

On the Coexistence of an Evaporation Minimum and Precipitation Maximum in the Warm Pool

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ABSTRACT

Climatologically, the equatorial western Pacific warm pool region is a local minimum in surface evaporation and a local maximum in precipitation. The moist static energy budget in this situation requires a collocated minimum in radiative cooling of the atmosphere, which is supplied by the greenhouse effect of high clouds associated with the precipitation. However, this diagnostic statement does not explain why the evaporation minimum should coexist with the precipitation maximum. A simple physical model of the Walker circulation is used as the basis for an argument that the surface heat budget and the radiative effects of high clouds are essential to the existence of this feature, while variations in surface wind speed are not, though the latter may play an important role in determining the sea surface temperature.

1. Introduction

Climatologically, the surface evaporation has a local spatial minimum in the equatorial western Pacific warm pool region (e.g., Zhang and McPhaden 1995) while the precipitation has a maximum there (e.g., Huffman et al. 1997; Xie and Arkin 1997). From the point of view of the moisture budget by itself, this is explained by the dominance of horizontal moisture convergence over local evaporation as the primary source term balancing loss by precipitation. From the point of view of the moist static energy budget, since the gross moist stability (Neelin and Held 1987) is positive, the net vertical flux convergence of moist static energy—the surface fluxes minus the radiative cooling of the atmospheric column—must be positive where there is large-scale upward motion, as must coexist with a precipitation maximum. To the extent that the gross moist stability can be considered roughly constant in deep convective regions, as recent estimates suggest (Yu et al. 1998), the coexistence of the evaporation minimum and precipitation maximum implies a minimum in atmospheric radiative cooling at the same location. This is provided by the greenhouse effect of high clouds that are invariably present in the same location (Sherwood et al. 1994; Tian and Ramanathan 2002). This is only a consistency argument, however; it does not explain why

there should be an evaporation minimum at the location of the precipitation maximum.

I will argue here that a more complete understanding of this feature is possible when one considers the atmospheric moist static energy budget and surface energy budget simultaneously, and includes both the shortwave and longwave radiative effects of high clouds. The argument is straightforward, but to my knowledge has not been stated previously. The argument is framed in terms of an explicit, if simple, physical model of the Walker circulation, a slightly modified version of the one used by Bretherton and Sobel (2002), coupled to a simple slab ocean mixed layer.

2. Model

a. Atmospheric model

The model is a one-dimensional reduction of the Neelin–Zeng quasi-equilibrium tropical circulation model (QTCM; Neelin and Zeng 2000; Zeng et al. 2000), incorporating the weak temperature gradient (WTG) approximation (Sobel and Bretherton 2000; Sobel et al. 2001, and references therein), similar to that employed by Bretherton and Sobel (2002, hereafter BS02) in their study of the Walker circulation. The equations are

$$M_s \delta = P - R, \quad (1)$$

$$\hat{b} \left(\frac{\partial q}{\partial t} + \frac{A_q}{\hat{b}} u \frac{\partial q}{\partial x} \right) - M_q \delta = E - P, \quad (2)$$

$$\hat{a} \frac{\partial T}{\partial t} = \int_0^L (P - R) dx, \quad (3)$$

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with T the heat capacity of air (C_p) times the temperature, q the latent heat of vaporization of water times the specific humidity (in practice, we then divide both T and q by C_p and express them in degrees), and u the horizontal velocity, each a function of x and t and implicitly multiplying a fixed vertical structure function in the vertical coordinate. To obtain the physical fields for T and q , specified reference profiles must be added to the variable components that solve the above equations. The basis functions for T and q have a single sign in the troposphere, while that for u changes sign once with height, the sign in the equations above representing its value in the upper troposphere [we use the structure functions, reference profiles, and constants and coefficients derived from them, from the version 2.2 of the QTCM; see Table 1 of Neelin and Zeng (2000) or Table 1 of BS02]. Here P is precipitation, R is radiative cooling (assumed positive), E is surface evaporation, and δ is horizontal divergence. We have neglected surface sensible heat flux since it is small compared to the evaporation over tropical oceans. Also, M_s and M_q are the dry static stability and gross moisture stratification, taken to be linear functions of T and q :

$$M_s = M_{sr} + M_{sp}T, \quad (4)$$

$$M_q = M_{qr} + M_{qp}q, \quad (5)$$

with M_{sr} , M_{sp} , M_{qr} , M_{qp} being constants derived from the structure functions for T and q . The constant coefficients \hat{b} , A_q derive from the projection of the advection term on the structure functions of u and q .

We have assumed a domain slab symmetric in y , reducing the problem to one spatial dimension, x . We assume walls ($u = 0$) at $x = 0, L$. Assuming periodic boundary conditions does not change the problem in any important way (for sufficiently symmetric lower boundary conditions the problem can be made mathematically identical to that with walls). The WTG approximation results in the lack of a horizontal advection term in (1), and the lack of any momentum equation whatsoever in this simplified geometry, since the flow is purely divergent and the divergence is diagnosed from the temperature equation (1). Consequently, the domain size L drops out of the problem entirely (BS02), so we choose $L = 1$. We do however compute the domain-averaged temperature interactively. By WTG, we do this through an ordinary differential equation (3) for the single number, T . Also, M_s is constant in space, though not time, by (4).

The horizontal velocity, u , is related to the divergence δ ,

$$\delta = \frac{\partial u}{\partial x},$$

which we can integrate, using the boundary conditions,

$$u = \int_0^x \delta \, dx'. \quad (6)$$

The precipitation is parameterized using the simplified Betts–Miller scheme

$$P = \frac{q - T}{\tau_c}, \quad q > T, \quad (7)$$

$$P = 0, \quad q \leq T,$$

with τ_c the convective timescale. The radiative cooling is parameterized by

$$R = \frac{T - T_e}{\tau_R} - rP, \quad (8)$$

where τ_R is a radiative timescale and T_e a radiative equilibrium temperature and r is a nonnegative dimensionless cloud-radiative feedback parameter. Evidence justifying this simple parameterization of the greenhouse warming effects of high clouds is presented by BS02. The evaporation is parameterized by the bulk formula

$$E = (a + b|u|) \left[\frac{q_s^*(T_s) - q_s}{\tau_E} \right], \quad (9)$$

where q_s^* is the saturation specific humidity at the surface temperature T_s , and q_s is related linearly to q by the vertical structure function and reference profile. Here τ_E is the basic surface flux timescale. The parameters a , b give us a way to adjust the dependence of E on the surface wind speed. By default we take $a = 1$, $b = 0$, so that there is no dependence on surface wind speed, but vary these choices for sensitivity experiments, which we briefly describe but do not show.

The lower boundary condition is an interactive bulk mixed layer, satisfying

$$C \frac{\partial T_s}{\partial t} = S - E, \quad (10)$$

where C is the heat capacity of the mixed layer and S represents the combined effects of shortwave radiation, longwave radiation, and ocean heat transport. The S will be modified as a spatially varying but time-independent background plus a cloud-radiative feedback

$$S = \hat{S} - rP, \quad (11)$$

with

$$\hat{S} = S_0 + \Delta Sx.$$

Note that (8) and (11) together imply that the net effect of high clouds (the only kind we attempt to represent in this model) vanishes in the top-of-the-atmosphere (TOA) radiation budget; high clouds warm the atmosphere and cool the ocean in equal measure, as seems to be approximately true in observations (Ramanathan et al. 1989; Harrison et al. 1990; see also Hartmann et al. 2001, and references therein). The variation ΔS can be thought of as representing the smaller ocean heat transport divergence in the western versus eastern equatorial Pacific (e.g., Seager et al. 1988; Gent 1991; Trenberth and Solomon 1994) as well as the greater reduction

TABLE 1. Model parameter values.

Parameter	Symbol	Value
Minimum surface forcing	S_0	130 W m ⁻²
Change in surface forcing across domain	ΔS	100 W m ⁻²
Clear-air radiative equilibrium temperature	T_e	-50 °C
Radiative timescale	τ_R	25 days
Convective timescale	τ_c	0.2 days
Surface evaporation timescale	τ_E	10 days
Cloud-radiative feedback	r	0-0.25, variable

in incident solar radiation at the ocean surface due to low clouds in the east versus the west. Besides neglecting surface sensible heat flux, we also implicitly assume that the surface longwave radiative energy flux is constant, which is reasonable since observations show the spatial variations in surface shortwave flux to be much larger than those in longwave flux (e.g., Seager and Blumenthal 1994).

Key model parameters are shown in Table 1 [see Zeng et al. (2000) or BS02 for basic QTCM parameters]. Note that the mixed layer heat capacity C drops out of the steady-state problem and so is not listed. The basic surface forcing parameters S_0 and ΔS are based very loosely on observations, but are tuned somewhat to yield solutions with SST varying in a range close to that observed near the equator.

Besides the use of a mixed layer at the lower boundary rather than fixed SST, the only significant differences between the above model and that of BS02 are the use here of a finite τ_c , as opposed to the “strict quasi-equilibrium” limit ($\tau_c = 0$, $q = T$ where $P > 0$) in BS02, and that we use time-dependent equations and look for steady solutions by direct time integration, while BS02 assumed steady solutions and solved a boundary value problem.

b. Numerical solution method

For time integration we use a leapfrog method with a Robert filter. The advection term in (2) is computed using first-order upwind differences, a simple and natural choice since our boundary conditions render u everywhere of one sign. We use 100 grid points in the horizontal. This is adequate to render numerical diffusion unimportant, as verified by sensitivity tests with much higher resolution.

3. Relationship between evaporation and precipitation

A simple analysis of the atmospheric model equations, the essential part of which was presented by Sobel and Bretherton (2000), forms the basis of our argument regarding the evaporation minimum and precipitation maximum.

Adding (1) and (2) and dropping the q tendency term gives us the steady-state moist static energy equation:

$$u \frac{\partial q}{\partial x} + M\delta = E - R, \quad (12)$$

where $M = M_s - M_q$ is the gross moist stability (Neelin and Held 1987). For small τ_c , q will be close to T in convective regions (where $P > 0$), which means that q will be nearly horizontally uniform in such regions, so we can drop the horizontal advection term there [though this term may be quite important in determining where the boundary between the convective and nonconvective regions lies (BS02)]. Observational evidence justifying this assumption can be found in Folkins and Braun (2003). Then, using (1) to substitute for δ , and solving for P , we have

$$P = \frac{M_s}{M}E - \left(\frac{M_s}{M} - 1\right)R. \quad (13)$$

Notice that this result does not depend on the parameterizations for E , P , or R , except that the quasi-equilibrium property of the convective scheme was used to justify dropping the horizontal advective term.

By the assumed near uniformity of q in the convective region, M_q and hence M can be taken constant there as well. Given this, Eq. (13) shows that in this model, if R is either constant or can be parameterized as decreasing with E (i.e., a scatterplot of R versus E has uniformly negative or zero slope), then P must also increase with E . In this case, there cannot be a coincident evaporation minimum and precipitation maximum. *The only way a coincident evaporation minimum and precipitation maximum can occur is if E and R are at least locally correlated*,¹ so that if evaporation is reduced, radiative cooling of the atmosphere is reduced also.

Thus one can say that, *given* the coexistence of an evaporation minimum and precipitation maximum, there must be a minimum in radiative cooling in the same place, which there apparently is due to infrared cloud-radiative effects (Sherwood et al. 1994; Tian and Ramanathan 2002). However, from this argument it is not clear why the evaporation minimum should be there in the first place. The surface energy budget dictates that it cannot be a result of the minimum in surface wind

¹ This argument reads oppositely here from in Sobel and Bretherton (2000), because they used the opposite sign convention for R , defining it as the (assumed negative) radiative heating.

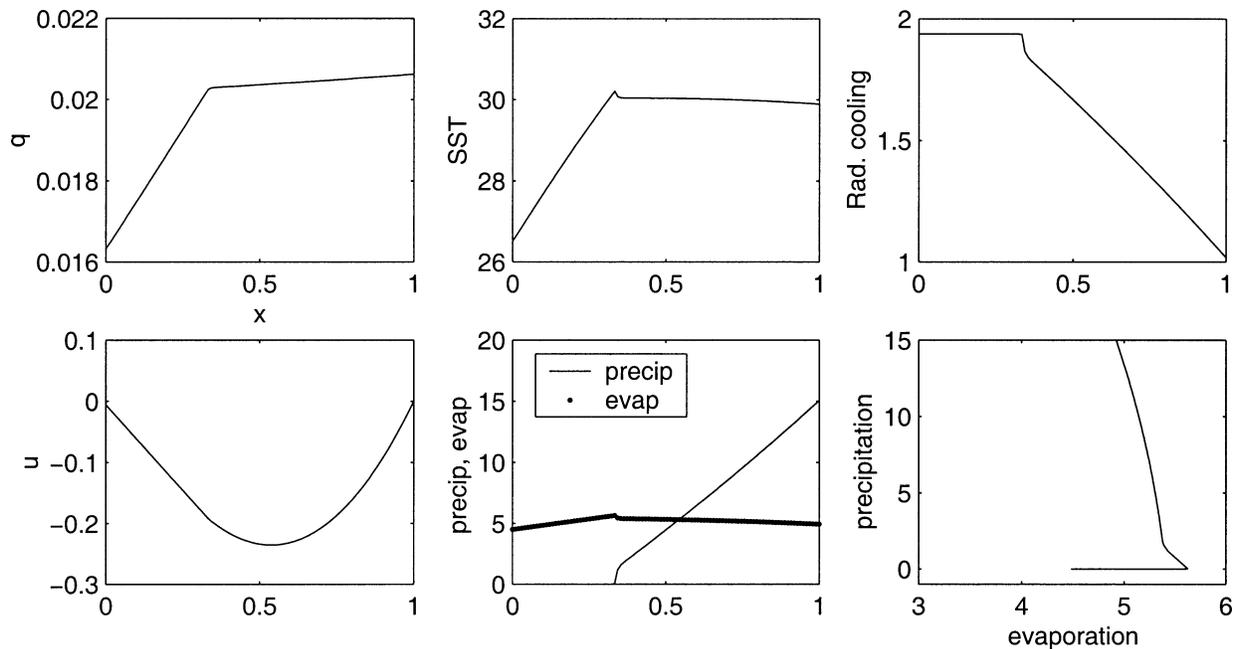


FIG. 1. Steady solutions of the model with the cloud-radiative feedback parameter $r = 0.2$. From left to right and top to bottom, the first five panels show surface specific humidity (kg kg^{-1}), SST ($^{\circ}\text{C}$), radiative cooling rate (K day^{-1}), wind velocity (arbitrary units), and precipitation and evaporation (mm day^{-1}), vs the horizontal coordinate x , while the last (bottom right) panel shows precipitation on the y axis and evaporation on the x axis.

speed. Since the divergence of ocean heat transport is weak in the warm pool (Seager et al. 1988; Gent 1991) and sensible heat flux small, the evaporation must approximately balance the net surface radiation, whose variations are dominated by the shortwave component. Therefore, given the surface radiation, the evaporation is fixed irrespective of the wind speed. Wind speed variations can, on the other hand, play an important role in controlling the SST. Once E is fixed, bulk formulas such as (9) show that larger wind speed $|u|$ requires a smaller difference $q_s^*(T_s) - q_s$, and we expect at least some of this difference to be accommodated by q_s^* and thus T_s .

If, for the sake of argument, we take the surface shortwave cloud forcing to be given, rather than interactive, we can consider the evaporation to be the given quantity in the problem for the atmosphere, since $E = S$ in the steady state. In this case, the coexistence of the evaporation minimum and precipitation maximum appears particularly difficult to explain, because the primary driver of the circulation will be the evaporation field. If atmospheric radiative cooling is fixed, features in P will have the same sign as those in E . If there is an infrared cloud feedback, as long as it is such that reduced atmospheric radiative cooling is associated with increased precipitation ($r > 0$), this cloud feedback will *amplify* those same features in P ; it cannot change their sign. Thus, E and R will be *anticorrelated*. The total moist static energy forcing of the column, $E - R$, and

thus the precipitation as well, will have maxima only where E does.

On the other hand, if we use both the shortwave and longwave components of our interactive cloud-radiative parameterization, with our interactive surface energy budget (10), then E and R can indeed be correlated in the steady state, because then we have $E = S$, and inspection of (11) and (8) shows that both R and E will decrease as P increases (for $r > 0$). This conclusion does not depend on the precise functional form of the cloud-radiative parameterization, but only that the presence of high cloudiness (represented by precipitation as a proxy) results in a net warming of the atmosphere and cooling of the ocean, of comparable magnitude.

4. Numerical solutions

Space prohibits presentation of a large number of numerical solutions. A few are shown and the general properties of a greater number are described in words.

Figure 1 shows surface specific humidity, SST, radiative cooling R , velocity u , precipitation and evaporation P and E plotted individually versus x , and evaporation and precipitation plotted versus one another, for a case with the mixed layer boundary condition and $r = 0.2$. The coexistent (local) evaporation minimum and precipitation maximum at $x = 1$ are evident. This feature exists for sufficiently large values of r . Plots of E versus P for a range of different values of r are shown

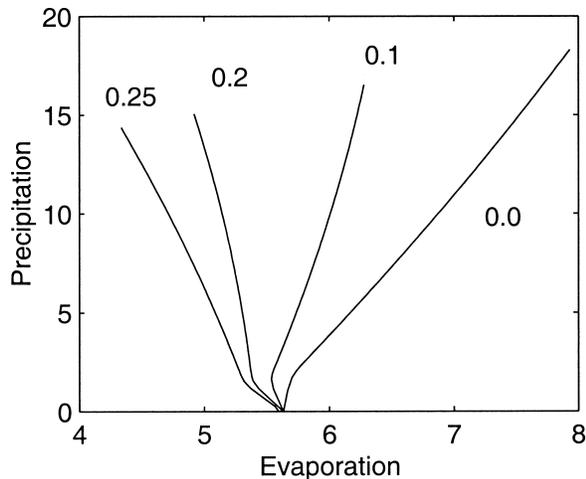


FIG. 2. Precipitation vs evaporation (mm day^{-1}) for four different values of the cloud-radiative feedback parameter r , labeled on the figure. Note that many (nonprecipitating) points lie along the x axis.

in Fig. 2. For r greater than some threshold (here about 0.3) a steady solution does not exist. The properties of the time-dependent solutions that do exist in this regime will be described in a separate study, but for the present purpose we simply state that the coexistent evaporation minimum and precipitation maximum persist robustly in the time mean of those time-dependent solutions.

BS02 found, with fixed SST and $\tau_c = 0$, that steady solutions were not possible if $M_{\text{eff}} < 0$, where the “effective gross moist stability” was defined by $M_{\text{eff}} = [M_s - a(1+r)M_q]/(1+r)$; a similar result was obtained by Su and Neelin (2002). Here, a steady solution is possible for M_{eff} slightly negative, because of the finiteness of both τ_c and the mixed layer depth (fixed SST corresponding to infinite mixed layer depth). The stability of steady solutions here depends not only on r , M_s , and M_q , but also on C and τ_c , and cannot be portrayed by an expression as simple as the above definition of M_{eff} . The stability problem will be addressed in detail, along with other aspects of the time-dependent behavior of the solutions in the unsteady regime, in a forthcoming study.

BS02 argued that the “observed” value of r based on monthly data should be around 0.2 with a significant uncertainty. This is close to the threshold value at which E starts to decrease with P in the convective region, according to Fig. 2. This raises the interesting question as to whether the existence of the evaporation minimum in the warm pool in fact gives us some indirect clue (in addition to more direct evidence of the sort shown by BS02) as to what the most appropriate value for r is.

The simulations shown above used fixed surface wind speed in the bulk evaporation formula. Using variable surface wind speed can change the SST field—for example, causing the weak local SST minimum at $x = 1$ in Fig. 1 to become a maximum, in better agreement with observations—but not the evaporation itself given

S , because of the constraint that $S = E$ in the steady state. From the discussion above, we expect that without both the longwave and shortwave cloud-radiative feedbacks, the coexistent evaporation minimum and precipitation maximum cannot be obtained, and numerical experimentation verifies that surface wind speed variations do not alter this conclusion. In combination with the cloud-radiative feedbacks, variable surface wind speed can modify the relationship between E and P , but since the cloud-radiative feedback can create the evaporation minimum by itself while variable surface wind speed cannot, the cloud-radiative feedback must be viewed as the key mechanism responsible for this feature.

5. Fixed SST

Since a consistent surface energy budget is a more physically appropriate boundary condition than fixed SST, we do not show any solutions of the model with fixed SST. However, solutions for a fixed SST distribution linear in x have been computed and warrant a few words of description and discussion. For fixed SST, the result (13) still holds, but E , rather than being determined by S in the steady state, has no a priori constraint, so the argument about variable surface wind speed in the preceding section has to be reformulated.

For $r = 0$, R is constant in the steady state and we can have no evaporation minimum at the precipitation maximum ($x = 1$). For $r > 0$ under fixed SST, clouds warm the atmosphere but have no effect on the ocean, resulting—unrealistically—in a significant net energy gain for the system. For r above a certain threshold, BS02 found that there is no steady solution. Our time-dependent simulations show that for fixed wind speed, past a certain threshold there is not even a well-behaved time-dependent solution. Rather, the convective region shrinks in size toward the limit of the numerical resolution and the precipitation in that region blows up rapidly, crashing the numerical algorithm. For variable surface wind speed, there is a regime with bounded time-dependent solutions.

For fixed wind speed, E in this model increases with increasing SST. This can be understood at least approximately, in the convective region (which is of most importance here), by noting again that for small τ_c we have $q \approx T$, hence constant, thus $q_s^* - q_s$ increases with T_s and because of fixed wind speed ($a = 1$, $b = 0$) so does E . We thus can have no evaporation minimum at $x = 1$, while the precipitation maximum stays there. In fact, the cloud-radiative feedback increases the positive correlation between E and P by creating a negative correlation between E and R , as discussed in the context of the fixed- E case in section 3.

By elimination, then, the only scenario in which it may be possible to get a coexistent evaporation minimum and precipitation maximum under fixed SST is to have variable surface wind speed in the bulk formula and (infrared) cloud-radiative feedback, since under this

scenario it is at least possible for E and R to become correlated. After considerable experimentation over broad ranges of a , b , r , etc., I have not found a case in which a coexistent evaporation minimum and precipitation maximum occur in the steady state. With variable surface wind speed and r sufficiently large that the solution is unsteady, I have found cases with a coexistent evaporation minimum and precipitation maximum in the time mean.

In any case, again, the fixed SST case, while of some interest theoretically, is clearly less relevant than the mixed layer case to the real climate, because the surface has finite heat capacity.

6. The possibility of a gross moist stability minimum

One possibly weak link in our argument is the assumption, built into the model by the smallness of τ_c , that the gross moist stability M be nearly constant within the convective region. If there were a deep enough minimum in M , coincident with an evaporation minimum that we assume exists for the sake of argument, a coincident P maximum could be obtained without cloud-radiative feedback. In fact, variations in M were assumed to be the dominant controller of spatial precipitation variations by Neelin and Held (1987). In a recent attempt to estimate the gross moist stability from observations, Yu et al. (1998) found M variations in convective regions to be small. More importantly, though, without high cloud-radiative feedback one again has a problem explaining why there should be an evaporation minimum in the warm pool to begin with. Diurnal average TOA insolation is constant along the equator, so longitudinal variations in ocean heat transport divergence and the albedo effect of low cloudiness must be balanced by evaporation variations of opposite sign in the surface energy budget. Since low cloudiness and ocean heat transport tend to cool the eastern Pacific relative to the warm pool, these effects would require an evaporation *maximum* in the warm pool in the absence of the shortwave reduction by high clouds.

7. Conclusions

I have used a simple model of the Walker circulation to examine the coexistent evaporation minimum and precipitation maximum in the western Pacific warm pool.

When the lower boundary condition of the model is a slab ocean mixed layer, and the simple cloud-radiative feedback parameterization assumes that precipitating clouds reduce (in equal measure) both the radiative cooling of the atmosphere and the radiative heating of the ocean, the model robustly produces a coexistent evaporation minimum and precipitation maximum—either in a steady state, or as the time mean of a time-dependent solution—for sufficiently large values of the parameter

that controls the strength of the cloud-radiative feedback. The model does this even with fixed surface wind speed in the bulk formula for the evaporation. Wind speed variations, if included, can be an important control on the SST in this model, but not on the evaporation or precipitation. Because the model surface wind speed is not particularly carefully simulated, no conclusions are drawn regarding what physical process limits the SST (e.g., Newell 1979; Ramanathan and Collins 1991; Wallace 1992; Waliser and Graham 1993; Hartmann and Michelsen 1993; Zhang et al. 1995; Waliser 1996).

With a fixed SST distribution as the lower boundary condition, it is much more difficult to obtain a coexistent evaporation minimum and precipitation maximum. A steady solution with this feature has not been obtained under any conditions for fixed SST. When the atmospheric warming effect of high clouds is included and the effect of wind speed on surface fluxes is taken sufficiently strong, it can be produced in the time mean of a time-dependent solution.

Because the coexistent evaporation minimum and precipitation maximum are less robustly produced for fixed SST than they are with the mixed layer, and (more importantly) because in reality the surface energy budget must be balanced, I conclude that the radiative effects of high clouds, but not surface wind speed variations, are the key ingredients leading to the existence of this feature.

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