Reply

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We thank Henderson-Sellers and Henderson-Sellers (herein abbreviated as H-S) for their suggestion that our diffusion cloud atmosphere (Weare and Snell, 1974; herein abbreviated as WS) might be useful in climate modeling of other planets, or Earth at times during "its evolution when both data and results are in globally and annually averaged form." In any such application, however, it must be made clear that the use of the model atmosphere demands knowledge of the thermodynamic and radiative transfer properties of its constituents especially those which may co-exist in two or more phases (such as water on Earth or carbon dioxide on Mars) with widely differing radiative transfer characteristics. However, it is the primary purpose in this "reply" to amplify and hopefully thereby clarify how the model atmospheric structure is also inextricably dependent upon nonthermodynamic, essentially kinetic quantities which we chose to parameterize in simple form in WS. Any extension of the model to other globally averaged atmospheres or to the zonally averaged Earth's atmosphere must thereby include a consideration of these quantities.

The three parameters under discussion are the surface relative humidity γ, the vertical diabatic heat transport parameter A, and the rainout factor f_r, each taken as globally determined constants in WS, independent of the primary variable, surface temperature. This assumption, we believe, necessarily restricts our global climatic model to a small range of surface temperatures ($\sim\pm 4$ K) about the global annual average of 288 K, contrary to the suggestion of H-S. Larger variations may well be associated with a change in the values of these parameters although we don't have either data or reasonable theory to functionally relate that change to surface temperature at this time. However, in extending the diffuse cloud structure to a zonally averaged climatic model it is clear that each of the parameters may be determined at the zonally averaged reference surface temperatures to reflect present-day observations. In this respect our model atmosphere is not restricted to the global case.

The surface relative humidity γ taken in WS as 0.75 appears to be an accepted reasonable value and requires no further comment.

The diabatic heating parameter A is defined by Eq. (2.6) of WS and as stated its value (properly given as -3.7 cal K^{-1}) was so chosen as to make the lapse rate correspond to the "standard" of 6.5 K km^{-1} at temperatures above the dew point (i.e., essentially equivalent to a dry atmosphere). Its value is calculated from Eq. (2.9) in WS.

Due to the release of latent heat at heights where the temperature falls below that of the dew point the overall lapse rate averages less than 6.5 K km^{-1}, being about 5.3 K km^{-1}. However our means of setting the value ot A leads to good agreement with data tabulated by Oort and Rasmussen (1971), as may be seen in Table 1. Certainly the agreement is best for the lower heights where most of the liquid and vapor water exists and thereby has the greatest influence on the radiative transfer calculations.

To further clarify the physical interpretation of the parameter A it may be seen from Eq. (2.9) of WS that A/n_α, where n_α is the number of moles, is equivalent to the magnitude by which the molar heat capacity of air must be increased, such that the lapse rate is reduced from the true adiabatic value of 9.75 K km^{-1}. This parameter may also be related to that portion of the diabatic heating required to reduce the lapse rate from the adiabatic value to 6.5 K km^{-1} (below heights of the dew point). Excluding that portion of the diabatic heating necessary to compensate for the radiative losses from the atmosphere we may write Eq. (2.6) of WS in terms of the total time derivative to obtain

$$T\frac{ds}{dt}=\frac{A}{n_\alpha}\frac{dT}{dt}=-\bar{v}\frac{\partial J_q}{\partial z}, \tag{1}$$

TABLE 1. Temperature differences between 1000 mb and other pressure levels.

	$T_{1000}-T_{850}$	$T_{1000}-T_{700}$	$T_{1000}-T_{500}$	$T_{1000}-T_{200}$
Oort & Rasmussen (Northern Hemisphere average)	6.2	13.9	29.4	69.0
Weare and Snell	7.2	14.3	27.5	53.5
Standard atmosphere	8.7	18.8	35.5	70.8

where s and $\bar{v}$ are the molar entropy and volume, T the temperature, and $-\partial J_q/\partial z$ the negative divergence of that portion of the convective heat flux altering the lapse rate. In the stationary state $\partial T/\partial t=0$, and then it follows that

$$\frac{A}{n_\alpha} w \frac{\partial T}{\partial z} = \frac{\bar{M}}{\rho} \frac{\partial J_q}{\partial z}, \tag{2}$$

where w is the vertical velocity of ascent, $\bar{M}$ the average molecular weight of air, and ρ the mass density. Of course, the assessment of w requires additional assumptions beyond those inherent in the global model presented.

The parameter f_r describing the fraction of condensate formed during a quasi-adiabatic expansion that is rained out from the atmosphere is absolutely necessary to produce an atmosphere which yields the proper radiative transfer. In the computational scheme there is, for a given incremental temperature decrease (at temperatures less than the dew point), an associated condensation of water vapor to liquid as given by Eq. (2.8) of WS. If all the condensate produced at each incremental temperature decrease were to remain in the atmosphere at that level and be distributed as droplets according to our assumed distribution function, the resulting scattering optical depth would be very large, and likewise the atmospheric albedo. Consequently, we remove a given constant fraction f_r of the condensate that is formed in each temperature interval leaving the fraction $1-f_r$ to be distributed as droplets. This fraction is adjusted to give the appropriate scattering optical thickness and which together with an assumed surface albedo provides the proper planetary albedo. The atmosphere so generated is assumed to be the steady state, and thus $1-f_r$ is related to the steady-state "holding capacity" of the atmosphere for liquid water. All subsequent evaporation from the surface must appear as rainfall, independent of the value assigned to f_r. Furthermore, the assumption that f_r is a constant independent of surface temperature means that the absolute holding capacity of the atmosphere increases with an increase in the surface temperature and thereby provides the negative feedback for absorbed solar radiation.

We thank H-S for bringing to our attention the sign errors in our Eqs. (3.9) and (3.10). This error was not within the programmed calculations and thus does not affect the validity of the published results. Also H-S are quite correct in pointing out that our method of averaging the hemispherical latitudinal extent of ice cover leads to inaccurate area-weighted albedo calculations for large variations in $\langle\theta\rangle_N$ or $\langle\theta\rangle_S$. For the surface temperature variations explored in WS, however, the discrepancy amounts to a maximum of only 1.03% in the surface albedo and when combined with the atmospheric transmissivities and reflectivities in Eqs. (3.7) and (3.8) of WS becomes entirely negligible.

The choice of proper surface albedos is a matter of continuing concern. Suffice it to reply that any values are subject to considerable uncertainty. Robinson (1966) lists possible values for snow ranging from 0.29 to 0.95, and for ice, from 0.12 to 0.41 depending on their state. Also Kondratyev (1969) gives some theoretical results for albedos of rough and calm sea at various solar elevations. These are reproduced in Table 2. In view of the great variability in estimates, our choice of 0.60 for snow-ice and 0.07 for land-sea giving an area-weighted average of 0.12 at a surface temperature of 288 K seems quite reasonable. In a global annually averaged model further adjustments to take account of the solar zenith angle dependence seem entirely unnecessary. However, with extension of the model atmosphere to zonally averaged quantities, the albedo dependence on zenith angle assumes greater importance. Also, the interplay of surface reflectivity and atmospheric absorbers becomes important in altering the absolute amount of solar radiation absorbed and the functional feedback for changes in snow-ice cover (Temkin *et al.*, 1975).

Finally we should reiterate that the diffuse thin cloud model atmosphere thermodynamically derived in WS is a horizontally homogeneous structure, and consequently there is no distinction between the environment and "cloud" lapse rate. Therefore, no defined "top" to the cloud is produced. Instead the absolute humidity asymptotically approaches zero in the limit of $z\rightarrow\infty$. However, in our computational scheme the vertical structure is truncated at the height where the temperature is 212 K. The small amount of residual water that would be condensed as $T\rightarrow 0$ or $z\rightarrow\infty$ is assumed to exist at this height, thus creating an artificial "top" or "tropopause."

TABLE 2. Albedos of sea.

Solar zenith angle (deg)	Rough sea	Calm sea
0	0.131	0.021
30	0.038	0.022
60	0.024	0.062

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Comments "On the Interaction between the Subcloud and Cloud Layers in Tropical Regions"

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The paper by Ogura and Cho (1974, hereafter O&C) presents a very interesting and comprehensive introduction to the problem of constructing a well-mixed type boundary-layer model for application to the tropical subcloud layer. However, there are four points with which I would like to take issue:

The cloud fluxes of $s=c_pT+gz$ and of specific humidity q are expressed by O&C (with minus signs added) as

$$(\overline{\omega' s'})_c=-M_c(s_c-s_m), \tag{2.4}$$

$$(\overline{\omega' q'})_c=-M_c(q_c-q_m), \tag{2.5}$$

where M_c is the upward cloud mass flux, s_c and q_c are representative values of s and q inside the clouds at cloud-base height, and s_m and q_m are subcloud (mixed) layer values of s and q. The latter are assumed constant with height in the mixed layer above the surface layer. (It should be noted that in keeping with the notation of O&C, $M_c \geqslant 0$ even though it has the dimensions of pressure change per unit time, while $\omega'>0$ represents a downward directed vertical velocity.) Inasmuch as the clouds are assumed to grow out of the top of the mixed layer, it follows that $s_c \approx s_m$ and $q_c \approx q_m$ at cloud-base height as shown schematically in Fig. 1. The right-hand sides of (2.4) and (2.5) are thus essentially zero except for the fact that the mixed-layer values of s and q are not strictly constant. The mixed layer in this figure is drawn with a locally variable height, of mean value h, and the thin transition layer depicted extends from the effective lower reaches of this height to the cloud-base height (which may itself be somewhat variable). Fig. 1 suggests that these equations should have the form

$$(\overline{\omega' s'})_c=M_c(\bar{s}_+-s_m), \tag{2.4$'$}$$

$$(\overline{\omega' q'})_c=M_c(\bar{q}_+-q_m), \tag{2.5$'$}$$

where $\bar{s}_+$ and $\bar{q}_+$ are the large-scale average values existing just at the top of the transition layer. To the extent that the cloud fraction is quite small, as assumed by O&C, $\bar{s}_+$ and $\bar{q}_+$ represent mean properties of the air existing between clouds at cloud-base height. It is rather important that vertical fluxes at cloud-base height be related to cloud-environment differences existing at cloud-base height and not to those existing at the underside of the transition layer as in (2.4), (2.5). The extension of the infinitessimally thin transition layer to one of finite thickness, as in Fig. 1, should make this point clear. Such an extension has been found useful by Betts (1974).

In Eqs. (2.4)′ and (2.5)′ subscript c refers to evaluation at cloud-base height. However, O&C let $(\overline{\omega' q'})_c$ be the moisture flux associated with the presence of cumulus clouds, and a remaining portion, F_q, be the vertical flux associated with all other small-scale eddies; the sum of these two being $\overline{\omega' q'}$. Judging from their discussion on p. 1852 and from the occurrence of $\partial(\overline{\omega' q'})_c/\partial p$ in (2.3), O&C intend this distinction to apply within the mixed layer as well as above. However, it then seems impossible to give a firm definition to their F_q (which O&C did not attempt to do). Perhaps the best that could be done is to state that their F_q is the moisture flux which would have occurred if cumulus clouds had not been present and all other conditions had been the same. Clearly this position is untenable if cumulus clouds do occur, because of the strength and

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