Regional and Seasonal Variations of the Clear Sky Atmospheric Longwave Cooling over Tropical Oceans

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ABSTRACT

The vertical distribution of the clear sky longwave cooling of the atmosphere over tropical oceans is inferred from three different datasets. Two of the datasets refer to the TIROS-N Operational Vertical Sounder (TOVS) NOAA/NASA Pathfinder project, PathA and PathB, and the last one refers to the ECMWF reanalysis (ERA-15). Differences are identified originating from the temperature and water vapor fields. They affect the geographical distribution of the longwave fields to various degrees. However, the three datasets lead to similar conclusions concerning the sensitivity of the clear sky total longwave cooling to SST variations. For the highest values of the SST (greater than 27°C), positively correlated to the increased efficiency of the longwave trapping (super-greenhouse effect), the atmosphere shows a lesser efficiency to cool radiatively. The atmosphere does reradiate the longwave radiation toward the surface as efficiently as it traps it. This is verified on regional as well as on seasonal scales. Such longwave cooling behavior is due to an increased mid- and upper-tropospheric humidity resulting from convective transports. The three datasets agree with the vertical distribution of the radiative cooling variations from normal to favorable to super-greenhouse effect conditions, except in the boundary layer, where the coarse resolution of the TOVS-retrieved data makes them not reliable in it. In “normal” conditions the cooling uniformly increases over the vertical with the SST. Over 27°C, the cooling is intensified above 400 hPa and reduced between 900 and 400 hPa.

1. Introduction

Although the earth–atmosphere system receives its energy from the sun in the shortwave band, the infrared radiation is also an essential component of the energy balance since it allows the system to lose its excess of energy and to regulate its temperature. According to the Clausius–Clapeyron relation, the atmospheric water vapor content is strongly dependent on the surface temperature: in the context of a possible climate warming, an increased surface temperature would lead to increased water vapor content, the gas that contributes the largest fraction to the greenhouse absorption in the present-day atmosphere. The existing observations allow the investigation of climate feedbacks, which may be involved in a long-term climate change. In particular, the study of the relationship between clear sky radiative budget and surface temperature in the actual climate provides an insight into the response of the earth–atmosphere energy balance to increased surface temperature, especially with respect to the thermodynamic water vapor feedback.

Until recently, owing to the global observations that are currently available, longwave (LW) radiative budget studies have focused on the top of the atmosphere (TOA) outgoing longwave radiation (OLR), on the total atmospheric absorbed radiation (the so-called greenhouse effect $G$), and on their variations as a function of sea surface temperature (SST), of the total precipitable water (see, e.g., Raval and Ramanathan 1989; Stephens and Greenwald 1991; Webb et al. 1993; Hallberg and Inamdar 1993; Sinha 1995), or of the large-scale atmospheric circulation (Bony et al. 1997). However, the ability of the atmosphere to lose its excess of energy and consequently to regulate its temperature, depends both on the OLR and on the LW radiation emitted by the atmosphere toward the surface. It is important to extend the previous results conducted on the OLR and

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on $G$ to radiative quantities taking into account both the upward and the downward LW fluxes.

The recent availability of global datasets of a variety of meteorological parameters makes it possible to estimate the vertical distribution of LW fluxes from the TOA to the surface with reasonable accuracy using radiative transfer models (Zhong and Haigh 1995; Slingo et al. 1998; Chevallier et al. 2000). It becomes feasible to study thoroughly the vertical distribution of the radiative energy balance. In this work, this question is addressed by studying the clear sky component of the infrared cooling and its vertical distribution, over tropical oceans. Here, the clear sky radiative description of the atmosphere is derived from three global datasets of vertical profiles of temperature and water vapor, and of fields of surface temperature, with a radiative transfer model. They are described in section 2. The expression “clear sky” is not restricted to the emission from the clear sky pixels only, but refers to the value that the LW fluxes would have if the clouds did not affect the radiative transfer. In section 3, the geographical variations of the clear sky OLR and these of the clear sky LW cooling, $C_{\text{LW}}$, are compared. The differences resulting from the various sets of meteorological parameters are discussed. In section 4 the geographical and seasonal variations of the clear sky LW cooling are analyzed as a function of the SST. Thermodynamic aspects as well as radiative aspects of the sensitivity of the clear sky LW cooling to SST variations are discussed. Variations in the vertical distribution of the LW cooling as a function of SST are studied in section 5. The last section (6) summarizes the results.

2. The data

a. The meteorological parameters

Various descriptions of the 3D thermodynamic structure of the earth atmosphere are currently available. Each of them has its own advantages and deficiencies that can affect the radiative parameters that are studied here. For this reason, three different datasets are inter-compared in the present study.

The first two datasets, respectively, called PathA (Susskind et al. 1997) and PathB (Scott et al. 1999), refer to the reanalysis of the TIROS-N Operational Vertical Sounder (TOVS) observations, taken on board the polar-orbiting National Oceanic and Atmospheric Administration (NOAA) satellites since 1979. TOVS combines a wide range of infrared and microwave sounding channels spread on three radiometers: the High-resolution Infrared Radiation Sounder (HIRS2), second generation, 20 channels, Microwave Sounding Unit, 4 channels (MSU), and Stratospheric Sounding Unit, 3 channels (SSU). In general two satellites with equator crossing times that are approximately 2:30 a.m. and p.m. and 7:30 a.m. and p.m. local time are flying simultaneously. The PathA and PathB datasets are archived as part of the NOAA-National Aeronautics and Space Agency Administration Pathfinder program (Maiden and Greco 1994). These include the temperature and the water vapor profiles, as well as the surface temperature fields, retrieved from the TOVS observations. Cloud parameters (cloud-top pressure and cloud effective emissivity) are also retrieved, but are not used in the present study. The main difference between the two Pathfinder datasets studied here is the nature of the a priori information that is used in conjunction with the TOVS observations. For the production of the PathB data, the improved initialization inversion (3I) procedure selects an initial guess in the Thermodynamic Initial Guess Retrieval (TIGR; Chédin et al. 1985) dataset. The current version of this database groups together about 2300 atmospheric situations selected from 450 000 radiosonde reports and satellite retrievals (Chevallier et al. 1998). In PathA, the first guess for atmospheric moisture and temperature profiles used by the retrieval system is the 6-h forecast produced by the Goddard Earth Observing System (GEOs). The algorithm for retrieving cloud properties, that substantially affect the quantity and the quality of the retrievals, is also different in the two datasets. The reader is referred to Susskind et al. (1997) and to Scott et al. (1999) for the details of both approaches. An additional processing is applied here to the PathB data. Indeed intrinsic limitations of infrared-based retrievals lead to relatively poor retrievals of water vapor in subtropical regions of subsidence off the western coast of the continents (Stephens et al. 1994; Chéruy et al. 1995). Therefore, the PathB data used here are corrected with independent data from the Special Sensor Microwave Imager (SSM/I) flown on the Defense Meteorological Satellite Project, in order to improve the retrievals [as explained in Chevallier et al. (2000)].

The third dataset used in the present study refers to the reanalyzed water vapor and temperature fields produced by the European Centre for Medium-Range Weather Forecasts (ECMWF) in conjunction with an estimation of the SST derived from blended satellite and in situ observations (Reynolds and Smith 1994). Output from a numerical forecast model is used together with observations, including the TOVS radiances, and forcing fields as input for the reanalysis (also called ERA-15, for 15-yr ECMWF reanalysis; see Gibson et al. 1997). The data assimilation scheme linked to the numerical forecast model allows information concerning the state of the global atmosphere to propagate from data-rich areas to data-sparse areas, according to the specified error statistics and to the physics and the dynamics that govern the system. Compared to the Pathfinder datasets, the ERA-15 data do not suffer from the sampling problem inherent in the infrared retrievals: the analysis is produced every 6 h.

Four months of data, January and July 1988 and 1990, grided on a $1^\circ \times 1^\circ$ grid for all three sets, are considered here. For ERA-15 the analyses at the
four synoptic times are used. Concerning the Pathfinder sets, only the evening paths of the NOAA-10 satellite, roughly 7:30 p.m. local time, are processed. As the diurnal cycle is expected to play a significant role over lands, the study is restricted to the oceans. It also focuses on the tropical belt: 30°N–30°S.

b. The radiative quantities

The ECMWF radiative code (Morcrette 1991) is used to compute the clear sky component of the LW flux profiles from the TOA to the surface. The water vapor continuum absorption used is described in Zhong and Haigh (1995). The computations are done without taking the radiative effect of clouds into account. The 3I procedure retrieves temperature information from the TOA to the surface. The water vapor profiles from the TOA to the surface. The water vapor to compute the clear sky component of the LW flux of specific humidities at six levels (1000, 850, 700, 500, 400, 300, 250, 200, 150, 100, 70, 50, 30, 20, 10 hPa) and precipitable water contents for the five lowest layers (below 100 hPa). An interface, called 3IFlux, described in Chéruy et al. (1996) and Chevallier et al. (2000), allows for retrieving vertical profiles to which the radiative transfer code can be applied. The PathA data are available under the form of temperatures at 15 levels (1000, 850, 700, 500, 400, 300, 250, 200, 150, 100, 70, 50, 30, 20, 10 hPa) and of specific humidities at six levels (1000, 850, 700, 500, 400, 300 hPa). An interpolation scheme linear in the logarithm of pressure is applied here in order to get the same inputs to the radiative transfer scheme as for PathB. Concerning ERA-15, the temperature and specific humidity profiles described on the 31 mandatory levels were extracted from the ECMWF archive and used as they were in input to the radiative transfer scheme. For all three cases, the climatology of McPeters et al. (1984) is used here to determine an ozone profile.

Various variables describe the LW radiative characteristics of the atmosphere. This study refers to six of them:

1) The OLR, already mentioned, is the only loss of energy from the earth–atmosphere system. It can be expressed as

\[
\text{OLR} = \pi \int_{-1}^{+1} \mu \, d\mu \int_{\nu} d\nu \left[ B_{\nu}(\text{SST}) \tau_{0}(0, \text{TOA}, \mu) + \int_{0}^{\text{TOA}} B_{\nu}(T_{s}) \frac{\partial \tau_{s}(z, \text{TOA}, \mu)}{\partial z} \, dz \right],
\]

where \( \nu \) is the frequency, \( \tau_{0}(\text{TOA}, z, \mu) \) is the monochromatic transmittance for isotropic radiation between altitude level \( z \) and the TOA, and \( \mu \) is the cosine of the zenith angle. Here \( B_{\nu}(\text{SST}) \) is the Planck function at frequency \( \nu \) for a temperature of SST.

2) The clear sky greenhouse effect, \( G \), or the total clear sky absorbed atmospheric radiation, is the difference between the LW flux emitted at the surface and the clear sky OLR:

\[
G = \sigma \text{SST}^4 - \text{OLR},
\]

where \( \sigma = 5.6698 \times 10^{-8} \text{ W m}^{-2} \text{ K}^{-4} \) and where the surface emissivity is set equal to the unity. Both the SST and the OLR are quasi-continuously available from the observations since the Earth Radiation Budget Experiment (ERBE) and since global-scale retrievals of SST, owing to Advanced Very High Resolution Radiometer data, are operationally produced.

3) The downward LW flux at the surface (DLR) represents the LW radiation emitted by the atmosphere toward the surface. It can be expressed as

\[
\text{DLR} = \pi \int_{-1}^{+1} \mu \, d\mu \int_{\nu} d\nu \int_{0}^{\text{TOA}} B_{\nu}(T_{s}) \frac{\partial \tau_{s}(z, \mu)}{\partial z} \, dz.
\]

4) The net LW flux at the surface (NETS) is the surface loss of LW radiative energy. It can be expressed as

\[
\text{NETS} = \sigma \text{SST}^4 - \text{DLR}.
\]

5) The total clear sky LW cooling, \( C_{T} \), measures the capacity of the clear sky atmosphere to cool by LW radiation:

\[
C_{T} = \pi \int_{-1}^{+1} \mu \, d\mu \int_{\nu} d\nu \left[ \int_{0}^{\text{TOA}} B_{\nu}(T_{s}) \frac{\partial \tau_{s}(z, \text{TOA}, \mu)}{\partial z} \, dz \right.
\]

\[
+ \int_{0}^{\text{TOA}} B_{\nu}(T_{s}) \frac{\partial \tau_{s}(z, 0, \mu)}{\partial z} \, dz
\]

\[
+ B_{\nu}(\text{SST}) (\tau_{s}(0, \text{TOA}, \mu) - 1)
\]

where \( C_{T} \) is the sum of a cooling-to-space term, a cooling-to-surface term, and a surface warming term. Equation (5) can also be easily broken down into the following components: (i) the difference between the OLR and the NETS or (ii) the difference between the downward LW flux at the surface, DLR, and \( G \). No direct observation of \( C_{T} \) exists yet on the global scale. But \( C_{T} \) can be estimated from radiative transfer.
6) The clear-sky LW cooling rates are the contribution of LW radiation to the temporal variations of the vertical temperature profiles. They are functions of the net radiative flux divergence:

$$C(z) = \frac{g}{C_p} \frac{\partial F}{\partial z};$$

where $F$ is the net LW flux at altitude level $z$, $C_p$ is the heat capacity at constant pressure, and $g$ the gravitational acceleration. Mean values of the vertical profiles of cooling rates in five coarse layers, roughly referring to the vertical resolution of the TOVS water vapor retrievals, are considered. The pressure levels at the boundaries are: 1013, 950, 850, 500, 300, 100 hPa. The 1000–950-hPa layer is isolated in order to take into account the dominating influence of the surface temperature in it over the oceans.

3. Geographical distribution of the OLR and of $C_T$

a. The clear sky OLR

Maps of the clear sky OLR are displayed for January and July 1988 for the PathA, PathB, and ERA-15 data in Fig. 1. They exhibit similar patterns that have been extensively described in the literature (see, e.g., Harrison et al. 1990). When comparing July and January maps, it is seen that the extrema follow the seasonal migration of the tropical dynamics: the intensification of the Hadley and Walker cells in the winter hemisphere and the meridional migration of the intertropical convergence zone.

The PathB OLR fields are seen to be more irregular than the two other fields. This may be due to the selection of the first guess in the retrieval process. Indeed, it is likely that the initialization of the retrieval with an analyzed field, as in the case of PathA, produces smoother retrievals than the initialization with the TIGR dataset.

If patterns are similar between the three datasets, values may significantly differ. The ERA-15 OLR has a positive bias of 5–10 W m$^{-2}$ with respect to PathA and PathB.

A possible source of bias between the datasets lies in their respective time samplings. This has two origins. The first origin is the cloud contamination problem discussed in section 2. Second, the polar orbit of the NOAA-10 satellite allows each tropical region to be seen only twice a day and at only fixed times (roughly 7:30 A.M. and P.M. for NOAA-10; besides, only the P.M. orbits are processed here—see section 2a). The hypothesis that a strong coherent diurnal cycle in the water vapor and temperature fields would bias the clear sky radiation field over ocean seems very unlikely. The effect of the cloud contamination on the sampling rate has been investigated by sampling the ERA-15 data according to the PathB sampling. This suppressed approximately two-thirds of the ERA-15 data, but remarkably only tiny differences in the OLR monthly means appear (not shown). Two hypotheses can be put forward to explain this result. Either there are no major differences in the large-scale structures of the temperature and water vapor...
fields when considering overcast and partly cloudy scenes, or if such differences exist, the ERA-15 system was not able to capture them. Thus, the ERA-15 system assimilated very few data appropriate to describe the humidity field in overcast or heavy precipitation-affected areas, especially over oceans. As a conclusion, the 5–10 W m\(^{-2}\) bias between the Pathfinder datasets and ERA-15 is linked to the difference in their respective descriptions of the 3D structure of the atmosphere, rather than to their respective time sampling. Now, swapping the PathB water vapor profiles for the ERA-15 ones in the PathB computations leads to a similar positive bias of 5–10 W m\(^{-2}\). This indicates that the differences in water vapor profiles are primarily responsible for the differences in the clear sky OLR.

The difference between the water vapor profiles from the three datasets are analyzed both in terms of total precipitable water vapor content (PWC) and in terms of water vapor vertical repartition. When comparing the PWC as a function of the SST for the three datasets, the ERA-15 PWC appears to be smaller than the PathA PWC up to about 27°C. It is also smaller than the PathB PWC (which is corrected with SSM/I—see section 2a) above 24°C. This contributes to give higher OLRs for ERA-15. In other respects, the vertical repartition of the water vapor can be characterized using the approach of Smith (1966). These authors approximate the water vapor profile by an exponential law:

\[
q(P) = q(P_0) \left( \frac{P}{P_0} \right)^\beta.
\]

The mean value of the \(\beta\) parameter has been computed for the three datasets over the tropical oceans. The \(\beta\) value obtained with ERA-15 is 6%–8% greater than the values obtained with PathA and PathB. Therefore, the ERA-15 water vapor is more concentrated in the lowest part of the atmosphere. This also contributes to increase the ERA-15 OLR compared to PathA and PathB, even though it is not possible to guess from this result which dataset is the most realistic.

Chevallier et al. (2000) show that the clear sky OLR from PathB has a negative bias of less than 2 W m\(^{-2}\) compared to ERBE, for instantaneous situations (ERBE S8). This may suggest that the ERA-15 OLR is overestimated. Now, independent calculations of clear sky OLR using the ERA-15 profiles and the radiative transfer code of Edwards and Slingo (1996) have been produced by Slingo et al. (1998), this product is known as the CLERA dataset. They show rather good agreement with the monthly ERBE clear sky product (ERBE S5): less than 5 W m\(^{-2}\) bias. This emphasizes the fact that slightly different results can be obtained using different radiative transfer models and justifies the use of a single radiative transfer scheme for all three datasets in the present intercomparison.

b. The clear sky LW cooling

Figure 2 shows the maps of \(C_T\) estimated from the various datasets for January 1988. The maxima occur in the descending part of the Hadley cells. Minimum occurs over subtropical regions of subsidence off the western coast of the continents. There is also a local minimum over the ascending branch of the Hadley cells, especially over the warm pool region. By comparing Figs. 1 and 2, some similarities appear between the geographical distribution of \(C_T\) and OLR. For instance, high values of the OLR occurring in the descending part of the Hadley cell in the western part of the North Pacific correspond to a maximum of cooling. Over the tropical ascending zones, local minima in both OLR and \(C_T\) are observed. This has been expected since Rodgers and Walshaw (1966) established that in numerous cases the cooling-to-space term dominates the total cooling [first term on the right side of Eq. (5)]. However, important discrepancies can be emphasized, especially in the South Pacific convergence zone, which does not show up in the map of \(C_T\), or over the Atlantic. In these regions the cooling-to-surface term and the surface warming are likely to play a very significant role.

By comparing the maps of \(C_T\) from the various datasets, no bias shows up, but the agreement between the general features is not as good as for the maps of the OLR. As for the OLR, the discrepancies can be related to differences in the water vapor profiles. Differences are identified in the ascending branch of the Hadley
cells, especially over the warm pool, where the local minimum of $C_T$ is less visible in the ERA-15 estimates than in the TOVS data, especially the PathA data. The above-mentioned discrepancies in the estimation of both the PWC and its vertical distribution explain this disagreement. The minimum of cooling occurring over the subtropical subsidence zones off the western coast of the continent does not appear in the PathA fields. This is related to the overestimation of the PWC by the infrared-based retrievals addressed in section 2a. As the PathB data have been corrected to account for this effect, the PathA estimates of $C_T$ may be less reliable in these regions than the PathB and the ERA-15 estimates.

4. Relationship between the LW cooling and the SST

a. Description of the variations of the clear sky cooling and the SST on a regional basis

As an extension of the previous studies of the relationship between $G$ and SST (quoted in the introduction), this section aims at analyzing the links between $C_T$ and SST. The variations of $C_T$ binned into 1°C intervals are examined in conjunction with the corresponding mean variations of the mean temperature and water vapor profiles for the months of January and July 1988.

As already mentioned in section 3, large differences exist in the absolute value of $C_T$ between the different datasets. A maximum bias of 15 W m$^{-2}$ is observed in Fig. 3 when comparing $C_T$ in January for PathB and PathA at 20°C. The bias decreases as the SST values increase up to 28°C.

However, two different regimes may be identified in all the datasets. For SST values less than 27°C, the cooling increases as the SST increases, and remains nearly constant or decreases for SST values greater than 27°C. The behavior of $C_T$ for SST values above 27°C is related to the super-greenhouse effect first observed by Raval and Ramanathan (1989) in ERBE data. The super-greenhouse effect, as defined by Ramanathan and Collins (1991), refers to the fact that for the highest values of the SST, the increase of the greenhouse effect with the surface temperature is larger than the increase of the surface emission. In Fig. 3, it appears that at the same time, the rate of increase of the atmospheric cooling is significantly reduced, if not inverted. This fact will be discussed in more details in sections 4c and 4d.

For SST values less than 27°C, the cooling is less pronounced in January (not shown), whatever dataset is considered. Indeed, when looking at the water vapor profiles, it appears that the upper troposphere is slightly drier in July compared with January, while the temperature profiles are nearly identical. These differences in the mean water vapor profiles may be due to variations in the tropical vertical motions, and particularly to a predominance of the subsiding motions in July in relation to the seasonal migration of the Hadley cells as well as to the land–ocean asymmetry between the two hemispheres. This illustrates the fact that, on a seasonal basis, the variations of the radiative cooling as a function of the SST variations are not direct. In particular, the large-scale atmospheric circulation plays a significant role in governing these variations independently of the SST variations (Bony et al. 1997).

b. Sensitivity of the variations of $C_T$ to SST variations on a seasonal basis

In the previous section, the spatial variations of the sensitivity of clear sky radiative cooling to geographic SST variations have been described. However, it is well recognized that the sensitivity of the radiative parameters varies according to the type of variability considered: geographic, seasonal, interannual, or even climatic (Duvel and Breon 1991; Bony and Duvel 1994). Indeed, as illustrated in section 4a, the variations of the radiative cooling as a function of the SST variations are not direct. Besides, due to the large thermal inertia of the oceans, the characteristic timescale of the ocean response to atmospheric changes is much longer than that one of the atmospheric response to oceanic changes. As a consequence, feedback processes can be or not be activated according to the timescale considered.

In this section, the sensitivity of the variations of $C_T$ to SST variations on a seasonal basis is investigated. Following the approach of Bony et al. (1995), linear regression coefficients between $C_T$ and SST seasonal variations are computed. For each region referenced by its mean SST value, a sensitivity parameter $S_{C_T}$ is computed. In the present case, the mean SST values co-
Fig. 4. Clear sky LW cooling sensitivity parameter computed on the basis of seasonal changes in $C_T$ (derived from PathA, PathB, and ERA-15 data) and SST. Seasonal changes correspond to regional change between Jan 1988 and Jul 1988 and between Jan 1990 and Jul 1990. The vertical bars represent the standard deviation around the mean values within each SST interval.

respond to the monthly means for July and January in 2°C intervals. Here $S_{CT}$ characterizes the changes of $C_T$ with regard to local SST changes. It is given by the least-squares fit to

$$\delta C_T = S_{CT} \delta SST,$$  \hspace{1cm} (8)

where $\delta$ corresponds to the variations between the January and the July values. The sensitivity parameter $S_{CT}$ computed on the PathA, PathB, and ERA-15 data for 1988 and on PathB and ERA-15 data for 1990, is reported in Fig. 4. While substantial differences exist between the various curves, their shapes are similar. Moreover, for a given dataset, the seasonal variations are consistent for both years. Here $S_{CT}$ progressively decreases from the cold to the warm tropical ocean regions. It is positive for the low values of the SST, but is significantly larger in ERA-15 than that in the Pathfinder datasets. For the highest SST values, $S_{CT}$ becomes negative, especially for PathA. This means that an increase of SST corresponds to a decrease of cooling. As a conclusion, on a seasonal basis as well as on the regional basis, the super-greenhouse effect is manifested in the total cooling as a loss of the capacity of the atmosphere to radiatively regulate its temperature.

c. Sensitivity study for the conditions favorable to the super-greenhouse effect

The thermodynamic conditions that allow the LW cooling not to increase for SST values greater than $27^\circ$C are analyzed in this section. The aim of this study is to find synthetic profiles, corresponding to SST values greater than $27^\circ$C, for which the slope of the curve of $C_T$ versus SST would be the same as the one observed for SST values less than $27^\circ$C.

On Figs. 5a and 5b, the binned thermodynamic profiles corresponding to a SST value of $25^\circ$C are presented
as an example of the “normal” conditions, while those corresponding to a SST value of 29°C illustrate the “favorable to super-greenhouse effect” (FSGE) conditions. Consistent with previous studies (e.g., see Inamdar and Ramanathan 1994), FSGE situations appear to be characterized by more unstable (Fig. 5a) and moister conditions (Fig. 5b), compared to the “normal” ones.

For SST values greater than 27°C, the binned profiles are modified as follows. (i) The temperature profile is modified such as to get the same vertical gradient as the binned temperature profile of which SST is 5 K colder (case A, see Fig. 6a for an example with the PathB data and Fig. 6b for ERA-15 data), the current SST being kept. For the ERA-15 profiles, the temperature perturbation does not affect the surface (1000–850 hPa) layer. The lapse rate is reduced in the 850–700-hPa layer, leading to a warmer profile above 850 hPa. For the Pathfinder profiles, the perturbation reduces the low troposphere lapse rate (1000–850 hPa), this leads to a warming of the whole troposphere. (ii) The water vapor profile is modified such as to get the water vapor profile corresponding to a 5-K colder SST but conserving the total content (case B, see Fig. 7 for an illustration in the case of the PathB data), the atmosphere remaining unsaturated. Radiative computations with the perturbed profiles including the temperature only perturbation (case A), the water vapor only perturbation (case B), and both temperature and water vapor perturbations (case C), are carried out.

Results of the sensitivity experiment conducted on PathB and ERA-15 for January 1988 are reported in Fig. 8.

The impact of the temperature perturbation is similar for the various datasets: it induces an overall increase of the cooling but does not modify its sensitivity to SST variations.

For all datasets, the water vapor perturbation (case B) is the most efficient to modify the sensitivity of the total cooling to SST variations, for the highest values of the SST. The modification of the water vapor profile results in a drying of the upper troposphere and a moist-
ening of the lowest troposphere (the case of PathB is reported in Fig. 7). Conversely, super-greenhouse situations are made possible by an increased humidity in the middle and upper troposphere. If the role of the convection is to increase the upper-tropospheric humidity, one may argue for a significant convection–radiation interaction over the warm tropical oceans.

d. Radiative aspects

In this section, the relative contributions of \( G \), of the DLR, and of the NETS to the variations of \( C_T \) as the SST increases, are analyzed. Their respective values are reported in Table 1 for all datasets for the month of January 1988. The overall rate of increase of the DLR with the SST is faster (about 30%) than the rate of increase of \( G \) with the SST. For instance, considering the PathB data, the DLR increases by 65 W m\(^{-2}\) for SST values increasing from 22°C to 30°C when \( G \) increases by only 49 W m\(^{-2}\). At the same time, the NETS decreases by only 17 W m\(^{-2}\). In other words, due to the faster increase rate of the DLR with the SST compared to \( G \), the atmospheric cooling tends to increase as the SST values increase. These results are consistent with those of Inamdar and Ramanathan (1994), who computed the clear sky radiative parameters from radiosonde reports grouped into “nonconvective” and “convective” categories using the SST as an index for the convective activity. Their first class refers to SST values between 20°C and 25°C and the second one to SST values greater than 27°C. Inamdar and Ramanathan found that the DLR increases by 62 W m\(^{-2}\) when \( G \) increases by 37 W m\(^{-2}\) and the net flux at the surface decreases by 29 W m\(^{-2}\).

Table 2 focuses on the variations of the components of \( C_T \) for a 3°C SST increase in each of the classes: normal (SST \(\leq 27°C\)) and FSGE (SST \(\geq 28°C\)). The agreement is excellent for \( G \) between the three datasets, but discrepancies appear for the other components especially the NETS, in which the rate of variation can be twice or three times as big from one dataset to another. The rate of variation of the DLR is also much smaller from the PathA dataset, especially for the high-

| Table 2. Increase rate of \( C_T \), of \( G \), of the total downward emitted flux (DLR), and of the net LW flux at the surface (NETS), for normal (SST values 22°C–25°C) and favorable to super-greenhouse effect (SST values 27°C–30°C) conditions (warmer temperature variable value minus colder temperature variable value), for Jan 1988. |
| SST | \(\delta C_T\) W m\(^{-2}\) | \(\delta G\) W m\(^{-2}\) | \(\delta DLR\) W m\(^{-2}\) | \(\delta NETS\) W m\(^{-2}\) |
| \(\delta SST\) | 5.0°C | 0.5°C | 5.0°C | 0.5°C |
| 22°C–25°C PathB | 12.9 | 13.2 | 26.0 | -8.3 |
| 22°C–25°C ERA-15 | 15.8 | 13.2 | 29.1 | -11.2 |
| 22°C–25°C PathA | 8.4 | 12.9 | 20.9 | -7.4 |
| 27°C–30°C PathB | -1.1 | 25.6 | 24.5 | -5.7 |
| 27°C–30°C PathA | -9.2 | 23.2 | 14.0 | 4.5 |
est values of the SST. However common behavior can be depicted: consistently with the previous studies, and according to the definition of the super-greenhouse effect, the rate of increase of $G$ grows from normal to FSGE conditions (13 vs 24 W m$^{-2}$, for PathB). PathB and ERA-15 agree with a decrease of the surface loss of LW radiative energy from normal to FSGE conditions, and therefore a tendency toward an increased surface warming. This is also consistent with previous studies (e.g., see Manabe and Wetherald 1967). The increase rate of the DLR decreases slightly (respectively, 1 W m$^{-2}$ for PathB, 3.5 W m$^{-2}$ for ERA-15 and 6.9 W m$^{-2}$ for PathA in January 1988) from normal to FSGE conditions. The weakness of the DLR variation rate modification with respect to the SST variations is explained by the following. The mid- and upper-tropospheric water vapor is mostly responsible for the super-greenhouse effect. Now, most of the DLR (50% in the case of the tropical McClatchey atmosphere) is emitted in spectral regions that are saturated in the lower troposphere at low latitudes: the 15-μm region of the carbon dioxide, and the 6.3- and 25-μm vibration–rotation bands of the water vapor. Only a small part (13% in the case of the tropical McClatchey atmosphere) is emitted in the 10–12-μm atmospheric window, in which the tropical DLR is sensitive to the variations of the middle and upper tropospheric water vapor. Therefore the super-greenhouse effect little affects the DLR. Moreover, it is remarkable that for the highest values of the SST, the variation rate of $G$ is nearly identical to the decrease rate of the DLR. Therefore, the faster increase rate of $G$ induces a canceling of the atmospheric radiative cooling.

5. Vertical variations of the cooling rates

In this section, the variation of the LW atmospheric cooling is analyzed as a function of both the SST and the vertical pressure. As indicated in section 2b, five vertical layers are considered. They are bounded by the following pressure levels: 1000, 950, 820, 500, 300, 100 hPa. The maps of the geographical distribution of the mean cooling rates (not shown) substantially differ from one another. This is obviously a consequence of the differences between the temperature and water vapor 3D structure as it is estimated from the three methods. However, common behaviors are observed when considering the vertical profiles of cooling binned as a function of the SST values, $C_{\text{SSTc}}(P)$. Figure 9a shows the vertical profiles of cooling $C_{\text{SSTc}}(P)$ for a mean SST value of 23.5°C (normal case) and $C_{\text{2Pc}}(P)$ for a mean SST value of 29.5°C (FSGE case) for the three datasets. In the surface layer, the cooling is maximum and increases as the SST value increases. When considering the middle troposphere (roughly 800–500 hPa), the cooling decreases with the altitude at a faster rate for the FSGE situations than for the normal ones, leading to a more pronounced cooling in the normal situations (about 2.5 K day$^{-1}$) than in the FSGE situations (about 2 K day$^{-1}$). Above 400 hPa, the FSGE atmosphere cools the most.

In section 4d, the variations of $C_T$ for a 3°C SST increase have been studied. Figures 9b and 9c present the vertical distribution of these variations, $\delta C_T(P)$, respectively, for the normal and FSGE conditions. One observes that $\delta C_T(P)$ is nearly constant along the vertical for the normal situations (about 0.2 K day$^{-1}$ increase for a 3-K SST increase). Concerning the FSGE situations, $\delta C_T(P)$ is positive in the upper troposphere and negative in the 900–400-hPa layer. Concerning the surface layer, there is no agreement between the datasets: for the PathA and PathB data $\delta C_T(P)$ is positive, and for the ERA-15 data $\delta C_T(P)$ is negative. This is due to the different vertical resolution of the estimated temperature profiles in the three datasets. In ERA-15, the mean difference between the SST and the atmospheric 1009-hPa temperature increases with SST above 20°C, the SST being warmer. At 28°C the difference is around 1°C. For the highest SSTs, the radiative cooling of the boundary layer is more than compensated by the warming from the surface. This cannot be observed in the Pathfinder data, due to the coarse resolution in the 850–1013-hPa layer, as retrieved from TOVS. However, without any comparison with independent dataset, it is difficult to assess the quality of ERA-15 in the planetary boundary layer over tropical oceans.

6. Summary and discussion

Since significant uncertainties exist in the presently available estimations of the vertical thermodynamic structure of the atmosphere, three different datasets have been used to study the clear sky LW cooling of the atmosphere over tropical oceans. Two of them come from the reprocessing of the TOVS data in the framework of the Pathfinder project. The third one is the ECMWF reanalysis archive. All of them have a far better spatial and temporal sampling than the radiosoundings, already used in previous studies.

It appears that the three datasets agree for the main geographical features of the LW cooling. However, important differences exist between them. They can be related to identified differences in the temperature and water vapor fields. Surprisingly, the sampling inherent to the cloud contamination problem of the TOVS Pathfinder data does not seem to induce substantial differences when comparing to the ERA-15 data. Two possibilities have been considered: either there are no substantial differences in the large-scale structure of the temperature and the water vapor fields for the clear/partly cloudy scenes and the overcast scenes, or the ERA-15 is not able to capture them. In the absence of a more reliable description of the vertical structure of the atmosphere, no firm conclusions can be drawn. Future progress is expected from the microwave sounding instruments such as those from the Advanced MSU,
which is part of the Advanced TOVS instrument and which allows it to partly overcome the difficulty linked with the infrared cloud contamination problem. Also, the assimilation of radiance measurements such as from the SSM/I and even the use of Precipitation Radar (PR) measurements in conjunction with the radiances of the multichannel passive microwave sensor, TMI, aboard the Tropical Rainfall Measuring Mission (TRMM) (Simpson et al. 1996), will help the future.

The basic features of the sensitivity of $C_T$ can be consistently derived from the three different datasets used here. For the low SST values, the radiative cooling increases as the SST increases. For the highest SST values, the radiative cooling tends to be constant or to decrease. This is closely connected with the super-greenhouse effect largely depicted in the literature but not equivalent. The differences between normal and FSGE conditions observed in the LW radiative budget are summarized in Table 3. Conclusions made possible by the availability of the vertical distribution of the LW fluxes are reported in the last five rows of the table.

On a thermodynamic point of view, the role played by the temperature profile seems to be minor. Consistent with studies conducted on the super-greenhouse effect, the vertical distribution of water vapor appears to be the key factor for governing the sensitivity of the LW cooling at low latitudes. The increased humidity of the middle and upper troposphere for high values of the SST is responsible for the different sensitivity of $C_T$ to SST variations for the highest values of the SST. Conse-
Table 3. Summary of the results on the rate of variation of $G$, $C_T$, DLR, and $C(P)$ obtained in this work. The arrows indicate how the variables evolve from normal to super-greenhouse conditions. For instance, the greenhouse effect increases from normal to super-greenhouse conditions. New results are reported in the last five rows.

<table>
<thead>
<tr>
<th>Variable/δSST</th>
<th>Thermodynamic conditions</th>
<th>Radiation observations required</th>
</tr>
</thead>
<tbody>
<tr>
<td></td>
<td>Normal</td>
<td>Super-greenhouse</td>
</tr>
<tr>
<td>$G$</td>
<td>$&lt;4\sigma T$</td>
<td>$&gt;4\sigma T$</td>
</tr>
<tr>
<td>$C_T$</td>
<td>$&gt;0$</td>
<td>$=0$</td>
</tr>
<tr>
<td>DLR</td>
<td>$&gt;\delta G/\delta SST$</td>
<td>$=\delta G/\delta SST$</td>
</tr>
<tr>
<td>$C$ (upper troposphere)</td>
<td>$&gt;0$</td>
<td>$&gt;0$</td>
</tr>
<tr>
<td>$C$ (middle troposphere)</td>
<td>$&gt;0$</td>
<td>$&lt;0$</td>
</tr>
<tr>
<td>$C$ (lower troposphere)</td>
<td>$&gt;0$</td>
<td>?</td>
</tr>
</tbody>
</table>

Fig. 10. Variations of $C_T$ binned into 1° SST intervals for PathB data for Jan 1988. The vertical bars shows the standard deviation computed in each SST interval over the month.

Consequently, if the convection is the process responsible for the humidification of the upper troposphere, the lesser efficiency of the radiative cooling at the highest SST values is made possible by the convection.

The sensitivity of the clear sky LW cooling variations to SST variations in super-greenhouse conditions has been shown to be due to the fact that, at high SST values, the atmosphere does reradiate the radiation toward the surface as efficiently as it traps it.

Considering the variations of the cooling rates in five coarse atmospheric layers, the vertical distribution of the total LW cooling has been studied for normal as well as for favorable to super-greenhouse effect conditions. Above 27°C, with increased SST the radiative cooling is intensified in the upper troposphere and decreases below 400 hPa. The coarse resolution of the TOVS-retrieved temperature profiles in the lower troposphere prevents us from deriving the radiative cooling rates with sufficient accuracy in the boundary layer. Future studies will have to link the vertical distribution of the clear sky radiative cooling to that of the other diabatic and adiabatic cooling/warming terms.

This study has been restricted to the clear sky contribution to the LW cooling, the case of the clouds being of even more complex issue. However, a test has been performed on the PathB data for which the cloud contribution has been tentatively introduced in the computation of the total LW cooling (Chérut et al. 1996). The curve of the variations of $C_T$ as a function of the SST is reported in Fig. 10. Compared to the clear sky case, the slopes are modified and standard deviation materialized by the vertical bars is higher. However, as for the clear sky case, the sensitivity of $C_T$ to SST variations is different for the lowest (less than 27°C) and for the highest values of the SST. This gives confidence on the possible enlargement of this work to the case of the cloudy skies. Progress in the retrieval of the vertical structure of the cloudiness is badly needed before the vertical distribution of the LW radiative budget taking the cloud contribution into account can be derived with sufficient accuracy.

Besides the direct analysis of one of the most important regulating mechanisms of the temperature of the earth–atmosphere system, the radiative cooling, this study can also be used in order to evaluate the ability of general circulation models to simulate it correctly. Since the radiative cooling is more sensitive to the vertical thermodynamic structure of the atmosphere than the more widely studied greenhouse effect, it can allow more constraining tests to be performed.

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