Week of 02-04 April

- **Monday**: Biosphere Atmosphere Lab
  - Simsphere: A Java Based Application for a Land-Surface Process Model Prototype

- **Wednesday**: Exam 2
  - In Class this time (Proctored at SDSU)
  - 2-hrs
  - (Lectures 11-17)
Exam 2

- Wednesday: Exam 2
  - In Class this time (Proctored at SDSU)
  - 2-hrs
  - (Lectures 11-17)
    - Numerical Stability through Modeling Reality
Lecture 18
Land Surface Modeling

Modeling Interactions Between the Atmosphere, Biosphere and Hydrosphere
Lecture 18
Land Surface Modeling

Modeling Interactions Between the Atmosphere, Biosphere and Hydrosphere
Agenda

- Exploring the ways in which Biospheric Surface Processes are represented by Physical, Atmospheric and Hydrologic Models
- Introduce key principles in Boundary Layer Meteorology and Land Surface Hydrology
- A formal introduction to “closure”
Our Challenge in NWP, Climate and Environmental Modeling

- We represent a large number of deterministic processes by approximations

- The Land Surface DEFINITELY not exception
Why it matters: The Boundary Layer

- Mixed Layer
- Residual Layer
- Nocturnal BL
- Surface Layer

Height

~1-5 km

~25-50m

Time

Sunset

Sunrise
Profiles of Temp, Humidity and Wind

From TR Oke’s *Boundary Layer Climates*

Figure 2.18  (a) Daily variation of the boundary layer on an ‘ideal’ day. (b) Idealized mean profiles of potential temperature ($\theta$), wind speed ($\bar{u}$) and vapour density ($\bar{\rho}_v$) for the daytime convective boundary layer. (c) Same as (b) for nocturnal stable layer. The arrows in (a) indicate times of sunrise and sunset.
What our atmospheric models need from the land surface

\[
\frac{\partial \bar{u}_i}{\partial t} + \bar{u}_j \frac{\partial \bar{u}_i}{\partial x_j} = -\delta^{i3}g + f\varepsilon_{ij3} \bar{u}_j - \frac{1}{\bar{\rho}} \frac{\partial \bar{p}}{\partial x_i} + \nu \frac{\partial^2 \bar{u}_i}{\partial x_j^2} - \frac{\partial (\bar{u}_j' \bar{u}_i')}{\partial x_j}
\]

\[
\frac{\partial \bar{\theta}}{\partial t} + \bar{u}_j \frac{\partial \bar{\theta}}{\partial x_j} = -\frac{L_v \bar{E}}{\bar{\rho} c_p} - \frac{1}{\bar{\rho} c_p} \frac{\partial \bar{Q}_j}{\partial x_j} + \nu \frac{\partial^2 \bar{\theta}}{\partial x_j^2} - \frac{\partial (\bar{u}_j' \bar{\theta}')}{\partial x_j}
\]

\[
\frac{\partial \bar{q}}{\partial t} + \bar{u}_j \frac{\partial \bar{q}}{\partial x_j} = \frac{E + S}{\bar{\rho}} - \nu \frac{\partial^2 \bar{q}}{\partial x_j^2} - \frac{\partial (\bar{u}_j' \bar{q}')}{\partial x_j}
\]

\[
\frac{\partial \bar{u}_j}{\partial x_j} = 0
\]
What our atmospheric models need from the land surface

- OK. In English.
  - Drag $u'w'$ & $v'w'$
  - The transfer of momentum between the lower layers of the boundary layer and the surface
  - Evapotranspiration $w'q'$
  - The transport of water from the land surface and atmosphere (AKA: Latent Heat Flux, ET, LE, E)
  - Sensible Heat Flux $w'\theta'$
  - The transport of heat between the land surface and atmosphere (AKA: H)
Why this is a challenge

The Problem with Drag.

- Turbulence (which is what drag is), cannot be modeled explicitly to its full extent.

- The more detailed we get in representing our velocity fields, (going from representing the mean velocity to its turbulent perturbations \( u' \) here), the more terms we wind up with that need to be solved.

- We call this growing hole in the water bag “CLOSURE”

\[
\frac{\partial \overline{u_i}}{\partial t} + \overline{u_j} \frac{\partial \overline{u_i}}{\partial x_j} = -\delta_{i3} g + f e_{ij3} \overline{u_j} - \frac{1}{\rho} \frac{\partial p}{\partial x_i} + v \frac{\partial^2 \overline{u_i}}{\partial x_j^2} - \frac{\partial (\overline{u'_i} \overline{u'_j})}{\partial x_j}
\]
Closure Simplified

- Consider our old Lotka-Volterra Eqs
- We have two unknowns
- We have two diagnostics equations for each unknown
- Thus the system is “Closed”
- (you can solve it)

\[ \frac{dH}{dt} = aH - bCH \]
\[ \frac{dC}{dt} = cCH - dC \]
Closure Simplified

- Now add something like variable plants
- We have three unknowns: $C(t)$ & $H(t)$ and $P(t)$
- We still have two prognostic equations
- And we must now have a third for $P(t)$

\[
\frac{dH}{dt} = aPH - bCH
\]
\[
\frac{dC}{dt} = cCH - dC
\]
Closure Simplified

- Now we have a prognostic equation for $P(t)$...

- ... which may depend on $N(t)$ which we can define as nutrient availability or other parameter.

- See how our closure problem grows?

$$\frac{dP}{dt} = gPN - fPH$$

$$\frac{dH}{dt} = aPH - bCH$$

$$\frac{dC}{dt} = cCH - dC$$
Closure Simplified

- We can keep going.
- Or we can do what the original LV equations do (which is “punt” at some level).
- We can represent “N” as a implicit system based on other observables.
  - We call this a “closure scheme.”

\[
\begin{align*}
\frac{dP}{dt} &= gPN - fPH \\
\frac{dH}{dt} &= aPH - bCH \\
\frac{dC}{dt} &= cCH - dC
\end{align*}
\]
Ok, back to momentum

- The problem with Drag
  - Turbulence (which is what drag is), cannot be modeled explicitly (the more you seek its explicit solution, the more complex it gets)
  - The more detail we get into $u'$ the more terms we need
  - This growing hole in the water bag is “closure”

\[
\begin{align*}
\frac{\partial \bar{u}_i}{\partial t} + \bar{u}_j \frac{\partial \bar{u}_i}{\partial x_j} &= -\delta_{i3} \bar{g} + f\varepsilon_{ij3} \bar{u}_j - \frac{1}{\rho} \frac{\partial p}{\partial x_i} + \nu \frac{\partial^2 \bar{u}_i}{\partial x_j^2} - \frac{\partial (u'_i u'_j)}{\partial x_j} \\
\end{align*}
\]
One way to fix Drag

- One way to do this is to rely on some empirical principles backed up with physical reasoning!
- All we need to do is represent the transport of momentum between the ground and lowest ATM grid cell.

\[
\frac{\partial \overline{U}}{\partial t}_{\text{drag}} = - \frac{\partial (u_j u_i')}{\partial x_j}
\]
The Log-Wind Law in a nutshell

- To the right is a vertical profile of mean observable wind speed (not turbulent wind speed).
- The blue represents what’s in the ATM model.
- Now, we know that wind decreases as you approach the ground in the surface layer.

\[
\frac{\partial \bar{U}}{\partial t} \bigg|_{\text{drag}} = -\frac{\partial (u'_i u'_i)}{\partial x_j}
\]
The Log-Wind Law in a nutshell

- The blue represents what’s in the ATM model, the red is what the LSM or BLM handles.

- Near the interface between the surface layer, we should be able to see a mean wind speed with height as a slope
The Log-Wind Law in a nutshell

- Now the more turbulence in the region (i.e., the more drag), the more the mean profile’s wind will slow down.

- The measure of this mixing will also be key to our evaporation and heat fluxes.
The Log-Wind Law in a nutshell

Now with a Log!

- For a neutral atmosphere (where a parcel neither rises nor falls when pushed up or down), our mean wind profile approximates a straight line when plotted in LogZ-\( \bar{U} \) space.

- Note that this is a diagnostic approximation!

- This is good! We can do this inside of a model time step!

\[
\frac{\partial \bar{U}}{\partial \ln z} = \text{Const}
\]
The Log-Wind Law in a nutshell
Now with a Log!

- Notice also that the mean wind speed hits zero before it hits the ground

- Two reasons:
  - Obstacles (roughness)
  - Turbulence winds exceed the mean wind speed down below.

\[
\frac{\partial \bar{U}}{\partial \ln z} = \text{Const}
\]
The Log-Wind Law in a nutshell
Now with a Log!

- If we put these together, we can create an obstacle scale height for surface obstacles (aka $z_o$, the “roughness height”), and a scaled turbulent velocity, which represents the turbulent velocity ($u^*$) or the effective strength of the turbulent eddies at this height.

\[
\frac{\partial \bar{U}}{\partial \ln z} = \text{Const}
\]
The Log-Wind Law in a nutshell
Now with a Log!

If we put these together, we can create an obstacle scale height for surface obstacles (aka \( z_o \), the “roughness height”, and scaled turbulent velocity, which represents the turbulent velocity (\( u^* \)) or the effective strength of the turbulent eddies at this height.

\[
\frac{\bar{U}}{u^*} = \frac{1}{k} \ln \left( \frac{Z}{z_o} \right)
\]

\( k \sim 0.4 \)

“von Karman Constant”
So after all these slides, where is the payoff?

- Well... we want $u'w'$ (which we get from $u^*$) for our atmospheric calculations.
- Our atmospheric component gives us $U$ at the lowest layer ($z_{atm}$)
- All we need is a “roughness” which is an idealized value and is available through community ascribed lookup tables.

\[
\frac{\bar{U}}{u_*} = \frac{1}{k} \ln \left( \frac{Z}{Z_o} \right)
\]

$k \approx 0.4$

“von Karman Constant"
Sample values for roughness height

- Ice Flats: $10^{-5}$ m
- Calm Sea: $10^{-4}$ m
- Grassly Plain: $10^{-2}$ m
- Forest: 1 m

These values tend to vary from publication to publication.

\[
\left(\overline{u'w'}\right)^2 = u_* = \bar{U} \left[\frac{1}{k} \ln \left(\frac{Z}{Z_o}\right)\right]^{-1}
\]
And for Stable and Unstable conditions?

- For stable conditions, large scale sinking air tends to increase the wind speed near the surface (reducing turbulence).

- For unstable conditions, the increased lifting environment increases turbulence and slower winds near the roughness height.

\[
\frac{\bar{U}}{u_*} = \frac{1}{k} \ln \left( \frac{z}{z_o} + \phi \right)
\]
How do you use the Log Wind Law?

- Where have you used this before?
- Any time you’ve used an aerodynamic method to calculate evaporation.
- Because the more turbulence the more evaporation!

**WHY?**

\[
Evap = \frac{0.102 U^2 [e_s(T) - e]}{\ln(z/Z_o)^2}
\]

Chow, Maidment & Mays, *Applied Hydrology* (one of my old texts)

---

28 March 2007

ATM 515: Intro Env Modeling : Land Surface Modeling

28
So… Notice that we have a convinient segue into…

Surface Energy and Mass Budgets!
Conservational Budgets

Energy, Momentum and Mass Budgets
“should” balance
So… from our first Global LSP with a few… “enhancements.”

- Energy In = Energy Out
- Energy In
  - Absorbed Insolation \[S(1-a)\]
- Energy Out
  - Outgoing Radiation
    - Net Radiation \(Rn = S(1-a)\)
    - ADD Departing Evapotranspiration (LE)
    - ADD Departing Sensible Heat Flux (H)
    - ADD Departing Ground Flux (G) to lower layers in the ground and soil
Or in Cartoons

\[ R_n = S(1-a) - L = LE + H + G \]
The Diurnal Energy Budget

Figure 1.10 (a) Energy balance components for 30 May 1978 with cloudless skies at Agassiz, B.C. (49°N) for a moist, bare soil, and (b) temperatures at the surface, in the air at a height of 1.2 m and in the soil at a depth of 0.2 m (after Novak and Black, 1985). The following table gives the energy totals for the day (MJ m⁻² day⁻¹).

<table>
<thead>
<tr>
<th>Energy balance</th>
<th>Derived terms</th>
</tr>
</thead>
<tbody>
<tr>
<td>( Q^* )</td>
<td>( \beta )</td>
</tr>
<tr>
<td>( Q_{11} )</td>
<td>( Q_e/Q^* )</td>
</tr>
<tr>
<td>( Q_E )</td>
<td>( E(\text{mm}) )</td>
</tr>
<tr>
<td>( Q_G )</td>
<td></td>
</tr>
</tbody>
</table>

From TR Oke’s *Boundary Layer Climates*
The Meteorologist’s Migraine

- Given all of the parameterizations we have in the atmosphere, we are often frustrated with representing the land surface (which happens to be where our clients spend most of their time).
- We want to be as true as possible to physical principles as possible.
- But we cannot represent many of these complex biological processes explicitly.
- Therefore we must grossly simplify the process while getting a bulk estimate that is somewhat workable.
Sensible Heat Flux (H)

- Fortunately, this is fairly simple, compared to the other parameters we have.
  - Fluxes tend to go “down gradient.”
  - From High Temperatures to Low Temperatures
  - A Warmer Surface than Air in the presence of turbulence sends H upward.

\[ H = f(U, z_o) \left[ T_s - T_a \right] \]
Sensible Heat Flux (H)

\[ H = f(U, z_o) [T_s - T_a] \]

But we use an alternative analogy representation (which is where you probably should have been paying attention in Physics 2)
Sensible Heat Flux (H)

The resistance analogy

\[ H = f(U, z_o)[T_s - T_a] \]

\[ I = \frac{\Delta V}{R} \]

\[ H = \rho_a c_p \left( \overline{w'\theta'} \right) \]

\[ H = \rho_a c_p \frac{\Delta T}{R(u_*)} \]

\[ H = g(u_*) \Delta T \]
Ground Flux (G)

Conceptually, Ground Flux is the transfer of heat from the surface layer into deeper ground reservoirs (deep soil layers).

\[ G = K_{grnd} \frac{\Delta T}{\Delta z} \]

(We often solve for G as a “residual”
\[ Rn - LE - H = G \]
Free Evaporation (LE)

- Evaporation from open sources of water is fairly simple compared to that from a contained medium (soil or canopy)

\[ LE = L \rho_w \left( \overline{w'q'} \right) \]
\[ LE = L \rho_w \frac{q_s(T_s) - q_a}{R(u_*)} \]
\[ LE \approx PE \]
Representing Evapotranspiration (LE)

- When extracting water from a mitigating surface, things get more interesting.

\[ \text{PE} = L \rho_w \frac{q_s(Ts) - q_a}{R(u_*)} \]

\[ LE = L \rho_w \frac{q_{sfc}(Ts) - q_a}{R(u_*)} \]

\[ LE = L \rho_w \beta(W) \frac{q_s(Ts) - q_a}{R(u_*)} \]

\[ LE = L \rho_w \frac{\alpha(W)q_s(Ts) - q_a}{R(u_*)} \]

\[ LE = L \rho_w \frac{q_s(Ts) - q_a}{R(u_*) + R_{sfc}} \]
Representing Evapotranspiration (LE)

- **Land Surface Factors limiting LE from PE**
  - **Bare Soil**
    - Surface Soil Moisture
    - Soil Textural Properties
      - Adherence of water to a soil particle
  - **Vegetation**
    - Root Zone Moisture
    - Vegetation Type
    - Leaf Temperature
Enter Land Surface Parameterizations

- In the beginning, we represented the surface in the simplest way possible.

- We didn’t

- We simply paved it over.
The First LSP

- “The Bucket”
  - Manabe (1969)
  - A fixed capacity “soil” zone in a bucket, where if stored water is above a given threshold, the excess “runs off” – from the soil bucket to the bit bucket.
The First LSP

- “The Bucket”
  - Evaporation Rates are calculated as a function of its capacity

\[
LE = L \rho_w \left[ \min \left( \frac{W}{W_{fc}}, 1 \right) \right] PE \\
LE = L \rho_w \left[ \min \left( \frac{W}{W_{fc}}, 1 \right) \right] \frac{q_s(T_s) - q_a}{R(u_*)}
\]

S = P - E - RO
The First LSP

- “The Bucket”
  - This became the core template for most LSPs

\[ LE = L \rho_w \left[ \min \left( \frac{W}{W_{fc}}, 1 \right) \right] PE \]

\[ LE = L \rho_w \left[ \min \left( \frac{W}{W_{fc}}, 1 \right) \right] \frac{q_s(T_s) - q_a}{R(u_*)} \]
More Biophysical Parameterizations

- After the bucket came a continuing series of increasingly complex models
  - Deardorff (with a simple soil model)
  - BATS (Biosphere Atmosphere Transfer Scheme)
  - SiB (Simplified Biosphere Scheme)
    - SiB2 (SiB V2)
    - SSiB (Simplified SiB) – they call these models “SiBlings”
  - LSM and CLM
  - NOAH (used in WRF and MM5)
  - And more!
The contemporary LSM Setup

- These more advanced models are also called SVATS (Soil-Vegetation-Atmosphere-Transfer Schemes)
  - Distributed Soil Zone (multilayer)
    - Temperature
    - Soil Moisture Hydrology
  - Simple to Complex Vegetation Systems
    - Plant Water Capacity
    - Leaf Temperature
    - Seasonal Phenology
Water and Energy Balance in a Canopy

- Evapotranspiration
- Canopy Temp
- Stomatal Aperture
- Leaf Water Potential
- Soil Water Potential
- Root Uptake
- Root Water Potential
- Sensible Heat Flux
- Turbulent Flux
- Upward Stem Transport
- PBL Temp

28 March 2007
ATM 515: Intro Env Modeling : Land Surface Modeling
Next Time

- Exploring SIMSPHERE (the graphical front end for PSUBAMS)

- Alfred Blackadar’s classic boundary layer model
  - (Downloading to windows needed for SDSU)