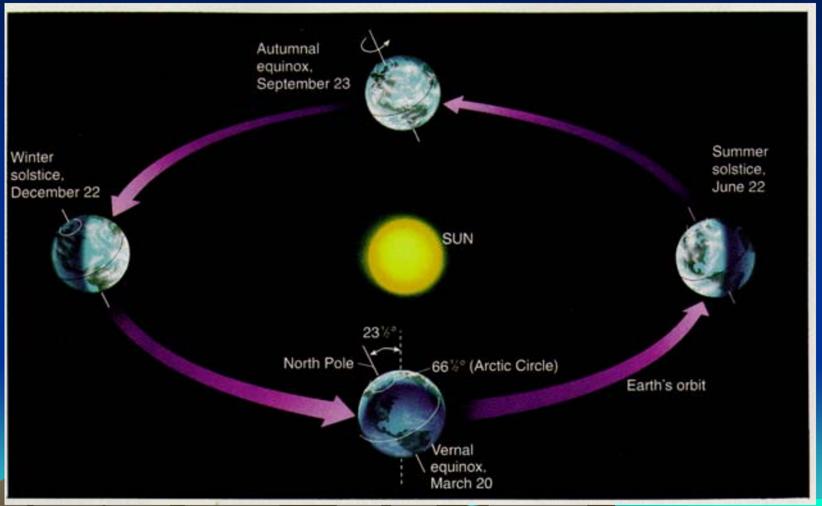
Energy Budget

- Solar energy drives the dynamics and thermodynamics of the atmosphere
- Energy received from sun balanced by energy emitted by earth system
  - Otherwise, earth would warm/cool at alarming rate
- Energy flux: rate at which energy is transferred across a surface (per unit area). Units: W/m^2
- Solar flux is greater at high altitude because less atmosphere present to absorb/scatter solar radiation

Revolution of Earth around sun



N Hem terminology

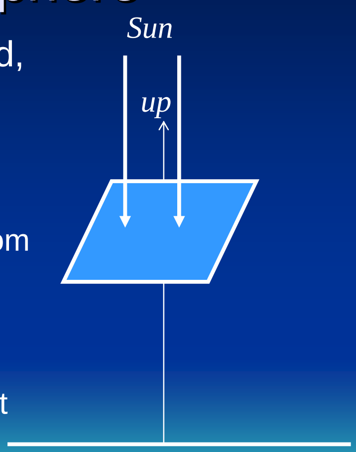
Ahrens (1994)

Earth Orbit: Animations

- http://www.ncsu.edu/scivis/lessons/earthinspace3d/eSpace.avi
- http://www.classzone.com/books/earth_science/terc/content/visualizations/es0408/es0408page01.cfm?chapter_no=04

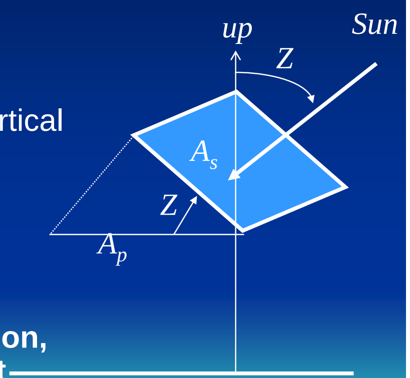
Solar Radiation At the Top of the Atmosphere

- When sun directly overhead, amount of solar radiation received is:
- $Q = S\,(\underline{d}/d)^2$
  - Where $\underline{d}$ – mean distance from the sun
  - d – distance from sun on particular day
  - d largest in January; smallest in July by ~ 5%



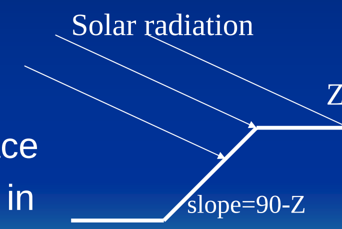
Zenith Angle

- When sun is at some other zenith angle, Z (angular distance of sun from local vertical), how much solar radiation is received?
- A_s- Area perpendicular to Sun
- A_p- Area perpendicular to local vertical
- $Q_s A_s = Q_p A_p$; $A_s = A_p \cos Z$
- $Q_p = S (\underline{d}/d)^2 \cos Z$
- Z=0 cos Z = 1; Z=45 cos Z= .71
- Z=60 cos Z = .5; Z=90 cos Z= 0
- **As sun becomes lower on horizon, amount of solar energy on a flat surface decreases per unit area**



Impact of Terrain

- If slope of terrain is 90º – Z, then area on slope is perpendicular to incoming solar radiation
- South facing slopes receive more solar radiation at low zenith angles than flat surface
- North facing slopes may be in shade (no direct solar flux)



Zenith Angle

- Zenith angle depends on:
  - latitude (φ north +, south -)
  - time of year. Solar declination angle, δ - angular distance of the sun north (+) or south (-) of the equator
    - δ = 23.5º on June 21 Summer Solstice
    - δ = -23.5º on Dec 22 Winter Solstice
    - δ = 0º on Sept 23 (Autumnal Equinox) and March 20 (Vernal Equinox)
  - Time of day. Hour angle, h, is 0 at solar noon, when sun is directly north or south of observation point
    - h increases by 15 for every hour before or after solar noon
    - h = -90º at 0600 LT and h = 90º at 1800 LT

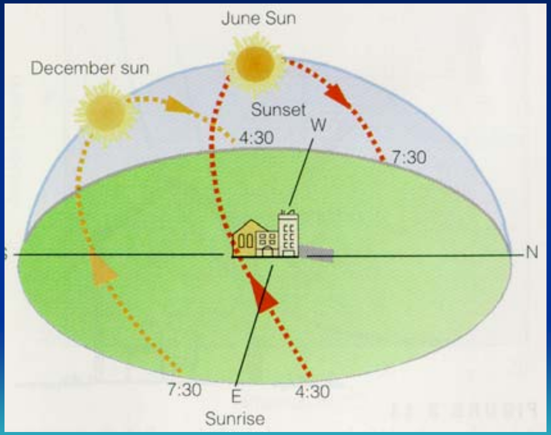
Zenith Angle

- $\cos Z = \sin \varphi \sin \delta + \cos \varphi \cos \delta \cos h$
- Special cases:
  - Solar noon: cos h = 1. $Z = \varphi - \delta$
    - On June 21, sun directly overhead at 23.5ºN
    - On the same day at 40ºN, Z = 40 – 23.5º = 17.5º
    - At 90ºN, Z = 67.5º

Zenith Angle

- $\cos Z = \sin \varphi \sin \delta + \cos \varphi \cos \delta \cos h$
- Special cases:
  - At sunrise or sunset, $\cos Z = 0$, $h = H$ = half-day length
  - $\cos H = - \tan \varphi \tan \delta$
  - At the equator, $\tan \varphi = 0$ so $H = 90^o$ which is 6 hours
  - On the equinoxes, $\tan \delta = 0$ so $H = 90^o$ or 6 hours
  - Latitude of polar night ($H=0$), $90^o - |\delta|$ in winter hemisphere

Sun path, summer and winter



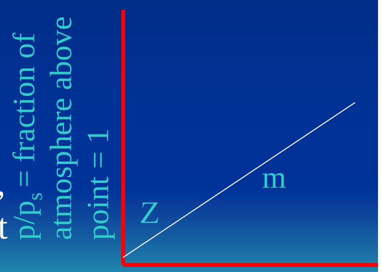
Northern Hemisphere

Ahrens (1994)

Transmission of Solar Radiation Through Atmosphere

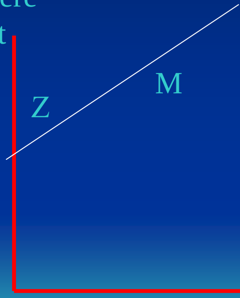
•m = non-dimensional optical depth at surface $m = 1/\cos(Z)$

- Greater optical depth implies more scattering and absorption of solar energy by atmosphere
- $m = 1$, $Z = 0^o$
- $m = 2$, $Z = 60^o$
- $m = 4$, $Z = 76^o$
- When sun at zenith angle of 76^o, light travelling through equivalent of 4 atmospheres



Effect of Elevation on Available Solar Radiation

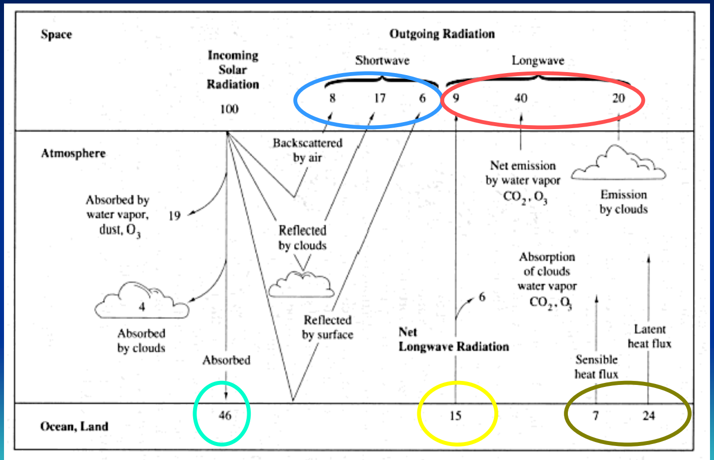
- $M = m\, p/p_s$ = non-dimensional optical depth at pressure p
- At 500 mb ($p/p_s = .5$):
  - for zenith angle = 0^o, half as much atmosphere
  - for zenith angle = 76^o, equivalent to path length at sea level for zenith angle of 60^o
  - So, more solar radiation at higher elevation



Solar Radiation- Ideal Atmosphere

Pressure (mb)	M=1; zenith angle= 0^o	M=2; zenith angle= 60^o
0 mb	1370 W/m^2	1359 W/m^2
500 mb	1299 W/m^2	1238 W/m^2
1000 mb	1244 W/m^2	1146 W/m^2
500-1000 mb	Diff = 55 W/m^2	92 W/m^2

Global Energy Balance



Definitions

- *Not only for global average, we can determine the transfer of energy for any location*
- *The surface radiation budget*: The radiation budget at the earth's surface, considered in terms of the fluxes through a plane at the earth-atmosphere interface. The radiation budget includes both solar (shortwave) and terrestrial (longwave) radiation fluxes.
- *The surface energy budget:* The energy or heat budget at the earth's surface, considered in terms of the fluxes through a plane at the earth-atmosphere interface. The energy budget includes net all-wave radiation, and sensible, latent, and ground heat fluxes.

Surface Energy Budget Measurements

Whiteman photo

Radiation measurements, Hardheim forest (Freiburg Met. Inst.)

UL and LL: CM21 pyranometer

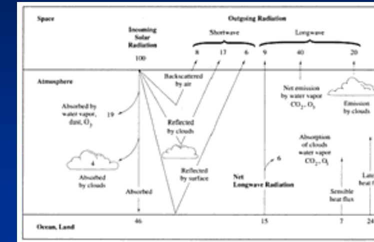
UR and LR: CG1Schultze pyrgeometer

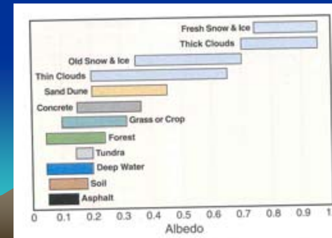
All 4 have upper and lower domes

Whiteman photo

Absorbed Solar Radiation

- Absorbed solar radiation depends strongly on albedo α
- α = for globe, blue/incoming Absorbed Solar = Solar $(1 - \alpha)$
  - Snow cover reflects solar radiation and diminishes absorbed solar radiation
  - Annually, snow cover at high elevation later in season than in valleys tends to cause absorbed solar radiation to diminish with elevation





Whiteman (2000)

Solar radiation reduced in lower atmosphere by:

--absorption of radiation by water vapor

-- Scattering by aerosols

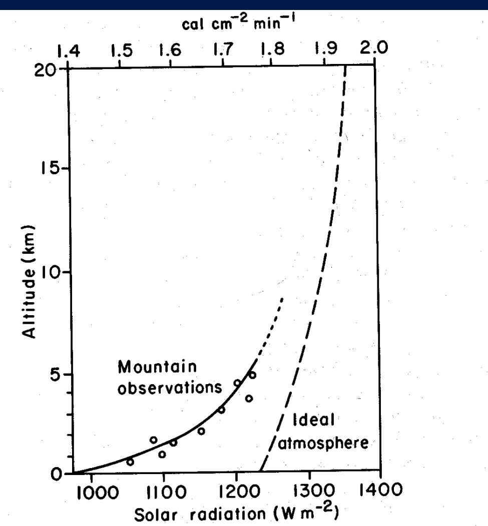
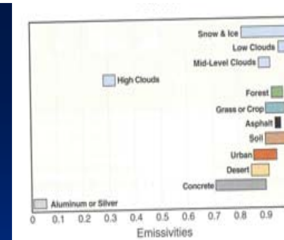


Figure 2.5 Direct solar radiation versus altitude in an ideal atmosphere for m = 1 (after Kastrov, in Kondratyev, 1969, p. 262) and as observed at mountain stations. (Based on Abetti, 1957, Kimball, 1927, and Pope, 1957.)

Outgoing Infrared Radiation



Whiteman (2000)

- Stefan-Boltzmann Law: IR = $\varepsilon\sigma T^4$
- As elevation increases:
  - temperature decreases, IR radiation decreases
  - Optical thickness of atmosphere decreases (less greenhouse gases, including water vapor), atmospheric transparency to outgoing radiation increases, more IR escapes
- Emissivity (ε) of snow and ice is high

Infrared Radiation: December, Austrian Alps

Altitude (m)	Bare ground W/m^2	Snow cover W/m^2
200	289	301
1000	270	287
2000	255	274
3000	240	255

Barry (1992)

Incoming (to surface) Longwave Radiation Function of temperature of atmosphere

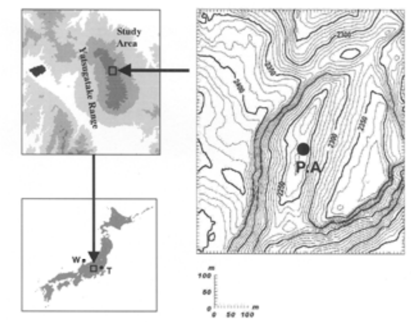


Fig. 1. Topographical map showing study site with the location of the observation point (solid circle).
The point P.A is located at the bottom of the hollow. The maps on the left show the location of the study area (square) and the location of the upper-air observation stations at Tateno (T) and Wajima (W)

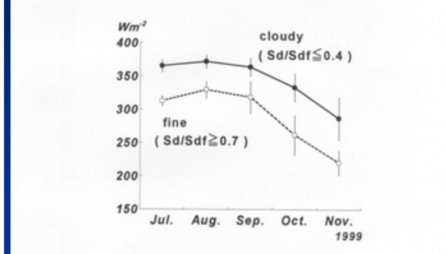
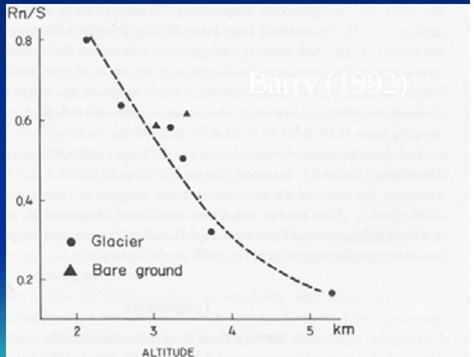


Fig. 3. Monthly averaged downward longwave radiation measured under fine and cloudy weather conditions taken from July to November 1999. Bars indicate standard deviation.

Iijima and Shinoda (2002)

Net Radiation

- Net Radiation = Absorbed solar radiation + downwelling IR – IR
- Effect of altitude on net radiation is variable and depends most strongly on decrease with height of absorbed solar radiation
- In this case, albedo effect of ice dominates



ratio of net radiation (R_n) to solar radiation (S) versus height in the Caucasus in summer

Barry (1992)

Surface radiation budget

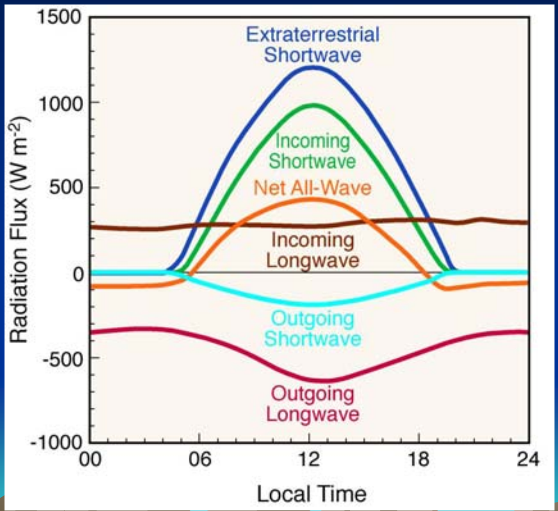


Figure 4.8.

Typical diurnal evolution of the four components of R at the earth's surface under cloudless conditions on a June day at a semiarid site near the Columbia River in eastern Oregon. The extraterrestrial shortwave radiation is the theoretical radiation that would be received outside the earth's atmosphere on a plane that is parallel to the earth's surface below.

Whiteman (2000)

Surface Energy Budget

- Source + sinks = 0 or R = - (G + H + L) (Watts/m^2)
- R = net solar and terrestrial radiation at the earth's surface (+ into surface). Contributions from:
  - Absorbed solar radiation (incoming – reflected)
  - Incoming longwave radiation emitted by gases and clouds in the atmosphere
  - Outgoing longwave radiation emitted by the earth's surface
- G- storage of energy into the deep soil (- into soil)
- H – heat flux into atmosphere(- into atm)
- L – Latent heat (evaporation) flux into atmosphere (- into atm)

Surface Energy Budget

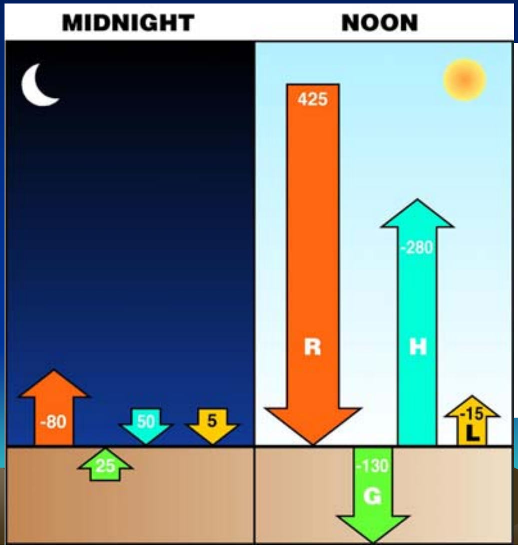


Figure 4.10 Typical values of the surface energy budget components at midnight and noon. The numerical values shown are from figure 4.9.

Whiteman (2000)

Heat Storage

	Specific heat (j/(kg K)	Density (kg/m3)	Heat storage per unit volume per K
Soil inorganic	733	2600	$1.9x10^6$
Soil organic	1921	1300	$2.5x10^6$
Water	4182	1000	$4.2x10^6$
Air	1004	1.2	$1.2x10^3$

Hartmann (1994)

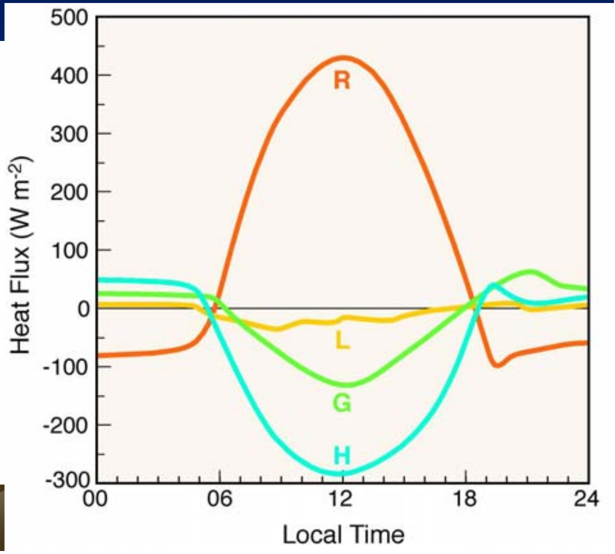
Soil Temperature

- high intensity of solar radiation at high altitudes can result in high surface temperature
- Austria- 80C humus at 2070 m; air temp at 2 m 30C
- New Guinea 60C at 3480 m air temp 15C

Barry (1992)

Surface Energy Budget

igure 4.9 The surface energy budget erms vary diurnally. The data shown re for a June day at a semiarid site near the Columbia River in eastern Dregon. The net radiation curve in this igure is repeated from figure 4.8, but s plotted on a different scale.



Whiteman (2000)