Chapter 7: The impact of atmospheric dust on climate

7.1 Introduction

Chapter 6 linked dust production in the Sahel and Sahara with elements of the regional-scale climatology of northern Africa. It was demonstrated that interannual variability in atmospheric dust loadings over the Sahel-Sahara zone is related to changes in the prevailing monthly and seasonal synoptic situation and the frequency of easterly wave activity.

It is likely that mineral dust aerosols in turn influence the regional circulation (to an extent determined by their concentration, distribution and residence time), resulting in a feedback cycle in which changes in climate will alter the amount of dust produced, which will in turn alter the circulation. This has been suggested by a number of authors (Nicholson, 1995; Andreae, 1996), but until now sufficient data relating to mineral aerosols have not been available to test such hypotheses for continental northern Africa. Nonetheless, there is sufficient theoretical and empirical evidence to strongly suggest that atmospheric dust alters the thermal structure of the atmosphere. It is a small conceptual step to suggest that such changes may result in changes in patterns of atmospheric divergence and convergence, and hence modify the regional circulation.

This chapter describes correlation analyses of daily IDDI data with data representing three types of climatological variable obtained from the NCEP/NCAR reanalysis dataset. The analyses are complementary in that together they address the relationship between airborne Saharan and Sahelian dust and atmospheric motion on a variety of spatial scales, from several degrees up to scales representative of the Sahel-Sahara zone as a whole. The aim is to investigate the direct impacts of atmospheric dust loadings on temperature, vertical motion and specific elements of the general circulation. The climatological context of the studies is described insofar as it is directly relevant to the results.

Firstly, temporal relationships between IDDI and reanalysis temperature values at different levels of the atmosphere are examined. Daily IDDI data and 6-hourly reanalysis data facilitate such an analysis, which examines dust-temperature relationships on timescales comparable with, or shorter than, those which characterise major dust events.

Any impact of dust on the thermal structure of the atmosphere will affect the vertical temperature gradient, which influences the intensity and extent of convective activity. In order to assess any impacts of dust on the local-scale dynamics of the regional climate in a
more direct fashion, relationships between IDDI and vertical velocity are investigated. The
analysis is analogous to that performed using the temperature data. The aim of this analysis
is to determine whether there are any regions in which dust loadings may be associated with
large-scale reductions or increases in the intensity of the vertical motion of the atmosphere.

Large-scale modification of atmospheric vertical motion will alter the regional general
circulation to some extent. The most important question within the context of this thesis is:

   Do large amounts of atmospheric dust over West Africa act to suppress the West
   African Monsoon circulation through a weakening of the vertical temperature
   gradient and consequent reduction of large-scale convection over continental
   West Africa?

The monsoon air-flow may be represented by the meridional wind across the Gulf of Guinea
and into the sub-Saharan and Sahelian regions of West Africa. Having addressed the issues
of temperature and vertical motion, the final analysis investigates the relationship between
dust and monsoon circulation by examining correlations between areally averaged local
IDDI values and zonally averaged meridional wind indices for different latitudes over West
Africa and the Gulf of Guinea.

The above analyses are described in detail in the following sections. To begin with, the
hypotheses to be tested in the chapter are outlined. This is followed by a description of the
data, methodology and processing employed in the analyses. The results are then described
and interpreted, before being synthesised into a coherent description of the impact of dust on
the climate of the Sahel-Sahara zone.

7.2 Hypotheses to be tested and aims of the analyses

7.2.1 Dust-temperature relationships

The effects of dust aerosols on atmospheric temperatures will depend on their vertical
distribution, particle size, concentration and geographical location, and the time of year and
time of day (Chapter 2). Dust events are episodic in nature, and the timescales associated
with such events are of the order of hours or days. In any given month in which dust
production is significant, it may be expected that regions in the vicinity of dust sources, or
along the trajectory of dust clouds, will experience a number of dust events. If such events
are of sufficient intensity they will result in intermittent perturbations to the ambient
temperature fields at certain levels of the atmosphere. Such perturbations will manifest
themselves as correlations between dust and temperature series at particular locations. Performing correlation analyses between daily timeseries of dust, as represented by the IDDI data, and temperature will determine whether perturbations of the temperature field by dust are of sufficient magnitude to be detected over the periods of correlation. Because of the short response time of the local atmospheric temperatures to dust loadings, correlations are most likely to reveal a dust signal if performed between contemporaneous data.

Theoretical considerations suggest that under a dust layer, daytime temperatures will be reduced and night-time temperatures increased, while temperatures at levels in the vicinity of the dust layer will be increased throughout the day (Chapter 2). The predicted decrease in surface temperature under a dust layer is observed in the example field in Figure 7.1, which shows fields of IDDI anomalies upon which are superimposed 1000 hPa temperature anomaly contours at midday for 24 April 1984. This is a particularly clear example of cooling under a dust layer over the eastern Sahel and Sahara. Nonetheless, a region of negative temperature anomalies occurs over West Africa in a region characterised by negative IDDI anomalies, representing low dust loadings. Thus, in order to identify any dust signal in the temperature data, it is necessary to examine local variations in the IDDI and temperature fields over time, instead of relying purely on individual case studies to infer a relationship between dust and temperature.

![Figure 7.1: IDDI anomalies (coloured scale) with contours of 1000 hPa temperature anomalies superimposed for midday on 24 April 1984. Temperature data are from the NCEP/NCAR reanalysis dataset. All units are standard deviations.](image)

Of course, near-surface temperatures will be controlled by many factors other than dust loadings, such as the synoptic-scale meteorology and the amount of cloud cover. Also at issue is the reliability of the NCEP temperature data which, it must be remembered, are generated by a model. The results from the reanalysis simulations are constrained by the input fields of available observations, which drive the model to produce what is effectively a
realistic interpolation of the observed data. The model in question does not contain parameterisations of dust generation, transport or the impact of atmospheric dust on the thermal structure of the atmosphere. For this reason it might be expected that, where observational data are few (such as over the Sahara), the model may fail to simulate atmospheric temperature, as modulated by the presence of dust, accurately. However, where the interpolation is constrained by a higher density of observations, the simulation is likely to be more realistic, despite the inadequacies of the model in its representation of dust. The NCEP/NCAR temperature data are classified as class A data (Chapter 3) indicating that they are defined primarily by the observations; they are among the most reliable of the reanalysis data products.

The main aim of this analysis is to determine how fields of atmospheric temperature respond to atmospheric dust. An attempt will be made to identify any existing dust signals in the NCEP temperature fields by examining the sign and magnitude of correlations between IDDI and temperature values over multi-year periods representing single months. The length of the timeseries used (representing some 300 daily pairs) should be sufficient to isolate the impact of dust from those of other processes which lead to the ambiguity in the daily anomaly fields discussed above. The extent, and by extension the importance, of dust in determining the temperature throughout the atmosphere will be inferred by the spatial extent and coherence of the significant correlations. The monthly and seasonal variation of dust impacts will also be deduced. The issue of the reliability of the NCEP reanalysis data will be discussed further in the section dealing with the interpretation of the correlation patterns.

7.2.2. Dust-vertical motion relationships

It is expected that vertical motion will be affected by dust loadings in a similar fashion to temperature, with regions of dust-induced cooling being associated with reductions in convection and hence in vertical motion. It might also be expected, however, that the dust signal in the vertical motion fields will be less obvious that in the temperature fields, as vertical motion is determined to a great extent by the distribution of regions of convergence and divergence in the atmosphere, which are associated with the large-scale atmospheric flow. It should be noted that vertical motion data are classified as class B, indicating that they are influenced to a large extent by the properties of the model. It is therefore expected that the NCEP vertical velocity data will be less representative of reality than the temperature data.

The above hypotheses will be tested in the same manner as those concerning the impact of dust on atmospheric temperatures. The same considerations of the reliability of the NCEP
data will be borne in mind in the interpretation.

7.2.3. Dust impacts on the West African Monsoon

The monsoonal flow is modulated by the differential heating between continental West Africa and the ocean region of the Gulf of Guinea (Barry and Chorley, 1995), although this is not the sole cause of the monsoon (Stringer, 1972). The geography of West Africa dictates that such differential heating will be most pronounced west of 10° E, where the orientation of the African coastline changes from approximately north-south to east-west. It is at these longitudes that the southwesterly summer monsoon flow is dominant.

If such low-level atmospheric cooling and reductions in vertical motion are of sufficient magnitude and extent, they may lead to a reduction in the upward motion of air over large areas. If such a phenomenon occurs west of 10° E during the onset months of the West African Monsoon (WAM), the reduction in large-scale convective motion may be sufficient to significantly reduce the low and mid-level flow of moist air from the Gulf of Guinea across the Guinea Coast and into the Sahel. Such a reduced northwards transport might act to delay the onset of the monsoon and/or reduce the intensity of the monsoon flow, which would produce conditions less conducive to rainfall generation over the Sahel. The hypothesis to be tested here is that regional dust loadings exert a discernible influence on the meridional monsoon flow, with high dust loadings over West Africa being associated with reduced northwards transport of moist air over the Sahel. It is anticipated that such changes will be manifest as reductions in the meridional wind velocities at the atmospheric levels at which the majority of monsoon transport takes place.

7.3. Methodology and data processing

7.3.1. In situ correlations with temperature and vertical velocity

Temporal relationships between IDDI values and temperatures, and between IDDI values and vertical velocity, at various levels of the atmosphere, were assessed using 6-hourly NCEP reanalysis temperature data for different pressure levels. The analyses were identical for temperature and vertical velocity; the description below therefore refers to ‘reanalysis’ data to represent both quantities. The levels used were 1000, 850 hPa, 700 hPa, 600 hPa and 200 hPa. These levels were chosen in order to represent the atmosphere near the land surface and in the vicinity of the jet streams which characterise the summer circulation over northern Africa. They approximately correspond to altitudes of 0 - 80 m, 1.5 km, 3 km, 4 km and 12 km respectively. These values are based on the thickness values for the various model levels
given in Kalnay et al. (1996). Apart from 1000 hPa, the values are the middles of the ranges, given to the nearest 500 m. The three middle levels correspond to heights at which dust transport is often a maximum (Kalu, 1979; Prospero, 1981; Schütz et al., 1981).

The daily 2.5° latitude x 2.5° longitude reanalysis data were interpolated to a 5° latitude x 5° longitude grid commencing at 20° W and the Equator and terminating at 45° E and 35° N (see Results, Figure 7.3). Daily IDDI data were averaged onto a 5° latitude x 5° longitude grid coincident with the grid of the interpolated reanalysis data. A 5° latitude x 5° longitude grid was chosen to maximise the potential for interpreting variations while producing a field which could be easily visualised. All the daily fields for a given month were concatenated into a three-dimensional array representing the period 1984-93, containing some 300 daily geographical fields. Each 5° latitude x 5° longitude box was therefore characterised by a timeseries of some 300 daily IDDI and reanalysis values. Missing data in each IDDI series due to cloudiness or satellite errors were identified and removed from the IDDI series, along with the corresponding daily reanalysis values. For each box, if the number of data in the resulting shorter series was greater than one hundred, the paired IDDI-reanalysis series representing the box were correlated, and the correlation placed in a two-dimensional geographical field. If the series contained fewer than a hundred pairs of data, a missing data value was returned to the correlation field. The figure of one hundred data is arbitrary, representing a criterion that at least one third of the data in the series are present. It ensures that multiple years are represented by the correlations, without being so high that many grid-boxes are excluded from the analysis.

The statistical significance of each correlation was assessed using a Monté Carlo (MC) style randomisation procedure, based on two thousand randomisations. This method of significance testing was assessed alongside the random phase (RP) method of Ebisuzaki (1997), described in Chapter 3. The RP method is, as expected, a much more conservative way of determining significance (Chapter 3). The more liberal MC approach yields patterns of significant values with a wider geographical coverage and a higher degree of spatial coherence than the RP method. Therefore the MC method is more appropriate for visual interpretation of fields in which significant correlations are highlighted, particularly if the feature of primary interest is the broad distribution of relatively high correlations of a particular sign. The consequences of using the MC procedure, as opposed the RP method, are discussed further in the results section (Section 7.4.1., Figure 7.2), where distributions of significant results determined by the two methods are compared.

Correlations were performed at the 6-hourly intervals for which the reanalysis data are available: 12:00, 18:00, 00:00 and 06:00 hours. The IDDI data are derived from
measurements made between 11:35 and 12:00, or approximately at midday. The IDDI data may therefore be viewed as being temporally coincident with the midday (12:00 hours), temperature data. It would be desirable to perform correlations between contemporaneous IDDI and reanalysis data at all times. However, as IDDI data only exist for midday, analyses of reanalysis values at other times must be based on lagged relationships with IDDI values. Correlations with reanalysis data at 18:00 hours are by definition lagged, with IDDI leading temperature/vertical velocity by six hours. When performing correlations between reanalysis data at 00:00 and 06:00 hours and the IDDI, the data were lagged by one-day, so that the IDDI values preceded the corresponding reanalysis values by twelve and eighteen hours respectively.

The degree to which such lags affect the results of this study depends on the timescales of individual dust events and the response time of the atmosphere in reacting to the presence of dust. As the large dust events, which will produce the biggest signals in the temperature fields, last for the order of days, it is reasonable to assume that dust levels at 18:00, 00:00 and 06:00 hours might reflect those of the preceding midday. The transport of dust, however, may act to reduce the correlations at the later times; a particular volume of dust associated with a midday cooling of the surface, for example, may have travelled to a region represented by a different grid-cell by, say, midnight. The midnight temperature value for a given box is therefore not necessarily associated with dust representing the same box, but with dust that has now travelled to a different location. The temporal lag may therefore give rise to a non-systematic spatial lag in the correlation. However, McTainsh reports a timescale of some 48 hours for dust generated in the Bilma-Faya Largeau region to travel to northern Nigeria. This represents a north-south and an east-west translation of approximately one cell-width, suggesting that over the eighteen hour lagging period, such dust events may be viewed as being approximately stationary. Nonetheless, the potential for spatial lags to have an impact on the correlation values is borne in mind when interpreting the results, and the interpretation is biased towards the unlagged midday correlations.

7.3.2. IDDI relationships with the monsoonal air-flow

Six-hourly meridional wind data at 700, 850 and 1000hPa for April to October were extracted from the global NCEP v-wind dataset for the window 1.25° S -23.75° N; 20° W - 10° E. The data were not interpolated as the correlations in this analysis were not performed in situ; the reanalysis data therefore did not need to be spatially coincident, or represented on the same grid as, the IDDI data. The wind data therefore remained at 2.5° latitude x 2.5° longitude resolution, with the mid-point of the grid-squares in the most southerly row of the
extracted window occurring at 0º N. Zonally averaged meridional wind values at midday were created for each 2.5º latitude band by averaging the values for each grid square in a given row of extracted data, creating ten zonal indexes of the southerly component of the monsoonal flow for each pressure level. Absolute v-wind values, rather than magnitudes, were averaged, as the quantity of interest is the strength of the transport from south to north, associated with the prevailing monsoon flow in the months in question.

Daily IDDI data were spatially averaged to the same 5º latitude x 5º longitude grid as used for the temperature and vertical velocity analyses. For each individual month, the series of daily IDDI data were smoothed using a boxcar averaging function with a width of five days. The first and last two values in each smoothed monthly IDDI series were discarded, as were the first four values in the unsmoothed monthly wind series, resulting in a smoothed IDDI series and an unsmoothed wind series of equal length for each month. Each IDDI value then represented a five-day average, and the corresponding value in the wind series represented an instantaneous value at midday on the final day of the five day IDDI smoothing period. For each variable, the single-year monthly series were pooled over ten years (1984-1993) for each month. The resulting ten-year series were then correlated for each month to obtain a value representative of the relationship between local cumulative dust loadings and subsequent southerly winds over different latitudinal bands.

This approach was adopted in order to examine the short-timescale (approximately daily) response of the wind to dust loadings, while minimising the extent to which the correlations represented the response of dust levels to wind values. It is important to decouple these signals, representing causal relationships in opposite directions, as dust levels are ultimately a manifestation of strong winds at the land surface. Dust is unlikely to be correlated strongly with wind values as the relationship between wind strength and dust mobilisation is not linear, due to the requirement that the wind breaches a threshold velocity before mobilisation occurs. In addition, a strong dependence of the IDDI values on the wind values used here would indicate a strong coupling of the surface winds with winds at higher altitudes, which is not necessarily to be expected. The use of smoothed, lagged data, alongside these considerations, means that there is a high degree of confidence that the correlations reflect the impact of dust on the wind values.

In order to account for the in-built autocorrelation of the dust series, resulting from the smoothing process, the RP method was used to assess significance. This approach will also take account of any trends due to seasonal variations, which might otherwise result in artificially inflated correlation values.
7.4. Results I: IDDI-temperature correlations

Before individual monthly fields are discussed, the statistical distributions of all the correlation values for different seasons over the Sahel and the Sahara are analysed in a quantitative assessment of variations in the areally-averaged dust signatures with altitude and time of day. Such an analysis yields information concerning the bulk transport of dust and its broad impact on the temperature structure of the lower and middle troposphere.

In the monthly fields, coherent regions characterised by significant correlations are apparent for certain months at various pressure levels. Such areas are described throughout this section as constituting dust “signals” in the correlation fields. These signals are interpreted in terms of geographical, seasonal and vertical variations in the inferred impact of dust on the atmospheric temperature. The assessment of statistical significance is discussed, followed by a broad description of the variations in the dust signal in the temperature fields. Where appropriate the discussion associates elements of the correlation fields with independently determined dust activity and features of the atmospheric circulation. Example fields are shown for some months; fields representing every month from midday to 06:00 hours for all five pressure levels are presented in Appendix III. Because of the inherent lag in the correlations representing times after midday, the geographical and seasonal variations in the fields are described principally in terms of the midday values. Correlations over the following 18 hours tend to reflect those at midday; any notable differences are interpreted as diurnal variations.

7.4.1. Statistical distribution of correlations

Relationships between dust and temperature on regional spatial scales and seasonal timescales may be explored by plotting the distributions of the correlations against altitude (represented by pressure levels) in the form of box-plots for given regions and periods. Figure 7.2 shows such distributions over the Sahel for the seasons April-June and July-September. These seasons are the most interesting from the point of view of studies of Sahelian climate as they represent the transition months immediately before the wet season, and the wet season itself. Any large-scale impacts of dust on atmospheric temperatures in these periods would have potentially important implications for atmospheric stability and ultimately for the ability of the atmosphere to foster conditions conducive to rainfall generation.
It should be noted that the line joining the median values in Figure 7.2 displays an exaggerated curvature in some instances, particularly between 600 and 200 hPa. The degree of curvature is a characteristic of the spline function used to create the curve (an in-built feature of IDL) and has no physical significance.
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**Spring and summer Sahel distributions**

The midday Sahel profiles for AMJ and JAS (Figure 7.2) exhibit similar distributions of correlations at 1000 and 850 hPa. The median and upper quartile values at both levels are negative and of similar magnitudes, indicating that more than 75% of the correlations represent cooling associated with increased dust loadings, and that dust-related cooling of the lowest 1500 m of the atmosphere is widespread. The range of values in JAS is smaller than in AMJ; JAS exhibits lower median values and higher values of the lower quartiles. This indicates that fewer extreme dust-related cooling events occur in JAS, but that the overall impact of dust on the lower troposphere is greater in this period. More than 90% of the JAS correlations at 1000 and 850 hPa represent cooling, as indicated by the negative values of the upper quartiles.

At this stage, it is worth addressing the question of the significance of these cooling signals. The standard deviation, $\sigma$, of the Sahel correlation distributions representing midday and 6 a.m. at all levels in AMJ and JAS ranges from 0.06 to 0.16. The Sahel region comprises 13 grid-boxes in the E-W direction and 2 grid-boxes in the N-S direction. Therefore, each three-month period is represented by $13 \times 2 \times 3 = 72$ grid-boxes, or 72 correlation ($r$) values. Using the largest standard deviation value, the largest standard error in the correlations is $0.16/\sqrt{72} = 0.019$. Applying the t-test, the critical t-value for the 95% confidence level is approximately 2. Thus, a conservative estimate of the one-sample t confidence interval for the spring and summer correlations at all levels is $r \pm 0.038$. Applying this conservative confidence interval to the AMJ and JAS median values, we find that the cooling signals over the Sahel are significantly different from zero at the 5% level. They may therefore be interpreted as representing a real physical effect; indeed, the confidence intervals indicate that the degree of cooling represented by the distributions is large and highly significant.

Confidence intervals are calculated for all the sets of profiles. For each set of plots involving two seasons, two different times, all pressure levels and a given geographical region (i.e. the Sahara or Sahel), an interval (centred on zero) is marked, within which correlation values are not significantly different from zero. This interval is liberally defined and based on the highest standard deviation value occurring in the set of distributions.

A lower, but still significant, degree of cooling than at 1000 and 850 hPa is apparent at 700 hPa in both AMJ and JAS, with a small significant warming occurring at 600 hPa in JAS. In AMJ, the 600 hPa values are fairly evenly distributed about zero, whereas in JAS, almost 75% of these values are positive. The dust-related warming of the 600 hPa layer is therefore greater in the core wet-season months. At 700 hPa, the magnitudes of the upper quartiles are small, indicating that approximately three-quarters of the values represent dust-related
cooling. However, JAS exhibits a slightly lower median value than AMJ. These results, and those for 1000 and 850 hPa, suggest that the overall thermal impact of dust on the Sahelian atmosphere is greater in the summer than in the spring, although the latter period is more likely to experience extreme dust events which will temporarily result in large perturbations to atmospheric temperatures. The bulk of the transport over the Sahel in both periods takes place between 700 and 600 hPa, with cooling below, and warming above, the layer of transport.

During AMJ, the eastern and north-central Sahelian sources are active, and dust production is intense over the northern Sahel west of 10° E (Chapter 5, Figure 5.6). The situation is one in which dust events originate south of 20° N in an episodic manner, travelling in a southeasterly direction with the prevailing Harmattan circulation. In JAS, the mechanisms of mobilisation and transport are different in the Sahel, embedded as they are in the monsoonal air mass. Much of the dust will originate north of the Sahel, and will be transported in the SAL above the monsoon air. The confinement of dust above the discontinuity between the air masses will play an important role in determining the nature of the thermal impact on the atmosphere. For example, the pronounced layering of the dust is likely to lead to more well-defined regions of cooling and warming than if the dust was distributed throughout the atmosphere. Long residence times (of the order of days) and the preferential transport of small particles far from their sources will lead to a weaker, and less episodic, dust signal in the Sahelian temperature fields.

At 200 hPa there is a slight warming in AMJ, (significantly different from zero at midday) and a negligible deviation from zero in JAS. The near-zero values in JAS are to be expected, as coupling between dust in the middle troposphere and temperatures in the upper levels will be weak or non-existent. The mechanisms responsible for the bias towards positive values in AMJ are unclear at present.

At 6 a.m. the AMJ values at altitudes above 1000 hPa have changed very little since midday. A very slight cooling has occurred at 700 hPa; there is a similar magnitude shift in the opposite sense at 850 hPa. The median of the 1000 hPa correlation has increased by about 0.2, resulting in a change of sign from negative to positive. However, this value is not significantly different from zero, so the interpretation must be that the net cooling effect disappears at night, rather than being replaced by a net warming effect. The diurnal variation in JAS is very similar, although the increase in magnitude of the 1000 hPa median value is slightly less, and this value remains negative. The slight shift in the negative direction at 700 hPa is also apparent at 600 hPa and 200 hPa; the shift in the positive direction at 850 hPa is still present. The vertical profiles indicate that the expected dust-related night-time near-
surface warming is very weak over the Sahel in spring and non-existent in summer. However, the magnitude of the summer cooling is reduced at night. These effects are limited to the atmosphere near the surface, with the cooling signal remaining strong at and above 850 hPa.

Saharan distributions and the Sahel in autumn and winter

In contrast with the case over the Sahel, the Saharan distributions are almost invariant with level. The median values are near-zero for all levels at all times of day, suggesting that the overall impact of dust on temperature over the Sahara is much lower than over the Sahel. The only significant deviations from zero in the median values occur in AMJ, representing a small night-time warming at 700 and 600 hPa. The JAS distributions suggest a bias towards cooling at 850 hPa, but do not indicate any diurnal variation.

The Sahelian and Saharan distribution profiles for OND and JFM (Figure 7.3) are qualitatively very similar to those for AMJ and JAS. However, the degree of cooling of the lowest two levels over the Sahel is much greater in OND and JFM than in AMJ and JAS. The median values increase at 6 a.m. as for the spring and summer profiles, but remain strongly negative. The 6 a.m. values for OND and JFM are similar to the midday values for AMJ and JAS, indicating that night-time cooling in the dry-season is as strong as the daytime cooling in the onset and wet seasons.

Over the Sahel, the 600 hPa values are significant and positive in JFM at 6 a.m., and significant and negative at midday in OND. In the latter case, there is no net cooling at 700 hPa. One possible interpretation of this profile is that dust transport is a major feature at two ranges of altitudes: above 600 hPa and just below 700 hPa. This would result in two separate signals, most likely characterising different geographical regions within the Sahel. The aggregating of the correlations would then result in a composite signal including two different populations of correlation values. An alternative explanation is that the mechanisms of dust production which dominate in this period are associated with meteorological processes linked with cooling at 600 hPa, with the transport occurring below 700 hPa. Such mechanisms may include intrusions of cold air at 600 hPa which generate near-surface conditions conducive to deflation.
Summary

The box-plots in Figures 7.2 and 7.3 provide compelling evidence for the widespread thermal modification of the Sahelian atmosphere by mineral dust. The impact is generally one of cooling in the lowest 3 to 4 km of the atmosphere (i.e. below 600 hPa) and warming above these levels to about 5 km. Although this effect is greatest in the dry-season months, the signal is present in the monsoon onset and the wet-season months, indicating that dust may increase atmospheric stability and lead to conditions less conducive to the generation of the strong convective events associated with rainfall. The fact that the cooling signals at 1000 and 850 hPa are strongly significant adds further weight to this hypothesis. The relationship between dust production and convective disturbances inferred in Chapter 6 may well constitute a negative feedback, whereby convective disturbances mobilise dust which then acts to damp the convective activity. Such a hypothesised feedback process will be discussed further in Chapter 8.

The lack of a similarly dramatic effect of dust on the vertical profile of Saharan temperature distributions is striking. The largest signals occur in JFM, and represent cooling below 600 hPa, reduced at 1000 hPa at night. There is also some evidence of night-time cooling around 850 hPa in JAS and night-time warming at 700 hPa in AMJ. However, these effects are slight in comparison to those over the Sahel. The relative lack of a dust signal over the Sahara is most likely to be due to the presence of dust throughout the atmospheric column, unlike the more layered structure over the Sahel. Dust near to the Saharan land surface is likely to have a more complex effect on the atmosphere, with cooling being offset by greenhouse warming.

The above discussions have examined the large-scale general relationships between dust and temperature. They have indicated that IDDI-temperature correlations can reveal information concerning altitudes of dust transport. The following sections examine individual monthly correlation fields in order to determine the degree of spatial and temporal variability in the relationships between dust and atmospheric temperatures, before examining direct impacts of dust on the vertical velocity fields.
Figure 7.3: Distribution of all correlations over the Sahara (top) and Sahel (bottom) in OND and JFM for midday and 6 a.m. Values outside the dotted lines are statistically significant at the 5% level.
7.4.2. Interpretation of significance of individual correlations

Figure 7.4 shows the correlations between local IDDI and temperature values at midday for January at 850 hPa. Values which tested as statistically significant at the 5% level in both the MC and RP tests are distinguished from those which tested as significant with only one of the tests or which were not significant at this level according to either test. This field is chosen for the comparison of the MC and RP methods as it contains a high number of large-magnitude correlations, particularly over the Sahel. All the RP-significant values, and most of the MC-significant values, occur between 5° and 20° N., with the highest frequency occurring in the 10° - 15° N. band. There are eighteen RP-significant correlations, the smallest of which has a magnitude of \(|r| = 0.21\), and thirty four MC-significant correlations, the smallest of which has a magnitude of \(|r| = 0.12\). Therefore in this case, using the MC method results in about twice as many statistically significant values as using the RP method, with the smallest magnitude in the former case being just over half that in the latter.

Figure 7.4: Midday in situ IDDI-temperature correlations (x100) for January at 850 hPa, for the period 1984-93. Values marked with an asterisk are significant at the 5% level when tested for significance using a simple Monté Carlo procedure; values in bold are significant at the 5% level when tested using the Random Phase method of Ebisuzaki (1997).

In the case of the IDDI-temperature correlations, the RP method is much more conservative than the MC method. This is not surprising, as the results produced using these two methods diverge as the autocorrelation in the data increases. Dust events often last for several days, and, under average conditions for any given month or part thereof, temperatures at specific times of day will not vary greatly with time over the regions in question. Both these factors will tend to inflate the autocorrelation values in the daily timeseries.
The RP-significant results may be viewed as representing a strong likelihood that there is a physical relationship between the data under examination, while the MC-only significant values represent a lower, but not negligible, likelihood of a physical relationship. However, the fact that the “weakly significant” MC-only values are associated with the “strongly significant” RP values means that there is some justification in viewing both sets of values as part of a coherent pattern of correlations representing a broad physical link between the variables in question. This consideration, coupled with the greater ease of interpretation of the liberal MC values, means that interpretation of the fields on the basis of the MC significance test is justifiable.

7.4.3. Vertical distribution of correlations

At 1000 and 850 hPa, statistically significant correlations tend to be negative and form large coherent patterns covering much of the Sahel and extending into the South Sahara in some months (Figure 7.5). Such patterns suggest a large-scale association of dust aerosols with a cooling of the lower troposphere south of about 25° N. North of 25° N, correlations are generally not significant and often positive, reflecting the signals apparent in the Saharan vertical profiles (Figure 7.2 and 7.3). The geographical and seasonal variations in the correlation patterns at these and other levels are discussed in more detail in Section 7.4.5.

Figure 7.5: Midday in situ IDDI-temperature correlations (x100) at 1000 hPa for February. Values in bold are statistically significant at the 95 per cent confidence level when tested using a Monté Carlo procedure.

At 700 and 600 hPa, the frequency of significant correlations decreases notably, and coherent dust signals in the temperature fields are more localised. The 700 hPa values are predominantly negative, while positive values predominate over many regions in the 600 hPa fields. The 600 hPa signals reflect those at 700 hPa to a limited extent in some regions,
particularly in the winter months. These cases represent dust above 600 hPa, exerting a similar cooling influence at both 600 and 700 hPa.

The north-south division between regions of positive and negative correlations is less apparent at 600 hPa; positive values are common, and often occur as coherent signals over Sahelian regions (Figure 7.6), representing a warming of the 600 hPa level by dust at lower levels.

Figure 7.6: Midday in situ IDDI-temperature correlations (x100) at 600 hPa for February. Key as for Figure 7.5.

Significant correlations are fewer in number at 200 hPa, and those that do occur tend not to form coherent patterns. Exceptions are the Guinea Coast region in October, which is characterised by positive correlations, and the central Sahel in November, where correlations are negative.

The vertical distribution of significant correlations supports the notion that dust typically extends to an altitude of between 700 and 600 hPa during most of the year, and cools the underlying atmosphere by reducing the amount of incoming solar radiation at the ground surface. Within and just below the dust layer the cooling will be offset to a certain extent by re-radiation in the infra-red from the layer itself, accounting for the lower number of significant negative values at 700 hPa than at the two lowest levels. The more localised signals at 700 hPa are most likely to be due to regional variations in the height of the base of the dust layer, which may be closely related to the height of the base of the trade wind temperature inversion (Kalu, 1979). Such relationships will be discussed in more detail in the following sections concerning regional and seasonal variations.

The dust layer will act as an emitting surface for long-wave radiation and will therefore warm the atmosphere above it in much the same way as the ground-surface acts to warm the
whole atmosphere. This explanation accounts for the increased number of positive correlations at 600 hPa; the presence of dust at or below 600 hPa will act to warm the atmosphere at this level via an enhanced greenhouse effect (Andreae, 1996; Tegen et al., 1996; N’Tchayi et al., 1997). However, the relatively low frequency of significant correlations, when compared with the 1000 and 850 hPa fields, suggests that the warming of the atmosphere above the dust layer is much weaker than the near-surface cooling. It is likely that such warming will be confined to levels near to or above the top of the dust layer, where the cooling effect due to a reduction in solar radiation at ground-level is low or non-existent.

Although the correlations at 200 hPa do not form large coherent signals, and may be either positive or negative, there is some geographical grouping together of values of the same sign, suggesting that the fields at this level are not simply the consequence of random, physically meaningless statistical noise. The much reduced frequency of significant values in most regions at 200 hPa when compared with the lower levels is as expected, indicating a far weaker link between temperature at this level with dust loadings, which will be confined to lower altitudes. Nonetheless, the suggested association between conditions at the top of the troposphere and dust levels in the middle and lower troposphere must be explained.

It is possible that dust-induced warming at lower levels (e.g. 600 hPa) may extend to 200 hPa in some cases, resulting in positive correlations, while under certain conditions the dust layer may act to trap heat in the lower layers of the atmosphere, reducing the radiative heating of the upper levels. However, an alternative explanation is that the causal mechanism operates in the opposite sense, with particular temperature anomalies at 200 hPa being associated with atmospheric motion which ultimately impinges on the lower troposphere and is coupled with dust production. Specific examples of such cases are discussed in Section 7.4.5.

7.4.4. Diurnal variation of correlations

The weak tendency for the correlations at 1000 hPa to become less negative with time (i.e. from midday to 06:00) throughout the Sahel, as noted in Section 7.4.1, is most noticeable in May, June and October. During these months a switch from negative correlations at midday to positive correlations at 06:00 occurs over parts of the Sahel. The effect is most pronounced in the western Sahel in May (although the negative midday values in this region are low) and over the South Sahel and Guinea Coast in October (Figure 7.7). These situations are not typical; during most months large coherent negative signals at night-time are common at 1000 hPa.
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Figure 7.7: In situ IDDI-temperature correlations (x100) at 1000 hPa for October at midday (top) and 6 a.m. (bottom). Key as for Figure 7.5.

At 850 hPa strong negative correlation signals are maintained over much of the Sahel at all times, suggesting that dust-induced cooling persists throughout the 24-hour period. A similar phenomenon is observed for most months at 700, 600 and 200 hPa. Exceptions occur in January at 600 hPa, when the number of significant positive correlations increases at 06:00, and in August at 700 hPa, when the number of negative values increases after midday. In neither case is there a change in the sign of the correlations over the areas where increases in the magnitude and/or extent of the signals occur. These isolated exceptions may be a result of the inherent time-lag in the correlations after midday. Signals may be stronger later in the day due to the time-integrated effects of dust on the radiation flux through the atmosphere. Alternatively, relatively low dust levels at midday (with a relatively weak temperature signature) may be a precursor of enhanced dust activity later in the day, perhaps related to strong convection in the afternoon, which will result in a stronger dust signal in the temperature field. In this case, the midday dust series would effectively be a proxy for a later
series in which variations in dust levels are more pronounced, as are the associated temperature anomalies.

These results suggest that the predicted dust-related warming of the near-surface layers at night does occur locally in some cases, but that the night-time warming effect is much less pronounced than the daytime cooling. The night-time warming effect appears to be confined to the lowest level under investigation, which represents an altitude of some 100m. Even at this level the warming signal is generally weak, suggesting that this phenomenon is restricted to altitudes of the order of tens of metres above the ground-surface.

7.4.5. Geographical and seasonal variations and altitudes of transport

The strong cooling signals manifest at 1000 and 850 hPa in the vertical profiles are most widespread in winter (December – February), when the Sahelian dust sources are most active (Chapter 5). They also characterise November, when dust loadings are generally low. Under such conditions, any relatively large dust events are likely to result in significant perturbations of the atmospheric temperature structure.

Figure 7.8: Midday in situ IDDI-temperature correlations (x100) at 1000 hPa for April. Key as for Figure 7.5.

In March and April (Figure 7.8) the signal at these levels is strongest in the east, migrating towards the central Sahel in May and then to western regions over the period June to October (Figures 7.7 and 7.9), although the strongest signal in August is in the central Sahel. This migration of the correlation signal reflects the migration of the regions of highest dust loadings, as described in Chapter 5. The westward migration during the wet season is consistent with the maximum in dust production over the western North Sahel and Saharan regions. These large cooling signals at the two lowest levels are indicative of the presence of dust above the 850 hPa level, which results in cooling of the underlying atmosphere due to
the reduction in the incident solar radiation at the surface, and hence in the outgoing
longwave radiation which heats the troposphere.

Strong 1000 hPa negative signals also occur in the South Sahara (between 20° and 25° N) in
the winter months (e.g. Figure 7.5). In December, correlations over this region at 850 hPa are
positive, albeit generally not significant south of 25° N. This indicates the presence of dust
between 1000 and 850 hPa, cooling the former level and warming the latter level as a result
of re-radiation in the infra-red from the dust layer. Where the correlations are positive and
small, it is likely that cooling is offset by radiative warming, indicating that the level under
examination lies just below or within the layer.

Differences between the pressure levels in the patterns of warming and cooling may be used
to infer the altitude of dust transport, as described above for the South Sahara at the lowest
two levels. Such an approach may be used to describe the seasonal and geographical
variation of dust transport over northern Africa.

\[\text{Figure 7.9: Midday in situ IDDI-temperature correlations (x100) at 1000 hPa for June.}
\text{Key as for Figure 7.5.}\]

**Thermal impact and transport south of 25° N**

For most of the winter, the widespread negative signal at 1000 and 850 hPa extends
throughout the Sahel, South Sahara into the Sudanian region (5° - 10° N). At 700 hPa the
negative signal is largely confined to eastern regions. At 600 hPa, some negative values
occur in the east, with positive correlations over the central Sahel and Sahara in February and
March (Figure 7.6). These distributions indicate dust transport above 850 hPa throughout the
winter, with transport occurring above 700 hPa, and in a few places above 600 hPa, in
eastern regions. Over the central Sahel and Sahara transport occurs just below 600 hPa in
February, when it results in a warming of the atmosphere at this level. (Figure 7.6
demonstrates the geographical variation in the height of transport during February, with negative values indicating dust above 600 hPa in parts of the eastern Sahara and Sahel.) The lack of significant correlations in the central regions at 700 hPa suggests that dust exists at or near to the 700 hPa level, resulting in no net heating or cooling (Section 7.4.3.). In the other winter months, the strong negative correlations at the two lowest levels indicate the presence of dust, which must be present at or in the vicinity of 700 and 600 hPa in order to explain the lack of a signal at these two levels. A similar situation to that apparent in February occurs in March. The positive 600 hPa values in February terminate further south than in March, consistent with the more southerly February dust sources.

Significant positive correlations at 600 hPa are widespread over the Sahel and Sahara in May. The lack of any notable signal at 700 hPa suggests transport at or around the 700 hPa layer. The positive signal diminishes through June and July as the Sahel comes under the influence of the summer monsoonal air mass and as the Saharan air layer (SAL), which transports dust out of Africa over the Atlantic, migrates northwards. The frequency of positive values at this level increases in August (Figure 7.10), and negative values at 700 hPa indicate that the dust is present between these two levels (Figure 7.10).

A positive 600 hPa signal is also apparent over the western Sahel in September. This signal diminishes in October, when a widespread negative signal is apparent south of 15° N (Figure 7.11). The October signal indicates dust transport below 600 hPa over the northwestern Sahel, rising to above 600 hPa over the South Sahel and Sudan-Guinea Coast region. At this time of year the monsoon air-mass is relatively far south, and the Sahel experiences the last phase of the wet season. It is likely that the negative correlations in October represent dust lying over the monsoon air-mass. The correlation distributions in the subsequent two months suggest that similar altitudes of transport are maintained in November, with transport above 600 hPa over the eastern Sahel and western Sahara established in December.
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Figure 7.10: Midday in situ IDDI-temperature correlations (x100) at 600 hPa (top) and 700 hPa (bottom) for August. Key as for Figure 7.5.

Figure 7.11: Midday in situ IDDI-temperature correlations (x100) at 600 hPa for October. Key as for Figure 7.5.
Figure 7.12: Midday in situ IDDI-temperature correlations (x100) at 600 hPa for March. Key as for Figure 7.5.

Regional variations in transport south of 25°N

It may be inferred from the vertical distributions of positive and negative correlation signals that dust transport takes place above 600 hPa over the Guinea Coast in March, October and November (Figures 7.11 and 7.12). During October this high-altitude transport extends throughout the region between 5° and 15° N. Similar transport occurs over the southwestern Sahara in December, and over the southeastern Sahel in January. Dust above 600 hPa is present over Egypt and Libya and parts of Sudan and the Red Sea region in February (Figure 7.6).

By the same token, transport occurs between 700 and 850 hPa over the southeastern Sahara in May, with the dust probably being closer to the 850 hPa level. This is consistent with the 850-900 hPa transport described by Kalu (1979). This is also the case in the western Sahel-Saharan in January. A similar situation is observed in June over the western and eastern South Sahara and the eastern Sahel, although the lack of negative values at 850 and 1000 hPa indicate the possibility that dust extends to the surface. In March, transport is between 700 and 850 hPa over the coast from Senegal to the Western Sahara, with positive and negative signals at these levels respectively. The same situation arises in October and November, but with the latitudes of transport displaced some 5° to the south. These results are consistent with those of Alpert et al. (1998), who found evidence that dust heats the atmosphere over the eastern tropical North Atlantic between 1.5 and 3 km.

In December, similar reasoning places a dust layer between 700 and 850 hPa over the North Sahel, at around 700 hPa in the west and central South Sahel, and above 700 hPa in the eastern South Sahel. This interpretation is complicated by the existence of a negative signal
at 600 hPa over the western South Sahara and (weakly) over the western North Sahel, where lower-level patterns would suggest the existence of a positive or negligible signal. It is possible that this situation arises due to the existence of two dust layers at different altitudes, arising from different sources.

**Features of the 200 hPa correlation fields**

The most notable features in the 200 hPa fields are a region of significant positive correlations over the southwestern Sahel and Guinea Coast in October (Figure 7.13), and a negative signal over the central North Sahel in November (Figure 7.14). There are no corresponding features unique to these months in the lower-altitude fields, although at both 1000 and 850 hPa the corresponding regions exhibit significant negative correlations.

**Figure 7.13: Midday in situ IDDI-temperature correlations (x100) at 200 hPa for October. Key as for Figure 7.5.**

**Figure 7.14: Midday in situ IDDI-temperature correlations (x100) at 200 hPa for November. Key as for Figure 7.5.**
It is plausible that the October and November 200 hPa signals represent a causal relationship in the opposite direction to those discussed so far. For example, interactions between the tropical and extra-tropical air-masses in the vicinity of the tropopause may have an impact on lower levels of the troposphere. An intrusion of cold air from mid-latitudes at 200 hPa would lead to high-level subsidence which would result in anticyclonic activity near 850 hPa, and low-level divergence (Pye, 1987). This situation is associated with strong surface winds and dust production. The resulting relationship between the IDDI and the 200 hPa temperature would therefore be represented by a negative correlation, as is seen over the North Sahel in November. Pye describes such a situation occurring on 9 February 1974.

The above scenario offers an explanation for the November signal, but not for that described for October. Although the positive correlations do not tend to form coherent patterns (October notwithstanding), they are more frequent than the negative values. The mechanisms which might link warming of the upper troposphere with dust levels in the lower and middle troposphere at present unknown.

**The role of dust north of 25° N**

Significant correlations are much fewer in number north of 25° N, and produce fewer coherent signals than values south of 25° N. The proportion of positive values north of 25° N is high at all times of day, suggesting that the role of dust in determining the radiative structure of the troposphere over the northern Sahara is different from that over the Sahel and southern Sahara. However, there are exceptions; at 1000 hPa, negative signals extend throughout the northern Sahara in January and February (Figure 7.5) and, to a lesser extent, in September. Negative values are widespread in April and September at 850 hPa in the eastern Sahara (Figure 7.15).

![Figure 7.15](image_url) **Figure 7.15:** Midday in situ IDDI-temperature correlations (x100) at 850 hPa for September. Key as for Figure 7.5.
Coherent positive Saharan signals occur at low levels in June, August and October at 1000 hPa, and in December at 850 hPa. These signals are strongest west of about 15° E (Figure 7.16)

The differences between the Sahel and Sahara are not wholly unexpected, as the character, distribution and, most importantly, the mechanisms of transport of the dust vary with latitude. Dust originating in the Sahara will be distributed throughout the lower layers of the atmosphere and extend to ground level near its source. Some of this dust will be transported to the south and west in the SAL. Daytime warming of the atmosphere due to re-radiation in the infra-red from dust aerosols will occur in the immediate vicinity of the dust layer. Such a warming may therefore occur near the ground in the vicinity of the source regions, but will be restricted to higher altitudes where dust is transported aloft over large distances, particularly where the SAL is undercut by the monsoonal air mass.

Figure 7.16: Midday in situ IDDI-temperature correlations (x100) at 1000 hPa for June. Key as for Figure 7.5.

7.4.6. Summary of IDDI-temperature relationships

The IDDI-temperature correlations indicate that dust has a distinct, and large scale, impact on the thermal structure of the lower and middle troposphere. This impact generally consists of a cooling of the lower levels and a warming of the mid-levels. Such a reduction in the vertical temperature gradient is most apparent over the region from 5° to 25° N, with the Sahel comprising the area most widely and dramatically affected. Although the effect is most pronounced in winter, both the statistical distributions of all the correlations and the spatial distributions of the statistically significant correlations indicate that it also occurs in spring and summer, with a consequent potential to modify rain-bearing convective events.
The distribution of significant correlations reinforces the conclusions drawn from the box-plots of all the correlation values, indicating that even small values of the statistical quantities plotted (median, upper quartile etc.) are likely to have a physical significance. The correlation fields demonstrate a distinct variation in the sign and geographical distribution of the dust signal with altitude, and a well-developed seasonality in the spatial extent of dust-induced heating and cooling of the atmosphere at different pressure levels. This is likely to reflect geographical and seasonal variations in the altitude of maximum dust transport, which is in turn dependent on the height and extent of features such as the trade inversion and the activity of the various source regions.

7.5. Results II: IDDI-vertical motion correlations

Results analogous to those representing the relationship between IDDI values and atmospheric temperatures (Section 7.4) are given in Appendix IV for correlations of IDDI with atmospheric vertical motion, at the same pressure levels. Vertical motion is defined as the change in pressure with time (dp/dt) at a given location, and has units of Pa/s. Negative values are indicative of rising motion, and positive values of sinking motion, in the atmosphere. Therefore the more negative the values are, the more intense is the convection they represent. An association of enhanced dust levels with reduced convective activity is therefore represented by positive correlations between vertical velocity (VV) and IDDI. The vertical motion-IDDI correlation fields yield much less information than the temperature-IDDI correlation fields, with signals being much weaker and less coherent in the former than in the latter. For this reason, the discussion of these results is much briefer than for the IDDI-temperature analysis.

7.5.1. Statistical distribution of correlations

Box-plots of the distributions of IDDI-VV correlations versus pressure level (Figures 7.17 and 7.18) reveal an apparently less dramatic impact of dust on VV than on temperature (Figures 7.3 and 7.4). It should be recalled that the VV data fall in class B of the NCEP/NCAR data description, meaning that they are less reliable than the temperature data and less likely to capture any changes to the atmospheric dynamics related to the presence of dust. This is so even over regions where observational data are numerous, as the VV is determined to a large extent part by the model parameterisations, which do not include dust. Nonetheless, certain features of the box-plots are worthy of note, and complement the findings of the previous section.
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Spring and summer Sahel distributions

In AMJ, the Sahelian midday median values are uniformly close to zero and not significant (Figure 7.17). However, more vertical structure becomes apparent at 6 a.m., when the 1000 hPa values exhibit a bias towards enhanced upwards motion (negative values), while those at 700 and 600 hPa indicate reduced upwards motion (positive values). These results are consistent with the IDDI-temperature correlation box-plots; AMJ is the only season in which the median value indicates a dominance of night-time warming near the surface, which will decrease atmospheric stability. The temperature values suggest that the dominant altitude of dust transport in AMJ is near 600 hPa. The increased stability at 700 hPa is consistent with the cooling of this layer; the slightly lower increase in stability at 600 hPa, despite a tendency towards increased temperatures, is a little more complex. It is expected that a locally enhanced greenhouse effect will raise atmospheric temperatures within the dust layer at night, strengthening the temperature inversion which exists over parts of the Sahel. This will increase the stability of the atmosphere in general, reducing the potential for air from lower altitudes to penetrate the base of the inversion.

In JAS, atmospheric stability is increased at 1000 and 850 hPa, and decreased at 600 hPa. This is consistent with the cooling associated with dust below 600 hPa and the warming at 600 hPa apparent in Figure 7.3. At 6 a.m., the median values are closer to zero than at midday, although increased stability is still evident at 850 hPa, and the 700 hPa median has changed from very slightly negative and not significant to positive and bordering on significance. The 700 hPa distributions suggest that the dynamical impact of dust is more complex at this level, which is generally in the vicinity of the dust layer.

Autumn and winter Sahel distributions

The association between dust and changes in atmospheric stability is more pronounced in the autumn and winter, particularly at night. At midday, stability is increased below 700 hPa and decreased at 600 hPa in OND, and increased at 850 hPa in JFM (Figure 7.18). This interpretation is based on the significant median values, which are, nonetheless, small. Their magnitude, and the remaining near-zero values indicate that the relationship between dust and vertical motion over large areas at midday is weak at some levels. The largest effects occur at 850 hPa. It should be remembered, however, that the VV data are less likely to capture the effects of dust than the temperature data, due to the greater influence of the model in generating the former.
Figure 7.17: Distribution of all correlations between IDDI and vertical velocity over the Sahara (top) and Sahel (bottom) in AMJ and JAS for midday and 6 a.m. Values outside the dotted lines are statistically significant at the 5 % level.
Figure 7.18: Distribution of all correlations between IDDI and vertical velocity over the Sahara (top) and Sahel (bottom) in OND and JFM for midday and 6 a.m. Values outside the dotted lines are statistically significant at the 5 % level.
At 6 a.m., atmospheric stability is increased at 600, 700 and 850 hPa in both seasons, with non-significant median values occurring at 1000 hPa (Figure 7.18). In both cases the maximum in suppression of vertical motion occurs at 850 hPa. The effect of dust on the Sahelian dry-season night-time atmosphere is therefore one of large-scale stability enhancement via a cooling below 700 hPa (OND) or 600 hPa (JFM) (Figure 7.3). The fact that the reduction in upwards motion extends above and below the inferred altitudes of dust transport again suggests a general stabilisation of the entire atmospheric column resulting from a strengthening of the temperature inversion.

**Saharan distributions**

In all seasons, the association between dust and VV over the Sahara is weak. There is some evidence for an increase in near-surface stability in AMJ (Figure 7.17) and at midday in JFM (Figure 7.18), when median values are small and positive. In AMJ the bias towards positive values increases at 6 a.m., when nearly 75% of the correlations are greater than zero. A bias towards negative values occurs in AMJ at 700 and 600 hPa. A stronger association with decreased atmospheric stability and increased rising motion occurs at 850 hPa in OND (at midday and 6 a.m.) and JFM (at 6 a.m. only). In these three cases, some 75% of the correlations are negative, with the largest deviations from zero occurring at 6 a.m. in JFM. The sign of the Saharan dust signal in the VV distributions in these cases is opposite to that of the Sahelian signal. As is the case with the temperature analysis, these results suggest that dust is distributed throughout the lower layers of the Saharan troposphere, where they give rise to an enhanced greenhouse effect. Below the temperature inversion, this heating leads to increased convective activity, particularly at 850 hPa, or around 1500 m. This is again consistent with the findings of Alpert et al. (1998) of dust-induced atmospheric heating at this level off the West African coast.

**7.5.2. 1000 hPa fields**

At 1000 hPa, the most coherent patterns of significant correlations occur from December to January, although the spatial coherence of the significant correlations is much less than in the case of the IDDI-temperature correlation fields. In December the signal extends throughout the Sahel and into the South Sahara. In January and February the signal extends into the Sudan region (5° - 10° N), east of 10° E. The correlations are negative in the most easterly regions, with the negative signal extending further west in the North Sahel than in the South Sahel, and positive elsewhere (Figure 7.19). These patterns suggest that dust loadings are associated with increased subsidence in winter, over most of the region in which dust is
linked with widespread reductions in temperature (Figure 7.5). The signal in the vertical motion fields is, however, considerably weaker than that characterising the temperature fields, and the negative correlations in the east indicate that the relationship between dust-related temperature and vertical motion anomalies is not geographically invariant.

Small regions within the central Sahel exhibit a residual winter signal in November and March. Throughout the rest of the year, no coherent signals are apparent; significant correlations are few and occur in isolation. Most of these values are positive, with negative values tending to be confined to the southeastern parts of the study area. The lack of widespread significant values means that it is difficult to determine any strong geographical trends from these data.

7.5.3. 850 hPa fields

Dust signals in the 850 hPa field are much more apparent than at 1000 hPa. The signal described for the winter months at 1000 hPa is evident to varying degrees in most months, with some geographical variation. From November to March significant positive correlations characterise much of the Sahel, indicating an association of dust with increased subsidence. The signal is confined to central regions in November and February (Figure 7.20), and extends from 25° - 30° E to the West African coast in December and January. Throughout this period, negative values occur in the east, most commonly in the North Sahel or South Sahara. The strength and extent of the signal tend to remain constant or increase over the course of the day. For example, the easterly negative signal at midday for February extends across much of the South Sahara at midnight and 6 a.m., indicating reduced subsidence in

Figure 7.19: In situ February IDDI-vertical velocity correlations (x100) at 1000 hPa for midday. Key as for Figure 7.5. c.f. Figure 7.6 for spatial relationship between dust-related cooling and subsidence signals.
this region (Figure 7.20). The same region exhibits a persistent cooling in the 850 hPa temperature fields. Such a relationship between changes in the temperature and VV fields is at first sight counter-intuitive. Changes in the vertical motion of the atmosphere may arise from larger-scale changes in the atmospheric overturning which override the local effects of heating or cooling due to the presence of dust. An alternative explanation is that the reanalysis data do not accurately represent the climatological situation due to lack of data and deficiencies in the model with which they were generated.

In March and April the geographical extent of the positive Sahelian signal diminishes, while a similar positive signal exists over the western Sahel near the coast. In March a negative signal is evident over parts of the Guinea Coast region. In May, significant negative values persist throughout the day over the eastern Sahel and Sahara and the Red Sea region.

![Figure 7.20: In situ February IDDI-vertical velocity correlations (x100) at 850 hPa for midday (top) and midnight (bottom). Key as for Figure 7.5.](image-url)
In June, a positive Sahelian signal is evident west of 10° E., with a negative signal prevailing east of 20° E. A negative signal exists over the Gulf of Guinea at all times of day. At nighttime, this negative signal is complemented by a positive signal over the South Sahel west of 0°. The strongest such dipolar pattern occurs at midnight. This pattern is indicative of enhanced convection over the Guinea Coast, with reduced convection immediately to the north. This will tend to reduce the northwards flow of air over this region, as the tendency for air to sink inland and rise over coastal regions is enhanced. This situation is most likely to occur at night, when the land is coolest. If such a “sea-breeze” effect does operate, the implication is that dust might reinforce it, leading to enhanced convergence over the Guinea Coast. A similar, but weaker, influence is apparent at daytime, when convergence will be stronger inland. The dust signals in the VV fields for June suggest that such inland convergence may be reduced by the presence of dust, with possible implications for the strength of the southerly monsoon flow in June, which represents the early part of the wet-season, and may be viewed as an onset month for some years.

In July, daytime correlations are very infrequent and isolated. However, localised positive signals are apparent at midnight and 6 a.m. in the southwestern Sahel. The strongest signal occurs over the southwestern part of West Africa near the coast at 6 a.m. (Figure 7.21), indicating reduced upwards motion. Scattered significant values are apparent in August, mostly throughout the Sahel, with the most widespread signal occurring at midnight (Figure 7.22). This consists of reduced upwards motion over most of the South Sahel west of 10° E. Similar values extend into the central and northeastern Sahel and the Guinea Coast regions. Negative values occur in the North Sahel and North Sahara between 5° W and 5° E. Correlations with temperature in these regions of enhanced upward motion are positive, indicating warming, but are not statistically significant. Over the regions of increased subsidence, the IDDI-temperature correlations are generally negative, and significant in places.

Coherent positive signals occur again over the South Sahel in September, after midday, and strengthening after 6 p.m. The signal is strongest to the north of the Guinea Coast, over which correlations are negative, and significant in places. This negative signal over the Guinea Coast is stronger in October at 6 p.m. (Figure 7.23) and midnight, but absent at midday and 6 a.m. A positive signal over the coastal areas of the South Sahel exists at midnight and 6 a.m. In September and October, significant negative correlations exist east of 10° E., with the most coherent and persistent signals occurring between 10° and 30° E in the South Sahel.
Figure 7.21: 6 a.m. *in situ* IDDI-vertical velocity correlations (x100) at 850 hPa for July. Key as for Figure 7.5.

Figure 7.22: Midnight *in situ* IDDI-vertical velocity correlations (x100) at 850 hPa for August. Key as for Figure 7.5.

Figure 7.23: Midnight *in situ* IDDI-vertical velocity correlations (x100) at 850 hPa for October. Key as for Figure 7.5.
7.5.4. 700 hPa fields

Widespread areas within the Sahel in which dust loadings are associated with increased subsidence are apparent in the 700 hPa fields from November to April, reflecting the 850 and 1000 hPa fields to a certain extent (Figure 7.24). However, at 700 hPa these signals are generally only apparent at midnight and 6 a.m., apart from in the case of February, and are confined to central and eastern Sahelian regions except in the case of January. The positive correlations associated with this situation tend to occur in the eastern regions which are characterised by negative values at the lower levels. This suggests that the lower levels are responding to near-surface dust-induced warming, while dust reduces instability at 700 hPa. The latter effect may be the result of a strengthening of a temperature inversion above 850 hPa by dust-induced warming. From September to March there is a tendency for negative values to occur over the far southwestern Sahel and/or the Gulf of Guinea, indicating dust transport and warming at or below 700 hPa.

Figure 7.24: 6 a.m. in situ IDDI-vertical velocity correlations (x100) at 700 hPa for April. Key as for Figure 7.5.

In May, significant values are mostly negative and confined to the central Sahel or the eastern regions discussed above. These negative values persist in June, when a negative signal is again apparent over the Gulf of Guinea. At midnight this signal is associated with positive values immediately to the north to produce the dipolar pattern seen in June at 850 hPa. From July to August the 700 hPa fields tend to reflect the 850 hPa fields. In July, enhanced subsidence at 6 a.m. is apparent throughout West Africa from Mauritania to Côte d’Ivoire. In August and September, negative values exist over the South Sahel at midday; after midday, negative values occur over the North Sahel, with positive values over the South Sahel. These variations suggest that dust over the South Sahel at 700 hPa increases instability during the day via local atmospheric warming, and reduces instability at night due
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to the strengthening of the temperature inversion above the boundary between the monsoon air and the warm, dry Saharan air. Over the North Sahel this boundary will be at a lower altitude due to the wedge-like shape of the monsoon airmass, and increased instability will result from dust-induced warming above the inversion.

7.5.5. 600 hPa fields

Positive signals indicative of enhanced subsidence are apparent from December to April in the same locations as at 700 hPa, but are less coherent, frequent and widespread in the 600 hPa fields. The strongest such signals occur in April.

Negative signals over the Gulf of Guinea are more common than at 700 hPa, occurring in various fields from September to March and, to a lesser extent, in June and July. The dipolar pattern, with a positive signal immediately to the north, is not apparent. In September and October, association of dust with increased convection at the 600 hPa level extends from the Guinea Coast into the North Sahel (Figure 7.25).

Figure 7.25: 6 p.m. in situ IDDI-vertical velocity correlations (x100) at 600 hPa for September. Key as for Figure 7.5.

7.5.6. 200 hPa fields

The 200 hPa VV fields tend to exhibit a greater number of coherent signals than the 200 hPa temperature fields. These signals tend to be negative, indicating an association of dust with increased upward motion at this level. However, reasonably coherent positive signals are apparent in January, February and September. These signals occur over the southeastern (sub) Sahel, west-central Sahel and the eastern Sahel-Sahara respectively, at various times after midday. A weak signal occurs over the central northern Sahara at midday in November. Less coherent positive patterns and isolated significant positive correlations also occur
throughout the year.

The strongest negative signals occur between 15° and 25° N. in eastern regions in April (at midday and 6 p.m.) and in the western Sahel in September and October (at 6 p.m.) (Figure 7.26). Significant negative correlations are also widespread throughout the Sahel in August (at 6 p.m.), and in the region 10° -25° N; 30°- 40° E in July at 6 a.m. They also occur relatively frequently in various Sahelian regions in May and June. The implications of the widespread occurrence of statistically significant values at 200 hPa is discussed in the next section.

Figure 7.26: 6 p.m. in situ IDDI-vertical velocity correlations (x100) at 200 hPa for April. Key as for Figure 7.5.

7.5.7. Direction of the causal relationship between dust and vertical velocity

The relationship between dust loadings and anomalies in fields of VV is likely to be more complex than that between dust loadings and temperature anomalies. Temperature has no direct effect on dust transport, although temperature will be influenced by processes that affect the transport of atmospheric aerosols. In contrast, VV will influence vertical dust transport, but may also be affected as a result of the thermal impact of dust. A reduction in temperature will affect the buoyancy of the local or regional atmosphere, and reduced buoyancy will lead to a decrease in upward motion. However, increased upward motion will facilitate the vertical transport of dust aerosols. It is therefore entirely plausible that enhanced convective activity might be associated with increased dust loadings.

Such a dichotomy between the different types of interaction between dust and the vertical motion field leads to a higher degree of complexity in the VV results than in the temperature results. Under certain conditions, transport of dust to middle/high altitudes by strong
convection might be followed by suppression of convection by the resulting dust layer, via its impact on the thermal structure of the atmosphere. Such a negative feedback system will lead to a weak, or non-existent, signal in the IDDI-VV correlation fields. If strong convection enhances the generation and/or vertical transport of dust particles, without a subsequent dust-related cooling and subsidence, the signal in the correlation fields will be negative. If an existing dust layer suppresses convection, the correlation signal will be positive.

A further, and highly likely, possibility, is that thermal heating of the atmosphere in the vicinity of a dust layer may reduce the stability of the overlying atmosphere at the same time as the reduction in solar radiation increases stability below the dust layer. This would lead to a change in the sign of the correlations above and below the layer.

The results suggest that all of the above situations are represented. Very little can be said about the frequency of occurrence of the situation in which the positive feedback occurs as, by definition, there is no coherent signal to be interpreted. Such situations may in principle be inferred from the existence of coherent regions of strong, statistically significant negative correlations between IDDI and temperature (large-scale cooling), which do not correspond to coherent regions of positive correlations between IDDI and VV. However, the degree of cooling required to produce a notable decrease in the convective motion of the atmosphere is not well understood. Case studies of situations in which such a correspondence is observed would be useful in determining such relationships. Such studies would need to take account of the local and regional atmospheric dynamics, and should be carried out only when the reliability of the NCEP data has been more fully assessed. Such analyses are outside the scope of this thesis.

However, the existence of coherent negative and positive signals are highly likely to represent cases in which strong vertical motion results in the elevation of dust particles, and in which an existing dust layer increases subsidence or reduces convection, respectively. For either case, the spatial coherence of the signal and/or the strength of the correlations might be reduced by the influence of the other, less dominant, process.

The above considerations lead to the conclusion that atmospheric dust is associated with enhanced atmospheric stability over some Sahelian regions in the winter, and over southwestern coastal regions of West Africa in some spring and summer months. From the point of view of studies of monsoon intensity and the associated rainfall, it is the latter period which is of interest. The extent and frequency of occurrence of the enhanced stability signal is limited enough to suggest that the impact of dust on large-scale spring and summer
convection patterns may be weak, or confined to certain periods within the onset and wet seasons. However, further analysis is required to ascertain whether there is any notable impact on the monsoon flow.

The association of increased convective activity with dust at 200 hPa suggests that at certain times, and at specific locations, dust mobilisation and elevation is the result of an anomalously unstable atmosphere. In September and October, enhanced instability at 200 hPa over the western Sahel is associated with a similar signal at 600 and 700 hPa, but no signal at lower levels. The temperature fields exhibit cooling up to 700 hPa, but warming at 600 and 200 hPa. A plausible conceptual model is one in which dust in the vicinity of the 700 hPa level cools the underlying atmosphere, but causes no net change in stability in the lower layers. The dust layer will warm the atmosphere in its immediate vicinity, leading to increased instability and enhanced upwards motion. If such upward motion extends to 200 hPa, the warming at this level might be due to advection of warmer air from lower layers. A slight problem with this explanation is the existence of both increased convection and cooling at 700 hPa. However, the former could be associated with the processes that elevate dust to this layer, rather than with the impacts of the dust layer itself. The spatial coincidence of the 700 hPa signals is not exact, and the apparent contradiction between temperature and VV fields might simply represent variations in the height or thickness of the dust layer.

7.6. Results III: Meridional wind indices and dust loadings

The aim of this analysis was to examine whether atmospheric dust levels exert an influence on the northwards transport of moist air from the Gulf of Guinea to the Sahel. The correlations were performed between local daily IDDI values representing the 5-day mean for two days either side of each value, and unsmoothed data representing the average daily southerly component of the wind between 20° W and 10° E for various latitudinal bands on the final day of the 5-day IDDI averaging period. If, as postulated, large dust loadings over the course of a period of several days lead to a reduction of the southerly flow, the correlations of the wind index with the local IDDI values should be negative in the regions whose dust loadings have modified the northwards flow. Such modification will take place via regional teleconnections involving changes in the fields of divergence and hence in the regional circulation.

Such a signal is not widely apparent in the monthly correlation fields, produced by pooling ten years of smoothed IDDI and unsmoothed wind data, with the lag built in to the individual monthly timeseries. However, a convincing pattern of statistically significant negative
correlations exists in the May 1000 hPa fields. The correlations are negative throughout most of West Africa and the Sahara when local IDDI values are correlated with the wind averaged over zones 3 and 4 (Figure 7.27). These zones represent the latitudes from 6.25° to 8.75° N and 8.75° to 11.25° N respectively. The pattern is most widespread and coherent over the latter zone; it also exists, but is much weaker and more broken (with many non-significant negative values) for zone 5, from 11.25° to 13.75° N (Figure 7.27).

For zone 4, the correlations become negative and statistically significant immediately north of the band over which the meridional wind averages are calculated. The 1000 hPa May correlations are reflected at 850 hPa, although the signal is only apparent for zones 4 and 5, and is much weaker for zone 4 than at 1000 hPa.

The May signal indicates that high dust levels which persist for several days are associated with subsequent reductions in the strength of the southerly component of the wind across West Africa south of 14° N. These results make it pertinent to examine the variation in the daily wind and IDDI values throughout May. Figure 7.28 shows timeseries of daily v-wind and IDDI values, averaged over the period 1984-1993 and for the individual years of 1984 (dry) and 1988 (wet). The wind series represents zone 4 (see Figure 7.27), and the IDDI series represents values spatially averaged from 10° W to the zero meridian, and 10° to 20° N. The area chosen for the IDDI averaging is the area in which the correlations have the largest magnitude, and in which significant correlations are also seen in zones 3 and 5 (Figure 7.27), indicating the strongest regional association with the wind. It should be noted that the meridional wind is positive (i.e. towards the north) only for the last third of the month in the mean timeseries. This identifies May as a key transition month, during which the southwesterly monsoon flow begins to establish itself over the far south of the Sahel.

The change in sign in the wind values suggests that the high correlations might be the result of trends in the series. However, this change occurs in the mean timeseries, and the correlations are calculated from pooled data from individual months. In a given month, there are large variations in both wind and IDDI on timescales of several days. The smoothing of the IDDI values will increase the autocorrelation in these series and result in inflated correlations between the two quantities. However, the large variation in the wind values, and the lack of an overall trend in the IDDI series, should ameliorate this effect. The use of the random phase method also means that the identification of statistically significant values is conservative; this method compares the correlation between the original data with correlations constructed from the same series with the trends redistributed, making large
values much more likely in the comparison, or “control” correlations. Finally, it should be noted that significant correlations with IDDI values in other regions are few, despite the fact that the IDDI signals over the north-central Sahel and the eastern Sahel-Sahara diminish throughout May, suggesting that trends are likely to occur in the associated series.

Figure 7.27: 1000 hPa correlations between meridional wind averaged over zones 3 (top), 4 (middle) and 5 (bottom) and local daily IDDI values for May. Zones over which the meridional wind is averaged are shaded.
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Figure 7.28: Timeseries of daily meridional wind over the band from 8.75°-11.25° N (top) and daily IDDI averaged over 20°-30° N; 10°-0° W (bottom) for May. The averaged data represent the mean daily series for May from 1984-1993. All data are unsmoothed.

While the May values are not conclusive proof of an impact of dust on the regional circulation, it is likely that the signals in these fields do have a physical significance. A plausible interpretation is that they represent a decrease in the likelihood that winds over the southern Sahel and northern Sudanian zone will change from northerly to southerly if dust levels are high and persistent to the north of these regions. The mechanism involved is most likely to be a decrease in the insolation over the western Sahara and Sahel, resulting in a comparatively weak land-sea differential heating and a reduction in the large-scale convection over West Africa.

A notable, but less extensive, negative signal occurs in September at 1000 hPa, when local IDDI is correlated with southerly wind over zone 3 (3.75° - 6.25° N). However, this signal is confined to the West African coast. For zone 2 (1.25° - 3.75° N), the negative coastal signal is slightly weaker, and significant positive correlations extend throughout parts of the Guinea Coast and Sahel. The zone 2 signal at 850 hPa resembles that at 1000 hPa, with a slightly
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stronger negative coastal signal (Figure 7.29). This signal weakens for zones 3 and 4. The positive 850 hPa signal extends throughout the Sahel and sub-Sahel for zones 2 and 3, becoming more localised near the Guinea Coast for zone 4.

The September signal could represent a similar situation to that postulated for May, but transplanted to the West Africa Coast, where southwesterly winds cross the coast during the wet-season. The positive values are more difficult to explain; it is possible that they are part of a large-scale modification of the circulation by dust extending from the south-central Sahel to the West African coast, in which processes in the latter region drive the modulation, diverting the circulation so that southerly flow is intensified over the Gulf of Benin.

Figure 7.29: 850 hPa correlations between meridional wind averaged over zone 2 and local daily IDDI values for September.

An extensive 1000 hPa negative signal is apparent throughout the Sahara for zone 3 correlations in June (Figure 7.30). For zones 1 and 2, significant positive values occur over the western Sahel, south of the significant Saharan values, which are more localised than for zone 3. Vestiges of these patterns are seen at 850 hPa. Significant positive values occur over the Guinea Coast from zones 4 to 7 in June at 700 hPa. As the monsoon airmass has penetrated further north in June than in May, it might be expected that the latitudes in which dust production in the Harmattan regime affects the southerly monsoon flow would be located further north. The June situation may therefore be analogous to that suggested for May. This does not explain the positive correlations over the Sahel for zone 1. A speculative explanation is that high dust levels over the Sahel reduce the northwards penetration of the monsoon air over land, leading to increased convergence over the Gulf of Guinea. If the resulting convergence is of sufficient intensity, it may produce a localised zonal circulation which intensifies the surface meridional wind to the south.
Localised regions of significant positive correlations are apparent in August at 1000 and 850 hPa away from the coast. At 700 hPa these values form a relatively coherent signal throughout much of the Sahel for zones 1 to 4 (Figure 7.31). Significant positive values also occur in October at 850 and 700 hPa at various Saharan and coastal locations. Similar processes to those outlined for June may also operate in August and October.

**Figure 7.30: 1000 hPa correlations between meridional wind averaged over zone 3 and local daily IDDI values for June.**

**Figure 7.31: 700 hPa correlations between meridional wind averaged over zone 1 and local daily IDDI values for August.**

### 7.7. Summary

In the discussion of the hypotheses to be tested in this chapter, the question was raised as to whether atmospheric dust had the potential to modify the monsoon circulation in such a manner as to inhibit the processes associated with rainfall generation. The subsequent analyses have provided ample evidence in support of this hypothesis.
It is evident from the results presented here that atmospheric dust causes widespread thermal modification of the troposphere over northern Africa, and particularly over the Sahel. An assessment of the significance of the correlations between IDDI and temperature values indicates that the impact of dust is important over large areas of northern Africa. The dominant effect is one of cooling in the first one to two kilometres of the atmosphere, with less pronounced warming of the mid-troposphere. Although this pattern is strongest in the winter months, it is also widespread in the spring and summer. The mid-level warming has implications for the dynamics of the African easterly jet and the associated easterly waves and convective disturbances (c.f. Chapter 2, Section 2.14). Although this warming is likely to increase the mid-tropospheric convection, such convection occurs in the dry Saharan air layer, and therefore is unlikely to be associated with the generation of convective rainfall. The enhanced stability of the lower layers will act to impede the development of such events, suggesting a likely suppression of rainfall by large dust loadings. The vertical distribution of cooling and warming is in good agreement with the patterns of dust transport and the effects of dust aerosols described in the literature reviewed in Chapter 2.

Evidence that the thermal impacts of dust are associated with changes in the intensity of the vertical motion of the atmosphere is also provided by the correlation analyses between IDDI and vertical velocity (VV) values. Although the VV fields are not expected to capture dust signals to the same extent that the temperature fields will (due to the greater influence of the model used to create the reanalysis data in determining the former), they do exhibit signals indicative of dust-related modification of the Sahelian and Saharan atmosphere.

The Sahelian signals in the IDDI-VV correlation fields are strongest and most widespread in winter at the 850 hPa level, where they indicate enhanced subsidence. A very strong cooling signal is also seen in the winter IDDI-temperature correlation fields at 850 hPa. At this time of year, dust is transported from West Africa over the Atlantic in a shallow layer above 1.5 km (850 hPa) and below 3 km (700 hPa) (Schütz et al., 1981; Hastenrath, 1991; Alpert et al., 1998). If such a layer extends over the Sahel, it would be likely to cause cooling, and therefore enhanced subsidence, below 850 hPa. Cooling is much less widespread at and above 700 hPa, supporting the notion that these levels lie within or above the dust layer over many regions. The winter signal in both the temperature and the VV correlation fields therefore seems realistic, suggesting that both variables capture the effects of dust to some extent.

In summer, the cooling extends to higher levels, consistent with the increase in the height of the base of the Saharan air mass. The IDDI-temperature correlations suggest that the Saharan air is often restricted above 700 hPa, or some 3.5 km. Above this level, the tendency for
warming apparent from the correlation distributions is consistent with the presence of dust up to 500 hPa (6 km) (Prospero et al., 1981). Warming above the boundary between the monsoon and Saharan air will strengthen the temperature inversion which exists where these two air masses meet (Riehl, 1979; Prospero et al., 1981). This will lead to the suppression of large-scale convection below, and in the vicinity of, the average altitude of the inversion, but may increase instability within the dust layer. This effect is seen in the vertical profiles of the IDDI-VV correlations in JAS.

The above effects may have a significant impact on the development of the convective disturbances that generate most Sahelian rainfall. An investigation into the relationship between dust and the dynamics of individual convection events is outside the scope of this thesis, although it is reasonable to suppose that a cooling of the lower troposphere may inhibit the formation and/or development of disturbance lines. Another means by which dust may influence rainfall amounts is via the modification of the monsoon dynamics, and hence on the wet-season atmospheric environment. Evidence that high dust levels can modulate the regional circulation in some instances has been presented. This evidence suggests a reduction in the strength of the southerly monsoon flow near the land surface between about 6° and 14° N in May and June following widespread and persistent dust events. May and June are key months as they represent the monsoon onset period. It is likely that the meridional flow will be most sensitive to perturbations arising from dust-modification when the meridional winds are weak. This condition will characterise the onset months as winds change from northerly to southerly. The apparent relationships between dust loadings and southerly wind strength in May and June therefore make good physical sense. The broader implication is that large dust loadings over the western Sahara and Sahel may have the potential to delay the onset of the monsoon. In addition to the transition-season impact, there is some evidence of modification of the monsoon circulation in other wet-season months.