On Mean Meridional Circulations in the Tropics

STEFAN L. HASTENRATH

Dept. of Meteorology, The University of Wisconsin, Madison

(Manuscript received 28 March 1968, in revised form 15 June 1968)

ABSTRACT

Various models of the mean meridional circulations are examined in terms of the vorticity equation. These include: 1) the traditional model of two Hadley cells joining in a region of ascending motion near the equator; 2) a model with two "equatorial cells" enclosed between the traditional Hadley cells, implying subsidence over the equator; and 3) a model with a single "equatorial cell" enclosed between the two Hadley cells, with ascending motion on one side, and subsidence on the other side of the equator. All three models are considered compatible with the vorticity equation, under specific conditions of latitudinal and vertical distributions of absolute vorticity, frictional force, and vertical and zonal wind components. The existence of twin equatorial cells requires the eastward directed frictional force in the lower layers to decrease from the kinematical equator poleward, within the subsiding portion of either of the two equatorial cells. It is suggested that this may be related to the existence of (eastward directed) equatorial undercurrents in the Pacific and Atlantic Oceans. The requirements for twin equatorial cells do not seem to be met over the interior of the tropical continents.

1. Introduction

Satellite observations and large-scale field experiments in the tropics during recent years are again drawing attention to the atmospheric circulation in equatorial latitudes. In a recent issue of this JOURNAL, Asnani (1968) presented an interesting contribution to this topic, discussing various models of the mean meridional circulation in the light of the vorticity equation. Asnani uses the vorticity equation in the form

$$\frac{d\eta}{dt} = -\eta\nabla\cdot\mathbf{V}, \qquad (1)$$

that is, the twisting, solenoidal, and frictional terms are disregarded. Following conventional notation, **V** signifies the horizontal wind vector, $\zeta=(\partial v/\partial x)-(\partial u/\partial y)$ the vertical component of relative vorticity (hereafter referred to as relative vorticity), f the Coriolis parameter, and $\eta=(f+\zeta)$ the vertical component of absolute vorticity (hereafter referred to as absolute vorticity).

Asnani discusses the conventional model in a meridional vertical plane of two Hadley wheels joining in a region of ascending motion in equatorial latitudes, for the case of a single minimum of $|\eta|$ near the equator, where absolute vorticity changes sign; this would correspond to the "kinematical equator" in Kruger's (1960) terminology. Asnani finds this model (Fig. 1 in Asnani's paper, hereafter referred to as model I) incompatible with Eq. (1). A model of two indirect cells enclosed between the conventional Hadley cells, with subsidence over the equator (Fig. 2 in Asnani's paper, hereafter referred to as model II), has been suggested by Fletcher (1945), Bryson (1948) and Rossby (1949). Asnani considers this model also to be in contradiction with Eq. (1). For better orientation, this model II is schematically represented in Fig. 1. Asnani finally proposes two alternate models (Figs. 3a and 3b in Asnani's paper, hereafter referred to as models IIIa and IIIb), which should satisfy Eq. (1). These models include two maxima of $|\eta|$ on either side of the equator, coinciding with the equatorward limits of the conventional Hadley cells. A single circulation wheel straddling the equator is enclosed by the two maxima of $|\eta|$; ascending motion is implied on one side, and subsidence on the other side of the equator, with zero vertical motion at the equator itself.

For the subsequent discussion it appears desirable to obtain an idea of the characteristic latitudinal distribution of absolute vorticity. For a treatment in a mean meridional-vertical plane, the relative vorticity reduces to $\zeta=-\partial u/\partial y$. Representative values for the 900- and 200-mb levels during the winter and summer seasons

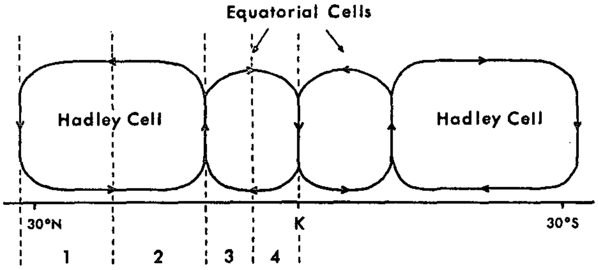

FIG. 1. Model II of mean circulation in a meridional-vertical plane (not to scale), showing two "equatorial cells" enclosed between the traditional Hadley cells, with subsidence over the (kinematical) equator K. (c.f. Fletcher, 1945; Bryson, 1948; Rossby, 1949). Regions 1, 2, 3 and 4 refer to Table 2.

TABLE 1. Latitudinal distribution of Coriolis parameter f and latitudinal and height variations of absolute vorticity $\eta = f - \partial u/\partial y$, all units 10^{-5} sec^{-1} (from Mintz, 1954).

	Latitude, north						
	0°	5°	10°	15°	20°	25°	30°
f	0	0.64	1.26	1.89	2.49	3.08	3.64
				η, winter			
200 mb	−0.5	+0.1	+0.5	−1.1	−0.5	+1.6	+4.1
900 mb	+0.1	+1.1	+1.3	+1.4	+1.8	+2.9	+3.4
				η, summer			
200 mb	+0.5	+0.7	+1.2	+1.2	+1.5	+2.4	+2.8
900 mb	−0.3	+0.6	+1.3	+2.0	+2.5	+3.0	+3.3

were estimated from Mintz' (1954) mean meridional cross sections, for various latitudes between equator and 30N. Estimates of absolute vorticity thus obtained are given in Table 1. In general, Table 1 shows a monotonic increase of absolute cyclonic vorticity from the equator poleward; at the 200-mb level in winter, however, anticyclonic vorticity is indicated for latitudes 15 and 20N, due to the strong shear south of the westerly jet stream. The mean meridional cross sections of Mintz (1954) are not detailed enough to identify the maxima of $|\eta|$ near the equator, suggested by Asnani. These maxima would require values of the relative cyclonic vorticity between about 0.5 and 1×10^{-5} sec^{-1}, well equatorward of 5°. From the maps presented by Sandoval (1967) for the tropical western Pacific, this would appear possible even on a climatic time scale.

2. Discussion in terms of the simplified vorticity equation

a. Model I. Asnani considered model I only for the case of a single minimum of $|\eta|$ at the kinematical equator. If we assume a single maximum of $|\eta|$ near the equator, for example in the Northern Hemisphere (as realistic an assumption for certain longitudes as Asnani's two maxima of $|\eta|$), then the Northern Hemisphere Hadley wheel satisfies Eq. (1) in the same way as in Asnani's model IIIa. On its poleward side, i.e., north of the relative minimum of $|\eta|$, convergence in the upper layers would produce the correct sign for the right-hand term of Eq. (1) in the case of positive absolute vorticity. Convergence in the lower and divergence in the upper layers would be required between the minimum and the maximum of positive absolute vorticity. The Southern Hemisphere Hadley wheel would satisfy Eq. (1) in a similar way as Asnani's "equatorial cell" in model IIIa. Lower-layer convergence with upper-layer divergence are required between the kinematical equator and the maximum of positive absolute vorticity. The pattern is reversed between the kinematical equator and the southern flanks of the Hadley cell. Similar considerations apply for a single maximum of $|\eta|$ near the equator, but in the Southern Hemisphere.

b. Model II. Model II does indeed present difficulties with Eq. (1), for all three latitudinal distributions of absolute vorticity considered here; namely, a minimum of $|\eta|$ at the kinematical equator, two maxima, or a single maximum of $|\eta|$ near the equator.

c. Model III. It appears appropriate to examine the model proposed by Asnani, for example, alternative IIIa, in the light of the latitudinal distribution of absolute vorticity brought out by Table 1 which indicates anticyclonic absolute vorticity in the upper layers and on the poleward flank of the Hadley cell. Under such circumstances, model III would be incompatible with Eq. (1). Similar reservations apply to our discussion of model I for the case of a single maximum of $|\eta|$ near the kinematical equator.

3. Use of the complete vorticity equation

a. General theory. Following Asnani, we have so far used the vorticity equation only in the very simplified form of Eq. (1). This form, however, may not be adequate for the problem discussed here. Starting from the complete form of the vorticity equation, and disregarding all terms involving a derivative with respect to x for reasons of zonal symmetry implied here, one obtains

$$\frac{d\eta}{dt} = -\eta \nabla \cdot \mathbf{V} + \frac{\partial w}{\partial y}\frac{\partial u}{\partial z} - \frac{\partial F_x}{\partial y}, \qquad (2)$$

where F_x is the frictional force per unit mass acting in the x direction and includes turbulent effects. Orientation of the x, y, z axes and the wind components u, v, w follow conventional notation. In contrast to Eq. (1), Eq. (2) contains two additional terms on the right-hand side. The second and third right-hand terms can only be disregarded if they are small compared to the divergence term. However, as will be seen later, this cannot be generally assumed in equatorial latitudes, where η becomes small. The various models of the mean meridional circulation should therefore be considered in relation to Eq. (2), rather than the simplified Eq. (1).

It was mentioned above using Eq. (1) that models I and IIIa conflicted with the latitudinal distribution of η at the 200-mb level during winter, as shown in Table 1. This discrepancy could conceivably be accounted for by the second right-hand term of Eq. (2). Model I has been considered above in relation to Eq. (1) for the case of a single maximum of $|\eta|$ near the equator; vertical motion was found to change sign at the kinematical equator. More realistically, the second and third right-hand terms of Eq. (2) should be taken into account; it is then seen that the region with lower-layer convergence and upper-layer divergence may extend from the maximum of $|\eta|$ well beyond the kinematical equator.

b. Model I. Model I with one minimum of $|\eta|$ at the kinematical equator, as discussed by Asnani, should be reconsidered in the light of Eq. (2). For reasons of symmetry, the discussion can be limited to the region north of the kinematical equator. Conditions on the poleward side of the Hadley cell offer no problems in relation to Eq. (1) except for the latitudinal distribution of absolute vorticity at the 200-mb level during winter, as indicated above. A more critical situation arises on the equatorward side of the Hadley cell, where air flows toward smaller values of cyclonic vorticity in the lower, and toward larger values in the upper layer.

Since frictional effects can be considered comparatively small in the upper troposphere, this would essentially require a positive sign for the second right-hand term of Eq. (2) in the upper layers. Typically, upward motion decreases poleward within the Hadley cell, and an east wind maximum appears to be characteristic of the upper troposphere and lower stratosphere in equatorial latitudes (e.g., Mintz, 1954; Palmer *et al.*, 1955; Joint Task Force Seven Meteorological Center, 1960–1961). This would indeed result in a positive sign for the second right-hand term of Eq. (2). Available meridional transects (e.g., Mintz, 1954) indicate that the increase of easterlies with height in the upper layers is of the order of 2 m $\sec^{-1}$ (2 km)$^{-1}$. Considering a meridional variation of vertical velocity of 5 mm $\sec^{-1}$ (100 km)$^{-1}$, the twisting term is found to be of the order of 0.5×10^{-10} $\sec^{-2}$. This should be compared to the magnitude of the divergence term. As seen from Table 1, absolute vorticity is of the order of 1×10^{-5} $\sec^{-1}$ in 5–10° latitude region, while it drops to less than 0.3×10^{-5} $\sec^{-1}$ within 2° from the equator. Even for divergence values as large as 50×10^{-7} $\sec^{-1}$, the divergence term would thus be only of the order of 0.5×10^{-10} $\sec^{-2}$ at 5–10°, and drop to less than 2×10^{-11} $\sec^{-2}$ within 2° from the equator. This shows that in equatorial latitudes the twisting term is by no means negligible when compared with the divergence term.

In the lower layers, the right-hand side of Eq. (2) must be negative. The meridional variation of vertical velocity can be thought of as having the same general magnitude as in the upper layers where the return flow is concentrated. However, available meridional transects (e.g., Mintz, 1954; Palmer *et al.*, 1955) suggest that the increase of easterlies with height tends to be much smaller in the lower layers. Monthly radiosonde averages for equatorial stations (Joint Task Force Seven Meteorological Center, 1960–1961; U. S. Navy Hydrographic Office, 1944) frequently indicate even an upward decrease of easterlies across the surface layer. The twisting term in the lower layers would thus seem to be positive, but much smaller than in the upper layers, or it may even be negative.

Frictional effects must be considered in the lower troposphere. The last term on the right-hand side of Eq. (2) would give a contribution of the proper sign, if it were negative. This would correspond to an increase of the eastward directed frictional force with latitude. Other things being equal, this could result from an increase of the easterly wind component in the lower layers, as one proceeds from the kinematical equator poleward. Available meriodional cross sections of zonal winds (e.g., Mintz, 1954; Palmer *et al.*, 1955) indicate that this indeed tends to be the case.

TABLE 2. Sign of the various terms of Eq. (2) in characteristic regions (1, 2, 3 and 4) of model II, as specified in Fig. 1. Bold print denotes the dominant terms.

	Regions			
	1	2	3	4
Upper layers				
$d\eta/dt$	**+**	**+**	**−**	**−**
$-\eta\boldsymbol{\nabla}\cdot\mathbf{V}$	**+**	−	**−**	+
$(\partial w/\partial y)(\partial u/\partial z)$		**+**		**−**
Lower layers				
$d\eta/dt$	**−**	**−**	**+**	**+**
$-\eta\boldsymbol{\nabla}\cdot\mathbf{V}$	**−**	+	**+**	−
$(\partial w/\partial y)(\partial u/\partial z)$		+(−)		−(+)
$-\partial F_x/\partial y$		**−**		**+**

Only very crude estimates appear possible regarding the magnitude of the frictional term in Eq. (2). If we assume that the surface stress affects the lower 100 mb of the atmosphere, we may use the following relationship between surface stress τ_0, drag coefficient C_D, air density ρ_0, and wind near the surface V_0, i.e.,

$$\tau_0 = C_D\rho_0 V_0^2, \tag{3}$$

where C_D may be of the order of $1\text{–}10\times10^{-3}$ for various natural surfaces (e.g. Priestley, 1959), and ρ_0 is of the order of 10^{-3} gm cm^{-3}. With values of $C_D=2\times10^{-3}$, and $u_0=-5$ m $\sec^{-1}$, the frictional force per unit mass $F_x=5\times10^{-3}$ cm $\sec^{-2}$. The meridional transects presented by Mintz (1954) and Palmer *et al.* (1955) indicate that the easterly wind component in the lower layers in equatorial latitudes may vary from 2–5 m $\sec^{-1}$ over a distance of 5° latitude. The frictional term in Eq. (2) would thus have the magnitude of 10^{-10} $\sec^{-2}$, and is therefore by no means negligible compared with the divergence and twisting terms. This shows that model I for the case of a single minimum of $|\eta|$ at the kinematical equator can indeed be compatible with Eq. (2).

c. Model II. Satellite cloud photographs in recent years (e.g., Kornfield *et al.*, 1967) and field observations during the Line Islands Experiment in spring 1967 suggest a pattern of meridional circulations over the equatorial Pacific of the type of model II, proposed two decades ago by Fletcher (1945), Bryson (1948) and Rossby (1949). This model was found incompatible with Eq. (1), but it should be reexamined in the light of Eq. (2). For reasons of symmetry with respect to the kinematical equator, only two latitudinal distributions of absolute vorticity have to be considered; namely, a single minimum of $|\eta|$ at the kinematical equator itself, and two maxima of $|\eta|$ on either side of the kinematical equator, for convenience coinciding with the equatorial limits of the Hadley cells. Also, the discussion can be limited to one hemisphere, say the region north of the kinematical equator. For better orientation, the sign combination of the various terms of Eq. (2) in characteristic regions of model II, with a single minimum of $|\eta|$ at the kinematical equator, is schematically summarized in Table 2.

First consider only the domain of the Hadley cell. The case where its equatorial limit coincides with a maximum of $|\eta|$ has, in principle, been discussed by Asnani, and it was found compatible even with the simplified Eq. (1). In the case of a monotonic decrease of $|\eta|$ towards the kinematical equator, conditions within the Hadley cell correspond to model I with one minimum of $|\eta|$ at the kinematical equator; conditions which would satisfy Eq. (2) have been discussed in Section 3b.

Most interesting is the circulation wheel between the equatorward flank of the Hadley cell and the kinematical equator. For this region, the two latitudinal distributions of $|\eta|$ specified above are identical; namely, $|\eta|$ decreases towards the kinematical equator. In the poleward portion of this "equatorial cell," motion in the direction of increasing η in the lower layers is associated with convergence, and motion toward smaller η in the upper layers with divergence; this is compatible even with the simplified Eq. (1).

In the subsiding, equatorward portion of the "equatorial cell," conditions are more critical and should be considered in relation to the complete Eq. (2). In the upper layers the right-hand side of Eq. (2) must be negative. This must be achieved essentially by the second right-hand term, since the divergence term is positive, and the frictional effects can be considered comparatively small in the upper layers. Within the "equatorial cell," ascending motion increases poleward and, as indicated above, there is a tendency for the easterly wind component to increase towards the upper troposphere and lower stratosphere, in the equatorial belt. This would indeed make the second right-hand term of Eq. (2) negative. Its general magnitude has been estimated in connection with model I to be of the order of 0.5×10^{-10} sec^{-2}.

However, an upward increase in easterlies does not seem to prevail in all equatorial regions and at all seasons. Aerological soundings over the equatorial Pacific performed mainly in connection with atomic tests (Joint Task Force Seven Meteorological Center, 1960–1961; Ramage, 1960), as well as experience during the Line Islands Experiment, indicate that upper tropospheric westerlies may extend into equatorial latitudes, this apparently being most pronounced over the western Pacific and in late winter. The twisting term in the upper layers would consequently be negative. In this respect, the development of meridional circulations should be studied in relation to the zonal wind regime in different equatorial regions and during specific synoptic situations. Also, the vertical extent of the 'equatorial cells" is of interest. However, upper-air data presently available do not appear to be adequate for a detailed study.

It can be realized that both models I and II can be compatible with Eq. (2) even for the case of upper-tropospheric westerlies in low latitudes. However, then the latitudinal and vertical distribution of absolute vorticity is less simple.

In the lower layers, the right-hand side of Eq. (2) must be positive. As discussed in Section 3b, the twisting term in the lower layers seems to be $<0.5\times10^{-10}$ sec^{-2}; with an upward decrease of easterlies across the surface layer, as indicated by various radiosonde stations in the equatorial Pacific (Joint Task Force Meteorological Center, 1960–1961; U. S. Navy Hydrographic Office, 1944), the twisting term would even be positive.

Considering the magnitude and sign of the twisting term in the lower layers, it would seem that the frictional term should play a major role in making the right-hand side of Eq. (2) positive. It should be noticed that the subsiding portion of the "equatorial cell" may lie within only 2° latitude from the equator, where the absolute vorticity $<0.3\times10^{-5}$ sec^{-1}. A positive sign in the last right-hand term of Eq. (2) would require the eastward directed frictional force to decrease poleward. It will be recalled that the inverse latitudinal variation was required in model I, in the equatorward portion of the Hadley cell.

In this connection, the Pacific equatorial undercurrent (e.g. Cromwell *el al.*, 1954; Knauss, 1960) appears to deserve particular attention. The Cromwell current is described as straddling the equator between 2N and 2S, with a maximum eastward velocity of about 1 m sec^{-1} at a depth of about 100 m. This undercurrent can be thought of as exerting an eastward directed stress on the lowest layer of atmosphere (c. f. also Lettau)[1]. For a crude quantitative estimate, assume again that surface friction affects the lower 100 mb of the atmosphere. Knauss (1961) found the velocity shear $\partial u/\partial z$ above the core of the Cromwell current to be of the order of 2×10^{-2} sec^{-1}, and Charney (1960) worked with a kinematic eddy viscosity of 10–100 cm^2 sec^{-1}. From this, the stress on the surface, $\tau=(\nu\rho)\partial u/\partial z$, is crudely estimated to be of the order of 0.2–2 gm cm^{-1} sec^{-2}, and the eastward directed frictional force F_x per unit mass, acting on the lower atmosphere, to be of the order of 2×10^{-2} to 2×10^{-3} cm sec^{-2}. Considering the latitudinal extent of the Cromwell current, this yields for the third right-hand term of Eq. (2) in this region a magnitude of 10^{-9}–10^{-10} sec^{-2}, giving the required positive contribution. The frictional term can therefore not be neglected compared with the divergence and twisting terms; in fact, it appears to play a major role in the origin of the "equatorial cells."

---

[1] Lettau, H., 1967: Physical coupling between the dry-belt of the lower atmosphere and the Cromwell current of the upper ocean along the Pacific equator. Paper presented at the Fifth Tech. Conf. on Hurricanes and Tropical Meteorology, Caracas, Venezuela.

4. Concluding remarks

Asnani (1968) has shown that his models IIIa and IIIb satisfy the simplified vorticity equation (1), when specific assumptions on the meridional and vertical distribution of absolute vorticity are made. The traditional model I of two Hadley cells joining in a region of upward motion near the equator was found compatible with the complete form of the vorticity equation (2); requirements with regard to the latitudinal distribution of absolute vorticity and the twisting and frictional terms do not seem very restrictive. The simplest latitudinal distribution of absolute vorticity appears to be that of a single minimum of $|\eta|$ at the kinematical equator. The existence of a single maximum or of two maxima of $|\eta|$ in the lower layers near the equator, as assumed by Asnani, is considered possible. However, some difficulty arises inasmuch as a similar latitudinal pattern is required for the upper layers.

Model II with two "equatorial cells" enclosed between the traditional Hadley cells, and subsidence over the (kinematical) equator, is found possible for two latitudinal distributions of absolute vorticity; namely, two maxima of $|\eta|$ on either side of the kinematical equator and a single minimum of $|\eta|$ at the kinematical equator itself.

An important point in this model II of the mean meridional circulation is the decrease of the eastward directed frictional force from the kinematical equator poleward, within the subsiding portion of either of the two "equatorial cells." It is suggested for the equatorial Pacific, that this could be related to the Cromwell current. The existence of an equatorial undercurrent has been reported also for the Atlantic and Indian Oceans (Neumann, 1960; Metcalf *et al.*, 1962; Metcalf and Stalcup, 1967).

Daily mosaics of ESSA III and ESSA V satellites during 1967 show a belt of scarce cloudiness right over the equator, across a great part of the Pacific Ocean. This belt appears also on monthly composite photographs (Lettau, *loc. cit.*; Kornfield *et al.*, 1967) Interestingly, in the daily mosaics of ESSA III and ESSA V, a belt of scarce cloudiness right over the equator can occasionally be recognized also over the Atlantic, and much weaker and less frequently also over the Indian Ocean. Contrarywise, an equatorial dry zone does not seem to be well developed over the interior of the tropical continents. This would agree with constraints imposed by the vorticity equation, which indicate that the twin "equatorial cells" may be a phenomenon limited to certain oceanic areas.

It will be noted that Eqs. (1) and (2) were considered only for zonal symmetry and steady state; the possible effect of space and time correlations is thus not accounted for. With these restrictions, the present discussion suggests that all three meriodional circulation models are possible in terms of the vorticity equation, under specific conditions of latitudinal and vertical distributions of absolute vorticity, frictional force, and vertical and zonal wind components. More observational evidence would thus seem desirable for examining where the respective conditions are satisfied, and whether all three meridional circulation patterns do exist in various equatorial regions.

Acknowledgments. This study was supported through National Science Foundation Grant GA-1010. Comments by G. C. Asnani, Institute of Tropical Meteorology, Poona, India, and R. A. Bryson, The University of Wisconsin, are gratefully acknowledged.

REFERENCES

Asnani, G. C., 1968: The equatorial cell in the general circulation. *J. Atmos. Sci.*, **25**, 133–134.

Bryson, R. A., 1948: On disturbances in the easterlies. Ph.D. thesis, University of Chicago, 104 pp.

Charney, J. G., 1960: Non-linear theory of a wind-driven homogeneous layer near the equator. *Deep-Sea Res.*, **6**, 303–310.

Cromwell, T., R. B. Montgomery and E. D. Stroup, 1954: Equatorial undercurrent in Pacific Ocean revealed by new methods. *Science*, **119**, 648–649.

Fletcher, R. D., 1945: The general circulation of the tropical and equatorial atmosphere. *J. Meteor.*, **2**, 167–174.

Joint Task Force Seven Meteorological Center, 1960–1961: Mean upper tropospheric circulation over the tropical Pacific. JTFMC TP-19, Pearl Harbor, Hawaii, Vols. 1–4, 1956–59.

Knauss, J. A., 1960: Measurements of the Cromwell current. *Deep-Sea Res.*, **6**, 265–286.

——, 1961: The structure of the Pacific equatorial countercurrent. *J. Geophys. Res.*, **66**, 144–155.

Kornfield, J., A. F. Hasler, K. J. Hanson and V. E. Suomi, 1967: Photographic cloud climatology from ESSA III and V computer produced mosaics. *Bull. Amer. Meteor. Soc.*, **48**, 878–883.

Kruger, E., 1960: Generalized gradient wind equations and contour analysis in the tropics. *Proc. Symp. Tropical Meteorology in Africa*, 1959, Munitalp Foundation, Nairobi, 168–176.

Metcalf, W. G., and M. C. Stalcup, 1967: Origin of the Atlantic equatorial undercurrent. *J. Geophys. Res.*, **72**, 4459–4975.

——, A. D. Voorhis and M. C. Stalcup, 1962: The Atlantic equatorial undercurrent. *J. Geophys. Res.*, **67**, 2499–2508.

Mintz, Y., 1954: The observed circulation of the atmosphere. *Bull. Amer. Meteor. Soc.*, **35**, 208–214.

Neumann, G., 1960: Evidence for an equatorial undercurrent in the Atlantic Ocean. *Deep-Sea Res.*, **6**, 328–334.

Palmer, C. E., C. W. Wise, L. J. Stempson and G. H. Duncan, 1955: The practical aspects of tropical meteorology. Air Force Surv. Geophys. No. 76, AFCRC-TN-55-460, Air Force Cambridge Research Center, Bedford, Mass.

Priestley, C. H. B., 1959: *Turbulent Transfer in the Lower Atmosphere.* University of Chicago Press, 130 pp.

Ramage, C. S., 1960: Notes on the meteorology of the tropical Pacific and southeast Asia. Air Force Surv. Geophys. No. 126, AFCRC-TN-60-455, Air Force Cambridge Research Center, Bedford, Mass., 158 pp.

Rossby, C. G., 1949: On the nature of general circulations of the lower atmosphere. *Atmospheres of the Earth and Planets*, University of Chicago Press, 16–48.

Sandoval, A. R., 1967: Background studies for a climatology of the intertropical convergence zone in the western central Pacific area. Ph.D. thesis, University of Wisconsin, 127 pp.

U. S. Navy Hydrographic Office, 1944: Weather summary for naval air pilot, central Pacific Hawaiian Islands area. Washington, D. C., 182 pp.