Layout

- Introduction
- General remarks
- Model development and validation
- The surface energy budget
- The surface water budget
- Link with CO2 budget
- Soil heat transfer
- Soil water transfer
- Snow
- Initial conditions
- Conclusions and a look ahead
Thermal budget of a ground layer at the surface

\[ \varepsilon_g \sigma T^4 \]

\[ \varepsilon_g R_T^\downarrow \]

\[ \alpha R_s^\downarrow \]

\[ D \]

\[ T_g \]

\[ \epsilon \]

\[ 4 \]

\[ R \]

\[ \rho \]

\[ \partial \]

\[ = \]

\[ + \]

\[ \frac{\partial T_s}{\partial t} \]

\[ R_n + H + LE + G \]
Energy budget: Summer examples

Dry

Fir canopy

Fig. 2.3 Observed diurnal energy balance over a dry lake bed at El Mirage, California, on June 10 and 11, 1950. [After Vehrencamp (1953).]

Fig. 2.5 Observed energy budget of a Douglas fir canopy at Haney, British Columbia, on July 23, 1970. [From Oke (1987); after McNaughton and Black (1973).]

Barley field

Fig. 2.4 Observed diurnal energy budget of a barley field at Rothamsted, England, on July 23, 1963. [From Oke (1987); after Long et al. (1964).]

Arya, 1988
The surface radiation

Surface albedo
Surface emissivity
Skin temperature

- In some cases (snow, sea ice, dense canopies) the impinging solar radiations penetrates the “ground” layer and is absorbed at a variable depth. In those cases, an extinction coefficient is needed.

<table>
<thead>
<tr>
<th>Surface type</th>
<th>Other specifications</th>
<th>Albedo ((a))</th>
<th>Emissivity ((e))</th>
</tr>
</thead>
<tbody>
<tr>
<td>Water</td>
<td>Small zenith angle</td>
<td>0.03–0.10</td>
<td>0.92–0.97</td>
</tr>
<tr>
<td></td>
<td>Large zenith angle</td>
<td>0.10–0.50</td>
<td>0.92–0.97</td>
</tr>
<tr>
<td>Snow</td>
<td>Old</td>
<td>0.40–0.70</td>
<td>0.82–0.89</td>
</tr>
<tr>
<td></td>
<td>Fresh</td>
<td>0.45–0.95</td>
<td>0.90–0.99</td>
</tr>
<tr>
<td>Ice</td>
<td>Sea</td>
<td>0.30–0.40</td>
<td>0.92–0.97</td>
</tr>
<tr>
<td></td>
<td>Glacier</td>
<td>0.20–0.40</td>
<td></td>
</tr>
<tr>
<td>Bare sand</td>
<td>Dry</td>
<td>0.35–0.45</td>
<td>0.84–0.90</td>
</tr>
<tr>
<td></td>
<td>Wet</td>
<td>0.20–0.30</td>
<td>0.91–0.95</td>
</tr>
<tr>
<td>Bare soil</td>
<td>Dry clay</td>
<td>0.20–0.35</td>
<td>0.95</td>
</tr>
<tr>
<td></td>
<td>Moist clay</td>
<td>0.10–0.20</td>
<td>0.97</td>
</tr>
<tr>
<td></td>
<td>Wet fallow field</td>
<td>0.05–0.07</td>
<td></td>
</tr>
<tr>
<td>Paved</td>
<td>Concrete</td>
<td>0.17–0.27</td>
<td>0.71–0.88</td>
</tr>
<tr>
<td></td>
<td>Black gravel road</td>
<td>0.05–0.10</td>
<td>0.88–0.95</td>
</tr>
<tr>
<td>Grass</td>
<td>Long (1 m)</td>
<td>0.16–0.26</td>
<td>0.90–0.95</td>
</tr>
<tr>
<td></td>
<td>Short (0.02 m)</td>
<td></td>
<td></td>
</tr>
<tr>
<td>Agricultural</td>
<td>Wheat, rice, etc.</td>
<td>0.10–0.25</td>
<td>0.90–0.99</td>
</tr>
<tr>
<td></td>
<td>Orchards</td>
<td>0.15–0.20</td>
<td>0.90–0.95</td>
</tr>
<tr>
<td>Forests</td>
<td>Deciduous</td>
<td>0.10–0.20</td>
<td>0.97–0.98</td>
</tr>
<tr>
<td></td>
<td>Coniferous</td>
<td>0.05–0.15</td>
<td>0.97–0.99</td>
</tr>
</tbody>
</table>

*Compiled from Sellers (1965), Kondratyev (1969), and Oke (1978).*

Arya, 1988
The other terms

Sensible heat flux

\[ H = \rho C_h u_L (C_p T_L + g z - C_p T_{sk}) \]

\[ C_h = f (Ri_B, z_{oh}, z_{om}) \]

\[ z_{oh}, z_{om} \quad \text{specify the surface} \]

Evaporation

\[ E = \rho C_h u_L \left[ q_L - q_s \right] = \rho C_h u_L \left[ a_L q_L - a_s q_{sat} (T_{sk}, p_s) \right] \]

\[ a_{L,s} = f (q_L, T_s) \]

Ground heat flux

\[ (\rho C)_g \frac{\partial T_s}{\partial t} = - \frac{\partial G}{\partial z} = \frac{\partial}{\partial z} \lambda_T \frac{\partial T}{\partial z} \]

\[ (\rho C)_g, \lambda_T = f (\text{soil type, other soil characteristics}) \]
Recap: The surface energy equation

\[(1 - \alpha)R_s \downarrow + \varepsilon_g R_T \downarrow - \varepsilon_g \sigma T_{sk}^4 + \rho C_h u_L (C_p T_L + g z - C_p T_{sk}) + \rho C_h u_L [a_L q_L - a_s q_{sat} (T_{sk}, p_s)] + G(T_s, T_{sk}) = (\rho C)_g D \frac{\partial T_s}{\partial t}\]

- Equation for $T_s, T_{sk}$
- For:
  - a thin soil layer at the top $(\rho C)_g D \frac{\partial T_s}{\partial t} \approx 0$
  - $G(T_s, T_{sk})$ is known, or parameterized or $G \ll R_n$

we have a non-linear equation defining the skin temperature
<table>
<thead>
<tr>
<th>Land</th>
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</tr>
</thead>
<tbody>
<tr>
<td>High vegetation</td>
<td>Open sea / unfrozen lakes</td>
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<td>Sea ice / frozen lakes</td>
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<tr>
<td>High vegetation with snow</td>
<td></td>
</tr>
<tr>
<td>beneath</td>
<td></td>
</tr>
<tr>
<td>Snow on low vegetation</td>
<td></td>
</tr>
<tr>
<td>Bare ground</td>
<td></td>
</tr>
<tr>
<td>Interception layer</td>
<td></td>
</tr>
</tbody>
</table>
## TESSEL geographic characteristics

<table>
<thead>
<tr>
<th>Fields</th>
<th>ERA15</th>
<th>TESSEL</th>
</tr>
</thead>
<tbody>
<tr>
<td>Vegetation</td>
<td>Fraction</td>
<td>Fraction of low</td>
</tr>
<tr>
<td>Vegetation type</td>
<td>Global constant</td>
<td>Fraction of high</td>
</tr>
<tr>
<td></td>
<td>(grass)</td>
<td>Dominant low type</td>
</tr>
<tr>
<td></td>
<td></td>
<td>Dominant high type</td>
</tr>
<tr>
<td>Albedo</td>
<td>Annual</td>
<td>Monthly</td>
</tr>
<tr>
<td>LAI</td>
<td>Global constants</td>
<td>Annual, Dependent</td>
</tr>
<tr>
<td></td>
<td></td>
<td>on vegetation type</td>
</tr>
<tr>
<td>$r_{\text{min}}$</td>
<td></td>
<td></td>
</tr>
<tr>
<td>Root depth</td>
<td>1 m</td>
<td>Annual, Dependent</td>
</tr>
<tr>
<td></td>
<td></td>
<td>on vegetation type</td>
</tr>
<tr>
<td>Root profile</td>
<td>Global constant</td>
<td></td>
</tr>
</tbody>
</table>
High vegetation fraction at T511 (now at T1279)

Aggregated from GLCC 1km
Low vegetation fraction at T511 (now at T1279)

Aggregated from GLCC 1km
High vegetation type at T511 (now at T1279)

![Map showing high vegetation types aggregated from GLCC 1km](image)

Aggregated from GLCC 1km
Low vegetation type at T511 (now at T1279)

Aggregated from GLCC 1km
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Evaporation: Idealized surfaces

- Lake
- Wet vegetation
- Snow
- Dry vegetation
- Bare ground
Potential evaporation, bucket model

- **Potential evaporation**
  - The evaporation of a large uniform surface, sufficiently moist or wet (the air in contact to it is fully saturated)

  \[ E_{pot} = \rho C h u_L \left[ q_L - q_{sat} (T_{sk}, p_s) \right] \]

- **Evaporation efficiency**
  - Ratio of evaporation to potential evaporation

  \[ E = \beta E_{pot} \]

  \[ 0 < \beta < 1 \]

- **Bucket model** Budyko 1963, 1974 Manabe 1969

\[ \beta = \begin{cases} 
1 & \theta > \theta_{crit} \\
\frac{\theta}{\theta_{crit}} & \theta > \theta_{crit}
\end{cases} \]

Ideally, \( \beta \) should be a well defined function of soil cover (vegetation type vs. bare ground) and soil properties.
A general, algebraic formulation

\[ E = \rho C_h u_L \left[ a_L q_L - a_s q_{sat}(T_{sk}, p_s) \right] \]
\[ a_{L,s} = f(q_L, T_s, \text{state and nature of the soil, soil cover}) \]

● Two limit behaviours

- **Bare soil**: Evaporation dependent on soil water (and trapped water vapor) in a top shallow layer of soil (~ 20 mm).

- **Vegetated surfaces**: Evaporation controlled by a canopy resistance, dependent on shortwave radiation, water on the root zone (~ 1-5 m deep) and other physical/physiological effects.
Bare ground evaporation

- Soil (bare ground) evaporation is due to:
  - Molecular diffusion from the water in the pores of the soil matrix up to the interface soil atmosphere ($z_{0q}$)
  - Laminar and turbulent diffusion in the air between $z_{0q}$ and screen level height

- All methods are sensitive to the water in the first few cm of the soil (where the water vapour gradient is large). In very dry conditions, water vapour inside the soil becomes dominant

\[ E = \rho \frac{q_L - \alpha q_{sat}(T_{sk})}{r_a} \]

\[ \alpha = f(\theta_1) \quad "Relative humidity of the soil" \]

\[ \theta_1 \quad Top soil layer (a few cm) water \]
TESSEL bare ground evaporation

\[ i \] is the bare ground tile

\[ E_i = \frac{\rho_a}{r_{soil} + r_{a,i}} [q_a - q_{sat}(T_{sk,i})] \]

\[ r_{soil} = r_{soil,min} f_3(\theta_1) \]

\[ f_2(\theta_1)^{-1} = \begin{cases} 
0 & \theta_{1,\text{liq}} < \theta_{\text{pwp}} \\
\frac{\theta_{1,\text{liq}} - \theta_{\text{pwp}}}{\theta_{\text{cap}} - \theta_{\text{pwp}}} & \theta_{\text{pwp}} < \theta_{1,\text{liq}} < \theta_{\text{cap}} \\
1 & \theta_{1,\text{liq}} > \theta_{\text{cap}}
\end{cases} \]
New bare soil evaporation
(Balsamo et al. 2011, Albergel et al. 2012)

The introduction of bare ground evaporation revision (green-line) is quite
effective in reducing the soil moisture below the wilting point in non-vegetated
area (upper panel of figure above, at 79% bare ground, SCAN site in Utah).

\[ R_c = R_{soil} f_2(w_1) \]
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<td></td>
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</table>
Transpiration: The big leaf approximation

- **Sensible heat \((H)\), the resistance formulation**

\[
H = \rho C_p C_h u_L (T_L - T_{sk}) = \rho C_p \frac{T_L - T_{sk}}{r_a}
\]

\[r_a = \frac{1}{C_h u_L}\]

- **Evaporation \((E)\), the resistance formulation (the big leaf approximation, Deardorff 1978, Monteith 1965)**

\[
E = \rho \frac{q_L - q_c}{r_a + r_c} = \rho \frac{q_L - q_{sat}(T_{sk})}{r_a + r_c}
\]

\[q_c = q_{sat}(T_{sk})\]  
Specific humidity for the interior of the stomata, i.e., for saturated conditions 

\[r_c\]  
canopy resistance
Some plant science

\[ r_c = \frac{\langle r_s \rangle}{L} \]

\[ \langle ( ) \rangle = \frac{L}{\int dL / ( )} \]

\( L \) Leaf area index (see later for a definition)

- \( r_s \), the stomatal resistance of a single leaf. Physiological control of water loss by the vegetation. Stomata (valve-like openings) regulate the outflow of water vapour (assumed to be saturated in the stomata cells) and the intake of CO\(_2\) from photosynthesis. The energy required for the opening is provided by radiation (Photosynthetically Active Radiation, PAR). In many environments the system appears to be operate in such a way to maximize the CO\(_2\) intake for a minimum water vapour loss. When soil moisture is scarce the stomatal apertures close to prevent wilt and dessication of the plant.
Jarvis approach (1)

\[ r_c = r_{c_{\text{min}}} f_1(PAR) f_4(T_s) f_3(D_a) f_2(\theta) \]

- \( T_s \): soil temperature
- \( D_a = e_{\text{sat}}(T_a) - e \): near-surface saturation deficit
- \( f_i > 1 \)

![Graphs showing the inverse of stomatal resistances](image)

Fig. 4. Environmental dependencies of the inverse of stomatal resistances (i.e., conductance) in BATS model: (a) dependence of conductance on temperature; (b) dependence of conductance on vapor pressure deficit; (c) dependence of conductance on PAR.

Dickinson et al. 1991
Jarvis approach (2)

**Shuttleworth 1993**

![Graph showing moisture extraction function](image)

**FIGURE 4.4.3** Typical variation of the moisture extraction function $f(\theta)$ which modifies the potential crop coefficient in response to changes in the volumetric soil moisture content $\theta$ in (portions of) the plants' rooting zone. $\theta_s$, $\theta_f$, and $\theta_w$ are the values of $\theta$ at saturation, field capacity, and wilting point, respectively, and $(\theta_d/\theta_f)$ is typically 0.5 to 0.8. These values are determined by soil type.

- $\theta$ soil water in the root zone
- $\theta_d = \theta_{cap}$ in many models (ECMWF)
- $\theta_{ava} = \theta_d - \theta_{pwp}$ availability
- $\theta_{ava} \cdot d$ available soil water (water holding capacity)
TESSEL transpiration

\( i \) is the high/low vegetation tile

\[
E_i = \frac{\rho_a}{r_{c,i} + r'_{a,i}} \left[ q_a - q_{sat}(T_{sk,i}) \right]
\]

\[
r_c = \frac{r_{s,\text{min}}}{LAI} f_1(R_S^\uparrow) f_3(\bar{\theta}) f_4(D_a)
\]

\[
f_1(R_S^\uparrow)^{-1} = \min \left[ 1, \frac{a(1 + bR_S^\uparrow)}{bR_S^\uparrow + c} \right]
\]

\[
f_3(D_a)^{-1} = \exp(-g_D D_a) \quad g_D = 0 \quad \text{fc}
\]

\[
f_2(\bar{\theta})^{-1} = \begin{cases} 
0 & \bar{\theta} < \theta_{\text{pwp}} \\
\frac{\bar{\theta} - \theta_{\text{pwp}}}{\theta_{\text{cap}} - \theta_{\text{pwp}}} & \theta_{\text{pwp}} < \bar{\theta} < \theta_{\text{cap}} \\
1 & \bar{\theta} > \theta_{\text{cap}}
\end{cases}
\]

\( \bar{\theta} \) is a root weighted average of the liquid soil water
TESSEL root profile

Integrated root fraction

1–19 BATS+Zeng et al. Solid: low veg  Dash: High veg

- 1 Crops, mixed farming
- 2 Short grass
- 3 Evergreen Needleleaf Trees
- 4 Deciduous Needleleaf Trees
- 5 Deciduous Broadleaf Trees
- 7 Tall Grass
- 8 Desert
- 9 Tundra
- 10 Irrigated Crops
- 11 Semidesert
- 16 Evergreen Shrubs
- 17 Deciduous Shrubs
- 18 Mixed Forest
- 19 Interrupted Forest
**TESSEL evaporation: shaded snow**

\[ E = E_{\text{veg}} + E_{\text{snow}} \]

\[ E_{\text{snow}} = \frac{\rho_a}{r_{a,s} + r_a} [q_a - q_{\text{sat}}(T_{\text{sn}})] \]

- Evaporation is the sum of the contribution of the snow underneath (with an additional resistance, \( r_{a,s} \), to simulate the lower within canopy wind speed) and the exposed canopy. The former is dominant in early spring (frozen soils) and the latter is dominant in late spring.
Interception (1)

- **Interception layer** represents the water collected by interception of precipitation and dew deposition on the canopy leaves (and stems).

- **Interception** (I) is the amount of precipitation (P) collected by the interception layer and available for “direct” (potential) evaporation. I/P ranges over 0.15-0.30 in the tropics and 0.25-0.50 in mid-latitudes.

- **Leaf Area Index** (LAI) is (projected area of leaf surface)/(surface area) 0.1 < LAI < 6

- **Two issues**
  - Size of the reservoir
  - $C_I$, fraction of a gridbox covered by the interception layer

- $T = P - I$; **Throughfall** (T) is precipitation minus interception
Interception: Canopy water budget

\[
\frac{\partial w_l}{\partial t} = e_i P^* + c_l E_l + D = I + c_l E_l
\]

- \( w_l \): Intercepted water
- \( e_i \): Efficiency of interception
- \( P^* \): Modified precipitation
- \( c_l E_l \): Evaporation of intercepted water
- \( D \): Rate of drainage at the bottom of the canopy
**TESSEL: interception**

- Interception layer for rainfall and dew deposition

\[
\frac{\partial w_i}{\partial t} = I + c_i E_i
\]

\[
I = \max \left( e_i c_i P^*, \frac{w_{lmx} - w^*_l}{\Delta t} \right)
\]

\[
P^* = \frac{P}{k} \quad \text{modified precipitation}
\]

\[
k \quad \text{fraction of grid-box covered by precipitation}
\]

\[
T = P - I \quad \text{Throughfall (input to top soil)}
\]
Example: Deep tropics interception

ARME, 1983-1985, Amazon forest
Accumulated water fluxes

Viterbo and Beljaars 1995
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Towards an increased vegetation complexity

“old/standard” LSMs

“new/photosynthesis-based” LSMs

(Calvet et al. 2005)

The stomatal aperture controls the ratio:

\[ A_n = \frac{\alpha}{r_{cc}} (C_s - C_i) \]

\[ E = \frac{\beta}{r_c + r_a} (q_a - q_{sat}), r_c = f(r_{cc}) \]
For vegetated area the evapotranspiration is parameterized as:

Where the canopy resistance $r_c$ is defined following Jarvis (1976) as:

$$r_c = \frac{r_{s,\text{min}}}{LAI} f_1(R_s) f_2(\theta) f_3(D_a)$$

Where $r_{s,\text{min}}$ is the minimum stomatal resistance, $LAI$ is the leaf area index and $f_1$, $f_2$, $f_3$ are respectively function of the downward shortwave radiation $R_s$, soil moisture $\theta$ and vapour deficit $D_a$.

If $LAI \downarrow$ then $r_c \uparrow$ and $E \downarrow$ so $T_{2m} \uparrow$

If $LAI \uparrow$ then $r_c \downarrow$ and $E \uparrow$ so $T_{2m} \downarrow$
Obtained by the inversion of a 3D radiative transfer model which compute the LAI and FPAR based on the biome type and an atmospherically corrected surface reflectance thanks to a look-up-table derived 8 years (2000-2008) climatological time serie (Jarlan et al. 2008).
Results from forecast experiments using MODIS LAI relative to the fixed LAI case for MAM at FC+36 (valid 12UTC), 2m temperature [K]

Impact of the usage of the MODIS based monthly LAI climatology on the 2m Temperature

\[
\text{Sensitivity (warming)} = T_{\text{MLAI}} - T_{\text{ctl}}
\]

\[
\text{Impact (error reduction)} = |T_{\text{ctl}} - T_{\text{an}}| - |T_{\text{MLAI}} - T_{\text{an}}|
\]
Land carbon/photosynthesis-based parameterisation (CTESSEL)

\[ \text{ISBA-A-g}_s / C-\text{TESSEL} \]

- Met. forcing
- LAI
- Active Biomass
- LE, H, Rn, W, Ts...
- CO₂ Flux

\[ [\text{CO}_2]_{\text{atm}} \]

\[ \text{Nitrogen} \]

\[ \begin{align*}
\text{Carbon flux} & \quad \text{Water vapour flux} \\
\text{Stomatal evaporation} & \quad \text{Stomatal CO}_2 \text{ flux} \\
\text{Cuticular evaporation} & \quad \text{Guard cell} \\
\text{Cuticle} & \quad \text{Cuticular CO}_2 \text{ flux} \\
\text{Intercellular spaces} & \quad \text{Mesophyll cells} \\
\end{align*} \]

\[ g_{sc}^1 = A_n - A_{\text{min}} \left( \frac{D_s}{D_{\text{max}}} \cdot \frac{A_n + R_d}{A_m + R_d} \right) + R_d \left( 1 - \frac{A_n + R_d}{A_m + R_d} \right) \frac{A_n}{C_s - C_i} \]

\[ E = \beta (g_a + g_{sc}) (q_a - q_{sat}) \]

\[ \text{NEE} = A_n - R_{\text{soil}} \]
Soil Respiration parameterization (1)

\[ R_{soil} = R_0 Q_{10}^{(0.1(T_{soil} - 25))} f_{sm} \]

The Berry and Raison (1982) Q10 approach

Q10 represents the proportional increase of a parameter for a 10 degree increase in temperature (Berry and Raison, 1982)

Including a snow attenuation effect on the soil CO2 emission

\[ R_{soil} = R_0 e^{-\alpha Z_{snow}} Q_{10}^{(0.1(T_{soil} - 25))} f_{sm} \]

Including a temperature dependency on the Q10 parameter (McGuire et al., 1992)

Q10 dependance on Temperature regime
Soil Respiration parametrization (2)

Example of NEE (micro moles /m²/s) predicted over the site Fi-Hyy taking the cold process into account (right) and previous simulation (left) by CTESSEL (black line) and observed (red dots).

Feedback from the atmosphere can contribute to improve the physical understanding and adjust the contribution from the surface.
Jarvis Vs photosynthesis-based evapotranspiration

CTESSEL improves the LE/H simulations (Photosynthesis-based vs Jarvis approach).
CO2 exchanges as part of the daily forecasts

Example of Average 10 days forecasted NEE from the 1st of June 2011 extracted from the pre-operational run (e-suites) [micromoles/m²/s] – Operational from November 2011
Impact of climate anomalies on CO2 flux: The summer 2003

- Summer 2003 heat-wave/drought hitting western Europe. The effect on NEE was to turn land into a CO2 source due to vegetation stress conditions, consistently with findings of P. Ciais et al. (2005, Nature)
The 2010 Russian summer heat wave

Impact on 2m T and Rh forecast