

RELATIONSHIPS BETWEEN SEA SURFACE TEMPERATURE, LARGE-SCALE ATMOSPHERIC CIRCULATION, AND CONVECTION OVER THE TROPICAL INDIAN AND PACIFIC OCEANS

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Abstract

In this paper, the quantitative estimates of the effect of large-scale circulations on the sea surface temperature (SST)-tropical convection relationship and the effect of SST on the large-scale circulation-convection relationship over the tropical Indian and Pacific Oceans are presented.

Although convection tends to maximize at warm SSTs, increased deep convection is also determined by the divergence (DIV) associated with large-scale circulation. An analysis of the relationship between SST and deep convection shows that under subsidence and clear conditions, there is a decrease in convection or increase in Outgoing Longwave Radiation (OLR) at a maximum rate of $3.4 \text{ Wm}^{-2} \text{ }^{\circ}\text{C}^{-1}$. In the SST range of 25°C to 29.5°C , a large increase in deep convection (decrease in OLR) occurs in the tropical Indian and Pacific Oceans.

The OLR reduction is found to be a strong function of the large-scale circulation in the Indian and western Pacific Oceans. Under a weak large-scale circulation, the rate of OLR reduction is about $-3.5 \text{ Wm}^{-2} \text{ }^{\circ}\text{C}^{-1}$ to $-8.1 \text{ Wm}^{-2} \text{ }^{\circ}\text{C}^{-1}$. Under the influence of strong rising motions, the rate can increase to about $-12.5 \text{ Wm}^{-2} \text{ }^{\circ}\text{C}^{-1}$ for the same SST range. The overall relationship between large-scale circulation and deep convection is nearly linear. A maximum rate of OLR reduction with respect to DIV is $-6.1 \text{ Wm}^{-2} (10^{-6} \text{ s}^{-1})$ in the western Pacific Ocean. It is also found that the DIV-OLR relationship is less dependent on SST. For example, the rate of OLR reduction over the western Pacific Ocean for $26^{\circ}\text{C} < \text{SST} \leq 27^{\circ}\text{C}$ is $-4.2 \text{ Wm}^{-2} (10^{-6} \text{ s}^{-1})$, while that for $28^{\circ}\text{C} < \text{SST} \leq 29^{\circ}\text{C}$ is $-5.1 \text{ Wm}^{-2} (10^{-6} \text{ s}^{-1})$.

These results are expected to have a great importance for climate feedback mechanisms associated with clouds and SST and for climate predictability.

1. INTRODUCTION

The largest part of the tropics is occupied by oceans, i. e., the tropical Pacific, Atlantic, and Indian oceans. Therefore, the large-scale dynamics of the tropical oceans and atmosphere is closely related. Energy is transferred from the atmosphere to the ocean surface mixed layer driving the circulation of the upper ocean. In turn, energy from the ocean is fed back to the atmosphere affecting the atmosphere circulation, the weather and the climate.

Many studies have shown that convection in the tropics is closely associated with sea surface temperature (SST). The annual cycle in SST in the eastern Pacific Ocean exhibits many similarities with annual cycle of rainfall (Horel 1982). Shukla and Wallace (1983) found during the mature phase of the ENSO event an eastward shift of the belt of heavy convective precipitation in the western Pacific Ocean is associated with positive SST anomalies in the eastern Pacific.

Some evidences that a certain SST range produces convection are given by previous investigators. Gadgil et al. (1984) found that over the Indian Ocean deep convection increases linearly with SST in range of $24.5^{\circ}\text{C} < \text{SST} \leq 28^{\circ}\text{C}$, whereas $\text{SST} > 28^{\circ}\text{C}$ no apparent relation between SST and deep convection could be found. Gutzler and Wood (1990) also showed that over the tropical Pacific Ocean the linear correlation between SST and Outgoing Longwave Radiation (OLR) is statistically significant only in limited areas with SST near 27.5°C but not in region with $\text{SST} > 28^{\circ}\text{C}$.

Waliser and Graham (1993) used the statistical analysis of the relationship between SST, OLR, and HRC to infer the role of convection in limiting ocean surface temperatures. They showed that the intensity of organized convection rises sharply as SSTs increase from 26.5°C to 29.0°C , reaches a maximum at 29.5°C , and then declines at higher temperatures. The presence of organized convection tends to limit upward excursions of SSTs. On the basis of a 'thermostat' hypothesis of SSTs being limited by the solar shielding effects of thick cirrus clouds and observed values of radiative feedback, Ramanathan and Collins (1991) calculated upper limits on SST of about 29°C - 32°C .

Zhang (1993) found that tropical deep convection remains weak and rarely observed for $\text{SST} < 26^{\circ}\text{C}$, the frequency and mean intensity of deep convection substantially increase with SST from 26°C up to about 29.5°C - 30°C , and then decay for further increasing SST. Meanwhile, in the warm pool region with $\text{SST} > 27^{\circ}\text{C}$, situation with no deep convection and vigorous deep convection can both be observed. In his study, there is no evidence of a 'critical SST' as suggested by Gadgil et al. (1984) and Waliser and Graham (1993). The increase in deep convection with SST is smooth and continuous and no distinct change in deep convection occurs at any particular SST. However, the above results appear to contradict other studies that suggest convection anomalies are primarily associated with anomalous low-level moisture

convergence rather than anomalies in evaporation. Carnejo-Garrido and Stones (1977) showed that over the tropical Pacific Ocean near 10°S in the region where condensation heating is a maximum, the evaporation and atmosphere heating by the surface are a minimum. This implies that the region of enhanced condensation is associated with a region of moisture convergence rather than with a region of enhanced evaporation, and that the SST gradients only play a secondary role in forcing the Walker circulation.

Although it is frequently assumed that the flow generated by the latent heat release in cumulus towers is an important component of the low-level convergence, Lindzen and Nigam (1987) argued that SST along with its gradients is a major contributor to the low-level tropical flow and convergence. These SST gradients are communicated to the atmospheric boundary layer by turbulent exchange from the surface, giving rise to low-level density gradients, and thus to low-level pressure gradients. These pressure gradients force the low-level tropical wind field and thus help determine the distribution of deep convection and rainfall.

Graham and Barnett (1987) found that over the Indian and Pacific oceans, SSTs in excess of 27.5°C are required for large-scale convection to occur. However, SSTs above 27.5°C are not a sufficient condition for convection and further increases in SST appear to have little effect on the intensity of convection. In addition, when SSTs are above 27.5°C , surface wind divergence is closely associated with the presence and absence of deep convection.

Lau et al. (1997) computed the influence of the large-scale atmospheric circulation on the relationship between SST and tropical convection inferred from OLR. They found that under subsidence and clear sky conditions, there is an increase in OLR with respect to SST at a rate of 1.8 to $2.5 \text{ Wm}^{-2} (^{\circ}\text{C})^{-1}$. In region of large-scale ascending motion, the rate of OLR reduction is found to be a strong function of the large-scale motion field. An intrinsic OLR sensitivity to SST is

found approximately -5 to $-4 \text{ Wm}^{-2}(\text{°C})^{-1}$ in SST range of 27°C - 28°C under condition of weak large-scale circulation, but under the influence of strong ascending motion the rate can be increased from -15 to $-12 \text{ Wm}^{-2}(\text{°C})^{-1}$ for the same SST range. On the other hand, deep convection and large-scale circulation exhibit a nearly linear relationship that is less dependent on SST and geographic locations.

The relative roles of the large-scale wind circulations on the relationship between SST and tropical convection and the relative roles of SST on the relationship between large-scale circulations and tropical convection have great importance for climate feedback mechanisms associated with clouds and SST and for climate predictability. In this study, a quantitative estimate of the effect of large-scale circulations on the SST-tropical convection relationships is derived from the linear regression coefficients (partial derivatives) between SST and convection at a constant wind divergence category. Similarly, an estimate of the effect of SST on the large-scale circulation-convection relationships is obtained from partial derivatives between large-scale circulations and convection at a constant SST category. The areas of interest will be the tropical Indian and Pacific Oceans.

2. DATA AND METHOD

The data used for this analysis are monthly SST, 200-mb wind divergence (DIV), and OLR. The analysis is based on 264 months of period from January 1975 to December 1997 (with a gap in 1978). The SST data are constructed from the Reynolds in situ analysis (Reynolds and Gemmill 1984) from January 1975 to December 1981 combined with the Optimum Interpolation data (Reynolds and Smith 1994) from January 1982 to December 1997. The in situ data sets are available at a $2^\circ \times 2^\circ$ resolution from the National Centers for Environmental Prediction-National Center for Atmospheric Research (NCEP-NCAR). The Optimum Interpolation data sets, which utilize in situ data from ships and buoys and from satellite data, are available at a $1^\circ \times 1^\circ$ resolution from the Climate Prediction Center (CPC)-NOAA.

The 200-mb DIV data sets are used as a measure of the large-scale circulations of the tropical atmosphere since they are closely related to the vertical velocity at 500 mb through the mass balance requirements. These data are derived from the monthly NCEP global reanalysis winds and available at a $2.5^\circ \times 2.5^\circ$ resolution.

The OLR data is available globally at a $2.5^\circ \times 2.5^\circ$ resolution and obtained from CPC-NOAA. The OLR data set is one of the most widely data used as a proxy for tropical convection (Waliser et al. 1993; Vincent et al. 1998). In the tropics, OLR is governed primarily by cloud top temperatures, so that low OLR values are related to high cloud tops or deep convection and thus to the ITCZ/SPCZ. All three data sets are then processed to provide monthly values on the same $2^\circ \times 2^\circ$ resolution from 25°N to 25°S between 0° and 358°E .

The analysis procedures are as follows. The relationship between SST and OLR is first examined by a scatter plot of collocated SST and OLR grid point values from the 264 months of period for each domain specified in Table 1 and shown in Fig. 1. It should be noted that the chosen Indian Ocean domain comprises central and eastern Indian Ocean. Each of plots comprises a total of over 130,000 data points. This large number of data points can significantly measure the relationship between these two variables.

The bulk SST-OLR regimes are then identified by calculating the mean OLR values as a function of every 0.5°C SST bin. Next, to examine the SST-OLR relationship as a function of the large-scale wind circulations. The upper-level divergence regimes of the tropical atmosphere are divided into 4 categories according to the strength of the wind divergence. Category DIV1 represents strong upper-level convergence (descending motion) with $-2 \times 10^{-5} \text{ s}^{-1} < \text{DIV} \leq -1 \times 10^{-5} \text{ s}^{-1}$, category DIV2 represents weak descending motion with $-1 \times 10^{-5} \text{ s}^{-1} < \text{DIV} \leq 0$, category DIV3 represents weak upper-level divergence (ascending motion) with $0 < \text{DIV} \leq 1 \times 10^{-5} \text{ s}^{-1}$.

¹, and category DIV4 represents strong ascending motion with $1 \times 10^{-5} \text{ s}^{-1} < \text{DIV} \leq 2 \times 10^{-5} \text{ s}^{-1}$. The dependence of SST and OLR on DIV is then obtained by calculating the

partial rate of change of OLR with respect to SST from linear regression equation between SST and OLR for each DIV category.

Table 1. Domains and longitude limits for the analysis.
Latitude limits are 25°N and 25°S.

Domain	Longitude limits
Indian	60° - 100°E
West Pacific	110° - 150°E
Central Pacific	160°E - 160°W
East Pacific	100° - 140°W

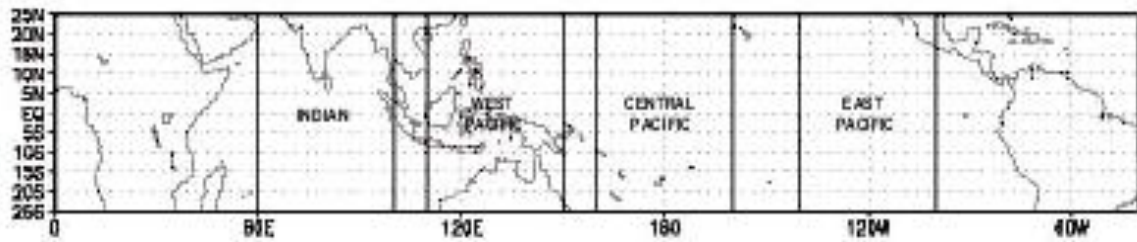


Figure 1. The analysis domains mentioned in Tabel 1.

To examine DIV-OLR relationship, scatter plots of DIV versus OLR are constructed for each domain. The mean OLR values as a function of every $0.5 \times 10^{-5} \text{ s}^{-1}$ DIV bin is intended to identify the DIV-OLR regimes. To examine the DIV-OLR relationship as a function of SST, the SST regimes are divided into five categories: Category SST1 represents $25^\circ\text{C} < \text{SST} \leq 26^\circ\text{C}$, category SST2 represents $26^\circ\text{C} < \text{SST} \leq 27^\circ\text{C}$, category SST3 represents $27^\circ\text{C} < \text{SST} \leq 28^\circ\text{C}$, category SST4 represents $28^\circ\text{C} < \text{SST} \leq 29^\circ\text{C}$, and category SST5 represents $29^\circ\text{C} < \text{SST} \leq 30^\circ\text{C}$. The partial rate of change of OLR with respect to DIV for each SST category is then computed to estimate the linear influence of SST on DIV-OLR relationship.

3. RESULTS AND DISCUSSION

The overall long-term means of the SST, OLR and DIV are shown in Figs. 2a-c. The mean SST values range between 15.7°C and 29.1°C , the mean OLR values range between 201.4 Wm^{-2} and 292.9 Wm^{-2} , and the mean DIV values range between $-3 \times 10^{-6} \text{ s}^{-1}$ and $7 \times 10^{-6} \text{ s}^{-1}$. The positive (negative) DIV values are associated with the rising (sinking) motions. Comparison of Fig. 2a and 2b reveals obvious

similarity between the spatial patterns of SST and OLR, with deep convection located over the warm waters ($\text{SST} \geq 27^\circ\text{C}$). However, there are some significant differences between the SST and OLR patterns. For example, the spatial gradients of the OLR patterns are much stronger than the SST patterns. Also, within the convective zone in the western Pacific, secondary maxima of deep convection are found over Kalimantan. Such conditions are not found in the SST field, but in the DIV field (Fig. 2c).

3.1 SST-OLR Relationships

Figures 3a-f show the scatterplots of collocated monthly SST vs OLR grid point values for the tropical Indian, western Pacific, central Pacific, and eastern Pacific oceans, respectively. Superimposed are the mean of OLR values at every 0.5°C SST bin and the associated standard errors given as error bars. The most noticeable feature from each domain is the elbow-like pattern marked by the rapid up-turn of the SST-OLR curve at about 26°C - 30°C . Based on the mean OLR values, the SST-OLR regimes for each domain are, furthermore, identified as:

- SST-OLR1 regime : an increasing OLR with increasing SST
 SST-OLR2 regime : a steep decreasing OLR with increasing SST
 SST-OLR3 regime : a drop-off of OLR with increasing SST

Previous studies (e.g. Lau et al. 1997) have demonstrated that the increase in OLR with SST in the SST-OLR1 regime is primarily due to the increasing SST minus the opposing effects of increasing water vapor under clear-sky /subsidence conditions. In the SST-OLR2 regime, the steep increase in the SST-OLR relationship can be attributed to the rapidly increasing occurrence of deep convection over warm waters or to large-scale moisture convergence. The SST-OLR3 regime is associated to the presence of hot spots in

isolated regions of forced subsidence particularly over the warm pool region.

Due to different dynamic and thermodynamic forcings, it is found that the SST range in the SST-OLR2 regime is different for different domains (Table 2). For example, the Indian Ocean has a rapid decrease of OLR (increase of convection) for SST between 25°C and 29°C. Meanwhile, a sharply skewed OLR distribution in the western Pacific Ocean occurs over warmer SST range from 26°C to 29.5°C. It should be noted that the statistics of high SSTs ($SST \geq 30^\circ\text{C}$) over the central and eastern Pacific oceans are contributed by relatively small number of data points, and hence may be affected by sampling errors.

Table 2. The SST ranges within each SST-OLR regime.

Domain	SST-OLR1	SST-OLR2	SST-OLR3
Indian	$SST \leq 25^\circ\text{C}$	$25^\circ\text{C} < SST \leq 29^\circ\text{C}$	$SST > 29^\circ\text{C}$
Western Pacific	$SST \leq 26^\circ\text{C}$	$26^\circ\text{C} < SST \leq 29.5^\circ\text{C}$	$SST > 29.5^\circ\text{C}$
Central Pacific	$SST \leq 26.5^\circ\text{C}$	$26.5^\circ\text{C} < SST \leq 31^\circ\text{C}$	$SST > 31^\circ\text{C}$
Eastern Pacific	$SST \leq 25.5^\circ\text{C}$	$25.5^\circ\text{C} < SST \leq 30^\circ\text{C}$	$SST > 30^\circ\text{C}$

Table 3 shows the partial rate of change of OLR with respect to SST ($\partial\text{OLR} / \partial\text{SST}$) in each domain for each SST-OLR regime. Under the subsidence and clear-sky conditions (SST-OLR1 regime), a near constant slope of SST-OLR relationship can be found in all domains with a minimum rate of $1.8 \text{ Wm}^{-2} \text{ }^\circ\text{C}^{-1}$ occurs in the Indian Ocean and a maximum

rate of $3.4 \text{ Wm}^{-2} \text{ }^\circ\text{C}^{-1}$ in the eastern Pacific Ocean. In this regime, increased occurrence of clear sky conditions will allow more insolation to reach the ocean leading to warmer SST. Over the eastern Pacific Ocean, this slope is more representative of the subsidence branch of the Walker circulation.

Table 3. The partial rate of change of OLR w.r.t. SST ($\partial\text{OLR} / \partial\text{SST}$) over different SST-OLR regimes.

Domain	$\frac{\partial\text{OLR}}{\partial\text{SST}} \left(\text{Wm}^{-2} \text{ }^\circ\text{C}^{-1} \right)$		
	SST-OLR1	SST-OLR2	SST-OLR3
Indian	1.8	-11.6	7.4
Western Pacific	2.6	-15.2	8.2
Central Pacific	2.7	-13.1	10.3
Eastern Pacific	3.4	-10.4	12.6

In warm SST regions (SST-OLR2 regime), there is a large decrease in OLR with respect to SST associated with increase in deep convection, with a minimum rate of $-10.4 \text{ Wm}^{-2} \text{ }^\circ\text{C}^{-1}$ in the eastern Pacific Ocean and a

maximum rate of $-15.2 \text{ Wm}^{-2} \text{ }^\circ\text{C}^{-1}$ in the western Pacific Ocean. Most of the net decrease in OLR can be attributed to the increased occurrence of deep convection overshadowing the background radiation. The

sensitivity of convection to local SST is strongly enhanced by the rising motion associated with the large-scale circulation. In the central and eastern Pacific oceans, the highest OLR sensitivity to SST is found during ENSO events. The migration of convection from the western Pacific to the central and eastern Pacific oceans is responsive to changes in the entire ocean-atmosphere structure.

A reduction in deep convection begins to occur for higher SSTs (SST-OLR3 regime) with a minimum rate of $7.4 \text{ Wm}^{-2} \text{ }^{\circ}\text{C}^{-1}$ in the Indian Ocean and a maximum rate of $12.6 \text{ Wm}^{-2} \text{ }^{\circ}\text{C}^{-1}$ in the eastern Pacific Ocean. However, as described earlier, the occurrence of relatively higher SSTs ($\text{SST} > 30^{\circ}\text{C}$) in the eastern Pacific Ocean is highly rare and questionable. A possible explanation of this behavior is that increased subsidence induced from nearby and remote convection provides warming and drying of the atmosphere above the boundary layer through adiabatic processes. In the Indian and western Pacific oceans, these conditions may be due to the formation of 'hot spots' (Waliser 1996). In the

eastern Pacific Ocean, these may be influenced by the insolation that warms the ocean and the vigorous ocean dynamics that cools it by upwelling.

Table 4 shows the $(\partial\text{OLR} / \partial\text{SST})$ values in the SST-OLR2 regime after the linear dependence of the SST-OLR relationship on DIV is removed. The increased convection is found to be a strong function of the large-scale motion field, particularly in the Indian and western Pacific oceans. Under condition of weak large-scale circulations (DIV2 regime), the $(\partial\text{OLR} / \partial\text{SST})$ values are $-3.5 \text{ Wm}^{-2} \text{ }^{\circ}\text{C}^{-1}$ and $-8.1 \text{ Wm}^{-2} \text{ }^{\circ}\text{C}^{-1}$ in the Indian and western Pacific, respectively. In contrast, under the influence of large-scale rising motions (DIV3 regime), the $(\partial\text{OLR} / \partial\text{SST})$ values can enhance to $-12.5 \text{ Wm}^{-2} \text{ }^{\circ}\text{C}^{-1}$ and $-12.8 \text{ Wm}^{-2} \text{ }^{\circ}\text{C}^{-1}$ or increase by about 257% and 58% for the Indian and western Pacific oceans, respectively. Meanwhile the effect of the rising motion in increasing deep convection is relatively much smaller in the central Pacific (35%) and eastern Pacific (11%) oceans.

Table 4. The $(\partial\text{OLR} / \partial\text{SST})$ at constant DIV categories for the SST-OLR2 regime.

Domain	$\frac{\partial\text{OLR}}{\partial\text{SST}} \left(\text{Wm}^{-2}\text{ }^{\circ}\text{C}^{-1} \right)$			
	DIV1	DIV2	DIV3	DIV4
Indian	4.9	-3.5	-12.5	-12.4
Western Pacific	-	-8.1	-12.8	-3.7
Central Pacific	10.5	-8.7	-11.7	-
Eastern Pacific	-	-7.5	-8.3	-

3.2 DIV-OLR Relationships

Figures 4a-f display the scatterplots of collocated monthly DIV vs OLR grid point values for the tropical Indian, western Pacific, central Pacific, and eastern Pacific oceans, respectively. The mean OLR values and the associated standard errors for every $0.5 \times 10^{-5} \text{ s}^{-1}$ DIV bin are also shown in these figures. Unlike the SST-OLR relationship, the relationship between DIV and OLR for all domains generally appears to be quite linear, particularly for $-0.5 \times 10^{-5} \text{ s}^{-1} \leq \text{DIV} \leq 1 \times 10^{-5} \text{ s}^{-1}$ range indicating an increase in rising motion results in an enhance in convection.

The partial rate of change of OLR with respect to DIV $(\partial\text{OLR} / \partial\text{DIV})$ in $-0.5 \times 10^{-5} \text{ s}^{-1} \leq \text{DIV} \leq 1 \times 10^{-5} \text{ s}^{-1}$ regime for each domain is shown in Table 5. As expected, the largest value is found in the western Pacific Ocean ($-6.1 \text{ Wm}^{-2} / 10^{-6} \text{ s}^{-1}$). In the eastern Pacific, the ocean-atmosphere interaction is most effective because of prevailing easterly trade winds and thus shallow thermocline. The ocean-atmosphere interactions are ineffective in the Indian and western Pacific oceans whereas cross-equatorial monsoon winds are more prominent than the trade winds. Therefore, the large-scale circulations have relatively

stronger control than local SST in these regions.

Table 5. The partial rate of change of OLR with respect to DIV ($\partial \text{OLR} / \partial \text{DIV}$) in $-0.5 \times 10^{-5} \text{ s}^{-1} \leq \text{DIV} \leq 1.0 \times 10^{-5} \text{ s}^{-1}$ regime.

Domain	$\frac{\partial \text{OLR}}{\partial \text{DIV}} \left(\text{Wm}^{-2} / 10^{-6} \text{ s}^{-1} \right)$
Indian	-4.8
Western Pacific	-6.1
Central Pacific	-5.1
Eastern Pacific	-4.3

Table 6 presents the ($\partial \text{OLR} / \partial \text{DIV}$) at constant SST categories derived using same regression procedure. In general, the relationship between DIV and OLR is nearly independent of SST. For example, over the Indian and western Pacific oceans, the ($\partial \text{OLR} / \partial \text{DIV}$) in the SST2 category ($26^\circ\text{C} < \text{SST} \leq 27^\circ\text{C}$) is $-3.5 \text{ Wm}^{-2}(10^{-6} \text{ s}^{-1})$ and $-4.2 \text{ Wm}^{-2}(10^{-6} \text{ s}^{-1})$, respectively. In SST4 category

($28^\circ\text{C} < \text{SST} \leq 29^\circ\text{C}$), the ($\partial \text{OLR} / \partial \text{DIV}$) is $-4.2 \text{ Wm}^{-2}(10^{-6} \text{ s}^{-1})$ and $-5.1 \text{ Wm}^{-2}(10^{-6} \text{ s}^{-1})$ or increase by only 34% and 21%, respectively. The DIV-OLR relationships suggest a fundamental link between the large-scale circulation and tropical convection that are less dependent on the spatial domains compared to the SST-OLR relationships.

Table 6. The ($\partial \text{OLR} / \partial \text{DIV}$) at constant SST categories in $-0.5 \times 10^{-5} \text{ s}^{-1} \leq \text{DIV} \leq 1.0 \times 10^{-5} \text{ s}^{-1}$ regime.

Domain	$\frac{\partial \text{OLR}}{\partial \text{DIV}} \left(\text{Wm}^{-2} / 10^{-6} \text{ s}^{-1} \right)$				
	SST1	SST2	SST3	SST4	SST5
Indian	-1.5	-3.5	-4.5	-4.2	-4.4
Western Pacific	-3.3	-4.2	-4.5	-5.1	-4.9
Central Pacific	-3.6	-3.6	-4.0	-4.4	-4.3
Eastern Pacific	-3.6	-3.6	-3.2	-3.9	-4.2

4. CONCLUSION

In summary, depending on the strength of large-scale upward motion or SST in each domain of the tropical oceans, the transition occurs at a SST range of 25°C to 26°C . Below this SST range namely under subsidence and clear sky conditions, an increase in OLR with respect to SST is at a rate of $1.8 \text{ Wm}^{-2} \text{ }^\circ\text{C}^{-1}$ to $3.4 \text{ Wm}^{-2} \text{ }^\circ\text{C}^{-1}$. Above this SST range to about 30°C , there is a large apparent decrease in OLR (increase in convection) with SST.

The SST-OLR relationship is strongly enhanced by the rising motions associated with the large-scale circulation. Under conditions of weak large-scale circulation, the rate of OLR reduction with respect to SST in the Indian and western Pacific oceans is about

$-3.5 \text{ Wm}^{-2} \text{ }^\circ\text{C}^{-1}$ and $-8.1 \text{ Wm}^{-2} \text{ }^\circ\text{C}^{-1}$, respectively. For the same SST range, under the influence of strong rising motions, the respective rate is increased to $-12.5 \text{ Wm}^{-2} \text{ }^\circ\text{C}^{-1}$ and $-12.8 \text{ Wm}^{-2} \text{ }^\circ\text{C}^{-1}$. On the other hand, for each domain the relationship between large-scale circulation and deep convection is nearly linear and less dependent on SST.

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