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Opposite responses of the diurnal amplitude of sea surface temperature to the Madden-Julian Oscillation

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ABSTRACT

Both observations and empirical model results show that the responses of the diurnal amplitude of sea surface temperature (DSST) to the Madden-Julian Oscillation (MJO) in the tropical regions are opposite on the two sides of a pivot longitude, which varies seasonally. This pivot longitude occurs at 180° in boreal winter, but moves to 150°E in boreal summer. When the active convection center of the MJO is located to the west of the pivot longitude, reduced solar insolation and enhanced wind speed decrease DSST; when the active convection center moves to the east of the pivot longitude, reduced wind speed increases DSST. Due to the opposing wind speed anomalies across the pivot longitude, opposite responses to the MJO can be seen in latent heat flux anomalies, evaporation rate anomalies and surface roughness anomalies. These findings suggest that opposite responses in the tropical regions on the west and east of the pivot longitude to the MJO be common for variables that are closely related to wind speed.

1. Introduction

As one of the fundamental periodic variations of sea surface temperature (SST), the diurnal cycle of SST has a significant influence on the ocean and atmosphere (Kawai and Wada, 2007; Guemas et al., 2013). In general, the diurnal amplitude of SST (DSST) is O(0.1 K) (Stuart-Menteth et al., 2003; Kennedy et al., 2007), but it depends primarily on solar insolation and surface wind speed, and can reach 5-7 K on a clear, calm day (Gentemann et al., 2008). Larger DSST is caused by either stronger solar radiation, which causes larger insolation absorbed at the sea surface, or lower wind speed, which causes weaker turbulent mixing in the upper ocean and less heat loss to the atmosphere (Kawai and Wada, 2007).

The Madden-Julian Oscillation (MJO), which is characterized by the eastward propagation of planetary-scale atmosphere circulation in the tropical regions coupled with deep convection over the tropical Indian and western central Pacific oceans, dominates the intraseasonal variation (30–90 days) in the tropics (Madden and Julian, 1971, 1972; Zhang, 2005). The seasonality of the MJO is featured by a latitudinal migration across the equator between the primary peak season in boreal winter (December-March) and the secondary one in boreal summer (June-September) (Zhang and Dong, 2004; Zhang, 2005). The MJO affects the degree of tropical cloudiness and hence solar radiation, zonal wind field and the ocean beneath it (Madden and Julian, 2005).

Both observations and model results demonstrate that the MJO has a large impact on DSST in the tropical Indian and western Pacific oceans (Bernie et al., 2005; Kawai and Kawamura, 2005; Bellenger and Duvel, 2009; Mujumdar et al., 2011; Li et al., 2013). Using moored observations in the western Pacific Ocean (156°E, 1°45′S) from December 1992 to March 1993, Bernie et al. (2005) found that DSST was relatively small (large) when the active (suppressed) convection center of the MJO was located in this region. This behavior was also observed in the tropical Indian Ocean (Mujumdar et al., 2011). Bernie et al. (2005) and Mujumdar et al. (2011) suggested that the MJO decreases (increases) DSST by reducing (increasing) solar insolation and enhancing (reducing) surface wind speed.

In contrast to the tropical Indian and western Pacific oceans, surface wind speed anomalies over the central tropical Pacific Ocean are negative (positive) when the active (suppressed) convection center of the MJO moves to the region (e.g., Hendon and Glick, 1997; Woolnough et al., 2000; Jin et al., 2013). This is because the climatological winds change from westerly to easterly near 180° (145°E) in boreal winter (summer) (Fig. 1). To our best knowledge, the response of DSST to the reversed wind speed anomalies associated with the MJO in the central tropical Pacific Ocean has not been reported previously. In this study, we examine the responses of DSST to the MJO over the entire tropical

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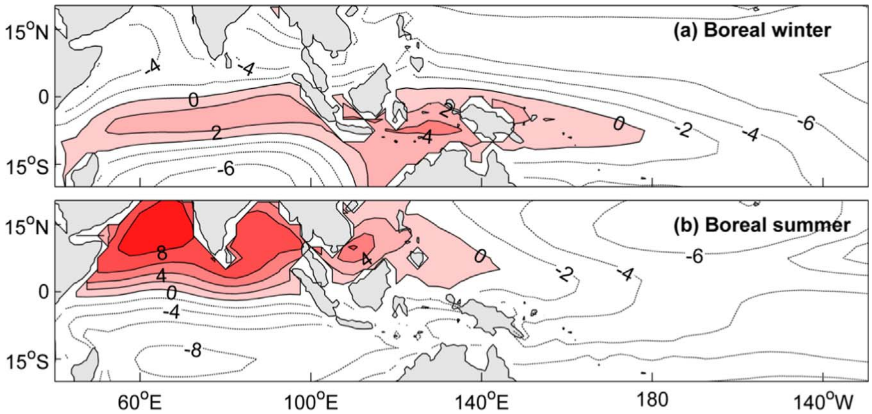

Fig. 1. Mean zonal wind (m/s) at 10 m in (a) boreal winter (December-March) and (b) boreal summer (June-September) during the period 1979–2013 derived from the ERA-Interim dataset.

Table 1
Comparison of intraseasonal signals of DSST in 2002–2003 estimated from the empirical model with the TAO/TRITON buoy data.

		137°E	147°E	156°E	165°E	180°	170°W	155°W	140°W
STD (Buoy)	5°N	0.128	0.147	0.101	0.134	0.115	0.127	0.083	0.063
	5°S				0.132	0.151	0.101	0.108	0.074
STD (Model)	5°N	0.124	0.120	0.106	0.112	0.114	0.102	0.092	0.066
	5°S				0.138	0.145	0.096	0.092	0.077
RMSE	5°N	0.051	0.052	0.060	0.052	0.053	0.050	0.050	0.039
	5°S				0.070	0.068	0.065	0.046	0.026
Corr.	5°N	0.92	0.94	0.83	0.92	0.89	0.93	0.84	0.82
Coef.	5°S				0.87	0.89	0.78	0.91	0.94

Indian and western central Pacific oceans, especially the response of DSST to the reversed wind speed anomalies in the central tropical Pacific Ocean. We use both the Tropical Atmosphere-Ocean/Triangle Trans-Ocean Buoy Network (TAO/TRITON) data and an empirical model.

2. Data and method

In this study, DSST at 1 m depth is estimated from the empirical model developed by Kawai and Kawamura (2003):

$$dSST = a(MS - H_l + e)^2 + b[\ln(U)] + c(MS - H_l + e)^2[\ln(U)] + d \tag{1}$$

where MS, H_l and U are, respectively, the daily mean solar insolation (positive downward, W/m^2), latent heat flux (positive upward, W/m^2) and 10 m wind speed (m/s); a, b, c, and d are regression coefficients; and e is 300 W/m^2 to prevent $MS - H_l + e$ from being negative. In this estimation of DSST, the daily mean wind speed below 0.5 m/s is set to 0.5 m/s, and the estimated DSST is set to zero when it becomes negative. Kawai and Kawamura (2003) validated the model-estimated DSST with both drifting and fixed TAO buoys in the tropics within 14°S–14°N, and concluded that the model estimate of DSST has a bias of less than 0.1K and a root mean square error (RMSE) of about 0.2 K. Compared to the empirical models of Webster et al. (1996) and Kawai and Kawamura (2002), latent heat flux is introduced into this model to reduce DSST bias in the tropics (Kawai and Kawamura, 2003, 2005). The daily mean solar insolation, latent heat flux and wind speed inputs to the model are derived from the ERA-Interim global atmospheric reanalysis with a horizontal resolution of 2.5°×2.5° over the period 1979–2013 (Dee et al., 2011).

To investigate the response of DSST to the MJO in the tropical Indian and western central Pacific oceans, the Real-time Multivariate MJO (RMM) index is used to identify the strength and phase of the MJO (Wheeler and Hendon, 2004). The (RMM1, RMM2) phase space consists of eight phases. Phases 1–3, 4–5 and 6–8 indicate the enhanced convection of the MJO over the Africa and the Indian Ocean, the Maritime Continent and the Pacific Ocean, respectively. In this study, composites of DSST anomalies with $\sqrt{RMM1^2 + RMM2^2} > 1$ for each MJO phase are constructed during both boreal winter and summer. The anomaly field is calculated by subtracting the time mean and the first three harmonics of the annual cycle for the 1979–2013 period.

To further verify the response of DSST to the MJO, the TAO/TRITON buoy observations, including hourly mean SST and wind speed, along 5°S and 5°N in 2002–2003 (strong MJO activity evidenced by large $\sqrt{RMM1^2 + RMM2^2}$) are selected for analysis. DSST is estimated as the difference between the maximum temperature during the diurnal cycle and the minimum temperature in the morning. The TAO/TRITON buoy data are also used to re-evaluate the empirical model. A comparison of intraseasonal signals of DSST in 2002–2003 estimated from the empirical model with the TAO/TRITON buoy data is listed in Table 1. We can find that the differences of the standard deviation (STD) of the intraseasonal signals are less than 0.03 K with RMSEs less than 0.07 K and their correlation coefficients greater than 0.78, which illustrates that the empirical model is capable of reproducing the intraseasonal signals of DSST.

3. Results

Fig. 2a shows the STD of DSST intraseasonal variability in the tropical Indian and western central Pacific oceans. Generally speaking, the STD of the intraseasonal DSST is relatively large in the Indo-Pacific warm pool (greater than 0.1 K), with its maximum values located in the eastern equatorial Indian Ocean, off the coast of northwestern Australia and in the southeastern South China Sea. The variance contribution rate of the intraseasonal DSST is mostly greater than 15% in the Indo-Pacific warm pool. The intraseasonal DSST signal exhibits obvious differences in its spatial pattern between boreal winter and summer, which are characterized by a northward shift across the equator. In boreal winter, the large STD in the intraseasonal DSST is mainly confined to the Southern Hemisphere, and its maximum values appear in the Timor and Arafura Seas off the coast of northwestern Australia (Fig. 2b), while in boreal summer the large STD shifts to the Northern Hemisphere with its maximum value located in the southeastern South China Sea (Fig. 2c). Further, a comparison of the intraseasonal signals between boreal