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ONE DIMENSIONAL CLIMATE MODEL

The one-dimensional climate model is composed of coupled atmospheric and land surface components. The atmospheric module includes convective adjustment, cloud and radiative parameterizations, as well as vertical diffusion processes. The land surface module is designed to be optional from three alternatives: a) a simple parameterization of hydrology based on the so-called bucket hydrology; b) a more complex parameterization based on the BATS scheme; and c) a statistical-dynamical representation of the land surface hydrology that accounts explicitly for the subgrid scale variability.

ATMOSPHERIC MODULE

The atmospheric component consists of a twelve layer atmospheric column whose pressure levels correspond identically to those used in NCAR's CCM. Each layer is characterized by its pressure, temperature and specific humidity. Model forcing is in the form of incoming solar radiation at the top of the column and seasonal cycles of lateral heat and moisture convergences. The thermodynamic state of each layer is propagated in time through radiative, convective, cloud, and diffusive processes whose parameterizations are briefly described below. See Williamson et al., (1987) for more details.

Once the new temperatures and mixing ratios are evaluated at the end of each time step, convective and non convective adjustments are performed to create mutually adjusted temperature and specific humidity fields. Net radiative fluxes at the surface, and heating rates and net radiative heating for each atmospheric layer are computed. Once the energy and mass balances at the surface are solved, the resulting fluxes of mass, and heat are used as boundary conditions for the vertical diffusion parameterization. Surface energy and water balances are computed with three different options as presented below.

CONVECTIVE AND STABLE CONDENSATION ADJUSTMENTS

At every time step, the temperature and specific humidity of each layer are adjusted to create neutrally stratified atmospheric layers. Three types of adjustments are carried out:

a) dry convective adjustment takes place when the predicted atmosphere is not saturated but the lapse rate exceeds the dry adiabatic lapse rate; temperature is then adjusted to the dry adiabatic lapse rate, and the humidity field is reset assuming complete mixing;

b) moist convective adjustment takes place when the atmosphere is supersaturated and the lapse rate exceeds the moist adiabatic lapse rate; in this case the moisture and temperature fields are simultaneously adjusted so that the resulting atmosphere is just saturated and neutrally stable; excess water is assumed to fall as precipitation,

$$PREC_c^n = \frac{1}{\Delta t}\sum_{k=1}^{K} \Delta p_k (T_k - \hat{T}_k) C_p \; / \; (Lg\rho_{H_2O}) \qquad \text{(A1)}$$

where the subscript k refers to a specific atmospheric layer, and the superscript n refers to the specific time step. L is the latent heat of condensation, C_p is the specific heat at constant pressure, and the $\wedge$ stands for variable before the modification at the given time step. T is temperature and p is atmospheric pressure.

c) stable condensation occurs when the lapse rate is stable but the atmosphere is supersaturated; in this case the moisture field is adjusted to be just saturated and the temperature field is modified to reflect the latent heating; excess water is assumed to fall as precipitation. The new specific humidity is given by,

$$q_k = \left[\hat{q}_{sk} + \frac{d\hat{q}_{sk}}{dT}(T_k - \hat{T}_k) \right] \qquad \text{(A2)}$$

where $\hat{q}_{sk}$ is the saturation specific humidity. The temperature change due to the release of latent heat during condensation is,

$$(T_k - \hat{T}_k) = \frac{L}{C_p}(\hat{q}_k - q_k) \qquad \text{(A3)}$$

and the corresponding stable precipitation is given by,

$$PREC_s^n = \frac{1}{\Delta t}\sum_{k=1}^{K} \Delta p_k (\hat{q}_k - q_k) \; / \; (g\rho_{H_2O}) \qquad \text{(A4)}$$

Cloud Parameterization

Convective clouds form in the columns in which the moist convective adjustment takes place. The fractional cloud cover in each layer involved in convective adjustment, A_k, is given as,

$$A_k = 0.3 / n_{ad} \tag{A5}$$

where n_{ad} is the number of layers with convective clouds. The total area of sky covered with these clouds, A_T, is computed under the assumption of random overlap as,

$$A_T = 1 - \prod_{k=Kl}^{Ku} (1 - A_k) \tag{A6}$$

where K_l and K_u are the lowest and highest level with convective clouds. Emissivities for convective clouds are assumed to be equal to unity.

Non-convective clouds are formed where the stable condensation adjustment takes place. These clouds are assumed to induce a 0.95 cloud cover in the respective layer, and their emissivities are computed as a function of liquid water content.

Radiation Parameterization

Radiation flux in the solar spectral region is divided into ultraviolet-visible radiation (UVV) and near infrared radiation (NIR). Fractional absorption by ozone in the UVV region, and by water vapor, molecular oxygen, and carbon dioxide in the NIR region are also accounted for. The resulting solar fluxes at the surface are computed taking into account the surface albedoes for both direct and diffuse radiation in each of the two spectral regions. The clear sky flux at the surface accounts for direct-beam solar radiation as well as reflection of diffuse radiation between the surface and molecular (Rayleigh) scatter from the atmosphere.

Solar heating within the atmosphere is separated into three regions, above clouds, within clouds, and below clouds. Heating occurs through three processes, direct beam absorption of the downward flux, absorption of radiation directly reflected off the cloud tops, and a contribution from solar radiation that is scattered between clouds and the surface.

Longwave radiation flux is calculated at each level in both the up and down directions by solving the radiative transfer equations and considering ozone, water vapor, and carbon dioxide.

Surface Temperature Calculation

The surface temperature is computed from the surface energy balance equation,

$$\sigma_B (T_s^n)^4 - F^{S^n} - F_d (p_s)^n + H_{K+1/2}^n + LR_{K+1/2}^n + Q_{Ice}^n = 0 \tag{A7}$$

where σ_B is the Stefan-Boltzmann constant. F^S is the net downward solar radiation flux at the surface, and $F_d(p_s)$ is downward longwave radiation flux at the surface which are provided by the radiation parameterization. $H_{K+1/2}$ is the vertical flux of sensible heat at the surface, $R_{K+1/2}$ is the evaporation mass flux, L is the latent heat of condensation/evaporation, and Q_{Ice} is heat conduction from sea ice (where present) to water below. The latent heat flux (or evaporation) is computed as,

$$R^n_{K+1/2} = \overline{\rho}^{n-1}_K C_H D_W \left|\overline{V}^{n-1}_K\right| \left(q_s(T^n_s) - \overline{q}^{n-1}_K\right) \tag{A8}$$

where K is the index referring to the first atmospheric layer above the surface, q_s is the saturation specific humidity for temperature T_s and pressure p_s. C_H is the drag coefficient and D_W is the evaporation factor, $|V|$ is the surface wind velocity. The density ρ is computed from the ideal gas law, modified to account for water vapor content. The energy balance equation above is implicit in temperature and is solved by the Newton-Raphson procedure.

Vertical Diffusion

The vertical diffusive fluxes of heat and water vapor are proportional to the vertical gradients of sensible and latent heat fluxes at each atmospheric layer, respectively. Only equations for water vapor are presented. Water vapor is diffused upward as,

$$\begin{aligned} q^{n+1}_k &= \overline{q}^{n-1}_k + 2\Delta t \frac{g(R^{n+1}_{k+1/2} - R^{n+1}_{k-1/2})}{(\overline{p}^{n-1}_{k+1/2} - \overline{p}^{n-1}_{k-1/2})} \\ R^{n+1}_{k+1/2} &= (\overline{\rho}^{n-1}_{k+1/2})^2 \frac{g\overline{K}^{n-1}_{k+1/2}(q^{n+1}_{k+1} - q^{n+1}_k)}{(\overline{p}^{n-1}_{k+1} - \overline{p}^{n-1}_k)} \end{aligned} \tag{A9}$$

where K is a function of the horizontal wind velocity at the given level, as well as of the Richardson number.

BUCKET HYDROLOGY PARAMETERIZATION

Precipitation resulting from moist convection and stable condensation processes increases either the soil moisture or snow cover depending on the surface temperature and the temperature of the first two model levels above the surface. If either of the two first atmospheric levels or the surface are at temperatures above freezing, precipitation is assumed to fall as rain. If all of these three temperatures are below freezing, precipitation is assumed to fall as snow. If as a result the soil moisture exceeds field capacity, the excess becomes runoff and soil moisture is set to field capacity.

The wetness factor for evaporation is calculated as follows,

$$D_W = \begin{cases} 1.0 & \text{if } W \geq W_c \\ W / W_c & \text{if } W < W_c \end{cases} \tag{A10}$$

with the critical soil moisture value W_c determined from filed capacity as,

$$W_c = 0.75 W_{FC} \tag{A11}$$

and W_{FC} is assumed to be equal to 0.15 m.

BIG-LEAF LAND SURFACE HYDROLOGY

The *big-leaf* land surface option, which is based on the BATS hydrology model (Dickinson, et al., 1986), uses the force-restore method to solve the energy budget equation. The soil and subsurface temperatures are obtained from,

$$\begin{aligned}\frac{\partial T_{g1}}{\partial t} &= \frac{2\sqrt{\pi} h_s}{\rho_s c_s d_1} - \frac{2\pi (T_{g1} - T_{g2})}{\tau_1} \\ \frac{\partial T_{g2}}{\partial t} &= -c_3 \left[\frac{(T_{g2} - T_{g1})}{\tau_1} + c_4 (T_{g2} - T_{g3}) + Q_{sf} \right]\end{aligned} \tag{A12}$$

where T_{g1} is the surface soil temperature, T_{g2} is the subsurface soil temperature, and T_{g3} is a fixed annual mean deep soil temperature. c_3 is the rate of subsoil relaxation which depends on how deep the soil thermal reservoir is. c_4 is the damping parameter to damp soil surface temperature to the annual mean value. τ_1 is the period of heating in seconds (1 day). Q_{sf} is the rate of subsoil temperature change due to melting or freezing. $\rho_s c_s$ is the specific heat of surface layer per unit mass. d_1 is the depth of soil influenced by a periodic heating. h_s is the heat energy balance at the surface.

Soil Moisture in the Absence of Vegetation

Three soil layers are considered in computing the mass water balance. Water diffuses through the soil with a diffusivity given by,

$$D = -K_w \frac{\partial \psi}{\partial s} = K_{Wo} B s^{B+2} \tag{A13}$$

In addition to the diffusive flow there is gravitational drainage or percolation given by,

$$R_g = K_{wo} s^{2B+3} \tag{A14}$$

where K_{wo} is the hydraulic conductivity at saturation, ψ is the soil matric potential, and s is the soil saturation. B is a constant with values in the range 3 - 11.

Evaporation is computed as the minimum of potential evaporation rate imposed by atmospheric conditions and the diffusion limited exfiltration capacity rate of the soil. The potential rate, F_{qp}, is computed as above but using a wetness factor equal to unity. The exfiltration rate, F_{qm} is computed as,

$$F_{qm} = C_k D s_o \;/\; (Z_o Z_1)^{1/2}$$
$$D = 1.02 D_{max} s_1^B s_o^2 (s_o \;/\; s_1)^{B_f} \quad \text{(A15)}$$

where Z_o is the depth of the active soil layer, and Z_1 is the depth of the surface soil layer. D is an average soil diffusivity. C_k is a function of soil diffusivity, saturated hydraulic conductivity, and saturated soil matric potential. s_i represents percent saturation in soil layer i.

Surface Runoff

In trying to account for both infiltration excess and saturation excess runoff mechanisms, runoff, R_s is parameterized as follows,

$$R_s = \begin{cases} (\rho_w \;/\; \rho_{ws})^4 G, & T_{g_2} \geq 0℃ \\ (\rho_w \;/\; \rho_{ws}) G, & T_{g_2} < 0℃ \end{cases} \quad \text{(A16)}$$

where ρ_{ws} is the saturated soil water density, ρ_w is the soil water density weighted toward the top layer, as defined by,

$$\rho_w = \rho_{ws} \frac{(s_0 + s_1)}{2} \quad \text{(A17)}$$

and G is defined as,

$$G = P_r + S_m - F_q \quad \text{(A18)}$$

where P_r is precipitation, S_m is snowmelt, and F_q is actual evapotranspiration. Only snow-surface processes are modeled. There is no explicit distinction between snow and soil temperatures. During rain on snow events, or during snowmelt periods, the model puts the water directly into the soil, neglecting percolation and freezing inside the snow pack.

The drag coefficient for the boundary layer parameterization is computed as a function of the drag coefficient for neutral stability conditions, and of the surface bulk Richardson number. The neutral drag coefficient is obtained using mixed-layer theory. Over vegetated areas, the neutral drag coefficient is estimated as a linear combination of drag coefficients over the vegetated, and bare fractions of the area.

ENERGY FLUXES WITH VEGETATION

Foliage variables are parameterized in terms of the leaf area index (LAI) and the stem area index (SAI). LAI has a seasonal variation, whereas the SAI is assumed constant. The fractional area available for transpiration is computed by accounting for the fraction of the leaves that is wetted (dew formation, interception) and thus unable to transpire. Interception of precipitation takes place before canopy drip and stem flow take place. The water stored by the vegetation canopy is computed based on the following equation,

$$\frac{\partial W_{dew}}{\partial t} = \sigma_f P_r - E_f + E_{tr} \quad \text{(A19)}$$

where E_f is water flux from the foliage (evaporation), E_{tr} is transpiration flux, and σ_f is the fractional vegetation cover. These fluxes are computed using the resistance formulation. The flux from wet foliage is parameterized in terms of the aerodynamic resistance. The flux from dry foliage (transpiration) needs consideration of the stomatal resistance.

The stomatal resistance refers to the total mechanical resistance encountered by diffusion from inside the leaf to the outside ambient air. The stomatal resistance factor is taken as,

$$r_s = r_{s\,\min} R_l S_l M_l \quad \text{(A20)}$$

where R_l is the stomatal dependence on solar radiation; S_l is the seasonal dependence of the stomatal resistance, which is expressed as a function of leaf temperature; and M_l represents the influence of soil moisture on the stomatal resistance.

STATISTICAL-DYNAMICAL PARAMETERIZATION

The statistical-dynamical option uses a derived distribution approach in order to account for subgrid scale variability of precipitation and soil moisture distribution. The heat and moisture balance equations for the soil layer are,

$$\begin{aligned} n\Delta z \frac{ds}{dt} &= P(1-R) - \beta e_p \\ (c_g + c_w \rho n s \Delta z)\frac{dT_g}{dt} &= R_s - R_o - L\beta e_p - H \end{aligned} \quad \text{(A21)}$$

where n is soil porosity, s is relative soil saturation, P is precipitation, R is the runoff ration, β is the evaporation efficiency function, R_s is the net absorbed solar radiation, R_o is the net upward longwave radiation, e_p is the potential evaporation rate, H is the latent heat flux, and Δz is the depth of the soil layer. The evaporation efficiency and the runoff ratio functions link the mass and energy balances at the ground surface.

The runoff ratio, R, is the ratio of actual to potential runoff. Assuming that the point rainfall intensity follows an exponential distribution with parameter $\kappa/E[P]$, and that the spatial distribution of soil moisture follows a two parameter gamma probability distribution function, $G(\alpha,\lambda)$, Entekhabi and Eagleson (1989) obtained the following expression for the runoff ratio,

$$R = \left[1 - \frac{\gamma\left(\alpha, \frac{\alpha}{E[s]}\right)}{\Gamma(\alpha)}\right] + \left[\frac{e^{-\kappa I(1-\nu)}\gamma\left(\alpha, \kappa I \nu + \frac{\alpha}{E[s]}\right)}{\left(\frac{\kappa I \nu E[s]}{\alpha} + 1\right)^{\alpha}\Gamma(\alpha)}\right] \quad \text{(A22)}$$

where κ represents the scaling necessary in order to distribute precipitation over the typical area of storm events; *E[]* is the expectation operator; $\gamma(,)$ and $\Gamma()$ are the incomplete and the complete gamma functions, respectively. *I* is the ratio of saturated hydraulic conductivity to the precipitation rate, *K(1)/E[P]*; and *v* is given by,

$$v = \left.\frac{d\psi}{ds}\right|_{s=1} \frac{1}{\Delta z} \tag{A23}$$

where ψ is the soil matric potential.

The evaporation efficiency function is the ratio of actual to potential evaporation and is given by (Entekhabi and Eagleson, 1989),

$$\beta_s = 1 - \frac{\gamma(\alpha, \alpha\xi^{-1})}{\Gamma(\alpha)} + \frac{\Omega'\,\gamma(\frac{1}{2m} + 2 + \alpha, \alpha\xi^{-1}) - \frac{1}{2}\,\gamma(\frac{2}{m} + 3 + \alpha, \alpha\xi^{-1})}{\left[\Omega'\,(\alpha\xi^{-1})^{\frac{1}{2m}+2} - \frac{1}{2}(\alpha\xi^{-1})^{\frac{2}{m}+3}\right]\Gamma(\alpha)}$$
$$\Omega' = \Omega[\frac{\alpha}{E[s]}]^{3/2m+1} \tag{A24}$$

for bare soil. For the vegetated fraction of the grid the evaporation efficiency function is given by (Entekhabi and Eagleson, 1989),

$$\beta_v = 1 + \frac{\gamma(\alpha+1, \alpha\xi^{-1}) - \alpha\xi^{-1}\gamma(\alpha, \alpha\xi^{-1}) - \gamma(\alpha+1, \alpha W^{-1}) + \alpha W^{-1}\gamma(\alpha, \alpha W^{-1})}{\Gamma(\alpha)(\alpha\xi^{-1} - \alpha W^{-1})}$$
$$W = \frac{E[s]}{s_w} \tag{A25}$$

where s_w is soil saturation at permanent wilting point. In the equations above ξ and Ω are given as,

$$\xi = e[s]\left[\frac{K(1)\Omega}{e_p}\right]^{\frac{2m}{1+4m}}$$
$$\Omega = \left[\frac{8n\psi(1)}{3K(1)(1+3m)(1+4m)\Delta t}\right]^{1/2} \tag{A26}$$

where *m* is the pore size distribution index; *n* is the soil porosity; *K(1)* is saturated hydraulic conductivity; and $\psi(1)$ is the soil matric potential at saturation conditions.