HEAT BALANCE
Energy and Temperature

• When an object absorbs short-wave energy, its temperature increases
• As the temperature increases, it emits more long-wave radiation out to space
• This process will continue until the rate of absorption of short-wave radiation equals the emission rate of long-wave radiation
The Earth's Heat Budget

Incoming solar radiation at top of atmosphere: 7 million calories per square meter per day, averaged for the Earth as a whole.

Outgoing radiation:

Light (shortwave radiation):
- Space: 100%
- Absorbed by water vapor, dust, CO2: 16%
- Back-scattered by air: 6%
- Reflected by clouds: 20%
- Reflected by water and land surface: 4%

Infrared (long-wave) radiation:
- Space: 38%
- Emission by water vapor, CO2: 26%
- Net surface emission of long-wave radiation: 21%
- Sensible heat: 7%
- Latent heat: 23%

Daily/Seasonal Radiation Patterns

- insolation peak vs. temperature
  - daily lag
  - seasonal lag
  - Lag is function of type of surface, wetness, wind, etc

- Temperature increases when input > output

- Temperature decreases when input < output
The Surface Energy Budget

• Determines amount of energy available for evaporation of surface water and to change the surface temperature
• Contributes to overall energy balance of Earth
• More complex than top of the atmosphere energy budget
  – Latent and Sensible heat fluxes
• Depends on insolation (incoming solar), surface characteristics, atmosphere characteristics
No two places on Earth have exactly the same surface.

Differences in surfaces will result in differences in local climate.
Surface Energy Budget

There is an imbalance in the energy budget with just the radiation fluxes present \((R_N)\)

Non-radiative fluxes (transfer of energy other than radiation) need to be included:

- Sensible Heat Flux \((H)\)
- Latent Heat Flux \((LE)\)
- Ground Heat Flux \((G)\)
Surface Energy Budget

\[ \Delta = R_N - (H + LE + G) \]

The \( R_N \) term involves only radiation.

\( H + LE + G \) represents the nonradiative fluxes.

On the globally averaged scale, \( \Delta = 0 \), which represents a planetary energy balance.

\[ R_N = H + LE + G \]
Seasonal variation of surface energy budget
Net radiometer (measure net radiation $R_n$)

Piranometer (measure shortwave radiation (sun))
Figure 9. View of the Net Radiation observed and the Net Radiation Estimated for two days.
Net Radiation $z = 2.5 \text{ m}$
SENSIBLE HEAT FLUX
Estimation of energy components …

- Sensible heat flow ($H$) is a diffusive process driven by turbulence (eddies) in the wind

$$H = \rho_a c_p K_H \left( \frac{T_s - T_a}{\Delta h} \right)$$

- $\rho_a =$ air density (kg m$^{-3}$)
- $c_p =$ specific heat of air (1005 J kg$^{-1}$deg$^{-1}$)
- $u =$ wind speed (m s$^{-1}$)
- $T_a =$ air temperature at height $\Delta h$
- $K_H =$ eddy diffusivity of heat
  - Depends on wind speed, roughness, atmospheric stability
  - Sometimes written as $K_H u$
Sensible Heat flux Bulk Aerodynamic Formula

\[ Q_H = c_p \rho C_{DH} U_r (T_s - T_a(z_r)) \]

- Wind speed at reference level
- Surface temperature
- Temperature of air at reference height \( z \)
- Specific heat (const. \( p \))
- Transfer Coeff. (temperature)
SENSIBLE HEAT FLUX

a) Inundated tidal flats (Zaker, 2003)

\[ Q_H = \rho c_p V C_h (T_w - T_a) \]

\( \rho = \text{air density (kgm}^{-3}) \)

\( c_p = \text{specific heat capacity (Jkg}^{-1} \circ C^{-1}) \)

\( V = (m / s) \)

\( C_h = 0.91 \times 10^{-3} \) (Friehe and Smitt, 1976)
b) Exposed tidal flats (Evett, 1994, 2002)

\[ Q_H = \rho c_p D_h (T_s - T_a) \]

\[ \rho = \text{air density (kgm}^{-3}\text{)} \]
\[ c_p = \text{specific heat capacity (Jkg}^{-1}\text{oC}^{-1}\text{)} \]

\[ D_h = k^2 V \left[ \ln \left( \frac{z}{z_{oh}} \right) \right]^2 \]

\[ k = \text{Von Karman cte} \]
\[ V = (m/s) \]
\[ z = \text{reference height} \]

\[ Dh = \text{heat exchange coefficient [m s}^{-1}\text{]} \]
LATENT HEAT FLUX
Latent Heat of Vaporization

wind speed at reference level

Specific Humidity

Transfer Coeff. (humidity)

\[ LE = L \rho C_{DE} U_r \left( q_s - q_a(z_r) \right) \]
Vapor pressure gradient

Lake surface

Saturated air
Vapor Pressure (e)

- Vapor pressure (e) is simply the amount of pressure exerted **only** by the water vapor in the air.

- The pressures exerted by all the other gases are not considered.

- The unit for vapor pressure will be in units of pressure (millibars and hectoPascals are the same value with a different name).
Saturation water vapor pressure

\[ e_s = 0.6108 \left[ \exp \left( \frac{17.27 \ T_a}{T_a + 237.3} \right) \right] \] [kPa]

Water vapor pressure \( e = (\text{RH} \ e_s) / 100 \)

Specific humidity

\[ q = \frac{0.622 \ e}{(P - 0.378 \ e)} \]

(Allen et al. 1998, Evett 2002)
SOIL HEAT FLUX
Ground Thermal Variations
Ground Temperatures Variations

- exponential decrease of diurnal wave amplitude \( A \) with depth
- phase shift of wave with depth
- For typical soil, \( A \) becomes insignificant below 1 m.

Source: The Hydrology Technical Group, Pacific Northwest National Laboratory (PNNL), Richland, Washington
Soil heat flux

• Primarily by conduction, i.e. no convection
• Found to be proportional to temperature gradient and thermal conductivity ($\lambda$).

\[ G = -\lambda \left( \frac{dT}{dz} \right) \]

• $\lambda$ - the **thermal conductivity**, is the amount of heat transferred through a unit area in unit time under a unit temperature gradient. For soil it depends upon its mineral composition and OM content, as well as on the volume fractions of water and air.

• For same porosity, $\lambda$ increases with an increase in volume wetness for a given soil. However, for same porosity and wetness: $\lambda$(sand) > $\lambda$ (clay).
Thermal diffusivity ($\kappa$)

\[ \lambda = \kappa C \]

$C$ = heat capacity = specific heat density = $c \rho$

$\kappa$ – thermal diffusivity (change in temperature produced in a unit volume by the quantity of heat flowing through the volume in unit time under a unit temperature gradient)

\[ K_s = \frac{(\Delta z/\Delta t)^2}{2\omega} \]
Thermal conductivity ($\lambda$). Thermal conductivity values were 1 W m$^{-1}$ K$^{-1}$ on 20 February and 0.79 W m$^{-1}$ K$^{-1}$ on 1 July. According to the granulometry of the tidal flat sediments (table 1), heat capacity was estimated as a mean value of saturated clay and saturated sand heat capacities (Oke, 1978), resulting in $3.03 \times 10^6$ J m$^{-3}$ K$^{-1}$. Thermal diffusivity, which varies with periodic changes of temperature and soil moisture content, was estimated at $3.31 \times 10^{-7}$ m$^2$ s$^{-1}$ and $2.61 \times 10^{-7}$ m$^2$ s$^{-1}$ for 20 February and 1 July, respectively. This range of values was similar to those estimated previously by Piccolo and Dávila (1991) in Ingeniero White: $5.7 \times 10^{-7}$ m$^2$ s$^{-1}$ in summer and $3.5 \times 10^{-7}$ m$^2$ s$^{-1}$ in winter (monthly mean values).