"Energetics"

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ENERGETICS. *[This article treats a technical aspect of climate and weather studies; some of it is intended for readers at an advanced level.]*

Atmospheric energetics, the study of the distribution and transformation of energy in the atmosphere, has long been a subject of fascination for meteorologists. Nonetheless, it may be argued that energetics serves more as an after-the-fact bookkeeping check than as a practical tool for developing a predictive understanding of weather and climate. A brief discussion of the ambiguous role of energetics in atmospheric science serves as our introduction to the topic. The remainder of this article is structured in three parts. First, definitions of potential, kinetic, and available potential energy are presented, along with a more advanced and mathematical discussion of how these are linked to internal energy, enthalpy, and entropy. We then examine transformations between various forms of energy, considering the atmosphere as a thermodynamic heat-engine. Finally, we consider the observed energy transports, focusing on meridional transport in the atmosphere and ocean.

Energy is one of the basic physical quantities for which we can write a local conservation law-an accounting that says that the change of a quantity inside a volume equals the net flux (flow) of that quantity through the walls of the volume. [See Conservation Laws.] This conservation law allows us to make powerful claims about the large-scale distribution and transport of energy without knowing the details of small-scale dynamics. For example, knowing that the energy of polar regions is approximately constant allows us to infer the oceanic energy transport into those regions from the difference between the poleward atmospheric transport and net energy radiated to space. Energy is the common currency that allows us to balance the books between fluxes of radiation, moisture, heat, and mechanical energy.

In many areas of physics, an understanding of energetics provides an efficient route to understanding and predicting a system's dynamics. In the atmosphere, the winds, which themselves contain only about 0.04 percent of the system's total energy, transport large fluxes of heat and moisture and thus strongly influence the distribution of energy. Although the maintenance of the winds themselves may be studied as a problem of energetics, consideration of other properties-such as momentum or potential vorticity-may provide a more powerful route to understanding the winds, and so to understanding the distribution of heat and moisture. [See Potential Vorticity.]

Comparing energy fluxes in the solar-driven planetary heat engine, which are of order 200 watts per square meter, with the fluxes typical of various biogeochemical cycles illustrates the limitations to the degree of insight that can be gained from energetic considerations. Nuclear decay in the planet's interior provides an upward flux of about 0.07 watts per square meter. This flux drives volcanoes and continental drift, which influence climate through emission of gases and reconfiguration of oceans and land with an importance that overwhelms the trivial direct impact of the geothermal heat flux. Photosynthesis converts solar radiation to chemical energy with peak fluxes of about 1 watt per square meter; as with geophysical energy fluxes, the relevant climatic impact of biology is not on energetics, but rather on the chemical composition of the atmosphere which indirectly regulates much larger energy fluxes. The energy fluxes associated with our fossil-fueled civilization amount to only a few tens of watts per square meter over large urban areas, and only 0.02 watts per square meter on a global average; yet the indirect effects of our modification of atmospheric gas and aerosol concentrations may have already changed global energy fluxes by a few watts per square meter. [See Global Warming; Paleoclimates.]

Definitions. Gravitational *potential energy* is the energy that mass has by virtue of its position in a gravitational field. It is the stored work that was done against the force of gravity in lifting an object from a reference surface. It does not depend on the composition of the body-only on its mass and position. In general, only changes in potential energy have significance; the actual magnitudes depend on the definition of the reference surface and have no physical meaning. However, when the planetary surface is used as a reference, the absolute value of the potential energy of an atmosphere in hydrostatic equilibrium is meaningful because of the partition between internal and potential energy (see the more detailed discussion below).

In atmospheric physics we assume that the gravitational field is constant with height, so that potential energy is a linear function of height as well as mass. We usually consider an effective gravitational field which is the sum of the gravitational field and the vertical component of the outward force due to the earth's rotation (the centrifugal force). The resulting acceleration, denoted g, varies by only about 0.3 percent between the equator and the poles. The potential energy of a mass m is mgz, where z is the height; and the potential energy per unit volume is ρgz, where p is the density.

Kinetic energy, the energy of motion, is the total work done on an object in accelerating it from rest. It is linearly proportional to an object's mass, and to the square of its velocity. In the atmosphere we write the kinetic energy density-the amount per unit volume—as $\rho v^2/2$, where ρ is the density and v is the speed. Kinetic energy is generated when a parcel of air is accelerated by gravitational or pressure-gradient forces; the energy supplied by the force comes from the conversion of potential energy in the former case, or internal energy in the latter. Kinetic energy can be destroyed by conversion back to internal or potential energy, which occurs when the corresponding gravitational or pressure-gradient forces act to decelerate the parcel. Alternatively, kinetic energy can be dissipated by viscous (frictional) forces, which always act to remove velocity gradients and convert kinetic energy into heat. [See Friction.]

Before discussing available potential energy we need to mention the connection between internal and potential energy, and to review the definition of adiabatic and diabatic processes. In addition, we will write the continuity equation for energy in the atmosphere, although it is not necessary for understanding the rest of this article.

Internal and potential energy are closely coupled. When air in a hydrostatic atmosphere is heated, it expands, doing work against the local pressure by lifting the air above it, and converting some of the heat into potential energy. Raising the temperature of a parcel by δT requires $C_v\delta T$ of internal energy and $R\delta T$ of work, where C_v is the specific heat at constant volume and R is the ideal gas constant. The sum $(C_v + R)\ \delta T = C_p\delta T$ is called *the enthalpy*, where C_p is the specific heat at constant pressure. The ratio R: C_v of potential to internal energy is constant in a column of air in hydrostatic equilibrium, and it is thus, to a good approximation, a constant ratio in the atmosphere as a whole. [See Enthalpy.]

When we add heat to a parcel of *air-diabatic heating*—we increase both its entropy and its enthalpy. A parcel that moves without exchanging heat maintains its entropy and is said to move *adiabatically*. Temperature change of a parcel due to adiabatic compression is called *adiabatic heating*. Dry air that moves adiabatically in a hydrostatic atmosphere conserves the sum of its enthalpy and potential energy-the dry static energy, $C_pT + gz$—which corresponds to its potential temperature. [See Adiabatic Processes; Diabatic Processes; Entropy.]

If air is ascending or descending while maintaining a constant temperature, then diabatic and adiabatic temperature tendencies are balanced. For example, in the tropics ascending air is confined to small regions of active convection, while most of the upper tropical troposphere experiences gradual descent. Where the air is descending, adiabatic heating is matched by radiative cooling. In the convective systems, adiabatic cooling is mainly balanced by latent heating.

We can now write the local conservation law for energy in the atmosphere. Ignoring moisture, the energy density is the sum of internal, potential, and kinetic energies.

$$\rho(C_vT + gz + \tfrac{1}{2}\nu^2)$$

However, the quantity which is conserved by a parcel moving in a hydrostatic atmosphere includes the enthalpy rather than the internal energy; it is called *the dry static energy*. The energy flux is the product of the dry static energy and the velocity vector.

$$\rho(C_pT + gz + \tfrac{1}{2}\nu^2)\ v$$

The conservation law for atmospheric energy density says that the rate of change of energy in a volume plus the flux out of the volume equals the diabatic heating rate, Q.

$$\frac{\partial}{\partial t} \rho(C_v T + gz + \tfrac{1}{2} \nu^2) + \nabla \cdot \rho(C_p T + gz + \tfrac{1}{2} \nu^2) \, \nu = \rho Q$$

Moisture can be included in this accounting by adding a latent heat term, Lq, to the expressions for energy density and flux, and subtracting the diabatic heating due to condensation from the right-hand side.

Only a small fraction of the atmosphere's total internal and potential energy can be converted to kinetic energy. This fraction is called the ***available potential energy;*** it may be defined as the maximum energy that can be extracted from an atmosphere at rest by adiabatic processes. Available potential energy is computed by considering a hypothetical atmosphere, called the ***reference state,*** which is in the minimum energy state that may be produced from the real state by adiabatic rearrangement of mass. The difference between the potential plus internal energy of the atmosphere and that of the reference state is the available potential energy. In a hydrostatic atmosphere, the available potential energy is proportional to the variance of temperature on pressure surfaces.

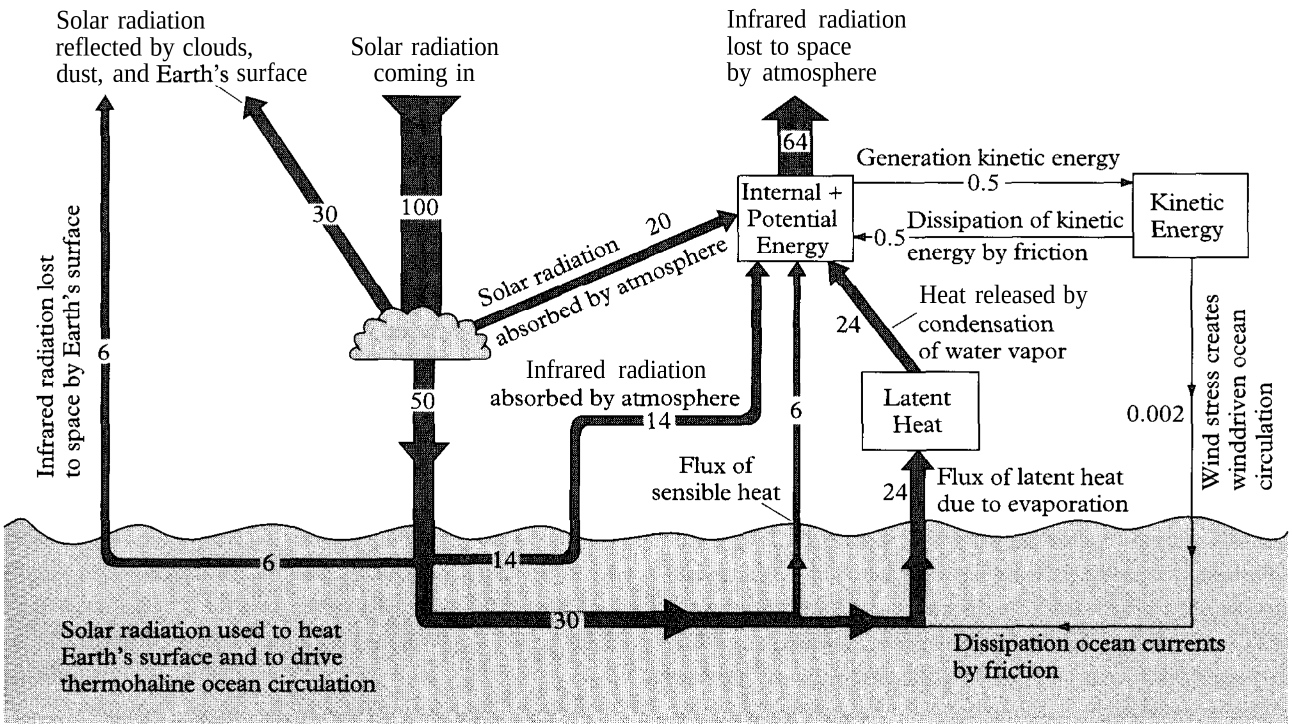
Unlike internal, potential, or kinetic energy-which may be defined for individual parcels of air-the available potential energy depends on the configuration of the atmosphere on a larger scale. Although it is sometimes convenient to discuss the available potential energy per unit mass or area, one must remember that it is not a local property and cannot be incorporated into a local conservation law.

A few examples may help to clarify the role of available potential energy. Consider an isothermal (uniform-temperature) atmosphere which is at rest and in hydrostatic equilibrium-that is, in which the pressure gradient force is balanced by gravity. This idealized atmosphere has zero available potential energy. Each parcel of air (which we may imagine as a small balloon) could do work either by expanding against the pressure that confines it (thereby converting internal to mechanical energy), or by being lowered in the gravitational field (thereby liberating potential energy). However, no net energy can be extracted from the system by such methods, because more work must be done on other parcels of air in compressing or lifting them to make room for the first parcel than can be gained by its expansion or lowering. This is both a necessary and a sufficient condition for hydrostatic equilibrium.

Now consider an atmosphere divided into two regions with the same surface pressure, each at rest and in hydrostatic equilibrium, which have uniform but different potential temperatures. Note that constant ***potential temperature*** implies a temperature that decreases linearly with height, and note that potential temperature rather than temperature is constant for a parcel of air moving in a hydrostatic atmosphere. Now imagine that the barrier between the two regions is removed. At a given altitude the pressure is smaller on the side with the higher temperatures. These pressure differences produce a net force which accelerates the gas, converting available potential and internal energy to kinetic energy. In order to compute the total amount of available potential energy, we must determine the minimum energy reference state, which in this case is the state where the high potential-temperature air lies atop the low in a situation of uniform horizontal stratification. [See Potential Temperature.]

Atmospheric Thermodynamics. The sunlight absorbed by the Earth, oceans, and atmosphere deposits energy, which must be returned to space as infrared radiation in order for the planet to maintain a roughly constant temperature. Although solar absorption and infrared radiation must be approximately balanced globally, they are not in balance locally. Solar absorption is concentrated at the surface and in the tropics; most of the infrared radiation to space, however, originates in the middle troposphere, and it is more evenly distributed between equatorial and polar regions than is solar absorption. In order to maintain the time-average temperature distribution, the system must transport heat from regions where solar heating dominates infrared cooling to regions where radiative cooling dominates. Thus the atmosphere transports heat from the ground to the upper troposphere, and the atmosphere and ocean together transport heat poleward from the tropics. Figure 1 illustrates the various processes considered under atmospheric thermodynamics. [See Thermodynamics.]

The difference between solar heating and infrared cooling, called the ***net radiative heating,*** may be regarded as the driving force for atmospheric thermodynamics. Solar radiation reaches the top of the atmosphere with a flux of 1,370 ± 10 watts per square meter, about 30 percent of which ***(the planetary albedo)*** is reflected back to space. The remainder is absorbed with a flux of 250 watts per square meter when averaged over the Earth's area, which is four times larger than the disk it presents to the sun. About 70 percent of the solar absorption happens



ENERGETICS. Figure 1. *Atmospheric thermodynamics: a schematic diagram of energy flows in the climate system.* (Adapted from Peixoto and Oort, 1992).

at the surface, and about 40 percent of that leaves the surface as net infrared radiation, leaving a net surface radiative heating of 150 watts per square meter. On a global and annual average, the net radiative heating of the surface is equal to the net radiative cooling of the atmosphere because of the balance between absorbed solar and outgoing infrared at the top of the atmosphere. [See Albedo; Radiation.]

Over the oceans or moist ground, most of the net radiative heating of the surface is used to evaporate water, which removes energy from the surface as *latent heat.* When the vapor condenses it deposits almost all of its latent heat in the air rather than in the condensed water, which returns to the surface as precipitation. The net result is to transport heat from the surface to the air where condensation occurs. On a global average, about 70 percent of the net radiative heating of the surface is removed by latent heat flux. The remaining 30 percent leaves the surface by conduction of *sensible heat* to the overlying air. [See Latent Heat; Sensible Heat.]

In summary, the atmosphere experiences diabatic heating (or cooling) from four processes: latent heating, sensible heating at the surface, solar absorption, and infrared heating or cooling. (There is additional heating due to the dissipation of kinetic energy, which is usually ignored.) The component of diabatic heating that is unevenly distributed horizontally produces available potential energy at a globally averaged rate of about 2 watts per square meter-only about 1 percent of the solar heating. Available potential energy is then transformed into kinetic energy, which is ultimately dissipated to produce heat.

A deeper insight into the process of dissipation may be gained by considering the spatial scale at which various energy transformations occur. First, a few general words about the definition of spatial scale are in order. The distribution of a quantity with respect to scale is defined by its *spatial power spectrum,* which is usually computed as a Fourier transform. We say that a quantity varies at a certain scale when most of its total spatial variance-the power in its power spectrum-is concentrated at that scale. Physical laws that are linear, such as Newton's $F = ma$ (force equals mass times acceleration), preserve the distribution of variance with respect to scale; that is, if F varies at a certain scale then so will a. Nonlinear laws allow energy to be transferred between scales, generally from a larger to smaller scale.

The generation of available potential energy by diabatic heating and its subsequent conversion to kinetic energy are concentrated at the spatial scale that characterizes the heating's variance-in the horizontal plane this is a few thousand kilometers and larger. Kinetic energy is dissipated when work is done against viscosity. Viscous forces depend on velocity gradients, which are larger in smaller eddies. The rate of dissipation in eddies is therefore a strong function of their scale; it is in fact inversely proportional to the square of an eddy's size. Thus kinetic energy generated at large scales is transferred to smaller and smaller scales, as allowed by the nonlinearity of the equations of fluid motion, until it reaches a scale where frictional dissipation balances the flow of energy from larger scales. In both the atmosphere and the ocean this dissipation scale, called the *Kolmogorov length,* is about 1 millimeter.

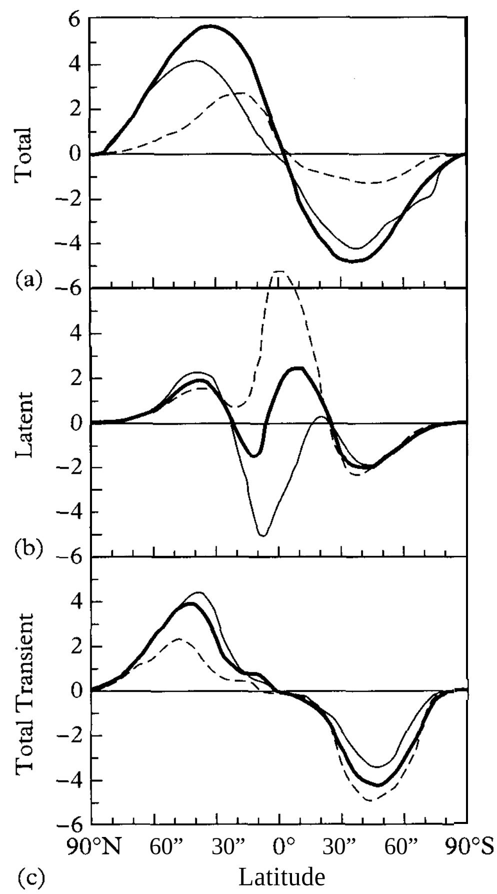
Although the net flow of energy is from available potential energy to kinetic energy and from larger to smaller scales, reverse flows also occur. Some energy is transformed from small to large scales; and at larger scales, some kinetic energy is transformed back to available potential energy. The net result is always to balance the generation of available potential energy by the dissipation of kinetic energy.

It is interesting to view atmospheric energetics from the standpoint of thermodynamics-that is, to analyze the atmosphere as a heat engine operating between regions of net radiative heating and cooling. A heat engine transforms heat into work by taking heat energy from a hot reservoir, transforming a fraction of it into work, and dumping the remainder into a cold reservoir. The maximum work that can be produced is limited by the requirement that entropy not decrease. The fraction of the thermal energy input which can be converted to work-the maximum thermodynamic efficiency-is given by the temperature difference between the two reservoirs divided by the temperature of the hot reservoir.

Because of the continuous nature of the atmospheric temperature distribution, we cannot precisely define hot and cold reservoirs; as a rough approximation, a pole-to-equator differential of 30 K and a hot temperature of 300 K limits the thermodynamic efficiency to about 10 percent. The generation of kinetic energy is thus more strongly constrained by limited production of available potential energy, which amounts to only about 1 percent of the solar heating.

Energy Transport. As discussed above, the most important global energy flows are vertical and meridional, upward from the surface and poleward from the tropics. There is a huge west-to-east (zonal) energy flow associated with the mid-latitude jet streams, but only the convergence or divergence of this closed loop of flow can alter local energy budgets, and so this is a small contribution to annual averages. [See Jet Stream.] Although its annual average contribution is small, zonal transport plays an important role in reducing the seasonal variation of mid-latitude continental climates by coupling the large oceanic heat reservoir to the much smaller terrestrial one.

It is instructive to divide the zonally averaged meridional transport into contributions from transient eddies, zonal mean flow, and zonally stationary eddies (Figure 2). Each component is caused by a different physical mechanism. The transient eddies are due to the covariance of energy density and meridional wind. A positive transient-eddy heat flux reflects the tendency of positive deviations from the mean wind to be correlated with pos-



ENERGETICS. Figure 2. *Meridional energy transport.* (a) Annual average energy transported by the atmosphere (thin line), ocean (broken line), and atmosphere and ocean combined (thick line). (b) Transport of latent heat on an annual average (thick line), transports during December-January-February (thin line), and transport during June-July-August (broken line). (c) Energy transport by transient eddies on an annual average (thick line), December-January-February (thin line), and June-July-August (broken line). (From Keith, 1995.)

itive deviations from the mean energy density. Transient eddy transport is principally associated with the traveling cyclonic or anticyclonic weather systems that prevail in mid-latitudes. On the ground in northern latitudes, we can observe this transport as the tendency for warm moist weather to come from the south, while cold dry air comes from the north. The zonal mean transport is due to the time-mean and zonal-mean wind acting on the time-mean and zonal-mean energy density. It is primarily associated with the tropical Hadley cell. Finally, the stationary eddies are due to covariances around a latitude circle between the time-average wind and energy density. Northward transport by stationary eddies at a given latitude is due to the tendency for winds that blow more strongly to the north than average at that latitude to be correlated with warmer-than-average temperatures. Stationary eddy transport is due to standing waves in the zonal circulation which are seen as stationary patterns in the jet stream. [See General Circulation; Meridional Circulations.]

In the tropics, within 20 degrees of the equator, solar heating exceeds infrared cooling by about 70 watts per square meter. About 50 watts per square meter of this is transported poleward by the atmosphere, primarily in the form of zonal-mean and stationary eddy flows, with the oceans removing the remaining 20 watts per square meter.

The annual-average tropical transports are small residuals of large seasonal transports which act in opposite directions. To understand the energetics of the Hadley cell better, consider the December-January-February season, the Southern Hemisphere summer. Air rises at around 10” south latitude, driven by the strong diabatic heating there; the ascending air cools adiabatically as it expands and then flows northward, carrying potential energy to the regions of descent around 20” to 30” north latitude. This northward transport of potential energy overwhelms the southward transport of internal energy “to produce an 8-petawatt (1 petawatt is 10^{15} watts) net transport of internal plus potential energy (dry static energy) to the winter hemisphere. Because the lower (southward-flowing) leg of the Hadley cell is much moister than the high-altitude flow, there is a net transport of latent heat (6 petawatts) from winter to summer hemisphere (Figure 2b), reducing the total northward energy transport across the equator to 2 petawatts.

Poleward of 30” latitude, meridional transport is dominated by transient eddies. The only exception is between about 50” and 60” north latitude, where stationary eddies transport about 40 percent of the energy during the Northern Hemisphere winter. Transient eddy transport is poleward for both latent heat and internal energy during all seasons; the fraction carried by latent heat declines from about 43 percent at 35° north or south, to about 10 percent at 75”. The ratio of latent to dry heat transport may be understood by assuming that the eddies mix the two forms of energy in the same way, while the fraction of atmospheric energy in the form of latent heat declines with the equilibrium vapor pressure.

Vertical heat transport lowers the surface temperature by moving energy from the surface, where very little energy can be radiated to space, to the middle troposphere, which is more effectively cooled by radiation. Despite its importance, the modes of vertical heat transport are much less well understood than are those of meridional transport. In particular, we do not know the relative importance of small-scale convection and large-scale circulation.

Understanding the factors that regulate meridional energy transport is of central importance in understanding the current meridional temperature distribution and in predicting the climate’s response to perturbations. If the meridional energy transport is controlled by the temperature gradient, then we may imagine a feedback loop in which increased transport warms the polar regions, causing a decreased temperature gradient which then reduces the meridional transport, thus maintaining the equilibrium. A plausible and commonly accepted mechanism for this regulation is to assume that mid-latitude eddy transport is governed by baroclinic instability, which depends on the meridional temperature gradient (among other factors). However, there are difficulties with this explanation. One must also consider the combined effects of oceanic and atmospheric transport in maintaining the temperature structure, and it appears that oceanic transports respond more strongly to differences in freshwater fluxes than to temperature.

[See *also* Dynamics.]

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