"Course on Ocean-Atmosphere Interaction in the Tropics"
29 July - 17 August 1991

"The Atmospheric Heat Engine"

E. SARACHIK
University of Washington
Department of Atmospheric Sciences
Seattle, Washington
USA

lease note: These notes are intended for internal distribution only
THE ATMOSPHERIC HEAT ENGINE

FIG. 7. Schematic diagram of the flow of energy in the climatic system. A value of 100 units is assigned to the incoming flux of solar energy. All values represent annual averages for the entire atmosphere. For simplicity no separate energy boxes have been drawn for the other subsystems.

\[ F_0 = 1370 \text{ W/m}^2 \quad \text{Average} \]

\[ \therefore \text{At top of atmosphere} \]
\[ F = \frac{1}{4} (1-\alpha) 1370 \text{ W/m}^2 \]
\[ = 343 (1-\alpha) \text{ W/m}^2 \]

Globally \( \alpha \approx 0.3 \)

The top of earth without \( \alpha T_s = 343 (1-\alpha) \text{ atmosphere} \]
\[ T_s = 255^\circ K \]
As in the case of the water balance, the torques are not a cause, requirement for the transport of angular momentum. If the circulation were unable to transport angular momentum to middle latitudes, the surface westerlies there simply would not occur.

The balance of total energy presents a more complicated problem. Not only does the atmosphere exchange energy with the underlying Earth, but both the atmosphere and the underlying Earth gain energy from the sun and lose it to outer space through radiation. On this account it is desirable to examine first the energy balance of the entire atmosphere-ocean-Earth system, and then the more complicated energy balance of the atmosphere alone.

The incoming solar energy, which is the ultimate driving force for the atmospheric and oceanic circulations, is more intense in low than in high latitudes. Some of this energy is reflected or scattered back to space and plays no further role in the energy balance. The remainder is absorbed by the atmosphere and the Earth’s surface; this portion, like the total, is more intense in low latitudes.

The energy re-radiated to space by the atmosphere and the Earth’s surface is also more intense in low latitudes, although not so much more intense as one might expect in view of the higher temperature. Much of the outgoing radiation takes place from the uppermost layers of water vapour in the atmosphere; these extend to great heights in low latitudes and are therefore about as cold as the uppermost water vapour in higher latitudes. The net result is therefore a considerable excess of heating in low latitudes. It follows that there must be a poleward transport or transfer of energy across virtually every latitude. This transport may occur within the atmosphere or the oceans.

In contrast to the scarcity of numerical estimates of the angular-momentum exchange, there are numerous estimates of the incoming and outgoing radiation. Figure 21 is again based upon the values compiled by Sellers (1966) from a number of sources. The upper curve shows the solar energy reaching

Figure 21. — Average solar energy reaching the extremity of the atmosphere (upper solid curve), average solar energy absorbed by the atmosphere-ocean-Earth system (lower solid curve), and average infrared radiation leaving the atmosphere-ocean-Earth system (dashed curve), as given by Sellers (1968). Values are in watts m⁻² (scale on left). (1 watt m⁻² = 1.605 × 10⁻³ cal cm⁻² min⁻¹ = 0.754 kilolangleye year⁻¹.)
0–30°N. Similar figures are given in Oort and Vonder Haar (1976). Flanders and Smith (1975) report a figure of 7 W m⁻² for the difference in global mean of net radiation at the top of the atmosphere during May 1970 versus 1971, but obtained from two different satellite systems. The interannual variability apparent from satellite data presumably differs from the "true" variability of global mean net radiation at the top of the atmosphere, which is not known, but is expected to be small. Different satellite sensing systems are presumably fraught with different random and systematic errors.

According to the classical rule of random error propagation an error of 10 W m⁻² in the global mean would correspond to errors of 20 and 35 W m⁻² for the bands 0–30°N or S, and a 10° latitude band of the tropics, respectively. However, a certain portion of the aforementioned error of 10 W m⁻² is likely to be systematic, and its relative contribution is not known. Only a fraction of this imbalance is likely to stem from a "true" interannual variability. Any adjustment aimed at achieving a zero global mean of radiation at the top of the atmosphere, as exemplified in Section 5, Tables 3–5 and Figs. 7 and 8, will still lead to errors that are at least as large as the "true" interannual variability. In terms of the measurement, SWLW ¼ is a small difference between the downward and upward directed shortwave and the upward directed longwave radiation. Accordingly, efforts at substantially improving over a 5–10 W m⁻² error in SWLW ¼ may involve a formidable task in the determination of component fluxes. This brief review of evidence cautions against optimistic estimates of errors in the presently available satellite-derived net radiation at the top of the atmosphere.

Errors in atmospheric transport related to interannual variability (Oort, 1977) have been estimated
Figure 12. — The time-and-longitude averaged temperature \( \bar{\bar{T}} \) in January as estimated by Palmen and Newton (1967). Values are in degrees K.

Figure 13. — The time-and-longitude averaged temperature \( \bar{\bar{T}} \) in July as estimated by Palmen and Newton (1967). Values are in degrees K.
Figure 1. — The time-and-longitude averaged zonal wind ($\bar{u}$) in northern winter and southern summer (October-March) as estimated by Buch (1954) and Obasi (1963). Values are in m sec$^{-1}$.

Figure 2. — The time-and-longitude averaged zonal wind ($\bar{u}$) in northern summer and southern winter (April-September) as estimated by Buch (1954) and Obasi (1963). Values are in m sec$^{-1}$.
Figure 20. — Average annual evaporation (solid curve) and precipitation (dashed curve) per unit area as given by Sellers (1966). Values are in centimetres of water per year, or $\varphi$ cm$^{-2}$ year$^{-1}$ (scale on left).

Figure 21. — Excess of evaporation over precipitation (solid curve) as given by Sellers (1966), in $\varphi$ cm$^{-2}$ year$^{-1}$ (scale on left), and northward transport of water in the atmosphere required for balance (dashed curve) in units of $10^{11}$ g sec$^{-1}$ (scale on right).
Figure 18. — The time-averaged meridional circulation in northern winter as estimated by Palmén and Vuorela (1963). The unit for stream function $\psi$ is $10^{12} \, g \, sec^{-1}$.

Figure 19. — The time-averaged meridional circulation in northern summer as estimated by Vuorela and Tuominiemi (1964). The unit for stream function $\psi$ is $10^{12} \, g \, sec^{-1}$.
FIG. 13. Global distribution of the surface pressure (in mbars) reduced to sea level. Also shown are vector plots of the surface wind. For geostrophic flow the arrows should parallel the isobars. Each barb on the tail of an arrow represents a wind speed of 5 m s$^{-1}$. (Note that a uniform value of 1000 mbars has been subtracted from the pressure field.)

Rev. Mod. Phys., Vol. 56, No. 3, Jul 1984

FIG. 15. Zonal mean cross sections of (a) the zonal wind component in m s$^{-1}$, (b) the meridional wind component in m s$^{-1}$, and (c) the inferred flow of atmospheric mass in 10$^{10}$ kg s$^{-1}$ for annual-mean and zonally averaged conditions. The (direct) energy-releasing tropical Hadley cells, the (indirect) energy-consuming midlatitude Ferrel cells, and the weak direct polar cells are evident in both hemispheres.
well over land, although a dense network of rain gauges is needed to capture the many local, topographically related anomalies. Over the oceans, quantitative estimates of precipitation are difficult to make from moving ships, while island data are often unrepresentative of open ocean conditions. Thus, unfortunately, two of the basic parameters of the hydrological cycle and climate, i.e., evaporation and precipitation, are poorly known over large portions of the globe. We will see later that the difference, evaporation minus precipitation, $E - P$, can be measured probably much more accurately than either component itself using the aerological method [see later discussion of Eq. (4.4) and Fig. 25(b)].

The global mean evaporation and precipitation are estimated to be on the order of 1 m/year. Combining this estimate with the earlier estimate of 0.025 m water in the atmosphere, we may derive a recycling time of almost 10 days for the atmospheric water vapor.

The vertical distribution of water vapor in the atmosphere is presented in Fig. 24(a). It shows that the water
Figure 2.4 Normal sea level pressure distribution for a) January and b) July; 1000 mb subtracted from the actual pressure values (after Godbole and Shukla, 1981).
Figure 2.6  Same as Figure 2.5 but for 200 mb flow.
Figure 2.5  Long-term 850 mb tropical flow, a) December-February and b) June-August (after Sanders, 1975).
Velocity potential $\chi$:

\[ V = V_\psi + V_x \quad \text{s.t.} \quad \nabla \cdot V_\psi = 0 \quad \nabla \times V_x = 0 \]

\[ \nabla \cdot V = \nabla \cdot V_x \quad \Rightarrow \quad V_x = -\nabla \chi \]

\[ \nabla \cdot V = -\nabla \chi \]

Figure 2.7 Isopleths of the northern winter (a) and northern summer (b) mean velocity potential at 200 mb, and streamlines of the divergent part of the wind shown with arrows (after Krishnamurti et al., 1973 and Krishnamurti, 1971).
Figure 2.8 Isopleths of the July mean velocity potential at 850 mb (a), and at 200 mb (b) (units: $10^7 m^2 s^{-1}$, interval every 10 units) (after van de Boogaard, 1977).
Figure 1. Sea surface temperature profile along Equator in January, April, July, and October (data from Schutz and Gates, 1971-74).
Figure 4a  Mean SST filtered climatology (in °C) on a one-degree grid for the Pacific Ocean for January.

REYNOLDS (1982)

Figure 4b  Same as 4a except for the Atlantic and Indian Oceans.
Figure 7a. Mean SST filtered climatology (in °C) on a one-degree grid for the Pacific Ocean for April.

Figure 7b. Same as 7a except for the Atlantic and Indian Oceans.
Figure 10a  Mean SST filtered climatology (in °C) on a one-degree grid for the Pacific Ocean for July.

Figure 10b  Same as 10a except for the Atlantic and Indian Oceans.
Figure 13a  Mean SST filtered climatology (in °C) on a one-degree grid for the Pacific Ocean for October.

Figure 13b  Same as 13a except for the Atlantic and Indian Oceans.
Figure 2.1a Same as Figure 2.2 but for DJF season.

Figure 2.1b Same as Figure 2.2 but for MAM season.
Figure 2.3c Same as Figure 2.2 but for JJA season.

Figure 2.3d Same as Figure 2.2 but for SON season.
Figure 6. Mean spring season OLR for the period June, 1974 through November, 1981. The contour interval is 20 W m⁻², with values > 280 W m⁻² dashed. The day and night observations have been averaged.
Figure 7. Same as Figure 6, except for summer season.
Figure 8. Same as Figure 6, except for autumn season.
Figure 2.14 Vertically and latitudinally averaged diabatic heating (thick line, units: K/day) and vertical omega velocity (thin line, units: $10^{-3}$ m/s) for Jan-Feb, 1979; zonal averages of diabatic heating (thick straight line) and vertical omega velocity (thin straight line) are indicated.
FIG. 1.4 WIND VECTORS AND ISOTACHS

APRIL

ONTOUR INTERVAL: 2 M/S  REFERENCE LINE: 6 M/S
bimonthly period and for the annual mean, as well as for surface topography relative to 3000 db, for 500 db relative to 1000 db, and for 1000 db relative to 2000 db, but only a few are shown here. The others have been published in a technical report (Wyrtki, 1974c).

3. The mean annual dynamic topography

While the general pattern of dynamic height on our map of the mean annual dynamic topography (Fig. 1) is quite similar to that shown by Reid (1961), there are some important differences. The few data points used by Reid show the North Pacific subtropical gyre as a rather smooth feature. Our maps clearly indicate the presence of a north equatorial ridge, south of which the North Equatorial Current flows westward, and north of which the Subtropical Countercurrent, described by Yoshida and Kidokoro (1967), flows eastward. There is also a pronounced ridge on the right-hand side of the Kuroshio, similar to the one outlined by Defant (1941) along the right-hand side of the Gulf Stream. In fact, the north equatorial ridge and the ridge along the Kuroshio form a continuous U-shaped ridge with its opening facing east.

When viewed on synoptic maps of dynamic topography (Japanese Oceanographic Data Center, 1967, 1968, 1969) the ridge is not always continuous but consists of several sections with eddies interspersed
from the area east of Luzon past the Hawaiian Islands to about 120°W, and from March to June it penetrates into the eastern tropical Pacific. The ridge is situated slightly north of 20°N, except during the period from March to June when it is south of 20°N. Fluctuations of the subtropical high are also evident in the eastward penetration of the 190 dyn-cm isobar.

Throughout the North Pacific Current, the pattern of dynamic topography changes little during the year, although absolute values of dynamic topography increase from winter to summer by about 12 dyn-cm in the western part of the current and by about 7 dyn-cm in the eastern North Pacific. No mean annual variation of the slope across the North Pacific Current could be conclusively detected from these maps.

In the subarctic region of the Pacific the existence of two gyres, the Alaska gyre and the Kamchatka gyre, is clearly evident. It is also weakly indicated in the map by Reid (1961). Both gyres exist throughout the year. They penetrate well below 300 m depth, which is apparent from the topography of the 300 db surface relative to 1000 db. The Alaska gyre is most intense during the first half of the year and is weakest in November and December. The Kamchatka gyre seems to be weakest during the summer from June to October and strongest at the end of winter in March and April.
from March to June, when the Countercurrent is weakest and large amounts of water are recirculated between Countercurrent and North Equatorial Current. During the second part of the year, when both currents are strong, this low is not as pronounced.

Along the equator, between 100°W and 180°, an east-west slope of about 40 dyn-cm is always present, as shown by Knauss (1963). The fact that this slope is slightly stronger in the second half of the year than in the first half compares favorably with sea level records at Canton Island and the Galapagos Islands being opposite in phase. It indicates that the annual variation of the east-west slope along the equator is in opposition to the transports of the Equatorial Undercurrent, the latter documented by Taft and Jones (1973).

The subtropical anticyclonic gyre of the North Pacific undergoes considerable fluctuation during the year. Highest dynamic heights are always found east of Taiwan, but the intensity of the high relative to 1000 db varies from 216 dyn-cm in January-February to 234 dyn-cm in July-August. This variation corresponds closely to that of surface layer temperature, so one can conclude that the annual variation of dynamic height in the subtropical gyre is due to the variation of density in the surface layer as associated with variations of heat content.

The north equatorial ridge stretches across the Pacific