The radiative impact of a simple aerosol climatology on the Hadley Centre atmospheric GCM

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Summary

Results from a new version of the Hadley Centre atmospheric general-circulation model are presented in which the sensitivity of the model to the inclusion of a simple aerosol climatology is investigated. Without aerosols, comparisons with clear-sky reflected short-wave radiation from the Earth Radiation Budget Experiment (ERBE) and the Scanner for Radiation Budget (ScaRaB) satellite measurements indicate a missing scatterer in the clear-sky atmosphere. In addition, recent estimates of global, annual mean absorbed short-wave radiation at the surface, which used the Global Energy Balance Archive (Geba) where possible for verification, indicate that the atmosphere did not absorb enough short-wave radiation. The insertion of a simple aerosol climatology based on World Climate Research Programme recommendations into the model was found to be the only plausible means of reducing its significant short-wave bias compared with ERBE and ScaRaB measurements at the top of the atmosphere, and with GEBa data at the surface. An increase in the global, annual mean atmospheric absorbed short-wave radiation accompanied these changes.

Keywords: Aerosol climatology General-circulation model Top-of-the-atmosphere radiation Surface radiation balance

1. Introduction

Atmospheric aerosols play an important role in the radiation budget of the planet (Hobbs and McCormick 1988). They scatter and absorb solar radiation, causing an increase in the planetary albedo and a reduction in the amount of radiation which reaches the surface. The impact on thermal radiation is generally smaller, although it can be significant (Lacis and Mishchenko 1995). Anthropogenic aerosols are also believed to reduce the magnitude of global warming resulting from the increasing concentrations of greenhouse gases (Houghton et al. 1995).

The first studies of the effects of aerosol loadings on simulations with general-circulation models (GCMs) were performed by Td\textsuperscript{e}ne et al. (1984) and by Coakley and Cess (1985). In each study the impact on radiative fluxes and heating rates was significant and there were changes in land surface temperatures, but the use of fixed sea surface temperatures meant that the impact on most of the model variables was small. Given these results, it is not surprising that aerosols have often been ignored in GCMs, except recently in models used for climate-change studies. For example, only seven of the 30 models contributing to the first phase of the Atmospheric Model Intercomparison Project (AMIP) included any representation of aerosols (Phillips 1994).

Until recently, the radiative effects of aerosols were not included routinely in the Meteorological Office Unified Forecast/Climate Model (Cullen 1993), but with the introduction of a new radiation code into the climate version of the model (Edwards and Slingo 1996) it became possible to treat aerosols realistically (Haywood et al. 1997). Independent evidence indicating that aerosols needed to be included came from comparisons with data from the Earth Radiation Budget Experiment (ERBE, Harrison et al. 1990) and the Scanner for Radiation Budget (ScaRaB, Kandel et al. 1998). As discussed in section 3, the clear-sky planetary albedo was lower than that from the ERBE and ScaRaB and the distribution of the deficit suggested that the lack of aerosols made a substantial contribution.

Much attention has been given recently to the question of how much of the incident solar irradiance is absorbed by the atmosphere and how much by the surface in GCMs, particularly since the development of the Global Energy Balance Archive (Geba, Ohnira et al. 1989; Gilgen et al. 1998). Garratt (1994) and Wild et al. (1995a) found an overestimate in incoming short-wave radiation at the surface for various GCMs, which appeared to be due to an underestimate of atmospheric short-wave absorption. Such discrepancies have been cited as evidence for the existence of enhanced absorption of solar radiation by clouds, although it is also likely that the use of out-dated water vapour absorption data and the neglect of aerosols make a contribution (Wild et al. 1996a; Li et al. 1997).

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In this paper we show the radiative impact that an aerosol climatology has on our climate model, with reference to the ERBE and ScaRaB data at the top of the atmosphere (TOA) and with the GEBA data at the surface. This climatology is based upon the World Meteorological Organization (WMO) (1982, 1983) recommendations, and despite its crude spatial and temporal resolution of aerosols, model biases at the TOA and surface are reduced significantly. Current work within the Hadley Centre, developing a prognostic means of specifying aerosol amounts and size distributions, should eventually replace this simple climatology.

Section 2 gives a brief outline of the Hadley Centre GCM. In section 3 the model performance without aerosols is compared with ERBE and ScaRaB clear-sky albedos over oceans and with best estimates of global, annual mean absorbed short-wave radiation at the surface, and these results are discussed. Section 4 gives an outline of the aerosol climatology inserted into the model, and the performance of this new model is presented in section 5. The significance of the remaining differences compared with observations is discussed in section 6, and possible causes are suggested. A summary of this work is given in section 7.

2. DESCRIPTION OF THE GCM

This study uses a development version of the latest Hadley Centre atmospheric GCM, known as HadAM3 (Hadley Centre Atmospheric Model—version 3). HadAM3 is a grid-point model with a horizontal resolution of 2.5° latitude and 3.75° longitude, using Eulerian dynamics as described by Cullen (1993). There are 19 levels in the vertical, specified using a hybrid vertical coordinate which moves from being terrain following (sigma) at the surface to isobaric at 40 hPa and above. One major difference from earlier versions of the model is the inclusion of the new radiation scheme, described by Edwards and Slingo (1996). A brief description of the version of the scheme used here is given below. Other major improvements compared with previous versions of the model include the treatment of convective momentum transport (Gregory et al. 1997) and a new land-surface scheme. A more complete description of the model and its climatology is in preparation.

The radiation scheme uses the two-stream approximation in both the short wave and long wave, following Zdunkowski et al. (1982): the Practical Improved Flux Method (PIFM) of Zdunkowski et al. (1980) has been chosen to solve these equations. Absorption due to H2O, CO2, O3, N2O, CH4, CFC-11 and CFC-12 are accounted for in the long wave, which has eight spectral bands, and H2O, CO2, O3 and O2 are accounted for in the short wave, which has six spectral bands. The foreign and self-continua of H2O are parametrized in the long wave using the Clough et al. (1989) formulation, version 2.1: the short wave makes use of the H2O foreign continuum alone. The Exponential Sun Fitting Technique used to generate the quasi-monochromatic absorption coefficients of the absorbing gases in the atmosphere, as described by Edwards and Slingo (1996), has been replaced for the major gases in the long wave and for H2O and O2 in the short wave by correlated k-distribution (c-k) theory (Cusack et al. 1998).

3. MODEL PERFORMANCE WITHOUT AEROSOLS

All previous versions of the Unified Model ignored aerosols, except in climate-change simulations where the effects of aerosols of anthropogenic origin were approximated. Here we describe the performance of HadAM3 without aerosols (HadAM3-NA), which provided the evidence that they needed to be included. HadAM3-NA was run in an AMIP-type mode (Gates 1992) with sea surface temperatures (SSTs) and the sea-ice distribution prescribed from the period December 1978 to February 1982. There are systematic biases in the radiation budget of HadAM3-NA when compared against observations, and these are discussed below.

(a) Clear-sky reflected short-wave radiation at the top of the atmosphere over oceans

The first bias concerns the clear-sky reflected short-wave radiation at the TOA in clear skies over oceans. Figure 1(a) shows the difference between the mean of three Julys from the HadAM3-NA (1979–81), and of three Julys from the ERBE (1986–88) over oceans only, whilst Fig. 1(b) shows the difference between the HadAM3-NA (1979–81) and that of the ScaRaB for July 1994 (only one year of data was obtained by the ScaRaB). The algorithm used to identify clear-sky scenes in the ScaRaB data is exactly the same as that used in the processing of the ERBE data. ERBE errors are given by Ramanathan et al. (1989) as being ±5 W m−2 on a monthly average on a regional basis. Biases over land are not plotted in Figs. 1(a) and (b) because of the possibility of sizable errors in the surface albedo, whereas the ocean albedo formulation, which uses the Briegleb et al. (1986) parametrization of the direct solar beam and a value of 0.06 for the diffuse beam, is known with greater accuracy (see below).
Figure 1. Difference in clear-sky reflected short-wave radiation over the oceans between (a) HadAM3 without aerosols and the ERBE data (mean of three Julys, 1986–88), and (b) between HadAM3 without aerosols and the ScaRaB data (July 1994). Both model results were for a 3-year July mean, 1979–81. Contours every 5 W m$^{-2}$, stippling for values less than $-10$ W m$^{-2}$. Missing data values have very little areal extent and are not shown.

Glew et al. (1998) compare many aircraft observations with ocean albedo models, and show that Briegleb et al.'s (1986) parametrization would underestimate ocean albedo by about 0.3% for solar zenith angles between 0$^\circ$ and 45$^\circ$, relative to their aircraft observations: however, Figs. 1(a) and (b) indicate that it requires an underestimate of about 3% in absolute terms to account for the differences. Therefore, deficiencies in the ocean albedo formulation cannot account for the differences in Figs. 1(a) and (b), and the explanation must be sought in the clear-sky radiative parametrization of the atmosphere. However, a comparison of the radiation code in the climate model with a reference code which used 220 spectral bands indicated that the reduction in spectral resolution necessary for climate models was not the cause of the error seen in Figs. 1(a) and (b).

The two-stream formulation used in the model must be considered as another potential cause of the disagreement with the ERBE and ScaRaB data. King and Harshvardhan (1986) investigated the accuracy of a number of two-stream approximations including the delta-Eddington approximation, which is very similar to the PIFM used here, and found that in optically thin layers the plane albedo was underestimated at small zenith angles, but overestimated at large zenith angles. Given the very low albedo of the atmosphere, the relative errors reported by King and Harshvardhan indicate that deficiencies in the two-stream approximation are not a sufficient explanation of the discrepancy between the model and satellite observations.

The clear-sky algorithm used to process satellite observations records any scene with less than 5% cloud as being clear sky: assuming that cloud cover can be identified accurately, cloud contamination cannot be responsible for the differences plotted in Figs. 1(a) and (b).

(b) The net downward short-wave radiation at the surface

A second bias concerns the net downward short-wave irradiance at the surface. Li and Leighton (1993) calculate a value of 157 W m$^{-2}$ as the global and annual average absorbed short-wave radiation by
the surface, using the algorithm extensively validated by comparison with the GBEA in Li et al. (1995). This algorithm, detailed in Li et al. (1993), empirically relates surface absorbed short-wave flux to the reflected short-wave flux at the TOA using solar zenith angle and precipitable water. Wild et al. (1996a) use a different method to calculate a global and annual average of the absorbed short-wave radiation at the surface, using GBEA data to determine the bias of the Max Planck Institute GCM (ECHAM4), then adjusting the estimate from ECHAM4 accordingly. They obtain a value of about 152 W m$^{-2}$ for the global, annual mean surface absorbed short-wave radiation using this method. Significant uncertainties exist in both of the above estimates, because GBEA sites are sparse in tropical and southern hemisphere land areas, and particularly so in oceanic regions. It seems that 155 W m$^{-2}$ is a reasonable estimate of the global, annual mean absorbed short-wave radiation by the surface, and that error limits of ±10 W m$^{-2}$ are justified by our lack of observations of this quantity over tropical land and all oceans. HadAM3-NA needs slight adjustment to produce radiative balance at the TOA, and if this is done it gives a value of about 171 W m$^{-2}$ for the global, annual mean absorbed short-wave radiation by the surface. This is higher than the GBEA estimate, indicating a clear lack of atmospheric absorption in HadAM3-NA relative to the estimates verified by the GBEA.

4. THE AEROSOL CLIMATOLOGY INSERTED INTO HADAM3

A very simple climatology was created, closely related to the recommendations of the WMO (1982, 1983). A much more complex aerosol climatology exists, see D'Almeida et al. (1991), but we have decided to explore the first-order effects of aerosols with a simple climatology, and to use the results as a guide for further work.

We describe here the small changes from the full climatology detailed in WMO (1982, 1983). We chose from two atmospheric profiles at all grid points, CONT-I and MAR-I. CONT-I is used at all land-surface points, except those points which have permanent snow cover which use the MAR-I profile. All points over the open ocean and sea ice use the MAR-I profile. These two profiles are split into three main sections vertically: the boundary layer, the free troposphere and the stratosphere. There are five aerosol components which are externally mixed to create the aerosol mixture in each atmospheric section of CONT-I and MAR-I; namely water-soluble, dust, soot, oceanic, and stratospheric sulphates. The complex indices of refraction and size distributions for each aerosol component (WMO 1982) were used to calculate the optical properties necessary for a two-stream radiation code using Mie theory. Table I lists the single-scatter albedo and asymmetry factor at 0.55 μm for the five aerosol components.

The boundary layer is defined as the region extending from the surface to the top of the fifth layer of the model, at approximately 750 hPa for standard surface pressure. The free troposphere extends from the top of the boundary layer to the tropopause, which is diagnosed from an examination of the temperature lapse rate, and the stratosphere is defined, for this climatology, as extending from the tropopause to the top of the second highest layer (i.e. 10 hPa). A surface-pressure scaling factor has been applied to aerosol profiles in the troposphere, by asserting that the mass mixing ratios as calculated using all the above information be scaled by a factor $p_0/1.013 \times 10^3$, at every grid point, where $p_0$ is the surface pressure in Pascals. This ensures that there is less aerosol optical thickness in the troposphere at grid points with an elevated surface, as is generally observed (e.g. Nakajima et al. 1996).

| TABLE I. OPTICAL PROPERTIES AT 0.55 μM FOR THE FIVE AEROSOL COMPONENTS OF THE GLOBAL CLIMATOLOGY |
|-------------------------------------------------|-----------------|-----------------|-----------------|-----------------|-----------------|
| | Water-soluble | Dust | Soot | Oceanic | Stratospheric sulphates |
| Single-scatter albedo | 0.963 | 0.728 | 0.209 | 0.9999 | 0.9999 |
| Asymmetry factor | 0.638 | 0.828 | 0.337 | 0.822 | 0.696 |

5. IMPACT AND ASSESSMENT OF THE AEROSOL CLIMATOLOGY ON HADAM3

(a) Clear-sky reflected short-wave radiation over oceans

The comparison of HadAM3 with satellite measurements of the clear-sky reflected short-wave radiation at the TOA over oceans is shown in Figs. 2(a) and (b), meaned over the same periods as for Figs. 1(a) and (b). There is a clear improvement in this quantity, globally and at almost all points locally, when the simple aerosol climatology is included. The global mean model error over sea points only has
decreased from 9.65 W m$^{-2}$ to 3.97 W m$^{-2}$ relative to the ERBE; the corresponding values for the ScaRaB are 7.82 W m$^{-2}$ and 2.08 W m$^{-2}$ respectively. Section 6 discusses the remaining differences between HadAM3 and satellite measurements. Table 2 indicates that the aerosol type over land, CONT-I, produces a significantly smaller signal in reflected short-wave radiation at the TOA compared with MAR-I.

(b) Global short-wave radiation absorbed by the surface

From Table 2 it can be seen that HadAM3 is not in radiative balance at the TOA. From our experience, to produce such a balance would reduce the global annual mean surface absorbed short-wave radiation from 167.7 W m$^{-2}$ (Table 2) to about 165 W m$^{-2}$. Therefore, the inclusion of a simple aerosol climatology has led to a significant improvement in the global and annual mean solar radiation budget at the surface, although this value still lies near the uppermost limit of possible values of this quantity derived from the GEBa.

Figure 3 shows the geographical distribution of the 3-year annual mean change in all-sky atmospheric absorption due to the aerosol climatology. The major feature is that of the contrast between the MAR-I and CONT-I profiles, with the latter absorbing considerably more solar radiation. Note that the reduction in short-wave heating of the surface is slightly compensated by an increase in heating in the long wave (see Table 2), and that this long-wave heating would be greater if SSTs were not prescribed as they were in these integrations. It could be anticipated that the long wave would counteract a significant part of the short-wave induced surface cooling if SSTs were variable.

(c) Direct comparison of downwards short-wave irradiance at specific GEBa sites

For the comparison of the GCM-calculated surface fluxes with land-based observations, the model data were interpolated to the measurement sites using the four surrounding grid points, weighted by their inverse spherical distance. The downwelling short-wave irradiance at the surface is used here, as this is the most frequently measured quantity, and is also substantially affected by aerosols. Approximately 700 sites
TABLE 2. Global, 3-year mean fluxes of HadAM3-NA (first three columns) and the flux change caused by the aerosol climatology (last three columns, calculated as HadAM3 minus HadAM3-NA). Units are W m\(^{-2}\).

<table>
<thead>
<tr>
<th></th>
<th>Global flux</th>
<th>Land flux</th>
<th>Ocean flux</th>
<th>Global ΔF</th>
<th>Land ΔF</th>
<th>Ocean ΔF</th>
</tr>
</thead>
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<tr>
<td>Values at the surface</td>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
</tr>
<tr>
<td>Downward SW</td>
<td>201.02</td>
<td>204.76</td>
<td>199.50</td>
<td>-8.27</td>
<td>-13.32</td>
<td>-6.21</td>
</tr>
<tr>
<td>Downward SW (clr)</td>
<td>261.45</td>
<td>256.57</td>
<td>263.45</td>
<td>-11.67</td>
<td>-17.76</td>
<td>-9.18</td>
</tr>
<tr>
<td>Absorbed SW</td>
<td>174.49</td>
<td>150.76</td>
<td>184.16</td>
<td>-6.81</td>
<td>-10.23</td>
<td>-5.41</td>
</tr>
<tr>
<td>Absorbed SW (clr)</td>
<td>228.01</td>
<td>191.63</td>
<td>242.84</td>
<td>-9.57</td>
<td>-13.69</td>
<td>-7.89</td>
</tr>
<tr>
<td>Downward LW</td>
<td>331.46</td>
<td>297.57</td>
<td>345.26</td>
<td>1.27</td>
<td>1.23</td>
<td>1.29</td>
</tr>
<tr>
<td>Downward LW (clr)</td>
<td>310.36</td>
<td>279.83</td>
<td>322.80</td>
<td>1.77</td>
<td>1.65</td>
<td>1.83</td>
</tr>
<tr>
<td>Net upward LW</td>
<td>64.12</td>
<td>71.96</td>
<td>60.92</td>
<td>-1.74</td>
<td>-3.00</td>
<td>-1.23</td>
</tr>
<tr>
<td>Values at the top of the atmosphere</td>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
</tr>
<tr>
<td>Outgoing SW</td>
<td>95.87</td>
<td>111.25</td>
<td>89.60</td>
<td>2.26</td>
<td>0.83</td>
<td>2.84</td>
</tr>
<tr>
<td>Outgoing LW</td>
<td>238.90</td>
<td>232.30</td>
<td>241.59</td>
<td>-0.22</td>
<td>-0.25</td>
<td>-0.21</td>
</tr>
</tbody>
</table>

The following abbreviations are used in column 1:

SW — short-wave radiation;
LW — long-wave radiation;
clr — clear sky.

Figure 3. The effect of aerosols on the all-sky atmospheric absorption (W m\(^{-2}\)), meaned over 3 years.

with long-term (more than 5 years) monitoring of the downwelling short-wave irradiance at the surface are used: the stations are representative of their larger-scale setting as outlined by Wild et al. (1995b, 1996b), and their geographical distribution is given by Wild et al. (1995a). The accuracy of such measurements is estimated as being about 4 W m\(^{-2}\) (or 2%, Gilgen et al. 1998).

Figure 4(a) shows the zonal mean difference of the all-sky (i.e. clear and cloudy sky) annual mean downwelling irradiance between the GBEA observations and the model for the integrations with and without the aerosol climatology. Model values were first interpolated to GBEA sites using the method mentioned in the previous paragraph and the differences calculated, then the zonal means were determined in 5° latitude belts. Figure 4(b) shows the number of stations per latitude belt. In the model run without aerosols there is a consistent overestimate of downward short-wave irradiance at the surface of about 20 W m\(^{-2}\), which peaks in the tropics with overestimates exceeding 30 W m\(^{-2}\). The inclusion of aerosols does reduce this bias by 10-15 W m\(^{-2}\) over most of the mid-latitude and tropical regions; however, a rather large bias remains near the equator. The aerosol climatology has its largest radiative impact over
the relatively dry sub-tropics, at about 30°N and 30°S of the equator: less cloud at these latitudes causes this greater signal.

Analogous plots to Fig. 4(a) (not included) between the models and the ERBE of the all-sky and clear-sky reflected short-wave radiation at the TOA have errors of a much smaller magnitude. Since the clear-sky reflected radiation is dominated by the surface-albedo value, we can infer that the net downwards radiation has a similar error to the downwards radiation error plotted in Fig. 4(a). Furthermore, the net downwards short-wave radiation error at the TOA must be small, since we use a solar-constant value of 1365 W m⁻². Therefore total short-wave atmospheric absorption (the difference between the net downwards at the TOA and surface) is more accurately predicted when the aerosol climatology is included, and the error in the tropics in Fig. 4(a) must be due to an underestimate of atmospheric absorption.

The response of the basic climate variables to the addition of this aerosol climatology was small, as was found by Coakley and Cess (1985) in a similar study. This was partly due to the fixed SSTs used in both model integrations, thus preventing any ocean temperature feedback. Land surface temperatures cooled by as much as 6 K in some areas in the seasonal mean, due to the reduced downwelling short-wave radiation caused by aerosol absorption.

The aerosol climatology has little sensitivity to the parameters of the log-normal size distribution, if optical thicknesses at 0.55 μm for each component are kept constant. Doubling or halving the mode radius, or altering the standard deviation by ±25%, produces little change in the global mean reflected short wave and the surface absorbed short-wave radiation.
6. Discussion

Figures 2(a) and (b) suggest that a bias in the reflected short-wave radiation at the TOA still exists, although there may also be a contribution from unknown biases in the satellite data. Alternatively, model errors could include an underestimation of the mean optical thickness of aerosols and the neglect of the dependence of aerosol optical properties on relative humidity. The latter effect has been explored in an experiment parallel to HadAM3: a parametrization of the relative humidity dependence of the optical properties of water-soluble, oceanic and soot aerosol components was developed, based on the method and data given by D’Almeida et al. (1991). In the experiment, the calculation of aerosol optical properties never allowed the relative humidity to exceed 99%. The mean reflected short-wave irradiances at the TOA increased by 6.57 W m\(^{-2}\) over sea surfaces in a clear sky. This value is, however, an overestimate, because in HadAM3 a clear-sky radiative calculation is done for every grid-box regardless of the presence of clouds, using the values of the variables such as H\(_2\)O densities but ignoring the cloud optical properties. This will tend to give higher relative humidity values than really exist in clear-sky conditions, which could significantly affect the above result.

A regional error of significant magnitude over the tropical Atlantic Ocean is present when the model is compared with both satellite data sets. This is a region of known high dust concentrations, carried from the Sahara Desert by the predominantly easterly winds, and a series of sensitivity experiments was performed with the model to investigate this signal further. The MAR-I profile was modified to include a desert dust aerosol component inserted between the top of the boundary layer and the mid-troposphere. The refractive index of the desert dust component as a function of wavelength used data suggested in WMO (1983); however, the size-distribution parameters used those of the dust component. From the mid-troposphere to the tropopause the aerosol type of the free troposphere had a reduced optical thickness set to 0.015 at 0.55 \(\mu\)m, with the boundary layer and the stratosphere remaining identical to that specified for the MAR-I aerosol type. This new aerosol vertical profile was inserted at all ocean points in a box from 10\(^\circ\) to 30\(^\circ\) W, and 10\(^\circ\) to 30\(^\circ\) N. Three sensitivity tests were performed, with the optical thickness of the desert dust component at 0.55 \(\mu\)m set to 0.1, 0.2 and 0.3, for a one-month integration with July forcing. There was a quite linear increase in the clear-sky reflected short-wave radiation at the TOA over the region for these different desert dust optical thicknesses, increasing by about 3.5 W m\(^{-2}\) for every 0.1 increase in optical thickness. Reference to Fig. 2 shows that a desert dust aerosol with an optical depth of about 0.2–0.4 could therefore explain the TOA irradiance differences over the tropical Atlantic Ocean. Moulin et al. (1997) plot geographical distributions of desert dust optical thickness over the eastern tropical Atlantic Ocean retrieved from Meteosat data, from which they show that mean optical thicknesses over a large region in the northern summer exceeds 0.2. We can therefore conclude that inclusion of the radiative effects of the wind-blown desert dust would reduce significantly the differences over the tropical Atlantic Ocean between the model and satellite measurements.

7. Summary

With no aerosols, the Hadley Centre atmospheric GCM produced differences from ERBE and ScaRaB measurements of the clear-sky reflected short-wave radiation at the TOA over oceans which could not be attributed to observational errors. Comparisons with aircraft measurements indicated that the ocean albedo parametrization could not be the cause, and gaseous absorption and Rayleigh scattering parametrizations in the atmosphere were in very close agreement with a narrow-band model which used modern spectroscopic data. This strongly suggested that the most likely cause of such discrepancies was aerosols.

The inclusion of a simple aerosol climatology based on WMO (1983) recommendations reduced considerably the bias of the new version of the model (HadAM3) relative to the ERBE and ScaRaB data. The aerosol climatology used only two profiles globally, one for land and the other for the oceans and permanently glaciated land. One remaining area of disagreement in clear-sky reflected short-wave radiation between the model results and both the ERBE and ScaRaB data was over the tropical Atlantic Ocean, where the model was about 10 W m\(^{-2}\) too dark: it was found that a quite reasonable representation of desert dust could explain the difference there. This suggests that broad-band measurements such as those from the ERBE and ScaRaB could be used to monitor aerosol distributions over the oceans, although cloud contamination will limit useful results to only the largest aerosol loadings.

The global annual mean estimate of 165 W m\(^{-2}\) for the short-wave radiation absorbed by the surface in HadAM3 contrasts with a corresponding value of about 171 W m\(^{-2}\) when the aerosol climatology is removed from the model (both estimates have had their TOA net radiation adjusted to zero in an approximate fashion). Relative to the best estimate based on GERA data (155±10 W m\(^{-2}\)), the inclusion of the simple aerosol climatology has significantly improved the model. The global, annual, mean atmospheric absorption is increased by 5 W m\(^{-2}\) to 76 W m\(^{-2}\) upon the addition of the aerosol climatology, and is
closer to the estimate of 83 W m\(^{-2}\) by Li et al. (1997). Total atmospheric absorption does still seem to be significantly underestimated over tropical land-surface regions, although it is not clear at this stage whether this is because the simple climatology represents the aerosol loadings poorly in these regions, for example by omitting the effects of biomass burning. Apart from aerosols, there are other factors not yet included in the model which could influence the value of this quantity, such as non-spherical ice crystals and cloud inhomogeneities. Work is proceeding on including such factors in the model.

This work has not attempted to model the full temporal and spatial variation of aerosol optical properties, but has instead shown that aerosols with spatially meaned quantities in general agreement with observations have the potential to alter the surface and TOA radiation budgets in GCMs, and to improve the agreement of these quantities with measurements significantly.

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