Lab Activity No. 3: Atmospheric Boundary Layer and Radiative forcing calculations

Crizzia Mielle de Castro

1 General and Specific Objectives

1.1 General Objective

- To determine the effects of air pollutants and low-level clouds on radiative forcing simulations

1.2 Specific Objectives

- To calculate the cloud base height and cloud top height from Skew-T diagrams of measurements in synoptic weather stations
- To determine the effects of differences in cloud base height and cloud top height on pollutant concentrations
- To model and simulate radiative forcing given initialized parameters from the sampling site and weather station measurements
- To observe whether aerosols in the atmosphere cause a warming or cooling effect

2 Description of Site/Sampling Area

Just like the previous activity, we focus our measurements of cloud base height and cloud top height in the Tanay, Rizal synoptic weather station ($14.56^o N$,$121.36^o E$; station no. 98433). On the other hand, our air pollution sampling sites are situated in the Manila Observatory (urban background station), Katipunan Ave. (roadside), and Taft Ave. (roadside) [1]. All the pollution sampling sites are located in Quezon City, Metro Manila. In this activity, I focused on the elevated sampling site in the Manila Observatory ($14.636^o N$,$121.078^o E$; 70 masl).

The Tanay weather station is situated in a rural mountainous area just outside of Metro Manila, about 50km from Katipunan Ave., and the Manila Observatory. The air pollution sampling sites were located in highly urban highways frequented by vehicles such as trucks, private cars, and public transportation.

3 Review of Related Literature

To determine the net radiative forcing in a specific location, we need to obtain the location's net radiant energy [2]. The net energy is given by the difference between the total solar energy absorbed by the Earth, and the total energy Earth radiates back to space. Positive radiative forcing may induce surface warming, while negative radiative forcing may induce surface cooling. Changes in solar energy, surface albedo, and concentrations of greenhouses gases and aerosols lead to varied radiative forcings. We call greenhouse gases and aerosols as radiative forcing agents. We will see in the Radiative Forcing Model discussed in the Methods that other parameters such as location, time, and cloud height also affect radiative forcing. Long terms rise or fall of the Earth's temperature due to imbalanced radiative equilibrium is called climate forcing—something that can last for decades [3].

In this activity, we focus our discussion on aerosol radiative forcing. Besides aerosol optical properties such as optical depth, single-scattering albedo, and the Angstrom extinction exponent, surface albedo and low-level clouds also affect aerosol radiative forcing [4]. Higher albedo or surface reflectivity increases the amount of energy absorbed by aerosols in the atmosphere. Aerosols can absorb both solar energy and energy reflected upward by the Earth's surface.

Low-level clouds—usually the bright fluffy clouds—effectively reflects solar radiation upward. Thus, aerosols found above low-level clouds absorb more radiation, thereby enhancing aerosol radiative forcing.

Aerosols may also serve as cloud seeds or cloud condensation nuclei [5]. Aerosol pollutants produce brighter clouds composed of smaller water droplets, while natural aerosols such as sulfates and sea spray produce grayish clouds composed of larger water droplets. Brighter clouds reflect more solar energy upward, thereby producing a cooling effect on Earth's surface. Brighter clouds can also reflect more energy emitted by the Earth back to the surface, thereby producing a warming effect. These two effects do not necessarily balanced each other out. Net cooling can typically occur during the day when the sun is out, while net warming can typically occur during cloudy nights.

However, not all aerosol pollutants produce longer-lasting bright clouds. Nonreflective aerosol pollutants such as black carbon formed from soot produce the opposite, thereby causing a net warming effect. The net warming effect of black carbon is known as the "semi-direct effect." Previous studies show that the net warming effect of black carbon produces a smoky haze in the atmosphere that suppresses low cloud formation and dissipates precipitation [6–10]. Since the air pollution sampling site measures black carbon concentrations, we expect the synoptic weather station to detect less low-level clouds, and the radiative forcing model to simulate positive radiative forcing (net warming).

4 Methods Framework

The radiosonde data for the Tanay weather station (station no. 98433) was obtained from the following link: http://weather.uwyo.edu/upperair/sounding.html. The data was summarized in Skew-T diagrams obtained from the same link. From the Skew-T diagrams, I obtained the Lifting Condensation Level (LCL) and the Convective Condensing Level (CCL). These two correspond to the pressure levels where the cloud base forms, and the cloud top ends, respectively.

To determine the LCL, I followed the following procedures illustrated in Fig. 1.

1. From the dew point temperature at the lowest surface level, I drew a line upward parallel to the saturation mixing ratio lines.

2. From the temperature at the lowest surface level, I drew a line upward parallel to the dry adiabat lines. The pressure where these two lines intersect denotes the LCL.

To determine the CCL, I followed the following procedure illustrated as well in Fig. 1.

1. From the dew point temperature at the lowest surface level, I drew a line upward

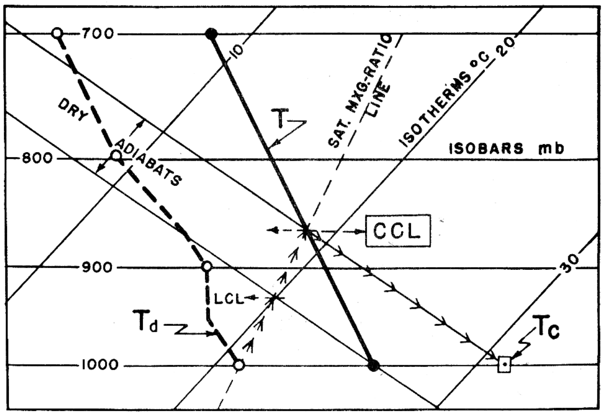


Figure 1: Procedures for determining Lifting Condensation Level (LCL) and the Convective Condensing Level (CCL) from a Skew-T diagram

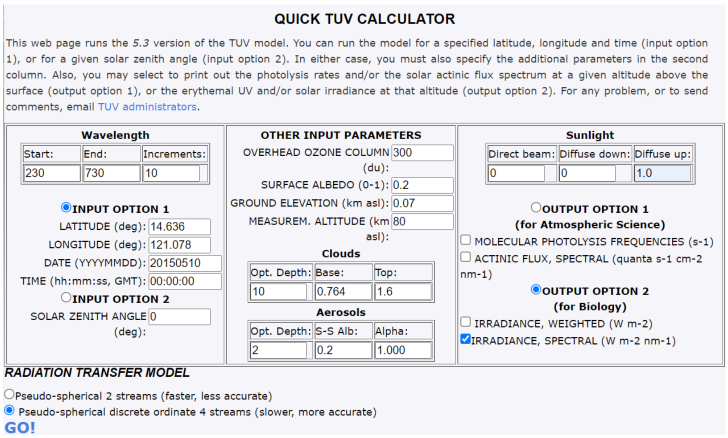


Figure 2: Radiative Forcing Model GUI from https://www.acom.ucar.edu with set parameters.

parallel to the saturation mixing ratio lines until it intersects the temperature curve. The pressure level of the intersection denotes the CCL.

To obtain the aerosol radiative forcing in the Manila Observatory station, I used the following link: https://www.acom.ucar.edu/Models/TUV/Interactive_TUV/. I used a fixed set of parameters (shown in Fig. 2), while only varying the date and time from May 10 00:00 GMT to May 16 12:00 GMT. I also varied the cloud base height and cloud top height from my calculations of LCL and CCL, respectively. On days with no cloud cover (LCL = CCL), I set the cloud base and cloud top to zero. Lastly, I set the surface albedo as 0.2, since the Manila Observatory has a grassy field around it. I varied the aerosol optical depth between 2 and 0 to compare the radiative forcing with and without aerosols, respectively. From the output of the model (shown in Fig. 3), I calculated the radiative forcing using the following equation

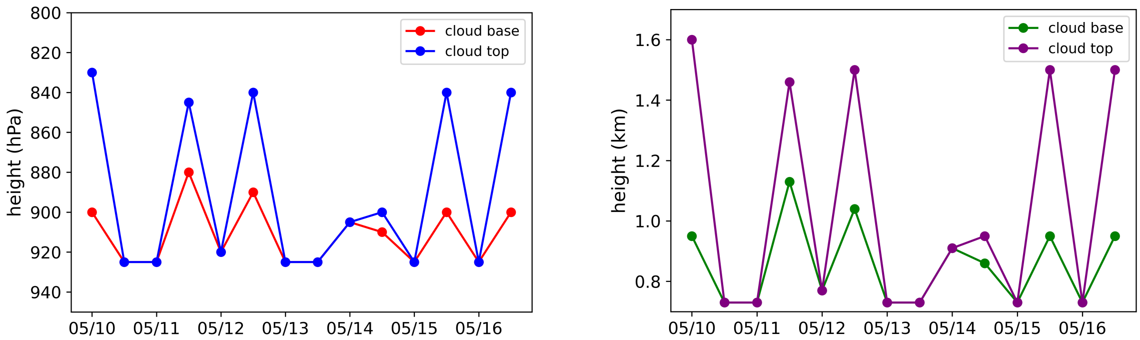
$$\text{radiative forcing} = \text{DIRECT} + \text{DIFFUSE DOWN} - \text{DIFFUSE UP} \tag{1}$$

5 Results and Discussion

The daily cloud base height and cloud top heights show that clouds tend to form more often during the evenings than the mornings (shown in Fig. 4). According to Alas et al., the black carbon concentrations increase in the Manila Observatory during the evening, SPECTRAL IRRADIANCE (W m-2 nm-1):

LOWER WVL	UPPER WVL	DIRECT	DIFFUSE DOWN	DIFFUSE UP	TOTAL DOWNWELLING
230.00	280.00	6.875E-02	0.000E+00	1.137E-04	6.875E-02
280.00	330.00	3.389E-01	3.578E-08	1.160E-03	3.389E-01
330.00	380.00	5.714E-01	-6.033E-08	3.818E-01	5.714E-01
380.00	430.00	8.165E-01	-4.311E-07	5.360E-01	8.165E-01
430.00	480.00	1.097E+00	-1.158E-07	7.004E-01	1.097E+00
480.00	530.00	1.063E+00	-1.122E-07	6.466E-01	1.063E+00
530.00	580.00	1.034E+00	0.000E+00	5.802E-01	1.034E+00
580.00	630.00	9.658E-01	0.000E+00	5.266E-01	9.658E-01
630.00	680.00	8.721E-01	0.000E+00	5.071E-01	8.721E-01
680.00	730.00	7.930E-01	-8.373E-08	4.822E-01	7.930E-01

Figure 3: Radiative Forcing Model sample output of spectral irradiance.



(a) (b)

Figure 4: Daily cloud base height (LCL) and cloud top height (CCL) in (a) hPa and (b) km from May 10, 8:00 LST to May 16, 20:00 LST obtained from Skew-T diagrams.

then decrease again during the mornings. Possibly the black carbon build-up during the evenings inhibits the formation of clouds during the mornings, then as the black carbon concentrations decrease during the day, clouds begin to form in the evening.

The aerosol radiative forcing obtained from the radiative forcing model varies depending on the presence of low-level cloud cover and aerosols (shown Fig. 5). Based on the synoptic weather measurements in the morning, only May 10 showed any low-level cloud cover. The other days of the week more or less follow the same radiative forcing pattern. From Section 3, low-level clouds can reflect more solar radiation, possibly cooling the Earth's surface. Thus, we expect the morning of May 10 to be cooler than the other days of the week.

The presence of aerosols increases the net radiative forcing (solid vs broken lines in Fig. 5, and also almost cancels out the cooling effects of low-level clouds. The difference between the radiative forcings with and without low-level clouds is much smaller in the presence of aerosols than the absence. We observe in order of increasing radiative forcing the following: no aerosols & w/ low-level clouds, no aerosols & cloudless, w/ aerosols & w/ low-level clouds, and w/ aerosols & cloudless. Thus, we expect days with higher black carbon concentrations and cloudless skies to warm the surface to higher temperatures. Since black carbon can inhibit the formation of low-level clouds, higher concentrations promote more days with warmer weather conditions.

We also observe that the difference among the four cases becomes more distinct within the visible light range. Cloudy days with no aerosols reflect more visible light, which we can observe in real life. Days with high amounts of low-level fluffy cloud cover are less bright and cooler than cloudless days.

The net radiative forcing peaks around the yellow wavelengths, with the lowest values in the UV range, which more or less follows the spectrum of solar radiation within the

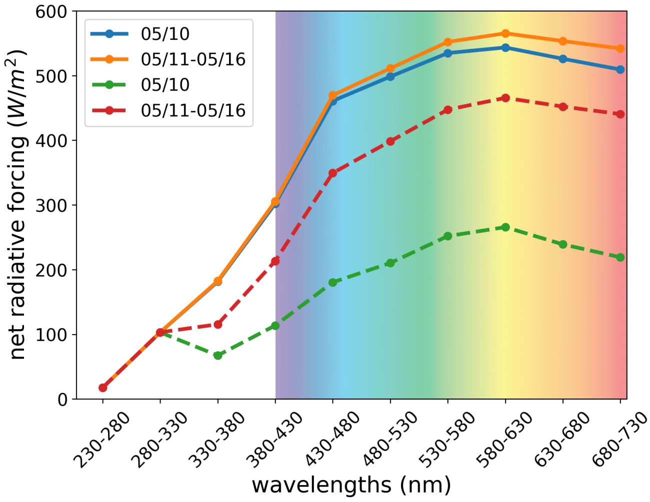


Figure 5: Radiative forcing during the mornings from 05/10/2015 to 05/16/2015 with (solid lines) and without (broken lines) aerosols obtained from the radiative forcing model. The background denotes the visible region from (400-700 nm). The ultraviolet region denotes all values to the left of 400nm.

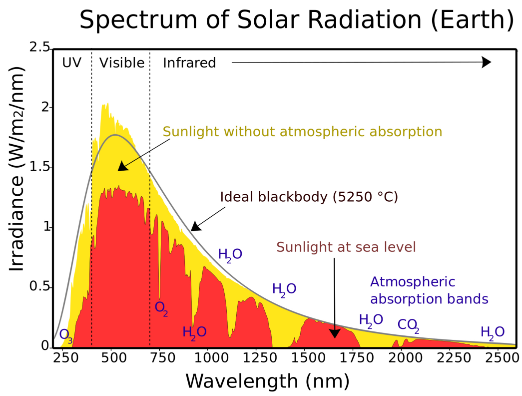


Figure 6: Spectrum of solar radiation. Credit: David Bice © Penn State University is licensed under CC BY-NC-SA 4.0

UV-visible range (shown in Fig. 6). Based on Eqn. 1, a higher radiative forcing means more energy absorbed, but less emitted back to space by Earth, and vice versa. Thus, we observe that since they produce lower radiative forcing, the absence of aerosols and/or the presence of clouds provide a certain amount of protection against UV light.

References

[1] H. D. Alas, T. Müller, W. Birmili, S. Kecorius, M. O. Cambaliza, J. B. B. Simpas, M. Cayetano, K. Weinhold, E. Vallar, M. C. Galvez, et al., Spatial characterization of black carbon mass concentration in the atmosphere of a southeast Asian megacity: An air quality case study for metro Manila, Philippines, *Aerosol and Air Quality Research* **18**, 2301 (2018), ISSN 20711409.

[2] C. D. Ahrens, *Third Edition Essentials of Meteorology* (2000), ISBN 9780123736154.

[3] L. Rebecca, Climate forcings and global warming (2014), last accessed 24 May 2021, `https://earthobservatory.nasa.gov/features/EnergyBalance/page7.php`.

[4] C. E. Chung, in *Atmospheric Aerosols*, edited by H. Abdul-Razzak (IntechOpen,

Rijeka, 2012), chap. 14, https://doi.org/10.5772/50248.

[5] P. Przyborski, Aerosols and clouds (indirect effects) (2010), last accessed 25 May 2021, https://earthobservatory.nasa.gov/features/Aerosols/page4.php.

[6] J. Hansen, M. Sato, and R. Ruedy, Tech. Rep. D6 (1997).

[7] U. Lohmann and J. Feichter, Tech. Rep. 1 (2001).

[8] A. S. Ackerman and O. B. Toon, Unrealistic desiccation of marine stratocumulus clouds by enhanced solar absorption, *Nature* **380**, 512 (1996), ISSN 1476-4687.

[9] A. S. Ackerman, O. B. Toon, D. E. Stevens, A. J. Heymsfield, V. Ramanathan, and E. J. Welton, Reduction of tropical cloudiness by soot, *Science* **288**, 1042 (2000), ISSN 00368075.

[10] J. Cook and E. J. Highwood, Climate response to tropospheric absorbing aerosols in an intermediate general-circulation model, *Quarterly Journal of the Royal Meteorological Society* **130**, 175 (2004).