Tropical Cyclone Environments over the Northeastern and Northwestern Pacific Based on ERA-15 Analyses

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(Manuscript received 19 November 1999, in final form 5 February 2001)

ABSTRACT

This study uses a 1° × 1° lat-long dataset, extracted from ECMWF reanalyses for the 15-yr period 1979–93 (ERA-15), to composite environmental characteristics and flow features in the vicinity of named tropical cyclones (TCs) in the eastern and western North Pacific Ocean basins. Tropical cyclones are partitioned into one of four classifications as they pass by selected locations along the axes of maximum frequency and TC tracks: weak (W), strong (S), intensifying (I), or dissipating (D).

Results of this study show that peak values of rising motion, within the same classification, are greater for TC composites in the western Pacific than for those in the eastern Pacific. The level of maximum rising motion was at or above the 500-hPa level for all locations and classifications, except for the Ss at our northernmost point (25°N, 130°E) in the western Pacific. Their maximum upward motion occurred at 700 hPa. It is also found that the latter systems, contrary to all other points, were located in a region of minimum large-scale convective instability. As one cause of stabilization, large-scale advection of drier air from the East China Sea into the western and southern vicinity of the composite storm is identified.

Anomalies of precipitable water (PW) were found to be related to the intensity of the storm, but not to the amount of available climatological “background” moisture. In the eastern Pacific, the monsoon-type southwesterly moisture flow across 10°N was much stronger and deeper for Is than for ws at point B (17.5°N, 112.5°W). On the other hand, the eastern Pacific ws were more impacted by westward transports originating off the Central American coast. When integrated from the surface to 700 hPa, the net effect was a change in the direction of moisture transport vectors and, therefore, in the major source region. Such a distinct directional change between classifications was exceptional to point B, and was not found for the three points in the western Pacific and South China Sea.

Finally, based on ERA-15 model-produced rain rates, it is found that, in the western Pacific, total precipitation rates for this study were compatible with those of earlier research by W. M. Frank, who used large-scale data. The fraction of ERA-15 stratiform precipitation to the total precipitation varies from 25% to 47% in composite samples used here. The representation of convective and stratiform rain in the ERA-15 model obviously favors the former when the systems are stronger and have a more intense and broader secondary circulation.

1. Introduction

Tropical cyclones are one of the world’s most feared and deadly circulation systems. Over the globe, six (seven, in some references) oceanic regions are generally recognized as being basins for tropical cyclone development and preferred tracks (e.g., Gray 1979, 1985; Elsberry et al. 1985; Frank 1985; Neumann 1993; McBride 1995; Fink and Speth 1998). Of these, the northwestern Pacific basin has, by far, the world’s greatest number of named tropical cyclones (TCs), while the northeastern Pacific basin has either the second or third highest number, depending on how one partitions the southeast Indian Ocean–Australian–southwest Pacific Ocean region [e.g., Neumann (1993) separates it into two areas, while Fink and Speth (1998) consider it as one]. In either case, the northeastern Pacific is one of the world’s most active TC basins. As will be shown later, the kinematic, thermodynamic, and moisture environments associated with TCs in these two basins can be quite different, especially when one partitions TCs according to their intensities. Thus, the overall goal of this paper is to document the environmental conditions that exist in the vicinity of TCs in the eastern and western North Pacific as a function of intensity. To accomplish this goal, we use a fine resolution (1° × 1°) dataset based on the European Centre for Medium-Range
Weather Forecasts (ECMWF) reanalyses for the 15-yr period 1979–93 (ERA-15). Details concerning this dataset, as well as our TC track dataset, are discussed in section 4. The rationale for our study is given in section 3. A few remarks about each basin’s climatology are provided in section 2. Results and discussion are presented in section 5, followed by a summary and concluding remarks in section 6.

2. Basin climatologies

In the two basins being investigated, the climatological settings in which TCs are located are quite different during their respective TC seasons (e.g., McBride 1995). To begin with, the sea surface temperatures (SSTs) in the northwestern tropical Pacific basin are among the warmest in the world, especially in the June–November period when values of 29°–30°C are common (Vincent and Schrage 1995). This plays an important role, through air–sea interaction, in accounting for the greatest annual total of TCs being observed in that region of the globe. The northeastern Pacific basin, on the other hand, has cooler SSTs, although a compact region of waters approaching 29°C occurs near the Central American coast in the summer and early fall period (Elsberry et al. 1985).

The low-level flow over the western basin generally consists of weak easterlies from June to November (Vincent and Schrage 1994), with the monsoon trough along or just south of the axis of maximum TC frequency. Briegel and Frank (1997), for example, note that nearly 75% of all TCs in 1988 and 1989 formed in the monsoon trough. Harr and Elsberry (1995) also note the importance of the monsoon trough in western Pacific TC formation. In contrast, the eastern Pacific basin is dominated by moderate easterly flow at low levels from June to October, with trade wind convergence along about 10°N, but mainly west of 120°W.

The upper-level flow over each basin is also quite different. At 200 hPa, the western basin is dominated by moderate east-northeasterly flow from June to November along the axis of maximum frequency (Vincent and Schrage 1994), while the corresponding axis in the eastern basin is located just north of an east–west zone of maximum divergence in the 200-hPa easterlies. The eastern Pacific divergence zone lies approximately along the intertropical convergence zone (ITCZ), which is located at about 10°N, and is most intense in June–October (Gruber et al. 1986; Vincent 1998). As will be shown in our results, TCs in this basin are often distinct from the ITCZ. In the western basin, however, the relationship between the upper-level flow, the ITCZ, and the TCs is not as clear.

3. Rationale for present study

In our study, statistics and results are compiled for a 5- (6-) month period, June–October (Jun–Nov), which represents the eastern (western) Pacific basin’s primary TC season. Figure 1 shows the frequency distribution in each basin for all named cyclones during the 15-yr period 1979–93. Frequencies were based on the total number of TCs compiled over 2.5° × 2.5° lat–long boxes and illustrate, together with the tracks shown in Fig. 2, that although the greatest number of TCs occurs in the western Pacific, TCs in the eastern Pacific are more concentrated in space. Similar results appear in publications by other authors (e.g., Elsberry et al. 1985; Gray 1985; Miller et al. 1988; Fink and Speth 1998). Also shown in Figs. 1 and 2 are the locations of three grid points, A, B, and C, in the eastern Pacific and four grid points, D, E, F, and G, in the western Pacific. These points, which lie along the respective axes of maximum TC frequency and preferred tracks, are the ones for which detailed examinations were performed in the present study.

One aspect of our study that is different from some previous studies is we partition TCs into one of four classifications according to their intensity or intensity changes (based on maximum sustained surface wind speeds given in 5-kt intervals) as they passed through a 5° × 5° lat–long box centered on each of the seven grid points shown in Figs. 1 and 2. Traditionally, there have been three criteria used to classify TCs: 1) intensity, which depends either on maximum sustained surface wind speeds or minimum central sea level pressure;
2) strength, which depends on outer core wind speeds averaged over a radial distance from the eye to about 2.5° from the eye; and 3) characteristics of the eye. Most studies (e.g., Gray 1975; Neumann 1993; Fink and Speth 1998; Cecil and Zipser 1999) have used intensity based on maximum sustained surface wind speeds to partition TCs into different classes. We also opted to use criterion 1 and selected one of the historical tropical cyclone track datasets (see section 4) to stratify TCs by intensity. The studies that have used strength or eye characteristics to classify TCs are generally based on aircraft data (e.g., Gray and Shea 1973; Frank 1977; Marks and Houze 1987; Weatherford and Gray 1988a,b; Raymond et al. 1998).

The four classifications used in the present study have not been applied previously to compare and contrast TCs in the two North Pacific basins. Our classifications are weak, if the TC’s maximum wind speed (when it was in one of our 5° × 5° boxes) remained nearly constant and never reached that required for hurricane or typhoon strength (i.e., its winds were always ≤60 kt); strong, if its maximum wind speed remained at hurricane or typhoon strength throughout its history within a box (i.e., ≥65 kt); intensifying, if its strength changed from a lower category to a higher category (i.e., from tropical depression (≤30 kt) to tropical storm (35 ≤ TS ≤ 60 kt]), tropical storm to hurricane/typhoon (≥65 kt), or hurricane/typhoon to superhurricane (≥100 kt)/super typhoon (≥135 kt)); and dissipating, if its maximum winds decreased to the point where the TC was downgraded by one or more categories. Once the individual TCs were classified, parameters were composited about each of the points, A–G, for each classification. The method for accomplishing this procedure is described in section 4. Table 1 shows the number of TCs used to construct composites for each class at each grid point.

Another novel aspect of the present study is the use of a relatively homogeneous reanalysis dataset (ERA-15) to obtain the composite structure of several kinematic, thermodynamic, and moisture variables associated with the environments of TCs. In particular, in contrast to most other studies, we include maps of moisture transports at several lower-tropospheric levels, area-averaged profiles of moisture convergence in the lower troposphere, maps of convective instability, and maps of precipitable water anomalies. We also provide estimates of convective and stratiform precipitation rates for each TC classification at all of our compositing points.

4. Data sources and methodology

a. Data

Two datasets were utilized in this study, TC track data and ERA-15 upper air and surface analyses. The TC data that we used to locate and track storms, as well as to monitor TC intensities, were obtained directly from the National Climatic Data Center revised tropical cyclone position dataset archived at the National Center for Atmospheric Research. Altogether, there are six datasets, one for each of the world’s TC basins. We extracted the datasets for our two regions for the 15-yr period 1979–93. The track data contain the following information for each TC at 6-h increments: latitude–longitude position to the nearest 1/10°, maximum sustained surface wind speed in increments of 5 kt that is mostly based on satellite estimates, and minimum central pressure for the most recent eastern Pacific TCs. A classification based on overall peak intensity, ranging

![Figure 2](http://journals.ametsoc.org/doi/pdf/10.1175/1520-0493(2001)129<1928:TCEOTN>2.0.CO;2)

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from tropical depression to hurricane (supertyphoon) in the eastern (western) Pacific, is also given.

The ERA-15 dataset used to produce our individual TC, as well as composite results, was obtained through ECMWF. Uninitialized pressure level analyses of several variables at the 10 standard pressure levels, plus 3 additional levels, 925, 775, and 600 hPa, were transformed from spherical harmonics space onto a $1^\circ \times 1^\circ$ lat–long grid. The ERA-15 surface variables, including total precipitable water, were interpolated from the Gaussian grid to increments of $1^\circ \times 1^\circ$. As for the TC track data, the ERA-15 gridded values were available four times daily at 0000, 0600, 1200, and 1800 UTC. Details concerning the ERA-15 dataset can be found in Gibson et al. (1997) and Uppala (1997).

It appears the ERA-15 dataset can resolve the larger-scale features of TCs in the western Pacific that are satisfactory for the purpose of this paper. For example, Serrano (1997) states that the “hit” rate for capturing TCs in the ERA-15, based on relative vorticity at 850 hPa, is 93%, with a mean positional error of 126 km, or approximately $1^\circ$ latitude. This is at or less than the distance that each TC in our study was from the grid point about which composites were compiled. In the eastern Pacific, however, the hit rate (61%) and mean positional error (180 km) are not as good. Serrano suggests that the inferior quality of the dataset in the eastern Pacific is due to poorer conventional and satellite data coverage and the fact that TCs generally are smaller in diameter than those in the western Pacific. Nevertheless, in a recently completed study (Pieper 2000), in which the large-scale characteristics of 11 individual intensifying TCs at point B were examined, we found that the center of the cyclonic circulation at 850 hPa was within $2^\circ$ of the TC track data location for all but one of the storms. This suggests that the ERA-15 dataset also performs satisfactorily in the eastern Pacific for our purpose. Also, the results, shown in section 5, support our contention that ERA-15 can be used to investigate the environments of TC composites in the northeastern and northwestern Pacific basins. In addition, several studies have shown that extensive insight into the behavior of TCs can be achieved when large-scale data are composited (e.g., Frank 1977; Gray 1979; McBride and Zehr 1981; Briegel and Frank 1997; Walsh 1997).

While the composite approach reduces random and sampling errors inherent to the analysis, the problem of large storm-to-storm variability can be partially overcome by stratifying storms with similar characteristics (Frank 1977). The composite technique we used to arrive at our results follows Frank’s argument and will be described now.

b. Methodology

As noted earlier, maps of TC frequency and tracks were used to identify seven grid points, A–G, that were located along axes of maximum TC activity. A list was compiled from the previously described TC track data that contained all the TCs passing through a $5^\circ \times 5^\circ$ box shown (centered on point B, large dot) during 1980. All TCs are propagating toward the northwest. Individual cyclones are Blas (B), Celia (C), Estelle (E), Frank (F), Georgette (G), Isis (I), Javier (J), Kay (K), and Madeline (M). The TCs selected for compositing were C, G, and J. See section 4b for details.

FIG. 3. Tracks of the named TCs that passed through the $5^\circ \times 5^\circ$ box shown (centered on point B, large dot) during 1980. All TCs are propagating toward the northwest. Individual cyclones are Blas (B), Celia (C), Estelle (E), Frank (F), Georgette (G), Isis (I), Javier (J), Kay (K), and Madeline (M). The TCs selected for compositing were C, G, and J. See section 4b for details.
Table 1, at least six individual TCs had to meet the criteria stated above before cases were listed and, subsequently, composited. It is worth noting that the proximity of the centers of the individual storms to their respective grid point, as well as the exclusion of rapidly moving cyclones and the stratification of the TCs according to their intensity (or intensity changes), are all factors that act to reduce interstorm variability in the composite ensemble.

It is seen from the description above that our compositing scheme is centered on a fixed point, whereas most studies either move the individual circulation patterns about each TC’s center to a common location, or else use a Lagrangian approach of moving with the TC center until a certain stage of development is achieved. We selected our scheme because it is better suited for the objectives of our study. For example, geographical features have an impact on the behavior and characteristics of TC environments in both basins. Our compositing scheme allows us to preserve the geography, while the other two approaches noted above do not. Also, our scheme allows us to focus on the most interesting and significant locales (e.g., points B and D and where maximum frequencies occur in each basin and points E and G are where large numbers of TCs in the western Pacific exhibit or do not exhibit recurvature).

5. Results and discussion

The horizontal and vertical distributions of several variables were computed about each of the seven points for the classifications listed in Table 1. To conserve space, results are presented at only four of the seven points: B, D, E, and G (see Figs. 1 and 2). The rationale for selecting these four points was noted at the end of section 4. First, we will compare the results at point B to those at point D. Table 1 shows that the only classifications that can be compared are W and I. Although point B had 13 strong cases, point D had only 5; thus, composites for this classification were not compiled at the latter point.

a. Results at points B and D

We start with vertical motion, $v$. This variable, particularly its vertical distribution, is strongly dependent on the ERA-15 model physics, as are all variables related to mass redistribution. In the case of our TC composites values have been modified through the assimilation of observations into the 6-h first guess. We checked the difference between first guess and uninitialized profiles of $v$ (not shown) and found that the shapes were similar, with the level of maximum rising motion occasionally varying by one, but no more than two, analysis levels. Figure 4 shows several interesting features concerning vertical profiles of $v$. The upper two panels depict profiles of the individual TCs that were used to construct composites for the Ws and Is at point D. Each profile was derived by computing an average over a $5^\circ \times 5^\circ$ lat–long box; thus, it is representative of the environmental flow surrounding the point. The composite profiles for Ws and Is at point D are similar in shape, but the I composite shows slightly greater upward motion at all levels than the W composite (Fig. 4c). Also, the level of maximum rising air is a little higher for the former (350 hPa) than the latter (400 hPa). As expected, there is considerable variation among the individual $v$ profiles at point D (Figs. 4a,b), especially with regard to magnitudes; however, the level of maximum rising motion is reasonably consistent, particularly for the Ws. For example, 7 of 8 Ws have maxima within 100 hPa of the composite maximum. This statistic for the Is at point D is 10 of 12.

The lower two panels of Fig. 4 show the composite profiles of vertical motion for the Ws and Is at point B and D, together with their climatologies. We computed climatology by averaging over a 31-day period, from 15 days before to 15 days after each TC passed by its compositing point. In this manner, we removed the possibility of a seasonal bias being introduced into our results. Also shown in Figs. 4c and 4d is the mean deviation from the composite average that was calculated and plotted separately for all values greater and less than the average, taking into account biases in the distributions (cf. Fig. 4b). Henceforth, this measure of the level-by-level variability about the composite profile will be shown with all vertical profiles.

The western Pacific composites are typical of deep convective systems over warm oceans and compare favorably to those of Gray (1979) and McBride (1981) for developing and mature typhoons. As will be seen later, the $v$ composite profile for strong systems at point E in the western Pacific is considerably different. The climatology profiles show that weak rising motion, of approximately equal strength, occurs for the Ws and Is at point D, whereas at point B, these profiles are near zero at all levels. The pattern at point D is, most likely, a reflection of the presence of the rising branch of the Pacific Walker circulation. Figure 4 not only illustrates that for the vertical motion profiles interstorm variability is large, but that the usefulness of the mean left and right deviation as a variability measure is limited in cases of a clustering of storms. Such a clustering is evident in the 500–250-hPa layer for I storms 3, 4, 6, 7, and 12 at point D (Fig. 4b). Fortunately, interstorm variability was less for all other parameters, and values were more equally distributed over the range of observed values. However, due to the small sample sizes, biases can be large, and an asymmetric variability measure is appropriate.

To provide an example of the horizontal distributions of vertical motion, composite maps of $v$ at 500 hPa are shown for the Is surrounding points B and D (Fig. 5). The W composite patterns were similar, with some changes in the magnitudes of the maximum values; thus, they are not shown. Upward maxima are located near
the center of each compositing point and, as expected, the peak value near point D is greater than that in the vicinity of point B, by almost a factor of 2. The pattern for the B composite shows an axis of maximum rising motion along and just north of 10°N. This axis denotes the ITCZ, which is known to be very active during the summer and fall seasons at this latitude (e.g., Gruber et al. 1986; Vincent 1998). Note that the TC center, near point B, has a region of maximum upward motion that is distinctly separate from the ITCZ. It is common for TCs in the eastern Pacific basin to “break away” from the ITCZ (Fuell 1997). In the western Pacific, however, there is no resemblance of the ITCZ in the vertical motion pattern, probably because the ITCZ meanders more during the longer TC season that occurs there.

Next, we examine the stability of the atmosphere in the vicinity of points B and D. Two measures are used, vertical profiles of equivalent potential temperature (EPT), and the difference between EPT in the boundary layer and saturated EPT in the middle troposphere. Vertical profiles of EPT indicate the degree of “potential moist instability” in the lower troposphere and are, therefore, widely used in the relevant literature (e.g., Gray 1977, 1979; McBride 1981). However, this criterion applies only to the stability of layers if they are lifted to saturation. Therefore, we used the second measure as an indicator of the buoyancy that rising saturated parcels, originating from the subcloud layer, would experience in the middle troposphere (Holton 1992). There has been occasional reference to the first measure in the
literature with respect to large-scale patterns associated with TCs [e.g., Gray (1968), for the western Pacific; Raymond et al. (1998), for the eastern Pacific] but, to our knowledge, no studies of the kind presented here have been published.

The top two panels in Fig. 6 show composite profiles of EPT for Ws and Is at points B and D, together with their climatological profiles. As for vertical motion, these profiles were obtained by averaging over 5° × 5° boxes centered on the compositing points. Each profile has a similar structure, with maximum values at the surface and minima at 600 hPa. As expected, the Is at the western Pacific point show the greatest conditional instability in the boundary layer (i.e., 1000–850 hPa). The individual profiles in the eastern Pacific (not shown) exhibited little variation about the composite mean, regardless of classification (Fig. 6b). In the western Pacific, the individual profiles for both Ws and Is showed even less variability, except in the lower layers (Fig. 6a). Finally, it is seen that, compared to our composites, the climatological profiles centered on points B and D are 3°–5°C less at all levels above 925 (Figs. 6a,b). The shape of the climatological profile at point D is similar to those for the W and I profiles. Moreover, it is nearly identical to Gray’s (1968) profile for TC environments in the northwest Pacific basin.

The bottom two panels show maps of EPT at 925 hPa minus saturated EPT at 600 hPa for the Is surrounding points B and D. Based on the available levels from our ERA-15 dataset, we selected 925 hPa as the most representative level for unsaturated boundary layer air, and we chose 600 hPa as our upper level because it is where minimum values of environmental saturated EPT are reached and rising air from the boundary layer has certainly become saturated. Our results (Figs. 6c,d) infer that the regions in the eastern and western Pacific that encompass the greatest potential for convective instability are centered on our I composites at points B and D. They also show that the western Pacific Is are located in a much more convectively unstable environment than are the Is in the eastern Pacific.

The last set of variables we discuss is related to the moisture environments encompassing points B and D. We begin with column precipitable water (PW). Composite maps of PW anomalies are shown in Fig. 7 for the Ws and Is in the vicinity of points B and D. Anomalies were calculated by subtracting climatology from the total PW values. We examined both the total and anomaly fields and decided to use the anomaly ones. The total fields (not shown) exhibited the expected pattern that maximum values occur near or at the centers of TC composites, values at point D were greater than those at point B, and values for the Is exceeded those for the Ws in their respective basins. The anomaly fields (Fig. 7) show the same patterns, with one important exception. The peak values at point B are comparable to those at point D for both the I and the W classifications. Thus, even though the western Pacific TCs at the point of maximum frequency have a higher moisture content than the eastern Pacific TCs, the eastern Pacific storms concentrate a higher percentage of the available moisture in their vicinity. Note that the eastern Pacific contains negative PW anomalies near the latitude of the ITCZ, (cf. Figs. 5b and 7d), suggesting that compensating subsidence to the south of the TC suppresses convection at the latitude of the ITCZ, and the air dries out in this region.

The next three figures show lower-tropospheric moisture transports in the eastern and western Pacific, based on calculations at the analysis levels of 1000, 925, 850, 775, and 700 hPa. It is important to examine the large-scale advection of water vapor at these levels since it may affect the outer rainbands. In addition, midlevel advection of dry air can influence the moisture distribution at the uppermost level (700 hPa) and, thus, contribute to intensity changes of the storm. It must be emphasized, however, that the moisture source for the inner core (e.g., eyewall) convection of TCs is supplied by the ocean through strong air–sea interactions that are certainly not well resolved in the ERA-15 model. The composite transport vectors for the Ws and Is at 925 and 700 hPa are illustrated in Fig. 8 for the region surrounding point B. The patterns at these two levels are sufficient to show the primary characteristics of the environmental flow that impact on the TCs composited about point B. Figure 8 shows that the climatological southeast trade winds transport moisture across the equator at 925 hPa for both Ws and Is. This flow turns clockwise and makes its way across 10°N into the TC environment (Figs. 8c,d). The southwest flow across 10°N is much stronger for Is than for Ws. By the time the 700-hPa level is reached, the southeast trade winds have vanished. At 700 hPa, only a small poleward component to the transport across 10°N exists for the Ws, and it appears to have little impact on the TC environment (Fig. 8a). For the Is, however, the effect of poleward moving air across 10°N is much greater (Fig. 8b). Thus, our results suggest that the Is at point B are associated with a deeper (not shown) and stronger mon-
soonal flow from the southwest. The latter fact is clearly evident from Figs. 9a and 9c, which show the north–south transport across 10°N in the 1000–700-hPa layer. In contrast, there is a significant westward transport into point B across 100°W at all levels from 1000 to 700 hPa for the W composite. The more intense westward moisture flow for Ws, which maximizes at 850 hPa and which is twice as strong as for Is, is also clearly depicted in the transport across 100°W (Figs. 9b,d).

A selection of maps of moisture transports, composited about point D, are shown in Fig. 10. As for point B, maps of transports were compiled at all levels and for both classifications, as well as for climatology; to conserve space, only four maps that help make the main points are shown. The latitude and longitude lines that appeared to be most representative of whether or not moist air would impact on the environment of TCs at point D were 5°N for poleward transports and 120°E for eastward transports. Our results show that both the W and I systems received low-level moisture that was transported across 5°N. The transport was greatest from the surface to 850 hPa, and considerably smaller from 850 to 700 hPa (Figs. 10a, 10b, and 10d). The transport across 120°E only impacted on the W composite at levels below 700 hPa. Figure 10a shows that none of the flow across 120°E at 700 hPa makes its way into point D’s environment. For the Is, however, the transport across 120°E impacts on the TCs at all levels (Figs. 10c, 10d).
The reason for showing the climatological map at 700 hPa (Fig. 10c) is to emphasize the changes to the mean large-scale moisture fluxes across 5°N and 120°E that are occurring when weak and intensifying TCs are present at point D in the Philippine Sea. We now use the transport vectors discussed above to compute profiles of moisture convergence in the lower troposphere. This quantity is, especially on the basis of a single grid point, governed by the ERA-15 model first guess. Nevertheless, we feel that a rough estimate of the vertical distribution of the larger-scale low-level moisture convergence for our TC composites is obtained by averaging over a 5° × 5° lat–long box. To the best of our knowledge, very few, if any, estimates of low-level moisture convergence in the environment of TCs appear in the literature.

Our profiles for the W and I composites at points B and D are shown in Fig. 11, together with their climatologies. At point B, it is seen that the Ws only have moisture convergence in the lowest layer. Above 900 hPa, weak divergence occurs up to 700 hPa. The Is at point B, as well as the Ws and Is at point D, show moisture convergence at all levels that decreases with height (Fig. 11). As expected, there is stronger convergence associated with the Is than with the Ws in each basin, and larger values of convergence for the TCs in the western Pacific, regardless of classification. Note that at point B the Is exhibit greater convergence than Ws at all levels (Fig. 11b), whereas at point D the Is only show stronger convergence from 1000 to 850 hPa (Fig. 11c).

The changes described above, with regard to the depth of the poleward flow south of point B, as well as of the layer of low-level moisture convergence result in a different major source region of environmental moisture for Is and Ws. This was evident from maps of the surface
Fig. 8. Composite maps of moisture transport vectors in units of \((g \, m)/(kg \, s)\) at (a) 700 hPa for 11 Ws at point B, (b) 700 hPa for 11 Is at point B, (c) 925 hPa for 11 Ws at point B, and (d) 925 hPa for 11 Is at point B. The length of the maximum vector is shown at bottom. Also shown are cross-sectional lines for Fig. 9. A tropical cyclone symbol locates point B.

to 700 hPa integrated moisture flux vectors that we examined (not shown). Table 2 provides a convenient summary of the vertically integrated moisture transports for point B across 10°N and 100°W, as well as similar transports at other points discussed later. It is seen that the poleward transport across 10°N, between 120° and 105°W, is more than twice as great for the Is as for the Ws. The reverse is true for the easterly flow across 100°W between 10° and 16°N. At point D in the western Pacific, the major low-level environmental moisture source for both classifications is the northward flow across 5°N that originates from the seas of the Maritime Continent (Table 2). Compared to this, the 1000–700-hPa integrated westerly moisture flux across 120°E is small, despite its large relative changes between classifications.

b. Results at points E and G

We now show some composite results centered about points E and G in the western Pacific basin (Figs. 1 and 2). As mentioned earlier, results surrounding these points can be considered to, respectively, represent the environmental patterns and characteristics for TCs that have or appear to be recurving and those that did not recurve (e.g., Schnadt et al. 1998). Table 1 shows that
we can only compare results for weak and strong systems at these two points. It is worth noting that we did a check, and all but 2 of the 20 TCs (6 W and 14 S) that passed by point E had a northward component to their track. Vertical profiles of $\omega$ are shown in Fig. 12 for the Ws and Ss at points E and G, as well as for their climatologies. The most obvious fact is that the upward motion at point G is much greater than its counterpart at point E for both W and S composites. Also, at point G the composite values of $\omega$ for strong TCs are more than twice as great as those for weak TCs at all levels, while the climatology profile shows weak upward motion throughout the troposphere.

Probably the most important difference among the profiles in Fig. 12 is the level of maximum rising motion. Both profiles at point G show maxima at 400 hPa, which is more typical of tropical TC environments over the northwestern Pacific, although peaks at higher levels (e.g., as high as 300 hPa) have been found in other studies (e.g., McBride and Gray 1980; McBride 1981). The composite profiles for the TCs at point E, however, show a much lower level of maximum upward motion, especially the S composite, where the sharp peak at 700 hPa is not so typical for tropical systems. The most appropriate results that can be compared with ours occur in papers by Gray (1979) and McBride and Gray (1980). Both papers show the same profiles of $\omega$ based on twice daily (0000 and 1200 UTC) composited rawinsonde data for TCs over the northwestern Pacific. The authors computed the average profile of $\omega$ at a radial distance from the center of the TC out to 3$^\circ$. They then partitioned their results into different classes with the most intense being a typhoon that has similar definitions and characteristics to our Ss at point E. For example, their typhoon composite was centered near 23$^\circ$N, 136$^\circ$E, whereas our composites at point E were centered at 25$^\circ$N, 130$^\circ$W. They found that their 1200 UTC (2200 local time) composite profile of vertical motion had maximum rising air at 700 hPa, whereas their 0000 UTC composite showed maximum upward motion at 350 hPa. We examined the 14 individual profiles of Ss at point E using our four times daily analyses and found no evidence of a diurnal variation. Nonetheless, their results, together with others based on mesoscale mea-
surements (e.g., Houze 1989; Black et al. 1996) reveal that maximum rising motion does occur at as low as 700 hPa in some TCs. More will be said later concerning the possible reasons for the low level of maximum upward motion found in the S composite at point E.

Following the same sequence of variables as we did at points B and D, we now examine our EPT results at points E and G (Fig. 13). As to be expected from the greater amount of moisture and heat in TC environments, the composite profile for Ws and Is show much higher values of EPT at all levels than their corresponding climatologies. Also, the profiles at point G show higher values of EPT at all levels for the Ss compared to the Ws. This feature is similar to the results in Figs. 6a and 6b, which show that EPT values at points B and D were greater for Is than for Ws. At point E, however, Fig. 13b shows that values of EPT for Ss are only greater than those for Ws above the 700-hPa level. Below 700 hPa, the values are less for Ss than for Ws due to a smaller increase for Ss from their climatological values. The net effect of this is to reduce the potential moist instability from the 925- and especially from the 850 hPa level for Ss at point E compared to Ws. In fact, the Ss at point E have the least unstable air for the 850–600-hPa layer of any of the profiles in Figs. 6 and 13. This result is corroborated by our alternative stability measure (Figs. 13c,d). Contrary to all other points, the 14 composited storms at point E are in the center of a minimum of parcel buoyancy (Fig. 13d). It is interesting to note that a tongue of anoma-
lously stable air with a small temperature excess of less than 3 K is present to the west and the south of the composite storm center suggesting an advection of drier and/or cooler air from the Yellow and East China Seas into the environment of strong TCs at point E. A comparison of points G and D located in the South China and Philippine Seas, respectively, reveals that the peak value of parcel instability is considerably smaller than its counterpart at point D (Fig. 6c) indicating that our storms in the Philippine Sea grow in an extremely convectively unstable environment.

The possible reasons for the more stable air between the 925- and 600-hPa levels, for the Ss at point E, are now discussed in terms of their relationship to the lower-than-usual level (700 hPa) of maximum rising motion found for this composite (Fig. 12). We examined maps and profiles (not shown) of several variables (e.g., temperature, specific and relative humidity, EPT, and moist static stability) at different levels, as well as vertical wind shear for the Ss at point E. Based on these figures, we have come to the following conclusion. The low EPT values from 1000 to 850 hPa are the combined effect of lower specific humidity, \( q \), and temperature in large parts of our 5\(^\circ\) × 5\(^\circ\) lat-long box. We found that the pattern of anomalously low \( q \) values was similar to the tongue of anomalously stable air in Fig. 13d. Thus, the large-scale advection of low-level dry air seems to have stabilized the western and southern TC environment. In contrast, cooler air off the coast of China did not reach the southern environment of the storm from 850 to 1000 hPa (not shown). Rather, and contrary to all other composites, an almost circular negative temperature anomaly was centered on the composite for point E at 1000 and 925 hPa. Since 1000-hPa low-level relative humidity in ERA-15 was on the order of 85% at the storm center (it exceeded 90% at all other western Pacific points) we speculate that the colder temperatures are attributable to evaporational cooling from raindrops in the subcloud layer. Certainly, the large-scale advection of drier air into the western and southern sectors of Ss at point E is a much more robust peculiar feature than is our explanation given for the temperature drop. The evaporational processes strongly depend on the parameterization of rainfall in the ERA model. In sum-

![Fig. 11](image)

**Fig. 11.** Profiles of composited moisture divergence from 1000 to 700 hPa in units of \([g/(kg s)] \times 10^{-2}\) for (a) Ws and their climatology (dashed), as well as for Is (solid) at point D; and (b) same as (a) except at point B. Also shown are indicators of the level-by-level variability about the profiles for Ws (marked X) and Is (marked O). A 5\(^\circ\) × 5\(^\circ\) average was used to construct profiles.

<table>
<thead>
<tr>
<th>Point D (kg s(^{-1}))</th>
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<tbody>
<tr>
<td>(zonal: 5(^\circ), 119°–135°E, meridional: 1°–4°N, 120'E)</td>
</tr>
<tr>
<td>Weak</td>
</tr>
<tr>
<td>2.98 × 10(^4)</td>
</tr>
<tr>
<td>Intensifying (D, B)</td>
</tr>
<tr>
<td>4.29 × 10(^4)</td>
</tr>
<tr>
<td>Strong (G)</td>
</tr>
<tr>
<td>1.44 × 10(^4)</td>
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<tr>
<td>3.15 × 10(^4)</td>
</tr>
<tr>
<td>1.98 × 10(^4)</td>
</tr>
<tr>
<td>2.70 × 10(^4)</td>
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<table>
<thead>
<tr>
<th>Point B (kg s(^{-1}))</th>
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</thead>
<tbody>
<tr>
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</tr>
<tr>
<td>Weak</td>
</tr>
<tr>
<td>1.23 × 10(^4)</td>
</tr>
<tr>
<td>Intensifying (D, B)</td>
</tr>
<tr>
<td>2.40 × 10(^4)</td>
</tr>
<tr>
<td>Strong (G)</td>
</tr>
<tr>
<td>1.42 × 10(^4)</td>
</tr>
<tr>
<td>0.75 × 10(^4)</td>
</tr>
<tr>
<td>2.24 × 10(^4)</td>
</tr>
<tr>
<td>2.46 × 10(^4)</td>
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<table>
<thead>
<tr>
<th>Point G (kg s(^{-1}))</th>
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</thead>
<tbody>
<tr>
<td>(zonal: 5(^\circ), 105°–114°E, meridional: 8°–15°N, 110'E)</td>
</tr>
<tr>
<td>Weak</td>
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<tr>
<td>1.44 × 10(^4)</td>
</tr>
<tr>
<td>Intensifying (D, B)</td>
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<td>2.46 × 10(^4)</td>
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<tr>
<td>Strong (G)</td>
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<td>1.42 × 10(^4)</td>
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<tr>
<td>0.75 × 10(^4)</td>
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<td>2.24 × 10(^4)</td>
</tr>
</tbody>
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Fig. 12. Profiles of composited vertical velocity, \( \omega_z \) in Pa s\(^{-1}\) \( \times 10^{-2} \) for Ws and their climatology (dashed), as well as for Ss (solid) at (a) point G and (b) point E. Also shown are indicators of level-by-level variability about the profiles for W (marked X) and S (marked O). A 5° \( \times 5° \) average was used to obtain profiles.

As for points B and D, moisture transport maps were compiled at lower-tropospheric levels for the composites about points G and E. Selected maps, which highlight the important features, are shown in Figs. 15 and 16, respectively. The W composite at 925 hPa for point G (Fig. 15a) shows that moisture is transported by south-westerly flow across 5°N, between 105° and 115°E, over the South China Sea. This flow carries moist air into the TC environment. This pattern, which is similar to the one for climatology, does not extend up to the next analysis level (850 hPa); thus, the monsoonal-type flow for the Ws at point G is quite shallow in this region. The moisture transport originating from Indochina appears to be much more important than that across 5°N for the W systems. The westerly flow across 110°E (Fig. 15a), for example, occurred at all levels up to the last level we analyzed (700 hPa), and most of this air appeared to make its way into the TC environment. The S composite at 925 hPa (Fig. 15b) shows a similar pattern to the W composite. The main differences are that the transport vectors are larger for the Ss than for the Ws, and the monsoonal flow across the equator and 5°N is deeper for the Ss than for the Ws. For the S composites, it extends to the 850-hPa level, but not above. Table 2 shows that the vertically integrated transport across 5°N is greater for Ss than for Ws. As for the W composite, the transport for the S composite across 110°E extended throughout the lower troposphere and made its way toward the TC environment. Thus, the integrated values for Ws and Ss at point G are similar (Table 2).

The maps that illustrate the key features of the moisture transport vectors for the composite about point E are shown in Fig. 16. One of the more salient features is that point E is climatologically located at the western
edge of a lower-tropospheric ridge, which is evident in the map at 925 hPa (Fig. 16a); this ridge persisted at all levels up to at least 700 hPa. When the TCs were present, however, they dominated the pattern of moisture transport and the ridge receded 5°–10° latitude toward the north (Figs. 16b–d), and little, if any, of the air from the ridge made its way into the TC environment. For the S composite, two levels of transport are shown, 925 and 700 hPa. At 925 hPa (and 850 hPa, as well), there is cross-equatorial transport south of the Philippines, as well as southwesterly flow over the South China Sea, that appears to have an impact on the TC composite. However, this pattern is not significantly different from climatology (see Fig. 16a). Due to large low-level southwesterly moisture flux anomalies over the Philippine Sea (Figs. 16c,d), the climatological monsoonal flow referred to above extends into the TC environment. This low-latitude moisture source does not extend above 850 hPa; instead a transport by westerly winds across the South China Sea becomes the dominant feature (Fig. 16b). The lower-tropospheric pattern for the W composites (Fig. 16c) was similar to that for the Ss at all levels.

c. Convective and stratiform precipitation rates

The precipitation rates adopted for this study come from the 6-h first guess produced by the ERA-15 model. This 6-h period is the one immediately preceding the synoptic time that corresponded to the date/time when a given TC was selected for compositing, that is, when the TC was nearest one of the seven points shown in Figs. 1 and 2. The model calculates both convective and stratiform precipitation in the following manner. First,
the model checks at a given grid point to see if deep convection occurs, and, if it does, the convective precipitation is produced by the Tiedtke (1989) scheme with a modification of the parameterization of the conversion of cloud water into rain in the convective updrafts (Tiedtke 1993). The convective precipitation that reaches the ground is computed from the vertical integral of rainwater generation and rainwater evaporation that takes place in the downdrafts and the subcloud layer. In order to account for anvil precipitation, the convective cloud condensate that has not been converted into rainwater is a source term in the grid-scale prognostic cloud water equation (Tiedtke 1993). The other relevant sources (sinks) that enter the cloud water equation are condensation (evaporation) through the large-scale lifting (descent) of moist air or radiative cooling (heating). Evaporation also occurs through turbulent mixing with unsaturated environmental air. The parameterization of the conversion of grid-scale cloud water to stratiform rainwater and the computation of the rain that reaches the ground is consistent with the way it is calculated in the convective scheme.

The stratiform rainfall is calculated for the same grid point in the same time step after the convective scheme has been completed. Since, in the ERA-15 model, the cloud condensate detrained from convective updrafts is the major source of cloud water in the Tropics (C. Jakob, ECMWF, 2000, personal communication), convective and stratiform rain are likely to occur at the same time at a grid point, and it is misleading to refer to the stratiform part as large-scale, grid-scale, or stable precipitation. It is worth noting that the stratiform precipitation produced by the ERA-15 model can only be compared to the observed stratiform buoyancy-generated precipitation discussed by Houze (1989, 1997), Marks and Houze (1987), and others, to the extent that it is produced in regions of weak updrafts. However, the ERA model is not able to simulate the cloud microphysical processes that lead to the growth of precipitation particles nor to the vertical profile of diabatic heating that

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**Fig. 14.** As in Fig. 7 except for Ws and Ss, with points G and E replacing points D and B.
occurs in mesoscale convective systems (cf. Houze 1997).

We computed averages over a $5^\circ \times 5^\circ$ box that was the same area used to compute profiles of $\omega$, EPT, and other variables in this paper. The reason we used 6-h first guess values was to be compatible with the $\omega$ values generated by the ERA-15 dataset. Our results are shown in Fig. 17 in the form of bar graphs at each of the four points about which we composited results. Each point has two sets of bar graphs, one for each type of intensity. Within a set, the 6-h precipitation rates in mm (6 h) are partitioned into convective and stratiform. As expected, W systems contain lower precipitation rates than their corresponding I or S systems. Also given in Fig. 17 is the ratio of convective to stratiform rain rates, a quantity that Raymond (1994) suggests might affect the strength of convective systems. The convective rainfall is greater than stratiform rainfall at all points and intensities. The fraction of ERA-15 stratiform precipitation to the total precipitation varies from 25% for Ws at point D to about 47% for Ss at point G and E. It is well known from radar and aircraft studies that the core region of tropical cyclones, out to the major rainbands, contains a considerable fraction of stratiform-type rainfall (e.g., Jorgensen 1984; Burpee and Black 1989). In fact, Marks (1985) found, for Hurricane Allen, that more than 50% of the rainfall within 111-km radius of the center occurred in the stratiform precipitation region outside the eyewall.

Three points are worth emphasizing with respect to Fig. 17. First, we can compare our precipitation rates with those of Frank (1977). He calculated a mean value of 51 mm day$^{-1}$ for nine typhoons over the northwest Pacific based on island station reports. Our composited values (if extrapolated from 6-h to daily rates) range from about 25 mm day$^{-1}$ for Ss at point E to 58 mm day$^{-1}$ for Ss at point G; thus, we bracket Frank’s value. It should be mentioned that, in the ERA-15 model, a spinup of precipitation is observed for both the convective and the stratiform part yielding about 15% higher precipitation amounts when rainfall is taken from 24-h instead of 6-h forecasts (Stendel and Arpe 1997). In any case, the results indicate that the ERA-15 model gives a reasonable estimate of the total diabatic heating released in TCs. Second, for the western Pacific storms, the ratio of convective to stratiform precipitation is always lower for the stronger systems, indicating that stratiform precipitation in the ERA-15 model becomes more important with TC enhancement. This, most likely is due to the fact that the azimuthal large-scale secondary circulation forces stronger ascent in ERA-15 around the storm center. Third, the total rainfall at point B for each TC classification is the smallest when compared to other points. Perhaps this is due, in part, to the fact that TCs are not as well depicted (see section 4) in the eastern Pacific basin as they are in the western Pacific where all of the remaining three points are located (Serrano 1997).

6. Summary and concluding remarks

To our knowledge, the approach of the present paper to composite environmental characteristics and flow features for different TC classifications using a relatively homogeneous reanalysis dataset (ERA-15) is a relatively novel one. It appears that past studies (e.g., Williams and Gray 1973; Ruprecht and Gray 1976a,b; Vincent and Waterman 1979) did not compare results between the eastern and western Pacific basins, nor did they include detailed examinations of moisture variables, as
was done in this study. Our results, however, must be considered to represent larger-scale environmental features of TCs, since ERA-15 did not resolve mesoscale hurricane features, like the eye, eyewall, and rainbands.

Several noteworthy differences occurred in the convective instability patterns and vertical velocity profiles between the two basins considered, as well as between different TC classifications composited about each fixed location. Peak values in the vertical motion within the same classification were greater for systems in the western Pacific basin than for systems in the eastern Pacific. This was probably due, in part, to the greater energy supplied to the western Pacific systems by the warmer sea surface temperatures in that region, but might also be influenced by analysis difficulties in the eastern Pacific (Serrano 1997). As expected, TC composites that were classified as S or I had greater upward motion than those classified as W at their respective locations.

The level of maximum rising motion was highest (400–350 hPa) in the Philippine Sea (point D) for both categories, and in the South China Sea (point G) for weak storms. At point E, on the other hand, the maximum upward motion for strong storms was at 700 hPa.

In comparing our results to those of others who derived vertical motion profiles kinematically from rawinsonde data in the Pacific, we found the following. McBride (1981) and McBride and Zehr (1981) showed that the level of maximum rising motion was highest (400–350 hPa) in the Philippine Sea (point D) for both categories, and in the South China Sea (point G) for weak storms. At point E, on the other hand, the maximum upward motion for strong storms was at 700 hPa.

In comparing our results to those of others who derived vertical motion profiles kinematically from rawinsonde data in the Pacific, we found the following. McBride (1981) and McBride and Zehr (1981) showed that the level of maximum rising motion was located in the 400–300-hPa layer, regardless of whether nondeveloping or developing cloud clusters, or typhoons were considered.
Thus, in general, our levels were lower. On the other hand, Gray (1979) and McBride and Gray (1980) showed several cases of western Pacific weather systems with a variable level of maximum vertical velocity between 700 and 300 hPa. Also, measurements within mesoscale convective systems by Houze (1989) and Black et al. (1996) yielded maximum upward motion as low as 700 hPa.

The lowering of the level of maximum upward motion for the Ss at point E is associated with the decrease in potential (convective) instability found for this composite in the 925–600-hPa layer. All other composites showed that TCs were embedded in regions of maximum instability; however, the Ss at point E were located in an environment characterized by minimum instability (Figs. 13b,d). We identified drier air and colder temperatures in the lowest levels as being the causes for stabilization. We found that the large-scale advection of drier air occurred from the East China Sea into the western and southern sectors of the TC composite. Also, the temperatures below 850 hPa were coldest near the center of the TC environment, most likely due to evaporation of falling rain. In summary, we hypothesize that the resulting changes in the profile of convective (parcel) instability is one reason for the low level of maximum rising motion for Ss at point E. However, that the fact that Ss at point E have one of the lowest ratios of convective to stratiform precipitation, $r = 1.18$ (Fig. 17), cannot be excluded as another reason for the lower level of maximum rising motion.

The thrust of our work was to provide some insight into the moisture environments surrounding TCs in the eastern and western Pacific. In this regard, we found that maxima of precipitable water were always located at or near the composite point for each class of TCs. We found that, at the points of maximum storm frequency in either basin, PW anomalies were of the same order despite the much lower climatological values in the eastern Pacific. Thus, the PW anomalies were found to be related to the intensity of the storm, but not to the amount of background moisture. Furthermore, based on horizontal distributions of columnar PW, $\omega$ at 500 hPa, and parcel instability, it was shown that both the weak and intensifying systems in the eastern Pacific basins were distinctly north of a well-established ITCZ. In the western Pacific, however, a relationship between the locations of the composited TC centers and the ITCZ was not obvious. This was also true after stratifying the storms at D and G into June–August and September–November periods.

With regard to lower-tropospheric moisture transport, we found that the cross-equatorial flow, which originated in the Southern Hemisphere as southeast trade winds, recurved to a monsoon-type southwesterly flow between 5° and 10°N. This pattern was important at all low levels in transporting moisture into the environment.
of intensifying TCs in the eastern Pacific. For weak systems, however, this southwesterly moisture flow was found to be weaker and shallower. On the other hand, the eastern Pacific Ws were more impacted by westward transports originating off the Central American coast. When integrated from the surface to 700 hPa, the net effect was a change in the direction of moisture transport vectors and, therefore, in the major source region. Such a distinct directional change between classifications was exceptional to point B, and was not found for the three points in the western Pacific and South China Sea.

In the western Pacific, the climatological cross-equatorial monsoonal flow was somewhat stronger for Is (Ss) at point D (G) than for W systems at both points. At point D in the Philippine Sea this meridional flow extending from the equator into the TC environment was much more important than the zonal advection, whereas this is not true for point G in the South China Sea. In this region, the zonal flow entering the South China Sea from Indochina and flowing into the TC environment was of similar magnitude as the cross-equatorial flow (Table 2). Finally, we note that changes occurred in the depth of the low-level moisture flux. For example, at point D, the westerly flow at 700 hPa across the Philippine Islands for intensifying storms was not evident in the W composite (Fig. 10).

The last variable we examined was ERA-15 model precipitation. Using the same areal boxes as for our \( \omega \) and EPT profiles, we calculated the mean 6-h rainfall rates from the ERA-15 dataset. We found that our total precipitation rates for Ss and Is in the northwestern Pacific were compatible with those of Frank (1977) who used large-scale data. It must be noted, however, that the weak systems in the eastern Pacific experienced an extremely low average precipitation rate of 9 mm day\(^{-1} \) which might indicate some difficulties in the analyses of weak storms in this area. The fraction of ERA-15 stratiform precipitation to the total precipitation varies from 25% to 47% in our composite samples. The representation of convective and stratiform rain in the ERA-15 model obviously favors the latter when the systems are stronger and have a more intense and broader secondary circulation. Unfortunately, we were not able to check the amount of stratiform precipitation that was due to anvil precipitation or to forced ascent of moist air, as is specified in the Tiedtke (1993) scheme. Tropical cyclones can have a non-negligible fraction of rainfall due to nonconvective processes, that is, forced ascent of moist air (Houze 1997). The fraction of stratiform rain in the ERA model is somewhat lower than what is known from radar studies (Jorgensen 1984; Marks 1985). It should be noted however that the contribution from stratiform rain will likely increase as the model resolution increases.

Finally, we are aware that it is difficult, from parameter to parameter, to assess the degree to which our results are influenced by the model, its assimilation scheme, and the input data used to produce the ERA-15 dataset. However, we consider the ERA-15 to be one of the best large-scale datasets presently available to conduct multiyear studies on the environmental features of tropical cyclones. Since the time when ERA-15 became available, model physics, model resolution, and the database have improved and will continue to improve in the ECMWF and other forecast models. For example, in 2002 the ERA-40 analyses will become available with higher horizontal and vertical resolution. In this regard, we consider the findings of our study as a useful basis for comparison with more advanced operational analyses or other large-scale datasets of the future.

Acknowledgments. First, the authors express their gratitude to the reviewers for their many constructive comments that, in our opinion, led to a greatly improved manuscript. They also thank Ms. Peggy Allard at Cologne for her assistance in producing many of the figures that appear in this paper and Ms. Dagmar Janzen for typing the manuscript. In addition, they acknowledge that some of the computational work was performed at the German Climate Computing Center in Hamburg, Germany. Finally, they express their appreciation for the assistance provided by Dr. Christian Jakob of ECMWF and Dr. Frank Marks from NOAA/AOML.

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