Radiative flux parameterizations

Radiative fluxes are used in meteorological models in a manner similar to turbulent fluxes; profiles of fluxes are converted to flux divergences for local heating/cooling rates. We can represent the radiative flux as the sum of solar and IR components

$$R_{no} = R_{s\downarrow} - R_{s\uparrow} + R_{l\downarrow} - R_{l\uparrow}$$

Complicated radiative transfer codes based on the absorption and scattering properties of the atmosphere (temperature, humidity, ozone, carbon dioxide, aerosols, clouds) are required. Realization of the fluxes at the air-surface interface can be handled with much simpler methods. A variety of simple parameterizations are available to estimate both IR and solar downward fluxes in cloud-free conditions (usually referred to as *clear-sky* algorithms). Upward fluxes are estimated through specification of surface temperature and emissivity (for IR flux) or albedo (for solar flux). Algorithms based on near-surface air temperature and humidity are quite effective for downward IR flux over the ocean, although regional or latitudinal adjustments are usually required. Clear-sky downward solar flux algorithms require code to account for the diurnal motion of the sun as a function of latitude and time of year.

The effects of clouds represent the greatest challenge for parameterizing surface radiative fluxes. A specification of *cloud fraction* is often sufficient to define most of the variability in IR flux associated with clouds. For solar fluxes, more information is usually required depending on the time resolution required (solar zenith angle, cloud thickness or aspect ratio, cloud liquid water). Broken cloud systems are the real problem for solar fluxes because if matters profoundly if the sun is obscured by a cloud or not obscured, a process that is difficult to represent statistically. The longer the time scale or the larger the spatial scale, the better the simple statistical representations work. It is also difficult to develop and/or verify parameterizations using point measurements. Thus a solar flux parameterization developed from a month of data taken in the Bahamas might be useless at another place (or another time in the Bahamas). Still, some success has been obtained with parameterizations of *daily-average* solar flux, a quantity of use in many climate applications

1 IR Flux

In horizontally homogeneous conditions the downward IR flux, R_{ld}, can be computed with considerable accuracy using a detailed radiative transfer model by specifying the profile of temperature and the various IR absorbing materials in the atmospheric column above the surface

$$R_{l\downarrow} = \sigma \int_o^\infty < T(z) >^4 \frac{\partial \varepsilon(z)}{\partial z} dz$$

where σ is the Stefan-Boltzmann constant and ϵ the emissivity of the absorbers in the path from the surface to height z. This is strongly affected by clouds, but in clear skies is dominated by water vapor and ozone. Downward IR flux can now be computed very accurately if detailed profile information is available. In the weather/climate model context, the profile is usually fairly well known in terms of the resolved scale profiles of temperature and moisture. Clouds are the major problem because there is considerable subgrid scale variability that is very difficult to parameterize.

In models that don't supply full specification of atmospheric profiles and in many measurement programs, detailed information is not available (i.e., a simple meteorological buoy in the middle of the Pacific Ocean), then approximations are used that are based on simplifications of this equation.

$$R_{l\downarrow} = \sigma T_{rad}^4 \int_o^\infty \frac{\partial \varepsilon}{\partial z} dz = \sigma T_{rad}^4 \varepsilon_\infty = \sigma T_{eff}^4 \varepsilon_{eff}$$

where T_{rad} is an effective radiating temperature and ϵ_∞ the total column longwave emissivity. Whereas T_{rad} and ϵ_∞ are computed from the specification of the atmospheric column, T_{eff} and ϵ_{eff} are parameterizations. Both of theses parameters must be expressed in terms of known quantities such as the near-surface air temperature and humidity or the column integrated water vapor density (precipitable water). This is an example of the *measurement* parameterization problem - we have a set of data that doesn't contain all the information we need so we must estimate missing pieces using what we do have.

Although these accurate models are readily available, we often do not have the profile information they require (in fact, we *rarely* have this information). For example, an unattended buoy at some remote location over the ocean will usually have only relatively simple local near-surface data. In this case, the flux is usually estimated in terms of surface parameters (e.g., the air temperature, T_a)

$$R_{ld} = \varepsilon_e \sigma T_a^4$$

For clear sky conditions, ϵ_e becomes an empirical function of atmospheric water vapor content - either the local humidity, q_a, or the total integrated column water vapor, IV. In fact, both parameters may contribute to the flux.

The literature is replete with examples of parameterizations of ϵ_e as simple functions of q_a (the simplest are for clear skies only). Using the ETL ship-based flux data base, we have constructed a parameterization of this type for clear-sky flux where the coefficients a crude functions of latitude.

$$\varepsilon_{eclr} = a + c\sqrt{q_a}; a = 0.52 + .13/60 * abs(lat); c = 0.82 - 0.03/60 * abs(lat)$$

The fit is based on data obtained between the equator and 60 N (thus the factor of 60 in the parameterization). As we move poleward, the sensitivity to local water vapor decreases and the constant term increases. This latitudinal dependence might be unnecessary if the parameterization included both q and IV. However, IV is usually obtained by integrating a water vapor profile (e.g., from a sounding) or through retrieval using microwave radiance (either surface-based or satellite-based).

If clouds are present, then we can compute the total flux using a quantity known as the maximum cloud forcing (MCF). CF is the difference between the clear sky flux and the mean flux while MCF is the difference in the clear sky flux and the flux when the sky is totally cloud covered (cloud fraction, f = 1.0). Thus, we can approximate total downward IR flux as

$$R_{ldn} = R_{ldn\_clr} + f * MCF$$

Unfortunately, MCF depends on the type of cloud, its altitude, the cloud water content, and the relative amount of column water vapor below cloud base. Typically, the higher the cloud and the lower the liquid water then the smaller its MCF. Boundary-layer clouds tend to have large MCF (50-150 W/m^2) while cirrus clouds have MCF near 0. For example, in the tropics most CF is caused by boundary layer clouds and MCF=65 W/m^2 gives a good fit to daily-average IR fluxes (see Fig. 1).

2 Solar Flux

Clear-sky parameterizations for solar fluxes based on specification of atmospheric profiles are available that are similar to the IR flux. Comparable accuracies are more difficult because of the greater effect of aerosols, which are considerably variable seasonally and globally. Unlike IR flux, which tends to have an isotropic angular distribution, the solar flux is highly anisotropic (obviously peaking in the direction of the sun). We are using a solar flux code from Iqbal that contains components to track the angular position of the sun as a function of time of day, season, and latitude. The solar flux constant at the top of the atmosphere is known fairly accurately (1360 W/m^2) and transmission coefficient of the atmosphere is estimated given specifications of total column ozone, aerosol optical thickness in two wave bands, and the column integrated water vapor IV (as above). We use the model by plotting the time series of measured R_{sd} and comparing it to computed values of clear sky values from the algorithm. The aerosol values are then adjusted to fit the peak solar on clear days (clear days are apparent in time series) on a cruise-by-cruise basis (see Fig. 2 for an example).

One issue with this approach is the specification of IV. In cases where IV is not available, we estimate it from q_a with a ratio relationship IV=q_a/b. This ratio is highly dependent on location. For example, in subsidence regions the atmosphere above the boundary layer is very dry and b will be large. We have taken cruises where we have measurements of both IV and q_a to determine b. In the eastern tropical Pacific b has a strong latitudinal dependence in fall but a weak dependence in spring (Fig. 3).

Things become much more complicated when clouds are present. The simplest case is a horizontally homogeneous cloud (stratus or stratocumulus). For purposes of surface flux parameterizations, we can represent the flux by characterizing the one-way transmission coefficient of the cloud, T_r, which is a function (at a minimum) of the solar zenith angle, θ, the integrated liquid water content of the cloud, IW, and the size of the droplets in the cloud. Stephens published a nice parameterization of T_r as a function of $\mu=\cos(\theta)$ and the optical thickness, τ. For clouds with liquid water concentration independent of height within the cloud, we estimate optical thickness by specifying the number of droplets in the cloud and IW:

$$\tau \approx const * (IW / \rho_w)^{5/6} N^{1/3}$$

A typical midlatitude stratus cloud with 100 droplets/cm^3 and IW=100 g/m^2 will have an optical thickness of 20 and a one-way transmission coefficient of about 0.5 when the sun is approximately overhead. Cloud droplet numbers vary from 20 cm^{-3} in very clean open ocean regions to a few hundred in coastal regions (overland values can exceed 500 cm^{-3}). For boundary layer stratus clouds the liquid water concentration tends to increase linearly withing the cloud so IW varies as the square of cloud thickness. Optical thickness parameterizations for linearly varying liquid water content are slightly different that the one given above.

If the cloud is over a surface with albedo, α, then the total downward flux will be given by

$$R_{sd} = R_{sd\_clr} \frac{T_r}{1 - \alpha R_e}$$

where R_e is the cloud one-way reflection coefficient. The extra term involving albedo comes from multiple reflections between the surface and the cloud; the clear-sky flux is given my the Iqbal algorithm. The albedo of the ocean is on the order of 0.05 but is a function of zenith angle, wind speed, and cloud amount.

Broken cloud situations are extremely complicated because (a) a huge variety of cloud geometries are possible, (b) the scatter properties of solar photons have strong angular dependencies, (c) solar geometry has a large effect. Broken clouds must be treated statistically which means that simple algorithms tend to work poorly for short time periods. We have had some success treating the cloudy and clear parts of the sky separately as an effective transmission coefficient

$$< T_r >= \beta f(\theta) T_r + [1 - \beta f(\theta)]$$

Here instead of using the vertical cloud fraction, f, we attempt to account for the fact that apparent cloud fraction is a function of zenith angle (i.e., it always seems more cloudy on the horizon). The factor β is about 0.8 and accounts for the photons that leak out the sides of individual clouds.

For daily-averaged solar flux, we can use an expression involving MCF as was done with IR flux. In that case, we must specify a value for solar MCF. In the tropics a typical value is 180 W/m^2 (see Fig. 1). This approach works fairly well in the tropics where seasonal variations are a minimum. Obviously, as we go poleward the MCF takes on an increasingly strong seasonal variation.

3 *Solar-IR Flux Connections*

In the previous section we saw how clouds affect the radiative flux at the surface in opposite manner in the IR versus the solar. This compensating effect depends strongly on latitude and season, but the balance tends to be correlated with latitude or climate regime. This is illustrated in Fig. 4, which shows the Cloud Forcing phase diagram where IR MCF and solar MCF are used as X-Y axes in a cloud forcing climate region phase diagram. Note that tropical regions have the strongest solar cloud forcing and polar regions the strongest (relatively speaking) IR cloud forcing. In storm track (upper midlatitudes) SCF tends to be small because solar and IR components cancel.

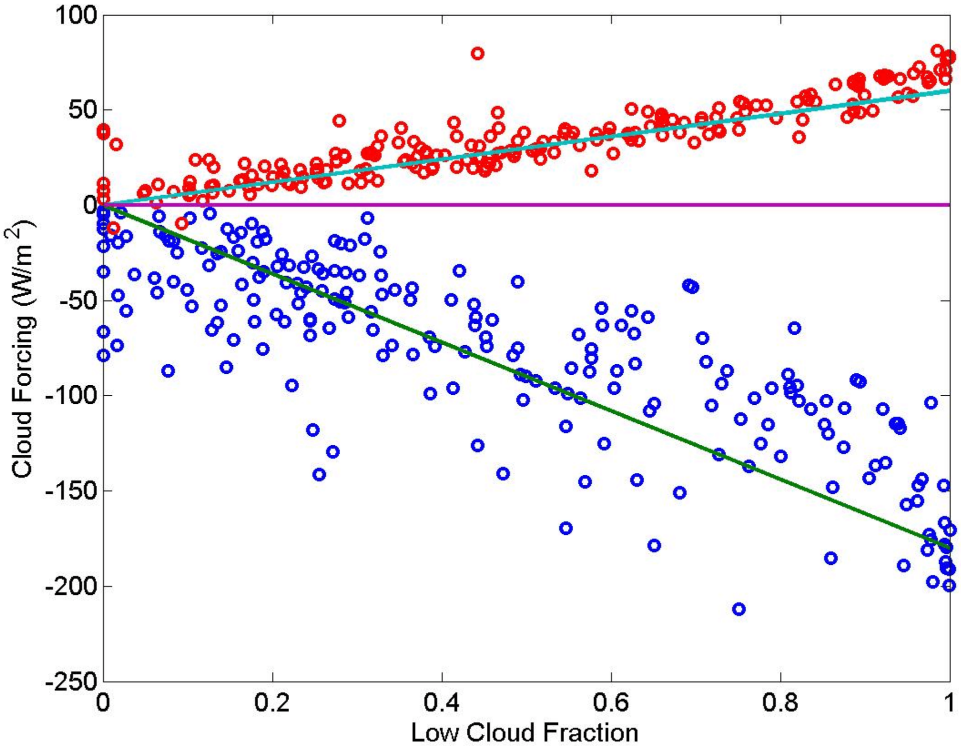

Figure 1. Daily averaged radiative surface cloud forcing as a function of cloud fraction for 6 PACS/EPIC cruises in the equatorial Eastern Pacific: red - IR flux; blue - solar flux. The lines represent MCF of +60 W/m^2 (IR) and -180 W/m^2 (solar).

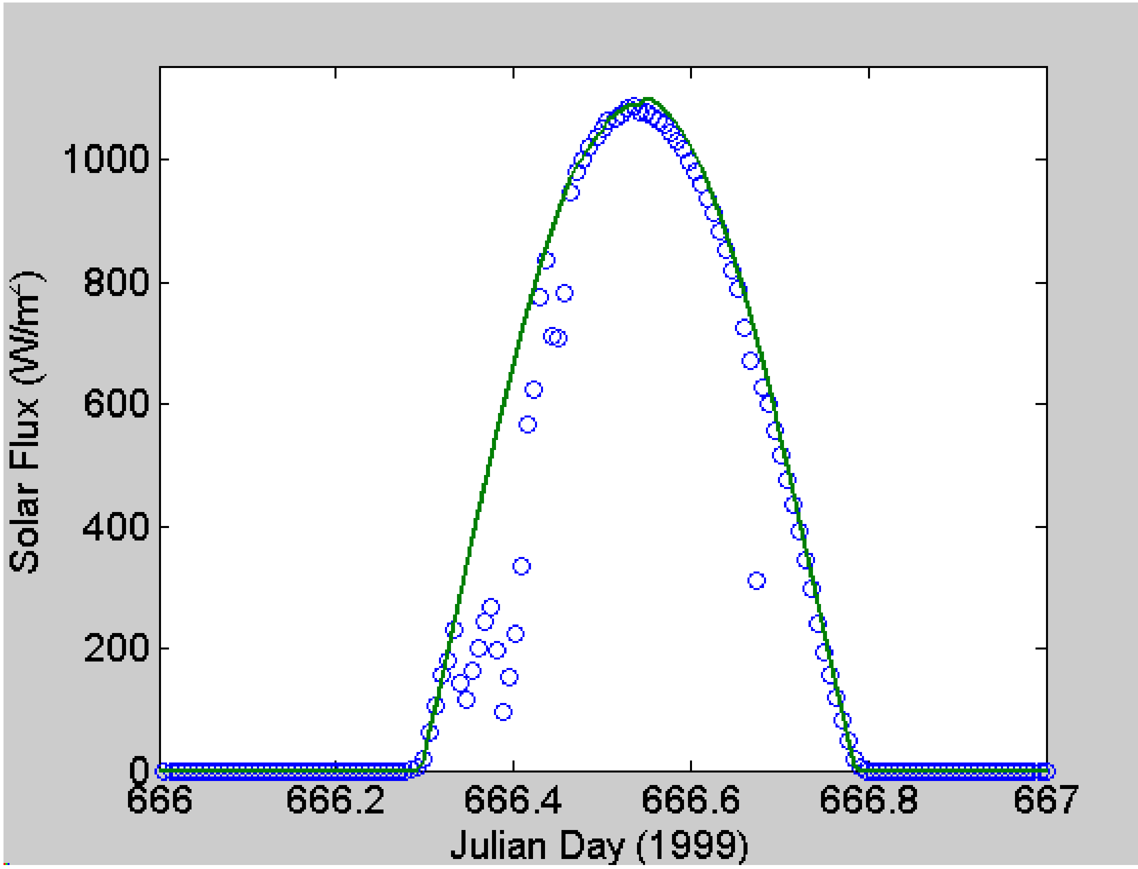

Figure 2. Time series of solar flux for one day in the fall 2000 PACS/EPIC cruise. The line is the clear sky function and the circles are measurements. Points well below the line are caused by clouds.

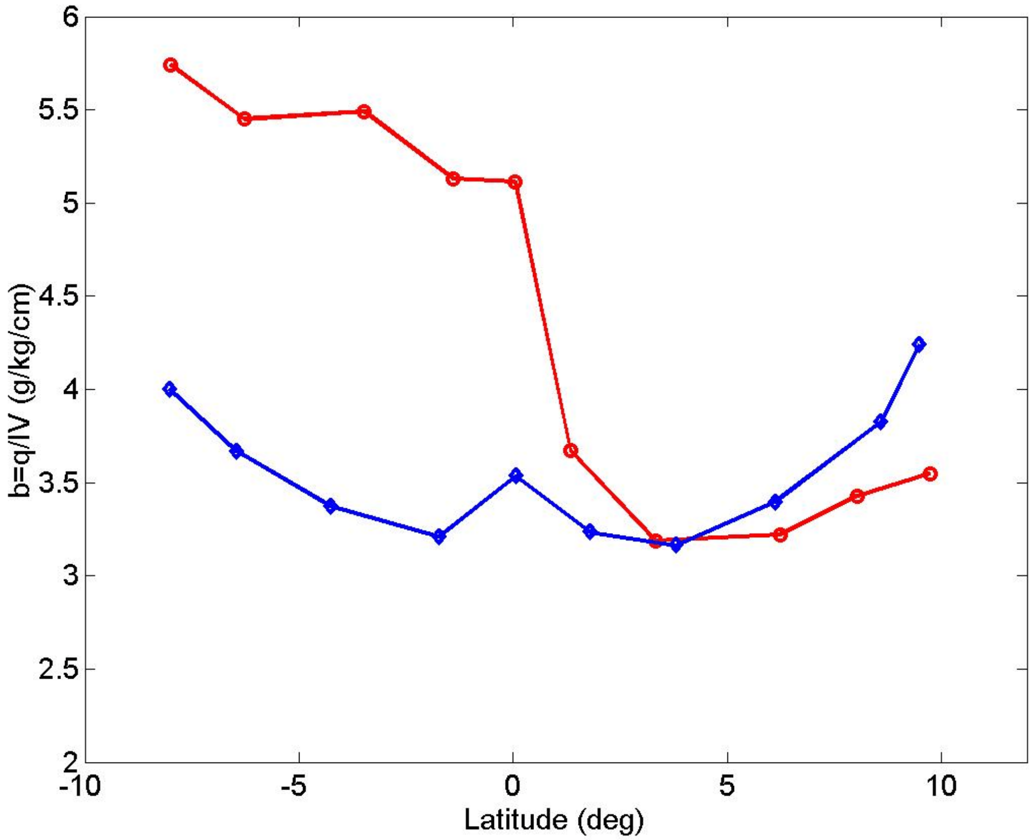

Figure 3. Average ratio of mixed layer specific humidity, q, and column integrated (precipitable) water vapor, IV: blue - spring; red - fall. The higher values in fall south of the equator are associated with strong subsidence believed to be caused by deep convection in the ITCZ north of the equator.

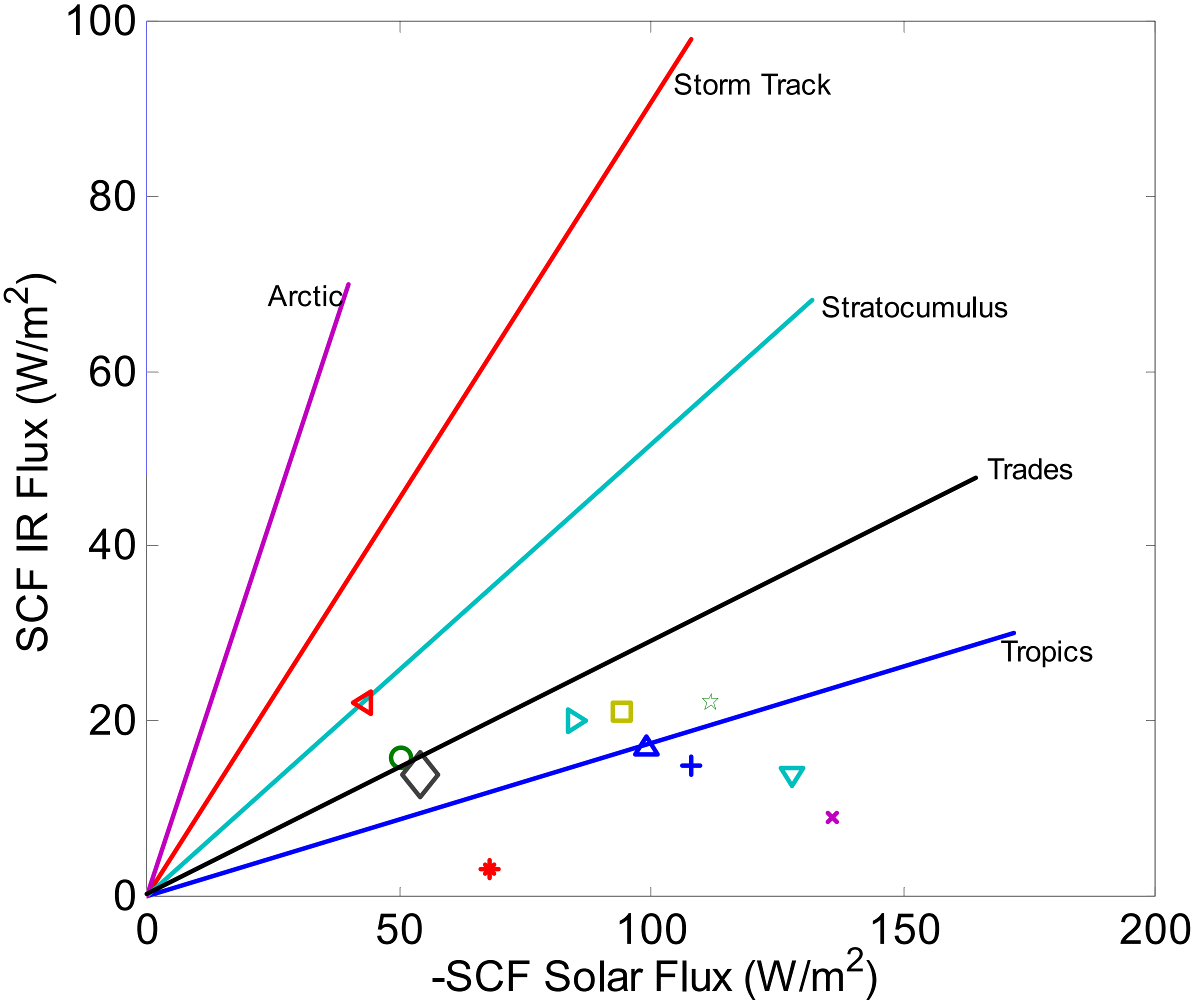

Figure 4. Phase diagram for surface cloud radiative forcing. The lines represent typical climate regimes with the nominal value at the midpoint of the line. The symbols represent individual cruise averages from 11 ETL tropical cruises. The large diamond is Nauru99. Others are: + - Coare Pilot, o - TIWE, * - COARE1, down-pointing triangle - COARE2, x - COARE3, square - JASMINE, upward point triangle - KWAJEX1, pentagram - KWAJEX2, left-pointing triangle - PACS99, and right-pointing triangle - CSP.