ATOC/ASTR 5560 Lab 5 Solutions
September 28, 2001

The purpose of this lab is to run a broadband longwave radiative transfer model and gain an understanding of longwave radiative processes that control spectral radiances and heating rates. You will use MDTERP, which is a narrowband longwave radiative transfer model with a graphical user interface written in IDL. MDTERP has three phases of operation: 1) it sets up one or more atmospheric profiles of temperature and gas concentrations according to user input, 2) it runs the Fortran narrowband radiative transfer model, and 3) it plots the radiative transfer results using the interactive graphical interface.

1. *First practice running MDTERP.*

2. *a) Plot the upwelling radiance spectrum (at 0°) for the midlatitude summer atmosphere. Explain the differences between the upwelling radiance spectra at 0 km, 5 km, and the top of the atmosphere. Refer to Planck function curves at the temperature of the surface and current level when interpretting the radiance values. Explain in terms of the absorber distribution, spectral bands of the absorbers, and the physics of thermal radiative transfer.*

   The upwelling radiance at the surface is simply a Planck function at 294 K because the surface is assumed to be a blackbody. At 5 km the upwelling radiance is still on the 294 K Planck function in the 800 to 1200 cm^{-1}window region because there is little absorption and hence high transmission. Below 400 cm^{-1}(H_2O rotational band), around 600 to 700 cm^{-1}(CO_2 band), and 1500 to 1800 cm^{-1}(H_2O vibrational band) the radiance is on the 267 K Planck function (temperature at 5 km) because there is so much absorption that the emission originates from close to the level. In the adjacent wavenumber regions the radiance is between the two Planck functions because some of the contributing radiation originates at lower altitudes or from the surface.

   The top of the atmosphere radiance shows substantial depressions from the surface radiance below 600 cm^{-1}from the H_2O rotational band, from the CO_2 band around 667 cm^{-1}, from the ozone band around 1000 cm^{-1}, and from 1200 to 2000 cm^{-1}from the water vapor vibrational band. The radiance decrease is greater for the CO_2 and O_2 bands because their weighting functions peak in the very cold stratosphere. The radiance decrease in the water vapor bands is less because the weighting functions peak in the middle to upper troposphere because nearly all the water vapor is in the low to mid troposphere.

   b) Plot the downwelling radiance spectrum for the midlatitude summer atmosphere. Explain the differences across the spectrum between the downwelling radiances at 20, 5, and 0 km.

   The downwelling radiance at 20 km is near zero except for the 15 μm CO_2 band and the 9.6 μm ozone band which have small radiance values due to the cold temperatures and because the emissivity is less than one due to the low absorber abundance above 20 km.

At 5 km the downwelling radiance reaches the 267 K Planck function below 400 cm^{-1}(water vapor), from 600 to 700 cm^{-1}(CO_2), and from 1500 to 1900 cm^{-1}(water vapor). This is due to the strong absorption causing the emission to originate close to 5 km. The radiance at the ozone band remains low because the ozone emission is from the stratosphere at cold temperatures and because the absorptivity of the ozone band is not 100%. The rest of the 8 to 12 μm window has radiance near zero because there is not much water vapor above 5 km.

At 0 km the downwelling radiance lies on the 294 K Planck curve except for the 8 to 12 μm window region. The absorption, mainly due to water vapor near the surface, is so strong that the transmission falls off very rapidly with height, so that the emission originates in the warm atmosphere near the surface. Even in the 8 to 12 μm window there is substantial downwelling radiance due to the water vapor continuum absorption.

3. *a) Plot the spectrally integrated fluxes and heating rate for the midlatitude summer atmosphere. Plot as a function of altitude and use 49 km for the upper level so only the troposphere and stratosphere are plotted. For a different perspective, change the y-axis of the plots to pressure.*

   b) Why does the downwelling flux have a change in slope around 200 mb?
   Why is the net flux slope and heating rate fairly constant from 300 to 1000 mb even though the water vapor density increases by two orders of magnitude?

   The downwelling flux increases much more quickly below about 12 km due to water vapor emission, since the water vapor concentration is much greater in the troposphere. Water vapor absorption occurs in a much larger fraction of the spectrum than does CO_2 and O_3, which are important in the stratosphere.

   The net flux slope and cooling rate are nearly constant below 12 km because there is a very wide range of water vapor absorption coefficients across the spectrum (especially in the pure rotational band). Thus for a particular altitude, and hence water vapor amount u above, there is a significant part of the spectrum for which the cooling to space term is large, i.e. where the transmission is intermediate. This is apparent in the spectral heating rate profile contour in question 4.

   c) Save the text output and report the upwelling and downwelling fluxes at the surface, tropopause (13 km), and top of atmosphere. Calculate the integrated net flux divergence (ΔF_{net} in W/m^2) in the troposphere and above. Note that the total atmospheric divergence does not equal the flux emitted at the top of the atmosphere – what else is cooling, i.e. what makes the longwave flux balance?

Pressure (mb)	Altitude (km)	Fup (W/m^2)	Fdown (W/m^2)	Fnet (W/m^2)
1013	0	423.53	348.21	75.32
179	13	284.81	23.00	261.82
0	103	278.87	0.00	278.87

The integrated net flux divergence is 17.1 W/m^2 in the stratosphere and 186.5 W/m^2 in the troposphere. The flux divergence in the stratosphere is relatively small because of the limited spectral range of the absorbers there (CO_2 and O_3). The total atmospheric flux divergence is thus 203.6 W/m^2, which is less than the outgoing top of atmosphere flux of 278.9 W/m^2. This is because the surface is also cooling, and loses a net flux of 75.3 W/m^2, which makes the longwave flux budget balance.

d) For what layer does the cooling rate peak? Explain, considering the results in c).

The cooling rate peaks in the upper stratosphere layer from 45 to 50 km due to emission by CO_2 and O_3 from the warm stratopause region (> 270K). The net flux divergence from this layer is small, but the density is about 1/1000 of the surface density, so the cooling rate is large.

4. *Look at the heating rate in more detail for midlatitude summer by making a Clough spectral heating rate profile plot (use an upper altitude of 49 km). Explain the variation in height of cooling across the spectrum by major absorption band. Consider the distributions of radiatively absorbing gases and the physics of longwave heating rate. Explain the major locations (wavenumber,height) of heating.*

   The major cooling region in the troposphere from 100 to 900 cm^{-1}is due to water vapor. As the water vapor absorption line strength decreases with wavenumber in this pure rotational band it takes more water vapor to reach transmission values around 50%, and hence the peak of the weighting function and thus the peak in the cooling to space term drops from the upper troposphere to near the surface. The altitude of water vapor emission (cooling) increases around the 1600 cm^{-1}vibrational band center, but the Planck function is small here so the cooling is less intense. There is also cooling due to water vapor in the upper stratosphere (below 500 cm^{-1}and around 1600 cm^{-1}) due to emission from the centers of very strong lines at the warm stratopause temperature.

   There are very high cooling rates in the upper stratosphere around 700 cm^{-1}from CO_2 and around 1000 cm^{-1}from ozone. The cooling rate is less in the lower stratosphere because the transmission to space, and hence the cooling to space term, is lower. Also the stratopause region is warm, so there is more emission to space. There is also significant cooling due to the 4.3 μm CO_2 band around 2400 cm^{-1}.

   The major region of heating is in the lower stratosphere around 1000 cm^{-1}. This is due flux exchange between the warm surface and the cold lower stratosphere in the ozone band. There is little ozone in the troposphere so there is high transmission between the lower stratosphere ozone and the boundary layer water vapor and surface. The the upwelling flux from the warm surface is larger than the downwelling flux from the cold stratosphere, so the flux exchange heats the ozone.

   There is a small region of heating in the 15 μm CO_2 band at 13 km, which is due to flux exchange with adjacent warmer layers.

5. *Now we'll look at the effect of single level temperature and water vapor perturbations on the heating rate profile. Go back to "Input data" and choose three profiles. Use an unmodified midlatitude summer for profile 1.*
For profile 2, modify the temperature (select T button under Modify). Add 5 K to the 7 km level by choosing "X=X+k", set k to 5, select "Choose levels", select "Individual", click on level 8, then "Done".
*For profile 3, modify the water vapor by doubling the value at 7 km. Use a similar procedure as for modifying temperature, but choose "X=k*X" and set k to 2.*
Run the longwave radiative transfer ("Compute results").
Go to the plot section and "Save text output", which saves results for all three soundings.

a) Calculate the change in outgoing longwave flux at the surface and top of atmosphere for the temperature and water vapor perturbations. Briefly explain the causes of the TOA flux changes. Why does the downwelling flux at the surface barely change?

The flux changes at the surface are:
Temperature: $\Delta F^{\downarrow}(0) = 348.24 - 348.21 = 0.03\ \mathrm{W/m^2}$
Water vapor: $\Delta F^{\downarrow}(0) = 348.27 - 348.21 = 0.06\ \mathrm{W/m^2}$

The flux changes at the top of atmosphere are:
Temperature: $\Delta F^{\downarrow}(\infty) = 279.80 - 278.87 = 0.93\ \mathrm{W/m^2}$
Water vapor: $\Delta F^{\downarrow}(\infty) = 277.58 - 278.87 = -1.29\ \mathrm{W/m^2}$

The temperature perturbation increases in TOA flux due to the larger Planck function at 7 km. Of course, only part of the increased emission at 7 km contributes to the flux change at TOA, because much of the spectrum has zero transmission from 7 km to space.

The water vapor perturbation decreases the TOA flux because the transmission from the surface and lower atmosphere is decreased by the larger water vapor absorption. Thus there is less radiance contribution from the warm surface and more from the cold 7 km layer.

The downwelling flux at the surface does not change because nearly all of the increased emission at 7 km from the perturbations is absorbed before it reaches the surface. The increased emission only occurs at wavenumbers where there is significant H_2O and CO_2 absorption, and radiation at these wavenumbers is heavily absorbed by the large absorber amounts in the lower atmosphere.

b) Plot the broadband heating rate profile of the difference between the temperature perturbed and the original sounding (plot "Sounding 2-1"). Select a altitude range from 0 to 16 km. Explain the change in the longwave heating rate profile using the flux exchange concept. It may be helpful to make a Clough spectral heating rate profile plot of the difference.

The longwave cooling rate change shows an increase in the cooling rate in the layer and a decrease in adjacent layers. The layer cooling rate increases because it emits more at the higher temperature, while it absorbs the same amount from other layers. The nearby layers absorb some of the extra emitted flux and thereby have less cooling than before. This is the flux exchange between layers concept. There has to be a temperature difference between layers, significant absorptivity/emissivity of the layers, and transmission between the layers. The transmissivity requirement often limits the flux exchange to nearby layers.

c) Plot the broadband heating rate profile of the difference between the water vapor perturbed and the original sounding (plot "Sounding 3-1"). Select a altitude range from 0 to 16 km. Explain the change in the longwave heating rate profile.

The longwave cooling rate is increased in the layer above 7 km as the emissivity is increased by the extra water vapor. The emitted flux is readily transmitted to space due to the dry atmosphere above. The cooling rate is decreased in the layer below 7 km as more upwelling flux from below is absorbed which counteracts the increase in emitted flux. The extra emitted downwelling flux is absorbed in the layers below the perturbation, which decreases the cooling rate there.

6. *Now look at the effect of doubling carbon dioxide. Go back to "Input data" and choose two profiles. Use midlatitude summer for the first one. Double the CO_2 concentration in the second profile by selecting the "2*X" button (make sure the mass mixing ratio is correct). Run the longwave radiative transfer ("Compute results").*
   Go to the plot section and save the text output for the two soundings.

   a) Calculate the change in outgoing longwave fluxes at the surface and top of atmosphere from doubling CO_2. Briefly explain the outgoing flux changes.

   The flux change at the surface is:
   2×CO_2: $\Delta F^{\downarrow}(0) = 350.00 - 348.21 = 1.79 \text{ W/m}^2$

   The flux change at the top of atmosphere is:
   2×CO_2: $\Delta F^{\downarrow}(\infty) = 276.13 - 278.87 = -2.74 \text{ W/m}^2$

   The increase in CO_2 absorption raises the altitude of outgoing flux emission to where the atmosphere is colder which decreases the upwelling TOA flux. The downwelling flux at the surface is increased because the increasd CO_2 absorption lowers the emission level to warmer temperatures.

   b) Plot the broadband heating rate profile of the difference between the doubled CO_2 and original sounding (plot "Sounding 2-1"). Select a altitude range from 0 to 49 km. Explain the change in the longwave heating rate profile.

   The most noticeable change is the increase in cooling rate in the upper stratosphere. This is due to increased emission to space by CO_2. Thus the largest temperature change in the atmosphere from doubling carbon dioxide is a decrease in the upper stratosphere.