The Dependence of the Low-Level Equatorial Easterly Jet on Hadley and Walker Circulations

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ABSTRACT

How the time-mean Hadley and Walker circulations affect the formation of a low-level equatorial easterly jet is investigated. Experiments are conducted for equinoctial conditions using a general circulation model, the Community Climate Model (CCM1), that includes a Kuo convective scheme and a lower boundary that is specified to be water at a fixed sea surface temperature (SST). Several zonally symmetric SST forcings are used to determine how various Hadley circulations affect the tropical zonal wind field. A zonal wavenumber one equatorial SST anomaly superimposed on a zonally symmetric SST distribution forces a wind field that includes both Hadley and Walker circulations.

The Hadley circulation experiments produce equatorial easterlies and low-level jets on the poleward sides of the intertropical convergence zone (ITCZ) 10° to 15° from the equator. In an experiment with a single, dominant off-equatorial ITCZ in the Northern Hemisphere, the Southern Hemisphere jet moves to within 7.5° of the equator; yet none of the Hadley circulation cases produce a low-level easterly jet on the equator because they lack a mechanism to vertically confine the flow.

The experiment that includes a zonally overturning cell on the equator produces a low-level equatorial easterly jet in the cold tongue region that is similar to the observed jet over the central to eastern Pacific. That case shows that east of the equatorial warm pool, the Walker circulation and its induced Kelvin wave response provide the necessary upper-level westerly flow and subsidence to vertically confine the low-level easterlies into a jet. Spring and fall climatological runs of the CCM1 with land surfaces, seasonally varying SSTs and insolation, and a moist convective adjustment scheme support the hypothesis that the Walker circulation provides the vertical confinement necessary to form a low-level equatorial easterly jet in the region east of the equatorial convective center, regardless of the Hadley circulation in that region.

The eddy vertical-flux convergence of moisture in the Kuo convective scheme produces a dry tongue in the Walker circulation simulation below the low-level equatorial easterly jet. The CCM1 climatologies show that the dynamics of the jet do not depend on this feature. Betts, Albrecht, and Kloesel have observed a similar feature just above the boundary layer in the central to eastern Pacific and, without referring to the low-level jet, they have hypothesized a mechanism in which convection forms this dry layer. Analysis of the simulations performed here suggests that the model's parameterized convective physics utilize the same mechanism to form the dry tongue in the vicinity of the low-level equatorial easterly jet; however, since the mechanism of Betts, Albrecht, and Kloesel has not yet been confirmed through observational studies, the relationship between the observed and modeled dry tongue remains speculative.

1. Introduction

Observations reported by Hastenrath (1971), Bunker (1971), and Sadler and Kilonsky (1981) indicate that low-level easterly winds in the central and eastern Pacific form an equatorial jet. Hastenrath and Bunker, using central Pacific upper-air soundings (ESSA, March and April 1967) and data from the Line Islands Experiment (LIE) at roughly 160°W in boreal spring of 1967, found easterlies exceeding 10 m s⁻¹ within the core of this jet centered around 850 mb and 0° to 5°N (Fig. 1). Sadler and Kilonsky constructed monthly horizontal plots of zonal wind based upon 2.5° latitude by 10° longitude gridded satellite-observed cloud motions averaged from 1975 to 1980. They estimated the height of the winds that corresponded to these cloud motions as 900 mb or 1000 m. For most months they found a ridge of maximum wind speed of nearly 10 m s⁻¹ centered on the equator in the central to eastern Pacific (Fig. 2). This jetlike feature was ill defined only in April and May, and, on average, it occurred between 5°N and 5°S, 120° and 150°W. Sadler and Kilonsky referred to this feature as an easterly jet primarily because it is a sharply confined feature in the meridional plane (narrower than the equatorial Rossby radius). They did, however, offer some evidence for its vertical confinement, as Carr (1983) noted—namely, mean soundings from three of the Line Islands taken during LIE in March and April 1967 that showed stronger

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Fig. 1. Observed zonal winds in the central to eastern Pacific. (a) Time-average zonal wind during March and April 1967 in a meridional transect along approximately 160°W (Hastenrath 1971). (b) Monthly mean zonal winds at 850 mb for April 1967 (redrafted from Bunker 1971). Easterly velocities are negative.

winds and wind shear from 900 to 850 mb at lower latitudes and the inconclusive evidence that the ship climatologies of Wyrtki and Meyers (1975) did not show any jet at the surface. Sadler and Kilonsky speculated that this jet develops in response to higher atmospheric stability over the equator that is caused by a colder sea surface and by divergent, sinking easterly flow; presumably, the strong easterlies at these low levels at other locations are weakened as they become vertically mixed under less stable conditions.

Subsequent to the studies of Hastenrath (1971), Bunker (1971), and Sadler and Kilonsky (1981), the low-level easterly jet has escaped the attention of the climate modelers. Nonetheless, this jet is a feature of the climate simulated by the general circulation models (GCMs). As an example, we present in Fig. 3 the climatological February, March, and April (FMA) mean zonal wind from a 5-yr integration of the T42 control run (case 261) of the Community Climate Model (CCM1) performed at NCAR (history tapes for this run are documented in Hack et al. 1989). The FMA season is chosen because it corresponds to the time of Hastenrath and Bunker’s observations of the jet, and it is also the season when two near-equatorial intertropical convergence zones (ITCZs) commonly are observed in the eastern Pacific (though the southern one is considerably weaker than its northern counterpart). The zonal wind field for this season develops a jet comparable in speed, location, and horizontal extent to that observed by Sadler and Kilonsky (compare Fig. 2 with Fig. 3a).

If this jet is a robust feature of the time-mean tropical circulation, as suggested by Sadler and Kilonsky’s 6-yr data and by the simulations using the numerical climate model, then it should be influenced by two of the most robust features of the Tropics: the Hadley and Walker circulations. Both circulations vary in strength and location on timescales from the diurnal to the interannual due to the monsoons, ENSO, and tropical easterly waves, but in the time average the Hadley and Walker circulations are well defined. The time-mean Hadley circulation in each hemisphere consists of a thermally direct, meridionally overturning cell whose upward branch constitutes the ITCZ and whose lower and upper branches consist of equatorward and poleward flow, respectively. The lower branch of the Hadley cell in the Northern (Southern) Hemisphere forms the northeasterly (southeasterly) trade winds as equatorward flow deflects because of angular momentum constraints.

The time-average Walker circulation, as described by Bjerknes (1969), consists of the thermally direct zonally overturning equatorial cell in the Pacific. Some authors include other zonally overturning equatorial cells whose upward motions occur over Africa and South America in the Walker circulation; but the Pacific cell constitutes the dominant circulation. It includes an upward branch over Indonesia with its accompanying lower-(upper-) level zonal convergence (divergence) and a downward branch located over the central and eastern Pacific. Within this downward por-
Fig. 3. Climatological winds for FMA from the T42 control run of the CCM1 by Hack et al. (1989). (a) Zonal winds, $u$, at 811 mb over the eastern Pacific, and (b) winds zonally averaged from 120° to 150°W [the longitudes of the average low-level easterly jet as observed by Sadler and Kilonsky (1981)]. Contours and vectors in (b) are as in Fig. 5. Maximum vector in (b) represents 5.2 m s$^{-1}$.

tion of the Walker circulation, low-level easterly winds flow back toward the western Pacific to complete the Walker cell. Not surprisingly, this region contains the low-level equatorial jet, and this suggests that the Walker circulation could be important for its maintenance.

The effects of the Hadley and/or Walker cells on the tropical circulation have been studied by using the shallow water equations (see, e.g., Matsuno 1966; Gill 1980; among many others), simple two-dimensional models (e.g., Schneider and Lindzen 1977; Schneider 1977; Held and Hou 1980), linearized steady-state primitive equation models (e.g., Geisler 1981; Webster 1981), and general circulation models (e.g., Hess et al. 1993, hereafter referred to as HBR93; Numaguti and Hayashi 1991). However, the simple analytic and modeling studies have not produced a low-level easterly jet on the equator, perhaps because their simplified geometries poorly simulate the interactions between the Hadley and Walker circulations or, perhaps, because their parameterizations of boundary-layer physics, dif-
fusion, radiation, and/or convection are inadequate. In the Walker circulation studies by Matsuno (1966), Gill (1980), and Webster (1981), maximum easterly winds formed at the lower boundary and maximum westerly winds aloft to the east of the heating; the low-level easterly winds extend throughout the lower half of the troposphere, and the meridional scale of the winds is the Rossby radius. In the Hadley circulation modeling studies by Held and Hou (1980), Schneider and Lindzen (1977), HBR93, and Numaguti and Hayashi (1991), equatorial winds are generally easterly at low-levels but not vertically confined into a jet.

In an attempt to better understand the role of the Hadley and Walker circulations in defining the tropical circulation and, in particular, the low-level jet, a series of experiments was designed whereby a GCM is forced by equinoctial insolation conditions and by various prescribed sea surface temperature (SST) distributions. We demonstrate the low-level jet exists because of the Walker circulation that is forced by the zonally asymmetric SST distribution, and the jet is maintained by physics that is inherently 3D and by physical processes that have not been included in studies of the Walker circulation using simpler models. The paper proceeds following a brief description of the model in section 2. We attempt to clarify the effects of various Hadley cells on the tropical circulation in section 3 by analyzing three experiments from HBR93 that were forced by zonally symmetric SSTs. In section 4, using the same 3D model of HBR93, we investigate how the interaction of a Hadley and Walker cell provides both a mechanism for enhancing low-level equatorial easterlies and for confining them into a low-level jet by running an experiment forced by zonally asymmetric SSTs. A dry tongue that forms coincident with the low-level jet is investigated in section 5. Conclusions of this work appear in section 6.

2. Model description

The model used in this study (and in HBR93) is Version One of the Community Climate Model (the CCM1) provided by the National Center for Atmospheric Research (Williamson et al. 1987; Hack et al. 1989). The integrations are performed using the standard horizontal resolution, which truncates the first 42 spherical harmonics trapezoidally (T42 resolution), resulting in an equivalent horizontal grid spacing of roughly 2.8° in latitude and longitude. All experiments use the standard vertical resolution of the model: 12 sigma levels (σ = p/p_s, where p is pressure and p_s is surface pressure), three of which lie within the planetary boundary layer. The 12 sigma levels used in these experiments are 0.991, 0.926, 0.811, 0.664, 0.5, 0.355, 0.245, 0.165, 0.11, 0.06, 0.025, and 0.009. For ease of discussion, these sigma levels have been interpolated to the following pressure levels: 950, 910, 811, 664, 500, 355, 245, 165, 110, 60, 25, and 9 mb.

Boundary-layer effects are parameterized by vertical diffusion of momentum, heat, and moisture. The model employs the bulk aerodynamic parameterization of Deardorff (1972) at the surface, while throughout the rest of the atmosphere it parameterizes vertical stresses on a field, A, as −ρKδA/δz, where K is the eddy viscosity and ρ is density. Both the drag coefficient in the bulk aerodynamic parameterization and the eddy viscosity are allowed to vary with stability and wind shear. The radiation scheme includes parameterizations for solar zenith angle dependence of albedo on various surface types; absorption by ozone, water vapor, and oxygen; and emissivity of stratiform clouds due to their liquid-water content.

To eliminate asymmetries in heating so that the Hadley and Walker circulations will remain in place, insolation in the model is set at perpetual equinoctial conditions, the surface is set to ocean everywhere—an aqua-planet—and the SSTs are fixed. In addition, the standard version of the CCM1 does not include a diurnal cycle. The experiments conducted with this model also include a modified Kuo convective scheme [see section 5b (also Donner 1986) or, for further references and a description of the convective scheme, HBR93] in addition to the model’s standard convective scheme (Manabe et al. 1965).

3. Tropical circulation and the Hadley cells

Before looking at the combined effects of the Hadley and Walker cells, it is useful to note how the Hadley cell alone affects the tropical zonal circulation in this model. The steady-state Hadley circulation in this time-dependent and fully three-dimensional model is isolated in the experiments of HBR93 that ran integrations with constant, zonally symmetric forcing. When averaged for a sufficiently long time period, the circulations develop approximate steady-state conditions. The three experiments from HBR93 discussed here produce significantly different Hadley cells and tropical circulations by utilizing different prescribed meridional SST profiles, displayed in Fig. 4. We will show that, as in the axially symmetric (2D) studies discussed in section 1, the zonally symmetric 3D simulations presented in HBR93 (and in Numaguti and Hayashi 1991) demonstrate a robust near-equatorial zonal circulation but do not simulate the low-level easterly tropical jet.

In the first experiment, called the Kuo–Std experiment in HBR93, the zonally symmetric SST peaks on the equator and decays symmetrically and monotonically to the poles (Fig. 4). With this forcing, the model

1 In the integrations analyzed in this paper, at least 80 days were used for these time averages.
2 The meridional profile of SST in the Kuo–Std experiment was determined from the March climatology (Alexander and Mobley 1976) that is zonally averaged, symmetrized about the equator, and modified near the poles to remove temperatures below freezing.
produces two Hadley cells, ITCZs between 4° and 7° latitude on each side of the equator, and one region of deep upward motion throughout the equatorial band extending from 10°N to 10°S. The zonal winds develop as nearly geostrophically balanced easterlies throughout the troposphere in the near-equatorial regions, and although easterlies greater than 7 m s⁻¹ form over the equator from 811 to 500 mb, no low-level equatorial jet forms (compare Fig. 1a with Fig. 5). Two low-level off-equatorial easterly jets form, however, near 910 mb and 12.5° latitude. These predominantly geostrophic jets form near the latitude of the maximum meridional pressure gradient. In addition to a weakening meridional pressure gradient, the winds blowing toward the equator within 12° latitude encounter other forces that hinder the formation of a low-level easterly jet on the equator. Equatorward of 12°, upward motions, which maximize between 4.2° and 7°, transport and mix zonal momentum vertically, opposing any vertical confinement necessary to form a jet. The vertical confinement of the low-level easterly flow also decreases with decreasing latitude as the friction layer, the depth over which surface drag on the wind is important, deepens. Ekman theory provides a simple qualitative framework for understanding why the friction layer deepens with decreasing latitude. The depth of the friction layer is inversely proportional to the Coriolis parameter, f, so decreasing latitude implies an increasing Ekman depth. A deepening friction layer implies a deepening layer of weak ageostrophic flow and weakening shear. Evidence of a deeper friction layer over the equator than at the latitudes of the off-equatorial jets can be seen in the cross sections of zonal wind (Fig. 5). Near 12°, significant vertical shear of the zonal wind occurs from the surface (σ = 1.0, where the flow is zero) to about 910 mb, and strong ageostrophic zonal wind (not shown) accompanies this in a shallow layer, whereas significant shear of the zonal wind extends to 811 mb over the equator, and weak ageostrophic zonal wind accompanies that in a deeper layer. Additional evidence for the increasing friction-layer depth with decreasing latitude and a further discussion of Ekman theory are provided in the appendix.

In another experiment, Kuo—Oct, the SST profile used in the Kuo—Std experiment was modified so there are maxima in SST off the equator at 7°N and 7°S (see Fig. 4). In this experiment, two Hadley cells form with their ITCZs centered over the warmest waters. With the ITCZs in this case more latitudinally separated than in the Kuo—Std case, the upper-level outflow from the ITCZs converges and forces subsidence over the equator. The low-level zonal flow in the near-equatorial region (Fig. 6) develops equatorial easterlies and off-equatorial easterly jets comparable to those in the Kuo—Std experiment, but the upper-tropospheric zonal winds over the equator form a deeper layer of easterlies that is less vertically confined than those in Kuo—Std. Hence, no low-level equatorial easterly jet forms in this experiment either.

In a third experiment from HBR93, Kuo—Poq, forced by a SST distribution with a single peak at 10°N that decreases monotonically to the poles (Fig. 4), a single, dominant Hadley cell develops with its upward motion off the equator (Fig. 7). An off-equatorial jet in the hemisphere opposite the ITCZ, again located at the latitude of the strongest meridional pressure gra-

![Fig. 4](Image)

**Fig. 4.** Sea surface temperatures for the various experiments. Values for Kuo—Std, Kuo—Oct, and Kuo—Poq are zonally invariant. The two Kuo—Walker profiles are taken through the centers of the warm pool at 90°E and the cold tongue at 90°W.

![Fig. 5](Image)

**Fig. 5.** Zonally averaged winds for Kuo—Std. Zonal wind, u, is contoured at 1 m s⁻¹ (2 m s⁻¹) intervals for negative (positive) speeds with dashed (solid) lines. Vectors represent u, meridional, and w, vertical, velocities. Vertical velocity is scaled by a factor of 500. Maximum vector represents 3.35 m s⁻¹.
dient, moves to within 7° latitude, but over the equator, the low-level equatorial easterlies still lack the vertical confinement to form a jet.

In each model integration forced by zonally symmetric SSTs, easterly winds develop throughout the depth of the troposphere in low latitudes when low-level meridional winds associated with the Hadley cell(s) flow toward the equator. The zonal wind field is largely geostrophic throughout the Tropics, though ageostrophic accelerations become important aloft and in the boundary layer due to advections and friction, respectively [see Ovens (1993) for a more complete discussion of the momentum balance]. Though an easterly jet forms near 910 mb off the equator in each experiment, no low-level tropical jet develops over the equator in any experiments with this 3D general circulation model under zonally symmetric forcing.

4. Tropical circulation with Walker and Hadley cells

a. Introduction

Though the Hadley circulations in the three experiments forced by zonally symmetric SSTs cannot form a low-level easterly jet on the equator, it remains to be seen if the other robust time-average tropical circulation, the Walker circulation, is sufficient to account for this jet. From previous studies of the Walker circulation by Geisler (1981), Gill (1980), and Webster (1981), one expects to find easterlies at low levels east of a strong equatorial heat source where the flow converges toward the heating and westerlies or weaker easterlies aloft where the flow diverges. But none of the previous studies predicted the spatial scale, speed, location, or even existence of a low-level easterly jet, possibly because of poor vertical resolution, linearized equations, or some other simplifications such as in the treatment of the boundary-layer physics. In the experiment to be discussed in this section, we use a general circulation model, the CCM1, identical to that used in all of the experiments discussed in the previous section, to examine how the Walker circulation affects the low-level flow in the Tropics. The conditions for the integration include prescribed equinoctial insolation and SSTs with a zonal wavenumber one component in the deep Tropics. In the integration of the model with the new SST distribution, which will be referred to as the Kuo–Walker case, a low-level easterly jet analogous to the one observed in the central to eastern equatorial Pacific does develop. In this section, the discussion focuses on the maintenance of this jet and the tropical circulation associated with the Hadley and Walker cells.

b. SST forcing for the Kuo–Walker case

In order to generate a Walker circulation in the same aquaplanet CCM1 that was used for the previous experiments, we construct an SST distribution with zonally asymmetric temperatures over the equator. The SST field that was used in the Kuo–Std case (Fig. 4) serves as the (zonally symmetric) basic state, and upon this we superimposed a zonal wavenumber one anomaly intended to resemble the warm and cold SSTs found in the equatorial Pacific Ocean. The resulting wave-number one anomaly includes: 1) a zonal peak to peak amplitude of 5°C on the equator, 2) a cold tongue centered at 90°W with a latitudinal e-folding scale of 6°, and 3) a warm pool at 90°E with a latitudinal e-folding
scale of 12°. The total SST field within 20° of the equator for this experiment and the observed SST field for September, after which this distribution is patterned, are shown in Fig. 8. Note that the SSTs along the meridians at the centers of the cold and warm anomalies, though they differ in magnitude, resemble the profiles from the Kuo–Oct and Kuo–Std cases, respectively (see Fig. 4).

c. Results

Vertical velocity (contours) and horizontal wind (vectors) at 811 and 165 mb (Fig. 9) summarize the time-mean tropical circulation that develops due to this zonally asymmetric forcing. The most important aspect of the time-mean tropical circulation in this Kuo–Walker study is that a single equator-centered ITCZ forms over the warm pool while two off-equatorial ITCZs form flanking the cold tongue. In the following discussion with its accompanying figures, the circulations over the warm pool and cold tongue, unless otherwise noted, refer to longitudinal averages from 60° to 120°E and from 60° to 120°W, respectively. This zonal averaging for these regions facilitates the discussion and highlights the contrasts between the regions by eliminating the intervening longitudes where the flow transits from one regime to the other.

Over the warm pool, the ITCZ is consistent with the circulation in the observed western Pacific and in the Kuo–Peq case of HBR93 (not to be confused with the Kuo–Poq case discussed in section 3), wherein an enhancement of the meridional SST gradient used in the Kuo–Std case forced a single ITCZ to form over the warmest water on the equator. Accompanying the ITCZ in the warm pool, two symmetric Hadley cells form (Fig. 10) whose downward branches are centered at 25° and whose lower-level equatorward and upper-level poleward branches produce meridional velocities up to 5 m s⁻¹ and 3.8 m s⁻¹, respectively. The equatorial precipitation averaged over 14 mm day⁻¹, which exceeds the values of over 8.5 mm day⁻¹ observed by Jaeger (1976) and over 9.3 mm day⁻¹ reported by Dorman and Bourke (1979) over the western Pacific. Low-level off-equatorial easterly jets, comparable to those in the experiments forced by zonally symmetric SSTs, emerge, but the zonal winds over the near-equatorial latitudes develop a significantly different profile from any in the symmetric cases (compare Fig. 10 with Figs. 5 and 7). Here, the vertical extent and speed of the equatorial easterlies decrease: zonal winds reach a maximum of only 2 m s⁻¹ at 910 mb, decrease to zero at roughly 500 mb, and then switch to weak westerlies until 150 mb, where they then switch back to easterlies.

In the cold tongue region, two ITCZs form above the warmest waters at 10°N and 10°S (vectors in Fig. 11). Meridional and large-scale vertical winds show the expected Hadley cell structure around the ITCZs: maximum upward motion at 10° with its coincident upper-

![Fig. 8. Tropical sea surface temperatures for (a) the Kuo–Walker integration and (b) climatological September values used by the CCM1. Contours are at 1 K intervals. Source for (b) is Hack et al. (1989) and is derived from Alexander and Mobley (1976). Note that the Kuo–Walker case is an aqua-planet, and the continental outlines in (a) are drawn only for continuity with (b).](image-url)
tropospheric outflow and low-level inflow. Over the equator, air subsides from 150 mb down to 800 mb. The equatorial subsidence coincides with a layer of convergent meridional winds from the tropopause to 550 mb. Though otherwise similar to the Kuo–Oct case, the subsidence caps a layer of shallow rising motion from the surface to roughly 850 mb. Between the rising and sinking motions lies a shallow layer of meridionally diverging flow from 850 to 550 mb. Precipitation under the ITCZs averages 7 mm day$^{-1}$, while over the equator it averages only 2.4 mm day$^{-1}$. Observed annual precipitation rates under the ITCZ and the equator at approximately 160°W are 12 and 1.5 mm day$^{-1}$, respectively (Hastenrath 1971), while over the eastern Pacific, Dorman and Bourke (1979) reported values of 9 and 3 mm day$^{-1}$, respectively. On average, deep convection in the double ITGZ region extends to only about 350 mb. The shallowness of the deep convection is due to the stabilizing effects associated with the subsiding air aloft, the source of which is in the deep convective region to the west.

As with all previously mentioned experiments with this model, zonal winds form low-level off-equatorial jets, and in the cold tongue region they are centered near 12.5° (Fig. 11). This figure shows that, in contrast to previous runs, however, a low-level easterly jet and strong upper-tropospheric westerlies develop on the equator over the cold tongue. Since the jet observed by Hastenrath (1971), Sadler and Kilonsky (1981), and Bunker (1971) (Figs. 1 and 2) could be defined by easterly speeds greater than or equal to 10 m s$^{-1}$, we define this jet also by easterlies with those speeds. As seen in Fig. 11, the jet thus defined extends from roughly 10°N to 10°S and 900 to 700 mb with maximum speeds at its 811 mb core of 12.5 m s$^{-1}$. Westerlies prevail above 500 mb in the latitudes between

Fig. 9. Time-average winds for Kuo–Walker. Vectors represent total zonal, $u$, and meridional, $v$, velocities. Large-scale vertical velocity, $w$, spectrally filtered to include only zonal wavenumbers zero to four, is contoured in (a) at 1 mm s$^{-1}$ intervals at 165 mb and in (b) at 0.6 mm s$^{-1}$ intervals at 811 mb with downward motions dashed. Labels multiplied in (a) by 10$^3$ and in (b) by 10$^4$. 

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the ITCZs and reach maximum speeds of 32.5 m s\(^{-1}\) at 165 mb.

In contrast to any of the cases mentioned in section 3, a Walker circulation (defined as the zonally overturning thermally direct cell between the warm pool and cold tongue regions) develops in this case, as shown by the vectors of vertical and zonal winds in Fig. 12. Upward motion in the cell, as expected, occurs in a longitudinally broad band over the warm pool with maximum velocities centered over the warmest SSTs at 90°E and 245 mb. The Walker circulation establishes its downward motion over the cold tongue with a less well defined maximum spread out roughly from 100° to 20°W and from 500 to 245 mb. The remaining portions of the Walker circulation consist of low-level zonal convergence (divergence) and its accompanying upper-level divergence (convergence) in the warm (cold) region. The zonal wind departures from the zonal mean (not shown) imply that two wide Walker cells form on either side of the convection, but, upon closer inspection of the wind fields, we find that only in a narrow zone from 10°W to 20°E do both zonal and vertical winds exhibit signs of a separate zonally overturning cell west of the convection. Thus, as found in the observations by Bjerknes (1969), Madden and Julian (1972), and others, and as found in the modeling work of Gill (1980) and Geisler (1981), the eastern cell with its low-level easterly flow and upper-level westerly flow dominates the simulated time-average Walker circulation.

d. Analysis

We discuss the warm pool and cold tongue regions separately and emphasize their Hadley cells and the forces responsible for the near-equatorial zonal wind field. The transitional regions will be discussed, as appropriate, when the overall Walker circulation and its relevance to the low-level easterly jet are analyzed.

As in the experiments forced by zonally symmetric SSTs, geostrophy accounts for the majority of the time-
average zonal wind in the Tropics [for more details, refer to Ovens (1993)]. Near the equator and the surface, nonlinear advection and friction become important, but ageostrophic wind speeds are still much less than geostrophic speeds. In this experiment, contrary to the zonally symmetric cases in section 3, westerly flow develops aloft east of an equatorial heating maximum and is nearly geostrophic. Hence, the circulations over the cold tongue and possibly the warm pool, which separately would not produce westerly flow aloft, will have westerly geostrophic components due to the equatorial Kelvin wave that is forced by the heating in the warm pool region. In the following analysis of the zonal wind in both the cold and warm regions, comparisons will be made to the winds in the Kuo–Std and Kuo–Oct cases to determine how important a role the Walker circulation plays in maintaining the tropical zonal wind field in this experiment.

1) ZONALLY AVERAGED CIRCULATION OVER THE WARM POOL

Though the SST forcing, the Hadley circulation, and the zonal winds poleward of 15° over the warm pool resemble those in the Kuo–Std case (see Fig. 4 and compare Fig. 5 with Fig. 10), the zonal winds near the equator differ significantly between the two cases. Weak low-level easterlies and upper-level westerlies develop over the warm pool in the Kuo–Walker case, while strong low-level and somewhat weaker upper-level easterlies form in the Kuo–Std integration.

High temperatures and low pressure near the surface result from the warm SSTs and the meridional circulation that they induce. Specifically, higher virtual temperatures occur on the equator than off the equator up to 245 mb because of greater latent heat released in the ITCZ over the equator than in the surrounding latitudes. In accordance with thermal wind, the upper-level (mostly geostrophic) westerlies over the equator are associated with a meridional pressure maximum on the equator, and the low-level easterlies are associated with a meridional pressure minimum. Above 245 mb, strong vertical motions continue in the ITCZ but without significant latent heating (since nearly all the moisture has condensed out by this level), so that adiabatic cooling causes layers above the equator to become cooler than their surroundings, thereby decreasing the zonal geostrophic wind with height in agreement with thermal wind balance.

Though predominantly geostrophically balanced in the near-equatorial regions, the upper-level zonal wind exhibits a strong ageostrophic component over the equator (see Fig. 13). An analysis of the momentum budgets over the warm pool (see Ovens 1993) shows that ageostrophic easterlies develop where the upper-level steady and transient meridional advection of easterly momentum and the vertical advection of westerly momentum are the main forces affecting the upper-level flow. Weak (0.5 m s⁻¹) westerly ageostrophic winds between 500 and 811 mb at the equator develop in a layer where a westerly acceleration by the zonal pressure gradient is balanced by easterly accelerations from meridional advection and horizontal diffusion of zonal momentum (Fig. 12). The profile of equatorial geopotential zonal anomalies in Fig. 12 shows that the zonal pressure gradient is strong over the warm pool in the lower troposphere (below 644 mb); further discussion of the sloping geopotential effect is found in section 4d.3.

The low-level off-equatorial easterly jets common to all three of the experiments with zonally symmetric SSTs also occur over the warm pool in this Kuo–Walker case (Fig. 10). Here again the zonal geostrophic flow strongly influences the speed and location of the jets, showing easterlies near 12.5° with speeds and meridional pressure gradients that reach a maximum at the lowest model level. As in the symmetrically forced experiments, significant frictional drag develops below these jets. In addition, below roughly 900 mb and within 5° of the equator, vertical advection of slower easterlies and a positive zonal pressure gradient become at least as important as friction in determining the strength of the flow.

The most prominent differences in the tropical wind fields between the Kuo–Walker warm pool region and the Kuo–Std case are: 1) that westerlies, balanced primarily by westerly accelerations by vertical advection and easterly accelerations by meridional advection and the zonal pressure gradient, form at upper levels in Kuo–Walker, not easterlies as in Kuo–Std. 2) that easterly speeds are over four times greater at 811 mb
in Kuo–Std, even though geostrophy accounts for the low-level equatorial easterlies in both cases, and 3) that ageostrophic lower-level westerlies develop in Kuo–Walker in a region with a strong zonal pressure gradient. Clearly, the Walker circulation strongly influences the tropical pressure distribution and the forces affecting the zonal wind throughout the troposphere.

2) Zonally averaged circulation over the cold tongue

The SST forcing and the residual meridional circulations in the cold tongue region of this experiment resemble those in Kuo–Oct (see Fig. 4); however, the zonal flows differ substantially in upper levels. The greatest difference between the tropical zonal winds in these cases is the strong westerly flow aloft in Kuo–Walker and the associated vertical confinement of easterlies below 500 mb. The westerlies, as expected from the work of Matsuno (1966), Gill (1980), Webster (1981), and others, develop in a predominantly geostrophic balance (the Kelvin signal) (cf. Figs. 11 and 14). Another important difference between the cold tongue and the Kuo–Oct cases is that in the cold tongue region the zonal flow well above the boundary layer shows greater westerly speeds over the equator than at the latitudes of the ITCZs (cf. Figs. 6 and 11). As the thermal wind equation indicates, this occurs because of a warmer air column over the equator. Both Kuo–Oct and Kuo–Walker cases display large subsidence warming over the equator, but in Kuo–Walker the precipitation rates in the ITCZs and over the equator differ by only 4.5 mm day$^{-1}$, as opposed to 10 mm day$^{-1}$ in Kuo–Oct. This contrast in rainfall rates and its associated latent heating contrast enable the air column along the equator and over the cold tongue in the Kuo–Walker case to be warmer than that in the ITZC, from roughly 850 mb to the tropopause. At 811 mb, the zonal geostrophic wind reaches maximum easterly speeds of over 11 m s$^{-1}$ near 5°N and 5°S (a 3 m s$^{-1}$ increase over the Kuo–Oct wind at the same location). The differences in geostrophic zonal winds in the Kuo–Oct case and over the cold tongue in the Kuo–Walker case clearly illustrate the effect of zonal asymmetry on the meridional pressure field on this idealized aqua-planet. Further discussion of the meridional pressure field and the Walker circulation appears in section 4d.3.

The greatest similarity in the tropical zonal winds among all the cases studied occurs in the low-level off-equatorial jets. Over the cold tongue, easterly geostrophic maxima near 12° and westerly ageostrophic flow throughout the tropical low levels combine to form jets centered at 12° and 910 mb with easterly speeds of 11 m s$^{-1}$ (Figs. 11 and 14). As expected, meridional pressure gradients, friction, vertical advection, and the changing depth of the friction layer help to define these jets.

Ageostrophic forces above 900 mb affect the time-average zonal wind field over the cold tongue within 10° of the equator in several ways: 1) they strengthen the easterly jet at 811 mb, 2) they provide westerly accelerations to the flow between 650 and 200 mb that further help to vertically confine the low-level easterlies, and 3) they decelerate the flow near 165 mb (Fig. 14). As indicated by momentum budget calculations (see Ovens 1993), the easterly ageostrophic flow at 811 mb develops where the dominant zonal accelerations are easterly from the zonal pressure gradient and westerly from zonal and vertical advections. As is the case over the warm pool, the zonal pressure gradient becomes important above the friction layer in the lower troposphere; it reaches its strongest value near 811 mb. Over the equator from 650 to 200 mb, ageostrophic westerlies, which enhance the vertical confinement of the low-level jet, develop as the vertical advection by subsiding, more westerly flow becomes the dominant term in the zonal momentum equation. Subsidence imports air with westerly momentum up to 165 mb, but near this level zonal and transient meridional advections of weaker westerlies become significant and easterly ageostrophic flow develops.

Over the cold tongue, the effects of introducing zonal asymmetry into the SST forcing appear in both the geostrophic and ageostrophic components of the time-mean zonal wind in the near-equatorial latitudes. The east–west asymmetry in SST ensures that geostrophic winds form westerlies in upper levels, as opposed to easterlies in Kuo–Oct, and easterlies in low levels stronger than those in Kuo–Oct. Additional easterly acceleration occurs at 811 mb over the equator from a zonal pressure gradient that exists because of the Walker circulation. Thus, in this model, a vertically

![Fig. 14. Kuo–Walker cold tongue ageostrophic zonal wind averaged from 60°W to 120°W and then symmetrized about the equator. Contour interval is 0.5 m s$^{-1}$.](image-url)
confined region of easterlies develops into a low-level jet over the equatorial cold tongue because the Walker circulation enhances the low-level easterlies and it caps the flow by (westerly) subsiding air.

3) THE WALKER CIRCULATION

The Walker circulation established in this experiment and represented by the vector winds in Fig. 12 significantly affects the distribution of atmospheric pressure throughout the Tropics. By comparing the time-average departures from the zonal mean for geopotential and horizontal winds over the warm pool and cold tongue in the Kuo–Walker case (Fig. 15) to the zero fields that would result in the zonally symmetric Kuo–Std and Kuo–Oct cases, it is clear that the addition of an equatorial zonal wavenumber one anomaly to the SST forcing field affects both the zonal and meridional pressure distribution in the Tropics and the associated nearly geostrophic zonal winds.

The pressure field adjustments to a Walker circulation forced by an imposed midtropospheric heating anomaly that Gill (1980) found in his model based on the shallow water equations consisted of two parts. First, Kelvin signals propagated east of the heat source and created, at low levels and centered at the longitude of the heating, easterlies with maximum speeds centered on the equator and a pressure trough that was also centered on the equator. Second, Rossby signals propagated west of the heat source and created in the lower troposphere westerlies with maximum speeds on the equator and a pressure ridge on the equator with two

**FIG. 15.** Time-average departures from the zonal mean for geopotential (contours) and horizontal winds (vectors) at (a) 165 mb and (b) 811 mb. Contour interval in (a) is 90 m² s⁻² and in (b) 30 m² s⁻². Maximum vectors in (a) represent 19.2 m s⁻¹ and in (b) 9.5 m s⁻¹.
symmetric low pressure centers off the equator. The resulting pressure and wind field (see Fig. 1b in Gill 1980) indicated that the Kelvin response had a larger spatial scale because it propagated three times faster than the Rossby wave. Similar solutions also appear in Matsuno (1966) and Webster (1981). In the Kuo–Walker simulation forced by an imposed zonal wave-number one SST anomaly in lieu of an elevated heating anomaly, the Kelvin wave response appears to dominate to an even greater extent (Figs. 15a and 15b). In this model, as in the Gill (1980), Matsuno (1966), and Webster (1981) solutions, the surface pressure reaches its lowest values on the equator and roughly 30° east of the center of the heating. Despite subtle discrepancies with the solutions by Gill and others for the pressure and wind fields, the low-level zonal wind and pressure field response in the Kuo–Walker simulation east of the heating corresponds well to the Kelvin response. Also, the Kuo–Walker low-level circulation agrees with the solutions of Gill and others in its production of maximum easterlies at low levels and westerlies at upper levels east of the heating. The Kuo–Walker case, however, shows that the low-level easterly return flow is confined into a jet near 800 mb and is meridionally confined to within the equatorial Rossby radius of the equator (see, e.g., Fig. 15b). One of the fundamental differences between the zonally asymmetric response in the Kuo–Walker case and the idealized Gill model response is found west of the maximum heating (and SST), where at low levels (e.g., 811 mb, Fig. 15b) there is no trace of the Rossby-induced cyclones in the GCM response.

In the previous discussions of the zonal wind over the warm pool and cold tongue, we noted that the zonal pressure gradient significantly affects the zonal winds only where the gradient reaches its largest values. Since the pressure pattern slopes easterward with height (Fig. 12), the longitudes around the centers of the warm pool (90°E) and cold tongue (90°W) include strong zonal pressure gradients only in the lowest levels. Gill’s model, with its two levels, does not account for the eastward slope of the pressure perturbations with height, but the slope can be easily explained. This slope results from the strong difference in tropospheric heating between the warm and cold regions and the location of the lowest surface pressure roughly 30° (20°) east of the center of the warm pool (precipitation maximum). Since everywhere in the vertical the geopotential thickness over the warm pool exceeds that over the cold tongue, and the minimum surface pressure occurs east of the center of the warm pool, the geopotential zonal anomalies must slope eastward with height.

e. Summary

When added to an otherwise symmetrically forced model, a zonally asymmetric tropical SST anomaly causes the model to form a low-level equatorial easterly jet that is similar in speed, scale, and location to that observed (compare Figs. 1 and 2 with Fig. 11). Even with a Walker circulation present, the time-average zonal winds develop, as they do when only the Hadley circulation is present, into predominantly geostrophic flows throughout the tropical troposphere above the boundary layer. In the upper troposphere over the equator, where the zonal winds depart most from the experiments forced by zonally symmetric SSTs, geostrophy explains less of the total zonal flow, as expected, but geostrophic westerlies form over both the cold and warm equatorial regions. Hence, the meridional pressure distribution induced by the Walker circulation differs greatly from the distributions produced by zonally symmetric SSTs.

Over the warm pool, the weak equatorial easterlies (westerlies) at lower (upper) levels reflect a near cancellation of the Walker and zonally symmetric Hadley circulation effects. Over the cold tongue at low levels, on the other hand, the Kelvin response induced by the Walker circulation enhances easterlies that are associated with the Hadley circulation. The greatest enhancement occurs in the longitudes where the easterlies experience accelerations by the zonal pressure gradient, that is, those longitudes just west of the cold tongue center. Aloft, however, the Hadley circulation’s tendency to form easterlies is overwhelmed by the westerly outflow from the convection over the warm pool, which can be described as a modified Kelvin response. This deep layer of westerly flow subsides over the equator, confining the maximum low-level easterly flow to the base of the subsidence near 800 mb. Thus, a low-level equatorial easterly jet forms in this model over the cold tongue only when a zonally asymmetric component is included in the prescribed SST field, ensuring the development of a Walker circulation. We conclude that the upper-level flow and the horizontal and vertical advection of westerly momentum that are required to vertically confine the low-level Hadley-circulation-induced equatorial easterlies into a jet can be provided only by the Walker circulation.

f. Discussion

The conclusion from the zonally symmetric and asymmetric aqua-planet studies that were conducted is that a low-level jet will form over the equator east of the upward branch of a Walker circulation. By examining the CCM1 climatologies, this conclusion can be examined in the same model but with 1) a different convective scheme [the moist convective adjustment scheme of Manabe et al. (1965) in lieu of the Kuo scheme used in the previous experiments], 2) prescribed climatologically varying SSTs, 3) the observed orography and surface types (land, sea, or ice), and 4) seasonal insolation. When averaged over the longitudes that define the average jet in Sadler and Kilonsky’s study, the zonal winds in a meridional cross section
show that an equatorial easterly jet at the third model level, $\sigma = 0.811$, also resembles that found over the cold tongue in the Kuo–Walker integration (Fig. 3a), even though the meridional circulation is much weaker in the Southern Hemisphere Hadley cell (compare Fig. 11 to Fig. 3b). Over the equator, the deep layer of subsidence from the tropopause to nearly 800 mb and the upper-level westerlies, which are both produced by the Walker circulation, provide the vertical confinement for the jet. An off-equatorial jet similar to those found in all the experiments with this model also develops in the FMA climatology at the second model level, $\sigma = 0.926$, and at 10°N; in the longitudes chosen for this zonal average, the southern jet is not well defined.

The zonal wind field averaged over August, September, and October (ASO) from the same climate model control run (CCM1, case 261) appears in Fig. 16. During ASO, the maximum heating associated with the Walker circulation shifts to near 170°E, a strong east Pacific cold tongue forms, a single ITCZ develops north of the equator, and one dominant Hadley cell appears. The Walker circulation in the ASO season also provides a deep layer of subsidence and upper-level westerlies to vertically confine low-level easterlies into a jet over the eastern Pacific cold tongue (Fig. 16). The meridional wind profile (Fig. 16b), when compared to that for the Kuo–Poq case (Fig. 7), suggests how theWalker circulation can sufficiently alter the tropical wind field in a region dominated by a single off-equatorial Hadley circulation to form a low-level equatorial easterly jet: by providing an easterly pressure gradient and a core of subsiding westerlies centered on the equator over the cold tongue. The comparison between the ASO climatology and the Kuo–Poq case also suggests that the southern low-level off-equatorial jet in a region with a Hadley cell that extends across the equator into the Northern Hemisphere can be modified enough by the Kelvin response from a Walker circulation so that it merges with the equatorial jet. Both seasonal climatologies support our conclusion that the Walker circulation is responsible for the low-level equatorial easterly jet east of its upward branch, regardless of the Hadley circulation in that region.

While we have focused on the importance of the Walker circulation for the regional low-level jet, there is a striking difference between the zonally averaged zonal flow in Kuo–Walker and Kuo–Std simulations (Fig. 17). The introduction of a zonally asymmetric forcing in the Kuo–Walker case has a fundamental impact on the zonally symmetric response. Consistent with the analysis of section 4, this impact is greatest at the upper-levels where the zonally symmetric winds are 21 m s$^{-1}$ more westerly in the Kuo–Walker case than in the Kuo–Std case (Fig. 17). The fundamental change in the zonal mean circulation is likely a result of the changes in the eddy activity in the model associated with the localized maximum in diabatic heating. This will give rise to enhanced baroclinic transients in the subtropics that will affect the tropical angular momentum balance. This possibility has been demonstrated and discussed in a two-level primitive equation model of the idealized tropospheric circulation by Saravanan (1990, 1993) and Suarez and Duffy (1992). A similar analysis to demonstrate that the qualitative change in the circulation is indeed driven by the changes in the eddy momentum flux is beyond the scope of this work. However, in our diagnostic studies of the Kuo–Walker case, we have seen nothing to contradict this hypothesis.

5. The moisture field over the cold tongue

a. The "dry tongue"

An intriguing feature of the time-mean moisture field in the Kuo–Walker integration occurs in the lower troposphere over the cold tongue. This feature, a horizontally and vertically confined region of minimum moisture, or a "dry tongue," develops at 910 mb, one layer below the core of the easterly jet (Fig. 18). Its average specific humidity, $q$, at 910 mb over the cold tongue is less than 3 g kg$^{-1}$! Though a relationship between this dry tongue and the easterly jet suggests itself merely from their nearly coincident horizontal and vertical dimensions, the formation of the jet does not depend on the existence of a dry tongue. As described in the previous section, the easterly flow is predominantly geostrophic and therefore linked to the meridional distribution of geopotential with its associated pressure, temperature, and moisture fields. As shown in Fig. 12, the zonal geopotential anomalies meridionally averaged over 4.2°N and 4.2°S display no significant deviation in the dry tongue. (A similar plot in a narrower latitudinal band along the equator at the core of the dry tongue shows an almost identical structure over the cold tongue, confirming that the virtual temperature is not altered significantly by the moisture minimum.) Further evidence that the dynamics of the jet are independent of any dry tongue can be found in the mean boreal spring and fall climatologies that were discussed in section 4f. These climatologies from the standard version of the CCM1, which uses a moist convective adjustment scheme instead of the Kuo convective scheme used in our experiments, reproduce the observed jet (Figs. 3 and 16) but not the localized dry tongue or $q$ minimum (see Ovens 1993).

The climatologies and the Kuo–Walker case indicate that the jet does not depend on the dry tongue and suggest that the dry tongue depends on the details of
the (parameterized) convective physics. The model forms a $q$ minimum only over the cold tongue in the Kuo–Walker case; no $q$ minimum forms over the warm pool or in any of the cases forced by zonally symmetric SSTs discussed in section 3. The zonally and meridionally symmetric cases, though, do show low-level meridional $q$ minimums at the equator; and in the Kuo–Oct case a layer of weaker vertical gradient of $q$ develops from 910 to 811 mb, though $q$ still decreases monotonically with height. Thus, though it may produce a dry tongue over the cold equatorial SSTs in the Kuo–Walker simulation, the Kuo convective scheme does not always produce a dry tongue. To determine how the Kuo scheme forms this dry layer, we first clarify some of the details of the Kuo convective scheme and then we examine all processes that affect moisture in the model.

b. The Kuo convective scheme

The scheme used in this model is derived from Kuo (1965, 1974), Anthes (1977), Donner et al. (1982),...
model that affect moisture. The moisture tendency equation for this model as modified by the Kuo convective scheme can be represented by

$$\frac{\partial q}{\partial t} = -U \cdot \nabla q - \dot{\sigma} \frac{\partial q}{\partial \sigma} + DQCOND$$

$$+ CMF + DQDADJ + F_{q,v} + F_{q,h}, \quad (5.1)$$

where $q$ represents specific humidity; $U$ and $\dot{\sigma}$, the horizontal and vertical velocity, respectively; $F_{q,v}$ and $F_{q,h}$, vertical and horizontal moisture diffusion; $DQCOND$, large-scale condensation; $CMF$, the sources and sinks of moisture due to Kuo convection; and $DQDADJ$, the dry convective adjustment. Here the cumulus moisture forcing, $CMF$, term represents both Kuo moisture effects, namely, net condensation and the eddy vertical flux convergence of moisture (or moisture flux convergence, for short).

A quantitative analysis of all the terms in (5.1) over the cold tongue within $4.2^\circ$ of the equator (Fig. 19) indicates that only the cumulus moisture forcing, or $CMF$, term dries at 910 mb in the time mean. All the other significant terms at 910 mb in Fig. 19 (horizontal advection, vertical diffusion, and vertical advection) act to moisten this layer. This result agrees with some of the characteristics of the Kuo scheme pointed out by previous authors: Anthes (1977), Donner et al. (1982), and Donner (1986) noted that one of the major effects of this Kuo-type cumulus parameterization is to dry the lower troposphere, particularly in the Tropics under deep cumulus clouds. The low-level drying, however, occurs over the cold tongue in the Kuo–Walker case in a region where subsidence prevails above 800 mb and nearly one-half of the convective events cap out below 500 mb. Of the two terms that make up the Kuo $CMF$ term in (5.1), the moisture flux convergence dries roughly five times more effectively than the cloud condensation process in the dry tongue. The moisture flux convergence also accounts for the strong moistening that occurs at the next model level, 811 mb (see Fig. 19). Though both condensation and moisture flux convergence dry the 910-mb level close to the equator in the cold tongue region, they dry that level even more at higher latitudes. Therefore, the dry tongue develops at 910 mb over the equator due to the drying by the Kuo scheme and due to the moistening contrasts between the equator and higher latitudes. The largest moistening differences between the equator and higher latitudes at 910 mb occur in the dry and moist convective adjustment terms: significantly smaller moistening occurs over the equator due to higher stability there. This higher stability is indicated by the deep layer of

c. Dry tongue causes

To determine what maintains the dry tongue in the Kuo–Walker case, we examine all of the terms in the

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4 Refer to HBR93 for a summary of the criteria utilized in the Kuo scheme used in these experiments.

5 The calculations for this analysis were made by determining the value of each term at every available time and then averaging these values over the last 120 days of the integration.
subsidence over the equator (Fig. 11) and the underlying relatively cold SST (Fig. 4).

d. The observed moisture field in the vicinity of the low-level easterly jet

As discussed in section 5a, while the convective parameterization does not affect the dynamics of the jet in this model, it does significantly affect the moisture profile in the jet region by producing a dry tongue one model level below the jet core. Does any dry layer accompany the jet in the observed atmosphere, however? Betts and Albrecht (1987) and Kloesel and Albrecht (1989) reported finding a vertical specific humidity minimum such as that shown in Fig. 20 in nearly 70% of the more than 2000 drop-windsondes and soundings over the central and eastern Pacific (from 5°S to 15°N, 170°E to 90°W) that they studied. The average q minimum in this region, which corresponds to the cold tongue region in the Kuo–Walker case, was found near 810 mb and was capped by a layer of air containing roughly 2 g kg⁻¹ more moisture. Kloesel, Betts, and Albrecht proposed that this q minimum forms due to convection whose updrafts penetrate an inversion and moisten the levels above that inversion and whose dry downdrafts sink down and spread out at the top of the inversion as a dry layer. While Betts, Kloesel, and Albrecht did not discuss their results in the context of the large-scale flow field, based on the climatol-
6. Conclusions

Zonal wind observations in the central and eastern equatorial Pacific indicate that a low-level easterly jet forms in the time-average for most months of the year. Since this jet, characterized by easterly wind speeds in excess of 10 m s⁻¹ between 900 and 850 mb and confined within an equatorial Rossby radius of the equator, was so prevalent in the six years’ worth of data that Sadler and Kilonsky (1981) reported, it would have been influenced by the most robust time-mean features of the tropical circulation, the Hadley and Walker circulations. Previous studies of the Hadley and/or Walker cells using the shallow water equations, linearized primitive equation models, 3D GCMs forced by zonally symmetric SSTs, or other simplified models (see, e.g., Matsuno 1966; Gill 1980; Geisler 1981; Webster 1981; HBR93; Numaguti and Hayashi 1991; Held and Hou 1980; Schneider and Lindzen 1977; Schneider 1977) did not develop a low-level equatorial easterly jet, although several studies produced easterlies over the equator at low levels with speeds comparable to the observed jet. These studies had zonally symmetric forcing and geometry and no orography, precluding a Walker-type zonal circulation from forming, or they focused solely on the Walker circulation and did not allow it to interact with a separately forced Hadley circulation.

To simulate the low-level easterly jet and to determine the dynamics responsible for maintaining it in the presence of both the zonal and meridional overturning cells, a simple zonally asymmetric experiment for a GCM was designed. The model used for this study, the
CCM1, was modified to include a Kuo convective scheme, equinoctial insolation, and a lower boundary covered by water only, and it was run with 12 vertical sigma levels and with T42 spectral truncation, giving it an equivalent horizontal grid spacing of roughly 2.8° in latitude and longitude.

To understand how the Hadley circulation, isolated from any time-average Walker circulation, affects the zonal winds in the model's equatorial region, several of the experiments from HBR93 were reanalyzed. Each case was forced by zonally symmetric SSTs with notably different meridional profiles, and the tropical winds that developed in these cases showed two significantly different time-mean Hadley circulations. In each of these integrations, however, easterly winds developed in low latitudes throughout the depth of the troposphere in those regions where low-level meridional winds associated with the Hadley cell(s) flowed equatorward. None of the experiments produced a low-level tropical jet over the equator; however, an easterly jet formed in each case at 910 mb near the center of the trade winds between roughly 10° and 15° latitude. Above the boundary layer, the zonal winds in the Tropics were primarily geostrophic. The near-equatorial pressure field was linked to the latent, radiative, and adiabatic heating contrasts between the equator and ITCZ(s) and by the meridional circulation in the Hadley cell(s). Though equatorial upper-level westerlies did not form in any of these experiments, they could form in this model with zonally symmetric SST forcing if the forcing produced a single, sharp ITCZ on the equator (see the Kuo–Peq case of HBR93), facilitating the transport of angular momentum into the Tropics by the eddies (e.g., Saravanan 1990). The upper-level westerlies would not confine the low-level easterlies into a jet, however, because the upward motion in the ITCZ over the equator would advect momentum vertically and ensure efficient vertical mixing of zonal momentum. Hence, the Hadley circulation, though it can produce equatorial easterlies, cannot vertically confine them into a low-level equatorial easterly jet.

To include a Walker circulation in the same model used for the Hadley circulation simulations, an experiment was designed in which the SST distribution near the equator included a zonal wavenumber one anomaly with a peak to peak amplitude of 5°C. The meridional e-folding widths of the anomaly in the warm and cold regions were 12° and 6°, respectively. With this SST field and equinoctial insolation forcing the aqua-planet model, a low-level equatorial easterly jet, comparable in speed and horizontal and vertical scale to the observed, develops east of the warm pool's convective center and over the cold tongue. The dominant, geostrophic component of the zonal flow in the jet results from both the Hadley circulation and the Walker-circulation-induced Kelvin wave meridional pressure distribution; ageostrophic easterlies associated with the zonal pressure distribution also enhance the easterly flow in the jet. The upper-level diverging and subsiding branches of the Walker circulation provide the essential vertical confinement to create an upper boundary for the jet, while surface friction creates a lower boundary.

In CCM1 climatologies that incorporate land surfaces, a moist convective adjustment scheme for convection, and seasonally varying insolation and SSTs, the Walker circulation forms a low-level equatorial easterly jet over the central to eastern Pacific during two seasons characterized by extremes in the meridional circulation. Together the experiments forced by zonally symmetric SSTs, the Kuo–Walker integration, and the CCM1 climatologies support our hypothesis that the mechanism to vertically confine the low-level equatorial easterly flow into a jet results from a Walker but not a Hadley circulation and that this jet should be a robust feature of the tropical eastern Pacific circulation. Each of these integrations also indicates that a low-level (910 mb) off-equatorial jet (or jets) develops in this model at roughly 10° to 15° latitude on the poleward side of an ITCZ with any meridional or zonal circulation. In the case of a single, dominant ITCZ, the jet in the opposite hemisphere moves closer to the equator, and since it is enhanced near the equator by strong easterly flow from the Walker circulation, this jet merges with the equatorial jet.

Though the CCM1 climatologies and our experiments imply that the dynamics of the low-level easterly jets on and off the equator are determined by large-scale dynamics and are likely to be insensitive to the unresolved, or parameterized, physics in the model, they indicate that the concomitant tropical moisture field, on the other hand, strongly depends on unresolved physics. An extremely dry layer along the equator and one model level below the jet develops in the Kuo–Walker circulation due to the Kuo convective scheme. Drying from the eddy vertical-flux convergence of moisture, which is the term in the Kuo scheme that accounts for the unresolved redistribution of cloud moisture by cloud updrafts and compensating downdrafts, and the lack of moistening by dry and moist convective adjustment processes cause the dry tongue to form in the Kuo–Walker integration. The same mechanism likely creates the meridional moisture minimums at the same level over the equator in the zonally symmetric cases; however, further analyses of the Kuo scheme and the large-scale convective environments are required to determine why the zonally symmetric case with an analogous meridional circulation to that over the cold tongue does not form a vertical moisture minimum at 910 mb.

In the majority of moisture profiles collected in the region of the low-level easterly jet, that is, the central to eastern Pacific, a low-level vertical moisture minimum was noted by Betts and Albrecht (1987) and Kloesel and Albrecht (1989). They proposed that this dry layer and its overlying layer of more moist air are formed by convection in which updrafts penetrate and
moisten the levels above an inversion, and dry downdrafts spread out at the top of the inversion as a dry layer. Their proposed mechanism requires an inversion and shallow clouds that are not necessarily present in the Kuo scheme. In addition, the model dry tongue occurs below the jet core, whereas the observed dry layer likely appears at or above the jet core. Hence, though both the model and the observations indicate the existence of a low-level dry layer in the vicinity of the equatorial easterly jet, it is not clear if the mechanisms responsible for this dry layer are the same.

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APPENDIX

Near-Equatorial Friction-Layer Depths

In the model used for all the experiments discussed in this paper, the CCM1, vertical stresses on a field, $A$, are parameterized by $-\rho K \partial A/\partial z$, where $K$ is the eddy viscosity. In this $K$-theory parameterization of vertical stresses, $K$ is allowed to vary with stability and wind shear. To explain why the depth of the friction layer increases with decreasing latitude, it is simplest to assume that $K$ and the geostrophic wind are constant so that Ekman theory applies. In Ekman theory, the depth of the friction layer is given by $\pi(2K/f)^{0.5}$. As the flow nears the equator, the Coriolis parameter, $f$, approaches zero, and the Ekman friction layer depth increases. However, $K$ and the wind speed vary in this model, so simple Ekman theory does not apply.

A useful diagnostic quantity to indicate the depth of the friction layer in these experiments is the amplitude of the vertical friction term $-1/\rho \partial /\partial z (-\rho K \partial /\partial z) U$. If the depth of the friction layer is shallow, then the vertical friction term will have large values in the lowest layers and then decrease rapidly to much smaller values with height, since the friction effects would be strongest in a shallow, low layer. In a deeper friction layer, the low-level vertical friction term would be smaller and it would not decrease as rapidly. Thus, in calculating this term for various latitudes at various levels, it is helpful to normalize the value at each level by dividing it by the value at the lowest level for that latitude. The resulting vertical profiles of the vertical friction term at various latitudes indicate a deeper (shallower) friction layer by a more (less) gradual decrease in magnitude with height. A representative plot for the meridional variation of the vertical friction term at low levels from the Kuo–Walker experiment is shown in Fig. A1. In this figure, the friction-layer depth in the latitude band near the equator over the cold tongue and warm pool regions in the Kuo–Walker experiment is notably deeper than that at higher latitudes.

REFERENCES


