Technical Memorandum

Understanding the Local and Global Impacts of Model Physics Changes: An Aerosol Example

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Abstract

This study demonstrates the use a package of diagnostic techniques to understand the local and global responses to a given physics change within a general circulation model. Here, the package is applied to the case of a change in the aerosol climatology in the forecast model of the European Centre for Medium-range Weather Forecasts (ECMWF). The largest difference between old and new climatologies is over the Sahara where, in particular, soil-dust aerosol is reduced. Conventional diagnostics show that the change lead to improvements in local medium-range forecast skill and reductions in seasonal-mean errors throughout the globe. To study the local physics response, short-range tendencies in weather forecasts are diagnosed. These tendencies are decomposed into the contributions from each physical process within the model. The resulting 'initial tendency' budget reveals how the local atmosphere responds to the aerosol change. The net tendencies also provide strong evidence to confirm that the new aerosol climatology is superior. Seasonal integrations demonstrate that the tropic-wide response can be understood in terms of equatorial waves and their enhancement by diabatic processes. The so-called 'Rossby-wave source' is made applicable to general circulation models and used to understand how the tropical anomalies subsequently impact on the global circulation. The mean response in the extratropics is found to be a stationary wave field. Precipitation anomalies that are co-located with extratropical divergent vorticity sources suggest the possibility for diabatic modification of the tropically-forced Rossby-wave response.

1 Introduction

It has long been recognised that localised tropical latent heating anomalies can have an influence on the global circulation. The tropical response to such localised heating has been discussed in terms of dynamical equatorial waves (Matsuno, 1966; Gill, 1980; Heckley and Gill, 1984). Midlatitude responses are often discussed in terms of 'teleconnection patterns' (Horel and Wallace, 1981; Hoskins and Karoly, 1981). A knowledge of global teleconnections is essential for understanding the global climate and is useful for identifying remote 'causes' of observed seasonal-mean anomalies (Hoskins and Sardeshmukh, 1987).

Teleconnection patterns can be well simulated in models by imposing a prescribed (tropical) convective heating anomaly (Webster, 1972; Hoskins and Karoly, 1981; Hoskins and Rodwell, 1995; Greatbatch and Jung, 2007) or, more directly still, by imposing the upper-tropospheric divergence anomaly associated with convective outflow (Sardeshmukh and Hoskins, 1988). These studies have been essential to separate the dynamics of teleconnections from the physical mechanisms involved in their initiation. However, if extended-range predictions, such as of monthly or seasonal-mean anomalies, are to benefit from the existence of teleconnectivity, then the (parametrized) physical mechanisms need to be well represented too.

Here two methodologies are brought together to gain a more complete understanding of the local and global impacts associated with a model physics change. Firstly the 'initial tendency' methodology Rodwell and Palmer (2007), a development of that used by Klinker and Sardeshmukh (1992), is used to understand the local physics within a model. The methodology incorporates data assimilation as used in numerical weather forecasting so that model physics can be tested at atmospheric states as close as possible to true atmospheric conditions. A second methodology is developed from the 'Rossby-wave source' approach of Sardeshmukh and Hoskins (1988). Assuming that the model change predominantly affects tropical physics, this second methodology is used to identify and understand the extratropical stationary wave response.

The methodologies are applied here to a change in aerosol climatology within the physics of the ECMWF model. It is well know that aerosol can have a significant impact on local and global climate (e.g. Miller and Tegen, 1998; Menon et al., 2002; Chou et al., 2005; Yoshioka et al., 2007). The aerosol change investigated in this study is predominantly a change in Saharan soil-dust. Plumes of this dust are occasionally seen over the subtropical north Atlantic (Goudie and Middleton, 2001) and these are of global climatological importance because of their radically different albedo to that of the underling ocean. At shorter timescales, Saharan aerosol has a very

direct impact on local dust storms associated with the Harmattan. However, Rodwell (2005) and Tompkins et al. (2005) have recently demonstrated that this Saharan aerosol can also have an impact on medium-range weather forecasts themselves. This paper applies the above methodologies to investigate the impact of the aerosol change on the local physics and the global dynamics of an atmospheric model in the presence of prescribed (realistic) sea-surface temperatures. At seasonal and longer timescales, interactions between the atmosphere and the ocean begin to modify the response to aerosol (Miller and Tegen, 1998; Miller et al., 2004a; Yoshioka et al., 2007) but these longer timescales and questions such concerning the coupled climate response to aerosol are beyond the scope of the present study.

Hence this paper can be viewed in two ways. It can be viewed as a demonstration of some key diagnostic methodologies that could be applied more widely to general circulation models. In this context, the advance made by the present study is in putting these methodologies together to form a 'seamless' package for model assessment. Alternatively, the paper can be viewed in more specific terms as an estimation of the impact of the (direct) effects of aerosol on local weather and on the global circulation.

In section 2, details are given of the data used and of the salient aspects of the ECMWF model. More specific details are then given on the aerosol change that is being studied here. The two types of model simulations made are then explained. Finally in this section there is a short note on the use of statistical testing. In section 3, the differences seen when the aerosol climatology is modified in the June–August season are documented. The initial tendency methodology is introduced and used to understand the local physics changes in the Saharan region. Equatorial wave theory and the possibility for interaction with diabatic processes is used to explain the tropic-wide response. The Rossby-wave source diagnostic is made applicable to general circulation models and used to understand the global impacts of the aerosol change. A short sub-section then documents the effects on surface fluxes of heat and momentum (in the atmospheric model). Section 4, discusses more briefly a parallel study of the December–February season. Finally conclusions are given in section 5.

2 Model and Data

2.1 Observational data

Upper-air fields for the period 1962–2001 come from the ECMWF 40-year Re-Analysis dataset (ERA-40, Uppala et al., 2005). This dataset is derived using the 3-dimensional variational data assimilation system. The data assimilation process ingests data from almost all available sources. These include top-of-the-atmosphere radiative fluxes at many different wavelengths obtained from satellites as well as radiosonde ascents, drop-sondes and 'SYNOP' station reports.

Precipitation observations for the period 1980–1999 come from Xie and Arkin (1997).

Top-of-the-atmosphere radiative fluxes come from the satellite-based "Clouds and the Earth's Radiant Energy System" (CERES) instrument (Wielicki et al., 1996).

Air-sea fluxes for the period 1980–1993 come from the modified version of the Southampton Oceanography Centre's climatology (Grist and Josey, 2003). In this version, hydrographic ocean heat transport estimates are used to help constrain the surface fluxes over the Atlantic and North Pacific oceans.

2.2 Model description

A detailed description of the ECMWF model can be found at *http://www.ecmwf.int/research/ifsdocs/*. In this section, a brief overview is given of the radiation and convection schemes that were operational in the model versions used in this study. These are the two physical processes most affected by the aerosol climatology change. Note that more recent versions of the model, not used here, include updates to both the radiation and convection schemes.

The radiative heating rate is computed as the vertical divergence of the net radiation flux. Long-wave radiation is computed for 16 spectral intervals using the 'Rapid Radiation Transfer Model' (RRTM, Mlawer et al., 1997). The short-wave radiation part, which is computed for 6 spectral intervals, is a modified version of the scheme developed by Fouquart and Bonnel (1980). Since the computation of the radiative transfer equation is very expensive, the radiation scheme is ordinarily called at 3-hourly intervals and on a lower-resolution grid. Temporal and spatial interpolation are used to get these calculations onto the model grid. In some of the experiments in this study (see below) it has been important to call the radiation scheme at every timestep.

Cumulus convection is parametrized by a bulk mass flux scheme which was originally described by Tiedtke (1989). The scheme considers deep, shallow and mid-level convection. Clouds are represented by a single pair of entraining/detraining plumes which describe updraught and downdraught processes.

The ECMWF model uses spherical harmonics to represent the prognostic fields. These harmonics are (triangularly) truncated at some total wavenumber, M. With the introduction of a two time-level semi-Lagrangian advection scheme in 1998, a linear, rather than quadratic, grid has been used for the calculation of physical tendencies. The triangular resolution is therefore defined as $T_L M$, and this equates approximately to a resolution in degrees of $180^o/M$ (the half wavelength of the shortest resolved zonal wave at the equator).

2.3 Aerosol changes

Until October 2003, the aerosol climatology used in the ECMWF operational forecasting model was based on that of Tanre et al. (1984). This climatology is specified as annual mean geographical distributions of various aerosol types: 'maritime', 'continental', 'urban', 'desert'. Each aerosol type has a fixed vertical profile that decays exponentially with height. The scale height is set at 3 km for all aerosol types except black carbon which has a scale height of 1 km. In addition, uniformly distributed stratospheric background aerosols are included. This aerosol climatology (based on Tanre et al., 1984) will be referred to here as the 'old aerosol'. Figure 1(a) shows the geographical distribution of the total optical depth for the old aerosol at 550 nm (an optical depth of d for a particular wavelength attenuates radiation at that wavelength by a factor e^{-d} as it passes through the atmosphere. This attenuation can be by scattering and absorption). The maximum optical depth (0.74) is seen to occur over the Sahara and this is dominated by desert (*i.e.* soil dust) aerosol.

In October 2003, a new aerosol climatology was implemented in the ECMWF forecast system (at cycle 26R3). This climatology is based on global maps of optical depths for a range of aerosol types compiled by Tegen et al. (1997). The aerosol types included are sea-salt, soil-dust, sulphate, organic carbon and black carbon. (Background stratospheric aerosol was left unchanged). Atmospheric loading for a given aerosol type is deduced from emission/transport modelling studies. For example, soil-dust loading is deduced with eight particle size classes between 0.01 and 10 μm. Uplift from the surface is a function of near-surface windspeed, soil moisture and soil texture. The smallest particles have an atmospheric lifetime of around 230 hours in the transport model, with rain-out being the primary deposition mechanism. The larger particles have a shorter lifetime of around 30 hours and deposition is primarily by gravitational settlement. The differing lifetimes lead to a particle size distribution in the atmosphere. The optical depth at a given wavelength is a function of atmospheric loading,
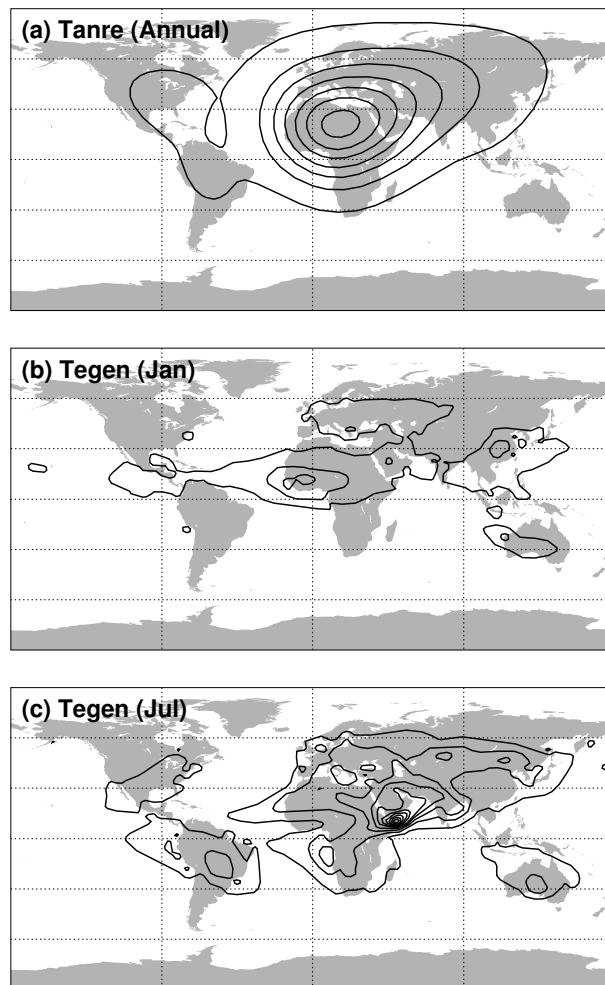
Figure 1: Optical depths at 550 nm associated with the model aerosol climatology. (a) The 'old' annually-fixed climatology of Tanre et al. (1984). (b) The 'new' January climatology of Tegen et al. (1997). (c) The 'new' July climatology of Tegen et al. (1997). The smallest contour is 0.1 and the contour interval is 0.1.

particle effective radius and refractive index. This climatology (based on Tegen et al., 1997) will be referred to here as the 'new aerosol'. In this paper, where a difference between two fields, forecasts or simulations is shown, it is in the sense of new aerosol minus old aerosol. This difference will be referred to as the 'anomaly' or 'anomalous field'.

For the new aerosol climatology in July (Figure 1(c)) the region of maximum total optical depth (maximising at a value of 1.05) is now located over the Horn of Africa and out into the Arabian Sea associated with the transport of dust by the monsoonal Somali Jet. The aerosol optical depth over the Sahara is more than halved. The January aerosol of the new climatology (Figure 1(b)) also shows major differences with the old annual-mean climatology (Figure 1(a)). The magnitude of these changes is comparable with the uncertainties in mineral dust loadings summarised by Zender et al. (2004).

For the short-wave, in addition to being able to scatter radiation, some aerosol types such as soil-dust and black carbon can also absorb. By absorbing short-wave radiation, these aerosols can have a very direct impact on atmospheric temperatures. A measure of the relative strength of absorption is given by the 'single scattering albedo'. This is the ratio of scattering efficiency to total light extinction (scattering plus absorption). The single scattering albedo and other aerosol optical properties used within the ECMWF forecasting system are

calculated following Hess et al. (1998). For desert aerosol, the single scattering albedo is around 0.888. For clean maritime air it is around 0.997. Since the differences between the new and old aerosol are particularly associated with desert aerosol, it is possible that absorption as well as scattering will be an important mechanism in the response. For the long-wave, there is no explicit representation of scattering by aerosol in the model. Absorption and emission of long-wave radiation is calculated over all long-wave spectral intervals.

The exponential vertical profiles of particle distribution are left the same in the new aerosol climatology as in the old climatology.

In the ECMWF model, the aerosol concentration does not impact the cloud microphysics. Hence indirect aerosol effects such as how larger numbers of cloud condensation nuclei can lead to more, smaller and longer-lived cloud droplets and thus changes in the radiation budget are not represented. Instead, the local and global impacts of the change in aerosol climatology discussed here must arise purely from the direct effect and the "semi-direct" effect. (The term "semi-direct" is used to describe the mechanism whereby radiation absorption leads to warming and less cloud, Hansen et al., 1997).

The last decade has seen further advances in aerosol estimation and, in this respect, the 'new' aerosol climatology cannot be considered as state-of-the-art. However, Tegen et al. (1997) show that comparisons with ground-based sun photometer measurements are reasonable and the 'new' aerosol climatology remains in ECMWF's operational forecast model. In this respect, it is clearly worth understanding the effects of this aerosol climatology within the global circulation.

2.4 Seasonal integrations

To assess the impact of the change in aerosol climatology on the atmospheric model's own climate, sets of seasonal integrations have been made for 40 December–February and June–August seasons for the period 1962 to 2001. The initial conditions for these integrations are based on 1 April and 1 October analyses from the ERA-40 dataset, respectively (the first two months of each forecast were discarded). Sea-surface temperatures and sea-ice cover are also taken from ERA-40. These are based on monthly-mean values from the HadISST dataset (Rayner et al., 2003) up to November 1981 and weekly-mean values from the NOAA/NCEP 2D-Var dataset (Reynolds et al., 2002) thereafter. For each season, a set of integrations is made with old and with new aerosol. The integrations use model cycle 26R3 and are run at T_L95 ($\approx 1.9^o$) horizontal resolution with 60 levels in the vertical and a timestep of 1 hour. In these seasonal integrations, the radiation scheme is called every three hours with computations made on a T_L95 linear grid.

It should be emphasised that the seasonal-mean climates and climate anomalies that will be shown are those from an atmospheric model in the presence of prescribed, realistic sea-surface temperatures. This approach allows a good investigation of the salient physics and dynamics of the atmospheric response to aerosol. A good representation of this atmospheric response is a pre-requisite for good atmosphere-ocean coupled simulations.

2.5 Weather forecasts

It is difficult to isolate the direct effect of a particular model change in seasonal or climate simulations because this direct effect will be obscured by interactions and feedbacks with the circulation. The use of weather forecasts can greatly help in this regard because these are initiated from atmospheric states where the circulation is much closer to a real state of the atmosphere. In this study, 10-day weather forecasts are started every 6 hours from 0 UTC on 26 June 2004 to 18 UTC on 26 July 2004 (31 days × 4 times per day). The day 5 verification times of these experiments exactly comprise July 2004.

The weather forecasts are initialised-with, and verified-against, analyses produced by ECMWF's 4-dimensional variational data assimilation system (4DVAR, Rabier et al., 2000). 4DVAR starts with a 'first guess' from a previous model forecast and essentially involves iteratively nudging the non-linear and tangent-linear versions of the model to the new observations. Hence the analysis can be quite strongly dependent on the model used within the data assimilation. Since forecast errors and tendencies will be diagnosed at very short lead-times, a fair comparison of aerosol climatologies requires that two sets of analyses are produced: one set for the model incorporating each aerosol climatology. This experimental design allows the same 'initial tendency' analysis (see below) to be conducted as in Rodwell and Palmer (2007). These experiments are also used to assess the impact on conventional measures of medium-range forecast skill.

These weather forecast integrations use model cycle 29R1 and are run at $T_L 159$ ($\approx 1.1^o$) horizontal resolution with 60 levels in the vertical and a timestep of $\frac{1}{2}$ hour. In these forecasts, the radiation scheme is called every timestep with computations carried-out on a $T_L 63$ linear grid.

2.6 Statistical testing

Where a statistical test is mentioned in this paper, this refers to a two-sided Student's t-test of the difference of means. In every case shown here both distributions are based on the same set of dates and times and so a more powerful paired t-test is performed. Since autocorrelation could reduce the effective number of degrees of freedom in a timeseries, this is taken into account by using an auto-regressive model of order one (von Storch and Zwiers, 2001). A 'dual colour palette' has been developed in this study to aid the display of statistical significance. This approach is used in some of the figures; with bolder colours indicating significant anomalies and the more pale colours indicating non-significance. Where confidence intervals are shown in plots, they are also based on the Student's t-distribution function and autocorrelation is taken into account in the same way. In general, an $x\%$ significance level can be thought of as a $(100 - x)\%$ confidence level.

3 The June–August season

This section first documents the climate and forecast skill differences seen when the aerosol climatology is changed. Later it is shown how the local physics can be understood by examining initial tendencies in weather forecasts. The tropical response is discussed in terms of physical coupling with equatorial waves and finally the vorticity forcing by tropical divergent flow anomalies is used to understand the extratropical response.

3.1 Documenting the differences

3.1.1 Seasonal-mean differences

Figure 2(a) shows mean June–August precipitation (shaded), low-level (925 hPa) wind vectors and 500 hPa geopotential height contours from the observational data. The summer monsoons of southern Asia, North Africa and Mexico, together with their associated low-level inflows are clearly evident. In the winter (southern) extratropics, a strong westerly jet is evident from the tightness of the geopotential height contours. In the summer (northern) hemisphere, the jet is weaker.

Figure 2(b) indicates the statistically significant mean errors for the model with the old aerosol. These errors include too much precipitation on the northern flank of the north African monsoon, strong wind biases over the sub-tropical north Atlantic and extratropical circulation biases to the south and southwest of South Africa. The
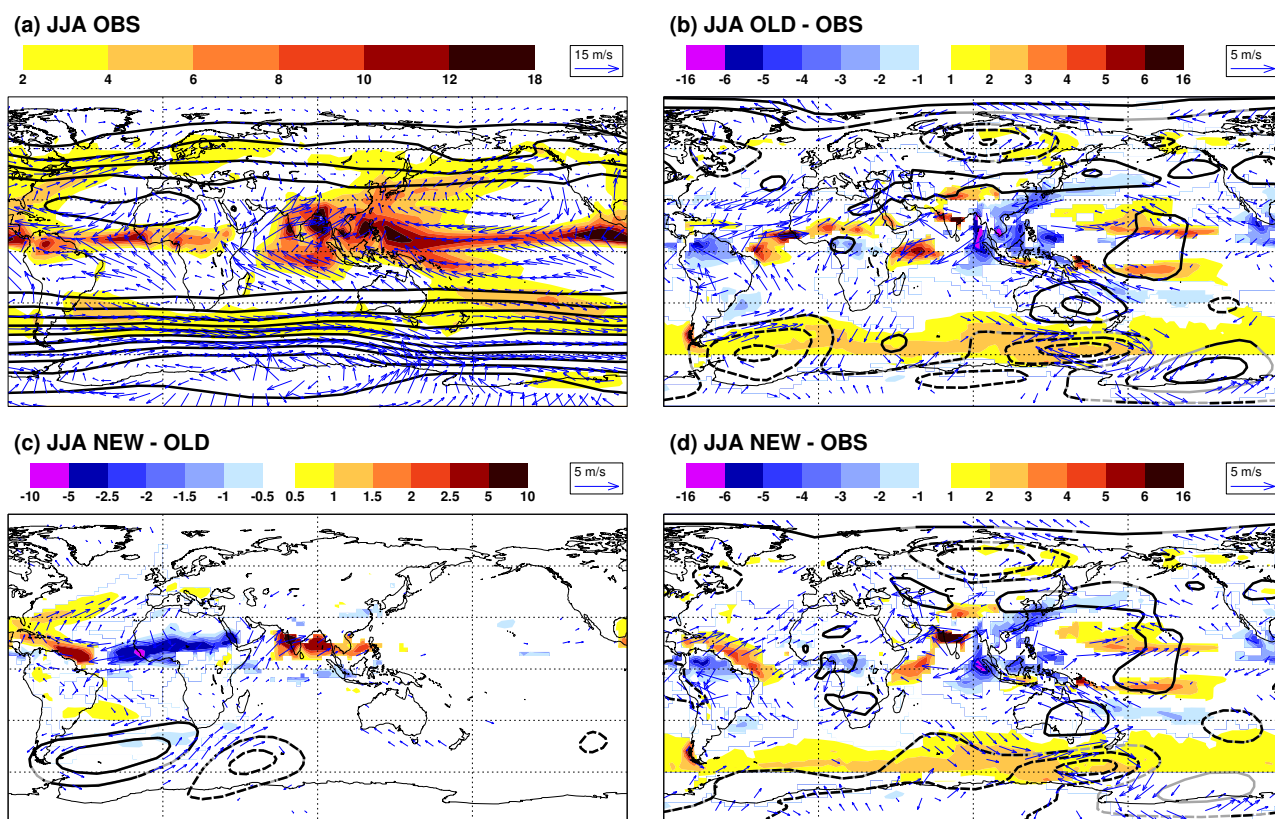
(a) JJA OBS

(b) JJA OLD - OBS

(c) JJA NEW - OLD

(d) JJA NEW - OBS

Figure 2: Diagnostics of June–August seasonal-mean total precipitation (shaded in mm day^{-1}), 925 hPa horizontal wind vectors (see scaling vector) and 500 hPa geopotential heights (see below for contour interval). Observational data come from Xie and Arkin (1997) for precipitation over the period 1980–1999 and from ERA-40 for the other fields over the period 1962–2001. Forecast data come from the seasonal integrations covering the same period as for the observations (see main text for details). (a) Mean observed. (b) Mean model error with the 'old' aerosol. (c) Mean difference: model with 'new' aerosol minus model with 'old' aerosol. (d) Mean model error with the 'new' aerosol. Precipitation and wind differences are only plotted where seasonal-mean differences are statistically significant at the 10% level. Height differences are contoured solid for positive, dashed for negative, grey where not significant and with contour interval of 10 dam in (a) and 2 dam in (b)–(d).

effect of the change in aerosol (Figure 2(c); note the change in shading interval for precipitation) is a reduction in these particular mean errors so that they are no longer apparent in the mean errors with the new aerosol (Figure 2(d)). Elsewhere, mean errors are largely unchanged. The main degradation is perhaps the increased mean error in precipitation off the north-east coast of South America.

The north African monsoon will be shown to be a key component in this global improvement. Miller et al. (2004b) show a similar response of north African monsoon precipitation to aerosol forcing, particularly when particles are made more absorbing. Yoshioka et al. (2007), who assume weaker absorption, produce a response of the opposite sign. It would appear that the response can be sensitive to the value of the single scattering albedo. At this point in the discussion, it is not possible to say whether the improvement in north African monsoon precipitation occurs for the right or wrong reasons. However, the 'initial tendency' discussion later will suggest that the improvement is, in fact, for the right reasons.

Assessments of the impact of aerosols often focus on top-of-the-atmosphere 'forcing' (e.g. Miller and Tegen, 1998). Such forcing is clearly a useful quantity to consider when investigating global climate change. Comparison (not shown) between the seasonal integrations and 10 months of CERES data (Wielicki et al., 1996) reveal

that, over the Sahara desert region, there is a reduction in net incoming short-wave radiation error (typically from +30Wm^{-2} to +20Wm^{-2}) and also a reduction in out-going long-wave radiation error (typically from -10Wm^{-2} to near 0Wm^{-2}). Hence, overall, the error in net radiative heating of the atmosphere is reduced (typically from +40Wm^{-2} to +20Wm^{-2})). Over the north African monsoon region, incoming short-wave errors are also reduced (typically from -15Wm^{-2} to -5Wm^{-2}). Out-going long-wave radiation errors, however, are increased in the monsoon region (typically from +20Wm^{-2} to +30Wm^{-2}). Overall the net atmospheric heating error in the monsoon region is unaffected by the aerosol change except along the borders of the Gulf of Guinea. It should be noted that the simulation of out-going long-wave radiation is more problematic in cloudy regions and so the increase in error in the monsoon region should not, on its own, be interpreted as reflecting underlying problems with the new aerosol. To emphasise this point further, the typical top-of-the-atmosphere 'forcing' by the aerosol change is an atmospheric cooling of around 20Wm^{-2} whereas the atmospheric cooling associated with the reduction in convective latent heating within the monsoon region (Figure 2(c)) is 4 times larger at around 80Wm^{-2}. This suggests the existence of positive feedbacks within the atmospheric response to aerosol. The apparent error in the out-going long-wave radiation within the monsoon region could easily be associated with errors in these feedbacks.

While seasonal-mean diagnostics can indicate changes and hopefully improvements in model climate, it is difficult to obtain a good understanding of how these changes come about. This is particularly the case when feedbacks are involved. Later the physics of the local response to the aerosol change will be examined by using very short-range forecasts. Next, however, the changes in weather forecast skill will be documented.

3.1.2   *Weather prediction differences*

It has been shown by Tompkins et al. (2005) that the use of the new aerosol climatology leads to a better representation of the *mean* African Easterly Jet in 5-day forecasts with the ECMWF model. Here the focus is on how the use of the new aerosol climatology affects the predictability of *variations* about the mean in the North African summer monsoon region. Such variations include fluctuations associated with the African Easterly Jet, including so-called African Easterly waves.

Here, the spatial 'anomaly correlation coefficient' is used to quantify the predictive skill of the weather forecasts. The anomaly correlation coefficient is defined as follows. Let the field $\mathbf{f}_t$ be the forecast for time t for a given parameter, region and forecast lead-time. Let the field $\mathbf{a}_t$ be the verifying analysis and let $\bar{\mathbf{f}}$ and $\bar{\mathbf{a}}$ be the temporal mean fields, averaged over the whole month of forecasts. Remove these mean fields (the part assessed by Tompkins et al. (2005) by writing $\mathbf{f}'_t = \mathbf{f}_t - \bar{\mathbf{f}}$ and $\mathbf{a}'_t = \mathbf{a}_t - \bar{\mathbf{a}}$. Next remove the area mean (signified by $\langle \cdot \rangle$) by writing $\mathbf{f}''_t = \mathbf{f}'_t - \langle \mathbf{f}'_t \rangle$ and $\mathbf{a}''_t = \mathbf{a}'_t - \langle \mathbf{a}'_t \rangle$. Then the spatial anomaly correlation coefficient is defined here as:

$$\rho_t = \frac{\langle \mathbf{f}''_t \mathbf{a}''_t \rangle}{\sqrt{\langle \mathbf{f}''^2_t \rangle \langle \mathbf{a}''^2_t \rangle}} \quad . \tag{1}$$

Figure 3(a) shows the seasonal-mean of the spatial anomaly correlation coefficient as a function of forecast lead-time for 700hPa meridional wind field in the African Easterly Jet region [15°W–35°E, 5°N–20°N]. Dots indicate statistically significant improvements, at the 5% significance level, when the new aerosol is used. Indeed, predictability in the medium-range is increased by about 1 day for each lead-time $\geqslant$ 4 days. Similar improvements are found for the eastern tropical Atlantic region [40°W–15°W, 5°N–20°N] (Figure 3b) and for the zonal component of the wind in both regions (not shown). Since the aerosol itself is fixed for any given month, these improvements in the prediction of variability must be associated with improvements in simulating the mean state. Further-a-field, the skill in forecasting variations about the mean is unaffected by the aerosol change (not shown).
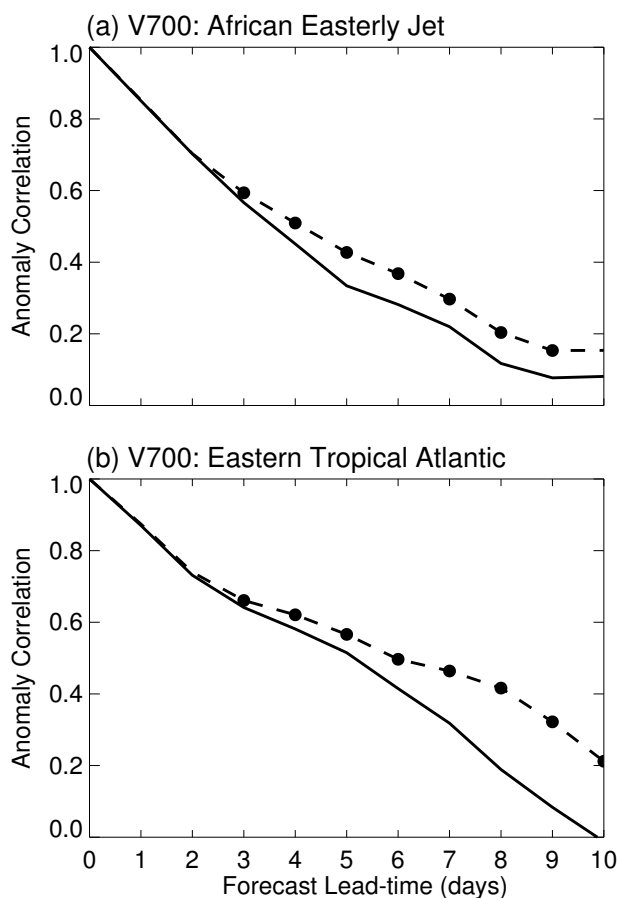
Figure 3: Average anomaly correlation coefficients (see main text for details) for forecasts of meridional wind variations at 700 hPa with the 'old' (solid) and the 'new' (dashed) aerosol climatology for (a) the African easterly jet region (15^oW–35^oE, 5^oN–20^oN) and (b) the eastern tropical Atlantic (40^oW–15^oW, 5^oN–20^oN). Forecast lead-times for which the score with the 'new' aerosol is significantly better (at the 5% level) are marked with filled circles. Results are based on the weather forecasts (see main text for details) started at 12 UTC on each day between 26 June to 26 July 2004.

3.2 Understanding the differences

It is of interest to understand how the improvements documented above are brought about in response to the aerosol change. The initial focus is on the local physics (in the North African region) and how this physics responds over the first few forecast timesteps to the change in aerosol. Later, the wider-scale tropical and extratropical responses are investigated.

3.2.1 Local physics: the 'Initial Tendency' methodology

The 'Initial Tendency' methodology represents a different approach to model assessment compared to more 'conventional' techniques involving, for example, comparison with observed top-of-the-atmosphere and surface fluxes. Hence, while there is nothing particularly difficult in 'initial tendencies', an understanding of the methodology can require a moment or two of reflection. Below, the methodology is motivated, explained and used in the context of the aerosol change. The dual aim of this section is to demonstrate the power of the initial tendency approach for assessing almost any model physics change.
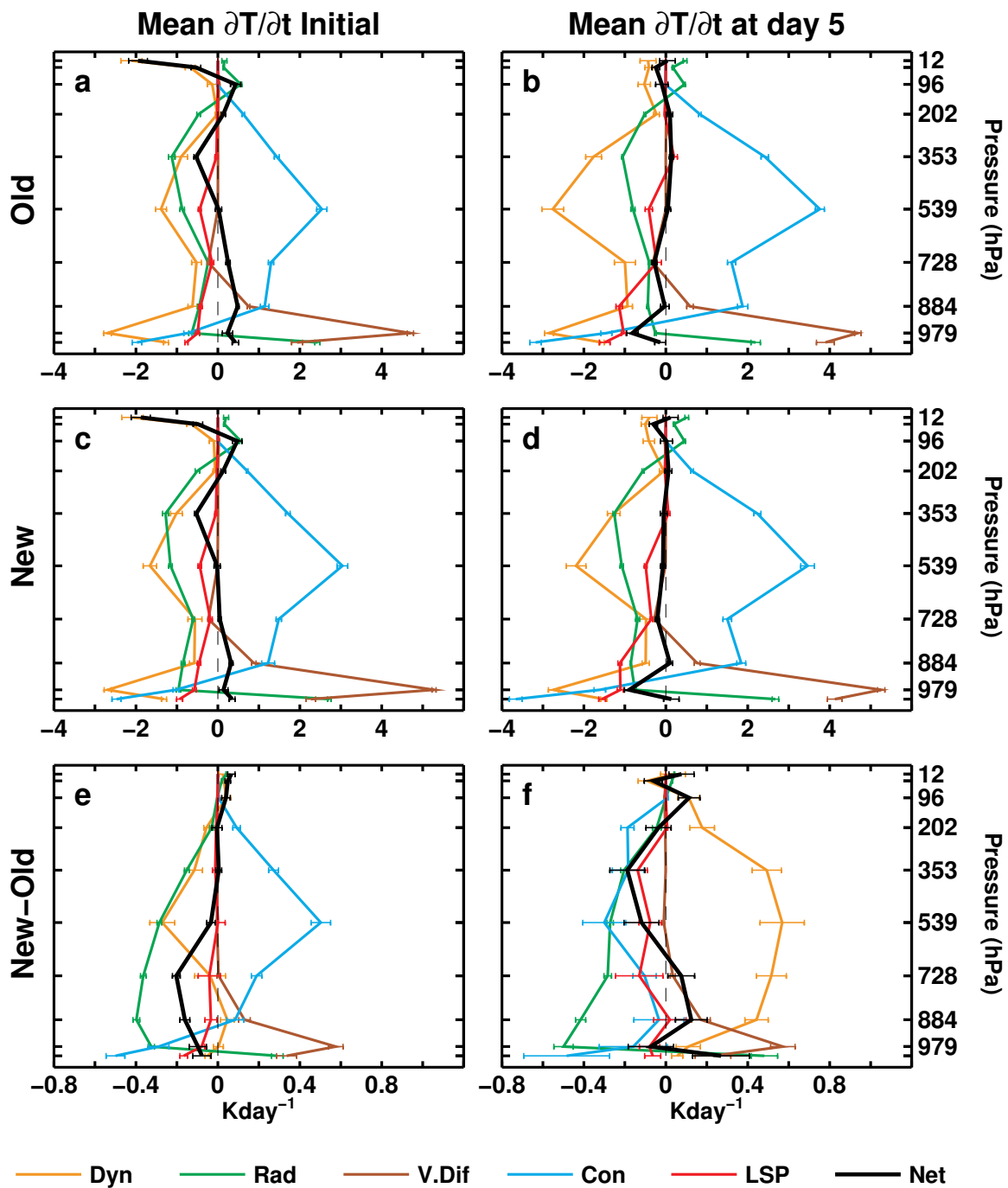
Figure 4: *Vertical profiles of 'initial tendencies' (left) and day 5 tendencies (right) of temperature averaged over the north African monsoon region. Tendencies from the dominant physical processes and the dynamics are shown, together with the net tendency (i.e. the total tendency over all processes). Results are based on 124 weather forecasts started every 6 hours from 26 June to 26 July 2004 (=31 days x 4 forecasts per day; see main text for details). (a) Initial tendencies with the 'old' aerosol (accumulated over the forecast lead-time 0.5 to 6.5 hours). (b) As (a) but for tendencies at day 5 (accumulated over the forecast lead-time 120.5 to 126.5 hours). (c,d) As (a,b) but for the 'new' aerosol. (e,f) As (a,b) but for the difference in tendencies: 'new' minus old aerosol. Tendencies are calculated on every 5th model level. The vertical axis shows approximate model level pressure and is linear in pressure. The indicated 70% confidence intervals are based on the 31 diurnally accumulated tendencies deduced by summing the tendencies from the 4 forecasts made each day. The north African region (5ºN–15ºN, 20ºW–40ºE) corresponds to the region of decreased precipitation found in the seasonal integrations.*

The initial tendency approach involves the use of the weather forecast experiments discussed in section 2.5. To recapitulate, there are 124 forecasts for each aerosol climatology (31 days × 4 times per day) and the analysis used to initiate a forecast with a particular aerosol is produced by the data assimilation system incorporating that same model aerosol.

Each physical and dynamical process within a forecast model contributes to a timestep increment. In this analysis, each process' contribution is accumulated over the first six hours of the four forecasts made each day. (Actually from lead-time $\frac{1}{2}$ hour to lead-time $6\frac{1}{2}$ hours. This is very different from Klinker and Sardeshmukh (1992) who examined just the very first forecast timestep. See Rodwell and Palmer (2007) for more discussion of these differences). In this way the diurnally-accumulated increment or tendency from each process is produced. These tendencies are then averaged over a chosen region (here the chosen region, 5°N–15°N, 20°W–40°E, encompasses the area of reduced June–August north African monsoon rainfall in the seasonal integrations; Figure 2c). Finally, these diurnally-accumulated, area-averaged tendencies are averaged over the 31 days' of forecasts. The resulting mean tendencies are referred to here as 'initial tendencies' as they reflect mean tendencies over the first six hours of a forecast. As will be seen below, initial tendencies are highly useful because they reflect mean tendencies within the model when it is initiated from a set of atmospheric states as close as possible to the true atmospheric states. (This true atmospheric state includes realistic boundary values such as sea-surface temperatures).

Figure 4(a) shows, for the old aerosol, vertical profiles of the initial tendencies of temperature from the dominant local processes. Before discussing these profiles it is worth briefly considering what one might anticipate in these profiles. The concept of 'radiative-convective equilibrium' embodies the idea that radiative processes act to destabilise the atmosphere (heat the surface and cool the mid-to-upper troposphere) and the convection induced by this destabilisation acts to restore balance by cooling the surface and heating the mid-to-upper troposphere. To some extent, Figure 4(a) does show this cancellation between convective (Con) and radiative (Rad) tendencies. However, other terms are also important in the overall balance. In particular, dynamical cooling associated with ascent (Dyn) and evaporative cooling of large-scale precipitation (LSP) act with the radiation to balance the convective tendency. In the lower troposphere, the vertical diffusion term (V.Dif), which includes the surface sensible heat flux, combines with the radiative tendency (Rad) to destabilise the vertical profile. (Because surface temperatures are strongly dependent on the surface radiation budget, this destabilisation by sensible heating can, in fact, be considered as an indirect destabilisation by the radiation process). The 70% confidence intervals shown in Figure 4(a) confirm that the 31-day mean tendencies are robust and not sensitive to the sampling of shorter timescale (*e.g.* synoptic) variability.

Net initial tendencies (*i.e.* the sum of initial tendencies from all individual processes) is also displayed in Figure 4(a) (Net). It is important to note that, in the real world, net tendencies should be almost zero when accumulated over the diurnal cycle and averaged over sufficient synoptic variability. This is because the net tendency (*e.g.* of temperature) in the real world is simply the temperature at the end of the period (*i.e.* month) minus the temperature at the beginning of the period, divided by the length of the period. Typically, for reasonably large area-means in the tropics and subtropics and for solstitial months, the net temperature tendency is $\sim$0.03 Kday^{-1}. Hence the fact that Figure 4(a) shows non-zero net initial tendencies in the forecast is indicative of physics errors in the model with the old aerosol (but see below for more discussion). Net initial tendencies are the seeds of what is termed 'model drift' in the climate community. As defined here, these tendencies are virtually equivalent to minus what is termed the 'analysis increment' in the data assimilation community.

To be more precise, to be able to assert that net initial tendencies are indicative of errors in model physics, one needs to assume that the climate of the initial state (*i.e.* the climate of the analysis) is not biased, or at least less biased than the climate of the model. This is a reasonable assumption to make since the observations that are used while forming the analysis are a sampling of the real world. As long as the observations are sufficiently unbiased (they can still be imperfect), so will be the climate of the analysis.

The net initial tendencies in Figure 4(a) (Net) show an erroneous destabilisation of the atmospheric column in the model with the old aerosol. There is erroneous warming of the lower troposphere (*e.g.* near 884 hPa) and erroneous cooling of the mid-to-upper troposphere (*e.g.* near 353 hPa). Referring back to the radiative-convective equilibrium discussion above, one might ask if the radiation scheme is destabilising too much or if the convection scheme is not stabilising enough? Of course, at this stage in the investigation the problem could be with one or more of the other processes within the atmosphere.

Figure 4(b) shows corresponding tendencies deduced at day 5 of the forecasts (accumulated from lead-time $120\frac{1}{2}$ hours to lead-time $126\frac{1}{2}$ hours). It can be seen that the convective heating (Con) and dynamical cooling associated with ascent (Dyn) have strongly increased relative to Figure 4(a). In contrast, the net temperature tendency at day 5 (Net) is smaller than it was in the initial tendencies (Figure 4a). This indicates that, in the mean, the forecast model has moved closer to its own preferred equilibrium position by day 5. Consistent with the seasonal integration results (Figure 2b), this equilibrium occurs at a state with an over-active north African monsoon.

Figure 4(c) shows the initial temperature tendencies for the forecasts with the new aerosol. Interestingly, the initial convective heating with the new aerosol (Con) is *stronger* than with the old aerosol (Figure 4a). At first sight, this is in apparent contradiction to the fact that overall seasonal-mean precipitation is less in the seasonal integrations with the new aerosol compared to those with the old aerosol (Figure 2c). Importantly, the net initial tendencies in Figure 4(c) (Net) show less erroneous destabilisation of the vertical column with, in particular, less erroneous lower-tropospheric warming than was the case for the old aerosol (Figure 4a).

By day 5 for the new aerosol (Figure 4d), the convective heating and dynamical cooling have again increased but the increase is less than with the old aerosol (Figure 4b). This decrease in convection by day 5 relative to the old aerosol resolves the apparent contradiction highlighted above. Clearly the change in seasonal-mean precipitation involves not just the changes in initial tendencies but also subsequent feedbacks with the large-scale circulation (as represented by Dyn).

Figure 4(e) shows anomalies (new minus old aerosol) in initial tendencies. Note the change in horizontal axis. The strengthened initial convection (Con) with the new aerosol is clearly evident. However, since indirect aerosol effects are not included in the model, the cloud (included in LSP) and the convection (Con) processes cannot be the root-cause of the reduced erroneous net initial tendencies. The root-cause must be associated with the interaction between radiation and aerosol.

The anomalous initial radiative tendencies in Figure 4(e) (Rad) show a (direct) cooling over much of the troposphere with increased warming near the surface. The vertical diffusion (V.Dif) also shows increased warming in the lower-troposphere. (It was mentioned above that the vertical diffusion scheme incorporates the sensible heat flux and this is itself strongly constrained by the surface radiation budget). The anomalous radiation and vertical diffusion tendencies do not change greatly by day 5 (Figure 4f) whereas the tendencies for the other dominant processes (Con, Dyn, LSP) do change as they respond to the anomalous aerosol radiative forcing.

There appear, therefore, to be two mechanisms by which the atmosphere responds to the direct radiative effect of the aerosol change:

(1) There appears to be a fast mechanism that is associated with the thermal destabilisation of the vertical profile. This would appear to lead to the initial enhancement of convection and allow the convection scheme to operate at lower specific humidity levels (the lower-tropospheric moist bias is reduced from 1gkg^{-1} with the old aerosol to 0.5gkg^{-1} with the new aerosol, not shown).

(2) There also appears to be a slower mechanism involving interactions with the large-scale dynamics. The aerosol in this region is primarily soil-dust and it was noted above that soil-dust is able to absorb as well as scatter radiation. The reduction in aerosol therefore leads to the direct cooling of the lower troposphere. This

cooling reduces the erroneous net warming tendency seen with the old aerosol (Figure 4a) and thus acts to reduce the erroneous strengthening of ascent (noted above). With less ascent, there is less large-scale convergence and thus the dynamics with the new aerosol provides less moisture to the profile (not shown). This slower mechanism appears to provide a limit to the strength of the convection so that by day 5, the convection is actually less with the new aerosol than it is with the old aerosol.

If these are the two primary response mechanisms then it may be important that the convection deposits its latent heat above the main aerosol layer. Otherwise the increased initial convection could counter-act the direct radiative cooling response to the aerosol change and thus prevent the slow dynamical feedback. Elements of the semi-direct response to aerosol (whereby radiation absorption by aerosol prevents condensation) could be important to ensure that the latent heating is above the main aerosol layer (regardless of which aerosol climatology is used).

The advantage of the initial tendency approach is that it allows complicated response mechanisms to be diagnosed before interactions can render the mechanisms impenetrable. Without the initial tendencies one would be forgiven for thinking that the direct effect of the reduced aerosol was reduced monsoon convection. In fact, the initial effect appears to be increased convection. It is the reduction in the erroneous feedback with the dynamics (Figure 4f, Dyn) that appears to lead to the eventual reduction in erroneous precipitation seen in the seasonal integrations (Figure 2c). Clearly the relative importance of these (and other) mechanisms may be model-specific. In addition, at seasonal and longer timescales, feedbacks with sea-surface temperatures could again modify the monsoon's response to the aerosol (Miller and Tegen, 1998).

Non-zero net initial tendencies imply errors in model 'fast physics' (*i.e.* physics that has an appreciable effect on numerical weather prediction). In this study the model with the old aerosol has clear physics errors in the lower troposphere over the north African monsoon region (Figure 4a, Net). (The model with either aerosol climatology appears to have physics errors in the upper troposphere; Figure 4a and c, Net). To make the converse inference; that smaller net initial tendencies imply smaller errors in model fast physics requires some further assumptions to be made.

Firstly, one has to assume that two or more errors cannot cancel each other to leave small net tendencies. Rodwell and Palmer (2007) did assert that cancellation was very unlikely because, over their global domain, the state-space of net initial tendencies has a very high dimension. An upper-bound on this dimension is the total number of model grid-points, levels and prognostic variables. Even if the horizontal dimension is reduced to take account of communication by the fastest internal gravity waves over the analysis window (here 6 hours) or simply reduced to the number of distinct climatic regions of the globe (taking into account the phase of the annual cycle), the dimension still remains very large. It would be very unlikely that two distinct model errors could cancel each other in such a large dimensional space. In the present study, the focus has been on a single climatic region and a single prognostic variable and so cancellation may not be quite so unlikely. However, while the discussion above has concentrated on the net initial tendencies of temperature, it is also the case that the net initial tendencies of moisture and momentum support the same conclusions.

Secondly, observations are rejected by the data assimilation system if they appear to be 'too far' from the first guess forecast. For an extremely bad model (and thus an extremely bad first guess forecast), many observations could be rejected. This would leave the analysis close to the first guess and the net initial tendencies would, therefore, be small. However, this is not the case in the present study as the number of tropospheric observations accepted by the data assimilation system in the north African monsoon region was actually 2% more for the model with the new aerosol climatology than for the model with the old aerosol climatology.

In summary, the smaller net initial tendencies over the north African monsoon region with the new aerosol are strong evidence that the new aerosol climatology is superior to the old aerosol climatology. The physical reasoning based on the initial tendencies is additional evidence. Other evidence has been given including

improvements in seasonal-mean biases and weather forecast skill.

In terms of the present study's discussion of methodologies for model assessment, it is clear that the initial tendency approach provides a very powerful way of assessing model errors and model changes. Unlike top-of-the-atmosphere fluxes for example, they enable a 3-dimensional examination of the processes involved. They ensure that the observations are used in a consistent manner for both forecast initiation and forecast error assessment. Importantly, they allow the assessment to be made at atmospheric states close to reality, before many interactions and feedbacks have had time to take place. The term 'close to reality' refers to the analysis, which is derived from the data assimilation system. Since the analysis can be different for each model under investigation, the initial tendency approach also offers something more than can be obtained from single column models forced with fixed boundary conditions. Referring back to the title of this section, initial tendencies allow a very *local* assessment of model physics. For example, the aerosol changes are predominantly over the north African region but there are differences further-a-field. These differences further-a-field (compare Figure 1a and c) will have impacts on the climate of long model simulations but will have minimal impact on the initial tendencies over north Africa.

3.2.2   *Tropical impacts: equatorial waves and diabatic interaction*

It has been shown that the direct radiative cooling effect of the aerosol change, and the much stronger negative precipitation response that it triggers, lead to strongly reduced diabatic heating in the June–August season within the north African monsoon region. Shallow water equation studies on the linearised β-plane (Matsuno, 1966; Gill, 1980) demonstrate that such heating anomalies can force equatorial waves. From this theory, one would anticipate that the weakening of the north African monsoon would force non-dispersive, eastward-propagating Kelvin wave anomalies. In the seasonal-mean, a signature of these waves would be anomalous upwelling over the Indian Ocean. The substantially increased rainfall seen in Figure 2(c) over the northern Indian Ocean / Asian monsoon region (5mmday^{-1} over the west coast of India) is consistent with a triggering of convection by these Kelvin waves. If true, this again highlights how the diabatic physics is able to enhance the dynamical forcing. Note that one can probably discount the possibility that the local increase in aerosol over the Arabian Sea and Bay of Bengal (compare Figure 1a and c) leads to this increased convection since studies show that aerosol increases over the ocean actually lead to a local *decrease* in precipitation (Miller and Tegen, 1998).

Shallow water theory also tells us that the cooling anomaly within the north African monsoon region and the heating anomaly over the northern Indian Ocean will force equatorial Rossby waves. The strengthened cross-equatorial and southwesterly low-level flow over the Arabian Sea in Figure 2(c), for example, is associated with the equatorial Rossby-wave response to the Asian monsoon heating anomaly (as in Rodwell and Hoskins, 1995). Similarly, the westerly low-level wind anomaly over the tropical Atlantic in Figure 2(c) is consistent with the reduced north African monsoon heating (Rodwell and Hoskins, 2001).

3.2.3   *Extratropical impacts: Rossby-wave forcing*

Using the initial tendency analysis, it has been possible to explain the June–August local physical response to the change in aerosol. The tropic-wide response has been discussed in terms of equatorial wave theory and the likely enhancement by the diabatic physics. One feature in Figure 2(c) remains to be examined. This is the June–August Southern Hemisphere extratropical response which appears as an equivalent barotropic anticyclone–cyclone pair centred to the south of South Africa, with strong southwesterly winds in-between. At 500hPa, this extratropical feature appears disconnected from the tropical changes further north. Two-layer shallow water theory, which was used above to interpret tropical, internal, baroclinic waves, is not well suited

to explaining this extratropical, external, equivalent-barotropic response. Instead, it is well known that Rossby-wave dynamics in the upper-troposphere provide the tropical-extratropical link for such a response. Here, Rossby-wave solutions to the vertical vorticity equation are considered. This equation can be written as:

$$\frac{\partial \zeta}{\partial t} + \boldsymbol{\nabla} \cdot (\mathbf{v}\zeta) = 0 \quad , \tag{2}$$

where ζ is the vertical component of absolute vorticity, t is time and $\mathbf{v}$ is the horizontal wind. In equation (2), the shallow atmosphere approximation has been made and vertical advection, tilting, friction and the solenoidal term have been neglected on scaling arguments for midlatitude synoptic systems. Traditionally (see, e.g. Sardeshmukh and Hoskins, 1988) the wind field is separated into divergent and rotational components, $\mathbf{v} = \mathbf{v}_\chi + \mathbf{v}_\psi$ (where $\mathbf{v}_\chi = \boldsymbol{\nabla}\chi$, the wind component parallel to the gradient of the velocity potential, χ, and $\mathbf{v}_\psi = \mathbf{k} \times \boldsymbol{\nabla}\psi$, the wind component parallel to the streamfunction, ψ, and $\mathbf{k}$ is the local unit vertical vector)

The components of this equation that are dependent on the divergent flow, $\mathbf{v}_\chi$, are moved to the right-hand side of the vorticity equation and regarded as a forcing from the tropics (associated for example with convective out-flow changes forced by sea-surface temperature anomalies or, as in this study, with aerosol changes). The remaining components, which are purely associated with the rotational flow, $\mathbf{v}_\psi$, are regarded as representing the extratropical barotropic response:

$$\begin{aligned} \frac{\partial \zeta}{\partial t} + \mathbf{v}_\psi \cdot \boldsymbol{\nabla}\zeta \ &= -\boldsymbol{\nabla} \cdot (\mathbf{v}_\chi \zeta) \\ &= -\zeta \boldsymbol{\nabla} \cdot \mathbf{v}_\chi - \mathbf{v}_\chi \cdot \boldsymbol{\nabla}\zeta \quad . \end{aligned} \tag{3}$$

To emphasis this separation, the right-hand side of equation (3) is sometimes known as the 'Rossby-wave source'. The second form of equation (3) splits the Rossby wave source into the divergence component and the component associated with advection by the divergent wind.

Previous studies (e.g. Ting, 1996) have attempted to identify equivalent barotropic levels that best represent the extratropical response to this Rossby wave source. The present study also uses equation (3) to investigate the vorticity balances in extratropical stationary waves. As in Qin and Robinson (1993), it is found (see later) that the divergence (*i.e.* 'stretching') 'source' of vorticity is not negligible in the extratropics and so the right and left-hand sides of equation (3) are not interpreted so strictly as tropical forcing and extratropical response, respectively.

All the terms in equation (3) except for the local time derivative are quadratic and so they are deduced here on a quadratic Gaussian grid. The calculation is done with the same spectral transforms as used by the ECMWF forecasting system. A scale-selective smoothing is applied that scales a spectral coefficient s_n^m, with zonal wavenumber m and total wavenumber $n > n_0 = 10$, by $n_0(n_0 + 1)/(n(n + 1))$. If, instead, one sets $n_0 = 1$, then the smoothing would map the Rossby-wave source onto spatial scales compatible with the anomalous streamfunction, $\psi = \nabla^{-2}\xi$, $\psi_n^m = -\xi_n^m a^2/(n(n + 1))$, where ξ is relative vorticity, a is the radius of the Earth and $\psi_0^0 = 0$. One could argue that these are the scales of interest when examining teleconnections. However, such a severe smoothing would not allow a good examination of the Rossby-wave dynamics and would, for example, smooth-away much of the zonal wavenumber 5 wave-guide-mode highlighted by Branstator (2002). With the $n_0 = 10$ smoothing applied here, it is sufficient to truncate the input vorticity and divergence data to T42 as smaller scales have negligible impact after smoothing.

Finally, to ensure that the vorticity analysis is applicable to general circulation models, each term evaluated is vertically integrated between 300 and 100 hPa. The rational for vertically integrating is that relatively minor changes to model physics can change the level of convective out-flow in the tropics. A Rossby-wave source
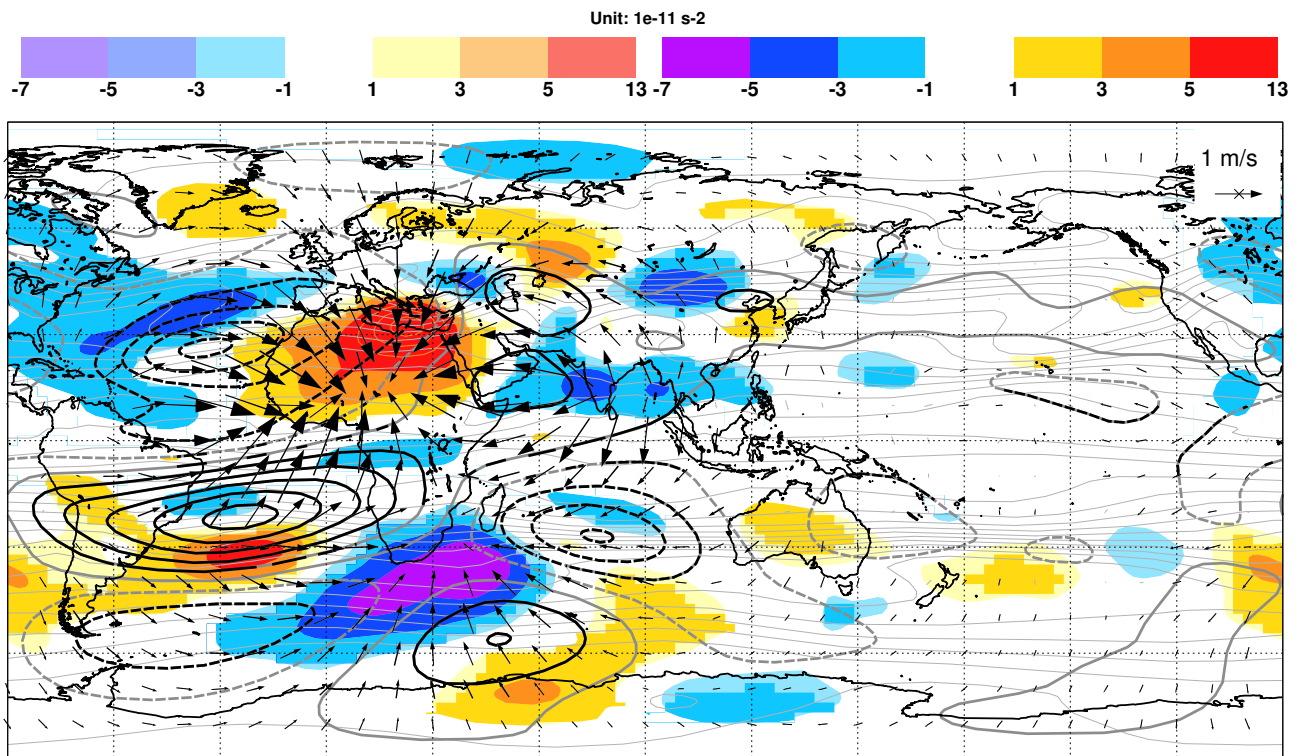
Unit: 1e-11 s-2

Figure 5: June–August mean change in upper tropospheric flow diagnostics from the 40 years of seasonal integrations. Arrows show the change in divergent winds. Shading shows the change in Rossby-wave source derived from daily data. Thick contours show the change in streamfunction. The change is in the sense of 'new' minus 'old' aerosol. Also shown in thin grey contours is the mean absolute vorticity (the full field is shown; not the change). The diagnostics are derived from the two sets of seasonal integrations (see main text for details). Daily data at 12 UTC is used throughout. The shading interval for the anomalous Rossby-wave source is generally $2 \times 10^{-11} s^{-2}$ but note that orange contours are used with the same interval to divide the most extreme (red) colour. The anomalous streamfunction contours are shown at ± 1, ± 3, ± 5,... $\times 10^6 \, m^2 s^{-1}$. Contours of the mean absolute vorticity are displayed at ± 5, ± 10, ± 15,... $\times 10^{-6} s^{-1}$. All quantities are vertically integrated between 300 and 100hPa. Black arrows, black contours and bold shading indicate 10% statistical significance for differences of seasonal-means.

diagnostic based on a single level could be strongly affected by this. On the other hand, the strength of the extratropical zonal-mean zonal flow generally maximises at around 200 hPa and is, therefore, reasonably constant over the 300-100 hPa layer. Hence the characteristics of stationary extratropical Rossby-waves are likely to be reasonably similar for all levels between 300 and 100 hPa.

Figure 5 shows the June–August mean of the *daily* anomalous Rossby-wave source (shaded) from the seasonal integrations along with the corresponding anomalous divergent wind vectors and anomalous streamfunction (thick contours). Significance, at the 10% level, is indicated by bold shading, black vectors and black contours, respectively. Also shown with thin grey contours is the mean absolute vorticity (averaged over both sets of seasonal integrations).

The anomalous divergent wind vectors in Figure 5 highlight the anomalous upper-tropospheric convergence associated with the weaker north African monsoon and also the increased divergence over southern Asia and the northern Indian Ocean (which were associated above with coupling between the convection and the upwelling Kelvin-waves). The divergent wind anomaly field associated with these tropical changes has a large-scale and can be seen to extend into the midlatitudes.

The upper-tropospheric 'quadrupole' streamfunction anomaly seen in Figure 5, with low streamfunction anoma-

lies over the subtropical north Atlantic and southern Indian Ocean and high anomalies over the subtropical south Atlantic and Arabian Peninsular (and Capsian Sea), virtually eliminate the mean errors in upper-tropospheric streamfunction in this region relative to the ERA-40 climatology (not shown). Before attempting to understand this response of the rotational flow to the aerosol change, the Rossby-wave source is first examined.

The mean Rossby-wave source anomaly shown in Figure 5 is calculated from daily data as

$$-\overline{\overline{\boldsymbol{\nabla} \cdot \left(\mathbf{v}_\chi \zeta \right)}}^{(N-O)} \quad , \tag{4}$$

where an overbar indicates a seasonal-mean over 92 days and a second overbar indicates a mean over the 40 years of simulations. The subscript $^{(N-O)}$ refers to the new aerosol minus old aerosol.

The strongest Rossby-wave source anomalies are centred over the north coast of Egypt ($> 11 \times 10^{-11}\text{s}^{-2}$) and to the south of South Africa ($< -5 \times 10^{-11}\text{s}^{-2}$). Both these anomalies also act to reduce the mean error in the Rossby-wave source relative to ERA-40 (not shown). In both these regions, the anomalous divergent wind is clearly advecting higher magnitude planetary vorticity towards the equator but it will be seen later that this is not the dominant component of the Rossby-wave source.

The Rossby-wave source is a quadratic term and so it is interesting (from a stationary-wave forcing perspective) to know how important transients are for the time-mean anomaly. Figure 6(a) shows the anomalous Rossby-wave source deduced from seasonal-mean data as

$$-\overline{\boldsymbol{\nabla} \cdot \left(\overline{\mathbf{v}}_\chi \overline{\zeta} \right)}^{(N-O)} \quad . \tag{5}$$

Taking the seasonal mean before calculating the Rossby-wave source should eliminate the effects of transients. The similarity between Figure 6(a) and the anomalous Rossby-wave source shown shaded in Figure 5 demonstrates that the time-mean anomalous Rossby wave-source is essentially unaffected by transient activity.

Equation (5) can be decomposed into four terms:

$$\left(\begin{array}{c} -\overline{\overline{\zeta}^{(O)} \boldsymbol{\nabla} \cdot \overline{\mathbf{v}}_\chi^{(N-O)}} - \overline{\overline{\mathbf{v}}_\chi^{(N-O)} \cdot \boldsymbol{\nabla} \overline{\zeta}^{(O)}} \\[2mm] -\overline{\overline{\zeta}^{(N-O)} \boldsymbol{\nabla} \cdot \overline{\mathbf{v}}_\chi^{(N)}} - \overline{\overline{\mathbf{v}}_\chi^{(N)} \cdot \boldsymbol{\nabla} \overline{\zeta}^{(N-O)}} \end{array} \right) . \tag{6}$$

Figure 6(b) shows the first of these four terms (the component associated with anomalous divergence). The similarity with the sum of all four terms (Figure 6a) demonstrates that the source of vorticity associated with anomalous divergence is dominant in the anomalous Rossby-wave source over most of the globe. The third and forth terms in equation (6) (not shown) are found to be negligible. Hence perturbations to the absolute vorticity are not important here for the seasonal-mean anomalous Rossby-wave source. The difference; Figure 6(a) minus (b) is therefore a good approximation for the second term in equation (6); the advection of absolute vorticity by the anomalous divergent wind. It can be seen (by comparing Figure 6a and b) that this term is generally also a small component of the Rossby-wave source although it has some contribution over the north coast of Egypt.

Sardeshmukh and Hoskins (1988) attribute almost the entire anomalous Rossby-wave source in the subtropical westerlies to the second term; the advection of absolute vorticity by the divergent flow anomalies. One may ask how this difference of emphasis arises and whether there is an impact on the total Rossby-wave source? The following integral constraint may partially address this question. As perturbations to the absolute vorticity have

(a) Rossby Wave Source (full)

(b) Rossby Wave Source (anomalous divergence component)

(c) Rotational Advection (by anomalous wind)

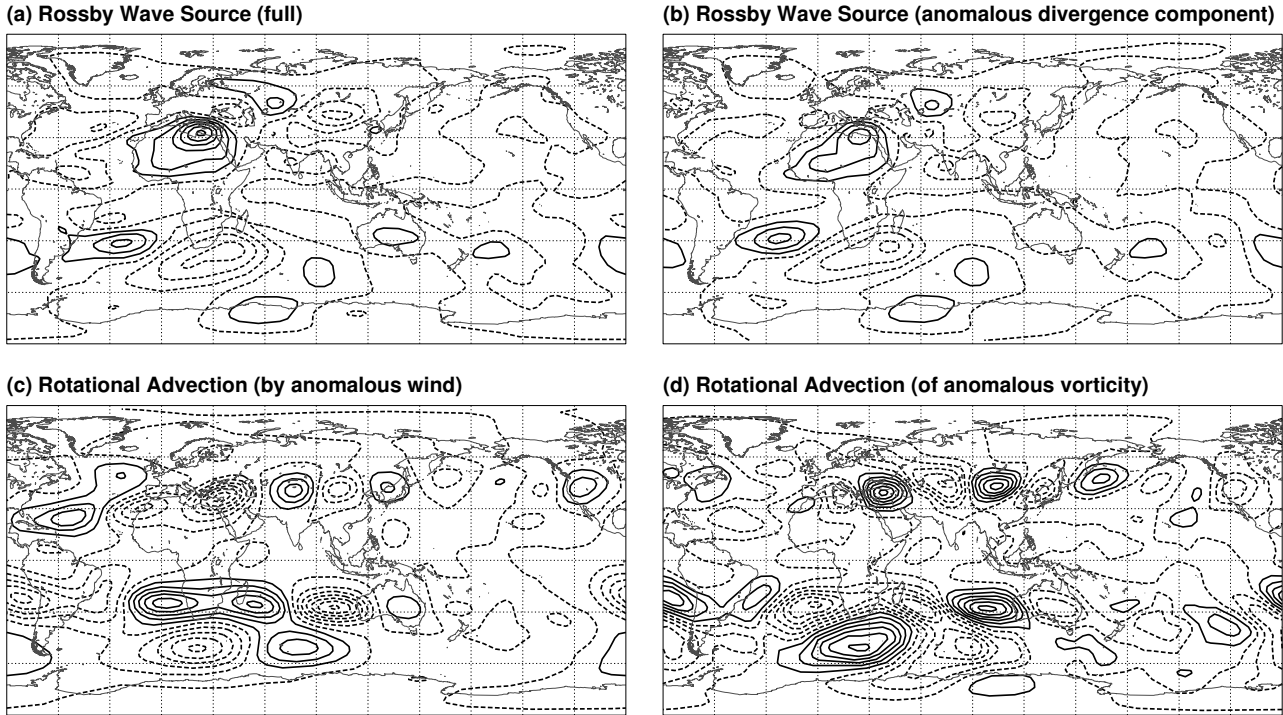**(d) Rotational Advection (of anomalous vorticity)**

Figure 6: June–August mean change in various terms in the vorticity equation when derived from seasonal-mean data from the 40 years of seasonal integrations. (a) The change in Rossby-wave source; $-\overline{\boldsymbol{\nabla}\cdot\left(\overline{\mathbf{v}}_\chi\overline{\zeta}\right)}^{(N-O)}$. (b) The change in the Rossby wave source due to changes in divergence; $-\overline{\overline{\zeta}^{(O)}\boldsymbol{\nabla}\cdot\overline{\mathbf{v}}_\chi^{(N-O)}}$, (c) The change in the advection of absolute vorticity due to changes in rotational wind, $-\overline{\overline{\mathbf{v}}_\psi^{(N-O)}\cdot\boldsymbol{\nabla}\overline{\zeta}^{(O)}}$, (d) The change in the advection by the rotational wind due to changes in absolute vorticity, $-\overline{\overline{\mathbf{v}}_\psi^{(N)}\cdot\boldsymbol{\nabla}\overline{\zeta}^{(N-O)}}$. The diagnostics are derived from the two sets of seasonal integrations (see main text for details). An overbar indicates a seasonal-mean and a second overbar indicates a mean over the seasons. A superscript $^{(N)}$ indicates the 'new' aerosol experiment, a superscript $^{(O)}$ indicates the 'old' aerosol experiment and a superscript $^{(N-O)}$ indicates 'new' minus 'old' aerosol. Contours are displayed at ±1, ±3, ±5,... $\times10^{-11}s^{-2}$. Almost all features in this figure are significant at the 10% level.

been found to have negligible impact on the Rossby wave source, the integral of the anomalous Rossby-wave source over a fixed area τ with boundary l can be written as:

$$\int_\tau \overline{\zeta}\boldsymbol{\nabla}\cdot\mathbf{v}'_\chi + \mathbf{v}'_\chi\cdot\boldsymbol{\nabla}\overline{\zeta}\ d\tau = \oint_l \overline{\zeta}\mathbf{v}'_\chi\cdot\mathbf{n}dl \quad, \tag{7}$$

where $\mathbf{n}$ is the unit (horizontal) vector normal to l. In the specific case considered here, $\overline{\zeta}=\overline{\zeta}^{(O)}$ and $\mathbf{v}'_\chi=\overline{\mathbf{v}}_\chi^{(N-O)}$.

Equation (7) indicates that if the anomalous divergent wind, $\mathbf{v}'_\chi$, can be modified within τ but not close to the boundary, l, then the relative contributions of the divergence and divergent advection can be modified but the area-integral of the anomalous Rossby-wave source will remain constant. Such a modification to $\mathbf{v}'_\chi$ can be made by altering the anomalous velocity potential, χ', within τ but not close to the boundary, l. For example, a local convergence centre (*i.e.* a χ'-maximum since $\boldsymbol{\nabla}\cdot\mathbf{v}'_\chi = \nabla^2\chi'$) could be moved within the area, τ, to a lower latitude where $|\overline{\zeta}|$ is smaller. This would reduce the divergence component and increase the advection component but leave the overall anomalous Rossby wave source unchanged. Hence it is possible that the

stronger emphasis on the divergence term seen here is compensated-for by a reduction in the relative importance of the advective term, with little impact on the total Rossby-wave source (which is essentially dictated by the tropical convection anomaly). Whether or not this is the full explanation, it is clear that the real world involves complexities that are not represented in classical barotropic vorticity equation studies where a non-interactive, idealised, divergence anomaly is used to represent the effects of a (hypothetical) local heating anomaly. The general conclusion would be that, without prior knowledge of which term is dominant, it is safer to calculate the total Rossby-wave source.

The response of the rotational flow to this Rossby-wave forcing is now examined. The seasonal-mean of the anomalous local tendency in (the anomaly version of) equation (3) should be negligible and so, where the assumptions made in equation (3) are valid, there should be a balance between the mean anomalous Rossby-wave source and the mean anomalous advection of absolute vorticity by the rotational flow. An interesting first question concerns whether such a time-mean balance can be achieved by a *stationary* rotational flow anomaly or whether transients are important for the mean anomalous 'rotational advection'. It has already been established that transients are not important for the mean anomalous Rossby-wave source. Using the same approach of calculating the quadratic term using seasonal-mean rather than daily values, it is found that transients also have a negligible impact on the mean rotational advection. This indicates that the rotational flow anomaly, displayed by the streamfunction contours in Figure 5, represents a stationary response to the mean anomalous Rossby-wave source.

To first-order, the vorticity balance can therefore be expressed as:

$$
\begin{aligned}
-\overline{\mathbf{v}_\psi^{(N-O)} \cdot \boldsymbol{\nabla} \overline{\zeta}^{(O)}} - \overline{\mathbf{v}_\psi^{(N)} \cdot \boldsymbol{\nabla} \overline{\zeta}^{(N-O)}} \\
-\overline{\overline{\zeta}^{(O)} \boldsymbol{\nabla} \cdot \overline{\mathbf{v}}_\chi^{(N-O)}} \approx 0 \quad .
\end{aligned} \tag{8}
$$

This balance is evident in Figure 6 where the first and second terms in equation (8) are displayed in Figure 6(c) and (d), respectively and the third term (the dominant component of the Rossby-wave source), which has already been discussed, is shown in Figure 6(b). In addition, in regions where the Rossby-wave source is relatively small, the first two terms in equation (8) should cancel each other. Physically, this cancellation can be thought of as the Rossby wave propagating upstream (first term in equation (8), Figure 6c) at the same speed as it is being advected downstream (second term in equation (8), Figure 6d). This balance is seen rather beautifully over Asia and into the North Pacific and reflects a Rossby wave being forced, primarily, by the Rossby-wave source (third term in equation (8), Figure 6b) centred over the north coast of Egypt. Meridional oscillations in the northern-most streamfunction contour over Asia and the north Pacific in Figure 5 reflect this stationary Rossby wave. This wave is in excellent agreement with the Asian 'waveguide' mode produced by Ambrizzi et al. (1995) using a barotropic vorticity equation model (their Fig 16). A balance between the first two terms in equation (8) is also seen over the subtropical north Atlantic (compare Figure 6c and d). The response in this region was discussed in section 3.2.2 in terms of classical equatorial Rossby wave theory. However, the present balance emphasises a conclusion of Rodwell and Hoskins (1996) that the westerly flow over the subtropical north Atlantic, and not simply the dissipation term used by Gill (1980), is important for the stationarity of this Rossby wave feature.

In the Southern Hemisphere, Figure 5 indicates that the Rossby-wave source to the southeast of South Africa initiates two Rossby waves, one centred at around 25^oS and another centred at around 50^oS. Again, this is highly consistent with Ambrizzi et al. (1995) who discuss a subtropical waveguide and a polar waveguide in their observational, theoretical and modelling results. It is clear that such waveguide theories are of great practical use for the investigation of remote responses to model physics changes. In this study, these two waves
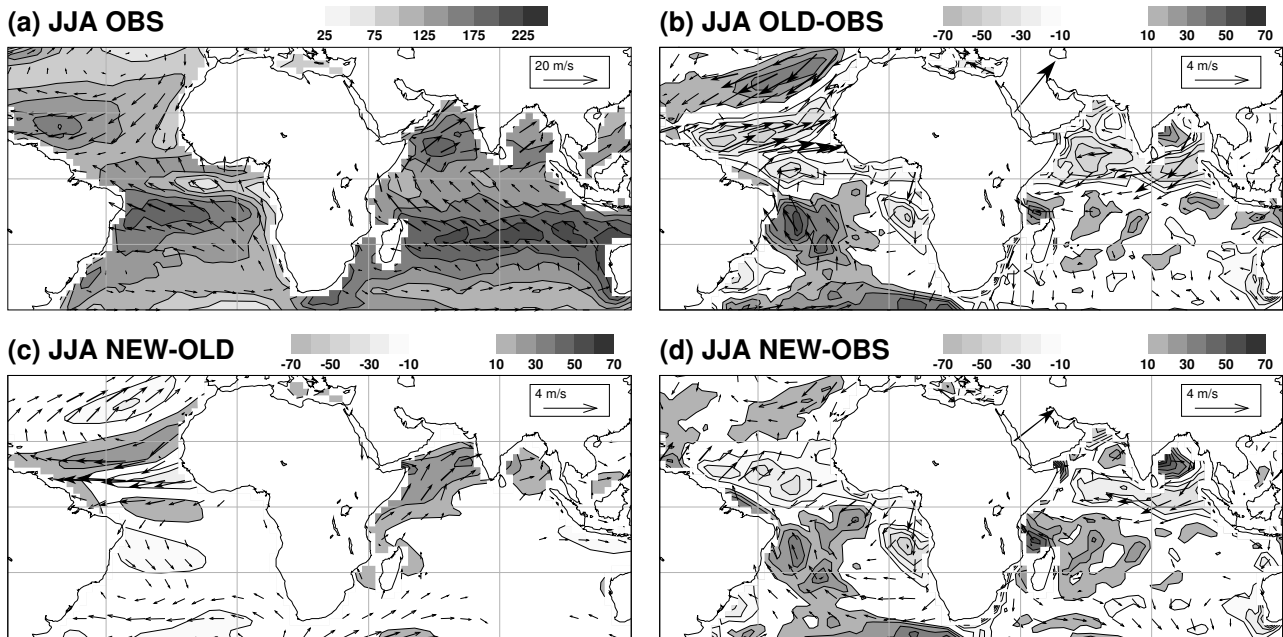
Figure 7: Diagnostics of June–August seasonal-mean surface latent heat flux (Wm^{-2}), shaded, and 10m wind vectors (ms^{-1}, see reference arrow). Observational data come from Grist and Josey *(2003) for surface latent heat flux over the period 1980–1993 and from ERA-40 for near-surface wind over the period 1962–2001. Forecast data come from the seasonal integrations covering the period 1962–2001 (see main text for details). (a) Mean observed. (b) Mean model error with 'old' aerosol. (c) Mean difference: model with 'new' aerosol minus model with 'old' aerosol. (d) Mean model error with 'new' aerosol.*

have opposite phases, with a high anomaly north of a low anomaly etc. At 50°S, the Rossby wave source (Figure 6b) can be seen to 'aid' the advection of mean vorticity by the anomalous rotational winds (Figure 6c). For the wave around 25°S, the Rossby-wave source (Figure 6b) tends to 'aid' the advection of anomalous vorticity by the mean rotational winds (Figure 6d). This is a partial explanation for the opposing phases of the two waves. Just how the three terms in equation (8) combine to produce a steady flow response may be sensitive to the mean gradient of absolute vorticity, which is stronger at 25°S than at 50°S. The wave at around 50°S is, as might be expected, more barotropic in nature and this is the wave that dominates the mean mid-tropospheric and lower-tropospheric response seen in Figure 2(c).

It is not obvious from these seasonal integrations how the anomalous flow would be set-up over time. Similar plots to Figure 5 have been produced (not shown) based on the 124 weather forecasts (31 days × 4 times per day). These show the evolution of the mean anomaly from the start of the forecast (when the small differences reflect analysis differences) to a leadtime of 10 days (when the differences are in good agreement with Figure 5). It is found that the divergent and rotational flow anomalies evolve together (rather than the Rossby-wave source being established first and the rotational flow responding sometime later). The time evolution of the rotational flow is consistent with the instantaneous imbalance between the Rossby-wave source and the rotational advection. These results imply that aerosol anomalies of the magnitude discussed here can have an impact on extratropical forecasts at lead-times of around 10 days.

Note that equation (8) is not appropriate to describe the vorticity balance over tropical Africa itself. One can see, for example, that the divergence term (Figure 6b) is not negligible around 15°N,0°E but neither component of the rotational advection (Figure 6c and d) is able to balance it. Clearly other terms that were neglected in the vorticity equation (2) are important in this small region.

3.2.4 Impact on the coupled ocean-atmosphere system

The aerosol change has been shown to lead to improvements in local atmospheric physics and global atmospheric teleconnections. Here, the consequences for coupled ocean-atmosphere models are briefly investigated by examining surface flux responses in the atmospheric model. There is already evidence that coupled models do respond differently to aerosol radiative forcing, particularly over the oceans (Miller and Tegen, 1998). However, it is difficult when analysing long integrations of coupled models to disentangle the impact *on* surface fluxes of (1) the purely-atmospheric response to the aerosol change from (2) the subsequent ocean-atmosphere feedbacks. Here, the purely-atmospheric response is diagnosed by calculating surface fluxes in the seasonal integrations of the atmospheric model forced with realistic sea-surface temperature. The hope would be that the purely-atmospheric response to the change in aerosol leads to an improvement in surface fluxes. Such an improvement would then suggest that the climate of the coupled model would not drift so quickly (or so far) from that of the real world. Before quantifying the atmospheric model's response, the observed surface (latent) heat fluxes are discussed.

Climatological near-surface wind vectors from ERA-40 and observed turbulent surface latent heat fluxes from Grist and Josey (2003) are shown in Figure 7(a) for the June–August season. In the tropical and sub-tropical Atlantic as well as in the Indian ocean, large latent heat fluxes out of the ocean generally coincide with strong near-surface climatological mean winds. An exception is in the oceanic upwelling regions, such as the eastern tropical Atlantic, where relatively small latent heat fluxes coincide with cold sea-surface temperatures.

Mean errors in simulating near-surface wind speed and surface latent heat fluxes are shown in Figure 7(b) for the model with the old aerosol climatology. By comparing with Figure 7(a) it can be seen that, where mean wind errors enhance the observed climatological winds, the mean latent heat flux errors are positive. Similarly, where mean wind errors oppose the observed climatological winds, the mean latent heat flux errors are negative.

Figures 7(c) and (d) show that the change in aerosol climatology leads to substantial improvements in near-surface wind speeds and, therefore, also the surface latent heat fluxes. This is particularly true for the tropical and subtropical North Atlantic and the Arabian Sea, where latent heat-flux anomalies amount to as much as ± 30 Wm^{-2} (even though sea-surface temperatures are the same for both sets of integrations).

The reduction of mean errors of the ECMWF atmospheric model in simulating turbulent surface heat fluxes may lead to a reduction of the climate drift (Moore and Gordon, 1994) of ECMWF's coupled ocean-atmosphere model, which is used to produce seasonal forecasts. The reduction of mean errors in near-surface winds may also lead to important improvements in surface momentum fluxes. Tropical Atlantic surface momentum flux improvements may, for example, lead to improvements in the tropical Atlantic thermocline tilt and thus improvements in low-frequency coupled variability.

4 The December – February season

For completeness, some of the sections above are briefly repeated for the December–February season. As would be expected, much of the reasoning given for the June–August season carries over to the December–February season.

4.1 Seasonal-mean differences

Figure 8 shows a similar plot to Figure 2 but for the December–February season based on the seasonal integrations started on 1 October for the years 1962–2001. Figure 8(a) shows mean December–February precipitation,

low-level (925 hPa) winds and 500 hPa geopotential heights from the observational data. The Southern Hemisphere summer monsoons over South America, Southern Africa and northern Australia together with their associated low-level inflows are clearly evident. In the winter (northern) extratropics, the westerly jet is stronger than it was in the June–August season (Figure 2a) while in the summer (southern) hemisphere, the jet is weaker than it was in the June–August season.

Some of the statistically significant mean errors for the old aerosol integrations (Figure 8b) are reduced when the new aerosol is introduced (Figure 8c). These improvements include a reduction in the erroneous precipitation over the Gulf of Guinea, a beneficial increase in mean precipitation over the equatorial Indian Ocean, a substantial reduction in the extratropical high geopotential height bias over the North Pacific and a reduction in the low geopotential height bias centred over the coast of California. These height biases had been longstanding problems for the ECMWF model (e.g. Jung, 2005). The height changes are reflected in the substantial improvements in mean low-level wind over the North Pacific and (not shown) improvements to synoptic activity in the North Pacific stormtrack region. Interestingly, Miller and Tegen (1998) also found a statistically significant mean response over the North Pacific to the radiative forcing by dust aerosol. In this study, it will be demonstrated that these extratropical anomalies are 'connected' to those in the tropics through the action of upper-tropospheric Rossby waves.

4.2 Local physics

The same physical reasoning as given for the June–August season is used to explain how the local physics in December–February responds to the change in aerosol. Hence the strong reduction in precipitation over the Gulf of Guinea in response to the reduced aerosol (compare Figure 1a,b) is likely to be triggered by reduced short-wave absorption. However, the strength of the change may be particularly strong over the Gulf of Guinea because the reduced short-wave absorption is *not* balanced, over a region of prescribed sea-surface temperature, by surface long-wave, sensible or latent heat fluxes. This explains the great improvement in precipitation in this region in terms of the atmospheric model being used here but it also highlights a potential difference between atmosphere and coupled ocean-atmosphere simulations. In atmospheric simulations, the ocean effectively has an infinite heat capacity whereas increased downward short-wave could be expected to increase sea-surface temperature in a coupled model and eventually lead to surface fluxes that compensate for the direct atmospheric cooling effect. One may speculate that, in the real world, short timescale fluctuations of aerosol are potentially much more powerful over the ocean than over the land but that this disparity diminishes at longer timescales.

The strong reduction in aerosol over the Sahara (compare Figure 1a and b) also leads to a radiative cooling anomaly (not shown). Although there is no precipitation in this region for this forcing to positively feed-back with at this time of year, the radiative cooling alone is apparently enough to force some descent and upper-level convergence.

4.3 Tropical impacts

With similar reasoning to that for the June–August season, the December–February decrease in precipitation over the Gulf of Guinea and equatorial Africa is likely to force eastward-propagating, upwelling Kelvin waves over the Indian Ocean. The increased precipitation seen over the tropical Indian Ocean in Figure 8(c) is consistent with a diabatic coupling with these waves. Such coupling was speculated for the June–August season above. However, for the December–February season, it is also possible that the reduction in aerosol over the tropical Indian Ocean itself (compare Figure 1 a and b) could contribute to the increase in precipitation (Miller and Tegen, 1998).
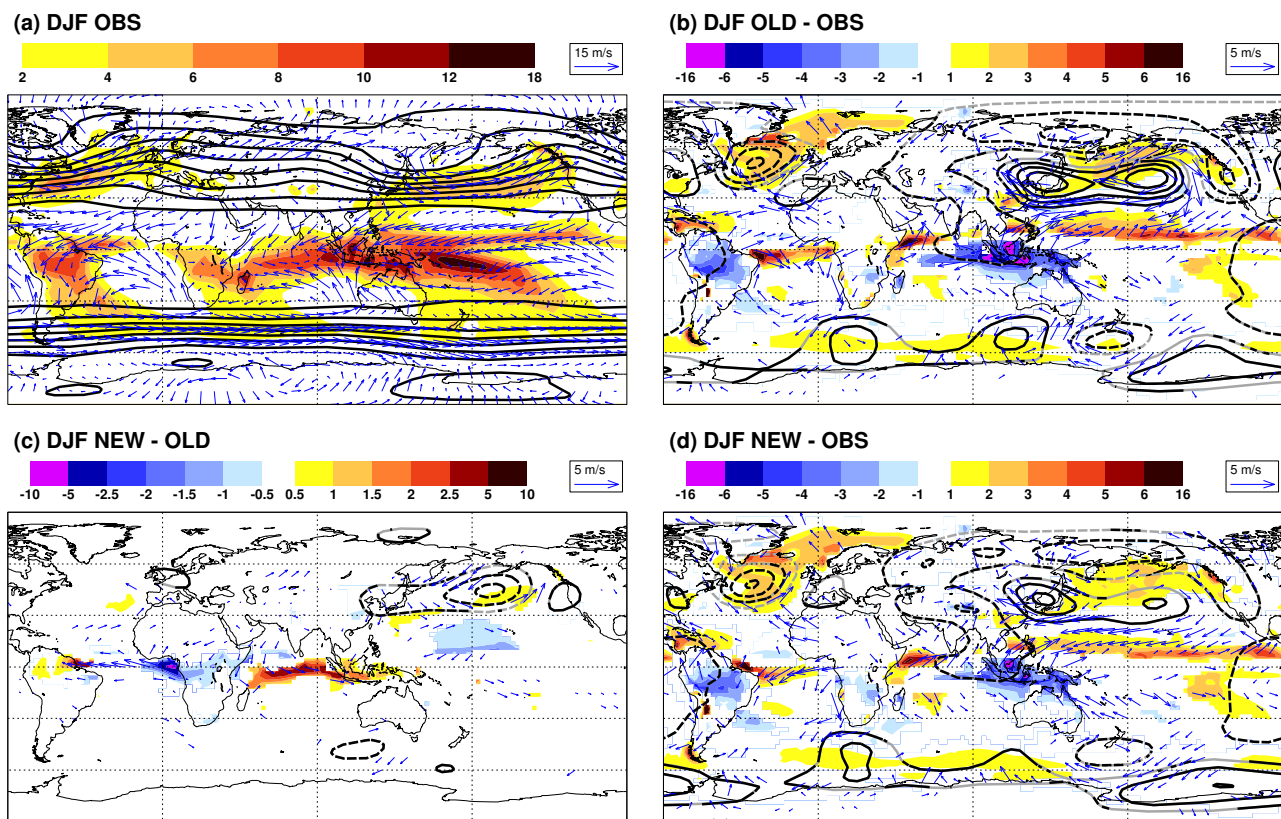
(a) DJF OBS

(b) DJF OLD - OBS

(c) DJF NEW - OLD

(d) DJF NEW - OBS

Figure 8: As Figure 2 but for December–February seasonal-means.

4.4 Extratropical impacts and feedbacks

Figure 9 shows the same upper-tropospheric diagnostics as in Figure 5, but for the December–February season. The tropical divergent wind anomalies, with anomalous convergence towards the Gulf of Guinea and divergence from the equatorial Indian Ocean are consistent with the latent heating changes associated with the anomalous precipitation. The Rossby-wave source centres located over Spain/Morocco and over Afghanistan appear to be related to the tropical response. Interestingly, these centres are on the northern flank of the subtropical jet and their signs are opposite to what one would expect from the vorticity advection by the divergent wind anomalies south of the subtropical jet. Again, the local divergence term must be dominating the Rossby-wave source in these regions but a full understanding has not been investigated here. Given that the Rossby-wave source is as it is in Figure 9, one can see that it is consistent with the wave-train in anomalous streamfunction (thick contours) that spreads along the jetstream to the east. Using the same methodology as used for the June–August season, it is again established that this wave is part of a stationary wave solution. The wave clearly connects to the North Pacific anomalies seen in 500hPa geopotential heights (Figure 8c). This stationary wave-train has strong similarities with the circumglobal wave of Branstator (2002). The wave's path and zonal wavenumber are highly reminiscent of the Asian jet-stream waveguide highlighted by Hoskins and Ambrizzi (1993). Indeed their barotropic vorticity equation model even predicts a southward turning of the stationary wave as it approaches the west coast of North America just as can be seen in Figure 9.

Interestingly, the amplitude of the wave-train in anomalous streamfunction in Figure 9 is relatively large over the North Pacific. The non-divergent barotropic vorticity equation model of Hoskins and Ambrizzi (1993) also shows that the largest amplitude (in relative vorticity anomaly) for their wave in the Asian waveguide is centred over the west Pacific (at around 140^oE, 30^oN). Although the location is not quite the same as found here,
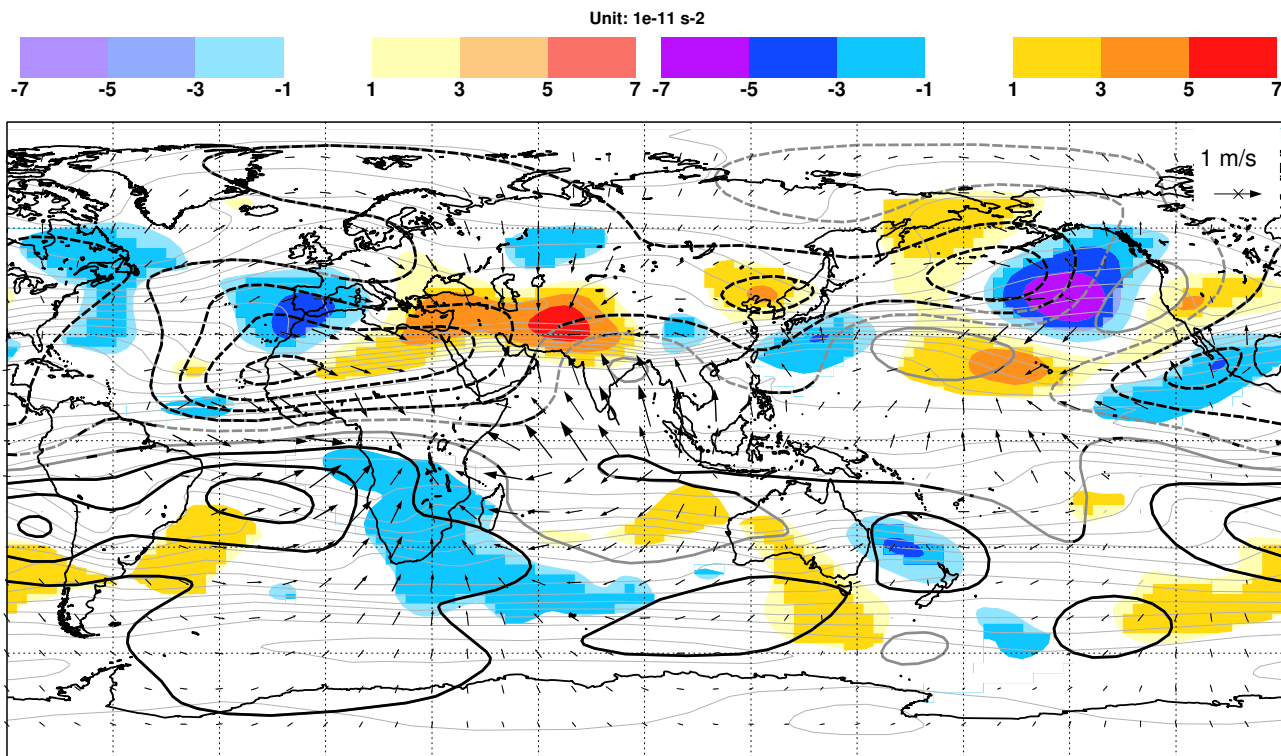
Figure 9: As Figure 5 but for December–February mean change.

it is clear that non-divergent dynamical processes alone can account for some downstream increase in wave amplitude. In the present study, however, Figure 9 shows that there is also a strong negative 'Rossby-wave source' anomaly centred at 210^oE, 40^oN. The inverted commas are used around 'Rossby-wave source' here because the location is far from the tropics and so this vorticity source cannot be viewed as an independent forcing. Instead, this 'Rossby-wave source' indicates that divergent processes also play a role in defining the stationary wave pattern over the North Pacific. The main component to this 'Rossby-wave source' anomaly is, as before, the divergence term rather than the advection by the divergent wind. One possible origin for the negative Rossby-wave source term here is adiabatic vortex-tube shrinkage in the anomalous northward flow. (The anomalous flow is northward because the rotational wind dominates the divergent component shown in Figure 9. Anomalous vortex shrinkage would occur because the isentropic surfaces get closer together towards the pole). However, it is intriguing to note that Figure 8(c) shows increased precipitation in this region. Increased precipitation would be consistent with upper-tropospheric potential vorticity destruction. Other regions in the north Pacific with reduced precipitation (one such centre is visible in Figure 8(c) at around 180^oE, 15^oN) coincide with the two positive Rossby-wave source anomalies. Hence it is possible, if a little speculative at present, that there is some local diabatic modification, or even enhancement, of the tropically-forced stationary Rossby wave taking place over the North Pacific.

5 Conclusions

This paper can be viewed as a sensitivity study and explanation of the effects, local and global, of a change in model aerosol climatology. Alternatively, the paper can be seen as the development and seamless application of previous methodologies for understanding the general circulation, that could form a diagnostic package to assess almost any physics change within a general circulation model.

The aerosol climatology change is predominantly a reduction in Saharan dust. This change leads to improvements in local medium-range forecast skill. In addition, there are many improvements in the general circulation. These include strong local changes in the north African monsoon, tropic-wide improvements in precipitation and a reduction in mean extratropical circulation errors. These improvements must arise from the 'direct effects' of the aerosol change through changes in scattering and absorption by the aerosol and the consequent changes in atmospheric temperatures and circulation. Indirect aerosol effects, arising from changes in the density of cloud condensation nuclei and consequent changes in cloud and cloud duration are not simulated interactively within the model.

The 'initial tendency' methodology has been used to assess the impact of the aerosol change on the local physics in the north African monsoon region. This approach, which focuses on the first few timesteps of a weather forecast, provides a powerful and objective method of assessing errors-in and changes-to the 'fast' physical processes within a model. The aerosol change appears to induce a stronger thermal destabilisation of the vertical profile which initially enhances convection. Later, however, the reduced short-wave absorption by aerosol leads to a reduction in an erroneous feedback with the dynamics (seen for the old aerosol) that otherwise increases large-scale convergence, moisture supply and convection. Clearly, the relative importance of each response process may be model dependent and sensitive to the form of the aerosol change. What can be concluded, however, is that the reduced net initial tendencies provide confirmation that the new aerosol climatology is superior to the old climatology. Rodwell and Palmer (2007) applied the same initial tendency methodology to 'perturbed model ensembles', which are increasingly being used to quantify uncertainty in climate change predictions (Murphy et al., 2004; Stainforth et al., 2005). By down-weighting poorer models, Rodwell and Palmer (2007) demonstrated that the use of the initial tendency methodology could reduce uncertainty in climate sensitivity to increasing atmospheric CO_2. The initial tendency methodology requires that the model comes with its own data assimilation system. Phillips et al. (2004) considered day 5 forecast errors where, arguably, there is less need for a data assimilation system. However, when applied to the north African monsoon, medium-range forecast errors would have emphasised the feedbacks with the dynamics rather than the initial physics problems responsible for the forecast error. While the need for the model to come with its own data assimilation limits the application of initial tendencies at present, one would hope that results such as those given here will encourage the development of more seamless weather prediction / climate forecasting systems in future.

Equatorial wave theory (which is already well understood) has been used to provide pointers to the tropic-wide response to the Saharan aerosol change. However this study indicates that diabatic processes interact with these dynamical waves to greatly enhance the total tropical response.

To understand the extratropical response, this study has examined stationary-wave vorticity balances within the general circulation model. Unlike some previous barotropic vorticity equation studies, the forcing for extratropical waves tends to be dominated by the divergence component of the Rossby-wave source (rather than the advection component). Stationary extratropical Rossby-waves are clearly excited by the tropical responses to the aerosol change. These Rossby waves explain the regional extratropical circulation improvements and, thus, the changes in other extratropical features such as precipitation and transient activity. Interestingly, the co-location of precipitation anomalies and 'Rossby-wave source' anomalies in the extratropical suggests the possibility that these waves are being diabatically modified.

Surface latent heat fluxes over the oceans are also improved by the aerosol change. This indicates that the mean climate and variability of coupled ocean-atmosphere models would be improved with better representations of aerosol. At longer timescales, these interactions with the oceans are likely to modify the global circulation response to aerosol forcing (Miller and Tegen, 1998; Miller et al., 2004a).

Uncertainty in the future atmospheric loading of Saharan mineral dust under anthropogenic forcing scenarios

is comparable in magnitude to the aerosol change investigated in this present study. For example, Jones et al. (2007) show that anthropogenic forcing associated with changes in sulphate, biomass-burning, black carbon and CO_2 emissions each lead to a 10-20% decrease in mineral dust burden in the west Sahara / east subtropical North Atlantic region. While the present study does not address the question of the coupled climate response to aerosol, the results tend to confirm that there remain substantial uncertainties in regional climate change predictions throughout the globe.

The ECMWF-coordinated 'Global and regional Earth-system (Atmosphere) Monitoring using Satellite and in-situ data' project, (GEMS), aims to (amongst other things) improve the knowledge of atmospheric aerosol composition (Hollingsworth, 2005). The availability of these real-time aerosol observations will allow the incorporation of prognostic aerosol schemes in future weather forecast models. Indeed, current efforts at ECMWF are focused on the prognostic modelling of aerosol. The results of the present study suggest that these efforts could lead to further numerical weather prediction improvements. The methodologies described here could, of course, be re-applied to assess the physics and the global responses of these new models.

Acknowledgements

References

Ambrizzi, T., B. J. Hoskins, and H.-H. Hsu, 1995: Rossby wave propagation and teleconnection patterns in the austral winter. *J. Atmos. Sci.*, **52**, 3661–3672.

Branstator, G., 2002: Circumglobal teleconnections, the jet stream waveguide, and the North Atlantic Oscillation. *J. Climate*, **15**, 1893–1910.

Chou, C., J. D. Neelin, U. Lohmann, and J. Feichter, 2005: Local and remote impacts of aerosol climate forcing on tropical precipitation. *J. Climate*, **18**(22), 4621–4636.

Fouquart, Y. and B. Bonnel, 1980: Computation of the solar heating of the earth's atmosphere: a new parameterization, Beitr. *Phys. Atmosph.*, **53**, 35–62.

Gill, A. E., 1980: Some simple solutions for heat-induced tropical circulation. *Quart. J. Roy. Meteor. Soc.*, **106**(449), 447–462.

Goudie, A. S. and N. J. Middleton, 2001: Saharan dust storms: nature and consequences. *Earth Science Reviews*, **56**(1-4), 179–204.

Greatbatch, R. J. and T. Jung, 2007: Local versus tropical diabatic heating and the winter North Atlantic Oscillation. *J. Climate*, **20**(10), 2058–2075.

Grist, J. P. and S. A. Josey, 2003: Inverse analysis adjustment of the SOC air-sea flux climatology using ocean heat transport constraints. *JC*, **16**, 3274–3295.

Hansen, J., M. Sato, and R. Ruedy, 1997: Radiative forcing and climate response. *J. Geophys. Res.*, **102**(D6), 6831–6864.

Heckley, W. A. and A. E. Gill, 1984: Some simple analytical solutions to the problem of forced equatorial long waves. *Quart. J. Roy. Meteor. Soc.*, **110**(463), 203–217.

Hess, M., P. Koepke, and I. Schult, 1998: Optical properties of aerosols and clouds: The software package OPAC. *Bulletin of the American Meteorological Society*, **79**(5), 831–844.

Hollingsworth, A., 2005: The GEMS project - making a contribution to the environmental monitoring mission of ECMWF. ECMWF Newsletter 103, ECMWF, Shinfield Park, Reading, Berkshire RG2 9AX, UK.

Horel, J. D. and J. M. Wallace, 1981: Planetary-scale atmospheric phenomena associated with the Southern Oscillation. *Mon. Wea. Rev.*, **109**(4), 813–829.

Hoskins, B. J. and T. Ambrizzi, 1993: Rossby wave propagation on a realistic longitudinally varying flow. *J. Atmos. Sci.*, **50**, 1661–1671.

Hoskins, B. J. and D. J. Karoly, 1981: The steady linear response of a spherical atmosphere to thermal and orographic forcing. *J. Atmos. Sci.*, **38**(6), 1179–1196.

Hoskins, B. J. and M. J. Rodwell, 1995: A model of the Asian summer monsoon. Part I: The global scale. *J. Atmos. Sci.*, **52**(9), 1329–1340.

Hoskins, B. J. and P. D. Sardeshmukh, 1987: A diagnostic study of the dynamics of the northern hemisphere winter of 1985-86. *Quart. J. Roy. Meteor. Soc.*, **113**(477), 759–778.

Jones, A., J. M. Haywood, and O. Boucher, 2007: Aerosol forcing, climate response and climate sensitivity in the Hadley Centre climate model. *J. Geophys. Res.*, **112**, D20211.

Jung, T., 2005: Systematic errors of the atmospheric circulation in the ECMWF forecasting system. *Quart. J. Roy. Meteor. Soc.*, **131**, 1045–1073.

Klinker, E. and P. D. Sardeshmukh, 1992: The diagnosis of mechanical dissipation in the atmosphere from large-scale balance requirements. *J. Atmos. Sci.*, **49**, 608–627.

Matsuno, T., 1966: Quasi-geostrophic motions in the equatorial area. *J. Met. Soc. Japan*, **44**(1), 25–43.

Menon, S., J. Hansen, L. Nazarenko, and Y. Luo, 2002: Climate effects of black carbon aerosols in China and India.

Miller, R. L., J. Perlwitz, and I. Tegen, 2004a: Modeling arabian dust mobilization during the asian summer monsoon: The effect of prescribed versus calculated SST. *Geophys. Res. Lett.*, **31**.

Miller, R. L. and I. Tegen, 1998: Climate response to soil dust aerosols. *J. Climate*, **11**(12), 3247–3267.

Miller, R. L., I. Tegen, and J. Perlwitz, 2004b: Surface radiative forcing by soil dust aerosols and the hydrologic cycle. *J. Geophys. Res.*, **109**, 1–24.

Mlawer, E. J., S. J. Taubman, P. D. Brown, M. J. Iacono, and S. A. Clough, 1997: Radiative transfer for inhomogeneous atmospheres: RRTM, a validated correlated-k model for the longwave. *J. Geophys. Res.*, **102**, 16663–16682.

Moore, A. W. and H. B. Gordon, 1994: An investigation of climate drift in a coupled atmosphere-ocean-sea ice model. *Clim. Dyn.*, **10**, 81–95.

Murphy, J. M., D. M. H. Sexton, D. N. Barnett, G. S. Jones, M. J. Webb, M. Collins, and D. A. Stainforth, 2004: Quantification of modelling uncertainties in a large ensemble of climate change simulations. *Nature*, **430**(7001), 768–772.

Phillips, T. J., G. L. Potter, D. L. Williamson, R. T. Cederwall, J. S. Boyle, M. Fiorino, J. J. Hnilo, J. G. Olson, S. Xie, and J. J. Yio, 2004: Evaluating parameterizations in general circulation models: Climate simulation meets weather prediction. *Bull. Amer. Meteor. Soc.*, **85**(12), 1903–1915.

Qin, J. and W. A. Robinson, 1993: On the rossby wave source and the steady linear response to tropical forcing. *J. Atmos. Sci.*, **50**(12), 1819–1823.

Rabier, F., H. Järvinen, E. Klinker, J.-F. Mahfouf, and A. Simmons, 2000: The ECMWF operational implementation of four-dimensional variational assimilation. I: Experimental results with simplified physics. *Quart. J. Roy. Meteor. Soc.*, **126**, 1143–1170.

Rayner, N. A., D. E. Parker, E. B. Horton, C. K. Folland, L. V. Alexander, D. P. Rowell, E. C. Kent, and A. Kaplan, 2003: Global analyses of sea surface temperature, sea ice, and night marine air temperature since the late nineteenth century. *J. Geophys. Res.*, **108**(D14), 4407.

Reynolds, R., N. Rayner, T. Smith, M. Thomas, D. Stokes, and W. Wang, 2002: An improved in situ satellite sst analysis for climate. *J. Climate*, **15**, 73–87.

Rodwell, M. J., 2005: The local and global impact of the recent change in model aerosol climatology. ECMWF Newsletter 105, ECMWF, Shinfield Park, Reading, Berkshire RG2 9AX, UK.

Rodwell, M. J. and B. J. Hoskins, 1995: A model of the Asian summer monsoon. Part II: Cross-equatorial flow and PV behavior. *J. Atmos. Sci.*, **52**(9), 1341–1356.

Rodwell, M. J. and B. J. Hoskins, 1996: Monsoons and the dynamics of deserts. *Quart. J. Roy. Meteor. Soc.*, **122**(534), 1385–1404.

Rodwell, M. J. and B. J. Hoskins, 2001: Subtropical anticyclones and summer monsoons. *J. Climate*, **14**(15), 3192–3211.

Rodwell, M. J. and T. N. Palmer, 2007: Using numerical weather prediction to assess climate models. *Quart. J. Roy. Meteor. Soc.*, **133**(622 A), 129–146.

Sardeshmukh, P. D. and B. J. Hoskins, 1988: The generation of global rotational flow by steady idealized tropical divergence. *J. Atmos. Sci.*, **45**, 1228–1251.

Stainforth, D. A., T. Aina, C. Christensen, M. Collins, N. Faull, D. J. Frame, J. A. Kettleborough, S. Knight, A. Martin, J. M. Murphy, C. Piani, D. Sexton, L. A. Smith, R. A. Spicer, A. J. Thorpe, and M. R. Allen, 2005: Uncertainty in predictions of the climate response to rising levels of greenhouse gases. *Nature*, **433**(7024), 403–406.

Tanre, D., J. F. Geleyn, and J. Slingo, 1984: First results of the introduction of an advanced aerosol-radiation interaction in the ECMWF low resolution global model. In: H. E. Gerber and A. Deepak, eds., *Aerosols and Their Climatic Effects*, pp. 133–177. A. Deepak, Hampton, Va.

Tegen, I., P. Hollrig, M. Chin, I. Fung, D. Jacob, and J. Penner, 1997: Contribution of different aerosol species to the global aerosol extinction optical thickness: Estimates from model results. *Journal of Geophysical Research*, **102**(D20).

Tiedtke, M., 1989: A comprehensive mass flux scheme for cumulus parameterization in large-scale models. *Mon. Wea. Rev.*, **117**(8), 1779–1800.

Ting, M., 1996: Steady linear response to tropical heating in barotropic and baroclinic models. *J. Atmos. Sci.*, **53**(12), 1698–1709.

Tompkins, A. M., C. Cardinali, J.-J. Morcrette, and M. J. Rodwell, 2005: Influence of aerosol climatology on forecasts of the African Easterly Jet. *Geophys. Res. Lett.*, **32**, L10891. Doi:10.1029/2004GL022189.

Uppala, S., P. W. Kallberg, A. J. Simmons, U. Andrae, V. Da Costa Bechtold, M. Fiorino, J. K. Gibson, J. Haseler, A. Hernandez, G. A. Kelly, X. Li, K. Onogi, S. Saarinen, N. Sokka, R. P. Allan, E. Andersson, K. Arpe, M. A. Balmaseda, A. C. M. Beljaars, L. van de Berg, J. Bidlot, N. Bormann, S. Caires, F. Cheval-lier, A. Dethof, M. Dragosavac, M. Fisher, M. Fuentes, S. Hagemann, E. Holm, B. J. Hoskins, L. Isaksen, P. A. E. M. Janssen, R. Jenne, A. P. McNally, J.-F. Mahfouf, J.-J. Morcrette, N. A. Rayner, R. W. Saunders, P. Simon, A. Sterl, K. E. Trenberth, A. Untch, D. Vasiljevic, P. Viterbo, and J. Woollen, 2005: The ERA-40 re-analysis. *Quart. J. Roy. Meteor. Soc.*, **131**, 2961–3012.

von Storch, H. and F. W. Zwiers, 2001: *Statistical Analysis in Climate Research*. Cambridge University Press. 484 pp.

Webster, P. J., 1972: Response of the tropical atmosphere to local, steady forcing. *Mon. Wea. Rev.*, **100**(7), 518–541.

Wielicki, B. A., B. R. Barkstrom, E. F. Harrison, R. B. Lee III, G. Louis Smith, and J. E. Cooper, 1996: Clouds and the Earth's Radiant Energy System (CERES): An earth observing system experiment. *Bull. Amer. Meteor. Soc.*, **77**(5), 853–868.

Xie, P. and P. A. Arkin, 1997: Global precipitation: a 17-year monthly analysis based on gauge observations, satellite estimates and numerical model outputs. *Bull. Amer. Meteor. Soc.*, **78**, 2539–2558.

Yoshioka, M., N. M. Mahowald, A. J. Conley, W. D. Collins, D. W. Fillmore, C. S. Zender, and D. B. Coleman, 2007: Impact of desert dust radiative forcing on sahel precipitation: Relative importance of dust compared to sea surface temperature variations, vegetation changes, and greenhouse gas warming. *J. Climate*, **20**(8), 1445–1467.

Zender, C. S., R. L. Miller, and I. Tegen, 2004: Quantifying mineral dust mass budgets: terminology, con-straints, and current estimates. *Eos Trans. AGU*, **85**(48), 509–512.