Spectral Model for Clear Sky Atmospheric Longwave Irradiance

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Purpose and Background

The Earth and its atmosphere absorb/scatter the shortwave solar radiation (< 4 μm) and absorb/emit/scatter longwave radiation (> 4 μm). The balance between the shortwave radiation and the longwave radiation determines the temperature in the atmosphere and on the Earth surface. Therefore, the surface downwelling longwave irradiance plays a critical role in the modeling of weather and climate variability as well as heat balance design of solar power plants and radiant cooling systems. The downward longwave atmospheric irradiance can be measured directly by pyrgeometers, but they are not equipped widely in weather stations because they are relatively expensive and require extensive calibration and adjustments to exclude the longwave radiation emitted by surroundings. Therefore, the focus of this work is to develop a model that can calculate the atmospheric downwelling longwave radiation at the ground level for different locations, and also examine the effects of water vapor and aerosols on the surface downwelling longwave irradiance with high spectral and spatial resolutions.

Methodology

A spectrally resolved radiative model is used to calculate surface downwelling longwave (DLW) irradiance ($0 \sim 2500\ cm^{-1}$) under clear sky (cloud free) conditions. The wavenumber spectral resolution of the model is $0.01\ cm^{-1}$ and the atmosphere is represented by 18 non-uniform plane-parallel layers with the pressure of each layer determined on a pressure-based coordinate system. The standard AFGL profiles for temperature and atmospheric gas concentrations have been adopted with the correction for current surface atmospheric gas concentrations. The model utilizes the most up-to-date HITRAN molecular spectral data for 5 atmospheric gases: H_2O, CO_2, O_3, CH_4 and N_2O. The MT_CKD model is used to calculate water vapor continuum absorption coefficients. The downwelling and upwelling monochromatic fluxes of each layer are calculated by integrating diffuse radiation from all directions via exponential integral.
For scattering atmosphere (with aerosols), the aerosol concentration is assumed to be constant below 2 km and has an additional plume between 2 km and 3 km. Aerosol size distribution is assumed to follow a bimodal distribution. The size and refractive index of aerosols change as they absorb water, therefore the size distribution and refractive index are corrected for different surrounding water vapor concentrations (relative humidities). The longwave absorption coefficients, scattering coefficients and asymmetry factor of aerosols are calculated from refractive index and size distribution by Mie scattering theory. With the aerosol absorption and scattering within each layer, the radiosity and irradiance of each layer is calculated by energy balance equations using transfer factors with the assumption of isotropic scattering (asymmetry factor is calculated to be smaller than 0.03). The monochromatic downwelling and upwelling fluxes of each layer are further calculated from radiosity and irradiance by modifying the transfer factors. Broadband fluxes are integrated using k-distribution method for both non-scattering and scattering atmosphere.

Results and Conclusions

For a non-scattering atmosphere, a resolution of 18 vertical layers is found to achieve grid convergence. The calculated surface DLW irradiance agrees within 2.95% with mean values from the InterComparison of Radiation Codes in Climate Models (ICRCCM) program, with spectral deviations below $0.035\ W/(m^2\ cm^{-1})$.
For a scattering atmosphere with typical aerosol loading (surface aerosol optical depth of 497.5 nm is 0.1), the DLW calculated by the proposed model agrees within 2.97% relative error when compared to measured values at 7 climatologically diverse SURFRAD stations. This relative

error is smaller than a calibrated parametric model regressed from data for those same 7 stations, and within the uncertainty (±5 W m^{-2}) of pyrgeometers commonly used for meteorological and climatological applications. The contributions of water vapor are mostly from the atmosphere below 1 km in the spectral range of 400 ~ 650 cm^{-1}. Contributions also come from the bands 750 ~ 1400 cm^{-1} and 1700 ~ 2300 cm^{-1}, for heights below 3.45 km. In some spectral regions above 3.45 km, more water vapor contributes to a decrease of transfer factor to the surface, because water vapor in the layer(s) below absorbs the emitted radiation, preventing the radiation coming from above from reaching the surface.
The DLW increases by 2.06 ~ 8.21 W m^{-2} when compared with non-scattering conditions, and this increment decreases with increased water vapor content due to overlap with water vapor bands. The aerosol contribution to the transfer factors is mostly felt within the spectral atmospheric windows: 400 ~ 650 cm^{-1}, 750 ~ 1400 cm^{-1} and 1700 ~ 2300 cm^{-1}, where the single scatter albedo is high. The latter two windows have negligible effect on the monochromatic flux density because the blackbody intensity in these bands is low. Aerosol forcing mostly comes from the layer between 2 to 3 km due to the additional aerosol plume resided in that region. Above 3 km, the band 1050 ~ 1150 cm^{-1} has a decrease in transfer factor and monochromatic flux density, because aerosols below prevent the emitted longwave radiation from reaching the surface.

Keywords: Discrete ordinates modeling, Line-by-line, Surface downwelling longwave irradiance, water vapor continuum absorption, aerosol scattering

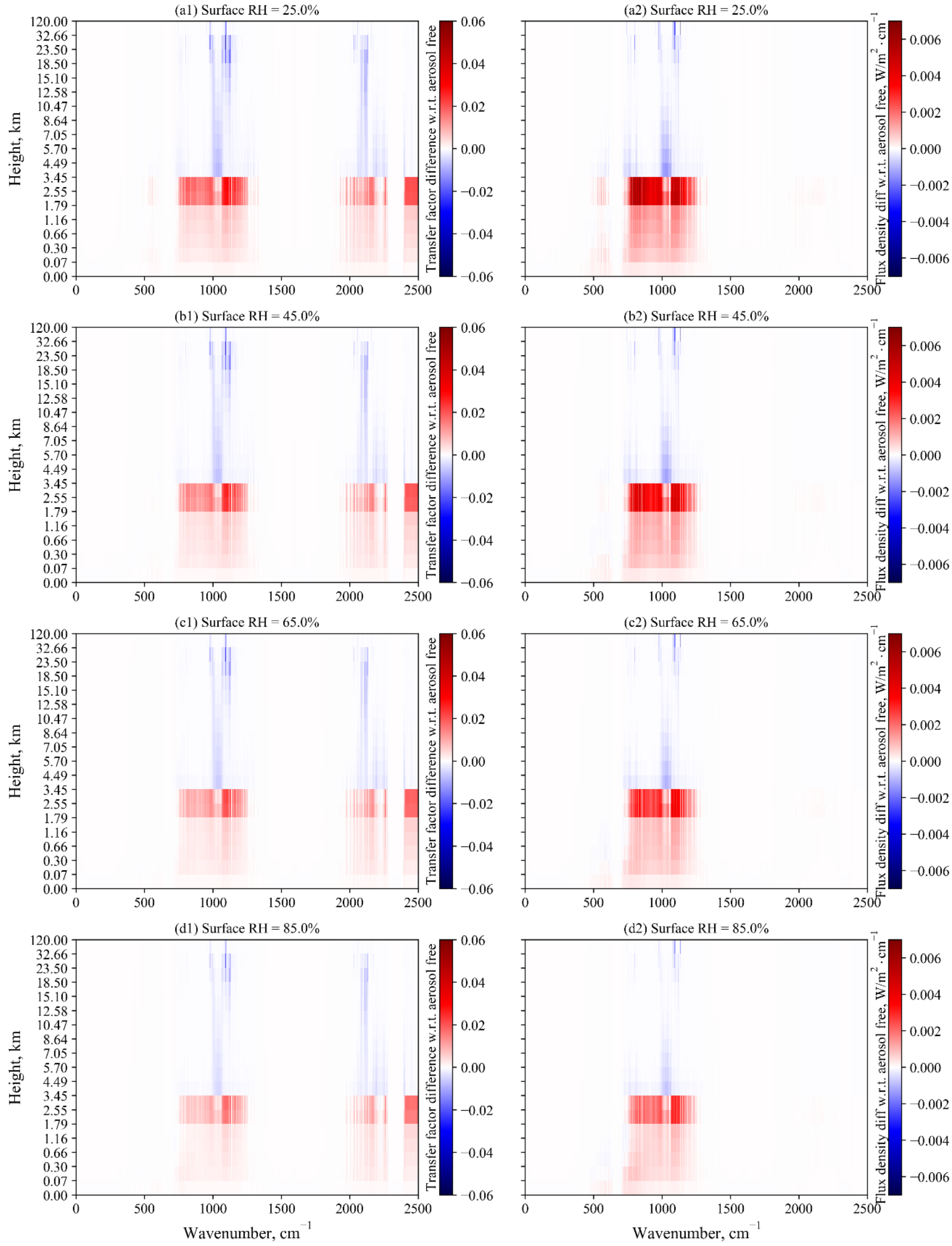

Figure 1. Illustration of the difference w.r.t. aerosol free of transfer factor and flux density to surface from different layers of atmosphere and at different wavenumbers. The left column shows the difference of transfer factor. The right column shows the difference of flux density.