Fog and rain in the Amazon

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The diurnal and seasonal water cycles in the Amazon remain poorly simulated in general circulation models, exhibiting peak evapotranspiration in the wrong season and rain too early in the day. We show that those biases are not present in cloud-resolving simulations with parameterized large-scale circulation. The difference is attributed to the representation of the morning fog layer, and to more accurate characterization of convection and its coupling with large-scale circulation. The morning fog layer, present in the wet season but absent in the dry season, dramatically increases cloud albedo, which reduces evapotranspiration through its modulation of the surface energy budget. These results highlight the importance of the coupling between the energy and hydrological cycles and the key role of cloud albedo feedback for climates over tropical continents.

Results and Discussion

We use the Weather Research and Forecasting (WRF) model at 2-km horizontal grid spacing (see Methods for details of the runs). This resolution has been shown to be sufficient to resolve deep convection (no convective parameterization is used) and to correctly represent the sign and magnitude of the land–atmosphere interaction feedback (18), contrary to GCMs, which tend to exhibit soil moisture–precipitation feedbacks of opposite sign to that observed (19). We parameterize the time-dependent large-scale vertical motion as a function of internally resolved convection using the weak temperature gradient (WTG) approximation (Methods). The WTG approach has been used in many previous studies of tropical atmospheric dynamics to represent the feedback between locally resolved convection and larger-scale circulation (Methods) (10, 11, 20–23). Under WTG, we diagnose the horizontal average large-scale vertical motion (Fig. S1) so as to induce a vertical advective potential temperature tendency, which relaxes the model’s domain-averaged potential temperature toward that of the target profile. The same large-scale vertical motion is then also used for vertical advection of moisture (Methods).

The target potential temperature profiles we use, representative of the wetter part of the Amazon, are taken from the 2014 Atmospheric Radiation Measurement mobile facility located at Manacapuru, near Manaus, Brazil (3°12′46.70″S, 60°35′53.00″W). The profiles are averaged over the month of February for the wet season, and September for the dry season, as shown in Fig. 1A. At the surface (where WTG is not directly applied, see Methods), the dry profile is warmer by about 5 K, reflecting the influence of higher surface heat fluxes (Fig. 1B). In the midtroposphere, the wet profile is warmer by more than 1 K. In the stratosphere, the wet profile is colder by more than 5 K due to the seasonal elevation of the tropopause. All these differences are consequences of the differences in the diurnal and monthly mean temperature profiles.

Significance

We here demonstrate that we can resolve the seasonality of the hydrologic cycle in the Amazon using an approach, opposite to general circulation models, in which we resolve convection and parameterize large-scale circulation as a function of the resolved convection. The results emphasize the key role of cloud albedo feedback and, in particular, of the morning fog layer in determining the diurnal course of surface heat fluxes and seasonality of the surface and atmospheric heat and water cycles. These results indicate that our understanding of tropical climates over land can be considerably advanced by using coupled land–atmosphere models with explicit convection and parameterized large-scale dynamics.

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of the seasonal cycle itself; the warmer troposphere and cooler lower stratosphere are typical differences between states of stronger and weaker deep convection (24). Here the ultimate cause of the wet–dry season temperature difference is presumably the stronger insolation in the wet season, but the temperature differences are in part an outcome of land–atmosphere interactions, as the tropospheric temperature difference between the two seasons in the nearby oceanic regions shows no such signal (Fig. S2). Because we assume this seasonal difference in temperature profiles, our simulations do not predict the seasonal changes purely as a function of external boundary conditions and forcing. Rather, we predict part of the solution—including precipitation, clouds, and surface fluxes—given another, the free atmospheric temperature. This allows us to understand the mediation of convection and land–atmosphere interaction by the atmospheric temperature, similar to what has been done in studies of tropical cyclones and the Madden–Julian oscillation (24, 25).

The increase in precipitation in the wet season relative to the dry season (Fig. 2) can be attributed to the reduction in the export of moist static energy by the large-scale circulation (see Supporting Information), which is a consequence of the more stable temperature profile (25), despite the absence of an increase in surface fluxes or radiative heating. This increase, in turn, moistens the land surface, an important factor leading to the differences in cloud between the wet and dry seasons. In agreement with observations [here Climate Prediction Center Morphing Technique (CMORPH) (26, 27) data averaged over 10 y (2004–2014) for the months of February and September for the wet and dry seasons, respectively], our simulated precipitation maxima occur in the early to late afternoon in both seasons, with greater absolute amplitude in the wet season.

In both seasons, rapid precipitation transitions from minima in the morning local hours to maxima in the afternoon are associated with transitions from shallow to deep convection (Fig. 3). In the dry season (Fig. 3A), convection starts in the morning (1000 hours time zone local time) manifesting first as a sharp increase in midlevel cloud at that time coverage, and then a more gradual transition to deep convection in the afternoon associated with the timing of rainfall maxima. Midlevel cloud cover dissipates overnight into the early morning. The overall cloud cover is relatively small during daytime and higher in the afternoon and evening, as observed, but contrary to what GCMs predict as mentioned above. The smaller cloud cover during daytime drives greater surface fluxes, and, in particular, evapotranspiration and photosynthesis, by allowing more solar radiation to reach the surface.

In the wet season (Fig. 3B), the diurnal cycle of cloud cover exhibits a different behavior. In addition to the increase in cloud cover, there is now a distinct layer of fog above the surface that is maintained by radiative cooling from the wet surface and lasts from midnight until noon, as observed in the Amazon (Figs. S3 and S4) and other rainforests. This morning fog layer blocks shortwave radiation from reaching the surface and reduces daytime net surface shortwave radiation and evapotranspiration. Such morning fog was also shown to be critical in the maintenance of distinct diurnal climate equilibrium regimes by a previous modeling study (28, 29).

As a result, the latent heat flux (Fig. 4A) and surface temperature are smaller in the wet season than in the dry season, because surface shortwave radiation is reduced by the fog layer (Fig. 4B), which more than compensates for the higher soil water availability and lower Bowen ratio in the wet season (from diurnal average of 0.4 in the dry season to 0.32 in the wet season based on eddy covariance data). We note that surface observations are difficult to correctly measure in the presence of dew (eddy covariance measurements cannot correctly record measurements) so that most days with fog are not captured by surface observations; thus clear days (infrequent in the wet season; Figs. S3 and S4) with higher latent heat flux are overrepresented in the observations. Because the diurnal rise of the surface turbulent fluxes is delayed in the wet season by the fog layer, the transition from shallow to deep convection occurs later in the day compared with the dry season.

Although the downward solar insolation at the top of the atmosphere in the wet season exceeds that in the dry season by more than 20 W m\(^{-2}\), the net shortwave radiation at the surface is greater in the dry season (Fig. 4B). This results from a strong negative cloud albedo feedback in the wet season and greater reflection of shortwave radiation to space. The fog layer is an important contributor to this albedo. We further quantify the cloud albedo feedback by computing cloud radiative forcing (CRF), as the clear-sky minus all-sky upwelling flux at the top of the atmosphere (Fig. 5). Weaker convection in the dry season induces a smaller longwave CRF component compared with that...
induced in the wet season. However, the shortwave CRF in the wet season is much more negative than in the dry season due to the presence of the fog layer. Because the shortwave CRF dominates in the wet season, and longwave CRF dominates in the dry season, the net CRF is negative in the wet season and positive in the dry. Comparison with CRF obtained from the Clouds and the Earth’s Radiant Energy System (CERES) [the CERES SYN1deg daily radiative fluxes (30, 31)] shows a reasonable agreement with our simulations, although some biases exist. Perhaps surprisingly, the seasonal difference in the top of the atmosphere seasonal mean insolation is not the dominant control that determines the differences in seasonal climate, even though it is ultimately what controls the seasonal cycle in nature. In our simulations, the differences in the target potential temperature profiles are primarily responsible for the seasonal differences described above. We performed sensitivity experiments in which the insolation from the wet season is used with the target potential temperature profile from the dry season and vice versa (see Figs. S5–S7). We also performed sensitivity experiments over Rondonia (10.7°S, 62.7°W), where the seasonal difference in insolation is greater (Figs. S8–S10). When insolation is varied while holding the potential temperature profile fixed, no significant difference is found in terms of the typical pattern of diurnal and seasonal patterns of precipitation. Again, the cloud albedo adjusts so that the surface shortwave radiation is always higher in the case of the dry season profile, leading to higher evapotranspiration flux (Fig. S9). This behavior differs from convection over oceans, where stronger seasonal insolation leads to higher surface fluxes (and sea surface temperature), and cloud albedo is not so tightly coupled to atmospheric convection because, unlike land, the ocean can both substantially store and transport heat reducing the coupling of evaporation with shortwave incoming radiation.

Methods

Model Configuration. We use the WRF model version 3.3, in three spatial dimensions, with doubly periodic lateral boundary conditions. The experiments are conducted with Coriolis parameter \( f = 0 \). The horizontal domain size is 192 x 192 km² with a grid spacing of 2 km. There are 50 vertical levels total, with the top level at 22 km, and 10 levels in the lowest 1 km. Gravity waves propagating vertically are absorbed in the top 5 km to prevent unphysical wave reflection off the top boundary by using the implicit damping vertical velocity scheme (32). The 2D Smagorinsky first-order turbulent closure scheme is used to parameterize the horizontal transports by subgrid eddies. The Yonsei University first-order closure scheme is used to parameterize nonlocal boundary layer turbulence and vertical subgrid-scale eddy diffusion (33). The surface fluxes of moisture and heat are parameterized following Monin–Obukhov similarity theory. Microphysics is simulated using the Purdue–Lin bulk scheme (34), which has six species: water vapor, cloud water, cloud ice, rain, snow, and graupel. Radiative fluxes are determined interactively using the National Center for Atmospheric Research Community Atmosphere Model version 3.0 scheme for shortwave and longwave radiation. Both surface and radiative fluxes are fully interactive. The atmospheric model is coupled to the Noah land surface model (LSM) (35) that has four soil layers at 10, 30, 60, and 100 cm depth. The LSM provides four quantities to the atmospheric model: sensible heat flux, latent

\( \text{Fig. 3. Composite of the WRF simulated 2-d cycle of fractional cloud cover for the (A) wet and (B) dry season. Note the fog layer above the surface in the wet season.} \)

\( \text{Fig. 4. Diurnal cycle of (A) latent heat flux, and (B) net shortwave at the surface, for WRF modeled (thick) and observed (thin) fluxes. Observed fluxes are taken from the climatology of eddy covariance fluxes observed at K34 station in Reserva Biológica do Cuieiras. We note that surface observations are difficult to obtain in the presence of dew (eddy covariance measurements typically cannot correctly record measurements) so that the observations typically omit fog situations, with an overrepresentation of relatively clear days compared with fog days.} \)
heat flux, upward longwave radiation, and upward shortwave radiation off the ground. The LSM prognostic land states are surface skin temperature, volumetric total (liquid and frozen) soil moisture at each layer, soil temperature at each layer, and canopy water content. Vegetation type is evergreen forest, surface albedo is 0.12, and the wind field is left unnuudged.

Parameterized Large-Scale Circulation and Initial Conditions. The large-scale vertical velocity is dynamically parameterized using the WTG method (20, 21). The WTG vertical velocity $W_{\text{WTG}}$ is obtained by a Newtonian relaxation with a relaxation time scale $\tau$ (over which gravity waves propagate out of the domain) taken here as $2 \, h$ (20, 21),

$$W_{\text{WTG}}(z) = \frac{1}{\tau} \left( \frac{\theta - \theta_0}{\partial \theta / \partial z} \right)$$

where $\theta$ is the domain mean potential temperature, and $\theta_0$ is the observed potential temperature profile obtained from radiosondes averaged over 1 mo of observations. $W_{\text{WTG}}$ is then linearly interpolated in the boundary layer to zero at the surface because gravity waves are not the main mode of buoyancy adjustment in the boundary layer. The impact is negligible in the daytime boundary layer as it is well-mixed so that the vertical gradients are null. Boundary layer height is determined interactively in the WRF model using the bulk Richardson number method and varies diurnally.

Once $W_{\text{WTG}}$ is obtained, it is used to define the domain-average large-scale tendencies of potential temperature and specific humidity over the computational domain,

$$\frac{\partial \theta}{\partial t}_{\text{Large Scale}} = \frac{\theta - \theta_0}{\tau}$$

$$\frac{\partial q}{\partial t}_{\text{Large Scale}} = -W_{\text{WTG}} \frac{\partial q}{\partial z}$$

respectively, where $\theta$ is the domain mean water vapor mixing ratio.

Initial conditions are seasonally averaged quantities in the dry and wet seasons using the ERA-Interim data set. The model is run for about 40 d and the analysis conducted on the equilibrium part only, which is the last 30 d.

We have performed a sensitivity test on the soil moisture initial conditions, by switching the seasonal magnitudes, and they had no influence on the results. The target temperature profile is time-independent in each season and does not include diurnal variations (21). The modeled diurnal cycles of surface fluxes, clouds, and precipitation are driven solely by the diurnal cycle of radiation at the top of the atmosphere. Daily mean simulations in the dry season is held fixed at 415 W m$^{-2}$ and, in the wet season, at 439 W m$^{-2}$, reflecting values at Manacapuru.

Surface Observations. We used eddy covariance data from the K34 station located in Estação Experimental de Silvicultura Tropical 02°37′S, 60°09′W, located around 60 km from Manaus. We have used data from 2000 to 2006 available from Oak Ridge National Laboratory (36). We used quality-controlled eddy covariance data, based on outliers comparison, wind speed (acceptable variation is two SD units from the linear regression), and unresponsive sensor checks. Eddy covariance measurements typically cannot correctly measure in the presence of dew; thus fog conditions are undersampled, which implies that the latent heat flux in Fig. 4 is an overestimate of the true latent heat flux (an average of fog conditions with little radiation and nonfog conditions with higher radiation).

Conclusion

We have shown that, when given the top of the atmosphere insolation and the monthly mean free-atmospheric temperature profile, the seasonal and diurnal cycles of cloud, precipitation, and surface fluxes can be simulated by a limited domain cloud-resolving model with parameterized large-scale forcing without the biases commonly found in global climate models.

We can view the seasonality in the Amazon as mediated by the atmospheric temperature profile. That profile itself results from the stronger convection in the wet season, but the increased wet season precipitation in the model is still a nontrivial prediction of the model. It can be explained by the reduced ventilation of moist static energy from the column, which is a consequence of the warmer tropospheric temperature profile in the wet season.

The surface fluxes, on the other hand, are strongly controlled by the diurnal cycle of cloud albedo, and especially of wet season fog, which blocks shortwave radiation from reaching the surface in the early morning. This fog is an essential regulator of the Amazon climate.

We demonstrate that a high-resolution cloud-resolving model with parameterized large-scale circulation offers a new window onto the dynamics of climate in the Amazon. In the future, this approach may allow new insights into the Amazon’s changes under anthropogenic influence and the capacity of the basin to act as a CO$_2$ sink in the future.

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Fig. 5. Cloud radiative forcing, shortwave (SWCRF), longwave (LWCRF), and net (NETCRF) for the wet and dry seasons as simulated (squares) and observed from CERES (circles).