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Variations in atmosphere-ocean solar absorption under clear skies: A comparison of observations and models

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Abstract. Satellite observations of the clear-sky, atmosphere-ocean solar absorption are compared to simulations from a global climate model. Variations in solar absorption are analyzed under conditions of constant solar geometry to highlight spatially coherent features. From this analysis, the observed zonal and interannual variability in clear-sky solar absorption is shown to be substantially larger than that predicted by the model. Explanations for the greater observed variability are considered in terms of aerosols, surface wind effects, and water vapor.

Introduction

Determining the amount of shortwave (SW) radiation absorbed by the Earth is fundamental for understanding its climate. Several recent studies have suggested that model calculations significantly underpredict the amount of SW radiation absorbed in the Earth's atmosphere (Cess et al., 1995; Ramanathan et al., 1995). This discrepancy was attributed to shortcomings in our current knowledge of the absorption by clouds, whose role in absorbing SW radiation has been of long-standing debate (Stephens and Tsay, 1990; Li et al, 1995). Similar discrepancies in model estimates of SW absorption have been identified in other studies, but attributed to clear-sky aerosols (Barker and Li, 1995) and water vapor (Arking, 1996) rather than clouds. Some field measurements also find discrepancies under clear skies (Charlock et al., 1996), while other studies find good agreement when aerosols are included (Waliser et al., 1996; Cess. et al., 1996; Conant et al., 1997).

In this report satellite observations of the clear-sky, top-of-atmosphere (TOA) SW absorptivity A are obtained from the Earth Radiation Budget Experiment (ERBE; Barkstrom et al., 1989). Here A is defined as $A = (S{\downarrow}-S{\uparrow})/S{\downarrow}$, where $S{\uparrow}(S{\downarrow})$ is the clear-sky upward (downward) SW flux at the TOA, or equivalently $A=(1-\alpha)$, where α is the clear-sky TOA albedo. The ERBE measurements are compared to observations of column integrated water vapor (W; Greenwald et al., 1993) and surface wind speed (V; Wentz, 1989) from the Special Sensor Microwave Image (SSMI), and aerosol optical depth (τ) from the Advanced Very High Resolution Radiometer (AVHRR) using the phase-2 retrieval of Stowe et al. (1997). This study focuses on periods of 1987 and 1988 when all three satellite data sets are all available and will be restricted to ocean surfaces since W, V, and τ are not retrieved over land. The observed variability is then compared to that predicted by a Global Climate Model (GCM; Wetherald et al., 1991) integrated using observed SSTs for 1979-1988.

If absolute values of SW absorption are compared, rather than variations, the level of agreement between the GCM and observations is relatively good (e.g., Fig. 1) and similar to that found by Kiehl et al. (1997). Such comparisons are largely a reflection of the model's ability to compute the surface albedo and its dependence on solar zenith angle. By examining the variability in A under constant solar geometry, this study performs a more critical assessment of the model's skill in accounting for atmospheric variables which influence the TOA SW absorption.

Observed and GCM-Simulated SW Absorptivity

The absorbed SW radiation at the TOA is determined by absorption both within the atmosphere and at the surface. The ocean surface albedo depends strongly on the solar declination angle which varies substantially with latitude and season. Therefore, to reduce the impact of changes in ocean albedo it is useful to consider variations in A under conditions in which the solar declination angle is constant. For this purpose we consider two sources of variability: (i) interannual variations associated with the 1987 El Nino/Southern Oscillation (ENSO) and (ii) zonal variations around individual latitude belts. It will be shown that both analyses lead to similar conclusions.

Figure 2 depicts the zonal deviations in the GCM-simulated (top) and satellite-observed (middle, bottom) quantities. The zonal deviations are computed by subtracting the zonal-mean from the value at each grid location; i.e. $\delta A = A - [A]$, where $[A]$ denotes the zonally-averaged value of A. All quantities are annual means for 1988. The analysis is restricted to latitudes equatorward of 45°, since ERBE clear-sky observations poleward of this region are poorly sampled due to persistent cloud cover. In the GCM, coherent large-scale patterns are evident with regions of positive δA coinciding with regions of positive δW and vice-versa. Other quantities affecting A (e.g., ocean albedo) do not vary zonally in the GCM, consequently the GCM-simulated δA is solely due to zonal variations in absorption by water vapor. The correlation between GCM-simulated δA and δW is r_{AW}=0.96, indicating that increased water vapor is associated with increased absorption of SW radiation. (To reduce the sensitivity to outlying data, all correlation coefficients are computed using the Spearman rank method and all slopes are computed using least absolute deviations).

The observed pattern of δA also resembles the pattern of δW (Fig. 2, middle) such that increased moisture coincides with increased SW absorption. The lower correlation (r_{AW}=0.43) with

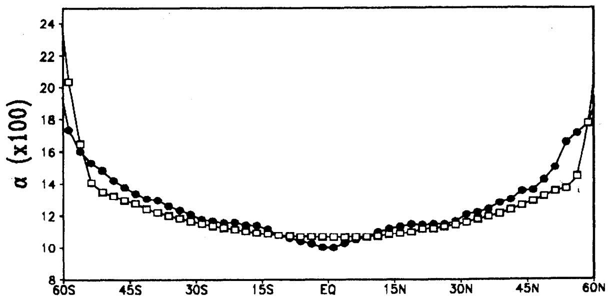

Figure 1. The zonal-mean TOA albedo (in %) from ERBE (filled) and the GCM (open).

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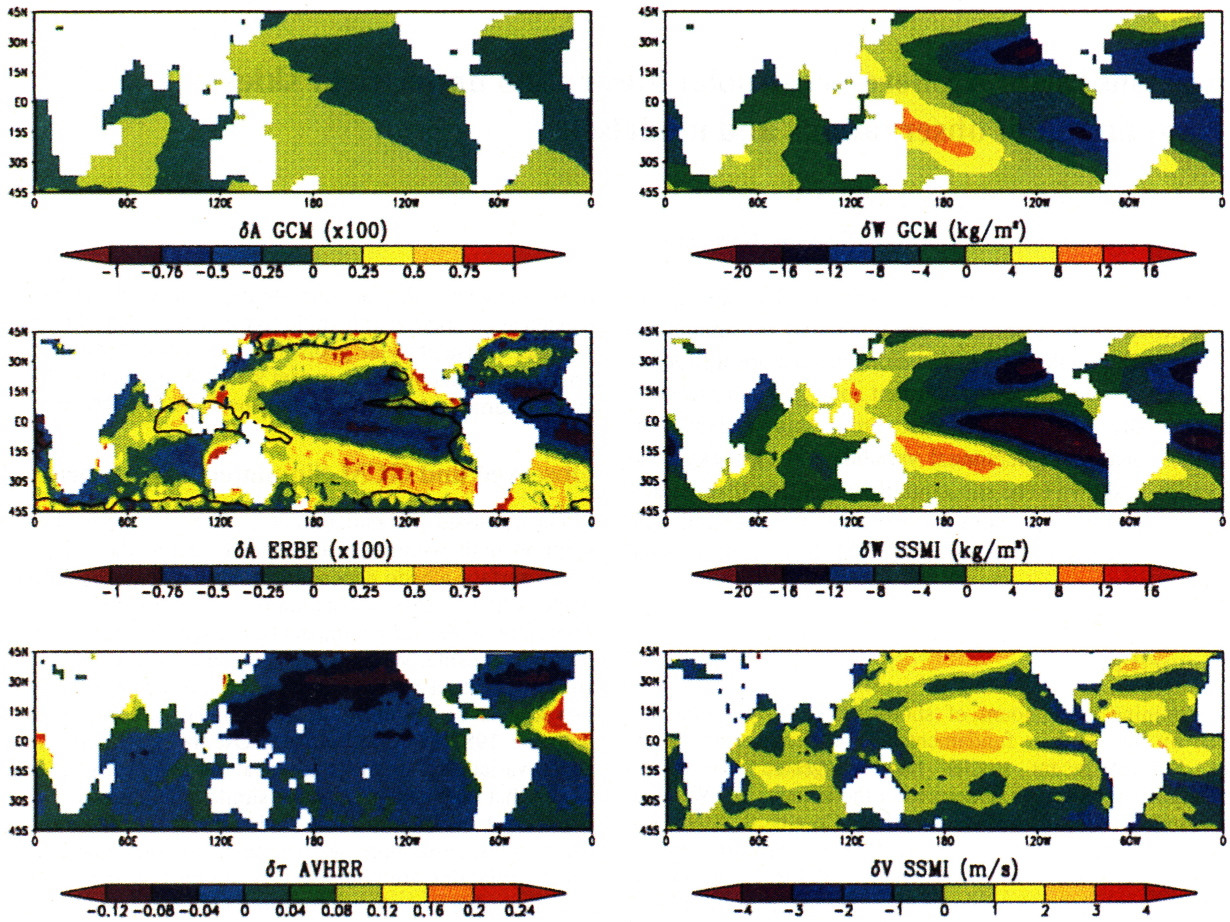


Figure 2. The zonal deviation in annual-mean: GCM absorptivity δA (top left), GCM water vapor δW (top right), observed absorptivity δA (middle left), observed water vapor δW (middle right), observed aerosols (bottom left), and observed wind speed δV (bottom right). The black line in the observed δA delineates regions which averaged fewer than one clear-sky observation every other day over the 1-year period and are excluded from the regression in Figure 3.

respect to the GCM's partly reflects the greater small-scale spatial variability in the ERBE δA. The observed variability in A is much larger than that predicted by the GCM, even for similar changes in W. Linear regression of δA versus δW (Fig. 3) yields a slope of 3.3 x10^{-4} m^{2}kg^{-1} for the satellite observations (black dots), but only 0.8 x10^{-4} m^{2}kg^{-1} for the GCM simulations (red dots). Note that any underestimate in atmospheric SW absorption would be largely compensated for by an increase in surface absorption. This must be considered when appraising the relatively small magnitudes of δA presented here.

The SW radiative transfer parameterization used in the GCM is based on that developed by Lacis and Hansen (1974). This parameterization slightly underestimates the SW absorption by water vapor compared to line-by-line (LBL) results (Fouqart et al., 1991). The dependence of δA on δW predicted by this model was compared to the GFDL LBL algorithm (Ramaswamy and Freidenrich,1992) and found to be in good agreement (slope of $\delta A/\delta W$ within 10% of LBL). Thus, the differing slopes in Figure 2 can not be attributed to a water vapor-related deficiency in the GCM's radiation scheme.

However other factors which influence A, such as tropospheric aerosols, are not accounted for by the GCM. Consider the observed variations in aerosol optical depth $\delta\tau$ (bottom left). Regional signatures of aerosol radiative effects are clearly present in the ERBE data. Enhanced SW absorption (reduced reflection) over the north Pacific and north Atlantic oceans coincides with bands of low optical depth. Likewise, enhanced aerosols over the equatorial Atlantic and Arabian Sea coincide with reduced SW absorption (enhanced reflection). However the agreement between δA and $\delta\tau$ is less impressive over the Southern Hemisphere where variations in SW absorption occur in the absence of any corresponding changes in optical depth.

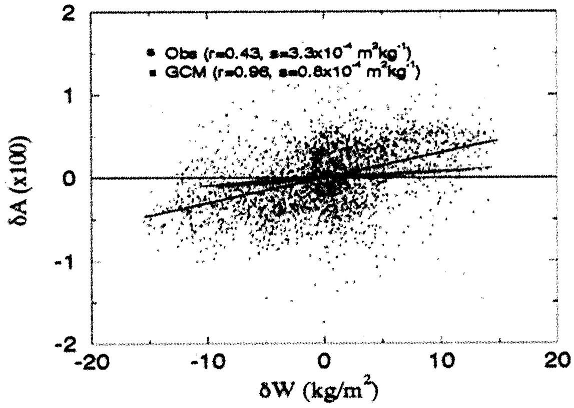
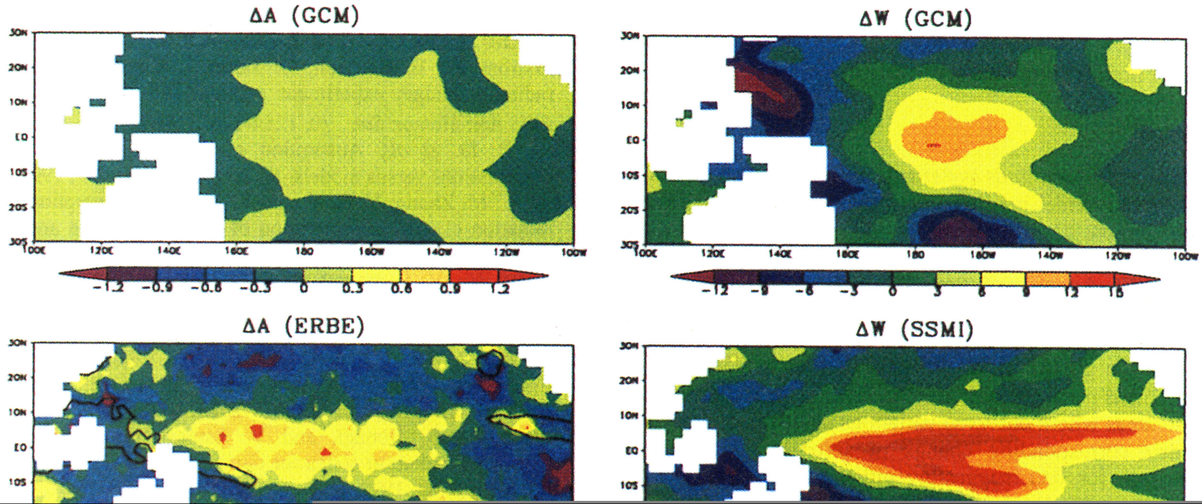


Figure 3. A plot of δA versus δW for the observations (black) and the GCM predictions (red).


ΔA (GCM)
ΔW (GCM)
30N
20N
10N
EQ
10S
20S
30S
100E
120E
140E
160E
180
160W
140W
120W
100W
−1.2 −0.9 −0.6 −0.3 0 0.3 0.6 0.9 1.2
−12 −9 −6 −3 0 3 6 9 12 15
ΔA (ERBE)
ΔW (SSMI)

salt. Although the possibility that wind-driven changes in the ocean surface albedo could introduce similar biases in the retrieval of both ΔA and $\Delta\tau$ can not be ruled out.

Summary

This investigation finds systematic discrepancies between the observed and model-predicted variability of atmosphere-ocean SW absorption. The discrepancies identified here occur under clear-sky conditions and are consistently manifest in both the interannual and zonal variations. We infer three possible explanations for the discrepancies:

(i) Aerosols. The presence of aerosols seems likely to contribute to the greater observed variability in A relative to the GCM. This is particularly evident in ΔA (Fig 4). However, there are also areas where observed variations in A are not associated with any corresponding change in aerosols (e.g. Southern Hemisphere for δA), suggesting either that aerosols are not solely responsible for the discrepancy or that their variability is underestimated in current satellite measurements.

(ii) Surface winds. The larger observed variability in A is also associated with changes in surface wind speed which can alter the ocean surface albedo. However, the sign of the relationship between wind speed and ocean albedo contradicts that deduced from both measurements and present theory. Little is known regarding the effects of whitecaps, although they are generally considered to have a negligible effect on α due to their small areal coverage (Monahan, 1970). A more likely explanation is that increased wind are associated with enhanced marine aerosols which decrease A. This is supported by field studies (O'Dowd and Smith, 1993) and, to some extent, by the satellite-observed aerosols (Fig. 4). However, there is little resemblance between δV and $\delta\tau$ for the Southern Hemisphere (Fig. 2), perhaps reflecting a difficulty in remotely sensing aerosols for this region. If true, this could explain some of the large observed variations in δA that are not present in the GCM or in the satellite aerosol measurements. The possibility of wind-speed dependent errors in the ERBE retrieval of SW fluxes should also be considered.

(iii) Water vapor. The larger observed variability in A is also associated with changes in water vapor, suggesting that the discrepancy may, in part, be attributable to a systematic underestimate of SW absorption by water vapor in current radiative transfer models. The extent to which water vapor is responsible for the discrepancy (if at all) can not be assessed here. While this explanation is unlikely, the similarity between the observed variations in A and W, suggests that its quantitative role warrants further scrutiny alongside aerosol and surface wind effects. Given the difficulty of unraveling the relative contributions of τ, V, and W to the observed variations in A, in-situ measurements of all four quantities will like be needed to fully resolve this dilemma. We note that analyses of recent field measurements (Conant et al., 1997; Zender et al., 1997) show no evidence to suggest that the magnitude of clear-sky water vapor absorption is deficient in existing models.

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