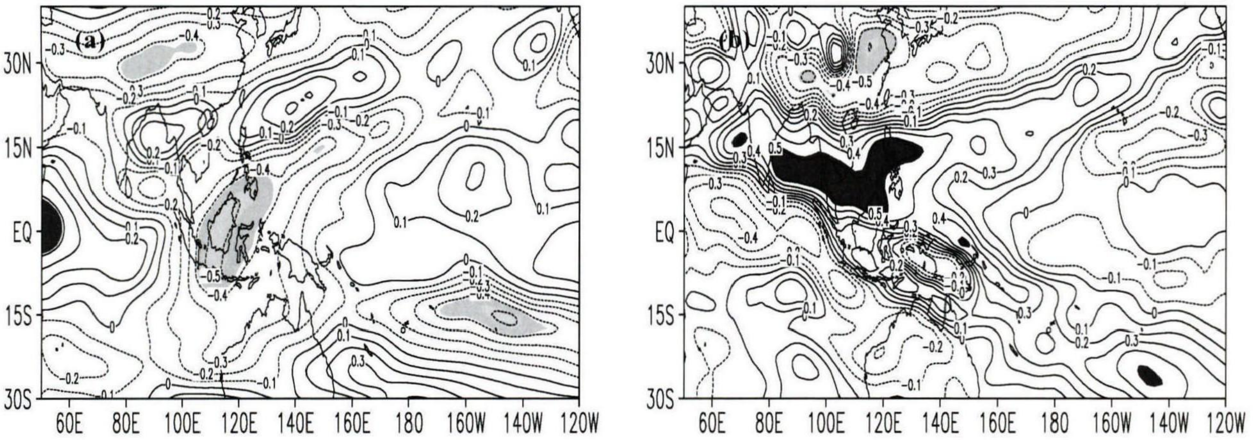

Figure 6. Correlation distribution between the time coefficients of EOF1 in previous winter and the zonal winds at 850 hPa (a) and 200 hPa (b) in summer. Shaded areas exceed the 95% confidence level.

significantly positive in the SCS (western Pacific Ocean and Kalimantan Island), implying that when the HC in the SCS increases (decreases) in the previous winter, the zonal wind increases (decreases) in these regions.

The above analysis confirms that the HC anomalies in the SCS affect the intensity of the SCS summer monsoon, probably by influencing the vertical movement above the tropical oceans and the divergence or convergence movements in the high and low levels of geopotential height. In turn, the high- and low-level convergence and divergence movements are closely related to the monsoon and Walker circulations. Simply speaking, the HC anomalies in the SCS can affect the intensity of the SCS summer monsoon by influencing the monsoon and Walker circulations. More specifically, when the HC increases in the SCS, upward movements prevail in the western Pacific Ocean, strengthening the low-level convergence and high-level divergence. Meanwhile, downward (sinking) movements prevail in the Indian Ocean, strengthening the low-level divergence and high-level convergence. The zonal wind, which centers at the divergence and convergence, abnormally blows westerly at the low level and easterly at the high level in the Pacific Ocean, and vice versa in the Indian Ocean. This conclusion is consistent with the Indo-Pacific gearing proposed by Wu (Wu and Meng[25, 26]) and the influence of the warm pool on the onset of the SCS summer monsoon (Gu et al.[27]; Li and Dai[28]). To verify the impact of HC anomalies on the SCS summer monsoon intensity, we performed a composite analysis on the time coefficients of EOF1. In this analysis, we selected years for which the standard deviation of the EOF1 coefficient equaled or exceeded 1. Under this criterion, 1989, 1999, 2000, 2001, 2006, 2009, and 2011 were selected as positively anomalous HC years and 1983, 1987, 1991, 1998, 2005, and 2010 were selected as negatively anomalous HC years.

In the 850 hPa wind anomaly field of the years showing positive abnormalities in the EOF1 time coefficients (Fig.7a), we find an anomalous cyclonic circulation near (25° N, 125° E), associated with a weakening and eastern withdrawal of the WPSH. South of the anomalous cyclonic circulation, an abnormal westerly prevails across the SCS, favoring northward movement and strengthening of the southwest monsoon. In years of abnormally negative EOF1 time coefficients, the 850 hPa circulation exhibits opposing patterns. Above all, the HC anomalies in the SCS largely affect the monsoon and Walker circulations. The latter circulation is closely related to the WPSH; in particular, a positive Walker circulation anomaly indicates a weak WPSH[8]. Chen and Hu[8] and Lu et al.[29] found that the HC anomaly can affect the convective activity, which is also closely related to the WPSH. Ultimately, the WPSH influences the SCS summer monsoon.

7 SUMMARY

Using the 1978-2012 Ishii 700 m HC dataset and the NCEP/NCAR reanalysis data, we investigated the interannual variability of the HC in the SCS and its impact on the intensity of SCS summer monsoons by EOF analysis, wavelet analysis, and a statistical method (composite analysis). The three main results of the study are summarized below:

(1) The first spatial pattern of the interannual variability of the SCS HC is consistent over most of the SCS; the second spatial pattern shows a "+ −" dipole signal. The time coefficient of EOF1 exhibits a 2-7 year interannual period consistent with the ENSO primary period. The time coefficient of EOF2 reveals a 1-4 year interannual period before 2000 and a 1-5 year interannual period thereafter.

(2) The SCS HC anomalies are closely related to the SCS summer monsoon intensities. When the greater part of the SCS HC increases (decreases) in previous winter, the SCS summer monsoon is strengthened (weakened). The HC changes in the SCS in previous winter well predict the intensity of the SCS summer

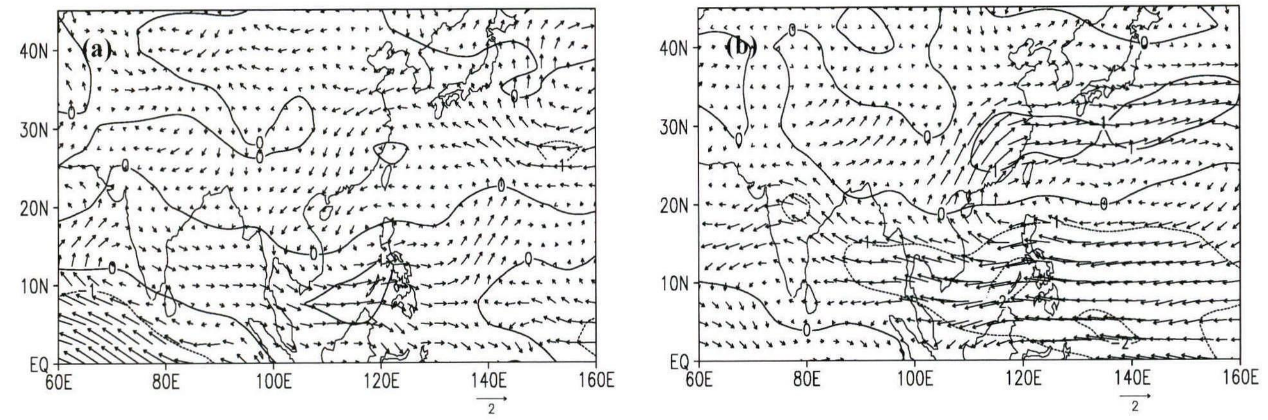

Figure 7. Composites of 850 hPa wind anomaly field, compiled from years of abnormally positive (a) and negative (b) time coefficients of EOF1 (unit: m/s). Contours denote the zonal winds, where positive and negative values indicate westerly and easterly anomalies, respectively.

monsoon.

(3) The HC anomalies in the SCS affect the SCS summer monsoon intensity chiefly by influencing the monsoon and Walker circulations. Specifically, the changes in these circulations are related to the WPSH, which ultimately affects the intensity of the SCS summer monsoon.

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