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# The effect of aerosols on long wave radiation and global warming

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ABSTRACT

The effect of aerosols on long wave (LW) radiation was studied based on narrowband LW calculations in a reference mid-latitude summer atmosphere with and without aerosols. Aerosols were added to the narrowband LW scheme based on their typical schematic observed spectral and vertical behaviour over European land areas. This was found to agree also with the spectral aerosol data from the Lan Zhou University Semi-Arid Climate Observatory and Laboratory measurement stations in the north-western China.

A volcanic stratospheric aerosol load was found to induce local LW warming and a stronger column "greenhouse effect" than a doubled $CO_2$ concentration. A heavy near-surface aerosol load was found to increase the downwelling LW radiation to the surface and to reduce the outgoing LW radiation, acting very much like a thin low cloud in increasing the LW greenhouse effect of the atmosphere. The short wave reflection of white aerosol has, however, stronger impact in general, but the aerosol LW greenhouse effect is non-negligible under heavy aerosol loads.



## 1. Introduction

The earth is in a near radiation balance, the outgoing long wave (thermal) radiation (OLR) closely balancing the absorbed solar radiation. The effect of clouds on OLR and on the cloud radiative forcing (CRF), can be estimated from satellite data by taking the difference between the clear sky scenes and all scenes. These observations indicate that clouds increase the planetary shortwave (SW) albedo by 15% to 30%, thereby reducing the absorbed solar radiation by about 50 W/m$^2$. This cooling effect is opposed by the warming effect of clouds on the longwave (LW) radiation (the "cloud LW greenhouse" effect), which reduces the OLR by about 31 W/m$^2$ (Hartmann, 1994) on the average.

This study concentrates on the analogous long wave radiative effect ("LW forcing") caused by airborne aerosols (other than clouds), which is less well-known than that of the clouds. During the 1970s, the influence of the aerosol layer height and the changes of surface albedo on the atmospheric radiation balance were investigated by Reck (1974, 1975). The results of those pioneering studies showed that like the clouds, aerosols produce two opposing effects in the atmosphere: they cause heating of the Earth's surface by enhancing the downwelling LW radiation, but they also increase the planetary SW albedo, which causes a cooling effect. The combined effect depends on many factors including the aerosol type, concentration and height. It also varies on time due to the diurnal and seasonal changes in incoming solar radiation. The cooling effect due to a reduction of the incoming solar radiation often dominates at daytime while the weaker warming effect due to the aerosol LW emission is present throughout the day and may be observed at night time.

In the 1980s, new methods were developed for investigating the aerosol characteristics and their effects on the albedo and climate. These include for example the multi-wavelength satellite extinction measurements (Lenoble, 1986), and balloon or aircraft measurements. At the same time the focus also turned towards the effects of volcanic

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aerosol loads in the stratosphere as well as to the effects of aerosols on the local climate in specific locations. The latter was investigated for example in the city of St. Louis, USA in (Method and Carlson, 1982). These studies showed that the effects of aerosols are similar to those of a thin cloud at the same height. The impact is small in magnitude, however and somewhat difficult to measure unless the aerosol concentration is extremely high. The 1990s saw significant increase of research on aerosols and their effect on climate. First computer models of the effects of aerosols on the radiation balance were developed. One such model is presented by Claquin et al. in (Claquin et al., 1997).

In the 21st century, the work of studying and understanding the effects of both natural and anthropologic aerosols on the radiation balance both globally (Dammann et al., 2000) and locally (Shaocai et al., 2001; Han et al., 2012) has continued. The effects of specific types of aerosols or effects of aerosols in specific locations have also been studied (Verma et al., 2006; Wendisch et al., 2008). Most recently, the radiative effects of aerosols have been studied in both urban and remote areas of western India in 2011 (Ramachandran and Kedia, 2011), and Europe (Péré et al., 2012), USA (Mickley et al., 2012) and China (Zhang et al., 2012) in 2012. The results show significant variability of the radiative effects due to the meteorological conditions as well as the aerosol loads themselves. Therefore the aerosol processes and meteorological processes appear to be coupled and they interact with each other. This is best studied by using atmospheric and radiative models equipped with aerosol schemes and using aircraft and satellite measurements. In this way, for instance, the major Saharan dust storms have been shown to imply considerable differences into the surface LW fluxes and OLR (Haywood et al., 2005; Slingo et al., 2006; Hansell et al., 2010). However, such coupled studies, although the most complete, are dominated by the strong daytime SW effects of dust, and so may not be optimal in isolating and characterizing the LW effects and mechanisms.

In this study the OLR differences, the LW surface budget differences and the internal LW heating/cooling rates are studied by comprehensive narrowband LW model calculations, using various aerosol loads in typical mid-latitude conditions. In particular, the observed aerosol loads of north-western China are used as an extreme example, because the wind-blown mineral dust from the surrounding deserts and the heavy industrial pollution in the city of Lan Zhou provide quite large natural and anthropologic aerosol loads for this region. This was analyzed by making LW calculations in a typical mid-latitude summer air column (MLS case) (Ellingson et al., 1991) with a narrow-band spectral LW scheme (Savijärvi, 2006), while introducing variable aerosol, cloud, and greenhouse gas loads into the scheme. The results for the different cases were compared. In particular the aerosol effects on the OLR, on the down-welling LW radiation at the surface (DLR), and on the internal LW heating rate in the atmosphere (LH) with different aerosol and greenhouse gas loads was analyzed.

## 2. Methodology

### 2.1. The long-wave radiation scheme

Long wave radiation in the Earth's atmosphere is defined as the electromagnetic radiation at wavelengths longer than 4 μm, usually from terrestrial origin. The short wave (SW) radiation wavelengths are less than 4 μm. It is usually from solar origin. The thermal LW and solar short wave radiation propagate in the atmosphere, experiencing absorption, emission, scattering and reflection. Unbalanced radiation will lead to variations in the atmospheric, ground and ocean temperatures and air movements (i.e. winds, weather and climate).

The ground and ocean are the main direct heat sources for the troposphere. They absorb solar radiation, for which the atmosphere is relatively transparent. The water vapour, $CO_2$ and other types of greenhouse gases in the atmosphere have varying abilities to absorb and emit LW radiation. Therefore the LW radiation emitted by the sun-heated ground will propagate through the atmosphere back to space only in the spectral LW window(s) of the greenhouse gases, since $O_2$ and $N_2$ are transparent to LW radiation. In general, 75% to 95% of the LW emission of the ground is absorbed by the water vapour, $CO_2$, $O_3$ and other greenhouse gases in the troposphere, and re-emitted at the air temperatures to all directions, hence partially back to the ground, creating the "greenhouse effect". How aerosols then impact this LW greenhouse effect is the subject of this study.

The LW radiation scheme (taken from (Savijärvi, 2006)) calculates the upwelling and downwelling LW fluxes ($F_{up}$, $F_{down}$) at each altitude from solutions to the plane-parallel equation of radiative transfer, using the absorption (nonscattering) approximation with a diffusivity factor of 1.66, and assuming local thermodynamic equilibrium. The spectral fluxes at each wave number $k = 1/\lambda$ for a narrow band $\Delta k$ around k were calculated with a statistical narrow-band model (NBM) for the gaseous transmissivity $t_{gas,k}$ at each band. The NBM covers the 0–1200 $cm^{-1}$ wave number range in 48 bands ($\Delta k = 25\ cm^{-1}$), the 1200–2100 $cm^{-1}$ range in 18 bands ($\Delta k = 50\ cm^{-1}$), and the 2100–2500 $cm^{-1}$ range in one band; so there are 67 bands in the LW range. The band parameters for water vapour, $CO_2$ and $O_3$ were taken from (Houghton, 2002). The Goody random band model was adopted for water vapour, the Malkmus model for $CO_2$ and $O_3$, and the Curtis-Godson method was used for line pressure broadening along inhomogeneous vertical paths. The Roberts *et al.* scheme (1976) (Roberts et al., 1976), augmented with foreign-broadened contribution, was used for the important water-vapour continuum effect in this study, where we concentrate on the boundary layer, although we recommend the more comprehensive Clough et al. (1992) continuum scheme (Clough et al., 1992) for studies that would concentrate on the upper troposphere.

The local LW heating rate of air is obtained as the vertical convergence of the total net LW flux $F_{net} = F_{up} - F_{down}$. Thus at each height *z*, LH is given by:

$$LH(z) = \left(\frac{\partial T}{\partial t}(z)\right)_{LWR} = -\frac{1}{\rho C_p}\frac{\partial}{\partial z}\left(F_{up} - F_{down}\right) = \frac{g}{C_p}\frac{\partial F_{net}}{\partial p} \quad (1)$$

Here $\rho$ is the density of air and $C_p$ the specific heat of air at constant pressure $p$. The last form follows from the hydrostatic relation.

The NBM should be validated before using it for aerosol-laden atmospheres. This was made in (Savijärvi, 2006), where the key LW flux values were compared with results from the International Comparison of Radiation Codes in Climate Models