I. Surface Forcing Fields

The uncoupled NCAR CSM ocean model is forced in a similar manner as when it is coupled to an atmospheric model. The open ocean surface boundary conditions are given as air-sea fluxes of momentum, heat and freshwater. The heat flux is comprised of shortwave solar and infrared longwave radiation, and of the turbulent latent and sensible heating. The freshwater flux includes evaporation and precipitation, in addition to a weak restoring to observed salinity. In areas of sea-ice, traditional strong restoring boundary conditions to observed temperature and salinity are used. In all cases these observed restoring temperatures and salinities are averages over the upper 50 meters of the Levitus (1982) climatologies.

The open ocean turbulent heat fluxes are computed from a prescribed atmospheric state using traditional, bulk formulae. The required surface winds, air temperature and air humidity are obtained from the 6 hourly National Meteorological Center (NMC) global reanalysis data set (Kalnay et al., 1995). These NMC fields are supplemented by cloud fraction (Rossow and Schiffer, 1991) and surface insolation estimates (Bishop and Rossow, 1991) from the International Satellite Cloud Climatology Project (ISCCP) and monthly precipitation computed from the Microwave Sounding Unit (MSU) data (Spencer, 1993). Sea surface temperature (SST) is either the model's prognostic upper layer temperature, or the monthly SST climatology of Shea et al. (1990), designated now as STR.

This option is activated by choosing the ifdef option *sflxcombo*. When used in coupled mode and communicating with the Flux Coupler, the ifdef option *sflxpvm* should be chosen. In addition, the model computes the heat and freshwater fluxes due to simple, local formation and melting of sea-ice. The presence of this sea-ice does not modify the other surface fluxes in any way. This is activated by choosing the ifdef option *sflxiceflx*.

Surface flux calculations

Table 1: Data sources used to construct the prescribed atmospheric state for forcing the uncoupled NCAR CSM ocean model. Data are taken from global numerical weather prediction and satellite data products.

	Variable	Source	Original time interval
U_{10}	zonal wind component	NMC	6 hours
V_{10}	meridional wind component	NMC	6 hours
T_a	air temperature	NMC	6 hours
q_a	specific humidity	NMC	6 hours
C	cloud fraction	ISCCP	daily
Q_{sw}	surface insolation	ISCCP	daily
P	precipitation	MSU	monthly

The NMC reanalysis data includes four times daily (0Z, 6Z, 12Z, 18Z) values of zonal (U_{10}) and meridional (V_{10}) 10m winds, surface 2m air (T_a) and 2m specific humidity (q_a) on a T62 grid (Table 1). The zonal and meridional wind stress components τ_x and τ_y are computed every 6 hours following Large and Pond (1981) using bulk turbulent formula:

$$\tau_x = \rho_a C_D U_{10} W_{10}$$

$$\tau_y = \rho_a C_D V_{10} W_{10}$$

where W_{10} is the 10m wind speed ($W_{10} = (U_{10}^2 + V_{10}^2)^{1/2}$), ρ_a is the air density and C_D is an empirical drag coefficient where the neutral, 10m drag coefficient is given by:

$$10^3 C_D = \frac{2.70}{W_{10}} + 0.142 + 0.0764 W_{10}.$$

Appropriate stability corrections are applied based on the air-sea temperature difference (e.g. Arya, 1988) using the STR ocean temperatures. Mean monthly climatologies of $\tau_x, \tau_y, W_{10}, T_a$ and q_a have been produced from the four times daily data from four years 1985 through 1988. Furthermore, mid-month climatologies of each of these quantities are formed, such that when linearly interpolated to each model time step the monthly means exactly equal the monthly climatological values.

These linear interpolation of τ_x and τ_y directly force the momentum equations. The heat equation is forced by the net open ocean air-sea heat flux Q_{oce}, which is the sum of the sensible, latent, net longwave, and net shortwave fluxes:

$$Q_{\text{oce}} = Q_{\text{sen}} + Q_{\text{lat}} + Q_{\text{lw}}^{\text{net}} + Q_{\text{sw}}^{\text{net}}.$$

The turbulent heat fluxes, Q_{sen} and Q_{lat}, are computed from standard air-sea transfer equations (Large and Pond, 1982):

$$Q_{\text{sen}} = \rho_a c_p^a C_H W_{10}(T_a - T_s)$$

$$Q_{\text{lat}} = \rho_a L C_E W_{10}(q_a - q_s)$$

where c_p^a and L are the specific heat of air and latent heat of freshwater, T_s and q_s are the sea surface temperature and implied saturated specific humidity from the model, and T_a and q_a are the linear interpolations from the mid-month climatologies. C_H and C_E are the transfer coefficients for heat and water, the neutral forms of which are given by:

$$10^3 C_H = 32.7 C_D^{1/2} \text{ unstable}$$

$$10^3 C_H = 18.0 C_D^{1/2} \text{ stable}$$

$$10^3 C_E = 34.6 C_D^{1/2}.$$

Monin–Obukhov similarity theory is used to adjust these coefficients for the local atmospheric stability based on model SST and the 2m and 10m heights of the atmospheric variables.

The net longwave radiation $Q_{\rm lw}^{\rm net}$ is calculated from a conventional bulk radiation formula of Berliand and Berliand (Fung *et al.*, 1984):

$$Q_{\rm lw}^{\rm net} = -\epsilon\sigma(T_a^4[0.39 - 0.05(e_a)^{0.5}]F(C) + 4T_a^3(T_s - T_a))$$

where T_a and T_s are in Kelvin, e_a (mbars) is the surface water vapor pressure found from the NMC q_a, ϵ is the surface emissivity taken as 1.0, and σ is the Stefan-Boltzmann coefficient equal to $567{\times}10^{-10}$W m^{-2} K^{-4}. The effect of clouds on the longwave flux is parameterized with a cloud correction factor $F(C)$ following Budyko (1974):

$$F(C) = 1 - a_c C^2$$

where the cloud fraction C varies from 0. to 1.0, and a_c is a latitude dependent empirical coefficient adapted by Fung *et al.* (1984) from Bunker (1976). The cloud fraction data are from the ISCCP C1 data set, which contains daily total cloud fraction (Table 1) estimates globally on a $2.5° \times 2.5°$ grid (Rossow and Schiffer, 1991). The ISCCP cloud data coverage is poor in the winter high latitudes because of low insolation and missing geosynchronus satellite data. The local annual mean is used to fill in missing data for regions with partial coverage. For a few areas, in particular the Indian sector of the southern ocean, the cloud fraction data was recomputed from the Bishop and Rossow (1991) surface clear and total insolation data using zonal mean linear regressions between C and the ratio clear sky to total insolation.

The presence of clouds has an opposite and partially compensating effect on longwave (warming) and shortwave (cooling) forcing of the ocean surface, and it is important, therefore, to have a consistent treatment of clouds for both radiative components. Bishop and Rossow (1991) present an algorithm, based on a one-dimensional atmospheric radiative transfer model, for computing surface shortwave irradiance from the ISCCP cloud data on the same grid as the C1 cloud data. The daily mean surface insolation from Version 2 (Bishop et al., 1995) is used here, and a constant albedo of 7% is applied to the downward shortwave flux to compute $Q_{\rm sw}^{\rm net}$. Shortwave radiation is allowed to penetrate below the model surface leading to sub-surface bulk heating. The subsurface profile for shortwave radiation is computed with the two band approximation of Paulson and Simpson (1977):

$$Q_{\rm sw}(z) = Q_{\rm sw}^{\rm net}[re^{z/\zeta_1} + (1-r)e^{z/\zeta_2}],$$

where the band fraction r and depth scales ζ are specified with spatially and temporally uniform values equivalent to Jerlov water type Ib: ($r = 0.67$, $\zeta_1 = 1.0$m, $\zeta_2 = 17.0$m). The sub-surface heating rate is then calculated from $\partial T/\partial t = (dQ_{\rm sw}/dz)/(\rho_w c_p)$, where ρ_w and c_p are the density and specific heat of water. This option is activated by specifying the ifdef option *sflxsw*.

The net ocean freshwater flux $F_{\rm oce}$ is the sum of the evaporation E and precipitation P rates and a weak restoring term $F_{\rm res}$ minus its global open ocean average $< F_{\rm res} >$:

$$F_{\rm oce} = E + P + F_{\rm res} - < F_{\rm res} >$$

The evaporation is given directly from Q_{lat} by dividing by the latent heat of vaporization L. Monthly precipitation (Table 1) estimates from the MSU satellite analysis (Spencer, 1993) are used to form mid-month climatologies as above. These MSU data, P_{MSU}, have been calibrated against coastal and island rain-gauge data. However, averaged over the ocean the annual mean MSU precipitation is 10 to 30 percent lower than given by other (eg. LeGates and Willmott, 1990) climatologies (D. Shea, personal communication, 1996). The MSU precipitation data is limited to 60°S to 60°N, and the high latitude regions are filled with the monthly, zonal average precipitation values from the poleward most latitude bands.

The local open-ocean salinity restoring term is given by:

$$F_{\mathrm{res}} = -R^{o}_{fw}(S_L - S_s)$$

where R^{o}_{fw} is the restoring coefficient ($23.0\ kg\ m^{-2}\ s^{-1}\ PSU^{-1}$), S_s is model surface salinity and S_L is the 50 meter average Levitus climatological salinity. This coefficient corresponds to a 2 year restoring timescale over the model upper layer thickness of 12.5m.

A partial sea-ice model was used for the uncoupled NCAR CSM model experiments comprised of two parts: strong restoring of heat and freshwater fluxes and an ice adjustment step. The under ice fluxes are given by:

$$Q_{\mathrm{ice}} = R^{i}_{q}(T_L - T_s)$$

$$F_{\mathrm{ice}} = -R^{i}_{fw}(S_L - S_s)$$

where the restoring coefficients are $R^{i}_{q} = 386\ Wm^{-2}K^{-1}$ and $R^{i}_{fw} = (27\ kg\ m^{-2}\ s^{-1}\ PSU^{-1})$, both equivalent to a 6 day restoring timescale over 12.5m.

The distribution of sea-ice is diagnosed from the STR temperature field using the weighting function w:

$$w = \frac{T_0 - T_f}{T_p - T_f} \quad \text{for} \quad T_f < T_0 < T_p$$

$$w = 0 \quad T_0 \leq T_f \quad \text{and} \quad w = 1 \quad T_0 \geq T_p$$

where T_f and T_p are -1.8°C and -0.8°C. The open-ocean and sea-ice fluxes are then blended through the transition region:

$$Q_{\mathrm{net}} = wQ_{\mathrm{oce}} + (1 - w)Q_{\mathrm{ice}}$$

and

$$F_{\mathrm{net}} = wF_{\mathrm{oce}} + (1 - w)F_{\mathrm{ice}}.$$

The presence of sea-ice does not alter the transfer of momentum to the ocean. Even when run in stand alone mode, the ocean model can form and melt ice locally, but not advect it around. How this is done is described previously in Section H. The changes in upper ocean temperature and net creation of sea-ice can, of course, be interpreted as ocean heat and freshwater fluxes. This is governed by ifdef option *sflxiceflx*. There is no Mediterranean Sea in the 3×3 model, but it is resolved in the 2×2 model. By setting ifdef option *restormed*, strong 6 day restoring fluxes in the Mediterranean are imposed.

Balancing The Annual Heat and Freshwater Fluxes

There is considerable uncertainty in all components of the heat and freshwater fluxes. However, a much more accurate assumption is that over an annual cycle the global average of the net fluxes are both about zero. When the heat fluxes are computed as outlined above using STR surface temperatures, the global average net heat flux, $< Q_{net} >$ is nearly 50 Wm^{-2}. Before an annual balance would be achieved the uncoupled ocean surface temperatures would need to be about $1°C$ higher. In order to achieve a balance with realistic surface temperatures the solar radiation is reduced and the latent heat flux is made more negative (more evaporation).

The Bishop and Rossow (1991) data show generally good agreement with long term, weathership surface irradiance climatologies (e.g. Dobson and Smith, 1988), with a positive bias of 10–20 Wm^{-2} at some stations (Bishop *et al.*, 1995). This bias is only one of a growing number of empirical indications (including comparisons with land stations) that radiative transfer equations transmit too much solar radiation to the surface (J. Kiehl, personal communication, 1996). Therefore, it is reduced by uniformly multiplying the solar radiation by a factor 0.875. Kent and Taylor (1995) correct observations from a select subset of North Atlantic Volunteer Observing Ships (VOS) to produce an annual cycle of latent heat flux for a portion of the North Atlantic between 30 and $55°N$. This annual cycle has systematically more negative latent heat flux by 10 to $20W/m^{-2}$ than the corresponding cycle from the NMC and STR data. The difference is too large to be attributed to uncertainties in the wind speed and the transfer coefficient. However, the two annual cycles are brought into good agreement by multiplying the NMC 2m humidity by a uniform factor of 0.93 to produce the q_a used in the above bulk formula for latent heat flux. This drying adjustment may reflect a tendency for model assimilated ocean humidity observations from (VOS) to be too moist (Kent et al., 1993).

Since precipitation rates over the ocean are generally poorly constrained, the freshwater budget is controlled by computing the annual average open ocean precipitation P_y and the annual volume integrated freshwater tendency, $d_t F_y$, for each year (y) of model integration. This requires setting the ifdef option *sflxcomboavan*, in order to calculate this annual tendency. The precipitation over the next year is then given by

$$P = f_{y+1} P_{MSU}$$

$$f_{y+1} = f_y (1 - d_t F_y / P_y),$$

so that the precipitation is reduced (enhanced) if over the previous year the ocean salinity decreased (increased). Before the 6 percent reduction of q_a, evaporation is about 0.90 P_{MSU} on a global annual average. After the correction it is increased to about 1.10 P_{MSU}. Therefore, the precipitation factor during year 1 of the model integration is $f_1 = 1.10$, making the precipitation forcing of the model more comparable with other climatologies, as noted above. In subsequent years this factor first decreases to about 1.05 then increases to an equilibrium value of 1.105. Note, the open ocean freshwater restoring has no impact on the precipitation, because the global open ocean mean restoring is subtracted. Without this subtraction restoring also responds to changes in precipitation, and even a two year restoring time constant is strong enough that an increase (decrease) in precipitation results in an increase (decrease) in global averaged ocean salinity over an annual cycle.

Table 2 shows the global annual mean heat, freshwater and momentum fluxes and flux components that result from the corrections and STR surface temperatures. As desired, the net heat and freshwater fluxes are near zero. The restoring fluxes in this calculation are zero, because everywhere the ocean and restoring temperatures and salinities are equal. This will not be true during a model integration, but the average open ocean restoring freshwater flux will be zero.

Table 2: Global annual mean heat, freshwater, and momentum fluxes for uncoupled NCAR CSM model averaged over ice-free region of model (total area: 3.523×10^{8} km^2, ice-free area: 3.285×10^{8} km^2).

	STR SST	NCAR CSM
Qnet (W/m^2)	-0.7	—
Qsw (W/m^2)	161.4	—
Qlw (W/m^2)	-53.0	—
Qlat (W/m^2)	-100.1	—
Qsen (W/m^2)	-8.9	—
Qice (W/m^2)	0.0	—
Fnet (m/y)	-0.001	—
Evap (m/y)	-1.290	—
Prec (m/y)	1.289	—
Fice (m/y)	0.000	—
Fres (m/y)	0.000	—
Taux (N/m^2)	4.06×10^{-3}	—
Tauy (N/m^2)	-0.66×10^{-3}	—

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