Satellite detection of dust using the IR imagery of Meteosat

1. Infrared difference dust index

M. Legrand, A. Plana-Fattori,1 and C. N’doumé2

Laboratoire d’Optique Atmosphérique, Université des Sciences et Techniques de Lille, Villeneuve d’Ascq, France

Abstract. The Infrared Difference Dust Index (IDDI) is a satellite dust product designed for climatological applications, designed specifically for dust remote sensing in arid regions such as the Sahel and Sahara. It is based on the atmospheric response to dust, extracted from midday Meteosat-IR imagery, and takes advantage of the impact of dust aerosols on the thermal infra-red radiance outgoing to space. Simulations show a quasi-linear relationship between satellite response to dust and shortwave optical depth, with a sensitivity depending on particle size distribution and radiative surface properties. Comparison of measured satellite response with photometric optical depth agrees with the simulations. Water vapor significantly affects the satellite signal for cases of large columnar amounts and oceanic air masses advected inland. Hence apart from possible coastal effects, the water vapor effect can be neglected in the Sahelian-Saharan zone north of the Intertropical Convergence Zone, coinciding with the major regions of African dust emission and transport. The construction of the IDDI involves the processing of reference images, theoretically representing the outgoing radiance obtaining under clear-sky conditions. Errors may arise from (1) dust remaining in the reference images and (2) seasonal shifts of the reference level; however, the latter error will be offset by averaging used in climatological processing. An error budget is presented for the station of Gao. A statistical comparison of IDDI data with visibility measured at synoptic stations results in (1) a validation of the product, and (2) a climatologically relevant visibility-IDDI relation, valid for the arid regions of northern Africa. The latter relation is consistent with both simulations and photometric measurements. IDDI maps over Africa compare successfully with optical depth over adjacent ocean regions derived from Meteosat-VIS imagery. The observed continuity of dust plumes across the African coast demonstrates the consistency between both products.

1. Introduction

In the first part of the desert dust cycle, during emission and the following atmospheric transport over the adjacent land, desert dust can be detected from space in the thermal infrared part of the spectrum. This was observed over the vast arid expanses of North Africa and Arabia, by Shenk and Curran [1974] using Nimbus THIR (temperature humidity infrared radiometer) data (10.5-12.5 μm), by Legrand et al. [1983, 1985] and Oliva et al. [1983] using Meteosat-IR channel (10.5-12.5 μm), and by Ackerman [1989] using 3.7 and 11 μm NOAA AVHRR (advanced very high resolution radiometer) data. The best sensitivity of the method was observed during the middle of the day with a decrease of the radiance outgoing to space, and also during the latter half of the night with an increase of this radiance [Legrand et al., 1988]. Quantitative comparisons between the daytime satellite response to dust presence and the aerosol shortwave (SW) optical depth derived from the well-known direct-Sun photometric technique, established the close correlation between the two parameters [Legrand et al., 1989; Tanré and Legrand, 1991] showing the potential of an infrared dust index for quantitative estimates of dust amount. Concurrently, the somewhat complex physics of this effect was analyzed and modeled [Cautenet et al., 1992; Legrand et al., 1992; Plana-Fattori, 1994]. Because of the agreement observed between the experimental and the numerical approaches, we relied on this infrared method, and we decided to use it to run a climatology of dust presence over North Africa. This was achieved by processing the daily midday IR images of Meteosat so that the dust effect component (referred to as the Infrared Difference Dust Index (IDDI)) is separated from the total satellite signal. The first version of this product was an 8 year (1984-1991) climatology obtained by N’doumé [1993]. Since then, the product has been improved, the region covered extended, and a data set covering the period 1983-1998 is now available at the Laboratoire d’Optique Atmosphérique (LOA). Its extension to 2000 is under way. In recent years the IDDI has been used to validate dust emission modeling [Marticorena et al., 1997, 1999], for the satellite determination of the dust emission wind speed threshold [Chomette et al., 1999] and to investigate the possible links existing between dust and rain in the Sahel [Brooks and Legrand, 2000].

At this point, it appears that a synthetic survey, based on various studies published and unpublished on the infrared method and the IDDI, dedicated to the description and understanding of this index, focusing on its performances and limitations, would certainly be of interest to the community of scientists involved in research on dust and connected topics.

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In this first part of the paper, a general presentation has been given of the infrared method and of the derived dust index. The physical processes involved and their simulation are described in section 2. In section 3, several significant results are derived from the simulations. Section 4 brings together results from various experiments of comparison between Meteosat-IR data and ground-based photometric measurements and draws conclusions from their confrontation with simulations. In section 5, a description is given of the technique of processing of the IDDI, designed to detect the presence of desert dust over Africa in view of climatological applications. Section 6 is devoted to the IDDI, focusing especially on its limitations in connection with the way the product is processed. Section 7 is dedicated to the results of various validation studies that examine the ability of the IDDI to detect airborne dust and to clarify its relationship to more conventional indicators of dust presence.

A forthcoming part 2 is dedicated to the setting up of a climatology of desert dust over Africa, based on 18 years of Meteosat-IR data representing the period 1983-2000. Seasonal and yearly dust distributions are presented. The locations of the main sources of dust emission and their seasonal activity are described.

2. Physics of Dust Detection in the Thermal Infrared

2.1. Description of Physical Processes

Since the late 1970s, images of Africa in the channels VIS (0.5–0.9 μm), water vapor (WV) (5.7–7.1 μm), and IR (10.5–12.5 μm) have been routinely yielded by the Meteosat satellites [Ma-son and Schmetz, 1992], suitably located over the Gulf of Guinea, at the longitude 0°. The size of the pixel at nadir is (2.5 X 2.5) km² in the VIS channel and (5 X 5) km² in the WV and IR channels. The measurements are digitized according to an 8 bit scale, in counts linearly related to the radiance at the radiometer entrance.

In the daytime IR images, dust presence is associated with a decrease of radiance (and of the associated brightness temperature). Hereinafter, results generally will be expressed in counts, defined according to a scale of 0.08 W m⁻² sr⁻¹ count⁻¹ valid for the Meteosat-4 radiometer. This standard corresponds to an approximate change of 0.5 K in brightness temperature at 300 K.

A temperature depression of 10 K corresponds to values commonly observed and several tens of degrees are reached for major dust storms, so a maximum of 32.9 K is recorded on February 16, 1985 nearby Lake Chad (14°N, 13°E) during a major dust event reported by Legrand [1990]. MéthoFrance [1991] exhibits in its pictures of the month a sandstorm image during which we determine a maximum of 29.3 K in northern Mauritania (24°N, 11°W) on March 14, 1991. Even larger values can be observed; a maximum of 45.6K has been registered SW of Air mountains (16°N, 7°E) on February 18, 1989. These examples are extracted from our IDDI imagery database derived from Meteosat-IR images at 1200 UTC.

This radiance decrease, which represents the radiative forcing of dust, results from a dual physical origin outlined in the diagrams of Figure 1.

1. The values of SW aerosol optical depth range usually from 0.5 to around 3 during the dust events (e.g., photometric determinations by Ben Mohamed and Frangi [1986], d'Almeida [1987],

![Figure 1](https://example.com/f1.png)

**Figure 1.** Mechanisms explaining the depression of thermal infrared radiance outgoing to space during daytime, in the presence of a dust layer over land. On left, temperature depression at the surface is due to a decrease of downward SW flux prevailing over the increased downward LW flux. In addition to radiative fluxes, sensible heat (SH), conduction heat (CH), and latent heat (LH) are involved in the equilibrium defining surface temperature. This temperature depression, damped by the presence of vegetation and soil moisture, is maximum for arid surfaces. The radiative parameters on right are integrated over the radiometric band pass. The dust layer absorbs more IR radiance than it emits, because it is colder than the ground surface. The radiance depression through the layer is ΔR = R_i - R_i = (1 - 0)(B(T_a) - B(T_s)), depending on the layer transmittance t, its mean temperature T_m and temperature T_s of the surface (assumed black); B means blackbody radiance.
that the thermal infrared radiance decrease (and the resulting heating rates are modified throughout the atmosphere [Cernich, 1980]. The resulting temperature change concerns the whole troposphere, and also the superficial soil layer. As illustrated in Figure 1, the mechanisms governing these changes are not only radiative but include also convective processes, evaporation, and heat conduction in the ground.

In summary, the thermal infrared radiance decrease (and the associated drop in brightness temperature) measured from the satellite in the presence of dust is due to (1) a ground surface cooled by the dust radiative impact in the SW and (2) the fact that the measurements are made through a cooler attenuating dust layer. The examples of large measured brightness temperature decreases previously reported result from these adding effects. It should be stressed that these high values are a major strength of the method. In addition, the physical processes involved prove clearly the specific relevance of the infrared method over arid environments.

2.2. Simulating the Physical Processes

The dust impact is not strictly limited to the surface temperature. The radiative fluxes, SW and LW (longwave), and the resulting heating rates are modified throughout the atmosphere [Carlson and Benjamin, 1980]. The resulting temperature change concerns the whole troposphere, and also the superficial soil layer. As illustrated in Figure 1, the mechanisms governing these changes are not only radiative but include also convective processes, evaporation, and heat conduction in the ground.

To obtain realistic quantitative estimates of the impact of dust on radiation and temperature and of the resulting effect on the satellite-detected IR radiance, we performed simulations of the physical mechanisms involved, using the mesoscale model of the Colorado State University (CSU) [Mahrer and Pielke, 1978]. Suitable codes of radiative flux transfer, developed earlier at the LOA [Cautenet et al., 1992], were merged with the model, allowing the radiative impact of dust to be suitably calculated.

For the model validation, results of the simulations were compared with experimental data [Cautenet et al., 1992] available from the field experiment ECLATS (Etude de la Couche Limite Atmosphérique Tropicale Sèche) carried out in the Sahel, near Niamey (Republic of Niger) in November and December 1980, at the beginning of the dry season [Druilhet and Tinga, 1982; Fouquart et al., 1987a, 1987b; Durand et al., 1988]. These data enabled the CSU model to be supplied with (1) parameters describing the properties of the surface and of the underlying soil, (2) vertical profile and size distribution of dust particles, as well as their derived optical parameters, (3) required boundary conditions at the 6000 m high top and 1 m deep bottom of the modelized domain, and (4) initial (sunrise) profiles of temperature, humidity, and wind. Daily cycles and vertical profiles for both radiative and nonradiative flux components, temperature, and wind of the ECLATS experiment were compared with the model computations for three well-documented cloudless days characterized by different dust amounts. The simulations show a satisfactory overall agreement with the experimental data. The simulated temperature of the ground surface shows a slight underestimate ranging from 0.8°C to 1.3°C. Inside the planetary boundary layer (PBL), up to 1000 m or so, the departure of the simulations from the experimental data is below 1°C. Only above the PBL is this departure greater, owing to synoptic advection which is not taken into account in the model.

This agreement means that this model, applied to a Sahelian environment during the dry season, provides reliable results. The signal of the Meteosat-IR channel calculated from these simulations therefore can be considered realistic. This fact offered the opportunity to perform senitivity tests on the satellite response to dust, by running the LOWTRAN-5 radiative transfer code [Kneizys et al., 1980] with the atmospheric profiles and the surface temperature calculated by the CSU model [Legrand et al., 1992]. To generalize these conclusions, a second set of simulations was conducted with the CSU model [Plana-Fattori, 1994]. The outgoing 10.5-12.5 μm radiances were computed through the LOWTRAN-7 radiative transfer code [Kneizys et al., 1988]. The influence of the meteorological initial conditions was minimized by running the mesoscale model over periods long enough to obtain quasi-periodic diurnal cycles of surface temperature $T_s$, $\Delta T_s < 0.5$°C at a 24 hour interval). These quantitative tests agree with the earlier ones of Legrand et al. [1992], confirming the main factors affecting the satellite response and allowing us to estimate the significance of their impact. The principal factor is the dust amount. The particle size distribution and the radiative properties of the surface, SW albedo, and LW emissivity must also be considered. Finally, fluctuations in atmospheric water vapor content can affect the satellite response.

3. Some Information Derived From Simulations

3.1. Incidence of Dust Particle Size Distribution

In addition to its extensive range, the size distribution of the mineral particles exhibits a large variability. The surface wind speed controls the relative production of large and fine particles of dust through the sandblasting process [Alfaro et al., 1998]. The main process acting on the size distribution of dust particles freshly lifted into a dry cloudless atmosphere is sedimentation under gravity, the efficiency of which increases with particle size. The
larger the particles the faster they settle. On the other hand, the
particles in the submicronic range are practically unaffected by
this process, so in a dry cloudless atmosphere the fraction of fine
particles increases with dust ageing. This evolution was observed
through measurements performed at various distances from the
sources of dust emission, between Libya and Barbados, along a
major westward transport path to the Atlantic and the Caribbean
Islands (reported by Westphal et al. [1987]).

This variability is taken into account in the models of desert
aerosol by including options relating to the size distribution, so
are the background and dust storm distributions of Shettle
[1984], the wind-parameterized distribution of Longtin et al.
[1988], and the background, wind-carrying dust, and sandstorm
distributions of d'Almeida [1987].

Legrand et al. [1992] investigated the effect of the particle
size distribution on the Meteosat-IR signal by comparing the re-
spective simulated effects of a fine desert aerosol referred to as
the ECLATS model and of a coarse one referred to as the Carlson
and Benjamin (CB) model, both derived from measurements and
described by Fouquart et al. [1987a, 1987b] and Prospero et al.
[1976]. Plana-Fattori [1994] extended the analysis by adding the
so-called background (fine) and severe dust storm (coarse) mo-
dels described by Longtin et al. [1988] (hereinafter referred to as
LSHP background and dust storm). The radiative parameters of
these models, shown in Table 1, are functions of the complex
index and of the size of the aerosol particles (assumed to be spheri-
cal). The complex index depends on the mineralogical composi-
tion of the aerosol. In the models of Longtin et al. the aerosol is
an external mixture with a dominant sand mode composed of
quartz with hematite inclusions. The models ECLATS and CB re-
fer to "Saharan dust". Their complex index is typical of quartz
silicates, but their mineralogical composition is not specified.
However, the dominant influence on aerosol radiative properties
appears to be particle size. Differences in particle size distribu-
tion are shown in Table 1 from the large differences between the
values of atmospheric columnar mass M for the same unit value
of $\delta_l$. Large values of M are due to large particles and charac-
terize coarse dust (for large particles, the ratio of mass to optical
depth increases with radius). The LW optical depth $\delta_l$ is mark-
edly smaller in the models of fine dust (ECLATS, 0.12; LSHP
background, 0.05) than in the models of coarse dust (CB, 0.73;
LSHP dust storm, 1.06). This results from the weaker contribu-
tion to $\delta_l$ than to $\delta_l$ of the submicronic particles predominant in
the fine dust models.

These features of the fine and coarse dust models will result in
very noticeable differences between their radiative impacts, both
in the SW and in the LW ranges. Figure 2 shows the overall ra-
diative impact of a dust layer during the daily cycle. In Figure 2a
the radiative impact on the surface temperature is quite different
with the fine dust models and with the coarse dust models. The
surface temperature (shown in Figure 2a), higher during the
whole diurnal cycle with the coarse dust models than with the
fine dust ones, results from the strength of the greenhouse effect
in the former, arising from the large emitted downward LW irra-
diance (due to the large optical depth $\delta_l$), and from its weak-
ness in the latter. The weak daytime temperature effect of the coarse
dust models is explained by this greenhouse effect counterbal-
cing the cooling effect due to attenuation of downward SW flux
reaching the surface. On the other hand, the weak nightime tem-
perature effect of the fine dust models is obviously explained by
the weakness of the greenhouse effect. Figure 2b shows that the
satellite response obtained at 1200 LST is larger with the coarse
dust models than with the fine dust models, in spite of the weak
impact of the former on the surface temperature. This fact means
a stronger attenuation of the ground-emitted upward IR radiance
through the atmosphere, again due to the greater value of $\delta_l$ for
course dust. The weak $\delta_l$ value of the fine dust models results in a
weak atmospheric attenuation due to little change of atmosphere
transmittance. This analysis is confirmed by Table 2, which pre-
ents the composition of simulated Meteosat-IR radiances wi-
out aerosol and for the aerosol models of Longtin et al. [1988].
We can summarize the effects of these two extreme dust models
stating that detection of dust of the background model (and any
fine dust model) is predominantly due to a surface cooling effect,
while detection of dust of the dust storm model (and any coarse
dust model) is predominantly due to attenuation of ground-
emitted radiance. Any realistic dust layer, with an intermediate
size distribution, will be detected through a combination of both
effects.

3.2. Satellite Sensitivity to Dust

The simulations are shown to verify linear or quasi-linear rela-
tionships between $\delta_l$ and the outgoing radiance arriving at the

Table 1. Compared Properties of Several Aerosol Models

<table>
<thead>
<tr>
<th>Model</th>
<th>$c_a$ ($\mu g m^{-3})$</th>
<th>$M (g m^{-2})$</th>
<th>$\sigma_l$</th>
<th>$\delta_l$</th>
<th>$\delta_l$</th>
<th>$\sigma_l$</th>
<th>$\delta_l$</th>
</tr>
</thead>
<tbody>
<tr>
<td>ECLATS</td>
<td>750</td>
<td>0.9</td>
<td>0.952</td>
<td>0.655</td>
<td>0.12</td>
<td>0</td>
<td>0.54</td>
</tr>
<tr>
<td>CB</td>
<td>3600</td>
<td>4.3</td>
<td>0.844</td>
<td>0.780</td>
<td>0.73</td>
<td>0.54</td>
<td>0.65</td>
</tr>
<tr>
<td>Background</td>
<td>1095</td>
<td>1.3</td>
<td>0.989</td>
<td>0.638</td>
<td>0.05</td>
<td>0.73</td>
<td>0.65</td>
</tr>
<tr>
<td>Dust Storm</td>
<td>219300</td>
<td>260</td>
<td>0.785</td>
<td>0.850</td>
<td>1.06</td>
<td>0.616</td>
<td>0.855</td>
</tr>
</tbody>
</table>

(1) the ECLATS and the CB models (adapted from Legrand et al. [1992, Table 4]), (2) the LSHP background and dust storm models used by
Plana-Fattori [1994]. Values are computed for a dust layer 1200 m thick with a unit SW optical depth. The variable $c_a$ is the mass concentra-
tion in the dust layer, and M is the atmospheric columnar mass of dust; $\delta_l$, $\sigma_l$, and $\delta_l$ are the optical depth, the single-scattering albedo,
and the asymmetry factor of dust, respectively; subscript s refers to the SW range (at 0.55 $\mu m$ for ECLATS model and for LSHP models, at 0.5 $\mu m$ for CB mo-
del) and subscript l refers to the LW range (at 8-14 $\mu m$ for ECLATS, at 10-13.2 $\mu m$ for CB, at 11.5 $\mu m$ for LSHP).

Table 2. Components (in %) of the Simulated Radiance Outgoing to Space, Measured Through the Meteosat-IR Channel at 1200
LST, for the Dust Models of Longtin et al. [1988]

<table>
<thead>
<tr>
<th>Components</th>
<th>Dust Free</th>
<th>Background</th>
<th>Dust Storm</th>
</tr>
</thead>
<tbody>
<tr>
<td>Ground emitted</td>
<td>88.6</td>
<td>84.4</td>
<td>32.3</td>
</tr>
<tr>
<td>Ground reflected</td>
<td>1.2</td>
<td>1.4</td>
<td>1.7</td>
</tr>
<tr>
<td>Atmospheric</td>
<td>10.3</td>
<td>14.2</td>
<td>66.0</td>
</tr>
</tbody>
</table>

Layer 1200 m deep, SW Optical Depth of 1, Surface Albedo of 0.25, Surface Emissivity of 0.90, Water Vapor Amount of 1.7 g cm$^{-2}$. 
Figure 2. Simulated diurnal cycles of (a) temperature of the ground surface and (b) radiance outgoing to space through the Meteosat-IR channel, and the corresponding radiometric output. The dust layer, homogeneous, lies between the surface and 1200 m. Its optical depth is 1 at a wavelength of 0.55 μm. The surface albedo and emissivity are 0.25 and 0.90, respectively. The columnar water vapor amount is 1.7 g cm⁻². Ancillary parameters can be found in the work of Plana-Fattori [1994].

Table 3. Simulated Satellite Sensitivity to Dust $\beta$ (Counts) and Corresponding Brightness Temperature Sensitivity $\beta_T$ (K), for Various Models of Dust, Surface Albedo $\lambda$ and Emissivity $\epsilon$

<table>
<thead>
<tr>
<th>Model</th>
<th>$\lambda$</th>
<th>$\epsilon$</th>
<th>$\beta$ (Counts)</th>
<th>$\beta_T$ (K)</th>
</tr>
</thead>
<tbody>
<tr>
<td>Background</td>
<td>0.25</td>
<td>0.90</td>
<td>15.0</td>
<td>6.5</td>
</tr>
<tr>
<td>$(\rho = 0.05)$</td>
<td>0.25</td>
<td>0.90</td>
<td>14.6</td>
<td>6.2</td>
</tr>
<tr>
<td></td>
<td>0.35</td>
<td>0.95</td>
<td>10.0</td>
<td>4.4</td>
</tr>
<tr>
<td>Dust Storm</td>
<td>0.25</td>
<td>0.90</td>
<td>21.1</td>
<td>9.2</td>
</tr>
<tr>
<td>$(\rho = 1.06)$</td>
<td>0.25</td>
<td>0.95</td>
<td>23.9</td>
<td>10.2</td>
</tr>
<tr>
<td></td>
<td>0.35</td>
<td>0.90</td>
<td>13.9</td>
<td>6.2</td>
</tr>
<tr>
<td>ECLATS</td>
<td>0.25</td>
<td>0.90</td>
<td>16.9</td>
<td>7.5</td>
</tr>
<tr>
<td>$(\rho = 0.12)$</td>
<td>0.25</td>
<td>0.90</td>
<td>13.8</td>
<td>6.0</td>
</tr>
<tr>
<td>CB</td>
<td>0.25</td>
<td>0.90</td>
<td>17.4</td>
<td>7.6</td>
</tr>
<tr>
<td>$(\rho = 0.73)$</td>
<td></td>
<td></td>
<td></td>
<td></td>
</tr>
</tbody>
</table>

The relation $\beta (W m^{-2} sr^{-1}) = 0.08 \beta$ (Counts) is valid with Meteosat 4.
Figure 3. Simulated values at 1200 LST against columnar water vapor amount of (a) temperature of the ground surface and (b) radiance outgoing to space through Meteosat-IR channel and corresponding radiometric output. The conditions of simulation are as for Figure 2. Ancillary parameters can be found in the work of Plana-Fattori [1994].

Meteosat-IR channel, for fixed properties of the dust layer and the underlying surface [Legrand et al., 1992; Plana-Fattori, 1994]. We can conveniently compare the various results by employing a linear approximation

\[ c = \beta \delta_s, \]

where \( c \) is the satellite response in counts (i.e., the difference between output signals with and without dust), and \( \beta \) is referred to as the satellite sensitivity to dust, a coefficient dependent on the size distribution of the dust particles, as shown in the previous section, and on surface albedo and emissivity. Thus the sensitivity to dust \( \beta \) defined in (1) can be conveniently described by the ratio \( \rho = \delta_s/\delta_{s0} \), a radiative parameter representative of the particle size distribution, and by the surface albedo \( A \) and emissivity \( e \). Therefore (1) can be more explicitly specified according to

\[ c(\delta_s, \rho, A, e) = \beta(\rho, A, e)\delta_s. \]

Table 3 shows the values of \( \beta \) computed with the previously described dust models (with \( \delta_s = 1 \)) and with values describing the ranges of albedo and emissivity typical of arid and semiarid surfaces in the Sahara and the Sahel. This parameter varies in a ratio of 2.4, between 10.0 and 23.9 counts, and the variability due to the dust model only (for an albedo of 0.25 and an emissivity of 0.90) is defined by a ratio of 1.5, with extreme values of 13.8 and 21.1 counts.

Finally, it should also be mentioned that if very large optical depths, say larger than 1-2, are met, which usually happens in dust storms, the sensitivity of \( c \) against \( \delta_s \) decreases significantly and the linear relation (1) holds no longer (a quadratic fit between these parameters is then more suitable).

3.3. Incidence of Water Vapor

The atmospheric water vapor amount needs to be taken into account for the assessment of the LW radiance outgoing to space. As does dust, the water vapor (1) modifies both the SW and the LW radiative fluxes, resulting in a heating of the ground surface and an increase of its emission, and (2) modifies radiative transfer by absorbing surface-emitted radiation and by emitting concurrently its own radiative component. So, the spatiotemporal changes of atmospheric water vapor content imply changes in the outgoing radiance. This effect, extraneous to dust, is a potential source of error, since dust is estimated by its impact on radiance through satellite remote sensing. For this reason, it was studied by Legrand et al. [1992, Figure 10], then by Plana-Fattori [1994], using simulations. Figure 3, from the results of Plana-Fattori, shows surface heating resulting from the water vapor greenhouse effect, against the columnar water vapor amount \( w \) (Figure 3a), and the corresponding response of the Meteosat-IR
It is obvious that for low or moderate values of $w$, the aforementioned greenhouse and radiative transfer effects balance each other approximately, so the radiance variations are quite limited. Only for high values of $w$ is a significant decrease of radiance observed, indicating a deficit in the greenhouse effect. If $w$ increases in the atmosphere, a radiative saturation occurs at the edges of the absorption bands, narrowing the $10\mu m$ spectral window. Hence this radiative saturation gives rise to a saturation of the surface heating, shown in Figure 3a by the decrease of the slope of the temperature curve at high $w$ values. On the other hand, the atmospheric radiative transfer in the Meteosat-IR channel, in the least absorbing region of the $10\mu m$ window, is not subjected to such saturation in the effective range of $w$. Hence it makes a greater contribution than the greenhouse effect for high values of $w$. In the range between 0.5 and $2.5\,g\,cm^{-2}$ of $w$ values, the maximum variation of radiance from a dust-free atmosphere is around 2%, corresponding to a simulated Meteosat-IR response not greater than four counts, equivalent to the effect of a variation of dust optical depth $\delta_s$ of 0.17-0.40 (derived from the simulations shown in Table 3). This effect is weaker in the presence of dust, corresponding, for a unit value of $\delta_s$, to the effect of a variation of 0.07-0.18 of $\delta_s$. If $w$ reaches $3\,g\,cm^{-2}$, the satellite response for a dust-free atmosphere exceeds six counts, and over this value, the effect of water vapor becomes quite important.

In conclusion, dust can be detected using the infrared method without significant contamination by water vapor, over the whole Sahara all the year-round, and over the Sahel and the Sudanian savannah during the dry season.

4. Results From Experiments

4.1. Comparing Satellite Response to Photometric Optical Depth of Dust

Concurrently with theoretical simulations of the relationship between IR satellite response to dust $c$ and SW optical depth $\delta_s$, ground-based photometric measurements have been performed at Sahelian sites during the dry season, and the derived values of $\delta_s$ can be compared to the coincident IR satellite data. Results from this experimental approach were published by Legrand et al. [1989] and Tanré and Legrand [1991]. Parts of these published results and unpublished ones are compiled hereinafter, especially from the point of view of their degree of consistency with the simulations.

4.1.1. Niamey measurements. Photometric measurements were performed in February 1985, at the Centre Agrhymet (13.53°N, 2.08°E), Niamey, Republic of Niger. The satellite data used are Meteosat-IR images in B2 format, obtained by sampling one pixel out of six, in both lines and columns from images in full resolution. Thus the successive pixels are 30 km apart at na-
Figure 5. Experiment on dust remote sensing, February 1985, Niamey, Republic of Niger. (a) Time variations of aerosol optical depth $\delta_s$ (dotted line) and of satellite signal $C$ (solid line). (b) Fit of signal difference $c$ against $\delta_s$. (c) Time variations of $\delta_s$ (dotted line) and columnar water vapor amount $w$ (dashed line).
Figure 5a compares the time variations of aerosol optical depth at 0.45 μm at 1200 UTC, to the coincidental satellite radiometric level \( C \) (counts) of the 3X3 pixel array centered at Niamey. A dust storm occurred at midmonth, starting on Julian day 45, peaking on day 48 and ending on day 52. The turbidity was so strong that photometric measurements were not performed on days 47, 48, and 49 because of the difficulty to track the Sun with the handheld photometer. In this period, optical depth estimates derived from visibility observations (crosses) reach values around 5. After a lull on day 53, a new severe event occurred at the end of February. The satellite signal shows a large response to the dust presence with a drop of 40 counts or so on day 48. Another observation illustrated by Figure 5a is a smooth increase of the satellite signal, independently of dust presence. It stems from the seasonal heating during this period of the year at this site. Taking this drift into account, the satellite signal is expressed as

\[
C(\delta t, t) = C_0 + st - \beta \delta t, \tag{3}
\]

where \( t \) is the date, \( s \) describes the seasonal heating rate of the Earth-atmosphere system, and \( \beta \delta t \) is the satellite response to dust \( \delta t \), in agreement with (1). The clean-air signal \( C(0, t) \), derived from a linear least squares fit of \( C \) against \( \delta t \) and \( t \), is plotted in the figure.

Figure 5b represents the linear least squares fit of \( C \) against \( \delta t \), after rejection of cloud contamination (one case). The vertical bars represent the standard deviation \( \sigma \) of the 3X3 pixel array, indicative of the error in \( c \) using B2 images, mainly due to inhomogeneity in the dust cloud and to cloud contamination (over six counts, \( \sigma \) indicates a cloudy case). The satellite sensitivity to dust \( \beta \) is 9.18 counts, and its 95% threshold confidence interval is \([7.6, 10.8]\) counts.

Figure 5c reveals the smooth aspect of the time variation of the columnar water vapor amount \( w \), with weak variations between 0.62 and 1.55 g cm\(^{-2}\), in contrast with the large variations of optical depth (between 0.36 and 5). In addition, according to the simulated results presented in section 3.3, the low values of \( w \) allow us to predict a weak influence of the water vapor on the satellite signal \( C \). Consistently, a tentative correction of \( C \) to ac-
count for the effect of water vapor, using the simulated results shown in Figure 3, yields negligible change. Similarly, fitting $c$ against $w$ in addition of $\delta_s$ brings no significant difference in the results, confirming the negligible effect of $w$.

4.1.2. Dakar measurements. Photometric measurements carried out in April and May 1987, at the Institut Français de Recherche Scientifique pour le Développement de la Coopération (ex-ORSTOM) from M’bour (16.90°W, 14.38°N) near Dakar, Senegal [Tanré et al., 1988], allowed the derivation of aerosol optical depth and of columnar water vapor amount using the differential absorption method. These data are compared to IR Meteosat images both in full resolution and in B2 format. Results from this comparison are presented by Tanré and Legrand [1991].

Figure 6a compares the time variations at 1200 UTC, of aerosol optical depth $\delta_s$ at 0.45 μm, columnar water vapor amount $w$ and “total” satellite response $(C_0 - C)$, where $C$ is the full resolution 3X3 pixel signal and $C_0$ is a fitted offset of 196.4 counts (taken from (4)). Both quantities $\delta_s$ and $w$ show large irregular variations, between 0.66 and 2.32 for $\delta_s$ and between 0.52 and 2.74 g cm$^{-2}$ for $w$. A large proportion of days show no measurements (66%), mostly due to the high frequency of cloud contamination occurrence. The time variations of the satellite response exhibit large discrepancies, compared to the variations of $\delta_s$, in agreement with the poor linear least squares fit between $(C_0 - C)$ and $\delta_s$ plotted in Figure 6b (correlation coefficient $r = 0.32$).

Taking water vapor into account according to a linear least squares fit of $C$ against $\delta_s$ and $w$,

$$C(\delta_s, w) = C_0 - \beta \delta_s - kw$$  \hspace{1cm} (4)

shows the significant contribution of water vapor to the satellite signal through an increase of $r$ to 0.975. Figure 7a shows the variations of the satellite response corrected for water vapor variations, $c'$ (representing the satellite response to dust with an offset $k\overline{w}$), in good agreement with the variations of $\delta_s$.

$$c' = (C_0 - C) - k(w - \overline{w}) = \beta \delta_s + k\overline{w} = c + k\overline{w}$$  \hspace{1cm} (5)

where $\overline{w} = 1.38$ g cm$^{-2}$ is the average value of $w$, and $k = 10.78$ counts cm$^{-2}$ g$^{-1}$ is the satellite sensitivity to water vapor.

Figure 7b confirms the agreement. The satellite sensitivity to dust $\beta$ is 15.06 counts, and its 95% threshold confidence interval is [11.5, 18.6] counts. The vertical bars represent the standard deviation for the 3X3 pixel arrays.

Figure 7. Dust remote sensing taking into account the effect of water vapor, April-May 1987, Dakar, Senegal. (a) Time variations of aerosol optical depth $\delta_s$ (dotted line), columnar water vapor amount $w$ (dashed line), and satellite response $c'$ corrected of water vapor variations (solid line). (b) Fit of satellite response $c'$ against $\delta_s$.  

\begin{align*}
\text{Satellite response } c' \text{ (counts)} \\
\text{Optical depth at 450 nm, column water vapor (g cm}^{-2}\text{)}
\end{align*}
Figure 8. Dust remote sensing from December 1986 to April 1987, Gao, Mali. (a) Time variations of aerosol optical depth $\delta_5$ (dotted line) and of satellite signal $C$ (solid line); the stepped line is the adjusted clean-air satellite signal. (b) Fit of signal difference $c$ against $\delta_5$. (c) Time variations of $\delta_5$ (dotted line) and $\omega$ (dashed line).
Table 4. Experimental Satellite Sensitivity to Dust $\beta$ (Counts)

<table>
<thead>
<tr>
<th>Location and Time</th>
<th>Photometric Data</th>
<th>Meteosat-IR Data</th>
<th>Data Processing</th>
<th>$\beta$ (Counts)</th>
</tr>
</thead>
<tbody>
<tr>
<td>Centre Agrhymet, Niamey, Niger, Feb. 1985, 1200 UTC</td>
<td>$\lambda = 0.45 \mu m$</td>
<td>B2</td>
<td>linear regression C versus $\delta_t$ and $t$</td>
<td>9.2 ± 1.9</td>
</tr>
<tr>
<td>Dakar, Senegal, April-May 1987, 1200 UTC</td>
<td>$\lambda = 0.45 \mu m$</td>
<td>full resolution B2</td>
<td>linear regression c versus $\delta_t$ and $w$</td>
<td>15.1 ± 2.2</td>
</tr>
<tr>
<td>Gao, Mali, Dec. 1986 to March 1987, 1200 UTC</td>
<td>$\lambda = 0.50 \mu m$</td>
<td>B2</td>
<td>stepped clean-air signal + linear regression c versus $\delta_t$</td>
<td>19.9 ± 3.1</td>
</tr>
</tbody>
</table>

Mean values and 95% confidence intervals are reported. Values of $\beta_t$ are obtained with $\delta_t \leq 1$ only.

The introduction of the fit of the date $t$ as a variable shows little seasonal effect on the results.

4.1.3. Gao measurements. Photometric measurements were performed between 1985 and 1989 in Mali (Sahel), with the aim of correcting satellite data for the aerosol radiative effect, for vegetation studies [Holben et al., 1991]. We use hereinafter the values of aerosol optical depth derived from the measurements at Gao (0.15°W, 16.32°N), at 1200 UTC, from December 1986 to March 1987 (dry season 1986-1987), at the wavelength 0.50 $\mu m$. This data set is compared to the Meteosat-IR data in the B2 format. The columnar water vapor amount over Gao for the corresponding period is provided by the European Centre for Medium-range Weather Forecasts (ECMWF) (Shinfield Park, Reading, England) from their initialized analyses, in a 1° square grid.

A question arising from the analysis of this 4 month long time series of satellite data deals with the definition of the clean-air signal, because of its obvious, but unknown a priori seasonal variations. In the current study, it is defined as a stepped function with a step of half a month. Each step level is defined as the highest level reached by the satellite signal $C$ during the period of concern. This definition coincides with the way the clean-air signal will be created in the image-processing technique of IDDI construction (section 5). As predicted, the day corresponding to this highest level shows an optical depth $\delta_t$ close to the lowest one observed in the period. Figure 8a shows the satellite signal $C$ of the 3X3 pixel array centered at Gao and the associated clean-air signal $C_0$. The optical depth $\delta_t$ is plotted for the sake of comparison. It shows a moderate mean level and a scattered aspect clearly synchronous with the fluctuations of $C$.

Figure 8b represents the linear least squares fit of the signal difference $c = (C_0 - C)$ against $\delta_t$. The vertical bars represent the standard deviation $\sigma$ for the 3X3 pixel array. The cases of cloud contamination are rejected according to the criterion of $\sigma > 6$ counts (a total of 26 cases, with 1 case in December, 5 cases in January, 9 cases in February, 11 cases in March). The satellite sensitivity to dust $\beta_t$ is 19.64 counts, and its 95% threshold confidence interval is [16.5, 23.0] counts.

Figure 8c shows the time variations of $\delta_t$ and of columnar water vapor amount $w$. The level of $w$ is moderately low on average, except at the very end of the period where it rises to 4 g cm$^{-2}$. The fluctuations of $w$ are significant during the whole period. A correction of $C$ to account for the water vapor effect, using the simulated results of section 3.3, yields weak changes, except for the few values of $w$ larger than 3 g cm$^{-2}$. Similarly, fitting $c$ against $w$ in addition of $\delta_t$ brings very small differences in the results, showing a negligible effect of $w$. However the errors on $w$ are probably significant (P. Källberg, personal communication, 2000) in this region where the network of meteorological stations is insufficiently dense to facilitate the control of ECMWF simulations; so further analyses will be necessary to confirm the current results.

4.2. Some Inferences From Simulations and Experiments

It is now possible to compare the theoretical and experimental results on dust remote sensing with Meteosat-IR imagery, for a Sahelian environment.

Firstly, the simulations indicate a quasi-linear decrease of the satellite signal $C$ with the increase of $\delta_t$. To make the results easier to present and discuss, we have approximated this law to a linear one, which does not involve large errors, provided the range of the values of $\delta_t$ is not very extensive. A similar linear behavior in the variations of $C$ against $\delta_t$ is revealed in Figures 5b, 7b, and 8b by the experimental data, after the extraneous effects of $w$ and $t$ on $C$ have been properly taken into account. As with any experimental data set, the scatter of the points results, at least partly, from the limited accuracy in the derivation of the data, including the determination of $\delta_t$, the calibration of the satellite data, the size of the pixels array, and the time lag between satellite data acquisition and photometric measurements. It also includes the effects of contamination by subpixel clouds which cannot be completely eliminated. In addition, it depends on the way the clean-air signal $C(0, w, t)$ is determined. On the other hand, it is also explained by (2), telling us that the satellite sensitivity to dust $\beta_t$ varies with ground surface properties and aerosol size distribution. While the surface properties can probably be considered as constant during the periods of study (an assumption perhaps more justified for Niamey and Dakar than for the 4 month data set of Gao), the size distribution of aerosol may have undergone changes, explaining a part of the scatter observed in the plots. Nevertheless the available scatter-plots suggest that these changes may have not been so dramatic as indicated by the results of the simulation (described in section 3). This may be due to an overestimation of the actual variability in particle size distribution by the models, or to a low actual variability arising from the fact that the measurements were performed at a single site and during a limited period of the year.

The experimental values of satellite sensitivity to dust $\beta_t$ are summarized in Table 4. For a better comparison with the theoretical values (Table 3), the values $\beta_1$ of $\beta_t$ limited to the values of $\delta_t$ between zero and 1, are included (except for Dakar where the number of measurements is too small). The sensitivity values obtained from the Niamey data set are clearly smaller than the values from Dakar and Gao. The differences between these values
are considered to arise from dust size distributions and from differences between the radiative properties of the surfaces at these sites. A significant result of the comparison of Tables 3 and 4 is that the ranges of values of the satellite sensitivity to dust derived from simulation and from experiment are quite consistent. Values between 8 and 25 counts must be considered as realistic (at least for a Sahelian environment, there is neither simulation nor experiment available for the Sahara).

The error due to the size of the 3X3 pixel arrays used in the experimental studies is estimated by the error bars shown with the scatterplots, representing their standard deviation σ. As predicted, this error is larger with the B2-sampled images (array size of 90 km at nadir) than with the full resolution data (15 km). Table 4 shows that replacing full resolution data by B2 ones for the site of Dakar results in a doubling of the confidence interval on the value of the satellite sensitivity to dust (from 2.2 to 4.3 counts). It can also be noted that the average of 19.0 counts with the B2 is 4 counts larger than the average with the full resolution. This fact stems mostly from the (unavoidable) shift inland of the B2 array, while the full resolution array is really near the coast (at the site of the photometric measurements).

For the experiments of Niamey and Gao the effect of water vapor on the satellite signal is negligible at the low levels of w typical of the Sahelian dry season, consistently with the simulations (Figure 3). This fact is more apparent for Gao, where the water vapor amount shows appreciable fluctuations (1.87±0.49 g cm⁻²) in the presence of moderate dust levels (0.41±0.34) than for Niamey where water vapor is approximately constant at a very low level (0.87±0.24 g cm⁻²) in the presence of a very high turbidity (1.27±0.94). In contrast, for Dakar the effect of the water vapor on the satellite signal is quite significant, and necessary to explain the signal variations (Figures 6b and 7b), even though the water vapor amount is at a lower mean level (1.38±0.82 g cm⁻²) than in the case of Gao. This fact, contrary to the simulations, relates to the coastal location of the site. The dry dusty air masses coming from the Sahara alternate with moister clean air masses arriving from the Atlantic (in the form of sea breezes). This alternation, illustrated by the anticorrelated variations of δ and w (Figure 6a), involves strong changes of w (standard deviation of 0.82 g cm⁻²). Contrary to the Saharan air masses, in equilibrium inside the continental environment, the oceanic air masses are cooler and out of equilibrium when they pass over the coast. This is the reason for the detection of their water vapor content by Meteosat IR. A similar but weaker effect was observed by N'dound [1992]. This algorithm is based on the simple assumption that the ranges of values of the satellite sensitivity to dust (counts), or alternatively of equivalent IR radiance or brightness temperature decrease. Such parameters must be extracted easily from raw Meteosat images for processing into the large image series required for climatological studies.

5. Image Processing Technique

5.1. Raw Satellite Data

A Meteosat-IR image of the Earth disk in full resolution contains 8 bit values for 2500X2500 pixels. To limit the amount of data to be processed and archived for climatological applications, sampled images in the B2 format, with 36 times fewer pixels (416X416), are used. A 10 year archive of daily images therefore requires 23 Gb with full resolution but only 634 Mb with B2 format. The latter format was created to support the International Satellite Cloud Climatology Project [Rossow et al., 1985]. Over a 17 or 20 day period the effect of the sampling procedure on the radiance average of 10X20 B2 pixel arrays is weak whatever the region investigated [Steere and Rossow, 1991, Table 4]. The difference between radiance averages with full resolution and with B2 never exceeds two counts (one count for North Africa and for the Sahel), including the clouds and surface contributions. Such effects are negligible for climatological results involving periods of a month or longer, derived from the IDDI product containing only data representative of atmospheric aerosols (clouds are masked and permanent surface features are eliminated). Joining the advantage of data volume compression to an acceptable spatial resolution, the B2 data are used to achieve a climatology of dust as described hereinafter.

The Meteosat-B2 images have been systematically archived by the European Space Operation Centre since April 1983, then relayed by Eumetsat (these data are currently delivered by Eumetsat, Darmstadt, Germany) since November 1995. The raw images are geometrically prerectified, so a given pixel corresponds to a unique geographical location. The raw radiometric counts can be corrected for the daily variations of sensitivity of the IR channel, using calibration coefficients associated to every image, as described by Legrand et al. [1989]. The corrected images are the so-called original images (OI).

5.2. Reference and Difference Images

The OIs are processed according to an algorithm reported for the first time by Legrand et al. [1985] and subjected to further analyses and improvements, as described by Legrand and N’doumé [1992]. This algorithm is based on the simple assumptions for the Earth-atmosphere system of a steady state of the ground surface in the presence of a varying atmosphere owing only to cloudiness and dust presence (the effect of water vapor is not considered). The presence of clouds and dust plumes is accompanied by a decrease of outgoing radiance, according to a time-varying pattern. This description is valid with daily images at a fixed time in the middle of the day, omitting the effect of the diurnal cycle. The technique consists of separating in the OIs, the steady contribution of the surface and the negative varying atmospheric contribution. The steady contribution is approximated
Plate 1. Illustration of the main steps of the image processing. An original image (top left) shows clearly clouds, colored in blue (cold) and white (very cold), and cloud-free areas comparatively warmer. The RI (top right) represents the corresponding cloud-free and dust-free images. The mountains and northern regions appear as cold areas. The DI (bottom left) shows a dust plume extending in the central Sahel over the Chad and the Niger Republics. Masking the clouds (in blue) separates the IDDI product, available only over the clear regions of the continent (bottom right).

Plate 2. Monitoring of a Saharan dust event in the second half of March 1988, using Meteosat [adapted from Legrand et al., 1994]. The images are composites, including oceanic and continental parts. The ocean scene shows visible aerosol optical depth ($\delta_\text{v}$) computed from Meteosat-VIS radiance. The continent scene shows infrared difference dust index (IDDI expressed in counts) derived from Meteosat-IR radiance. Over the ocean, clouds are replaced by interpolated $\delta_\text{v}$ over the continent they are not replaced by a mask (contrary to the IDDI full processing).
by the maximum radiance (or brightness temperature) met at
every OI pixel during a large enough period to reasonably ensure
the occurrence of at least one cloudless and dustless day for every
pixel. Hence a clear-and-clean (hot) reference image (RI) is con-
structed with the collection of these hottest values, picked out for
every pixel from the OI series covering the period.

This operation can be written out for a pixel \((i, j)\) and accord-
ing to the day number \(k\)

\[
P_{i,j}^r = \text{Max}(P_{i,j}^1, \ldots, P_{i,j}^n),
\]
with \(n\) being the number of days (with available satellite images)
constituting the reference period (superscript \(r\) means the refe-
rence value of \(k\) associated with the largest value of \(P_{i,j}^r\)).

Then, a difference image (DI), exhibiting the clouds and dust
plumes separated from the permanent surface pattern, is created
by subtracting every OI from its corresponding RI. Then

\[
P_{i,j}^d = P_{i,j}^r - P_{i,j}^1,
\]
with \(P\) and \(p\) holding for any relevant radiative parameter: radi-
ometric signal in counts, radiance through the IR channel, or
brightness temperature, and with \(P_{i,j}^r\), \(P_{i,j}^n\), and \(P_{i,j}^d\), representing,
respectively, these original, reference, and difference radiative
parameters.

Illustrating the previous steps, Plate 1 exhibits the OI, RI, and
DI covering Africa north of the equator and the Middle East, on
January 21, 1993, at 1200 UTC. The period used for constructing
the RI is half a month (January 16-31). A color scale with the
corresponding counts and brightness temperature is available for
the OI and RI. The nonlinearity between brightness temperature
and radiometric signal results in distinct corresponding DIs. The
DI displayed in Plate 1 shows the brightness temperature de-
crease.

An OI contains all the radiative information about surface and
atmosphere. The clouds are rather well distinguished due to their
brightness and structure. Conversely, dust plumes are not easily
discernable because their radiative effect is confused with the
pattern of the ground surface radiative effect.

The RI is a clear and clean image showing only the unvarying
surface contribution. The coldest surfaces (blue) lie in the north
and the hottest ones (red) tend to lie in the Sahel and eastern Sa-
hara. The mountains appear colder than the surrounding regions
(Ethiopia). The equatorial zone along the Gulf of Guinea and in
the Zaire Basin appears colder than the Sahel, owing to presence
of dense vegetation and large amount of atmospheric water va-
por.

The DI shows only the variable atmospheric radiative effects.
Clouds seen in the OI are restituted (over northwestern Africa, Saudi Arabia and the Middle East, northern Libya, Egypt, Ethiopia and Somalia). A strong dust plume, easy to discriminate from the clouds due to its smoothness and coherence, appears in the central Sahel, over Chad and Niger. Lake Chad is visible (in the RI and OI) south of the plume. Dust absence NNW of the plume reveals the Tibesti massif. Some dust is also apparent in central Sahara and Senegal, or partly obscured below clouds in northwestern Africa, Egypt, and Saudi Arabia.

5.3. Cloud Masking

The information content of a DI relates both to clouds and to dust. The next step in the processing therefore consists of identifying clouds in the Dis and masking them. Because of elimination of the permanent surface pattern in the Dis, clouds are observed over continental regions against a smooth, somewhat ocean-like, background, so it is more convenient to identify clouds in Dis than in OIs. The technique used, related to the spatial coherence method [Coakley and Bretherton, 1982] relevant to the detection of clouds over ocean in the thermal infrared, is therefore suited to this task. The operation of cloud identification is automated through an algorithm designed to be straightforward and fast. It is adapted for identifying clouds in a scene where they coexist with dust plumes.

The technique of cloud identification can be described considering the arch diagrams in Figure 9. They are made up using 3X3 pixel arrays. Every point represents the mean and the standard deviation of a given array. A complete set of running arrays fills a 12X12 pixel box, covering an area of (360X360) km$^2$ at nadir. The basic technique consists in separating the clear (or hot) foot of the arch, located near the origin, of the fully or partly cloudy points. This is realized by the algorithm which determines cutoffs ($c_r$, $\sigma_r$) delimiting the clear foot domain, as illustrated in Figure 9a. In Figure 9b the scatterplot is limited to the clear foot of the arch, and the whole 12X12 pixel box is declared clear. The case of a cloudy arch with no clear foot (Figure 9c) is correctly interpreted by the algorithm. Even though dust clouds appear smooth, some structure exists, especially in association with the dust front. In the diagram, a dust front is represented by an arch (Figure 9d), which is correctly interpreted as dust by the algorithm (cloud arches are higher than dust arches). When this first step is over, every box is associated with a set of cutoff values pairs ($c_r$, $\sigma_r$). In a final step, the DI is scanned and the mean and standard deviation of every 3X3 pixel running array are compared to the relevant cutoffs ($c_r$, $\sigma_r$). As a result the central pixel of the array is declared clear (if $c < c_r$ and $\sigma < \sigma_r$) or cloudy. In the latter hypothesis, the radiometric value is replaced by a mask-coding value. Misinterpretation from the algorithm may occur in a few particular or intricate cases. For instance, for a very steep dust front, or for a dust front associated with broken clouds having a size comparable with the 3X3-pixel arrays, dust can be erroneously classified as cloud (such errors cover dust fronts) and 12X12 pixel boxes can be filled incidentally with cloud masks. Few data are significantly affected by these casual errors whose impact is weak after climatological processing. A visual syste-

![Figure 10](image)

**Figure 10.** Monthly frequency of cloud occurrence (in percent) averaged over the period 1984-1993, derived from cloud masking applied in the IDDI processing. The gray scale increases linearly with the frequency of cloud occurrence. A geographical latitude-longitude projection of the images is applied. The covered region lies from the equator to 35°N and from 18°W to 47°E. The resolution is a square degree.
natic control of the daily IDDI images of 1986 has shown that 15% of the images are affected by such errors, resulting in 0.15% of the IDDI data set.

In Plate 1, the bottom right-hand image illustrates the resulting cloud discrimination obtained from application of the algorithm to the DI (bottom left-hand image). Figure 10 shows the monthly frequency of the presence of cloud masking, averaged over the period 1984-1993, derived from the algorithm described above. It describes the seasonal pattern of frequency of cloud occurrence. The main feature is the annual oscillation of the cloudy belt associated with the ITCZ, from the Gulf of Guinea in January to the southern Sahara in July and August. If this seasonal pattern of cloudiness is set against the seasonal pattern of water vapor shown in Figure 4, they are observed to match satisfactorily. Thus over these areas and during the seasons characterized by high water vapor and cloud amounts, it can be predicted that the quality of the derived IDDI data will be degraded, due to few clear days available for the construction of the RI and to IDDI contamination by humidity. However, it is important to emphasize that such conditions are also unfavorable for the presence of atmospheric dust.

6. Infrared Difference Dust Index (IDDI)

6.1. Theoretical Definition and Practical Realization

The unmasked part of the continent displayed in the bottom right-hand image of Plate 1 is an example of the IDDI product. It represents the decrease in the Meteosat-IR signal relative to the reference signal, in counts or alternatively in the corresponding outgoing radiance or brightness temperature (Plate 1 shows the brightness temperature decrease).

To what extent this operational definition resulting from the image processing coincides with the ideal definition of the IDDI to be the dust-induced satellite response over clear continental regions is a crucial point that must be addressed. With this in mind we examine the consequences of the various processing stages for the ability of the IDDI to detect a signal arising from mineral dust and to discriminate it between those originating from other sources.

Theoretically, the RI most relevant to dust detection would be made of the modified outgoing radiance resulting from the removal of mineral aerosol, keeping the radiative effects of non dust aerosols and water vapor (and adding the cloud masking). It is different from the aforementioned clear and clean (and dry) image (section 5.2), associated with an absolutely transparent atmosphere in the Meteosat-IR channel (one containing no cloud, no aerosol, no water vapor). However, if we limit the study to Africa north of the ITCZ, excluding in addition the coastal regions (section 4.2), the effect of water vapor is negligible. There, these two definitions of the reference differ only in the presence or absence of nondust aerosols. In the real Saharan atmosphere, mineral aerosol is ubiquitous and the other components (stratosphere and upper troposphere aerosols) are marginal, so the overall difference between both definitions of reference image is of little significance. In the Sahelian atmosphere north of the ITCZ, we expect more significant differences, with regard to nonmineral aerosols such as biomass burning emitted from the savannah at the onset of the dry season, and with regard to possible water vapor effects induced by oscillations of the ITCZ. In this region, an ideal RI would retain the aerosol component (possibly significant) representing the nondust content of the atmosphere. The actual RI approaches the ideal condition more closely in the Sahara than in the Sahel.

| Table 5. Frequency of Dust Contamination According to Period Length |
|-------------------|-------------------|
| Period Length (weeks) | Dust Contamination (%) |
| 1                 | 23                |
| 2                 | 13                |
| 3                 | 10.5              |

The contaminated periods are those with no visibility ≥ 10 km on the cloud-free days. Statistics from 43 stations.

6.2. Errors

A difficult question relating to the construction of the RI is the choice of a suitable length for the reference period [Legrand and N’doumé, 1992]. The longer this period, the cleaner the RI is expected to be. However, this procedure cannot be fully applied because the reference level varies continuously all year-round because of (1) changes in the relative positions of the Sun and the Earth, (2) seasonal changes of the land surface, especially in terms of vegetation cover in the Sahel, and (3) seasonal atmospheric evolution. If the period is too short, the RI can be strongly and frequently contaminated by appreciable amounts of dust and occasionally by cloud residues [N’doumé, 1993]. If the period is too long, the RI may be affected by the long-term (seasonal) shift of reference level. The best result will be obtained using a period of some intermediate length.

Data analyses comparing frequencies of cloud and of dust occurrence with seasonal rates of variation in the reference level derived from the RI time series show a posteriori that a reference period 2 weeks long is generally adequate. Table 5 shows the percentage of "poor" reference periods according to their length, for 43 Sudanian, Sahelian, and Saharan meteorological stations (Figure 13), during the months of the dry season in 1984. Dust is revealed by visibility routinely measured at ground level and is assumed to be present below a 10 km visibility threshold [N’tchayi et al., 1994]. Cloud presence is estimated by cloud masking of the IDDI images. The second column of Table 5 shows the percentage of unsatisfactory periods (i.e., without any clear day having a visibility value ≥ 10 km). It can be noted that the percentage of 13% for a period length of 2 weeks is not negligible and that extending the period length to 3 weeks does not improve substantially this result. This may be related to cases of stagnancy of dust-transporting harmattan air masses north of the ITCZ.

These results can be compared to the estimate of residual aerosol optical depth δ, affecting an RI 2 weeks long, based on photometric measurements made during the dry season, between 1985 and 1989, at the Malian stations of Gao, Barnako (7.98°W, 12.64°N), Gossi (1.33°W, 15.80°N), and Sevare, near Mopti (4.17°W, 14.48°N) [Holben et al., 1991]. A value of δ ≥ 0.4 is observed for 16% of a total of 186 periods. This result is not very different from the 13% of contaminated periods (visibility < 10 km). Both results are consistent, since a visibility of 10 km corresponds to an optical depth between 0.3 and 0.5 according to the various empirical relationships found in the literature [N’tchayi et al., 1994]. Both results confirm that an RI may retain a significant amount of aerosol, including dust, in the Sahelian zone.

The reference signal increases or decreases according to the season and depending on the location of the site considered, as illustrated in section 4.1. This signal increases steeply in February at Niamey (Figure 5a), according to a rate s = 0.73 counts/day.
periods definitely exceeds 10 counts (error equivalent 5 s of 0.55, rainy season. In several cases the shift between two consecutive April and August) and of the African monsoon giving rise to the effects of solar insolation (Sun passes at zenith twice a year, in the seasonal shift of the reference level is estimated by subtracting the actual reference signal, approximated by linear interpolation of the signal at the reference dates, from the stepped reference level created by the IDDI algorithm (dotted line). The error due to a large extent, through averaging of large numbers of individual images. On the other hand, it should be stressed that the residual aerosol remaining in the RI will result in a systematic underestimate of the IDDI. The case study of Gao, 1986, shows that the aerosol residue has an optical depth of 0.28±0.20, to be compared to the mean optical depth and its standard deviation of 0.73±0.57 and to the highest measured values over 2.5. This residual and the error it involves on the IDDI are significant, even though it contains a (unknown) nondust fraction. Analyzing this data set shows that this is due, to a large extent, to a high background component, as shown in the work of Holben et al. [1991, Figure 2]. To a lesser extent, this results also from the high frequency of cloud presence (47% throughout the year, 39% during the dry season, 62% during the rainy season). The significance of the residual aerosol error in the IDDI for the case study of Gao agrees with the high level of turbidity in the Sahel, reported in the literature with values of optical depth, Ångström turbidity parameter, and visibility [Cerf, 1980; d’Almeida, 1986; Holben et al., 1991; Ben Mohamed et al., 1992; N’tchayi et al., 1994]. This level is particularly high at Gao as shown by Holben et al. [1991, Table 2] and N’tchayi et al. [1994, Table 3]; the latter shows, in addition, that turbidity at Gao has been regularly and strongly increasing since the 1950s. This high level of turbidity is in agreement with our IDDI database, showing that this station lies close to a dust source particularly active at end of spring and beginning of summer [Legrand et al., 1994, Figures 4 and 5]; so the level of error obtained for Gao, 1986, must likely be considered as a maximum.

Figure 11. Seasonal cycle of the reference level for the station of Gao during 1986. The reference approximated by linear interpolation of the signal at the reference dates (solid line) is compared to the stepped reference level created by the IDDI algorithm (dotted line). The dashed stripe indicates the rainy season.
Atlantic derived from Meteosat-VIS imagery were used for this purpose. Details on these studies can be found in the work of Legrand et al. [1994].

7.1. Relation IDDI-Visibility

The meteorological visibility is one of the elements identifying air mass optical characteristics. By day, it is defined as the greatest horizontal distance at which a black object of suitable dimensions, located near the ground, can be seen and recognized when observed against a background scattering of sky, haze, fog, [WMO, 1996]. Thus the reduction of meteorological visibility measured in the meteorological stations has been widely used to estimate the local dust amount [Bertrand et al., 1973; Middleton, 1985; d’Almeida, 1986; Ben Mohamed and Frangi, 1986; N’chayi et al., 1994, 1997] and to map the footprint of dust plumes at ground level [Legrand et al., 1985; Legrand, 1990]. In addition, visibility observations and photometric dust optical depth are correlated (a short review is presented by N’chayi et al. [1994]), showing that visibility is suitable to estimate the optical depth of a dust event. It can therefore be used for statistical comparisons with the IDDI, its interest being that it is routinely measured in the many meteorological stations of the African synoptic network. However, it is necessary to be aware of some shortcomings of the visibility: (1) the measurements of visibility are of limited accuracy; (2) a fraction of observed low visibility values is associated with events such as smoke, fog, or local pollution without relation with dust events; (3) the visibility does not reveal the presence of dust for layers in altitude (occasionally reported by pilots). A discussion on the application of visibility to estimate the magnitude of dust haze in West Africa can be found in the work of N’chayi et al., [1994].

Visibility data from 1984 at 1200 UTC for 43 stations located in western and central Africa (Figure 13) are used statistically to perform the validation and assess the representativeness of IDDI as a dustiness parameter for climatological purposes. Most of the stations are Sahelian and Sudanian, some are Saharan.

For validation purposes, we have selected a subset of 12 stations lying in an area ±6° off Niamey in longitude and ±2° in latitude (frame in Figure 13), for a comparison with the simulations described in section 3 and the experiments described in section 4, which relate to this area (Niamey is at the center, Gao lies at its northern limit, only Dakar is outside). The selected measurements are restricted to November for the sake of comparison with the simulations.

Visibility measurements are compared with IDDI values for 3X3 pixel arrays centered on each station. Uncleared, i.e., partly and fully cloudy, cases have been eliminated from the statistics. In addition, a criterion of quality has been applied to the RIs. A minimum number of 3 days cloudless and dustless is prescribed for a RI period to be acceptable for the statistics, otherwise it is rejected. For application of this criterion, cloudiness was determined from IDDI masking and nondusty cases were associated with measured visibility over 10 km. Because of this strict selection, 71% of cases have been removed from the initial data set of 360. The selected data have been classified into four categories according to the range of visibility, and the means and medians (very similar) are reported in Figure 14 for every class. The fitted curve describes the empirical relation between visibility V and IDDI. Table 6 shows the values of IDDI derived from Figure 14 (taking the median) for the values of visibility of 25, 10 and 5 km. In addition, the various empirical relations between V and δt [N’chayi et al., 1994] are summarized in Table 6 as a range of optical depth associated with each visibility value.
Figure 13. Region covered by the IDDI product (indeed the satellite images extend farther eastward, showing Ethiopia, Somalia, and most of the Arabian Peninsula). The meteorological stations whose data are used in the study are plotted. The subset of 12 stations used for the IDDI validation is framed within a rectangle centered at Niamey (labeled N). The meteorological equator, northern limit of the ITCZ, is shown in its northmost situation in July and in its southmost situation in January (from Leroux's [1983] Atlas).

In Figure 15 the relation between aerosol optical depth and IDDI derived from the previous approach is plotted, including variability bars corresponding to the aerosol optical depth intervals reported in Table 6. The simulated results found in Table 3, corresponding to the minimum and maximum values of satellite sensitivity to dust, as well as the values presented in Table 4, resulting from the experiments of Niamey, Dakar, and Gao are reported in Figure 15. This synthetic drawing shows the general consistency resulting from the use of visibility, between the IDDI sensitivity and the satellite sensitivity to dust assessed from prior simulations and experiments relevant to the infrared method.

The IDDI-visibility relation shown in Figure 14, established for the region of Niamey during November, must not be considered a priori representative for the whole region of concern (displayed in Figure 13) and throughout the year. Thus the whole visibility data set of 1984 from a network of 43 stations is used and the data to be compared, IDDI and visibility, are processed as in the previous step of IDDI validation to determine a relation of climatological value. Owing to the strict selection applied, 70% of the cases have been removed from the initial data set of 12,152. The IDDI visibility relationship, of climatological relevance for West Africa and over the whole year, is shown in Figure 16. Both Figures 14 and 16 are very similar, but they reveal also some differences. For a given value of visibility the IDDI values are larger in Figure 16 than in Figure 14, by 1-3 counts. We can summarize from these results, climatologically valid for West Africa, that visibilities of 10 and 5 km correspond, respectively, to IDDI values of 10 and 15 counts (instead of 9 and 12.5 counts in Figure 14). The shape of the curve in Figure 14 is smoother than for the climatological case of Figure 16. Instead, the latter shows distinct aspects on both sides of the visibility value of 10 km. In addition, the median is distinctly larger than the mean, especially with the highest values of visibility, indicating an asymmetric distribution of IDDI in each class. These features may reflect the large variety of physical and geographical condi-
Table 6. Empirical Correspondence Between Visibility $V$ and IDDI. Statistics from 12 Sahelian Stations in the Region of Niamey, for November 1984

<table>
<thead>
<tr>
<th>$V$ (km)</th>
<th>IDDI (Counts)</th>
<th>$\delta_3$</th>
</tr>
</thead>
<tbody>
<tr>
<td>25</td>
<td>4.5</td>
<td>0.15-0.3</td>
</tr>
<tr>
<td>10</td>
<td>9</td>
<td>0.3-0.5</td>
</tr>
<tr>
<td>5</td>
<td>12.5</td>
<td>0.6-0.9</td>
</tr>
</tbody>
</table>

The correspondence between visibility and aerosol optical depth $\delta_3$ is derived from N'tchayi et al. [1994].

7.2. Satellite Observation of Dust Over Continent and Ocean

Over the ocean, the presence of aerosols increases the upward SW radiance by scattering the downward propagating SW radiation. Because of the darkness and uniformity of the ocean surface, this effect can be measured and modeled to retrieve values of optical depth. It has been used to map aerosol optical depth over ocean regions for climatological applications. Rao et al. [1988] were the first to derive a global aerosol distribution over the ocean from NOAA satellite data. More recently, Moulin et al. [1998] derived a dust climatology over the Mediterranean from Meteosat-VIS data. The latter satellite was also used to monitor Saharan dust outbreaks over the Atlantic and to construct a dust climatology of the tropical Atlantic [Jankowiak and Tanré, 1992]. Therefore dust plumes expelled over the Atlantic from the African continent can be viewed using a combination of continental IDDI imagery and oceanic aerosol optical depth derived from Meteosat-VIS imagery. The technique consists of merging the IDDI product over the continent and Meteosat aerosol optical depth over the ocean into a single composite image. An example is shown in Plate 2 of a huge dust expulsion across the coasts of Mauritania, Senegal, and Guinea, over the eastern Atlantic, at the end of March 1988. The construction of a single image from both products, derived from IR and VIS images of the same satellite, is straightforward. On the other hand, it is necessary to fit the color codes used with each product, to obtain the best visual agreement, or in other words, to get the best continuity of the structures across the coast. As shown in the legend of Plate 2, the IDDI and the aerosol optical depth are observed to match satisfactorily using a proportional law with a ratio of IDDI to optical depth of 24 counts (i.e., $\delta_3 = 1$ corresponds to IDDI = 24 counts, at the top of the range of sensitivity to dust $\beta$ derived from simulation). However, this agreement is approximate (e.g., the image of March 28 exhibits discrepancies along the Mauritanian and Guinean coasts).

8. Conclusions and Recommendations

The so-called infrared method applied to the IR-channel data of Meteosat is proven to be relevant for estimating, over land and during the middle of the day, the amount of dust originating from sources lying in the arid areas monitored by the satellite. The satellite IR signal relates to the aerosol SW optical depth, and this relation is characterized by a strong quasi-linear satellite response and hence by a large satellite sensitivity to dust. Because of its physical dependence on dust size distribution and on the radiative characteristics of the underlying surface, the sensitivity to dust can vary between 10 and 25 counts in a Sahelian environment. The effect of water vapor, if uncorrected, is a source of error limiting the regions where the method can be reliably applied, to the Sahara and the Sahel north of ITCZ. However, these regions where dust is generated and transported away from the sources,

![Figure 15](image-url)

Figure 15. Experimental relation between IDDI and aerosol optical depth for the region of Niamey, in November (thick solid curve). The curves corresponding to extreme values of simulated satellite response to dust (Table 3) are plotted as a function of aerosol optical depth for Niamey in November (dotted lines); the slight nonlinearity shown by simulation is included. The curves from photometric measurements at Niamey, Gao, and Dakar are plotted (from Table 4) as regression of IDDI against aerosol optical depth values between zero and 1 (thin solid lines).
are those of greatest interest for a number of applications. Care should also be taken to the "edge effect" of oceanic air masses advected inland along the coast. All these results were established using simulations as well as experimental data, two radically different approaches resulting in consistent relationships and conclusions.

The IDDI is derived from the infrared method. It is a dust product processed using Meteosat-IR imagery, designed for climatological applications (i.e., with averaging and statistical computations on a sufficiently large number of images). The reliability of the IDDI for dust remote sensing is shown using visibility data on a statistical basis; the values of IDDI sensitivity to dust agree with the satellite sensitivity to dust derived from the previous simulations and experiments dedicated to the infrared method.

However, users of IDDI data should be aware of limitations resulting from the processing of reference images. The removal of dust from the RIs is not always entirely successful. In 1984, over West Africa, 13% of the RIs are contaminated by residual aerosol, optically characterized by visibility \( < 10 \text{ km} \), i.e. \( \delta_+ > 0.4 \). The seasonal modulation of the reference level introduces errors into the RIs, especially during periods and over zones experiencing a steep seasonal evolution of brightness temperature of the Earth-atmosphere system. In 1986 at the Sahelian site of Gao, 7% of the RIs are contaminated by an error equivalent \( \delta_+ > 0.4 \) or \( \delta_- < -0.4 \).

The error due to reference level shift is unbiased, so it will vanish through statistical processing and it will be generally insignificant in the climatological products. In addition, this error could be efficiently reduced in the individual images through the processing of a running reference (instead of the current reference realized on a fixed 2 week period).

The error due to aerosol residue will spread to climatological products in the form of a bias (but the error derived from the case study of Gao must correspond to a maximum). What is more, it cannot be eliminated easily. A way to reduce this error would be to lengthen the reference period, which requires a reliable correction of the seasonal variations of the reference (in addition this would reduce the error of reference shift).

The easiest applications which can be realized using the IDDI concern the sources of dust emission, their identification, the description of their seasonal activity, of their year-to-year change and long-term evolution (to be addressed in the second part of this paper). The published results (reported in section 1) using IDDI statistically for validation of the physics of dust emission and for the retrieval of the source wind speed threshold fall into this category. In this case, satellite response is especially large, because of the high dust amounts emitted by the sources and of the particular efficiency in arid environment of the physical processes involved (section 2.1). In addition, several possible factors of contamination of the satellite response such as water vapor effect, high frequency of cloud occurrence, and seasonal change of the state of the surface (through vegetation) affecting the Sahel are absent in the source areas.

Dust is a major component of the climate over large regions of the planet, Africa north of the equator probably being the most dramatic example. The IDDI makes possible the incorporation of this component into climate studies as a whole, in order to analyze its interactions with precipitation, wind, cloudiness and atmospheric temperature, as well as its possible relationships with the atmospheric circulation. Again, the IDDI is suitable for statistical studies of this type, provided the results for regions and periods with large cloudiness and atmospheric humidity can be avoided or are considered with caution.

Pending the development of more sophisticated algorithms, the use of the IDDI for case studies of dust emission and transport is quite possible but requires more care with regard to the possible large errors associated to RI making. These difficulties can, however, be overcome (1) by multiplying the studied cases and (2) by using intense dust events (minor events can be significantly enhanced or deleted due to the errors involved in the IDDI processing technique). Quality control of the product is possible, considering the frequency of cloud masking, the rate of change of the reference level, and when available, extraneous data such as photometric optical depth.

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References

Ackerman, S. A., Using the radiative temperature difference at 3.7 and 11 \( \mu m \) to track dust outbreaks, Remote Sens. Environ., 27, 129-133, 1989.


