Atmospheric Precipitation in Response to Equatorial and Tropical Sea Surface Temperature Anomalies

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ABSTRACT

Three sets of numerical experiments based on a GFDL GCM were developed to investigate the response of the large-scale tropical circulation and precipitation to the tropical and equatorial sea surface temperature (SST) anomalies. Specified SST anomaly (SSTA) with a small latitudinal scale of 13.5° was imposed in different regions of the Pacific Ocean in different sets of experiments and added to the climate-mean August SST to form a lower boundary forcing. Each set is composed of two experiments in which the SSTA possesses the same coverage and intensity but opposite sign. Anomalies of meteorological fields are calculated as the differences between the results of the warm and cold SSTA experiments.

In all experiments, prominent anomalous low-level convergence and high-level divergence are observed over the warm SSTA regions. For experiments with warm SSTA placed at the equator, responses of the tropical streamfunction are similar to corresponding results found in other studies. When the warm SSTA is placed in the warmest SSTA region in the western North Pacific away from the equator, the excited anomalous streamfunction is different from that in the equatorial SSTA cases. A strong anomalous low-level cyclone and high-level anticyclone are generated, and strong anomalous westerly–southwesterly flow at lower levels and northeasterly–easterly flow at upper levels sweep through the southeastern part of the region.

Anomalous rainfall is shown to be balanced mainly by anomalous convergence of stationary water flux; transient flux and anomalous evaporation from the warm water surface are secondary. Advection of water vapor by the large-scale flow and its anomaly were found to be significant in determining the rainfall pattern. Anomalous precipitation occurs in regions where the mean flow is down the SSTA gradient, or the anomalous flow is down the mean SST gradient. Mainly due to advection of water vapor by the divergent wind component and its anomaly, abnormal rainfall near the equator is biased toward the hemisphere where near-equatorial SST is warmer. Advection of anomalous water vapor by strong low-level equatorial westerlies in the eastern equatorial Pacific causes anomalous rainfall associated with the warm SSTA in the region to shift westward. Away from the equator, advection of water vapor by the anomalous rotational wind becomes important. It is this contributor that causes anomalous rainfall to shift away from the warmest SSTA region in the western North Pacific.

1. Introduction

During the warm phase of the Southern Oscillation, the so-called El Niño event, warmer than normal SST is observed in the central or eastern tropical Pacific, or both. As sea surface pressure becomes abnormally low over the eastern Pacific and abnormally high over the western Pacific, the Walker circulation and the trade winds weaken (Bjerknes 1966, 1969). Enhanced equatorial precipitation is observed near the date line, west of the warm SST anomaly (SSTA) region (Flohn and Fleer 1975). The distribution of the anomalous rainfall is asymmetric with respect to the relatively symmetrical distribution of the SSTA. Such asymmetry appears to be seasonally dependent. Anomalous rainfall near the equator associated with the warm ENSO events are usually biased toward the summer hemisphere (Rasmussen and Carpenter 1982). These observations have been successfully reproduced by a number of numerical modeling studies (Rowntree 1972; Manabe et al. 1974; Julian and Chervin 1978; Keshavamurty 1982; Lau 1985; Ting 1990). For example, in their GCM experiments, Shukla and Wallace (1983, hereafter abbreviated as SW) used the GLAS climate model with January initial conditions based on observed data and an eastern equatorial Pacific positive sea surface temperature anomaly, based on the analysis of Rasmussen and Carpenter, to investigate the atmospheric response to equatorial Pacific SSTAs. Their results show that the heavy equatorial precipitation shifts eastward from near 150°E in the experiments without SSTA to near the date line in the experiments with SSTA. In all experiments, the major redistribution of precipitation

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takes place in the western half of the Pacific as observed, even though the largest SST anomalies are in the eastern half of the Pacific.

At the equator the SST difference over the eastern Pacific between the date line and the western coast of South America is only about 4°C in August. The Clausius–Clapeyron relation can only explain an increase of less than 25% between the water vapor content in the western region and in the eastern region. In observations, abnormal rainfall associated with a warm ENSO episode, which occurs in the western part of the warm SST region, can be several times as large as that which occurs in the eastern part (Sardeshmukh and Hoskins 1985). The estimate thus fails to explain the asymmetrical distribution of anomalous rainfall.

What is then responsible for the bias of the anomalous rainfall? Can enhanced rainfall be biased toward a colder SST area? Observational analyses have shown that the aforementioned anomalous rainfall is mainly associated not with increased local evaporation from the warmer seawater, but with intensified convergence of water vapor into the convective rainfall area (Cornejo-Garrido and Stone 1977; Ramage and Hori 1981; Philander 1989). Results from numerical modeling of SW also show that local changes in evaporation can account for only a small fraction of the changes in precipitation, and changes in moisture convergence should play a major role in the intensified rainfall. Since the latter depends much on large-scale circulation and its anomaly, in order to answer these questions we must study not only the direct impacts of SST on thermal features of the atmosphere, but also the large-scale circulation and its anomaly associated with changes in the underlying SST. Although SW have postulated that the increase in low-level westerlies to the west of positive SST region may play the leading role in accounting for the changes in the moisture budget associated with anomalous rainfall, which occurs in response to the SSTAs, this postulation needs to be verified by detailed budget calculations. This is one of the goals of the present study.

Some basic features of the response of the tropical atmosphere to diabatic heating have been elucidated by Gill (1980) based on a simple analytic model. Low-level easterly inflow and westerly inflow were generated to the east and west of the heating region, respectively. Similar features of the tropical atmospheric response to a given equatorial SST have been obtained by Keshavamurty (1982) based on a low-resolution version of the GFDL climate model. The latter study also showed that SST over equatorial central and western Pacific are more efficient in producing atmospheric circulation and rainfall anomalies than equal ones over the eastern Pacific. In these studies, meridional scales of the imposed heating or SSTAs fields are large. The e-fold scale in the north–south direction in Gill’s model is half of the width of the beta plane. The latitude span of the imposed SSTA in the study of Keshavamurty is about 40°. It covers not only the equatorial region but also some parts of the tropics. In order to differentiate the response of the tropical atmosphere to equatorial and tropical forcing, in the present study SSTAs with much smaller latitude scale of 13.5° were introduced into the same low-resolution version of the GFDL climate model used by Keshavamurty to initiate several numerical experiments.

As in the study of Keshavamurty, the equatorial SSTAs was carefully placed in different regions of the Pacific Ocean where either the climate-mean SST is the coldest, or the Walker circulation has its strongest ascent or descent. In addition to the equatorial SST experiments, a tropical SSTA with a similar size to that used in the equatorial forcing experiments was placed in the warmest SST region in the western North Pacific (WNP) to initiate tropical forcing experiments. These SSTAs were added to the August climate-mean SST and imposed at the lower boundary of the climate model to initiate a series of perpetual August integrations.

The model description and experiment design will be presented in the next section. Based on these experimental results, section 3 is devoted to the study of large-scale circulation and its anomaly in relation to the corresponding equatorial and tropical SST anomalies. Based on these results, the anomaly of the atmospheric water budget can be evaluated, and the anomalous rainfall in response to equatorial and tropical SSTAs can be studied. These will be discussed in section 4. Conclusions and discussions are presented in section 5.

2. Model description and experiment design

A low-resolution version of the GFDL global climate model with spectral harmonics truncated rhomboidally at zonal wavenumber 15 (R15) was used in this study. It has nine layers in the vertical. Orographic effects have been taken into account. The physical processes included in the model are solar and infrared radiation, cloudiness, boundary flux, and subgrid-scale diffusion of momentum, sensible heat, and moisture. Over ocean areas, the surface temperature is prescribed. Over land, the surface temperature is determined from an energy balance equation by assuming zero heat capacity. A description of the model and its performance can be found in Manabe et al. (1979).

Perpetual mean August climate has been imposed for the control experiment. The solar height angle, the length of day, and the SST were set at their mean August values. Because August corresponds to the developing phase of the El Niño/Southern Oscillation (ENSO), and because convective activities over the Northern Hemisphere oceans are the most active, the specification of the perpetual August in our experiments should highlight the relationship between anomalous convective rainfall and SSTAs more clearly.

In order to investigate the sensitivity of the tropical circulation and precipitation to equatorial and tropical
SSTA, six additional experiments, each with different geographical locations or signs of SSTA, have been developed (see Table 1). The SSTA in three different regions, as shown in Fig. 1, have been added to the climate-mean August SST used in the control run and used as the lower boundary forcing for the different experiments. The three regions are located, respectively, in the western equatorial Pacific (WEP), eastern equatorial Pacific (EEP), and western North Pacific (WNP). WEP and EEP were chosen because the centers of the specified SSTA in these regions are below the ascent and descent branch, respectively, of the Walker circulation in the control run, and closely linked to the ENSO. In addition, the SST in the EEP is the coldest in the tropical North Pacific in the boreal summer, and the climatic mean of the lower tropospheric easterlies over this region is the strongest (>5 m s\(^{-1}\)) in the equatorial region. The locations of the equatorial SSTA are very close to the corresponding SSTA locations in the study of Keshavamurty (1982). WNP was selected because it is located over the warmest SST area and coincides with the maximum frequency of tropical storm development (Gray 1968; Frank 1987). Since the center of the SSTA region is located at 18°N, this SSTA can then generate a diabatic heating asymmetric about the equator.

Comparison of the results obtained from this study with those obtained from the others will enable us to investigate the difference in atmospheric response to equatorial and tropical forcing. The areas of SSTA in each of the experiments have the same spans in both longitude (60°) and in latitude (13.5°). For each region, there is a pair of positive SSTA (experiment A, B, and E) and negative SSTA (experiment C, D, and F) experiments. In the outer region, SSTA is set to be 1°C in the WEP and WNP cases, and 2°C in the EEP cases. In the inner region, the SSTA specified is doubled that specified in the corresponding outer region, as shown in Table 1 and Fig. 1.

The observed extremes of July mean SSTA in the eastern equatorial Pacific in the El Niño year 1972 and in the La Niña year 1973 are 8°C and -6°C, respectively (taken from Fishing Information, U.S. National Marine Fisheries Service, see Fig. 1 of Keshavamurty for reference). The magnitudes of SSTA in the EEP region in experiments B and D were specified in reference to these observations. Those in the WEP region in experiments A and C were determined by experiments to obtain a precipitation difference between A and C in the region close to the precipitation difference between B and D in region EEP (see Figs. 3a and b). In this sense the latent heat difference in these two pairs of experiments may be considered comparable, and the sensitivity of atmospheric circulation in response to the equatorial latent heating in different regions can be investigated. The magnitudes of SSTA specified in experiments E and F in WNP are the same as those in experiment A and C, so that the differences in atmospheric response to equatorial and tropical SSTA can be compared.

In each experiment, the integration period was 1800 days or 60 model months. The SSTA was introduced on the second model day. The integration was started with an atmosphere in a state of thermal equilibrium. It reached quasi-equilibrium around the sixth model month. Therefore, only the results from day 181 to day 1800 were retrieved for the following analyses. Integration output has also been archived on the basis of 180-day means. In such circumstances, each parameter extracted for the analyses has a sample number of 9. The statistical significance of the response differences between two experiments can then be evaluated by using Student's t-test as in Chervin and Schneider (1976).

3. Response of large-scale flow to equatorial and tropical SSTA

We will concentrate on the very large scale atmospheric responses to low-latitude SSTA. For this purpose, the horizontal wind \( \mathbf{v} \) is divided into a divergent component \( \mathbf{v}_x \) and a rotational component \( \mathbf{v}_\psi \):

\[
\mathbf{v} = \mathbf{v}_x + \mathbf{v}_\psi = \nabla x + k \times \nabla \psi
\]
Table 1. Maximum SST anomalies specified for each experiment in different regions of the Pacific Ocean. Units are in degrees celsius.

<table>
<thead>
<tr>
<th>Expt.</th>
<th>WEP</th>
<th>EEP</th>
<th>WNP</th>
<th>Main characteristics</th>
</tr>
</thead>
<tbody>
<tr>
<td>A</td>
<td>+2.0</td>
<td>/</td>
<td>/</td>
<td>Below the ascent branch of the Walker circulation</td>
</tr>
<tr>
<td>B</td>
<td>/</td>
<td>+4.0</td>
<td>/</td>
<td>Below the descent branch of the Walker circulation</td>
</tr>
<tr>
<td></td>
<td></td>
<td></td>
<td></td>
<td>Coldest SST in tropical North Pacific</td>
</tr>
<tr>
<td>C</td>
<td>−2.0</td>
<td>/</td>
<td>/</td>
<td>Below the ascent branch of the Walker circulation</td>
</tr>
<tr>
<td>D</td>
<td>/</td>
<td>−4.0</td>
<td>/</td>
<td>Below the descent branch of the Walker circulation</td>
</tr>
<tr>
<td></td>
<td></td>
<td></td>
<td></td>
<td>Coldest SST in tropical North Pacific</td>
</tr>
<tr>
<td>E</td>
<td>/</td>
<td>/</td>
<td>+2.0</td>
<td>Warmer SST in tropical North Pacific</td>
</tr>
<tr>
<td>F</td>
<td>/</td>
<td>/</td>
<td>−2.0</td>
<td>Warmer SST in tropical North Pacific</td>
</tr>
</tbody>
</table>

where \( \chi \) and \( \psi \) are, respectively, the velocity potential and streamfunction, which are obtained by solving the following Poisson equations:

\[
\nabla^2 \chi = \nabla \cdot \mathbf{V}
\]

\[
\nabla^2 \psi = k \cdot \nabla \times \mathbf{V}
\]

The Lagrangian interpolation was used to obtain wind fields at pressure levels from those at the model \( \sigma \)-levels. Wind components at grid points below the ground surface were set to zero. Figure 2 shows, in the control run, the time-mean streamfunction \( \psi \) and the rotational wind component \( \mathbf{V}_r \) at 200 mb (a) and 950 mb (b), respectively. The Southern Hemisphere is in late winter and is characterized by strong subtropical jets in the upper troposphere and midlatitude westerlies at the lower level. In the Northern Hemisphere at 200-mb level, troughs over oceans and ridges over continents typify the upper-tropospheric northern summer circulation, with the huge and strong South Asian high over the Tibetan Plateau. At 950 mb, anticyclones dominate the eastern portions of the Atlantic and Pacific oceans. Troughs are observed along the western coasts of the continents. A strong monsoon low is simulated over South Asia. Easterlies prevail over the southern tropics, which deviate northward near the eastern coast of Africa, then cross the equator to reinforce the monsoon southwesterly flow. The R15 model, however, fails to reproduce the tilting and strength of the tropical upper-tropospheric troughs (TUTT). The model TUTTs over the northern Pacific and Atlantic are very weak and relatively symmetric in their longitudinal locations, whereas the observed climatology for August shows a narrow but strong TUTT oriented from northeast to southwest in the subtropical region over each ocean. Also, the subtropical ridges observed along 10° to 15°N over the eastern Pacific, Caribbean, and Atlantic is absent in the model climatology at 200 mb. At 950 mb, the model also fails to reproduce a realistic monsoon trough over the eastern Pacific–central American region, and the mean subtropical high in the Pacific is about 40°N longitude too far west, as compared to observations. Despite these deficiencies, gross features shown in Fig. 3a are in agreement with observations of the June, July, and August (JJA) mean climates (Oort 1983; Wallace 1983).

The lower-level divergent component (Fig. 3b) is characterized by divergence out of the anticyclonic regions over the eastern oceans, and convergence into the regions over the western North Atlantic (WNA), WNP, and northern Africa. Along the equator, \( \mathbf{V}_x \) is directed everywhere from the Southern Hemisphere to the Northern Hemisphere. The distribution of the divergent wind at the upper level (Fig. 3a) resembles the distribution at the lower level (Fig. 3b), except that the polarity of the velocity potential centers and the directions of the wind components are all reversed. The cross-equatorial flow is southward everywhere. The strong 200-mb divergence center over the WNP, however, is at about 140°E, displaced eastward from the low-level convergence center over the Indochina Peninsula by about 40°.

If Fig. 3 is compared to the SST distribution displayed in Fig. 1, one finds that, over cold SST, low velocity potential \( \chi \) is located at the lower level, and high \( \chi \) at the upper level, whereas over warm SST, high \( \chi \) is located at the lower level, and low \( \chi \) at the upper level. In general, therefore, low-level divergence and high-level convergence are located over cold SST area, whereas low-level convergence and high-level divergence are located over warm SST area. To a first-order approximation, the 950-mb \( \chi \) field may then be considered to be qualitatively proportional to the SST \( (T_c) \) distribution, that is,

\[
\chi_{950 \text{ mb}} \propto T_c.
\]

The distributions of the time-mean differences of velocity potential \( \chi \) and divergent velocity \( \mathbf{V}_x \) between the warm and cold SSTA experiments are shown in Fig. 4. A “difference” field is obtained here by subtracting the field of the experiment with cold SSTA from that of the corresponding experiment with warm SSTA. It then presents the amplitude of the field anomaly generated by a given SSTA. For convenience, this difference is also termed as an “anomaly” in the following. A common feature among these three sets of experiments (i.e., A–C, B–D, and E–F) is that warm SSTA corresponds to the low-level convergence and high-level divergence. Intensified ascent must occur over the warm SSTA pool. Let “*” denote the difference; then to a first-order approximation, the anom-
alous velocity potential at 950 mb may be considered to be qualitatively proportional to SST anomalies, that is,

$$x_{950}^* \propto T^*_f.$$  \hspace{1cm} (1b)

Besides the common feature, some differences among these three sets of experiments can also be identified:

1) Although the SST anomalies in the EEP are twice as strong as that over WEP and WNP, the difference in low-level convergence over EEP is only about one-third stronger than that over WEP and comparable with that over WNP.

2) The divergence difference in the upper layer over WNP is the strongest, whereas those over EEP and WNP are comparable.

These results imply that a warm SST anomaly imposed on a warmer SST region can result in stronger anomalously high-level divergence and high-level divergence in the troposphere than equal energy imposed on a colder SST region. This can be used to explain why, in experiments by Keshavamurty, anomalous rainfall in response to the SST located in the western or central equatorial Pacific is much stronger than the response to an SST with the same energy but located in the eastern equatorial Pacific, as we will see later.

The time-mean streamfunction difference and the time-mean rotational wind difference are shown in Fig. 5. For the two equatorial sets of experiments (A and C, and B and D), the anomalous streamfunction field at 950 mb is characterized by a pair of cyclones straddling the equator (Fig. 5b,d). They are over the warm SST region, but shifted westward by about 30° in the EEP case and 20° in the WEP case. At 200 mb, above the low-level anomalous cyclonic dipole, a pair of anticyclones is also straddling the equator. The 200-mb subtropical jet is then intensified on the poleward side of the anomalous anticyclone, particularly in the winter hemisphere (Fig. 5a,c). These differences are statistically significant over most parts of the tropics. The anomalous winds in the EEP case are weak, however, and occur mainly over the equatorial Pacific. On the other hand, in the WEP case, anomalies of rotational wind are much stronger. At 200 mb along the equator, in addition to the anticyclonic dipole, strong anomalous westerlies of 3 to 5 m s⁻¹ prevail over the whole tropics. They circulate around the anticyclone dipole and greatly enhance the subtropical westerly jet, particularly in the winter hemisphere.

Comparing the results obtained in this study to those obtained by Keshavamurty (1982), one finds that similar anomalous flow patterns between the two simu-
lations can be found in the equatorial zone, particularly in the EEP experiments. On the other hand, anomalous flow in the tropical and subtropical areas are different between the two simulations, particularly in the WEP experiments. In the study of Keshavamurty, the anomalous streamfunction in the northern tropics is much stronger than in the southern tropics. In the present study, the anomalous streamfunction is relatively symmetric to the equator, very similar to the solutions of Gill for the case of the heating symmetric about the equator [see Fig. 1 of Gill (1980)].

Comparing Fig. 5 to Fig. 1 of Gill, we find that the dipole pattern of the anomalous streamfunction in the WEP case (Fig. 5b) is close to the corresponding pattern obtained by Gill, whereas the dipole pattern in the EEP case (Fig. 5d) is noticeably shifted westward over the SSTA region. The same shift can also be found in the experiments of Keshavamurty. Gill's model neglects the time-mean flow, whereas in the climate model the 950-mb time-mean easterly in the EEP region is very strong (>5 m s⁻¹). These differences then suggest that advection of vorticity by the background flow may not be entirely unimportant in the vorticity budget in the lower troposphere. Quantitative evaluation of the importance of this advection process requires detailed budget analyses and will not be discussed here.

The response of the streamfunction to SSTAs away from the equator (Fig. 5e,f) is quite different from what we have seen in the equatorial SSTA cases. The symmetric polarity of the anomalous streamfunction over the SSTA region no longer exists. At 950 mb, strong characteristic westerly–southwesterly anomalous rotational wind sweeps through the southeastern part of the region. Weak anticyclonic and very strong cyclonic vorticity are found to its southeast and northwest, respectively. At 200 mb, strong anomalous northeasterly–easterly winds of more than 5 m s⁻¹ flow through the region from the midlatitude region east of the date line to equatorial Africa. Anomalous cyclonic and anticyclonic vorticity are found to the northwest and southeast of the SSTA region. The polarities of the anomalous streamfunction field in the upper troposphere are opposite to those in the lower troposphere, similar to what we have observed in the equatorial SSTA cases.

Many characteristics of the anomalous streamfunction pattern presented in Figs. 5e, f can also be identified from the results of Keshavamurty in the tropical and subtropical regions. As a matter of fact, the anomalous streamfunction in the Northern Hemisphere in the western Pacific SSTA case of Keshavamurty's modeling is very close to the summation of the streamfunction
in our WEP and WNP cases (Figs. 5a,b,e, and f). This is because the latitudinal scale of the SSTA in his study is large and gives both equatorial and tropical forcing as defined in this study. It is also interesting to note that the anomalous 950-mb streamfunction shown in Fig. 5f is rather similar to the solutions of Gill for the case of the heating antisymmetric about the equator (see his Fig. 2a). The latitudinal scale of the heating in Gill’s model is much larger than that in the present study. The aforementioned results then imply that the characteristic response of the tropical circulation to tropical forcing is rather insensitive to the meridional scale of the heating.

The response of the large-scale tropical flow to a warm SSTA can therefore be summarized as follows. At first, lower-level convergence and upper-level divergence are excited over a positive SSTA region. At the equator, an anomalous cyclonic dipole in the lower
layers and anticyclonic dipole in the upper layers are then generated at the equator. Anomalous low-level westerlies and high-level easterlies at the equator are produced over the western and central parts of the SSTA region. On the other hand, SSTA in the WNP generates anomalous westerly–southeasterly flow in the lower troposphere and anomalous northeasterly–easterly flow in the upper troposphere through most parts of the region. An anomalous low at the lower level and an anomalous high at the upper level are located to the northwest of the anomalous SST region.

In the next section we will see that this large-scale flow pattern is very important in producing an anomalous rainfall pattern over the warm SSTA region.

4. Anomalous moisture convergence

The time-mean atmospheric budget of water vapor can be estimated by

\[
MC = - \int_0^\infty \nabla \cdot (\mathbf{V} \mathbf{q}) \, dz = \bar{R} - \bar{E}.
\]
Here the overbar (\(\bar{\cdot}\)) denotes a time mean over a period of 1620 days (the 181st to 1800th model day), \(R\) and \(E\) are precipitation and evaporation rates of the atmosphere, respectively, and MC denotes the vertical integral of moisture convergence. The latent heat release to the atmosphere due to condensation can be calculated from the rate of rainfall \(R\). Equation (2) has been applied to our model output to evaluate the change in the water budget of the atmosphere due to the SST anomalies. It has been shown from the Control experiment that the mean \((R - E)\) in August is negative over the Southern Hemispheric oceans, eastern North Pacific, and central and eastern Atlantic. From there the extra atmospheric moisture obtained from the underlying surface is transferred to precipitation over the WNP, WNA, and the continents (figures not shown). As SST anomalies are introduced in the other experiments, anomalous evaporation and precipitation are observed.
over the warm SST anomaly regions. Usually anomalous evaporation (Fig. 6) is relatively symmetrically distributed and more evaporation is observed over warmer SST regions.

Anomalous rainfall over the SST anomaly region shown in Fig. 7 is about three to four times as large as the anomalous evaporation. As a result, the distribution of anomalous convergence of moisture MC (refer to Fig. 8) resembles that of anomalous rainfall. In other words, greater than normal rainfall over the warm SST region is not simply due to the increase in evaporation, but primarily due to an increase in convergence of water vapor flux. The ratio between these two contributors is about 1:3. Thus, these model results agree with what have been reported from data analyses of warm ENSO events (Cornejo-Garrido and Stone 1977; Ramage and Hori 1981), and from numerical modeling of SW, although the anomaly defined in the present study is not the same as that defined in the others. It is worthwhile to mention that, over WEP and WNP, the precipitation difference and evaporation difference between anomaly and control experiments are comparable and opposite.
for the opposite-signed SSTA, whereas over the EEP, these differences for positive SSTA are considerably larger than for negative SSTA. This is because in the Control experiment (also in the real atmosphere) the evaporation and precipitation over EEP are already very weak. Therefore, results presented in Figs. 6b and 7b are mainly due to the warm SSTA.

Another striking feature revealed by these sets of experiments is the prominent asymmetry of anomalous rainfall over the symmetrical SSTA regions. Increase in rainfall of 5–10 mm day$^{-1}$ is found in the eastern part of the WEP region, in the western part of the EEP region, and in the central and northeastern parts of the WNP region. Although anomalous rainfall in the equatorial SSTA case is biased toward the warmer SST area, in the WNP case anomalous rainfall is biased away from the warmer SST area. This cannot at all be interpreted in terms of the exponential relation between
temperature and saturation water vapor pressure; nor can it be explained by anomalous evaporation since the latter is relatively symmetrically distributed over the SSTA region. Since anomalous rainfall over the warm SSTA region is determined to a large extent by anomalous water vapor pumped onto the region by the anomalous circulation, it will be instructive to study anomalous rainfall by scrutinizing each of the separate factors contributing to the intensification of convergence of water vapor flux over the SSTA region.

Moisture convergence (MC) can be separated into contributions from stationary and transient flows, that is,

\[ MC = - \int_0^\infty \nabla \cdot (\nabla q) \rho dz = (F_q)_s + (F_q)_t \]  

in which

\[ (F_q)_s = - \int_0^\infty \nabla \cdot (\nabla q) \rho dz \]

\[ (F_q)_t = - \int_0^\infty \nabla \cdot (\nabla q') \rho dz \]  

Fig. 7. As in Fig. 6 except for the rainfall difference.
are the convergence of stationary and transient water flux, respectively. Here primes denote the time deviation. Let asterisks denote the difference between two experiments, then from (2) and (3), the water budget equation and its difference between the two experiments can be expressed, respectively, as

\[(Fq)_t + (Fq)_p \approx \bar{R} - \bar{E}\]  \hspace{2cm} (5)

and

\[(Fq)_t^* + (Fq)_p^* \approx (\bar{R} - \bar{E})^*.\]  \hspace{2cm} (6)

In the following analyses, the rainfall \(R\), evaporation \(E\), and the convergence of stationary water flux \((Fq)_t\), were calculated directly from the model output, whereas the convergence of transient water flux \((Fq)_p\), was calculated as a residue from the other terms in (5).

The spatial distributions of each term in (6) for each set of experiments are shown in Fig. 8. Distributions of the anomalous water sink \((R - E)^*\), in general, are similar to those of the anomalous rainfall \(R^*\) shown in Fig. 7, indicating the relatively small contribution of the local evaporation to the changes of precipitation. Results show that, in all cases, contributions of the transient water flux to changes of water budget are dissipated and are of secondary importance. The main balance is between the anomalous convergence of stationary water flux and the water sink/source. This indicates that asymmetry in the distribution of \((R - E)^*\) primarily results from anomalous convergence of stationary water flux. Since horizontal moisture convergence (MC) occurs mainly in the lower troposphere, and since horizontal distributions of MC at each of these lower layers are similar (figures not shown), it is feasible to use water flux at one lower level, for example, 950 mb, in presentation of the vertical integral for the

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**Fig. 8.** Distributions of the differences of water sink \((R - E, \text{ top row}), \) and the vertically integrated convergence of stationary water flux \([(Fq)_t, \text{ middle row}], \) and transient water flux \([(Fq)_p, \text{ bottom row}]\) between the warm and cold SST experiments in the regions of the WEP (column a), EEP (column b), and WNP (column c), respectively. Contour interval is \(2 \times 10^{-8} \text{ s}^{-1}\). Stippling indicates negative values.
purpose of discussions of its spatial distribution. Notice also that for arbitrary quantities $A$ and $B$, there exists the relation

$$(AB)^* = A^*\hat{B} + \hat{A}B^*,$$

where $(\hat{ })$ denotes the average of the quantity in the two experiments, which will be omitted in the following discussions for simplicity. Furthermore, if horizontal wind is separated into its rotational and divergent components, then the difference in water-flux convergence can be written as

\[
\frac{\mathbf{MC}^*}{\epsilon\text{UL}^{-1}} \sim \left[ -\nabla \cdot (\nabla q^*) - \nabla q^* \cdot \nabla q \right]_{50\text{ mb}} = \frac{-\nabla q \cdot \nabla q^* - \nabla q^* \cdot \nabla q}{\epsilon\text{UL}^{-1}}
\]

Let $\epsilon$ be a small number less than unity, $L_x$ and $L_y$ the zonal and meridional scales of the SSTA region, and $Q$ and $U$ the characteristic scale of specific humidity and velocity. Then, from the experimental output we have the following characteristic scaling:

\[
\begin{align*}
L_y & \sim L \\
L_x & \sim \epsilon^{-1}L_y \sim \epsilon^{-1}L \\
\bar{u}_x, \bar{u}_x^\epsilon, \bar{v}_x, \bar{v}_x^\epsilon, \bar{u}_y, \bar{u}_y^\epsilon & \sim U \\
\bar{v}_q, \bar{v}_q^\epsilon & \sim \begin{cases} 
U & \text{(away from the equator)} \\
\epsilon U & \text{(close to the equator)}
\end{cases} \\
q & \sim Q \\
q^* & \sim \epsilon Q
\end{align*}
\]

so that

\[
\begin{align*}
\nabla q, \nabla q^* & \sim \epsilon\text{QL}^{-1} \\
(\nabla x, \nabla y, \nabla z, \nabla \cdot V_x, \nabla \cdot V_y) & \sim UL^{-1} \\
(\nabla q, \nabla \cdot V^* x) & \sim UL^{-1} - \epsilon UL^{-1}.
\end{align*}
\]

By applying (8) and (9) to each term in (7), we then obtain its scaling, which is indicated under the corresponding terms in (7). The leading term in (7) is the term associated with anomalous divergence, that is, $(-q\nabla \cdot V^*_y)$. This term is very important for the increase of anomalous rainfall over the warm SSTA region in WNP. Our main interest here, however, is to study mechanisms that lead to the asymmetric distribution of anomalous rainfall over a given SSTA region. Since this term is directly linked to anomalous rainfall, it cannot be used to explain the causal relation between the two. Other terms on the right-hand side of (7) possess similar magnitude and must be retained for the budget analyses. Near the equator, however, if $V_q^*$ is small as in the WEP case, then the last two terms in (7) can be ignored and the first three terms associated with the divergent wind and its anomaly should basically determine the anomalous convergence of water flux.

By using 950-mb level data, each of the terms on the right-hand side of (7) and their summation have been calculated for each set of experiments. Results are shown in Figs. 9–11, respectively. Terms associated with the divergent wind and its anomaly are shown in the panels (a)–(c), whereas those terms associated with rotational wind and its anomaly are shown in panels (d) and (e). As anticipated, the fourth term $(-q\nabla \cdot V^*_y)$ shown in panel (f) of each figure dominates the total contribution [panel (g)] in each region. For the WEP case (Fig. 9), the rotational wind contribution is small. For the EEP case, however, advection of anomalous moisture by the rotational wind component $-V_q \cdot \nabla q^*$, see Fig. 10d) is large. This is because the equatorial rotational component $V_q^*$ in region EEP is very large (>5 m s$^{-1}$, see Fig. 2b). Strong equatorial easterlies advect anomalous water vapor downstream of the SSTA region, resulting in the westward shifting of anomalous rainfall from the SSTA region. For the WNP case (Fig. 11), as strong anomalous rotational southwesterlies (Fig. 5f) sweep over the warm and wet ocean surface, additional moisture convergence occurs to the northeast of the region. This dominates the process of anomalous water transfer and should contribute to the bias of anomalous rainfall away from the warmest SST region. In general, the divergent wind and its anomaly near the equator contribute both to enhance convergence of water flux over the northern parts of the SSTA region and to enhance divergence of water flux over its southern parts. Anomalous rainfall is then biased toward the Northern Hemisphere. All results suggest that advections of water vapor and its anomaly by the basic flow and its anomaly in the lower tropical troposphere are important in determining the anomalous rainfall pattern.

We now verify the SW postulation by using the experiment results presented above. First, the patterns of low-level flow difference between the warm and cold SSTA experiment in the EEP region (Figs. 4d and 5d) are very similar to those between composite El Niño and La Niña events, which were simulated by the same model as used in this study (Wu and Lau 1992). Second, despite the difference in seasons, near the equator wind anomaly pattern at 700 mb, corresponding to El Niño SST anomaly obtained by SW (see their Fig. 17) is similar to that at 950 mb obtained from the EEP anomaly experiments, that is, westfl to the west of the positive SSTA region increase in both cases. These anomalous westerlies do carry more moisture eastward and contribute to the increase of moisture convergence to the west of the SSTA region (Figs. 10b,e) as expected by SW. This contribution, however, is relatively small compared to advection of anomalous moisture by the climatic mean flow, particularly by the mean rotational flow, that is, $-V_q \cdot \nabla q^*$ (Fig. 10d). Therefore, the anomalous low-level westerlies cannot play the leading
role in accounting for the bias of precipitation, at least in the EEP case.

The contribution of each term in (7) to the rainfall pattern can be interpreted in terms of the distribution of SST and its anomaly. Figure 12 shows the distributions of the simulated mean specific humidity at 950 mb in the control experiment. Comparing this figure to the August SST distribution (Fig. 1), one finds good correspondence of moist air to warm SST; the area with specific humidity exceeding 0.012 is above the area where SST is warmer than 27°C. This feature can be found at other model levels below the mean tropical inversion (usually between 850 and 700 mb, see Lindzen and Nigam 1987) and is indeed in good agreement with the observational analyses provided by Oort (1983, Figs. A16, F16, 93, 110, 115, 117). This indicates that the 950-mb specific humidity \( q \) can be well parameterized by SST \( (T_s) \) (see also Neelin and Held 1987). The differences of the 950-mb specific humidity between the warm and cold SSTA case \( (q^*) \) for each set of experiments are shown in Fig. 13. In each set, positive \( q^* \) is also above the warm SSTA \( (T_s^*) \) region. Then, to a first approximation, 950-mb specific humidity and its anomaly can be expressed in terms of \( (T_s) \) and \( (T_s^*) \), respectively, that is,

\[
q \sim T_s, \quad q^* \sim T_s^*,
\]

so that the remaining terms in (7) can be expressed as

\[
MC^* \propto [-\nabla \cdot (Vq)]_{950 \text{mb}} \sim -V_{\text{f}} \cdot \nabla T_s^* - V_{\text{f}} \cdot \nabla T_s - T_s^* \nabla \cdot V_s - V_{\text{f}} \cdot \nabla T_s^* - V_{\text{f}} \cdot \nabla T_s.
\]

This means that the difference in convergence of water flux can be explained in terms of the distribution of

\[
(VX)Q^* \text{ EXPC. 950 MB}
\]

\[
(VP)Q^* \text{ EXPC. 950 MB}
\]

\[
(DVQ) \text{ EXPC. 950 MB}
\]
SST and the large-scale flow and their anomalies. Thus, the schematic interpretation for each term can be described as follows:

1) Advection of anomalous moisture by the divergent wind ($\sim -V_x \cdot \nabla T^*$).

In the boreal summer, as warm SST moves to north of the equator, the divergent wind $V_x$ at the equator is directed northward (Fig. 3b). For cases of equatorial SSTAs, $V_x$ is then downgradient of anomalous moisture in the area north of the equator, resulting in the bias of anomalous rainfall toward the warm SST area (as shown in the schematic diagram Fig. 14a).

2) Advection of water vapor by the anomalous divergent wind ($\sim -V_x \cdot \nabla T^*$).

Anomalous divergent wind is directed inward toward warm SSTAs (Figs. 4b,d, and f). The gradient of SST in August is northward along the equator (Fig. 1). For the equatorial SSTAs, such as in the WEP and EEP cases, downgradient moisture flux then occurs to the north of the SSTAs region, resulting in anomalous rainfall being biased toward warmer SST, as shown schematically in Fig. 14b.

Assuming the 950-mb divergence wind in the tropics is qualitatively proportional to the SST gradient [refer to Eq. (1a) and Fig. 3. See also Lindzen and Nigam 1987],

$$(V_x)_{950 \text{ mb}} \propto \nabla T^*$$

and the anomalous divergent wind is qualitatively proportional to the gradient of SSTAs [refer to Eq. (1b) and Fig. 4]

$$(V_x^*_{950 \text{ mb}} \propto \nabla T^*$$

then the two terms discussed above are both negatively proportional to the scalar product of the gradient of SST and of its anomaly SSTAs, that is,

$$-V_x \cdot \nabla q^* - V_x^* \cdot \nabla q \sim -V_x \cdot \nabla T^* - V_x^* \cdot \nabla T^*$$

$$\propto -\nabla T^* \cdot \nabla T^*.$$  \hspace{1cm} (11)
It is positive only over the region where warmer SSTA overlaps with warmer SST. Anomalalous rainfall should then be biased toward warmer SST.

3) Impact of large-scale divergence ($\sim T_x^{\uparrow} \nabla \cdot V_x$).
   Since
   \[
   \nabla \cdot V_x = \nabla^2 \chi \propto -\chi,
   \]
   according to (1a), we then have
   \[
   -T_x^{\uparrow} \nabla \cdot V_x \propto T_x^{\uparrow} \cdot T_y.
   \]
   This also indicates that anomalous convergence of water flux and rainfall should be biased toward warm SST. Therefore, the first three terms on the right-hand side of (10), which are associated with large-scale divergent flow and its anomaly, have the same effects: they all cause anomalous rainfall to be biased toward warm sea surface temperature.

4) Advection of anomalous moisture by the rotational wind ($\sim -V_y^{\uparrow} \cdot \nabla T_y$).
   As previously discussed, this term can cause anomalous rainfall to occur downstream of the high anomalous moisture region. It is, however, small unless the basic flow is sufficiently strong as in the EEP case, in which case it results in shifting anomalous rainfall toward the west (Fig. 10d) where SST is warmer than in the east (see Fig. 14c).

5) Advection of water vapor by the anomalous rotational wind ($\sim -V_x^{\uparrow} \cdot \nabla T_y$).
   Since, near the equator in the lower layer, the anomalous zonal component of the rotational wind is also directed inward over the warm SSTA region (Fig. 5), it should also cause anomalous rainfall to be biased along the equator toward warm SST. This effect, however, is small. Away from the equator this term becomes larger and comparable to the other terms listed above. In the WNP case, since the SSTA region is lo-
cated above the area of maximum SST, and the strong anomalous flow (~5 m s⁻¹) is northeastward, large downgradient transfer of moist air occurs to the northeast of the warmest SST area (Fig. 9d), resulting in large anomalous precipitation there (Fig. 14d). It is this anomalous rotational wind component that results in the shifting of anomalous rainfall away from the warmest SST center toward the northeast.

Overall analyses show that water vapor advection is very important in determining the rainfall pattern over the warm SSTA areas. In this aspect, there appears to be a common rule: anomalous rainfall associated with warm SSTAs events occurs in the region where the basic flow is downgradient of the SST field, or the anomalous flow is downgradient of the SST field.

5. Discussion and conclusions

In response to warm tropical SSTAs, anomalous low-level convergence and high-level divergence occur in the troposphere over the SSTA region. In the case of the equatorial SSTAs, the anomalous streamfunction is characterized by a cyclonic dipole in the lower troposphere and an anticyclonic dipole in the upper troposphere straddling the equator over the SSTA region.

In the atmospheric response to a tropical SSTA away from the equator, as in the WNP case, such symmetrical dipoles disappear. A large anomalous cyclone in the lower troposphere and an anomalous anticyclone in the upper atmosphere are excited over the warm SSTA region, resulting in strong anomalous westerly–southerly winds at lower levels and strong northeasterly–easterly winds at upper levels over most parts of this region.

Anomalous rainfall is asymmetrically distributed over a symmetric warm SSTAs region. The exponential relation between saturation water vapor pressure and temperature can explain only a minor part of such a distribution. The contribution from anomalous evaporation is relatively symmetric, and that from transient eddy flux of moisture is dissipated. They are both of secondary importance. The majority of the anomalous rainfall is from anomalous convergence of the stationary water flux. Therefore, the asymmetry in the distribution of anomalous rainfall should be understood as a result of the large-scale circulation and its changes related to the given SSTAs. Bias of the anomalous rainfall could be toward or away from the warm SST region, depending on the relative distribution of large-scale flow and SST, and their anomalies: anomalous rainfall occurs where the mean flow is down the SSTAs gradient.
or the anomalous flow is down the SST gradient. Generally, the large-scale divergent wind and its anomaly resulting from SST act together to shift the maximum anomalous rainfall toward warm SST regions. It is this effect that explains why anomalous rainfall near the equator associated with a warm ENSO episode is always biased toward the west of the SSTA region and the summer hemisphere. Tropical rotational wind tends to move anomalous rainfall downstream of the SST region and is important in the case of warm SST in the EEP. Advection of warm and moist air by the anomalous rotational wind is also important in determining the distribution of anomalous rainfall. Its effect at and near the equator, although smaller compared to the contribution from the divergent wind, is also to shift anomalous rainfall toward warmer SST. Away from the equator this term becomes larger, which can result in a prominent bias of enhanced rainfall away from the warmer SST region, as observed in the WNP experiments.

Anomalous rainfall is not only the result, however, but also the cause of the atmospheric circulation and its anomaly. Once rainfall bias occurs, the anomalous divergent wind pattern and the rotational wind pattern will also shift toward the anomalous rainfall region. The whole feedback procedure between anomalous
rainfall and anomalous flow will continue until both diabatic heating due to latent-heat release is balanced by adiabatic cooling of the ascending air, and generation of vorticity is roughly balanced by dissipation and the spindown process. Therefore, each of the terms discussed above should be better understood as an initiating mechanism responsible for the occurrence of the asymmetry of rainfall over a symmetric SSTA region.

It should also be pointed out that, although the magnitudes of SSTA specified for the EEP experiments are not far from corresponding observations during extreme ENSO events, those specified for the WEP and WNP experiments are quite large compared to observations. Also, the idealized SSTA in the WNP is not the shape that is normally observed over the region due to the prominence of the Kuroshio Current in this region. Therefore, the present study should be considered as a study of sensitivity of the atmosphere in response to SST anomalies located in different parts of the Northern Pacific. The conclusions obtained in section 4 that contribution of the transient water flux to the changes of water budget are dissipated and secondary was based on this very crude low-resolution model with SSTA fixed. In the real atmosphere, mesoscale and small synoptic-scale transient disturbances develop frequently in the tropics, and SSTA varies spatially as well as temporally. These are hardly depicted by such model experiments. Therefore, to what extent this conclusion is applicable to the atmosphere needs to be studied further by using observation data.

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