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Sensitivity of stable water isotopic values to convective parameterization schemes

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[1] Convective parameterization has been argued as a principal generator of inter-model differences in climate sensitivity, but it is difficult in practice to constrain simulated convective processes. Here we show how stable water vapor isotopes, which are sensitive to the convective condensation rates, may be useful for evaluating convective parameterizations. By varying one of the least constrained convection parameters in the NCAR Community Atmosphere Model (CAM), namely the timescale for consumption of convective available potential energy (CAPE), τ, the simulated precipitation experiences substantial changes in response to changes in both the deep and shallow convection schemes—increasing τ from the standard 2 hours to 8 hours increases the contribution from shallow convection. The lowest order effect of increasing τ is a decrease (increase) in lower (upper) tropospheric condensation rates, with approximately the opposite vertical structure for the change in simulated isotopic signature. Increasing τ from the standard 2 hours to 8 hours also provides a better match to satellite-observed water vapor isotope ratios, albeit with some uncertainty related to the quality of currently-available satellite measurements. Thus, the incorporation of water vapor isotopes into GCMs provides additional constraints on convective parameterizations, especially as more and better quality water vapor isotope measurements become available.

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1. Introduction

[2] Because of the inherently small spatial scales of convective physics, the representation of convection in Atmospheric General Circulation Models (AGCMs) requires parameterization onto the grid-scale variables simulated. Most of the widely used AGCMs invoke some aspects of the quasi-equilibrium approximation because convective elements tend to cover small area [*Stevens*, 2005]. Needless to say, convective parameterizations constitute a major source of intermodel discrepancy in climate sensitivity among current generation GCMs [*Held and Soden*, 2006], although it may be difficult to isolate the precise cause of model divergence [*Raymond*, 2007].

[3] Heavy isotopologues of liquid phase H_2O (i.e., HDO or $H_2^{18}O$) have lower saturation vapor pressure than the lighter one ($H_2^{16}O$); therefore, the isotopic ratios of heavy to lighter elements in condensate is higher than in vapor [*Gat*, 1996]. The isotopic content is reported as deviations from the ratios in standard mean ocean water (SMOW): $\delta D = \left[\left(\frac{D}{H}\right)_{sample} / \left(\frac{D}{H}\right)_{SMOW} - 1\right] \cdot 1000$. In equilibrium, the difference in deuterium between the isotopic compositions of liquid and vapor is $\sim$80‰. Because of this large fractionation signal, the isotopic compositions of the surface and meteoric water have regionally distinctive ratios dependent on climate conditions [*Dansgaard*, 1964]. The characteristic ratios of water isotopes have been used to estimate temperature [e.g., *Jouzel et al.*, 2007; *Lee et al.*, 2007] and precipitation amount [e.g., *Partin et al.*, 2007]. Stable water isotopes have also been used to diagnose the dehydration pathway in the upper troposphere and lower stratosphere [e.g., *Moyer et al.*, 1996; *Kuang et al.*, 2003; *Dessler and Sherwood*, 2003]. The increasing amount of in situ airplane [*Moyer et al.*, 1996; *Webster and Heymsfield*, 2003] and satellite measurements [*Worden et al.*, 2007] of stable water isotopes should facilitate the study of atmospheric hydrological processes.

[4] The usefulness of water isotopes as proxies for observed hydrological cycle processes has stimulated inclusion of water isotopes in atmospheric models of varying complexity [e.g., *Tindall et al.*, 2009; *Yoshimura et al.*, 2003; *Risi et al.*, 2008] because they exhibit a unique correlation length scale, different from either temperature or precipitation (Figure S1 of the auxiliary material).[4] A single column model analysis by *Bony et al.* [2008] demonstrated the sensitivity of upper tropospheric vapor to how the convection scheme treats microphysics, thereby suggesting the potential utility of water isotope tracers for diagnosing convective processes.

[5] In this paper, we demonstrate how water vapor isotopes can provide a useful constraint on the details of the convective parameterization. Although the adjustable parameters used in convection schemes may span a range of physically permissible values, the precise values chosen can have substantial impacts on the climate simulations [*Jackson et al.*, 2008]. One such parameter is the characteristic decay time for convective instability (τ). *Arakawa and Schubert* [1974] assumed that atmospheric instability decays on small temporal and spatial scales. *Betts* [1986] recognized that a convecting atmosphere tends toward

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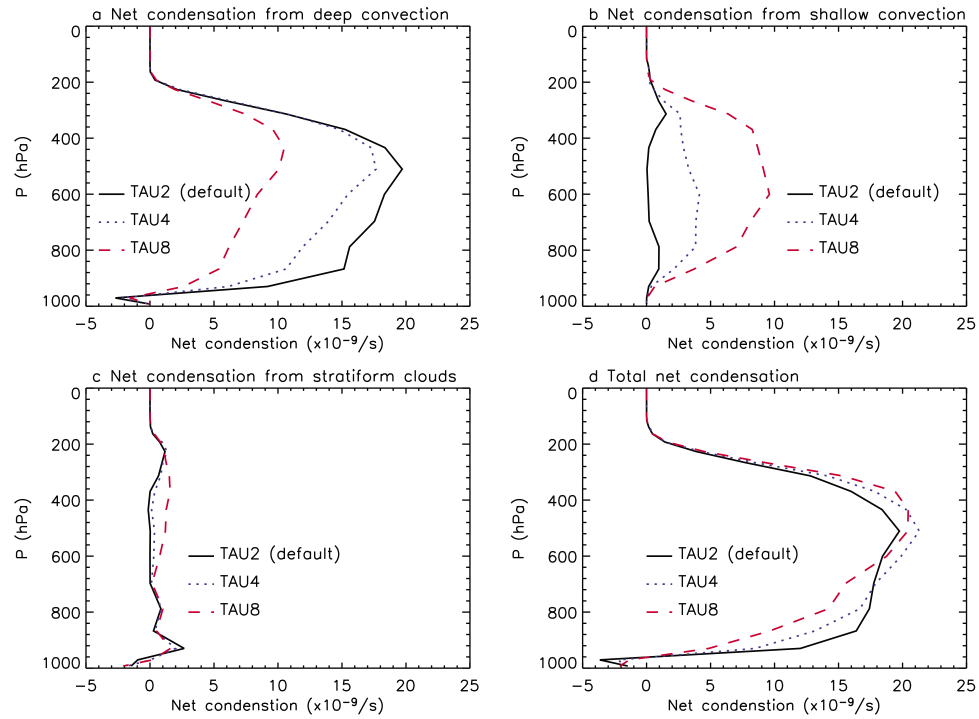

Figure 1. Vertical distribution of in-cloud condensation-evaporation from (a) deep convection, (b) shallow convection, (c) stratiform clouds, and (d) the sum of Figures 1a–1c, over Maritime Continents (80°E–160°E, 10°S–10°N) for a convective instability decay time (τ) of two hours (TAU2; black solid line), four hours (TAU4; blue dotted line), and eight hours (TAU8; red dashed line).

slightly unstable conditions relative to a moist virtual adiabat in order to keep the cumulus tower rising to the point where precipitation-sized particles are formed; thus a relaxation time of 2 hours state was assumed. However, using microwave precipitation and water vapor data, it has been shown that the convective adjustment time for the atmospheric moisture profile is strongly scale-dependent [*Bretherton et al.*, 2004]. Here, we consider the impact of varying τ and discuss how variations of water vapor isotope ratios with this parameter may provide insights on model convection schemes.

2. Method

[6] For this study we use NCAR CAM2, a detailed description of which is given by *Collins et al.* [2002]. *Lee et al.* [2008] describe the incorporation of the water isotopes into this model and note that the global distribution of water isotopes in precipitation is reasonably simulated, with discrepancies between observed and modelled $\delta^{18}O_p$ attributable to errors in the precipitation simulation [*Lee et al.*, 2008]. The isotope-enabled model has been used to interpret ice core and speleothem data [*Lee et al.*, 2007, 2009].

[7] NCAR CAM2 simulates three types of precipitation, deep convective, shallow convective, and large-scale by the Zhang and McFarlane (ZM) scheme [*Zhang and McFarlane*, 1995], the Hack scheme [*Hack*, 1994], and parameterized stratiform clouds, respectively. The ZM scheme invokes vertical mass fluxes based on the Arakawa-Schubert scheme [*Arakawa and Schubert*, 1974], with the convective closure obtained by relating instability (i.e., convective available potential energy, CAPE) to the cloud base mass flux. CAPE is consumed at a fixed time scale, τ, with time-evolution described by:

$$\frac{\partial A}{\partial t} = -\frac{A}{\tau} = -M_b F \tag{1}$$

Here, A is CAPE, M_b is cloud base mass flux, and F is CAPE tendency per unit mass. Numerically, the Hack shallow convection scheme is initiated if the atmosphere remains unstable following deep convective adjustment, while any moisture remaining after both deep and shallow convection is condensed as large-scale stratiform clouds.

[8] We obtained a suite of a sensitivity simulations with τ set to 2 hours (default: TAU2), 4 hours (TAU4), and 8 hours (TAU8). All experiments were forced with climatologically fixed sea surface temperatures (SSTs), and each simulation was integrated for 7 years, with the last 3 years analyzed below.

3. Results

[9] Vertical distributions of condensation for deep convection, shallow convection, and large-scale condensation from straform clouds are illustrated in Figure 1. As τ is increased, the net condensation from deep convection decreases while the net condensation from shallow convection increases. Overall, less (more) total condensation occurs in the lower (upper) troposphere with larger τ. Such changes can be understood by noting that in the ZM scheme, the

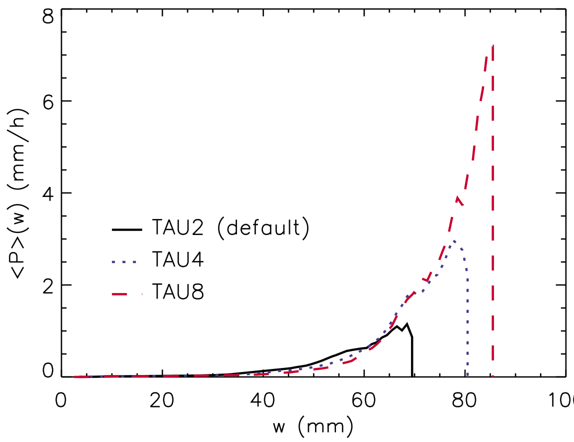

Figure 2. Mean precipitation rate at a give column integrated water vapor from model simulations with τ of TAU2 (black solid line), TAU4 (blue dotted line), and TAU8 (red dashed line) simulations. The data used are three hourly averages for tropical points (20°S–20°N). *Bretherton et al.* [2004] and *Peters and Neelin* [2006] show that column-integrated water vapor is tightly related to the surface precipitation rate.

cloud base mass flux determines how fast CAPE is consumed: with increased τ, the atmospheric instability decays more slowly by deep convection, and thus a larger portion of the initial instability is decayed by shallow convection.

[10] When τ is smaller, intense precipitation is limited because moist convection is triggered too often and the atmosphere cannot accumulate large amounts of water vapor. Thus, for the TAU2 run, the model simulates peak precipitation values of ~1 mm/hr, leading to what has been referred to as the "drizzle problem" of GCMs [*Dai*, 2006]. Given the greater accumulation of water vapor and instability with larger τ, the resultant precipitation is more intense (Figure 2), as is the mid-troposhere convective updraft velocity (Figure S2). As τ increased, the double ITCZ becomes less apparent, and some spurious features, such as precipitation over coastal Saudi Arabia, are diminished (Figure S3). However, the regions of most intense precipitation (e.g., the western Pacific warm pool) become too strongly convecting as τ is increased.

[11] Isotopes can be useful in diagnosing convection because strong fractionation occurs during phase changes, leading to a larger isotopic signal when condensation occurs. Because there is less condensation in the lower troposphere and more in the upper troposphere over the warm pool region around Indonesia as τ is increased (Figure 1d), δD is substantially higher in the lower troposphere (e.g., the difference between TAU2 and TAU4 cases is ~30‰ at 850 hPa) and lower aloft (Figure 3). However, because other processes including water vapor transport and post-condensation processes can also modify the vertical distribution of δD, the differences in δD do not exactly follow the condensation differences; in particular, in the upper troposphere, small changes in cloud water evaporation can substantially affect the δD in vapor [*Moyer et al.*, 1996]. In the lower free troposphere (850 to 700 hPa), re-evaporation has little impact because the layer is above the lifting condensation level (LCL).

[12] Figure 4 shows the spatial distribution of δD in vapor between 900 and 750 hPa from the three model cases (Figures 4a–4c) as well as the Tropospheric Emission Spectrum (TES) [*Worden et al.*, 2007] data at 825 hPa (Figure 4d). The latter comprises a 3-year mean covering April 2005 to March 2008. With the caveats that the TES analysis is in its early stages and there appears to be a high-bias over the ocean compared to the observations from field campaigns [*Lawrence et al.*, 2004], the observed data exhibit coherent features consistent with expectations based on the observed distribution of convection. For example, convection over tropical oceans tends to be more frequent and weaker than the convection over land regions [*Liu and Zipser*, 2005], and thus more condensation is expected at lower layers over tropical oceans and lower δD is observed there. Over the subtropical eastern Pacific, both simulated and TES δD values are extremely low, since condensation occurs around 825 hPa and there is a strong subsidence of low δD air from the upper atmosphere [e.g., *Lee et al.*, 2008].

[13] The largest differences among the three simulations are located over the Maritime continent, where the strongest simulated convection occurs. Overall, the TAU8 run provides the closest match to the TES distribution. Our analysis above suggests that the condensation level is too low for the TAU2 run and results in too low values of δD at 850hPa. Of course, there are discrepancies between the spatial structures in the simulations and TES data, e.g., all 3 runs exhibit local maxima at 20°S between Australia and Africa and in a region extending southeastward from 180°W, 0°S to 120°W, 20°S over the Pacific, both of which are absent in

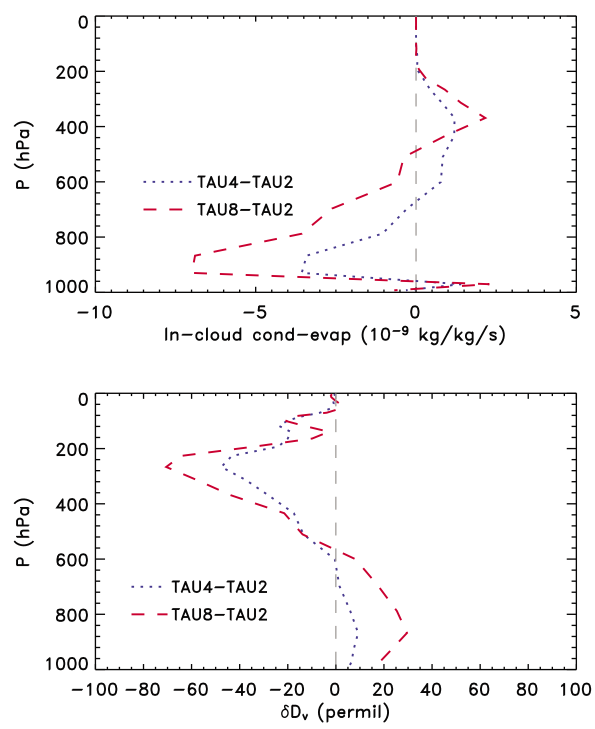

Figure 3. Vertical distribution of the differences in (top) condensation and (bottom) δD_v for TAU4-TAU2 (blue dotted line) and TAU8-TAU2 (red dashed line) over Maritime Continent (80°E–160°E, 10°S–10°N).

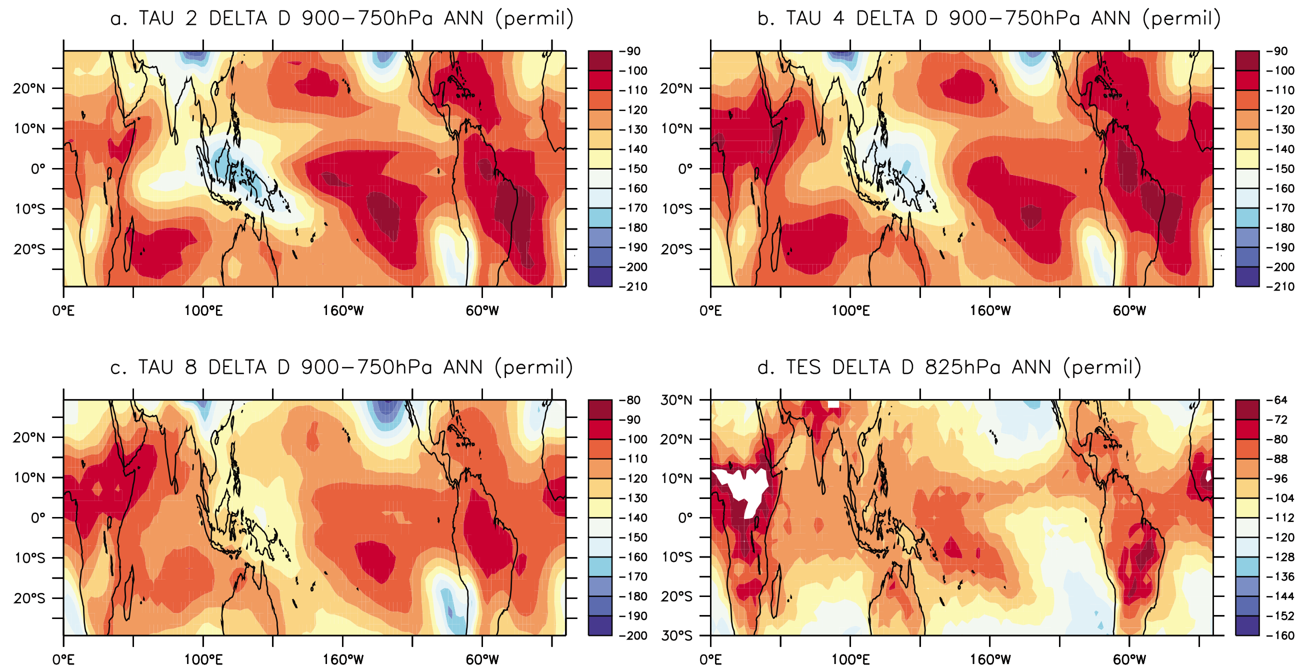

Figure 4. Mean annual δD_v in vapor between 900–750 hPa for the (a) TAU2, (b) TAU4, (c) TAU8 simulations and (d) TES observations (April 2005–March 2008) at 825 hPa. Note the differences in scales for the simulations and TES data. We speculate that the TES data capture the correct spatial distribution but that the absolute values may not be accurate. Missing data are represented as white area in the TES data.

the TES observations. The reasons for these discrepancies will be the focus of our future research.

4. Discussion and Conclusion

[14] The deep convective parameterization in NCAR CAM2, a modified Arakawa-Schubert convective scheme, is formulated based on the assumption that cloudy area occupies less than 5% of the grid. However, recent observations from Tropical Rainfall Measuring Mission (TRMM) show that a large fraction (>40%) of precipitation comes from meso-scale (>100 km in a least one direction [*Houze*, 2004]) organized convective systems, and overshooting area can be close to 5000 km^2 [*Liu and Zipser*, 2005]—much larger than 5% of grid area for the most widely used GCMs ($\sim$300 km). Thus, if most of the precipitation is simulated by a convection scheme based on the small cloudy area assumption—most of the precipitation around Indonesia comes from the deep convective scheme in the TAU2 run—the model cannot be reasonably expected to simulate properly the statistics of convection. In fact, to achieve reasonable precipitation rates over tropical regions, simulated convection must often be triggered too frequently. In turn, this can contribute to spurious behavior or biases such as the double ITCZ or too light rain [*Dai*, 2006]. We could eliminate some of these problems by increasing the contributions from shallow convection to stabilize the atmosphere in higher τ runs.

[15] Of course, optimizing a single climate variable may come at the expense of another; in fact, as *Jackson et al.* [2008] show, in the sense of a multivariate cost-function, a wide range of τ values are consistent with such a global optimization, suggesting the need for additional constraints. In this study, we have shown how the vertical distribution of δD, through its fractionation sensitivity to condensation rate, can provide insight into the vertical distribution of condensation, and in turn, the balance of convective processes necessary to achieve this distribution.

[16] Going forward, we note the challenge of measuring the vertical structure of δD on a global scale. However, the development and refinement of technologies for in situ measurements of isotopic composition of water vapor, together with mechanistic exploration such as described above, suggest strong potential for isotope-related diagnostics to constrain cumulus convection schemes.

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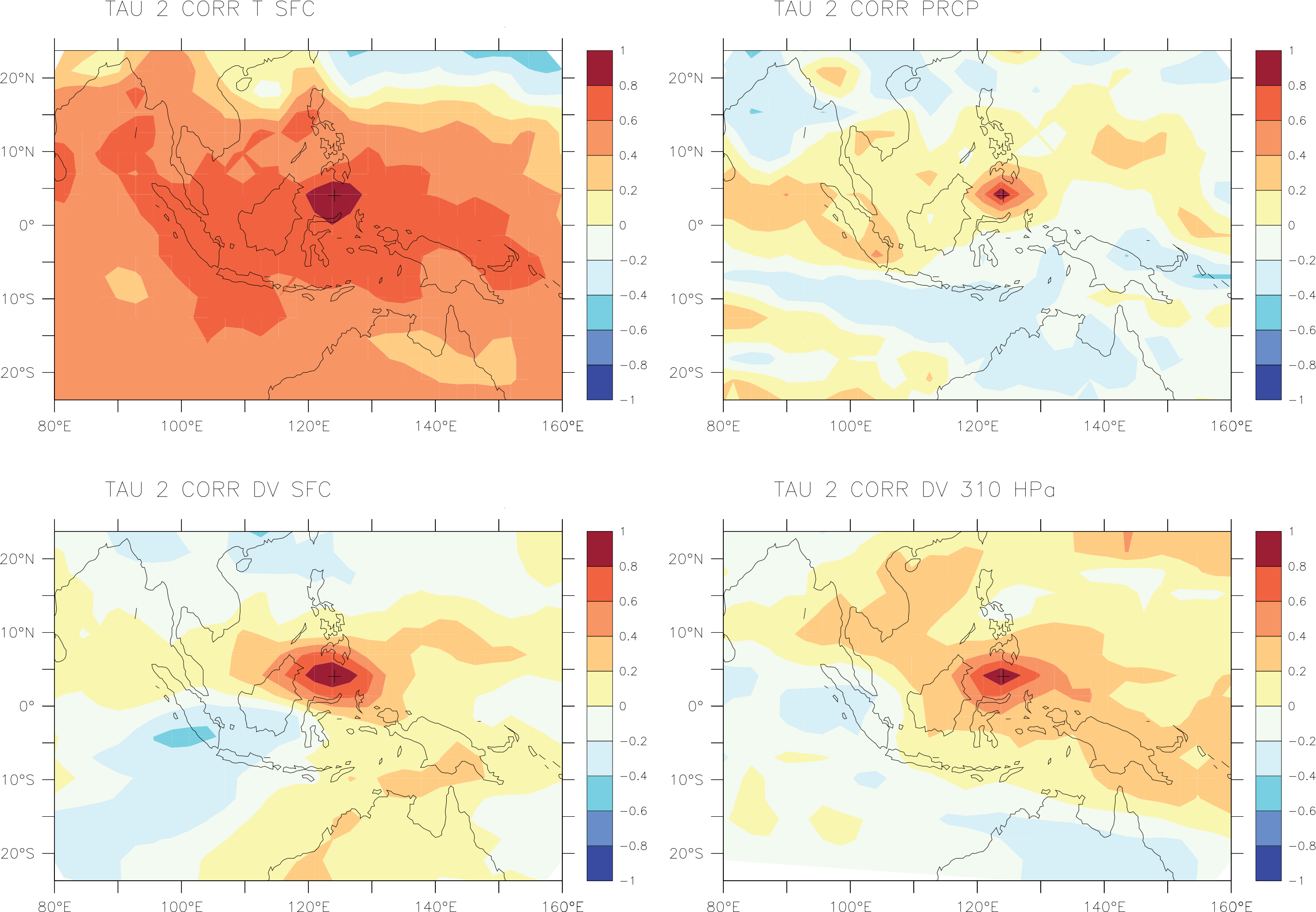
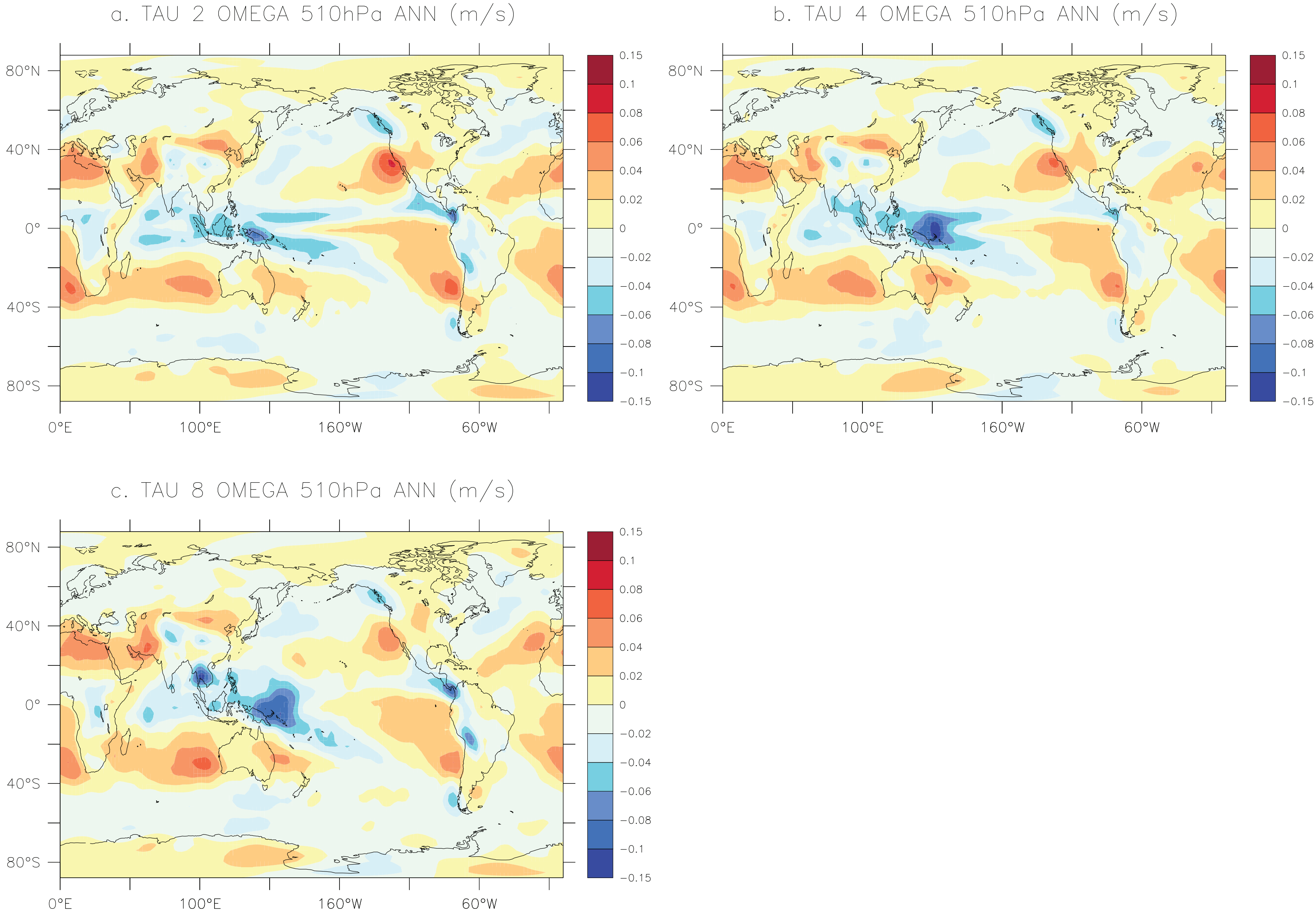
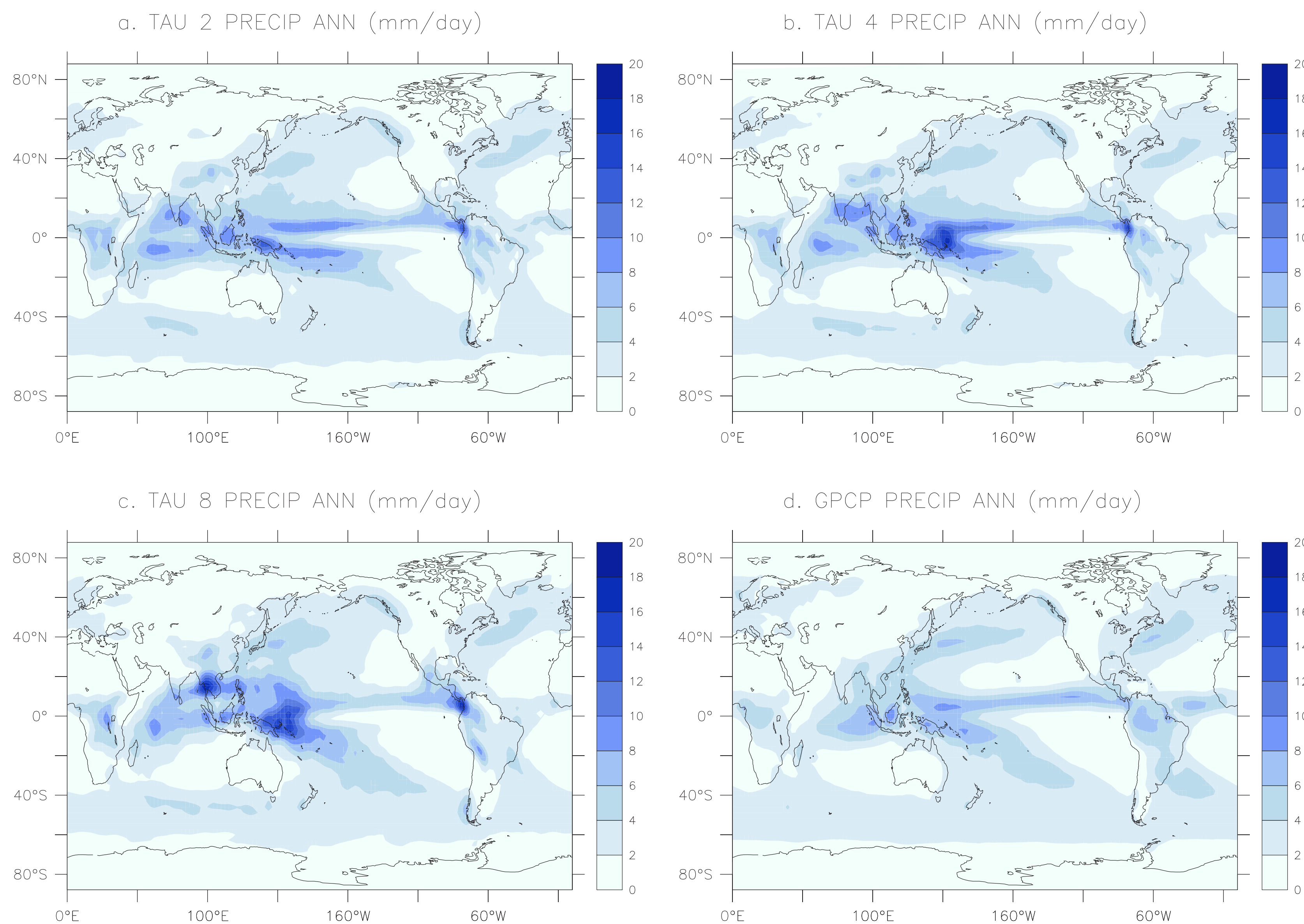

TAU 2 CORR T SFC
TAU 2 CORR PRCP
TAU 2 CORR DV SFC
TAU 2 CORR DV 310 HPa
20°N
10°N
0°
10°S
20°S
80°E
100°E
120°E
140°E
160°E
1
0.8
0.6
0.4
0.2
0
−0.2
−0.4
−0.6
−0.8
−1



a. TAU 2 OMEGA 510hPa ANN (m/s)
b. TAU 4 OMEGA 510hPa ANN (m/s)
c. TAU 8 OMEGA 510hPa ANN (m/s)
80°N
40°N
0°
40°S
80°S
0°E
100°E
160°W
60°W
0.15
0.1
0.08
0.06
0.04
0.02
0
−0.02
−0.04
−0.06
−0.08
−0.1
−0.15


a. TAU 2 PRECIP ANN (mm/day)
b. TAU 4 PRECIP ANN (mm/day)
c. TAU 8 PRECIP ANN (mm/day)
d. GPCP PRECIP ANN (mm/day)
80°N
40°N
0°
40°S
80°S
0°E
100°E
160°W
60°W
0
2
4
6
8
10
12
14
16
18
20