Lecture 7. 8 Feb 2006

Surface temperature (1)

References:

Martin, S., 2004, An Introduction to Ocean Remote Sensing, Cambridge University Press, 454 pp. Chapter 7.

Robinson, I. S., 2004, Measuring the Oceans from Space: The principles and methods of satellite oceanography, Springer Praxis Books/Geophysical Sciences, 669 pp. Chapter 7.

Physical Oceanography (PO) Distributed Active Archive Center (DAAC) at the Jet Propulsion Laboratory http://podaac.jpl.nasa.gov/sst/sst_links.html

Satellite SST observations contribute to an understanding of ocean circulation and climate variability:

- The upper 3 m of the ocean has about the same heat capacity as the entire overlying atmosphere

- The upper 10 m of the ocean has about the same mass as the overlying atmosphere

- SST plays a fundamental role in setting the air-sea fluxes of heat and water vapor between the atmosphere and ocean – essential to weather and climate

- Fine scales in SST patterns visualize ocean currents:
  fronts, jets, eddies, upwelling, El Nino, global change

- Satellite observed SST is used operationally for weather forecasting, ship routing, fishing, and climate variability prediction.

Satellite SST data have been gathered using passive infrared instruments since the 1980s.

Lecture outline:

- IR physics

- Thermal behavior of the ocean

- Atmospheric effects and calibration

- Satellite instruments and systems

Passive infrared (IR) observations of SST commenced with the *Advanced Very High Resolution Radiometer* (AVHRR) on the NOAA-7 polar orbiter (1981)

Accuracy of the AVHRR instrument is about ± 0.4 K for any single datum. Greater accuracy is possible with spatial and temporal averaging and consistency error checking.

AVHRR instruments have been deployed on all NOAA satellites (now up to NOAA-19) and more accurate instruments (MODIS instrument) have been deployed on Terra (AM) and Aqua (PM) satellites and European Space Agency satellites (ATSR and AATSR).

Atmospheric *sounding* instruments give information about the vertical profile of temperature within the atmosphere.

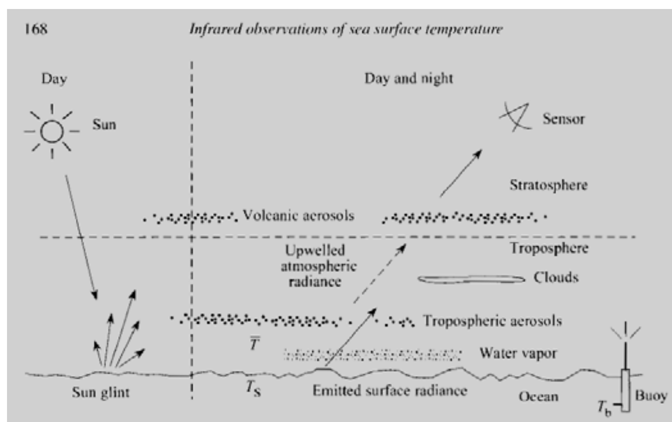
Satellites cannot observe the oceanic vertical profile of temperature.

<u>What does and IR radiometer see?</u>

Infrared wavelength EM radiation reaching the satellite at the top of the atmosphere, coming *from the direction* of the ocean

- affected by atmosphere
    - sources
    - scattering
    - absorption
    - clouds obscure ocean surface

- emission by ocean depends on temperature and ocean physics



(*Martin*)

Figure 7.2. Schematic drawing of the radiances and atmospheric and oceanic properties involved in the infrared SST retrieval. See text for further description.

Planck's radiation law gives the thermal emission as a function of wavelength for a body of a given temperature. Integrating over wavelength we get the total exitance, M:

$M = \sigma T^4$ where

- σ is the Stefan-Boltzmann constant = 5.669×10^{-8} W m^{-2} K^{-4}

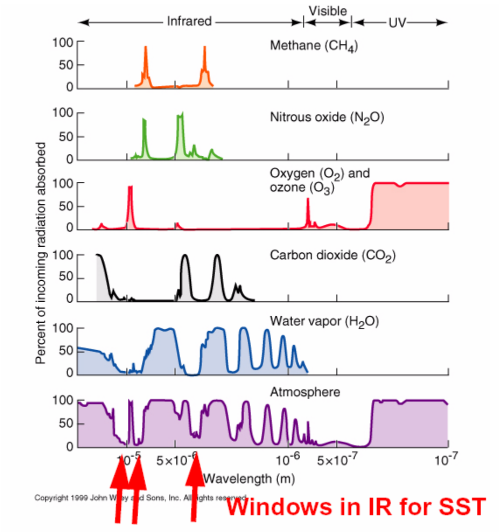
- T is the temperature in Kelvin

The emissivity of water is very close to 1, so M is very close to the exitance for ocean water of temperature T (in K).

The spectral peak of Planck's law occurs at the wavelength given by *Wien's Displacement Law*:

$$\lambda_{max} T = C_3 \quad = 2897 \ \mu m \ K^{-1}$$

A consequence of this is that ocean surface temperatures in the range 0°C to 40°C have an emission peak between 9 μm and 11 μm, with the radiation being spread over the range of about 4 μm to 20 μm.



An IR radiometer measures the *brightness temperature*, defined as the temperature of the blackbody that would emit the measured radiance, at a set of discrete wavelengths.

The actual temperature requires a correction for the true emissivity, and the intervening atmosphere.

Atmospheric absorption restricts IR radiometry of SST to spectral windows in the range of 3.5 to 4.1 μm and 10.0 to 12.5 μm.

During the day, reflection of light can influence the signal, and shallow heating of the ocean surface in a thin layer can mask the underlying mixed layer temperature of oceanographic interest.

Night-time IR SST data is generally a more reliable indicator of the ocean temperature than day-time data.

Received radiance is a combination of the surface-emitted radiance, and the atmospheric upwelled radiance (which depends on water vapor and aerosols, and atmosphere temperature).

Clouds can be opaque or thin, so cloud identification is an important step. Clouds are always significantly colder than the ocean surface.

What is SST?

Factors that affect water temperature near the ocean surface include:

- upper ocean heats up during the day and cools at night from radiation
- day/night heating and radiative cooling affects the difference between T_s and T_b within the mixed layer. ΔT can be ~1°
- air-sea heat exchange occurs through sensible (conduction) and latent (evaporation) fluxes
- wind speed and wave breaking affect these fluxes and the rate of mixing of heat through the ocean and atmosphere boundary layers

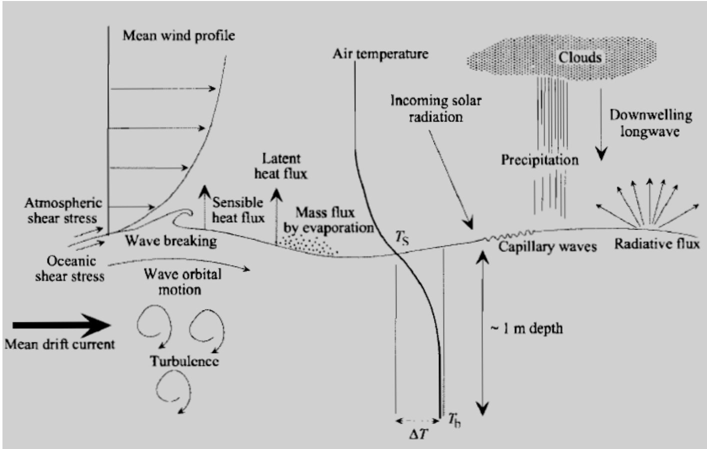


Figure 7.3. The factors that determine the surface temperature and the upper ocean interior temperatures, where ΔT is the temperature difference between the bulk and skin temperatures (Adapted from Figure 1 of Katsaros, 1980).

- net shortwave and longwave radiation heat/cool the ocean

But what temperature determines the upwelling radiance?

The absorption coefficient (an inverse length scale) for EM radiation varies with wavelength.

It is smallest for blue visible light, and increases rapidly to very high values ~10^7 m^{-1} for infrared.

This means net infrared exitance at 11 μm wavelength comes only from the water within 30 μm of the sea surface.

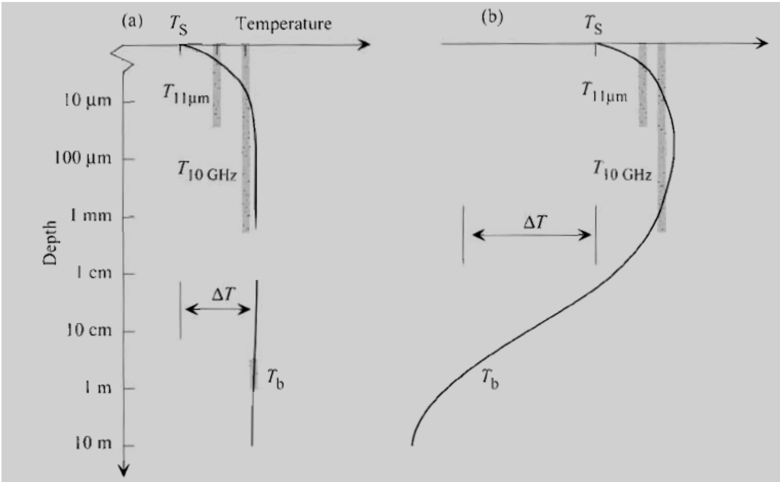


Figure 7.4. Comparison of characteristic temperature profiles for day and night conditions. Depth is on a log scale. (a) Day and night strong wind case, where $U > 6\,\mathrm{m\,s^{-1}}$. (b) Daytime weak wind and strong solar insolation case that yields a stratified upper ocean. On the figure, T_S is the surface temperature, the bar marked $T_{11\mu m}$ shows the depth range that contributes to the infrared 11-μm SST; the bar marked T_{10GHz} shows the depth range that contributes to the microwave surface temperature, T_b is the buoy or bulk temperature and ΔT is the temperature difference between the buoy and surface temperatures (Adapted from Figure 1, Donlon *et al.*, 2002).

The blackbody radiation emitted in water below this depth is absorbed by the neighboring water. Only at the sea surface can it escape.

The loss of heat by IR (longwave) radiation means there is a flux of heat from below toward the sea surface. This requires there be a vertical gradient in temperature getting warmer going down, i.e. there is a surface *cool bias* ΔT = skin temperature minus bulk temperature. Typically $\Delta T = -0.17 \pm 0.07$ K

This is because the heating due to shortwave is occurring at a deeper depth of, on average, about 3 m.

It is bulk temperature (at say 1 m to 3 m depth) that an oceanographer wants to know, but skin temperature that an IR satellite observes.

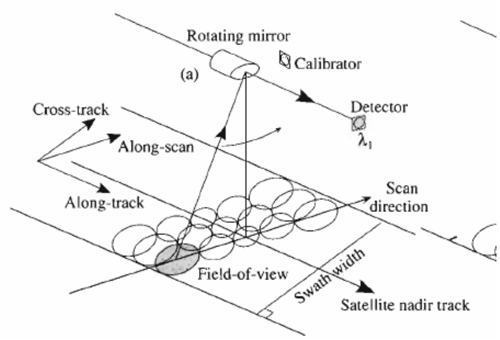
If winds are weak, the absorption of solar shortwave radiation during daytime can warm just the top few centimeters producing a difference between near surface temperature (and hence skin temperature) and the bulk temperature below. This presents another problem for calibration.

(If winds are stronger, this near surface gradient is mixed away by ocean turbulence.)

At night, with weak winds, the absence of solar heating to maintain stratification means that the surface cooling will cause the water column to convect, removing the surface warm layer. (The cool bias will remain).

AVHRR calibration is with respect to observations of bulk temperature, and uncertainty due to this temperature profile is a significant component of the AVHRR error.

<u>Satellite IR sensors, calibrations, and corrections</u>



AVHRR:

- A whisk-broom scanner
- Mirror rotating at 360 RPM = 6 scans per second
- resolution:

speed = $v = \sqrt{\dfrac{GM}{r}}$ = $(3.986 \times 10^{14} / (833 \times 10^3 + 6373 \times 10^3))^{1/2}$

$= 7.4 \times 10^3$ m s^{-1}

distance = $v \times 1/6$ second = $7.4 \times 10^3 / 6 = 1.2 \times 10^3$ m = 1.2 km

- angular resolution of 1.4 milliradian

  resolution: (Use $s=r\theta$) $833 \times 10^3 \times 1.4 \times 10^{-3} = 1.1 \times 10^3$ m = 1.1 km at nadir

- swath half-width of 1350 km (orbit altitude is 833 km)

- quantum detector – counts photon flux

  - semi-conductor: must be cooled to significantly less than the ambient temperature of the satellite to reduce noise (< 100 K or -173°C)

  - thermal connection from base-plate of detector to radiative heat sink (high thermal emissivity) facing open space

  - internal surfaces facing the detector must be cooled to reduce spurious IR radiation

  - condensation can degrade sensor performance – periodic outgassing step

- sensor calibration is by pre-launch calibration, and operational viewing cold space (assumed to be 3K) and an internal blackbody source

- sensors operate by observing 5 or 6 bands of different wavelength

  - 1 visible, 2 near IR, 3 in thermal IR

  - Band 2: ~0.8 μm = near IR … ocean emission is low
    - discriminates land/ocean boundary and cloud

  - Band 3A: ~1.5 μm day time reflectance discrimination of
    - Snow, ice, clouds, forest fires

  - Band 3B: ~3.7 μm night time SST

  - Bands 4 and 5: ~10.8 μm and ~12 μm day and night SST (split window algorithm)

- AVHRR transmits the 1 km resolution Local Area Coverage (LAC) data to HRPT receiving stations (High Resolution Picture Transmission) like the one on the roof of the IMCS building (over the loading dock).

- The satellite also stores Global Area Coverage (GAC) data and downloads this over a set of ground stations. GAC data are compiled from every 4th scan, with 4 of 5 adjacent samples averaged. This reduces data volume by factor of 10.

Lecture 8. 13 Feb 2006

Surface temperature (2)

Satellite borne radiometers observe radiance in selected wavelength bands at the Top of the Atmosphere (TOA)

Certain bands can be used for imaging Sea Surface Temperature (SST)

- Bands must have
  - high transmittance (low absorption) in the atmosphere
  - significant emittance at black-body temperatures typical of the ocean
- Wavelengths used for SST retrievals are principally
  - AVHRR Band 3B: ~3.7 μm (MODIS band 20) night-time SST
  - AVHRR Bands 4 and 5: ~10.8 μm and ~12 μm (MODIS bands 31 and 32) day and night SST using split window algorithm

Irradiance observed at TOA is the sum of

- radiation emitted and reflected by the sea surface … reduced by attenuation in the atmosphere (absorption and scattering)
- emission by the atmosphere itself

Reflectance of solar incoming radiation

- Band 3 ~ 12% of SST emittance (best used only for night-time SST)
- Bands 4 and 5 ~ 0.001% of SST emittance (but has more atmosphere emission)

<u>Converting multi-channel irradiances to SST data</u>

The radiance at the TOA is observed for a given wavelength λ_i is expressed in terms of a Planck function for that wavelength and the equivalent black-body temperature corresponding t the received radiances. The radiance for a given channel is sometimes referred to as the *brightness temperature* and reported in units of *Kelvin.*

$$L(\lambda_i) = f_P(T_i, \lambda_i) = f_P(T_s, \lambda_i)t_i + f_P(\overline{T}, \lambda_i)(1 - t_i)$$

The received radiance L or black-body temperature T_i is the sum of the radiance emitted by the ocean surface skin temperature T_s and average atmosphere temperature $\overline{T}$ with the factor ($1- t_i$) representing the atmospheric emissivity and is a function of the columnar water vapor, V.

The split window algorithm uses the brightness temperatures in bands 4 and 5: T_4 and T_5

These are approximately the same order as each other and the atmosphere temperature so we can express the Planck function with a Taylor series expansion with respect to the skin temperature we are trying to determine:

$$f_P(T_4) \cong f_P(T_S) + \left.\frac{\partial f}{\partial T}\right|_{T,\lambda}(T_4 - T_S)$$

After some algebra, a relationship between the channel 4 and 5 brightnesses w.r.t. the skin temperature can be derived:

$$T_4 - T_S = (\overline{T} - T_S)M_4$$
$$T_5 - T_S = (\overline{T} - T_S)M_5$$

$$T_S = T_4 + \Gamma(T_4 - T_5) \qquad \Gamma = \frac{(1 - t_4)}{(t_4 - t_5)}$$

What is happening in this algorithm is that the wavelengths λ_4 and λ_5 are both being viewed (temperature emission) and attenuated through the same atmosphere. This allows us to remove the atmospheric effects by having two independent looks.

Path length of the scan is included in the estimate of t

$$t_i = \exp(-m_i V \sec\theta)$$

where m_i is the dependence on transmittance on water vapor as a function of wavelength, V is the vertically integrated water vapor, and θ is the viewing angle.

In practice, the AVHRR algorithm formulated in this way but with coefficients adjusted empirically to agree with a climatology of in situ observations.

This means the estimate is not of skin temperature, but of bulk temperature.

SST (in °C) = 0.95876 T_4 + 2.564 (T_4 – T_5) – 261.68

Day-time retrievals can use band 3 in a triple-window algorithm.

This algorithm works well for the Multi-Channel SST (MCSST) product generated operationally for weather forecast models.

Nonlinear refinements to the algorithm are used to improve its accuracy. Γ is found to be a nonlinear function of SST because of water vapor effects. This is to be expected since moist atmospheres tend to occur over warm oceans. The coefficients are continuously adjusted to fit the match-up data base of observations from ships, drifters, and moored buoys. Since much of this data is quality-controlled in a delayed mode, reanalysis SST products readjust the algorithm coefficients to produce best-SST estimates for climate studies (as opposed to operational SST for forecasting).

Environmental sources of error

- High cirrus cloud
    - Thin semi-transparent ice clouds
    - Very cold so can introduce considerable errors (compared to low altitude clouds closer in temperature to the SST)
- Skin/bulk temperature difference
    - Diurnal cycle increases error in match-up data base
- Volcanic aerosols
    - Pinatubo eruption produced cool bias of satellite SST w.r.t. buoy data of 0.5°C too cold due to stratospheric aerosols – persisted for 2 years
    - Tropical biases were as much as 2°C
    - Adjustment of algorithm coefficients was used to correct for this effect in reanalysis SST
    -

Cloud detection algorithms

Clouds are colder and more reflective than the ocean surface

- Single band and single pixel threshold tests

- Multi pixel uniformity tests
    - Scattered clouds lead to high variance
- Multi-band test designed to detect certain types of clouds (e.g. cirrus)

Products with less resolution that the 1 km pixel size (e.g. 10 km MCSST) apply multiple tests:

- Warmest pixel or percentile value
- Consistency with neighboring values
- Comparison to climatology
- Comparison to running time mean

<u>GOES</u>

Resolution is less because of altitude of geostationary satellites.

Cloud detection can use observations every hour to reject moving features.

With multiple looks (as opposed to twice-daily for LEO satellites) it can be easier to build a daily composite with fewer cloud gaps – useful for weather prediction

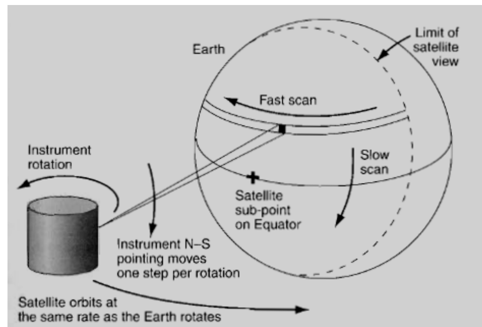


Figure 7.21. Meteosat radiometer scanning concept.

<u>ATSR and AATSR</u>

Dual look angle

Assume atmospheric properties are the same along each path (i.e. function of z only)

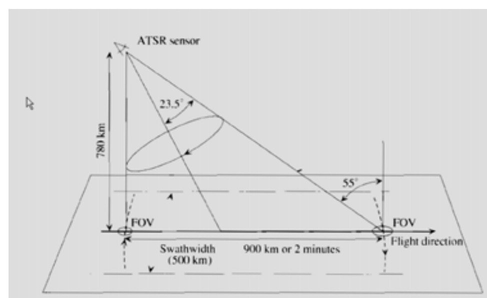


Figure 7.14. Schematic diagram of the ATSR and AATSR operation.

e.g. 60° path is twice as long as the nadir path

- The difference in radiance would be equivalent to the attenuation due to a single atmosphere – so subtract this from the nadir view
- Views are 2 minutes apart