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## Thermal Impact of Saharan Dust over Land. Part II: Application to Satellite IR Remote Sensing

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### ABSTRACT

The use of the mesoscale model described and qualified in Part I is arranged with radiative transfer codes for the simulation of the thermal infrared response of Meteosat from a Sahelian target. The sensitivity of the satellite response to various atmosphere and surface parameters, either relevant or extraneous to dustiness, is analyzed and physically interpreted throughout the daily cycle, considering especially the thermal impact of the dust at the ground surface. The most significant parameters, according to this criterion of sensitivity, are the amount of dust in the atmosphere and its radiative characteristics, and the ground surface emissivity in the satellite channel. If neglected, the atmospheric water vapor content may be a large source of error for the retrieval of dustiness from the satellite data. The theoretical results are discussed and compared with earlier published experimental work.

### 1. Introduction

The impact of the Saharan dust on the land temperature is a mechanism that has to be considered to explain the efficiency of the thermal infrared radiometry from satellite for remote sensing of dust over continental areas. A further reason is the screen effect, modifying the upward transfer of radiance emitted by the ground through the dust layer, when the ground and the dust have appreciably different temperatures.

Thus, the satellite infrared dust detection may by no means be considered as a result of simple processes, and it can be expected that a large set of parameters is needed for the purpose of expressing the satellite response.

Ground measurements have been performed at Sahelian sites during the dry season in order to tentatively fit the measured parameters with the Meteosat 10.5–12.5- $\mu\text{m}$  channel data. The results from these in-

vestigations have been reported by Legrand et al. (1989) and by Legrand and Desbois (1989), and are summarized in Table 1. For considered Sahelian sites during the dry season, the Meteosat thermal infrared data from the middle of the day have been approximated with acceptable accuracy by means of both the visible aerosol optical depth and the atmospheric water vapor amount for the site of Dakar, and the day number accounting for the seasonal effect for the site of Niamey. The visible optical depth has been preferred to the infrared optical depth or the dust amount, because of its experimental accessibility and its radiative importance to the climatic impact of the dust. Such results demonstrate that the optical depth could conversely be retrieved using thermal infrared satellite data and other parameters that differ according to the site and the season. Prior to a tentative generalization of experimentally derived relations to the whole arid and semiarid area constituted by the Saharan desert and its borders, it is of prime importance to identify the parameters whose variations have a strong impact on the satellite response value.

This study is a first attempt at getting some insight into this topic by testing the sensitivity of various parameters with model simulations. First, the modified Colorado State University (CSU) mesoscale model described and qualified in Part I (Cautenet et al. 1992) is used so as to simulate the thermal impact of the dust. The output atmospheric profiles and surface temperatures are then inserted into programs that are suitable

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TABLE 1. The summary of the statistical results derived from the regression of daytime Meteosat IR data upon "ground truth" photometrically obtained data. The linear or quasi-linear fits of satellite response  $C$  with ground parameters are realized with a correlation coefficient  $r$  and a residual standard deviation  $\Delta C$  corresponding to an equivalent optical depth error  $\Delta\delta_s$ . Reported intervals of values of  $r$ ,  $\Delta C$ , and  $\Delta\delta_s$  refer to the various values obtained according to the time of the day and the Meteosat image format. A radiance variation of  $50 \text{ mW m}^{-2} \text{ sr}^{-1}$  roughly corresponds to a 1-count variation.

Site and period	Ground measurements		Statistical results		
	Geographic and climatic features	Parameters	$r$	$\Delta C$ (count)	$\Delta\delta_s$
Niamey, Republic of Niger, February 1985	Sahelian site far inland; dry season, strong seasonal warming	aerosol optical depth at $0.45 \mu\text{m}$ (photometer); day number (i.e., time at seasonal scale)	0.98	1.3–2.3	0.1–0.2
Dakar, Senegal, April 1987	Sahelian site, near seashore; dry season, negligible seasonal warming	aerosol optical depth at $0.45 \mu\text{m}$ (photometer); vertical columnar amount of water vapor (photometer)	0.94–0.98	2.0–3.8	0.12–0.24

for the calculation of the radiative transfer through the atmosphere in order to compute the resulting radiance, which would be detected by the Meteosat 10.5–12.5- $\mu\text{m}$  channel. The amount of dust and its vertical distribution are successively varied in order to evaluate their respective impact on the satellite response, and the importance of the radiation properties of dust is assessed by comparing the satellite responses for two markedly different dust models. Then the impact of the surface properties of the target is investigated, and finally atmospheric variable parameters, extraneous to dustiness, such as humidity and wind, are varied in order to estimate the spurious variations they could induce on the satellite response.

## 2. The models

### a. The CSU mesoscale model

The version of the modified CSU mesoscale model of Mahrer and Pielke (1978) is run with data issued from the Etude de la Couche Limite Atmospherique Tropicale Sèche (ECLATS) experiment (Fouquart et al. 1987), depicting the general conditions for the Sahelian site of Niamey (Republic of Niger) during the second half of November (dry season). Every run is performed with a cloudless atmosphere in which the forcing by a dust layer vertically homogeneous and constant with time extending from the ground to some fixed height has been inserted. The reference parameters are gathered in Table 2 and the reference profiles, initialized at sunrise (0650 LST), are plotted in Fig. 1.

Most of the tests of sensitivity are defined by the modification of a single (several in only a few cases) parameter or profile. The initial atmospheric, surface, and soil temperatures have to be tuned so that the model outputs are approximately stationary (i.e., the initial conditions are restituted after a 24-h run).

### b. The radiative transfer codes

For an aerosol considered as nonscattering in the infrared range, as the ECLATS-type aerosol (see Table 2), the LOWTRAN-5 code (Kneizys et al. 1980) is used to compute the radiance incident on the

TABLE 2. Reference parameters used with the modified mesoscale CSU model, representative of a Sahelian site during the dry season (Niamey, second half of November). SW refers to the shortwave range (0.25–4.0  $\mu\text{m}$ ) and LW to the longwave range (4–50  $\mu\text{m}$ ).

Parameters	Values	Comments
General		
Ground level	230 m	flat
Date	26 November	
Latitude	14°N	
Surface		
Soil thermal inertia	1000 SI units	after Cautenet et al. 1986
Roughness length	1 mm	
SW albedo	0.26	from ECLATS experiment
LW emissivity	0.9	gray
Initial and synoptic		
Sea level pressure	1013 mb	
Surface temperature	289 K	
Surface specific humidity	$3 \times 10^{-3}$	
Soil wetness	$10^{-2}$	very dry
Dust (ECLATS model)		
Layer geometrical depth	1.2 km	
Layer SW optical depth	1	
SW single-scattering albedo	0.952	after Fouquart et al. 1987
SW asymmetry factor	0.655	id.
Layer LW optical depth	0.12	id.
LW single-scattering albedo	0	id.

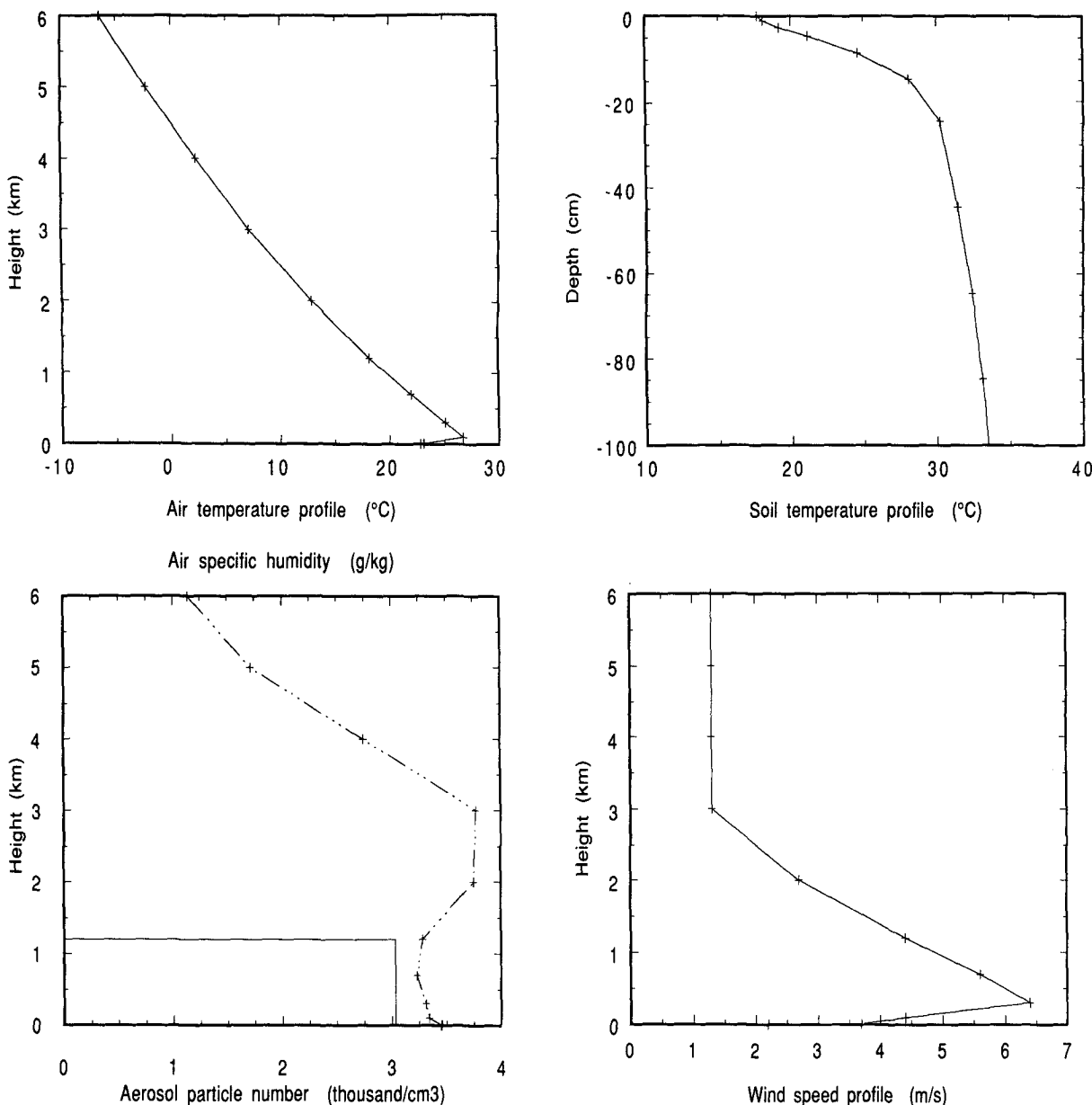


FIG. 1. (a) Initial reference profile of air temperature for the CSU mesoscale model (typical of a Sahelian site at sunrise during the dry season). (b) Same as (a) for the temperature in the soil. (c) Same as (a) for specific humidity (dash-dot line) and aerosol particle number density (solid line). (d) Same as (a) for wind speed.

*Meteosat-2* detector of the 10.5–12.5- $\mu\text{m}$  channel, inserting as input data the ground temperature and atmospheric profiles from the output files of the CSU mesoscale model.

The filtering function of the infrared channel has been introduced in the program for radiance computation. In addition, the computed radiance is digitized into radiometric counts of a satellite signal using a conversion constant of  $50 \text{ mW m}^{-2} \text{ sr}^{-1}$  per count (results can then be easily compared with previous experimental results of Legrand et al. 1988 and Legrand

and Desbois 1989). Moreover, the program has been adapted so as to account for the radiance reflected by a ground surface with an emissivity smaller than 1.

For a scattering aerosol, as considered in section 3a, the method of the successive orders of scattering (see Lenoble 1985) is used. In order to ensure the consistency with nonscattering aerosol results issued from LOWTRAN-5 code, the gaseous optical depth of the atmospheric sublayers has been first computed with the LOWTRAN-5, then inserted into the successive orders of scattering code.

TABLE 3. Impact of a thick aerosol layer ( $\delta_s = 4$ ) on radiation, surface temperature, and satellite response;  $F_{s\downarrow}$  and  $F_{l\downarrow}$  are the downward shortwave and longwave irradiances at ground level,  $T_S$  is the land surface temperature,  $C$  and  $T_{sat}$  are the radiometric counts and corresponding equivalent brightness temperature of the satellite response. Values between parentheses indicate the percent variations relative to the dustless values.

Time (LST)	$\Delta F_{s\downarrow}$ ( $W m^{-2}$ )	$\Delta F_{l\downarrow}$ ( $W m^{-2}$ )	$\Delta T_S$ ( $^{\circ}C$ )	$\Delta C$ (count)	$\Delta T_{sat}$ ( $^{\circ}C$ )
1200	-535 (-58%)	+81 (+26%)	-10.8	-21.8	-10.0
0600		+80 (+28%)	+6.5	+18.1	+9.5

### 3. The tests of sensitivity

Twenty-four-hour runs are performed with the CSU mesoscale model by varying the parameter of interest among the reference conditions of Table 2 and Fig. 1, and hourly outputs are collected. Then radiances in the 10.5–12.5- $\mu m$  Meteosat channel are computed with the previous outputs at 3-h intervals.

#### a. Sensitivity to dust

##### 1) AMOUNT OF DUST

Varying the amount of atmospheric dust consists of changing at the same ratio the values of the optical depths spectrally averaged on the shortwave and longwave ranges,  $\delta_s$  and  $\delta_l$ , respectively. These optical

depths are of climatological relevance since they concern the broadband irradiances, and they are introduced into the CSU mesoscale model. Moreover, the infrared optical depth  $\delta_{sat}$ , restricted to the Meteosat channel width, is varied at the same ratio. This optical depth affects only the radiative transfer of the narrowband radiance and is introduced into the relevant code (LOWTRAN, successive orders of scattering). For an aerosol of the ECLATS type, the relation

$$\delta_l = \delta_{sat} = 0.12\delta_s \quad (1)$$

is used. For this aerosol, a unit shortwave optical depth corresponds to a total volume of dust of 0.0036  $cm^3$  in a vertical column having a 1- $cm^2$  section; that is, a dust content of 0.009  $g cm^{-2}$  with a mean density of 2.5  $g cm^{-3}$ .

Table 3 illustrates the importance of the modifications involved at 0600 and 1200 LST by a thick dust layer with a  $\delta_s$  value of 4. The thermal impact at the ground surface is presented in Fig. 2a during the daily cycle, for values of  $\delta_s$  varying from 0 to 4, and in Fig. 2b against  $\delta_s$  at various times during both day and night. The presence of dust results in a ground surface temperature  $T_S$  lower during the day and higher during the night. As described in Part I (Cautenet et al. 1992), the nighttime effect results from the dust-induced increase of the atmospheric irradiance at the ground surface. The daytime effect results from the dust-induced decrease of the solar irradiance at the ground surface (prevailing over the increase of the atmospheric irradiance). The variations of  $T_S$  are large, agreeing with

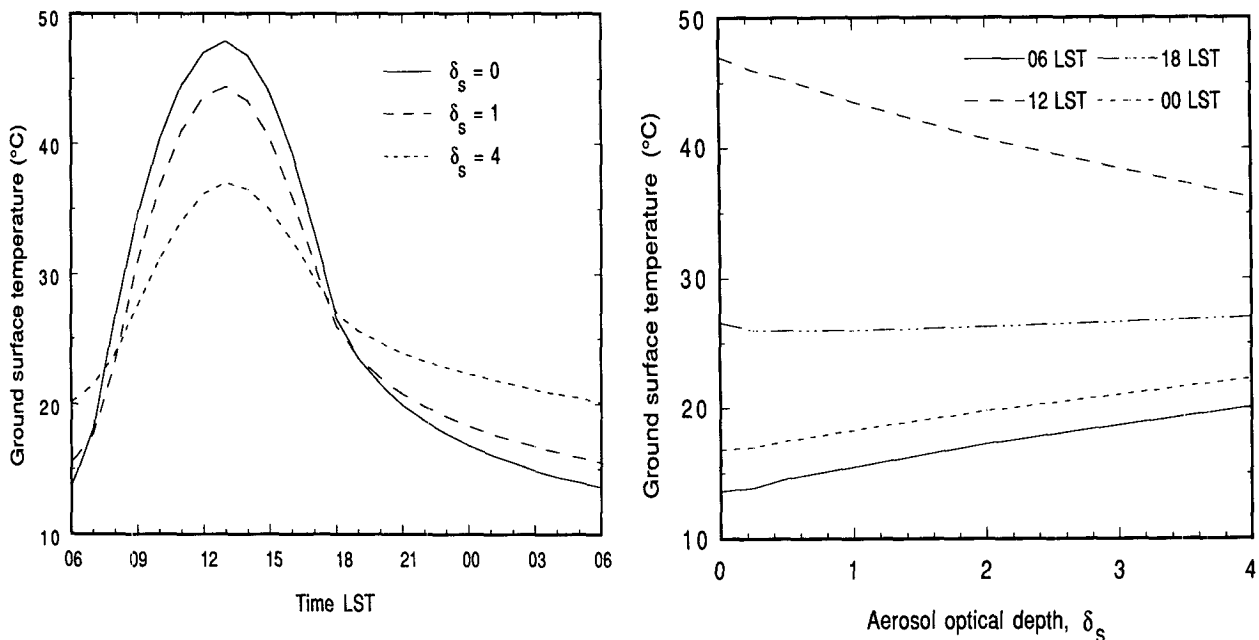


FIG. 2. (a) Daily variations of ground surface temperature  $T_S$  according to the atmospheric dust amount. (b) Ground surface temperature  $T_S$  against  $\delta_s$  at various times (LST) of the daily cycle.

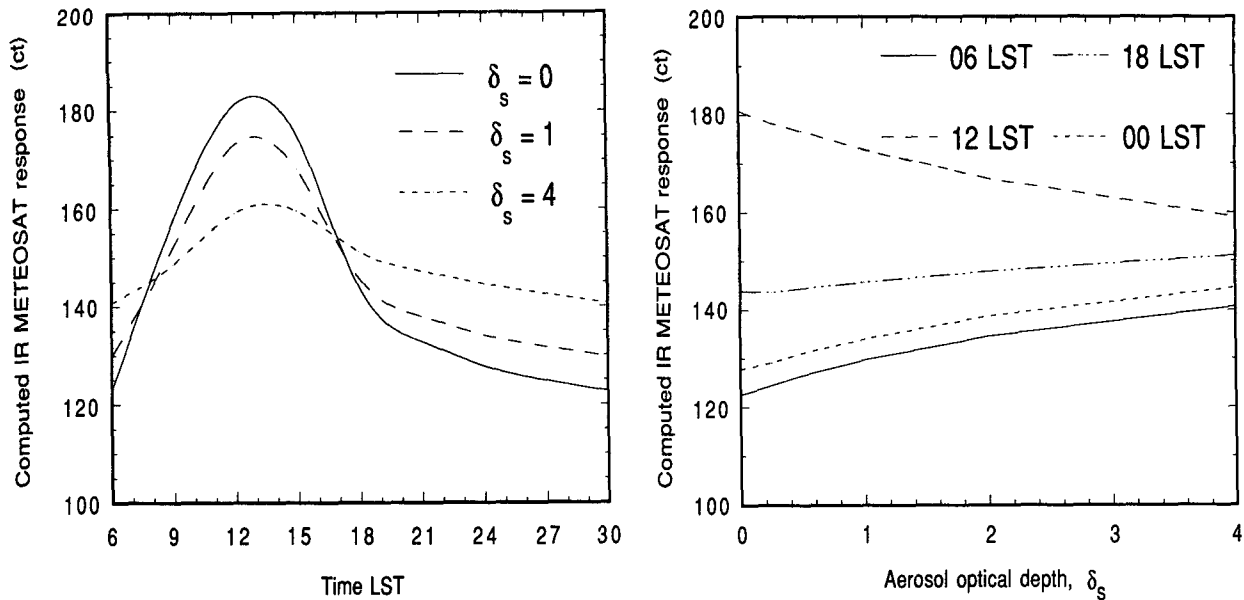


FIG. 3. (a) Daily variations of simulated Meteosat IR channel response  $C$  according to the atmospheric dust amount. (b) Simulated Meteosat IR channel response  $C$  against  $\delta_s$  at various times (LST) of the daily cycle.

the important daytime air temperature decrease observed with dust haze in Nigeria by Brinkman and McGregor (1983). The corresponding Meteosat response variations are presented in Figs. 3a and 3b. They result from the aforementioned impact of dust on  $T_s$  and from the transfer of the radiance emitted by the ground surface through a dust layer at a temperature

different of  $T_s$ . Figure 3b confirms the nonlinearity of the satellite response variation with  $\delta_s$ , experimentally observed and discussed in Legrand et al. (1989).

2) TYPE OF DUST

The incidence of the microphysical properties of an aerosol on the satellite response is investigated by comparing the respective impacts of the ECLATS aerosol with a quite different aerosol having a much greater fraction of large particles (see Fig. 4). The size distribution of this latter aerosol has been used by Carlson and Benjamin (1980) (abbreviated CB hereafter). For this aerosol type, relation (1) no longer holds. Instead, the optical properties in the longwave range are spectrally dependent, as reported by CB. The optical properties restricted to the Meteosat infrared channel are reported in Table 4 for a layer having a unit value of

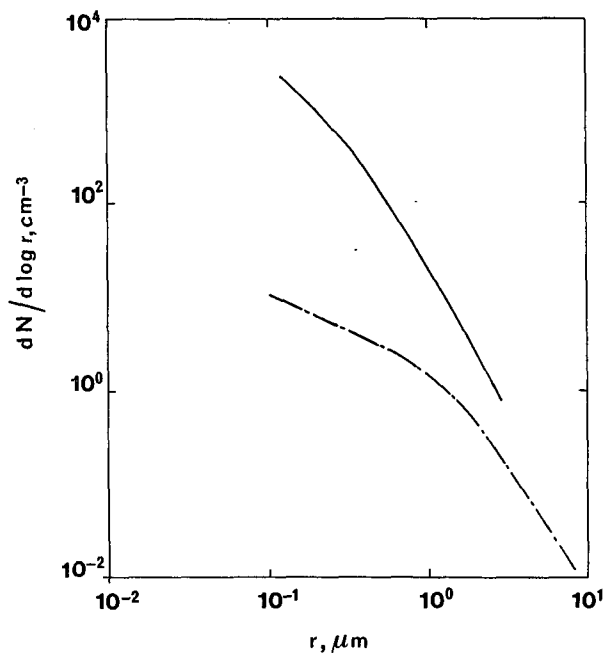


FIG. 4. Comparison of aerosol-size distributions observed during the ECLATS experiment, reported by Fouquart et al. (solid line); and reported by Carlson and Benjamin (dashed line).

TABLE 4. Compared properties of the ECLATS and CB (Carlson and Benjamin) aerosol models. Variable  $V$  is the total volume of dust in the vertical column of the atmosphere (cubic centimeters per  $1\text{-cm}^2$  section). Subscripts  $s$  and  $\text{sat}$ , respectively, refer to broadband shortwave and to  $10.5\text{-}12.5\text{-}\mu\text{m}$  Meteosat channel;  $\delta_s$ ,  $\omega_s$ , and  $g_s$  are, respectively, the optical depth, the single-scattering albedo, and the asymmetry factor.

Model	$V$ (cm)	$\delta_s$	$\omega_s$	$g_s$	$\delta_{\text{sat}}$	$\omega_{\text{sat}}$	$g_{\text{sat}}$
ECLATS	0.0036	1	0.952	0.655	0.12	0	
CB	0.0173	1	0.844*	0.780*	0.73	0.54	0.59

\* The values are averaged, after the spectral data given by Carlson and Benjamin (1980).

$\delta_s$ . It must be noted that quite different dust amounts correspond to the same  $\delta_s$ , as shown in Table 4, due to the differences in particle-size distribution. The sensitivity to the complex indices is not investigated here, since it is expected to be weak, as reported by CB.

The output data of the CSU mesoscale model for each aerosol type allow a quantitative analysis of the compared impacts. The major difference between the effects of the aerosol models under consideration is a greater enhancement of the atmospheric irradiance during the whole daily cycle for the CB model, owing to a significantly greater longwave optical depth, as expected from the larger CB dust volume. The resulting greenhouse effect then induces a large warming, both at surface and in the dust layer. An additional difference arises from the absorbing properties of the aerosol models in the shortwave range (as shown by the values of  $\omega_s$  in Table 4). In the case of the CB model, the downward irradiance at the ground is even more attenuated than with the ECLATS model, and higher shortwave heating rates affect the dust layer, involving an increase of the air temperature during daytime. Of course, a consequence of this is that the dust-layer temperature remains higher during the complete diurnal cycle in the case of the CB aerosol model.

In Fig. 5a the daily variations of the ground surface temperatures are compared first without aerosol and then with amounts of ECLATS and CB aerosol corresponding to a unit value of  $\delta_s$ . It is observed that the nocturnal temperature decrease is drastically reduced with the CB aerosol because the greenhouse effect is much stronger than for the ECLATS aerosol. On the contrary, during the daytime, the temperature increase is moderately reduced with the CB aerosol as compared with the ECLATS one, still owing to the large greenhouse effect partially compensating the large visible attenuation.

In Fig. 5b the corresponding Meteosat response variations are plotted, allowing the comparison of ECLATS and CB aerosol impacts. During the night, the CB aerosol impact is much higher than the ECLATS one due to the warmer ground surface and dust layer. During the day, the CB aerosol impact appears still higher—in spite of a lesser ground surface temperature drop—because of the considerable screening of the ground radiance through a widely colder dust layer characterized by a large infrared optical depth (see  $\delta_{\text{sat}}$  in Table 4).

### 3) GEOMETRICAL DEPTH OF THE DUST LAYER

The impact of the geometrical depth of a dust layer of the ECLATS type is investigated by varying the height of the top of the layer from the reference value of 1.2 km to a value of 2.0 km, a realistic variation representative of the variability for the area of Niamey, as inferred from ECLATS experiment and other observations. The computed variations are weak: for a

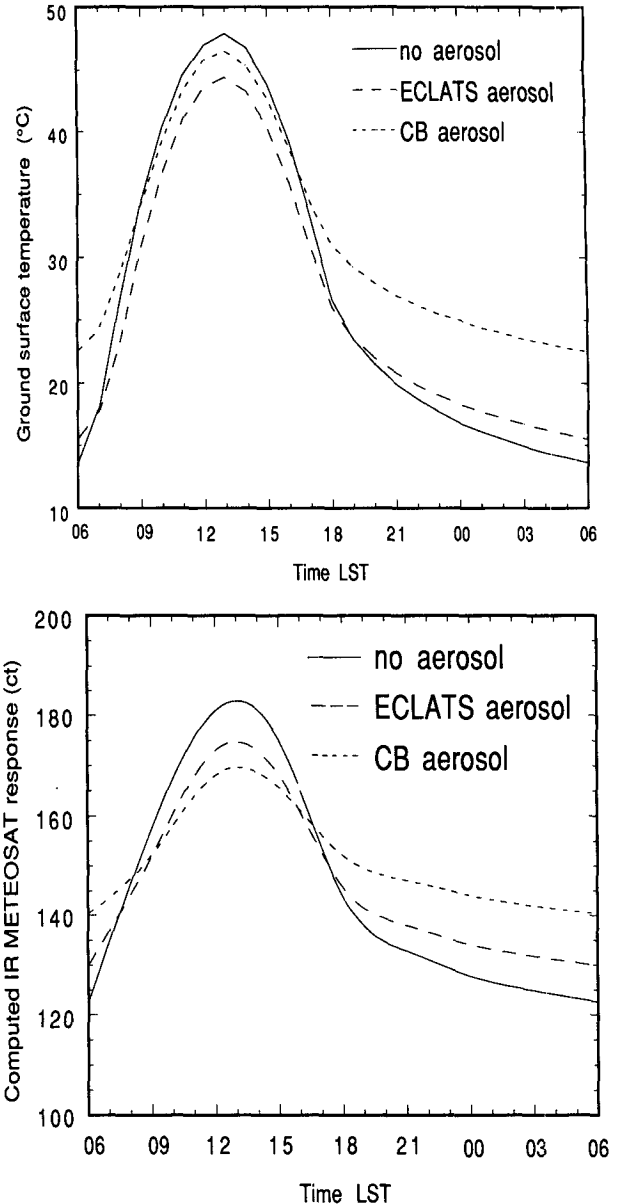


FIG. 5. (a) Daily variations of ground surface temperature  $T_s$  without aerosol, and with the ECLATS and CB models ( $\delta_s = 1$ ). (b) Daily variations of simulated Meteosat IR channel response  $C$  without aerosol, and with the ECLATS and CB aerosol models ( $\delta_s = 1$ ).

unit value of  $\delta_s$ , the surface temperature decreases by  $0.4^\circ\text{C}$  at 1200 LST and by  $0.1^\circ\text{C}$  at 0600 LST, and the Meteosat response decreases by 1.6 counts at 1200 LST and by 0.7 counts at 0600 LST.

### b. Sensitivity to the target characteristics

#### 1) LAND SURFACE ALBEDO

The effect of a modification of the land surface albedo has been evaluated by replacing the reference value of 0.26 by values of 0.20 and 0.30, successively

without dust and with a unit value for  $\delta_s$ . The appreciably modified variables are the upward shortwave irradiance and hence the net irradiance, involving in turn a variation of the surface temperature and of the satellite response. The effect is moderate during the day and vanishes during the night, as can be seen from the results presented in Table 5. Symbols  $\Delta T_S$  and  $\Delta C$  represent the ground surface temperature and satellite radiometric count variations involved by a variation of 0.01 on the albedo. For instance,  $\Delta C$  could be the realistic error associated with an uncertainty of localization of a target from a satellite image: it is less than one count at midday and becomes negligible during the night. Symbols  $k_T$  and  $k_C$  are the sensitivities of the ground surface temperature and satellite radiometric counts to dustiness; they are defined as the variations of  $T_S$  and  $C$  for a unit variation of  $\delta_s$ . The parameter  $k_C$  shows a moderate dependence to the albedo  $A$ , with enhanced absolute values over targets of lower albedo.

2) LAND SURFACE EMISSIVITY

The emissivity of the land surface affects the satellite response according to the two following mechanisms:

- (i) It controls the longwave energy exchange between the ground and the atmosphere, and hence the energy balance and the temperature at the surface.
- (ii) Given any surface temperature (i.e., independent from the former mechanism), it controls the upward emitted radiance, and thus the satellite response.

Indeed, it is appropriate to define distinct parameters to describe these two effects. The infrared energy exchange at the surface is controlled by the hemispherical spectrally averaged emissivity, conveniently defined for modeling the longwave irradiances emitted and reflected by the land surface. This emissivity  $\epsilon_l$  is used in the CSU mesoscale model to describe the emission from a surface assumed gray and Lambertian and its reflection of the atmospheric downward radiation.

TABLE 5. Impact of land surface albedo  $A$  on surface temperature  $T_S$  and on satellite response in radiometric counts  $C$ ; (a) describes the deviations involved by a variation of 0.01 of  $A$  according to  $\delta_s$ , (b) describes the variations against  $\delta_s$  (sensitivities) according to  $A$ .

(a)	$\Delta T_S$ (°C)		$\Delta C$ (count)	
	$\delta_s = 0$	$\delta_s = 1$	$\delta_s = 0$	$\delta_s = 1$
1200 LST	0.3	0.3	0.7	0.5
0600 LST	<0.1	<0.1	0.1	0.1
(b)	$k_T$ (°C)		$k_C$ (count)	
	$A = 0.2$	$A = 0.3$	$A = 0.2$	$A = 0.3$
1200 LST	-3.7	-3.0	-8.9	-7.0
0600 LST	+2.1	+1.9	+7.5	+7.1

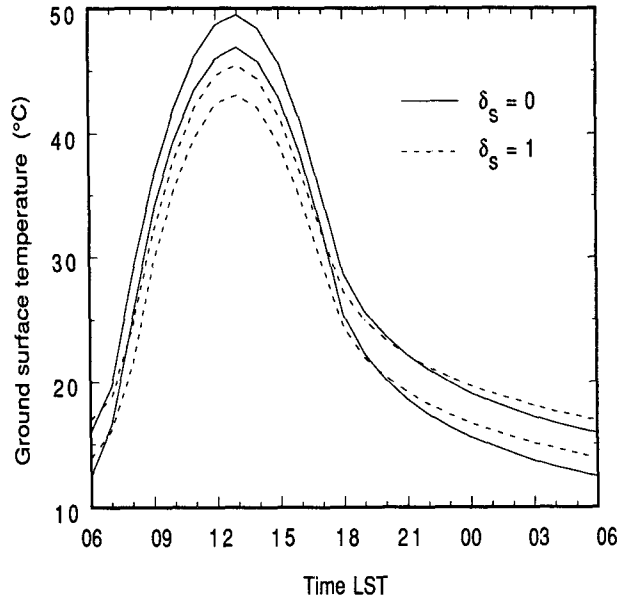


FIG. 6. Daily variations of the ground surface temperature  $T_S$  according to the hemispheric ground surface emissivity spectrally averaged on the longwave range  $\epsilon_l$  for a dust-free case and for  $\delta_s = 1$ . Upper and lower curves are for  $\epsilon_l = 0.8$  and  $\epsilon_l = 1$ , respectively.

On the other hand, the radiance emitted and reflected by the surface toward the satellite is controlled by a directional emissivity  $\epsilon_{sat}$  defined in the 10.5–12.5- $\mu\text{m}$  range and weighted by the Meteosat radiometer spectral response to account for the signal detection. Even though bare soils are not Lambertian reflectors, they are almost Lambertian emitters, as outlined by Becker et al. (1986). Since the absorption by atmospheric gases is quite strong out of the 8–14- $\mu\text{m}$  window region, one can expect that the spectrally averaged emissivity  $\epsilon_l$  does not depart significantly from  $\epsilon_{sat}$ .

Varying the parameter  $\epsilon_l$  in the CSU mesoscale model results in the modification of the ground surface temperature as illustrated in Fig. 6. Curves are plotted for emissivities of 0.8 and 1, which may be considered as the realistic limits for usual ground emissivity values. With a low emissivity, the exchanged infrared flux is reduced, thus involving a decrease of the radiative cooling and the increase of the temperature at the ground surface. As  $\epsilon_l$  decreases from 1 to 0.8,  $T_S$  increases about 3°C in the dustless case for the whole diurnal cycle. This effect is slightly reduced by the presence of dust, as can be observed in Fig. 6.

In Fig. 7, the modifications of the daily cycle of satellite response are presented for a gray Lambertian ground surface characterized by equal values of  $\epsilon_l$  and  $\epsilon_{sat}$ . For a dust-free atmosphere a decrease of the emissivities from 1 to 0.8 involves a reduction of 20–25 counts during daytime and 10–15 counts during nighttime. This variation is reduced in the presence of dust, but remains important with a satellite response decrease



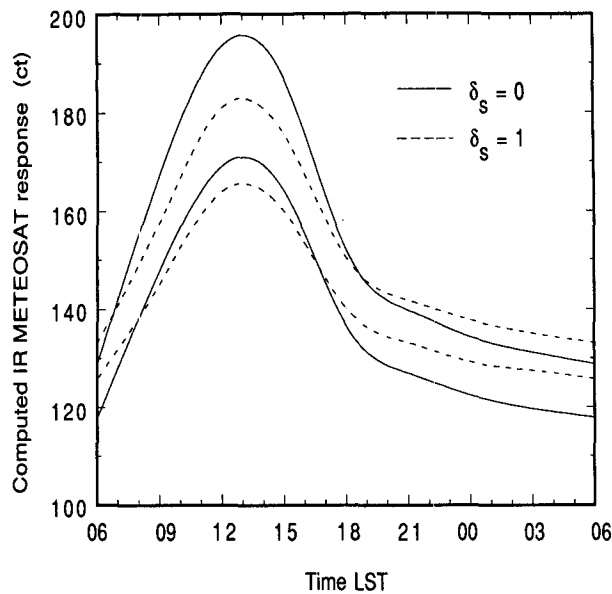


FIG. 7. Total impact for identical values of hemispheric spectrally averaged emissivity  $\epsilon_l$ , and of directional Meteosat IR channel emissivity  $\epsilon_{\text{sat}}$ , on daily variations of Meteosat IR channel response for a dust-free case and for  $\delta_s = 1$ . Contrary to Fig. 6, upper and lower curves are for  $\epsilon_l = \epsilon_{\text{sat}} = 1.0$  and  $\epsilon_l = \epsilon_{\text{sat}} = 0.8$ , respectively.

of 15–20 during daytime and 8–10 counts during nighttime.

Figure 8 allows us to estimate the respective parts of  $\epsilon_l$  and  $\epsilon_{\text{sat}}$  in the large variations of the satellite response observed in Fig. 7. The solid and dotted curves relate to different values of  $\epsilon_l$  (1 and 0.9, respectively) and the same unit value of  $\epsilon_{\text{sat}}$ . It is observed that for

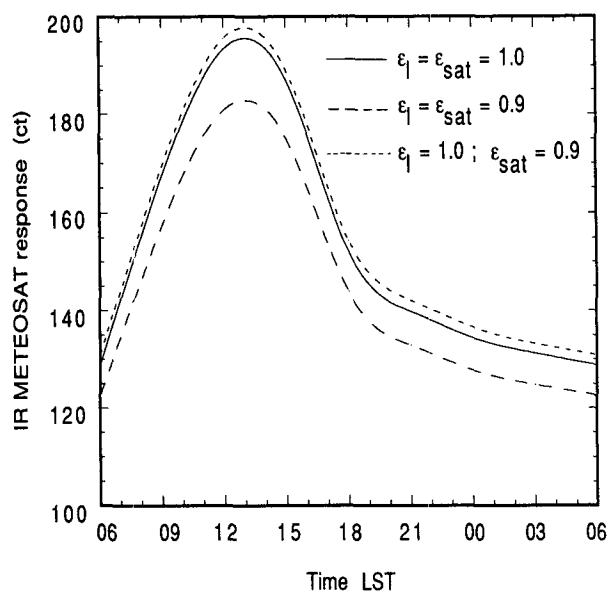


FIG. 8. Compared impacts of  $\epsilon_l$  and  $\epsilon_{\text{sat}}$  on Meteosat IR channel response  $C$  for a dust-free case.

TABLE 6. Impact of variations of 0.01 for  $\epsilon_{\text{sat}}$  and for  $\epsilon_l$  on the satellite response expressed in radiometric counts.

Parameter variation	Time (LST)	$\Delta C$ (count)	
		$\delta_s = 0$	$\delta_s = 1$
$\Delta\epsilon_{\text{sat}} = 0.01$	1200	+1.5	+1.1
	0600	+0.8	+0.6
$\Delta\epsilon_l = 0.01$	daily mean		-0.2

a given value of  $\epsilon_{\text{sat}}$ , the satellite response increases as  $\epsilon_l$  decreases, due to the corresponding increase of the surface temperature. However, this is a rather slight effect of about 2 counts, practically constant during the whole daily cycle (for any value of  $\delta_s$ ). The dotted and dashed curves relate to different values of  $\epsilon_{\text{sat}}$  (1 and 0.9, respectively) with a common value of 0.9 for  $\epsilon_l$ . For a decrease of  $\epsilon_{\text{sat}}$ , the satellite response decreases due to the corresponding drop of the ground-emitted radiance. This highly sensitive effect varies in the range 6–15 counts during the daily cycle, and is more important during daytime and for dust-free conditions. Table 6 summarizes the variations of the satellite response  $\Delta C$  as a function of  $\delta_s$ , for variations of 0.01 for  $\epsilon_{\text{sat}}$  and  $\epsilon_l$ , respectively.

Figure 9 presents the daily variations of the satellite IR channel sensitivity to dust  $k_C$  for  $\epsilon_l = \epsilon_{\text{sat}} = 0.8$  and 1. This sensitivity is defined as the variation of the Meteosat response (in counts) corresponding to a unit shortwave optical depth variation. It is negative during daytime and positive during nighttime, and it algebraically increases whenever  $\epsilon_{\text{sat}}$  decreases. So the most

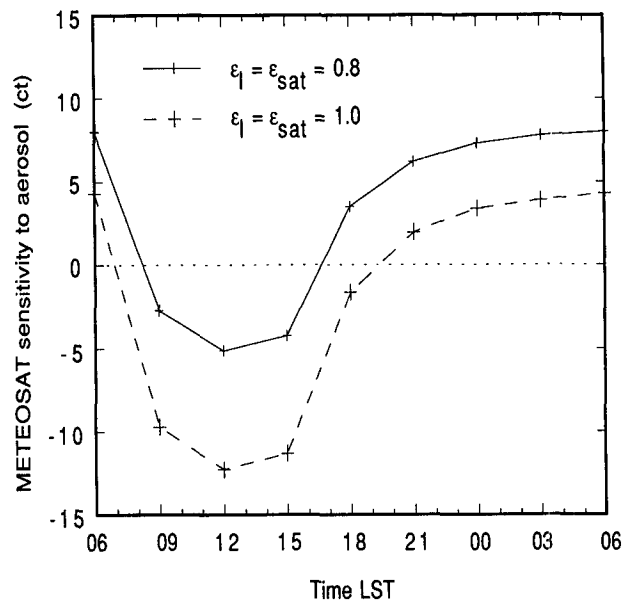


FIG. 9. Daily variations of simulated Meteosat IR sensitivity to dust  $k_C$  according to the emissivity.

favorable conditions to observe the dustiness, inferring the greatest absolute values of  $k_C$ , are met during daytime for a high emissivity and during nighttime for a low one.

### c. Sensitivity to parameters extraneous to dustiness

#### 1) ATMOSPHERIC WATER VAPOR

The impact of atmospheric moisture on the satellite response has been evaluated by multiplying the initial humidity profile shown in Fig. 1c successively by 0.5, providing dry atmospheric conditions, and by 2, giving moist atmospheric conditions. The soil wetness is kept unmodified for these various air moisture conditions so that evaporation and latent heat flux are practically unchanged. As the soil is very dry (see Table 2 for the wetness), and advection plays no part in the present study, evaporation is weak and the atmospheric water vapor amount is almost constant over the daily cycle, with mean columnar values of 0.9, 1.7, and 3.4  $\text{g m}^{-2}$ , respectively, for the dry, reference, and moist atmospheres.

Figure 10a presents daily cycles of the ground surface temperature relating to the three cases. The temperature is enhanced by the presence of atmospheric water vapor, as a result of the associated greenhouse effect. During nighttime at 0600 LST the deviations from the reference atmosphere reach  $-2.3^\circ$  and  $+4.3^\circ\text{C}$ , respectively, for the dry and moist atmospheres. During daytime, the greenhouse effect is for its greater part balanced by the absorption in the solar range, so that the total downward and net irradiances at the ground are only slightly modified and the temperature variations do not exceed  $1^\circ\text{C}$ .

The presence of atmospheric water vapor involves the screening of the upward spectral radiance throughout the daily cycle, a mechanism counterbalancing the impact of the greenhouse effect on the satellite response. However, the strength of this radiative transfer effect is greater during the day owing to the decrease of the temperature with the height in the lower atmosphere, so that it overcompensates the incidence of the small ground surface temperature increase on the satellite response. On the other hand, this radiative transfer effect is weaker in the presence of the nocturnal inversion, so that it is widely overcompensated by the large ground surface temperature increase resulting from the greenhouse effect. The daily variations of the satellite response plotted in Fig. 10b reflect these combined effects.

#### 2) WIND

The wind field controls the convective exchanges in the atmosphere and at the surface. According to the wind strength in the lower atmosphere, a variable part of the radiative energy absorbed at ground surface is vertically transported back to the atmosphere in the form of convective heat. This effect chiefly concerns

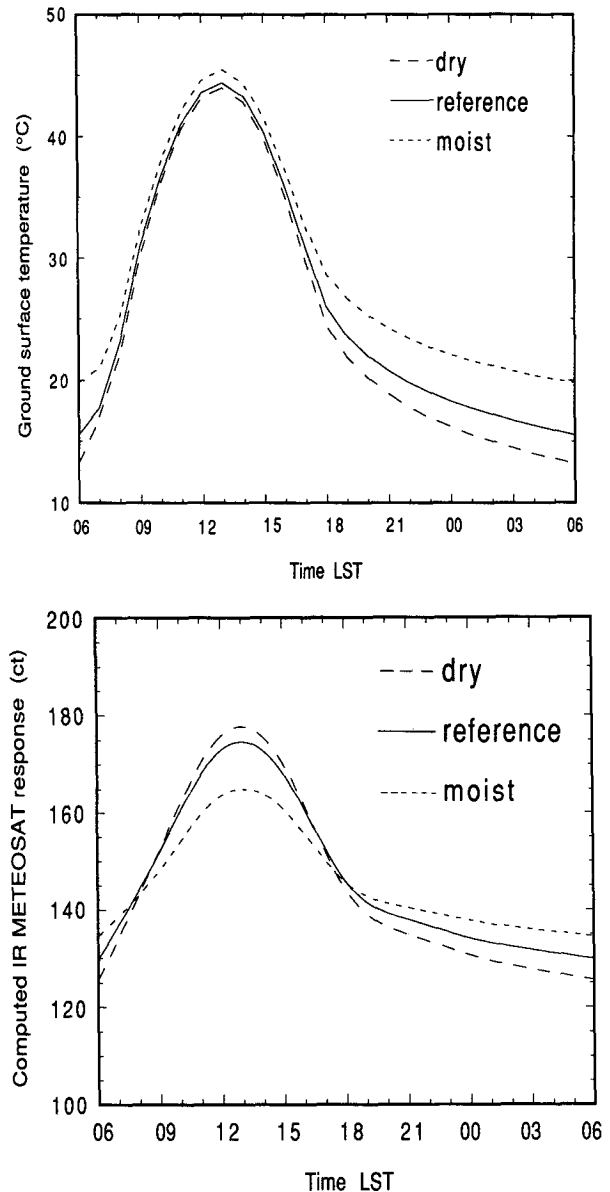


FIG. 10. (a) Daily variations of ground surface temperature  $T_S$  according to the atmospheric moisture ( $\delta_s = 1$ ). Dry, reference, and moist atmospheres contain 0.9, 1.7, and 3.4  $\text{g cm}^{-2}$  of water vapor, respectively. (b) Daily variations of simulated Meteosat IR channel response  $C$  according to the atmospheric moisture. Same values as for (a).

the sensible heat, which represents the major term of convective energy in the dry Sahelian conditions.

The wind field impact on the ground surface temperature has been investigated by multiplying the initial wind speed profile (shown in Fig. 1d) by 0.5 and 2, giving light wind and high wind conditions, respectively. The three resulting cases are characterized by the following daily mean wind speed at 2 m: 1.7  $\text{m s}^{-1}$  (light), 3.0  $\text{m s}^{-1}$  (reference), and 5.3  $\text{m s}^{-1}$  (high). Because the convective energy transfer from the surface

to the atmosphere is important only during daytime, the same can be said about the wind impact on the ground temperature and on the satellite response. This feature is depicted by the diurnal cycles of ground surface temperature and satellite response plotted in Figs. 11a and 11b.

**4. Discussion**

The results presented in the previous section allow the following statement: In addition to its predicted

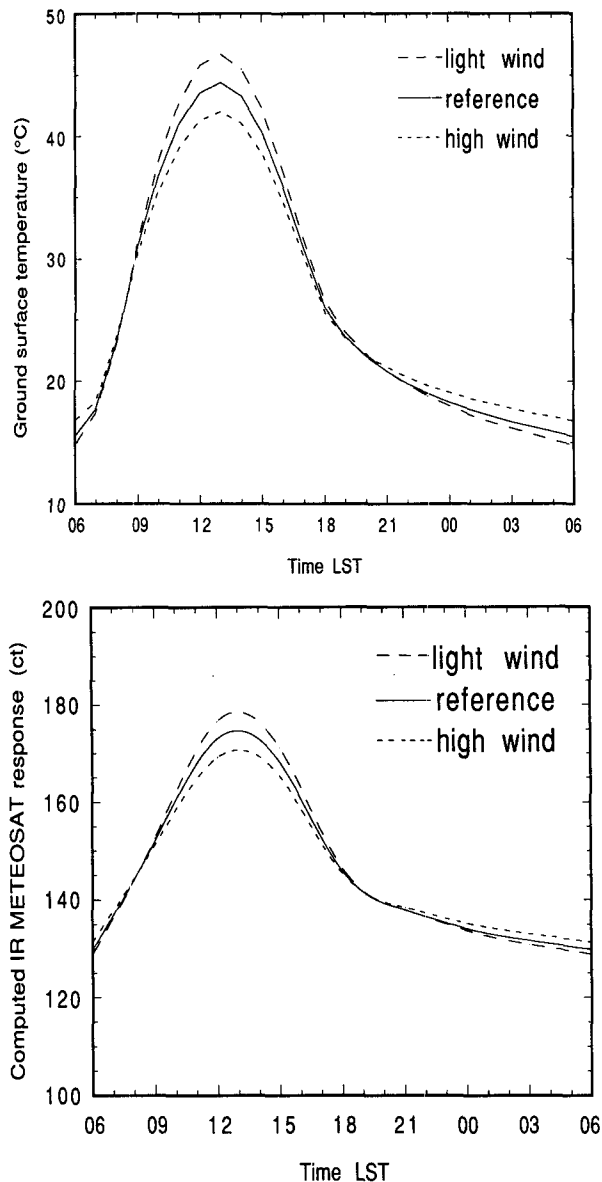


FIG. 11. (a) Daily variations of the ground surface temperature  $T_s$  according to the wind field ( $\delta_s = 1$ ). Light, reference, and high wind cases are associated to wind velocities at 2 m of 1.7, 3.0, and 5.3  $m s^{-1}$ , respectively. (b) Daily variations of simulated Meteosat IR channel response  $C$  according to the wind field. Same values as for (a).

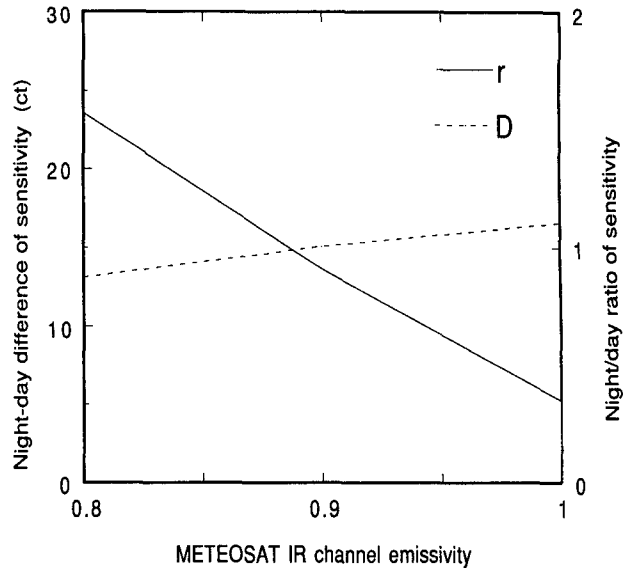


FIG. 12. Variations against  $\epsilon_{sat} (= \epsilon_t)$ , of the ratio  $r = |k_C^n / k_C^d|$  and the difference  $D = k_C^n - k_C^d$  (dotted line) of night ( $n$ , 0600 LST) and day ( $d$ , 1200 LST) Meteosat sensitivity to the ECLATS aerosol model.

high sensitivity to dust amount illustrated in Figs. 3a and 3b, the satellite response is highly sensitive to the ground surface emissivity  $\epsilon_{sat}$  in the satellite channel range and to the aerosol properties as shown by Fig. 8 and Fig. 5b, respectively.

The sensitivity of the satellite response to dust, presented in Fig. 9 for different land surface emissivities in the IR channel, agrees with earlier results reported for the nighttime by Legrand et al. (1988).

For a decrease of  $\epsilon_{sat}$ ,  $k_C$  increases approximately by the same amount during daytime and nighttime. However, as  $k_C$  is positive during nighttime and negative during daytime, its absolute value increases during nighttime and decreases during daytime. Hence, the night-to-day ratio  $r$  of the absolute values is drastically modified by an emissivity change, whereas in the same conditions the night-to-day difference  $D$ , is only slightly modified. Figure 12 shows the variations of  $r$  and  $D$  against  $\epsilon_{sat}$  for the ECLATS aerosol model, using values at 0600 LST for the night and values at 1200 LST for the day. It may then be inferred that the parameter  $r$  would be useful for the experimental determination of  $\epsilon_{sat}$ .

On the other hand, important variations of the parameter  $D$ , as it is rather insensitive to  $\epsilon_{sat}$ , would suggest primarily some departure in the aerosol properties. Indeed, the aerosol type is characterized in the used models by a set of radiation parameters describing the properties in the solar and infrared ranges, of either climatological or remote sensing relevance. For the ECLATS and CB aerosol models selected in the present study, the difference between the dust impacts on the satellite response results from the large difference be-

tween the longwave and shortwave optical depth ratios of aerosols;  $(1 - \omega_l)\delta_l/\delta_s$  is roughly three times greater for CB than for the ECLATS aerosol. As a consequence, replacing the ECLATS model by the CB model leads to a doubling of the night-to-day sensitivity difference (for  $\epsilon_{\text{sat}} = 0.9$ ,  $D = 15.1$  and 30.5 counts for the ECLATS and CB models, respectively).

However a more detailed description of the aerosol-type impact needs further sensitivity studies of the satellite response to suitably defined aerosol parameters and a realistic determination of the variability of these parameters.

Table 7 is presented for the purpose of comparing the calculated values of the satellite response sensitivity to dust  $k_C$  at 0600 LST and at 1200 LST with the experimental values reported by Legrand et al. (1989), derived from statistical comparison between Meteosat data and aerosol optical depth from ground photometric measurements in February 1985 at the site of Niamey. According to the aerosol model and to the value of  $\epsilon_{\text{sat}}$ , the computed sensitivity ranges from +4.8 to +17.7 counts at 0600 LST, and from -12.8 to -3.8 counts at 1200 LST. Changing the aerosol model, or tuning  $\epsilon_{\text{sat}}$  (or both together), would be clearly sufficient to have computed values of  $k_C$  located inside the 95% confidence interval of the experimental values, without attending to the influence of the other parameters and of the various sources of errors.

In Table 8 are summarized the variations of the satellite response obtained by varying the atmospheric water vapor amount and the wind strength in the model. They are interpreted as uncertainties  $\Delta C$  introduced in the determination of the aerosol optical depth and converted into  $\Delta\delta_s$ . These variations result in uncertainties in the satellite determination of the optical depth  $\Delta\delta_s = 0.6$  for a change of  $1 \text{ g cm}^{-2}$  of the columnar water vapor amount and  $\Delta\delta_s = 0.1\text{--}0.3$  for a change of  $1 \text{ m s}^{-1}$  of the daily mean wind speed at 2 m.

TABLE 7. Comparison of calculated values with experimentally derived results for the satellite response sensitivity to dust  $k_C$ ; superscript *n* (night) means 0600 LST, superscript *d* (day) means 1200 LST.

Results	Cases	$k_C^n$ (count)	$k_C^d$ (count)
Computed by model	ECLATS aerosol model $\epsilon_{\text{sat}} = 0.9$	+7.2	-7.9
	CB aerosol model $\epsilon_{\text{sat}} = 0.9$	+17.7	-12.8
	ECLATS aerosol model $\epsilon_{\text{sat}} = 0.8$	+9.5	-3.8
	ECLATS aerosol model $\epsilon_{\text{sat}} = 1$	+4.8	-12.7
	Derived experimentally	mean	+8.7
	95% lower limit	+6.1	-17.3
	95% upper limit	+11.3	-10.1

TABLE 8. Variations of the satellite response  $\Delta C$  and of the equivalent aerosol optical depth  $\Delta\delta_s$ , obtained by varying: (a) the vertical columnar water vapor amount and (b) the wind speed at level 2 m, below and above the reference values,  $1.7 \text{ g cm}^{-2}$  and  $3.0 \text{ m s}^{-1}$ .

Atmospheric conditions		$\Delta C$ (count)		$\Delta\delta_s$	
		0600 LST	1200 LST	0600 LST	1200 LST
(a) Moisture	dry ( $0.9 \text{ g cm}^{-2}$ )	-4.2	+2.9	-0.58	-0.37
	wet ( $3.4 \text{ g cm}^{-2}$ )	+4.7	-9.4	+0.65	+1.19
(b) Wind	light ( $1.7 \text{ m s}^{-1}$ )	-0.9	+3.6	-0.12	-0.46
	high ( $5.3 \text{ m s}^{-1}$ )	+1.6	-3.9	+0.22	+0.49

The values of  $\Delta\delta_s$  for variations of the atmospheric moisture apparently mean that a satisfactory retrieval of  $\delta_s$  using infrared satellite data would comply with the estimation of the columnar water vapor amount in the atmosphere. Such a requirement has already been predicted for the nighttime by Legrand et al. (1989), according to a result of satellite response sensitivity to water vapor reported in a previous paper (Legrand et al. 1988), and it has been directly confirmed for daytime on the basis of measurements at a site near the seashore (Dakar, April 1987), where moisture variations were important (Tanré and Legrand 1991). However, giving a definitive diagnosis on this point requires the knowledge of the likely variations of the atmospheric moisture, which depends on the season and on the location of the selected target. So, contrarily to the case of Dakar where the water vapor was needed for the retrieval of the aerosol optical depth, it was shown in Legrand et al. (1989) that the aerosol optical depth could be satisfactorily expressed using the infrared satellite data with an empirical formula that neglects the atmospheric moisture, for the site of Niamey located far inland. It is thus recommended that a climatological study of the monthly mean and standard deviation of the atmospheric water vapor amount be done when selecting a target for the monitoring of the optical depth using the infrared satellite data. Practically, favorable conditions would be encountered for targets located in dry areas, far from the ocean.

As shown in Table 8, the impact of the wind is in no way so drastic. However, corrections could be made by taking into account the wind speed provided by the meteorological stations (monthly means and standard deviations could also be useful).

The impact of the albedo is moderate. The albedo can be determined with methods using air or spaceborne measurements. However, the uncertainties relative to these determinations still seem high, owing particularly to the approximate knowledge of the ground spectral and directional reflectance properties (see Arino 1990).

It must also be noted that the Sahelian area probably undergoes large seasonal variations of albedo associated with the alternating of contrasted dry and rainy seasons. However, it is thought that during a large part of the dry season—the very one concerned with dust events—the albedo variation is weak because of the rather steady state of the withered vegetation and dry ground.

The satellite response variations resulting from a geometrical depth change of the dust layer—turbidity being unchanged—are negligibly small with the realistic values used for the Sahelian area. However, the dust layer is expected to be much deeper for very different atmospheric conditions, as can be observed in the northern part of Sahara. Especially, the occurrence of dust clouds lifted in altitude could give rise to a somewhat greater impact on the satellite response, as a result of the much lower dust-layer temperature.

## 5. Conclusions and prospects

The effect of the atmospheric dustiness on the Meteosat response in the thermal infrared range is studied using a mesoscale model and radiative transfer codes described and tested in Part I of the current study. The sensitivity of the satellite response is computed for a Sahelian site according to dust amount, radiative properties, and vertical distribution, to surface albedo and emissivity, and to atmospheric water vapor content and wind velocity. This sensitivity results in part from the variation of the ground surface temperature. Thus, this latter parameter is at the center of the present study. The theoretical results are observed to be in agreement with previously published results (Legrand et al. 1988; Legrand and Desbois 1989; Legrand et al. 1989).

The satellite response to dust results in large part from a strong dust impact on the surface temperature. This response is approximately proportional to the aerosol optical depth, defining a sensitivity coefficient  $k_C$  shown to depend primarily on the aerosol properties and the surface emissivity in the satellite channel  $\epsilon_{\text{sat}}$ . Large errors in the estimate of dustiness by means of the satellite data can be involved by variations of atmospheric water vapor content.

Therefore, the determination of the aerosol optical depth from a given variation of the satellite response is not unique. Thus, a necessary step in the development of a method of atmospheric dust remote sensing by means of infrared satellite data will obviously be to investigate the appropriate way to measure and take into account the major parameters identified in the present study:

(i) The aerosol type plays a prominent part in the determination of the satellite response sensitivity as shown in section 3a.2. Hence, a knowledge of the values and realistic ranges of variability of the aerosol optical properties is needed. Indeed, as is regularly stressed by

the experts (i.e., at the WMO-ISCU meeting, 1983; see also d'Almeida et al. 1989), the properties of tropospheric aerosols—dust and others—are insufficiently known, with too scarce and partial experimental datasets. These insufficiencies are particularly evident in the infrared range. This lack of data is a serious obstacle, especially because of a large space and time variability of these aerosol properties. In such conditions, one may wonder whether the IR method, that is, the method of retrieval of aerosol optical depth from IR satellite imagery, can be made operational.

However, experimental studies could monitor these variations. Legrand et al. (1989) and Tanré and Legrand (1991) present examples of regression between IR satellite response and optical depth over a 1-month period at Niamey and Dakar. The correlations computed independently of any parameter describing the aerosol properties are very strong; so it is obvious that these properties were approximately constant throughout the periods of concern. This fact agrees with the idea that the rather regular wind pattern over the Sahel may involve a unique—or very dominant—source area of generation for the dust events observed at a given site. Thus, an appreciable variation in the dust properties at this site could arise only from the seasonal change of the wind pattern and correlated source area.

Thus, it is expected that observational studies can enable us to specify, over a given area, a regular seasonal variation of the optical properties of the aerosol, so that the IR method applicability may be more clearly assessed. Many experimental investigations are still necessary to set up a time-dependent cartography of the Saharan aerosol properties. With that prospect, a workshop organized by the “Réseau des Sciences de l'Atmosphère en Afrique Francophone” (Dakar, November 1990) was intended to settle a network of photometric and complementary measurements in West Africa.

(ii) The ground surface emissivity  $\epsilon_{\text{sat}}$  in the satellite channel range is an important parameter for the determination of the satellite response sensitivity to dust, as already pointed out in Legrand et al. (1988). It is expected to vary negligibly with time at a given site during the dry season. However, this point has to be confirmed experimentally. As reported by Becker et al. (1986), the remote measurement of this parameter is difficult and attention will have to be paid to this question.

(iii) The atmospheric humidity needs to be taken into account in order to avoid large uncertainties in the determination of the dust optical depth from the satellite response. Monthly mean and standard deviation of atmospheric water vapor content derived from data routinely measured at meteorological stations could give a satisfactory solution if the standard deviation is small enough. An optional solution is the use of rawinsonde data collected simultaneously with

the satellite data acquisition. But the most adequate solution is clearly the satellite remote sensing of this atmospheric water vapor.

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