F. Flux Formulae

Flux computations in the Flux Coupler are denoted by a circle in Fig. 1. Many of the fluxes across the interfaces $i = 1,\ 2,\ 3$ and 4 are calculated from bulk formulae and general expressions are

$$\begin{aligned} \vec{\tau}_i &= \rho_A \ \ u_i^{*2} \ \ \Delta\vec{U}_i \ \ |\Delta\vec{U}_i|^{-1} \\ E_i &= \rho_A \ \ u_i^* \ \ Q_i^* \\ H_i &= \rho_A \ \ Cp_A \ \ u_i^* \ \ \theta_i^* \\ L\uparrow_i &= -\ \epsilon_i\ \sigma \ \ T_i^4 \ \ + \ \ \alpha_i^L \ \ L\downarrow, \end{aligned} \tag{F.1}$$

where the turbulent velocity scales are given by

$$\begin{aligned} u_i^* &= \ \ CD_i^{1/2} \ \ |\Delta\vec{U}_i| \\ Q_i^* &= \ \ CE_i \ \ |\Delta\vec{U}_i| \ \ (\Delta q_i) \ \ u_i^{*-1} \\ \theta_i^* &= \ \ CH_i \ \ |\Delta\vec{U}_i| \ \ (\Delta\theta_i) \ \ u_i^{*-1}, \end{aligned} \tag{F.2}$$

where ρ_A is atmospheric surface density, Cp_A is the specific heat, $\sigma = 5.67\times10^{-8}\mathrm{W/m^2/K^4}$ is the Stefan-Boltzmann constant, ϵ_i is the emissivity of the interface and α_i^L is the surface albedo for incident longwave radiation, $L\downarrow$. In (F.2) the differences $\Delta\vec{U}_i$, Δq_i and $\Delta\theta_i$ are defined at each interface in accord with the convention of fluxes being positive down.

The transfer coefficients in (F.2), shifted to a height, Z_i, ($Z_i = Z_A$ for $i = 1,\ 2,\ 3$), and to the stability, ζ_i, appropriate to the particular interface, are :

$$\begin{aligned} CD_i &= \kappa^2 \left[ln\left(\frac{Z_i}{Z_i^o}\right) - \psi_m \right]^{-2} \\ CE_i &= \kappa^2 \left[ln\left(\frac{Z_i}{Z_i^o}\right) - \psi_m \right]^{-1} \left[ln\left(\frac{Z_i}{Z_i^e}\right) - \psi_s \right]^{-1} \\ CH_i &= \kappa^2 \left[ln\left(\frac{Z_i}{Z_i^o}\right) - \psi_m \right]^{-1} \left[ln\left(\frac{Z_i}{Z_i^h}\right) - \psi_s \right]^{-1}, \end{aligned}$$

where $\kappa = 0.4$ is von Karman's constant and the integrated flux profiles, ψ_m for momentum and ψ_s for scalars, are functions of the stability parameter, ζ_i. These functions as used in the coupler are:

$$\begin{aligned} \psi_m(\zeta) &= \psi_s(\zeta) = -5\zeta && \zeta > 0 \\ \psi_m(\zeta) &= 2ln[0.5(1+X)] + ln[0.5(1+X^2)] - 2tan^{-1}X + 0.5\pi && \zeta < 0 \\ \psi_s(\zeta) &= 2ln[0.5(1+X^2)] && \zeta < 0 \\ X &= (1-16\zeta)^{1/4} \end{aligned}$$

Above the atmospheric interfaces $i = 1,\ 2$ and 3 the stability parameter

$$\zeta_i = \frac{\kappa\ g\ Z_A}{u_i^{*2}} \left(\frac{\theta_i^*}{\theta_v} + \frac{Q_i^*}{(Z_v^{-1} + q_A)} \right) \quad ,$$

where virtual potential temperature is computed as $\theta_v = \theta_A(1 + Z_v q_A)$, q_A and θ_A are the lowest level atmospheric humidity, and potential temperature, respectively, and $Z_v = (\varrho(water)/\varrho(air)) - 1 = 0.606$. Since ζ_i is itself a function of the turbulent scales (F.2), and hence the fluxes, an iterative procedure is generally required to solve (F.1).

In addition to surface fluxes, the atmospheric model requires effective surface albedos for both direct α (dir), and diffuse, α (dif), radiation at each wavelength. They are used in a single call to the computationally demanding atmospheric radiation routines. This call gives downward atmospheric albedo for diffuse radiation, $\alpha_a(dif)$. If direct and diffuse albedos, $\alpha^m(dir)$ and $\alpha^m(dif)$ for $m = 1, M$ different surfaces below an atmospheric grid cell are known, the multiple scattering coefficients, $R^m = (1 - \alpha_a(dif)\,\alpha^m(dif))^{-1}$ are computed, and with f^m the fractional coverage of each surface type, then the albedos are given by

$$\alpha(dif) = (1 - F_2) \,/\, (1 - F_2/, \alpha_a(dif))$$

$$\alpha(dir) = F_1\,(1 - \alpha_a(dif)\,\alpha(dif))$$

$$F_1 = \sum_{m=1}^{M} f^m\, R^m\, \alpha^m(dir)$$

$$F_2 = \sum_{m=1}^{M} f^m\, R^m\, (1 - \alpha^m(dif))$$

Use of these albedos ensures that the solar radiation exiting the bottom of the atmosphere is identical to the sum of M radiation calculations, each using an $\alpha^m(dif)$ and $\alpha^m(dir)$. At present, however, the albedo calculations are greatly simplified by assuming $\alpha_a(dif) = 0$, such that $R^m = 1$. This albedo then does not need to be passed from the atmosphere to the coupler, and the coupler simply computes the effective albedos to be passed to the atmosphere as

$$\alpha(dif) = \sum_{m=1}^{M} f^m\, \alpha^m(dif)$$

$$\alpha(dir) = \sum_{m=1}^{M} f^m\, \alpha^m(dir)$$

In general the partition of radiation among the M surfaces is a function of R^m, $\alpha^m(dif)$, and $\alpha^m(dir)$, and hence of wavelength. Hence, correct partitioning would need to be performed for each wavelength band within the radiation transfer code of the atmospheric model. This procedure is greatly simplified by partitioning the net solar radiation within the Flux Coupler.

I. Fluxes Across the Atmosphere - Cryosphere Interface (1)

The state variables required to calculate the fluxes (F.1) for $i = 1$, are the atmospheric wind velocity, potential temperature, density and specific humidity in (kg/kg) at a height Z_A, and the ice surface temperature $T_1 = T_C$. The ice surface humidity is assumed to be saturated :

$$q_1 = \rho_A^{-1} \;\; C_5 \;\; \exp(C_6/T_1),$$

where $C_5 = 640380.$ kg/m^3 and $C_6 = -5107.4$ K. The differences in (F.2) become

$$\begin{aligned} \Delta\vec{U}_1 &= \vec{U}_A - \vec{U}_1 \\ \Delta q_1 &= q_A - q_1 \\ \Delta\theta_1 &= \theta_A - T_1 \;\; , \end{aligned}$$

where the ice velocity, $\vec{U}_1$, is presently assumed to be negligible.

Presently the coupler assumes constant and equal roughness lengths for ice :

$$Z_1^o \;=\; Z_1^h \;=\; Z_1^e \;=\; 5 \times 10^{-4} m.$$

To start the iterative solver for the fluxes (F.1), ζ_1 is set to zero : $\psi_m = \psi_s = 0$, and the neutral coefficients are used to approximate the flux scales (F.2). The iteration begins by computing ζ_1, finding the non-neutral transfer coefficients and updating the flux scales. After two such iterations there is sufficient convergence to compute the fluxes (F.1). Also in (F.1), the reflected downward incident longwave radiation is simply accounted for by assuming an emissivity, $\epsilon_1 = 1$, and the ice surface albedo for incident longwave radiation, $\alpha_1^L = 0.0$.

II. Fluxes Across the Atmosphere - Biosphere Interface (2)

The fluxes across the Atmosphere - Biosphere Interface (2) are computed in the biosphere model. Their formulae can be written in the forms of (F.1) and are detailed in Bonan (1996), "A Land Surface Model (LSM version 1.0) For Ecological, Hydrological, and Atmospheric Studies : Technical Description and User's Guide".

III. Fluxes Across the Atmosphere - Hydrosphere Interface (3)

At the atmosphere-hydrosphere interface the near-surface air is also assumed to be saturated,

$$q_1 = 0.98 \;\; \rho_A^{-1} \;\; C_5 \;\; \exp(C_6/T_3),$$

where the sea surface temperature is $T_3 = T_H$, and the factor 0.98 accounts for the salinity of the ocean. The differences (F.2) become

$$\begin{aligned} \Delta\vec{U}_3 &= \vec{U}_A - \vec{U}_3 \\ \Delta q_3 &= q_A - q_3 \\ \Delta\theta_3 &= \theta_A - T_3, \end{aligned}$$

where the surface current, $\vec{U}_3$, is presently assumed to be negligible.

The roughness length for momentum, Z_3^o in meters, is a function of the atmospheric wind at 10 meters height, U_{10}:

$$Z_3^o = 10 \ \exp - \left[\kappa \left(\frac{C_1}{U_{10}} + C_2 + C_3 U_{10} \right)^{-1} \right],$$

where $C_1 = 0.0027$ m/s, $C_2 = 0.000142$, and $C_3 = 0.0000764$ m^{-1}s. The corresponding drag coefficient at 10m height and neutral stability is

$$C_{10}^N \ = \ C_1 \ U_{10}^{-1} \ + \ C_2 \ + \ C_3 \ U_{10} \ .$$

The roughness length for heat, Z_3^h, is a function of stability, and for evaporation, Z_3^o, is a different constant:

$$\begin{aligned} Z_3^h &= 2.2 \times 10^{-9} m \qquad \zeta_3 > 0 \\ &= 4.9 \times 10^{-5} m \qquad \zeta_3 \leq 0 \\ Z_3^e &= 9.5 \times 10^{-5} m \ . \end{aligned}$$

Since the roughness lengths are neither constant nor equal, the iterative solution for the fluxes (F.1) differs somewhat than that for sea–ice. First, ζ_3 is set incrementally greater than zero when the air–sea temperature difference suggests stable stratification, otherwise it is set to zero. In either case, $\psi_m = \psi_s = 0$, and the initial transfer coefficients are then found from the roughness lengths at this ζ_3 and $U_{10} = U_A$. As with sea–ice, these coefficients are used to approximate the initial flux scales (F.2) and the first iteration begins with an updated ζ_3 and calculations of ψ_m and ψ_s. The wind speed, U_A, is then shifted to its equivalent neutral value at 10m height :

$$U_{10} = U_A \ \ \left(1 + \frac{\sqrt{C_{10}^N}}{\kappa} ln(\frac{Z_A}{10} - \psi_m(\zeta_3)) \right)^{-1} \ .$$

This wind speed is used to update the transfer coefficients and hence the flux scales. The second and final iteration begins with another update of ζ_3. The final flux scales then give the fluxes calculated by (F.1). As for the sea–ice interface (1), the reflected downward incident longwave radiation is simply accounted for by assuming an emissivity, $\epsilon_3 = 1$, and the water surface albedo for incident longwave radiation, $\alpha_3^L = 0.0$.

IV. Fluxes Across the Cryosphere - Hydrosphere Interface (4)

Only the stress between the water and ice can be computed using bulk formulae similar to (F.1). The upper water column stability is assumed to be near neutral ($\zeta_4 = 0.0; \psi_m(0) = 0.0$), so that CD_4, and hence $\vec{\tau}_4$ depend only on a constant roughness length and the velocity difference,

$$\begin{aligned} Z_4^o &= \ \ 0.05m \\ \Delta\vec{U}_4 &= \vec{U_C} - \vec{U_H} \ \ , \end{aligned}$$

where $\vec{U_C}$ and $\vec{U_H}$ are, respectively, the ice velocity and surface current. The implied drag coefficient for $Z_4 = 10m$ is $CD_4 = 0.0057$.

However, with the present sea–ice model, the stress is not computed in this fashion, but is given by a linear function of $\Delta\vec{U_4}$, as described in Section G.

The heat and freshwater fluxes between the cryosphere and hydrosphere due to the freezing and melting of sea-ice are computed, respectively, in the hydrosphere (see ocean model documentation) and cryosphere (see ice model documentation) component models. Each column of the hydrospheric component needs to be scanned after each time step to see if any ice is formed. The cumulative heat of fusion associated with the freezing is output as Q_H. The sense of this heat transfer is downwards, so Q_H is positive. The heat and freshwater fluxes associated with freezing are applied directly by the hydrospheric model and not passed to the coupler other than by the cumulative $Q_H > 0$.

If there is no freezing,

$$Q_H = \frac{\rho_H \; Cp_H \;\; \Delta_1 \;\; (T_f - T_H)}{t_m} \quad < 0 \quad ,$$

is used to represent the potential of the upper ocean layer (thickness Δ_1) to melt sea–ice, where t_m is a melting time scale. Since the hydrosphere itself does not know how much ice is available to be melted it does not respond to $Q_H < 0$ in any way. Instead the heat and freshwater fluxes associated with melting, both due to warm surface water Q_H, and atmospheric heating are computed within the cryosphere and passed to the Flux Coupler as H_4 and F_4, respectively.

The solar radiation passing through the bottom of the ice, S_4, is found in the cryosphere component by reducing the atmospheric solar, S_1, to account for heat absorption by the ice and snow.

V. Fluxes Across the Biosphere - Hydrosphere Interface (5)

The only important direct exchange between the biosphere and hydrosphere across interface 5 is river runoff. The runoff from each of the land model grid points is accumulated into number $N_B (< 100)$ of drainage basins. At present N_B is 19, and this accumulation is done in the land model. These drainage basins must correspond to where the freshwater will be distributed in the ocean model. The runoff vector $R_n, n = 1, N_B$ is then passed to the Flux Coupler. The Flux Coupler then passes R_n to the ocean model without modification. Note that this implies no time delay in the land hydrological cycle between precipitation and its discharge into the ocean, and enables conservation of freshwater in the fully coupled model to be achieved.

At present river runoff is not utilized by the ocean model. The freshwater contained in the active ocean is conserved in the coupler by multiplying the precipitation over the ocean by the uniform factor needed to balance the global average ocean evaporation every coupling period, see section G. In the future, runoff will be redistributed within the ocean model to model gridpoints consistent with the N_B drainage basins.