ATMS 532. Homework 2. Due Wednesday 22 January 2014.

1. Petty 2.12 (eclipses)

2. Orbital Eccentricity of Earth and Mars

Define S as the solar flux (W m^{-2}) received at the top of the atmosphere.

(a) The Earth-Sun distance r_{es} at aphelion is 1.75% greater than the average distance; the distance at perihelion is 1.75% less than the average distance. This means the *eccentricity* of Earth's orbit is 1.75%, or 0.0175. [The eccentricity has varied between 0 and 7% during the last million years.] Using the present value of eccentricity, compute the percent difference in S between perihelion (January) and aphelion (July).

(b) Compute the percent difference in S between perihelion and aphelion for Mars, whose eccentricity is 9.3%.

3. Relative brightness of snow and sun

(a) Compute the fraction (f) of the sky occupied by the Sun. [Hint: in Problem 1 you have computed the solid angle occupied by the Sun in the sky.]

(b) Consider a horizontal snowfield on the Earth's surface, with albedo A, illuminated by the Sun at zenith angle 60°. Assuming that the snow reflects radiation isotropically, obtain the brightness of the snow I_{snow} relative to that of the Sun I_{sun} (where I_{sun} is the brightness averaged over the solar disk); i.e., obtain the ratio I_{snow}/I_{sun} in terms of f and A.

(c) Evaluate the expression you obtained in part (b), to obtain the relative brightness of sun and snow, assuming that the snow has albedo 1.0 for visible light. Assume that sunglasses will attenuate the incident radiation equally at all wavelengths, by a factor of 5, and that one pair of sunglasses is sufficient for looking at a snowfield. How many sunglasses should you put in front of your eye to look safely at the Sun?

4. Relative densities of molecules and photons

(a) For an overhead sun, the downward flux of solar energy onto the top of the atmosphere is 1370 W m^{-2} (the "solar constant"). Assuming the Earth's albedo is 0.3, and that the Earth is an isotropic ("Lambertian") reflector, compute the number-density of solar photons (photons/m^3) in the upper atmosphere (i.e. above all scattering and absorbing layers), when the Sun is overhead. Use the approximation that all solar photons are at the visible wavelength λ = 500 nm. The upward (reflected) photons are traveling at all zenith angles between 0° and 90°; you can simplify the problem by assuming they all travel at the "average" angle of 60°. Ignore the curvature of the Earth.

(b) Find the height in the atmosphere where the number-density of molecules is the same as the number-density of solar photons. Assume the atmospheric density ρ decreases exponentially with height z, as $\rho(z) = \rho(0) \exp(-z/H)$, with scale-height H=8 km.

ATMS 532. Homework 3. Due Monday 3 February 2014.

1. Transmission of sunlight through ice; color of ice

For this problem you will need the absorption coefficient for pure ice:
β_a=0.0014 m^{-1} at λ=400 nm (blue);
β_a=0.520 m^{-1} at λ=700 nm (red).
And you will need the refractive indices:
N=1.00 for air; *N*=1.31 for ice at both red and blue wavelengths.

In winter the ice on Lake Baikal is thick enough to support a railroad train. [Tracks were laid on the lake for a few winters ~1902-3 before the Trans-Siberian Railway was built around the lake.] Assume the ice is clear, bubble-free, snow-free, and 2 meters thick. Consider the sunlight transmitted through the ice to the water below.

(a) With the Sun at zenith angle 60°, what is the path length of the solar beam through the ice?

(b) What is the ratio of transmittance at λ=400 nm to that at λ=700 nm? If the incident sunlight is white, what will be the apparent color of the transmitted sunlight? [The reflectance *r* at an ice-air interface for zenith angle 60° at visible wavelengths, computed using the Fresnel equations, is *r*=0.05.]

2. Units of Planck function

If the spectral flux from the Sun at wavelength 500 nm is 2000 W m$^{-2}\mu$m^{-1} (Petty's Figure 3.2), then what is it (a) in units of W m^{-2}Hz^{-1}; (b) in units of W m^{-2}(cm^{-1})$^{-1}$?

3. Maxima of the Planck function.

Derive these relations for the maximum of the Planck function (you will obtain a transcendental equation that can be solved graphically):
(a) The maximum of B_λ occurs at λT = 0.290 cm deg.
(b) The maximum of B_ν occurs at $cT/\nu = \lambda T$ = 0.510 cm deg.
(c) The maximum of $dB/d\ln\lambda$ occurs at the intermediate value λT = 0.367 cm K
(d) The maximum of $dB/d\ln\nu$ occurs at the same wavelength as the maximum of $dB/d\ln\lambda$.

For Earth's effective temperature T_e=255K, where is the maximum for each of these three versions of the Planck function? Give all three answers in wavelength (μm).

4. Reflection of *thermal infrared* solar radiation

Petty Problem 6.14 (page 132).

ATMS 532. Homework 4. Due Monday 10 February 2014.

1. Brightness temperature (6 points)
The "brightness temperature" T_B is the temperature a blackbody would require, to emit the observed intensity I_ν. Thus $I_\nu = \varepsilon_\nu B_\nu(T) = B_\nu(T_B)$,
where T is the true thermodynamic temperature of the emitter and ε is the emissivity.
(a) Show that $T_B \approx \varepsilon T$ if $h\nu/kT \ll 1$.
(b) In the microwave "window" used for surface remote-sensing, radiation emitted from the earth's surface is transmitted directly to space with almost no absorption in the atmosphere. A satellite measures intensity at $\lambda = 1.55$ cm, and we infer the sea-surface temperature (SST), assuming no atmospheric absorption. If the true SST is 273 K, how much error (degrees) in inferred SST results from the assumption that $T_B = \varepsilon T$? Assume the surface is liquid water, ε=0.43.

2. Kirchhoff's law (3 points)
At night on a flat desert surface, the ground temperature is T_g=250K and ground emissivity is ε_g=0.9, independent of wavelength in the thermal infrared. A thick stratus cloud moves in to cover the sky. The cloud has infrared emissivity ε_c=1.0; the atmosphere between the cloud and the ground is transparent to infrared radiation. The upward radiation flux measured with a pyrgeometer at 2 meters above the surface is $1.05\ \sigma T_g^4$, where σ is the Stefan-Boltzmann constant. Find the temperature T_c of the cloud base. [Hint: Do not assume radiative equilibrium.]

3. Sensitivity of outgoing longwave radiation to surface temperature (3 points)
Define F_B as the total radiation flux emitted by a blackbody (W m^{-2}).
(a) From the Stefan-Boltzmann law obtain an expression for the derivative dF_B/dT.
(b) For a temperature increase of 1 K, what will be the increase of emitted radiation? Do this for the Earth's average surface temperature (+15°C) and for the temperature at the 6-km level in the atmosphere (-18°C).
(c) When satellite measurements of the outgoing longwave flux at the top of the atmosphere are correlated with surface temperature T_s, the slope is $dF/dT_s \approx 1.83$ W m^{-2}K^{-1} (see figure). Compare this value to your result and speculate on the reason for the difference.

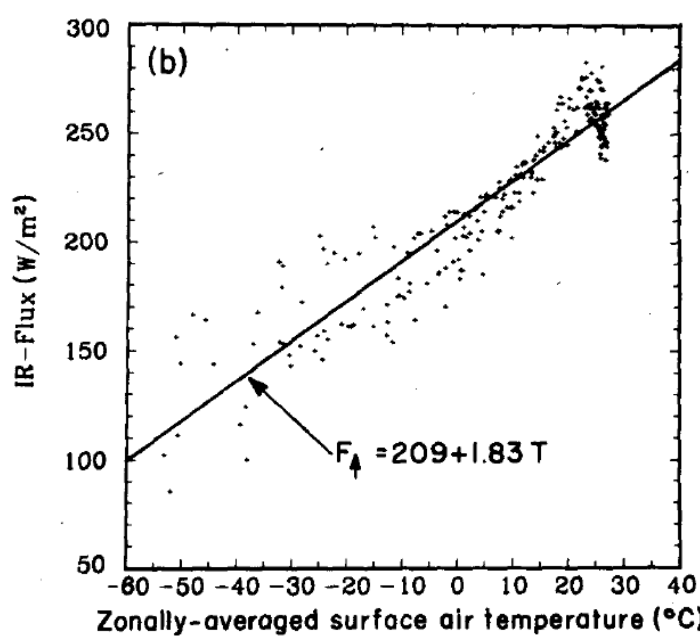


Figure 5b from
Warren and Schneider, 1979:
J. Atmos. Sci. 36, 1388.

ATMS 532. Homework 5. Due Wednesday 19 February 2014.

Three points each, total 15 points.

1. Petty 7.6 (page 173)

2. Petty 7.7 (page 179)

3. Ozone versus oxygen

The result of the previous problem may have disturbed you, if you worry about exposure to UV radiation. So now do the calculation for ozone instead of oxygen. At this same wavelength, 240 nm, the absorption cross-section of ozone is 8.2 x 10^{-22} m^2/molecule. For a total ozone amount of 300 Dobson Units (DU) (i.e., 3 mm at STP*), compute

(a) the number of ozone molecules per m^2 in a column,

(b) the optical thickness at 240 nm, and

(c) the vertical transmittance through the ozone layer at 240 nm.

4. Petty 7.11 (page 200)

5. Petty 7.12 (page 200). (Assume a monodispersion.)

Also compute the radii of the cloud droplets.

*STP = Standard temperature and pressure (273 K, 1 atmosphere)

ATMS 532. Homework 6. Due Monday 24 February 2014.
(3 points each; 9 points total)

1. Photon mean free path
Consider a stream of photons of frequency ν_o entering a semi-infinite* medium that absorbs photons of this frequency but does not scatter them. The photons will travel a variety of distances into the medium before being absorbed. We define the "mean free path" as the average of these distances. Write an expression for the mean free path in units of optical depth. Then evaluate this expression to find the numerical value of the mean free path (in optical depth units). You may need the series expansion:

$e^x = 1 + x + x^2/2! + x^3/3! + \ldots.$

Hint: To find the number of photons whose paths end in the interval $\Delta\tau$ between τ_i and $\tau_i + \Delta\tau$, compute the number of initial photons that reach τ_i and the number that reach $\tau_i + \Delta\tau$.

2. Absorption of solar energy
Consider a thin absorbing layer located above all scatterers. [For example, the result you get will be applicable to absorption of 700-nm radiation by the stratospheric ozone layer, which is located above all tropospheric clouds. And Rayleigh scattering is negligible at 700 nm.] Show that the rate of absorption (W m^{-2}) of solar energy by this layer is independent of the solar zenith angle θ_o if the layer is optically thin ($\tau^* \sec\theta_o << 1$).

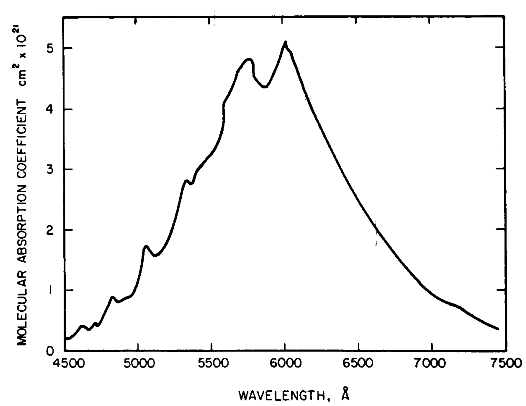


FIG. 5.13. The Chappuis bands of ozone at 291 K. After Vigroux (1953).

(Goody & Yung 1989)

3. Plotting
Plot the survival probability of an emitted photon in an infinite medium, as a function of the number of extinction events experienced, for these values of single-scattering albedo:
(a) $\varpi = 0.9$
(b) $\varpi = 0.999$
(c) $\varpi = 0.99999$
Think carefully about the best way to display these results on the graph.

*a "semi-infinite" medium has a top boundary but is infinitely deep.

ATMS 532. Homework 7. Due Wednesday 5 March 2014.

1. Exponential integrals (2 points)

Show that $\frac{dE_n(\tau)}{d\tau} = -E_{n-1}(\tau)$.

2. Flux-transmittance (4 points)

The intensity-transmittance t_I through an absorbing, non-scattering, plane-parallel layer is

$$t_I(\theta) = \frac{I(\tau^*, \theta)}{I(0,\theta)} = \exp(-\tau^* \sec\theta).$$

Consider a layer illuminated from above by isotropic radiation, so that the flux-transmittance t_F is

$$t_F = \frac{F(\tau^*)}{F(0)} = 2E_3(\tau^*).$$

For what angle θ is the intensity-transmittance equal to the flux-transmittance:

(a) If the layer is optically thin ($\tau^* \to 0$) show that $\theta = 60°$.

(b) If the layer is optically thick ($\tau^* \to \infty$) show that $\theta = 0°$.

[You may want to refer to Abramowitz & Stegun for properties of exponential integrals, in particular $E_n(\tau) \to e^{-\tau}/\tau$ as $\tau \to \infty$.]

3. Radiative equilibrium temperature distribution (3 points)

Using the radiative-equilibrium temperature distribution (from class notes pp 60-61 or from Houghton eqs. 2.12, 2.13), plot surface air temperature T_o, ground temperature T_g, and temperature-slip (T_g - T_o) as functions of the optical thickness τ^* of the atmosphere from $\tau^*=1$ to $\tau^*=90$. Use T_e=255K.

4. Excited state populations (2 points)

CO_2 has three fundamental vibrational modes, designated ν_1, ν_2, ν_3, corresponding to wavenumbers ν_1/c=1388.23 cm^{-1}, ν_2/c=667.40 cm^{-1}, and ν_3/c=2349.16 cm^{-1}. The excited levels of ν_2 have a degeneracy of 2; those of ν_1 and ν_3 have degeneracies of 1. Calculate the ratio of the number-densities of molecules in the lowest excited levels of each of these modes to the number density in the ground state, at T=200K and T=300K, assuming LTE.

continued over

5. e-type absorption (4 points)

(a) Use Goody & Yung's Figure 5.8 (it's in the handout notes, p. 75) for the water-vapor continuum to estimate the "binary absorption coefficient" k_n in the "atmospheric window" at λ=11 μm. This k_n is related to the linear absorption coefficient β_a as $\beta_a = n_{H_2O}^2 k_n$. If β_a has units cm^{-1}, and n_{H_2O} has units cm^{-3}, then k_n has units cm^5. You will also need Loschmidt's number n_s, the number of molecules per cm^3 at STP: n_s=2.687×10^{19} cm^{-3}.

(b) Assume the water vapor is uniformly mixed in the boundary layer (lowest 2 km of the troposphere), and that *all* the water vapor is confined to this 2-km layer. [You may neglect the decrease of density with height within this 2-km layer; the resulting error is only 2%.] Assume a mean temperature of 280 K. Estimate the vertical optical depth of the atmosphere at λ=11 μm due to e-type continuum absorption, for precipitable water amounts of 0.5 (winter), 2.5 (summer), and 5 g cm^{-2} (tropical). Also compute the corresponding transmittance (percent) between surface and top of atmosphere at λ=11 μm, using a diffusivity factor of 1.66.

6. Q-branch (3 points)

In typical spectra of outgoing longwave radiation at the top of the atmosphere (e.g. Petty's Figures 8.3a,c,d), there is an upward spike in the center of the CO_2 15-μm band and also in the center of the ozone 9.6-μm band. Explain how these upward spikes are caused by the *presence* of a Q-branch for CO_2 but the *absence* of a Q-branch for ozone.

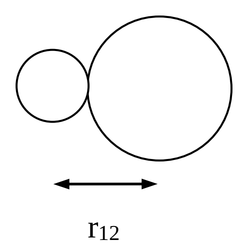
ATMS 532. Homework 8. Due Wednesday 12 March 2014.

Line shapes.

1. Calculate the mode velocity (the peak of the velocity distribution) at temperature 273 K
(a) for nitrogen molecules
(b) for ^{132}Xe.

2. Calculate the atmospheric pressure level (in millibars) at which the Lorentz half-width α_L is equal to the Doppler half-width α_D, for the following lines. Do it for two temperatures, 200 K and 300 K.
(a) CO_2 at 667 cm^{-1}.
(b) H_2O at 1600 cm^{-1}.
(c) H_2O at 100 cm^{-1}.
Assume for each case that α_L=0.1 cm^{-1} at STP conditions.
(STP means T=0°C and pressure = 1 atmosphere.)

3. (pp. 108-109 in notes).
Calculate the effective collision radius r_{12} which yields α_L(STP)=0.1 cm^{-1}, for the foreign broadening of an H_2O line by collisions with N_2.



4. (p. 97, p. 105 in notes).
Compare the ratio of the absorption coefficients at the line centers to the values 3 half-widths from line center for the Doppler and Lorentz shapes. Note the very slow falloff of Lorentz shape in the wings. As a consequence, radiative transfer involving line *wings* will be dominated by the Lorentz shape to very high elevations.

ATMS 532. Longwave Radiative Transfer Project, Winter 2014

Stephen Warren and Ryan Eastman

General Help: If you do not have an account on the Atmospheric Sciences department computer system, then contact Harry Edmon (room 418) or David Warren (room 425C) at support@atmos.washington.edu. The programs provided here can be executed on the Linux platforms in the 6th floor computer lab. There is a version of Streamer that works on OSx if you have a Mac.

To complete this assignment you will be required to be competent in:
Linux or OSx.
Matlab.

See me (Ryan Eastman) if you would like some references for either operating system or Matlab. The assignment is located in a file called radTransPro532.tar on the department ftp server. It contains the executable files for running both LBLRTM and Streamer in separate directories. There are example input data files for Streamer and actual input data for LBLRTM.

To access the files, ftp to ftp.atmos.washington.edu anonymously and go to the *rmeast/atm532* directory. Grab the files and put them in a directory in your Linux account where you would like to work on this assignment. Untar them with the command 'tar -xvf radTransPro532.tar'. Alternatively, if you are on one of the department computers you may grab the files by changing directories into /home/disk/ftp/rmeast/atm532/.

LBLRTM help:

We provide some M-files that will run LBLRTM. They take the Matlab input and turn it into LBLRTM input (the TAPE5 file). We do it this way because it is easy to mess up the formatting of the TAPE5 file, so it is better to do this automatically each time.

Steps to run LBLRTM:
1. Load your atmospheric profile into Matlab. Profiles are stored as .mat files or as text files and are located in the LBLRTM/data directory.
2. Execute the lblrtmRun.m script using a statement similar to:
   [nu,rad] = lblrtmRun(P,T,rh,o3gm3,alt,theta,v1,v2,res,TOA,Tsfc);

Make sure that all the other M-files and LBLRTM files provided are still in the same directories as when the tar bundle was unpacked.

Some other things to note:
1. Ignore the final segmentation fault on some of your runs. It occurs when one or both of the wavenumber intervals are not four digits, but it does not affect the results.

2. You will not be able to run LBLRTM from 0 to 2500 cm^{-1} in one statement. You must break the spectrum up into small pieces and put it together later. The maximum continuous spectrum allowed by LBLRTM is ~2000 cm^{-1}.
3. LBLRTM does not deal well with large zenith angles. Avoid $\mu < 0.2$.
4. For some of the problems you will be required to edit lblrtmRun.m. I recommend 'packaging' lblrtmrun.m into a function.
5. When radiances at the TOA are desired, the zenith angle (theta) must be between 90° and 180° (horizontal-vertical); for radiances at the surface theta must be between 0° and 90°.
6. LBLRTM produces radiances in units of mW m^{-2} sr^{-1} $(cm^{-1})^{-1}$.

Streamer help:

Your experience with Streamer will be more hands-on than with LBLRTM. You must edit the input scripts yourself in a text editor such as Emacs or Vi. The Streamer executable is not nearly as particular about the horizontal spacing of the input data as LBLRTM. However, it can be somewhat particular about the number of lines in an input file. Some lines that are blank should be left in the input file because Streamer will look for them even if it knows that there should be no input there.

It will be necessary for you to become familiar with parts of the User Manual, particularly pages 14-28, where the authors describe the input file for Streamer. The User Manual is very helpful for this application. The manual is located in the STREAMER/docs directory, and is called userman.pdf. I recommend starting with their test input file, then altering that file to suit the problem.

Steps to run Streamer:
1. Edit the input file using Emacs, Vi, or some other editor of your choice. (Text editor is fine)
2. Execute the Streamer application using the command: "./streamer_linux input.inp" (you don't have to name your input file "input.inp"; it can be whatever you like).
3. Open the output file, the name of which is specified in the input file, using a text editor. If you are adept at Matlab you can write a script to do all of this from Matlab and will then eliminate the step of entering the data yourself into Matlab for analysis.

Contact me if you have any questions!

Contact Info:
Ryan Eastman
rmeast@atmos.washington.edu
Room ATG 719

Project due date:
To Steve's mailbox in ATG 410 by 5 pm on Friday 14 March
or
by email attachment to sgw@uw.edu by 10 am on Friday 21 March.

1. Baseline Case (use LBLRTM)
Use the Mid-Latitude Summer (MLS) standard atmosphere.

a. Top of Atmosphere (TOA)

i. Compute the outgoing intensity at TOA using 20 cm^{-1} resolution, at a viewing angle of 60°.

ii. Approximate the flux as $F = \pi I$, and plot the flux versus wavenumber.

iii. Include Planck blackbody curves at 200 K, 250 K, and 300 K in the plot.

iv. Integrate the flux over wavenumber to obtain a table of partial fluxes for these seven spectral regions:

1.	water-vapor rotation band	0-560 cm^{-1}
2.	CO_2 15-μm band	560-780 cm^{-1}
3.	11-μm window	780-980 cm^{-1}
4.	ozone 9.6-μm band	980-1080 cm^{-1}
5.	9-μm window	1080-1240 cm^{-1}
6.	water-vapor 6.3-μm band	1240-2500 cm^{-1}
7.	total longwave	0-2500 cm^{-1}

b. Surface

For the same atmosphere, compute the downward intensity at the surface and obtain the partial fluxes as above. Plot flux versus wavenumber as before, including blackbody curves at 200 K, 250 K, and 300 K.

c. Interpretation. Explain the variations in flux from one spectral region to another, with the blackbody curves for reference.

Note: Be careful not to double-count when computing the total longwave flux if your spectral regions overlap.

2. Contributions to OLR from surface and atmosphere (LBLRTM)

In this exercise you will examine the contribution of the ground to the outgoing flux at the TOA. The TOA outgoing longwave flux is the sum of the upward fluxes from the surface and from the atmosphere. Compute the outgoing flux at the TOA for three atmospheres: tropical (TRP), midlatitude summer (MLS), and subarctic winter (SAW). By turning off the surface contribution, obtain the fractional contributions from the atmosphere and the surface separately. As in Problem 1, determine the spectral fluxes, but only in the following spectral bands.

(a) the 6.3-μm water-vapor band (1240-2500 cm^{-1})

(b) the 9-μm window (1080-1240 cm^{-1})

(c) the 15-μm CO_2 band (560-780 cm^{-1})

Why is the surface percentage contribution greater for the SAW than for the TRP atmosphere in each band? (Hint: examine the standard atmosphere profiles to explain your results.)

3. Angular variations of intensity (LBLRTM)

Compute the upward radiance at the TOA for MLS at 1 cm^{-1} resolution at viewing angles of μ = 0.2, 0.3, . . . , 0.9, 1.0.   Note: Do not use a μ less than 0.2 because LBLRTM doesn't deal well with large zenith angles.

(a) Plot *I* versus μ for three different frequencies: 667,725, and 900 cm^{-1}.

(b) Give an explanation for the wavenumber dependence of *I* versus μ.

(c) Compute the flux at each of these three frequencies by integrating the radiance over angle and compare it to the flux estimate obtained by simply approximating $F = \pi I(60°)$.

Since intensity was not computed at μ=0.1, in the integration you can extrapolate down from μ=0.2 to μ=0 using the trend from μ=0.3 to 0.2.

4. Errors due to coarse vertical resolution (LBLRTM)

(a) Compute the downward longwave flux ($F = \pi I(60°)$) at the surface in the Antarctic winter using the following two atmospheric profiles: winterAnt.mat and winterCoarseAnt.mat. These are winter profiles, one with fine vertical resolution of the near-surface temperature inversion and another with coarse resolution.

(b) Plot the spectral flux derived from both model atmospheres, as well as the difference in spectral flux between the two cases. Also, plot the vertical profiles of temperature and water vapor. Integrate the spectral fluxes to get the total flux and compare the two cases.

(c) In which of the spectral regions do you see the largest differences? Why?

(d) Double the CO_2 concentration in the model (specified in lblrtmRun.m). Use the winterAnt.mat atmospheric profile and compute the spectral flux again. Plot the difference in spectral flux: $F(2\times CO_2) - F(1\times CO_2)$. How is the flux in the spectral region between 560 and 780 cm^{-1} changed by doubling the concentration of carbon dioxide? How does this compare with the effect of coarse vertical resolution? Explain the positive and negative regions of the difference-spectrum.

5. The Greenhouse Effect (Streamer)

The OLR at TOA is reduced by the absorption of radiation by CO_2 ,water vapor, and other atmospheric constituents. Increased concentrations of some of these constituents will lead to increased absorption of longwave radiation. We define the *greenhouse effect* G as the difference between the upward LW flux at the surface and the outgoing LW flux at TOA. For example, if the surface emits 390 W m^{-2} and OLR = 240 W m^{-2}, then G = 150 W m^{-2}.

In this exercise, we make estimates of how much is contributed to the clear-sky greenhouse effect by water vapor, CO_2, and ozone. Streamer allows you easily to alter the concentrations of these gases in the model atmospheres using scale factors. Set the surface emissivity to 1, and the surface temperature to 294 K.

(a) What is the OLR if the "only" absorbing gas is water vapor? [To calculate this, set CO_2 and ozone amounts to zero, but the atmosphere will still contain some other minor greenhouse gases.]

(b) What is the OLR if the only absorbing gas is CO_2?

(c) What is the OLR if the only absorbing gas is ozone?

(d) What is the OLR if CO_2, ozone, and H_2O are all removed?

(e) What is the OLR if all gases are present at their standard amounts?

(f) Make a table of the separate greenhouse effects of water vapor, CO_2, and ozone in W m^{-2}. Then add to the table the *percent* contributions of each of these three gases to the total greenhouse effect. Note that because of overlap the three separate greenhouse effects will add to more than the total, so they need to be scaled to make their total equal 100%.

6. Cloud Radiative Effects - low vs. high clouds (Streamer).

Cloud radiative forcing (CRF) is usually defined as the difference at some level (e.g. surface or TOA) between (1) the monthly-average or seasonal-average net flux (including all scenes both clear and cloudy) and (2) the clear-sky net flux. However, in Streamer the CRF is instead defined as the difference in flux between a specified scene (which may be partly cloudy) and a clear scene.

(a) Using the MLS profile and the following specifications, compute CRF for the window region (780-1240 cm^{-1}) at TOA and at the surface for a water cloud (droplet effective radius $r = 10\ \mu m$), with cloud base 500 m above the surface and cloud thickness 100 m.

(b) Repeat (a) setting cloud base 5 km above the surface, but retaining all other values.

(c) Explain the differences in CRF you observe.

7. Cloud Radiative Effects - latitude and season.

(a) Perform the same computation as in 6a for the Arctic winter atmosphere.
(b) Explain any differences you see in CRF between 6a and 7a.

8. Cloud Radiative Effects - particle size.

(a) Perform the same computation as in 6a using an effective cloud droplet radius of 5 μm.
(b) Repeat using an effective radius of 50 μm.
(c) Explain the differences between 6a, 8a, and 8b.

9. Cloud Radiative Effects - liquid water path.

(a) Perform the same computation as in 6a but with a cloud thickness of 10 m.

(b) Perform the same computation as in 6a but with a cloud thickness of 1000 m, with the cloud base still at 500 m.

(c) Explain any differences you observe in CRF between 6a, 9a, and 9b.