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EVAPORATION AND HUMIDITY

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Introduction

Evaporation from the sea and humidity in the air above the surface are two important and related aspects of the phenomena of air–sea interaction. In fact, most subsections of the subject of air–sea interaction are related to evaporation. The processes that control the flux of water vapor from sea to air are similar to those for momentum and sensible heat; in many contexts, the energy transfer associated with evaporation, the latent heat flux, is of greatest interest. The latter is simply the internal energy carried from the sea to the air during evaporation by water molecules. The profile of water vapor content is logarithmic in the outer layer, from a few centimeters to approximately 30 m above the sea, as it is for wind speed and air temperature under neutrally stratified conditions. The molecular transfer rate of water vapor in air is slow and controls the flux only in the lowest millimeter. Turbulent eddies dominate the vertical exchange beyond this laminar layer. Modifications to the efficiency of the turbulent transfer occur due to positive and negative buoyancy forces. The relative importance of mechanical shear-generated turbulence and density-driven (buoyancy) fluxes was formulated in the 1940s, the Monin-Obukhov theory, and the field developed rapidly into the 1960s. New technologies, such as the sonic anemometer and Lyman-alpha hygrometer, were developed, which allowed direct measurements of turbulent fluxes. Furthermore, several collaborative international field experiments were undertaken. A famous one is the 'Kansas' experiment, whose data were used to formulate modern versions of the 'flux profile' relations, i.e., the relationship between the profile in the atmosphere of a variable such as humidity, and the associated turbulent flux of water vapor and its dependence on atmospheric stratification.

The density of air depends both on its temperature and on the concentration of water vapor. Recent improvements in measurement techniques and the ability to measure and correct for the motion of a ship or aircraft in three dimensions have allowed more direct measurements of evaporation over the ocean. The fundamentals of turbulent transfer in the atmosphere will not be discussed here, only the special situations that are of interest for evaporation and humidity. As the water molecules leave the sea, they remove heat and leave behind an increase in the concentration of sea salts. Evaporation, therefore, changes the density of salt water, which has consequences for water mass formation and general oceanic circulation.

This article will focus on how humidity varies in the atmosphere, on the processes of evaporation, and how it is modified by the other phenomena discussed under the heading of air–sea interaction. All processes occurring at the air–sea interface interact and modify each other, so that none are simple and linear and most result in feedback on the phenomenon itself. The role of wind, temperature, humidity, wave breaking, spray, and bubbles will be broached and some fundamental concepts and equations presented. Methods of direct measurements and estimation using *in situ* mean measurements and satellite measurements will be discussed. Subjects requiring further research are also explored.

History/Definitions and Nomenclature

Many ways of measuring and defining the quantity of the invisible gas, water vapor, in the air have developed over the years. The common ones have

Table 1 Measures of humidity

Nomenclature	*Units (SI)*	*Definition*
Absolute humidity	$kg\,m^{-3}$	Amount of water vapor in the volume of associated moist air
Specific humidity	$g\,kg^{-1}$	The mass of water per unit mass of moist air (or equivalently in the same volume)
Mixing ratio	$g\,kg^{-1}$	The ratio of the mass of water as vapor to the mass of dry air in the same volume
Saturation humidity	Any of the above units	Can be given in terms of all three units and refers to the maximum amount the air can hold at its current temperature in terms of absolute or specific humidity, corresponds to 100% relative humidity
Relative humidity (RH)	%	Percent of saturation humidity that is actually in the air
Vapor pressure	hPa (or mb)	The partial pressure of the water vapor in the air
Dew point temperature	°K, °C	The temperature at which dew would form based on the actual amount of water vapor in the air. Dew point depression compared to actual temperature is a measure of the 'dryness' of the air
Wet-bulb temperature	°K, °C	This is a temperature obtained by the wetted thermometer of the pair of thermometers used in a psychrometer[a] (see Measurements chapter)

[a]A psychrometer is a measuring device consisting of two thermometers (mercury in glass or electronic), where one thermometer is covered with a wick wetted with distilled water. The device is aspirated with environmental air (at an air speed of at least 3 $m\,s^{-1}$). The evaporation of the distilled water cools the air passing over the wet wick, causing a lowering of the wet thermometer's temperature, which is dependent on the humidity in the air.

been gathered together in **Table 1**, which gives their name, definition, SI units, and some further explanations. These quantitative definitions are all convertable one into another. The web-bulb temperature may seem rather anachronistic and is completely dependent on a rather crude measurement technique, but it is still a fundamental and dependable measure of the quantity of water vapor present in the air.

Evaporation or turbulent transfer of water vapor in the air was first modeled in analogy with down–gradient transfer by molecular conduction in solids. The conductivity was replaced by an 'Austaush' coefficient, A_e, or eddy diffusion coefficient, leading to the expression:

$$E = -A_e \rho \frac{\partial \bar{q}}{\partial z} \qquad [1]$$

where E is the evaporation rate, ρ the air density, $\bar{q}$ is mean atmospheric humidity, and z represents the vertical coordinate. Assuming no advection, steady state, and no accumulation of water vapor in the surface layer of the atmosphere (referred to as 'the constant flux layer'), the A_e is a function of z as the turbulence scales increase away from the air–sea interface and the gradient is a decreasing function of height, z, as the distance from the source of water vapor, the sea surface, increases.

Determining E by measuring the gradient of q has not proved to be a good method because of the difficulties of obtaining differences of q accurately enough and in knowing the exact heights of the measurements well enough (say from a ship or a buoy on the ocean). The A_e must also be determined, which would require measurements of the intensity of the turbulent exchange in some fashion. The so-called direct method for evaluating the vapor flux in the atmosphere requires high frequency measurements. This method has been refined during the past 35 years or so, and has produced very good results for the turbulent flux of momentum (the wind stress). Fewer projects have been successful in measuring vapor flux over the ocean, because the humidity sensors are easily corrupted by the presence of spray or miniscule salt particles on the devices, which being hygroscopic, modify the local humidity.

Evaporation, E, can be measured directly today by obtaining the integration over all scales of the turbulent flux, namely, the correlation between the deviations from the mean of vertical velocity (w') and humidity (q') at height (z) within the constant flux layer. This correlation, resulting from the averaging of the vapor conservation equation (in analogy to the Reynolds stress term in the Navier–Stokes equation) can be measured directly, if sensors are available that resolve all relevant scales of fluctuations.

The correlation equation is

$$\overline{\rho w \cdot q} = \bar{\rho}\bar{w}\bar{q} + \bar{\rho}\overline{w'q'}, \qquad [2]$$

where w and q are the instantaneous values and the overbar indicates the time-averaged means. The product of the averages is zero since $\overline{w} = 0$. Much discussion and experimentation has gone into determining the time required to obtain a stable mean value of the eddy flux $\bar{\rho}\overline{q'w'}$. For the correlation term $\bar{\rho}\overline{w'q'}$ to represent the total vertical flux, there has to be a spectral gap between high and low frequencies of fluctuations, and the assumption of steady state and horizontal homogeneity must hold. The required averaging time is of the order of 20 min to 1 h.

Another commonly used method, the indirect or inertial dissipation method, also requires high frequency sensing devices, but relies on the balance between production and destruction of turbulence to be in steady state. The dissipation is related to the spectral amplitude of turbulent fluctuations in the inertial subrange, where the fluctuations are broken down from large-scale eddies to smaller and smaller scales, which happens in a similar fashion regardless of scale of the eddies responsible for the production of turbulence in the atmospheric boundary layer. The magnitude of the spectrum in the inertial subrange is, therefore, a measure of the total energy of the turbulence and can be interpreted in terms of the turbulent flux of water vapor. The advantage of this method over the eddy correlation method is that it is less dependent on the corrections for flow distortion and motion of the ship or the buoy platform, but it requires corrections for atmospheric stratification and other predetermined coefficients. It would not give the true flux if the production of turbulence was changing, as it does in changing sea states. Most of the time, the direct flux is not measured by either the direct or the indirect method; we resort to a parameterization of the flux in terms of so-called 'bulk' quantities.

The bulk formula has been found from field experiments where the total evaporation E has been measured directly together with mean values of q and wind speed, U, at one height, $z = a$ (usually referred to as 10 m by adjusting for the logarithmic vertical gradient), and the known sea surface temperature.

$$E = \rho\overline{w'q'} = \bar{\rho} \cdot C_{E_a}\overline{U_a}(\overline{q_s} - \overline{q_a}) \qquad [3]$$

where $\overline{q_s}$ is the saturation specific humidity at the air–sea interface, a function of sea surface temperature (SST). Air in contact with a water surface is assumed to be saturated. Above sea water the saturated air has 98% of the value of water vapor density at saturation over a freshwater surface, due to the effects of the dissolved salts in the sea. C_{E_a} is the exchange coefficient for water vapor evaluated for the height a. Experiments have shown C_{E_a} to be almost constant at $1.1–1.2 \times 10^{-3}$ for $U < 18\,\mathrm{m\,s^{-1}}$, for neutral stratification, *i.e.* no positive or negative buoyancy forces acting and at a height of 10 m, written as $C_{E_{10N}}$. However, measurements show large variability in $C_{E_{10N}}$ which may be due to the effects of sea state, such as sheltering in the wave troughs for large waves and increased evaporation due to spray droplets formed in highly forced seas with breaking waves. Results from a field experiment, the Humidity Exchange Over the Sea (HEXOS) experiment in the North Sea, are shown in **Figure 1**. Its purpose was to address the question of what happens to evaporation or water (vapor) flux at high wind speeds. However, the wind only reached $18\,\mathrm{m\,s^{-1}}$ and the measurements showed only weak, if any, effects of the spray. Theories suggest that the effects will be stronger above $25\,\mathrm{m\,s^{-1}}$. More direct measurements are still required before these issues can be settled, especially for wind speeds $> 20\,\mathrm{m\,s^{-1}}$ (*see* Further Reading and the section on meteorological sensors for mean measurements for a discussion of the difficulties of making measurements over the sea at high wind speeds).

Clausius–Clapeyron Equation

The Clausius–Clapeyron equation relates the latent heat of evaporation to the work required to expand a unit mass of liquid water into a unit mass of water as vapor. The latent heat of evaporation is a function of absolute temperature. The Clausius–Clapeyron equation expresses the dependence of atmospheric saturation vapor pressure on temperature. It is a fundamental concept for understanding the role of evaporation in air–sea interaction on the large scale, as well as for gaining insight into the process of evaporation from the sea (or

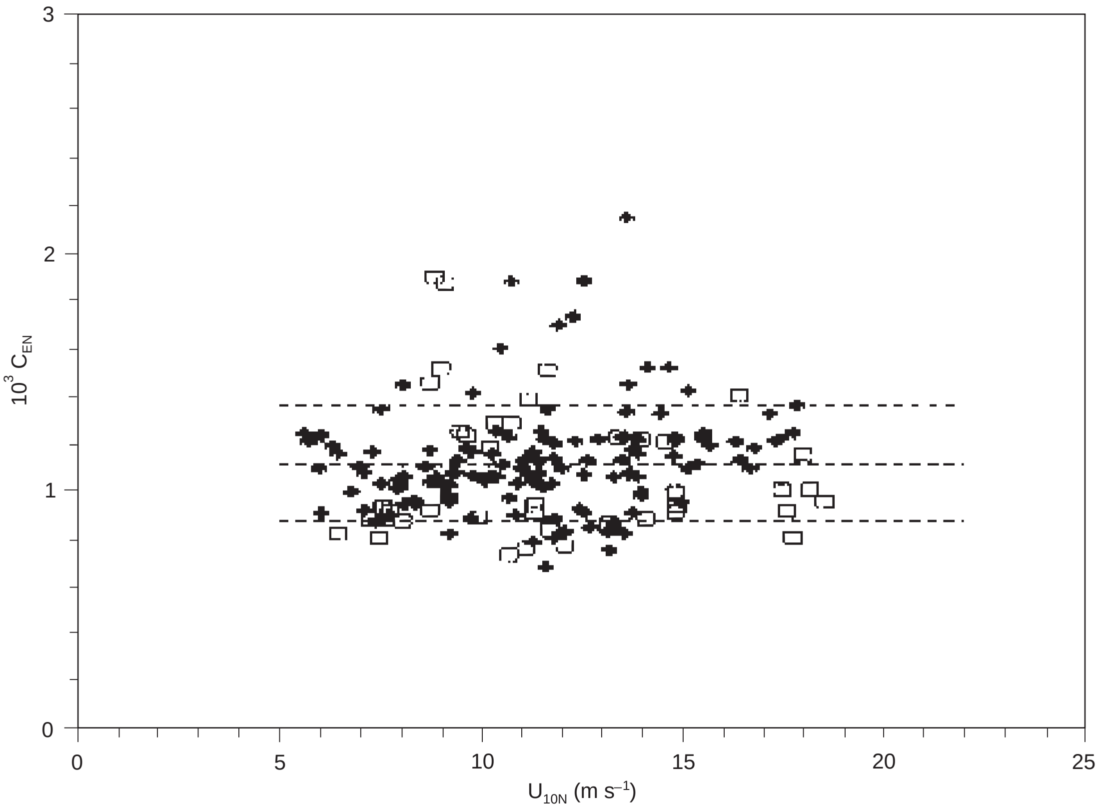

Figure 1 Vapor flux exchange coefficients from two simultaneous measurement sets: the University of Washington (crosses) and Bedford Institute of Oceanography (squares) data. Thick dashed line is the average value, 1.12×10^{-3}, for 170 data points. Thin dashed lines indicate standard deviations (from DeCosmo *et al.*, 1996).

Earth's) surface on the small scale. Note first of all that the Clausius–Clapeyron equation is highly non-linear, viz:

$$\frac{d \ln p_v}{dT} = \frac{\Delta H_{vap}}{RT^2} \qquad [4]$$

where p_v is the vapor pressure, T is absolute temperature (°K), and ΔH_{vap} is the value of the latent heat of evaporation, R is the gas constant for water vapor $= 461.53\,\mathrm{J\,kg^{-1}\ {}^\circ K^{-1}}$. The dependence of vapor pressure on temperature is presented in a simplified form as:

$$e_s = 610.8 \exp\left[19.85\left(1 - \frac{T_0}{T}\right)\right](P_a) \qquad [5]$$

where e_s is vapor pressure in pascals, T_0 is a reference temperature set to $0°\mathrm{C} = 273.16\ °\mathrm{K}$, and T is the actual temperature in °K which is accurate to 2% below 30°C (**Figure 2**).

Figure 2 displays the saturation vapor pressure and the pressure of atmospheric water vapor for 60% relative humidity. On the right-hand side of the figure, the ordinate gives the equivalent specific humidity values (for a near surface total atmospheric pressure of 1000 hpa). This figure illustrates that the atmosphere can hold vastly larger amounts of water as vapor at temperatures above 20°C than at temperatures below 10°C. For constant relative humidity, say 60%, the difference in specific humidity or vapor pressure in the air compared with the amount at the air–sea interface, if the sea is at the same temperature as the air, is about three times at 30°C what it would be at 10°C. Therefore, evaporation is driven much more strongly at tropical latitudes compared with high latitudes (cold sea and air) for the same mean wind and relative humidity as illustrated by eqn [4] and **Figure 2.**

Tropical Conditions of Humidity

By far, most of the water leaving the Earth's surface evaporates from the tropical oceans and jungles, providing the accompanying latent heat as the fuel that drives the atmospheric 'heat engines,' namely,

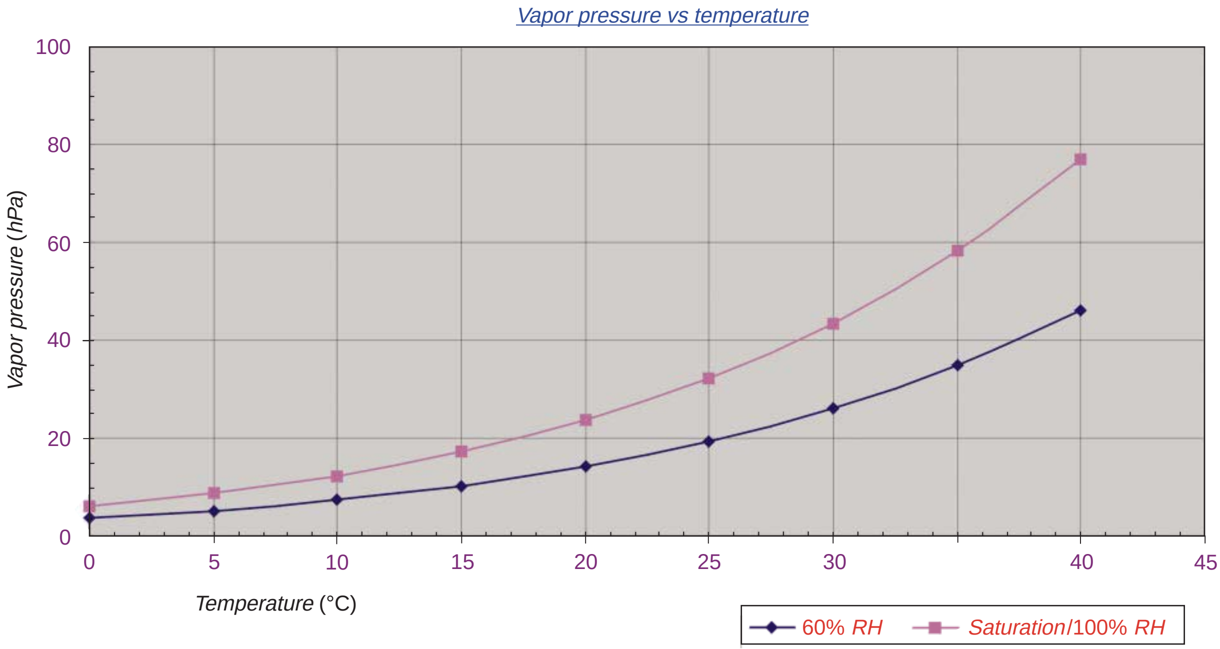

Figure 2 Vapor pressure (hPa) as a function of temperature for two values of relative humidity, 60% and 100%.

thunderstorms and tropical cyclones. Such extreme and violent storms depend for their generation on the enormous release of latent heat in clouds to create the vertical motion and compensating horizontal accelerated inflows. Tropical cyclones do not form over oceanic regions with temperatures $< 26°C$, and temperature increases of only 1° or 2°C sharply enhance the possibility of formation.

Latitudinal and Regional Variations

The Clausius–Clapeyron equation holds the secrets to the role of water vapor for both weather and climate. Warm moist air flowing north holds large quantities of water. As the air cools by vertical motion, contact with cold currents, and loss of heat by infrared radiation, the air reaches saturation and either clouds, storms and rain form, or fog (over cold surfaces) and stratus clouds. The warmer and moister the original air, the larger the possible rainfall and the larger the release of latent heat. Latitudinal, regional, and seasonal variations in evaporation and atmospheric humidity are all related to the source of heat for evaporation (upper ocean heat content) and the capacity of the air to hold water at its actual temperature. Many other processes such as the dynamics behind convergence patterns and the development of atmospheric frontal zones contribute to the variability of the associated weather.

Vertical Structure of Humidity

The fact that the source of moisture is the ocean, lakes, and moist ground explains the vertical structure of the moisture field. Lenses of moist air can form aloft. However, when clouds evaporate at high elevations where atmospheric temperature is low, the absolute amounts of water vapor are also low for that reason.

Thus, when the surface air is continually mixed in the atmospheric boundary layer with drier air, being entrained from the free atmosphere across the boundary layer inversion, it usually has a relative humidity less than 100% of what it could hold at its actual temperature. The exceptions are fog, clouds, or heavy rain, where the air has close to 100% relative humidity. The process of exchange between the moist boundary layer air and the upper atmosphere allows evaporation to continue. Deep convection in the inter-tropical convergence zone brings moist air up throughout the whole of the troposphere, even over-shooting into the stratosphere. Moisture that does not rain out locally is available for transport poleward. The heat released in these clouds modifies the temperature of the air. Similarly, over the warm western boundary currents, such as the Gulf Stream, Kuroshio, and Arghulas Currents, substantial evaporation and warming of the atmosphere takes place. Without the modifying effects of the hydrologic cycle of evaporation and precipitation on the atmosphere, the continents would have more extreme climates and be less habitable.

Sublimation–Deposition

The processes of water molecules leaving solid ice and condensing on it are called sublimation and deposition, respectively. These processes occur over the ice-covered polar regions of the ocean. In the cold regions, this flux is much less than that from open leads in the sea ice due to the warm liquid water, even at 0°C.

At an ice surface, water vapor saturation is less than over a water surface at the same temperature. This simple fact has consequences for the hydrologic cycle, because in a cloud consisting of a mixture of ice and liquid water particles, the vapor condenses on the ice crystals and the droplets evaporate. This process is important in the initial growth of ice particles in clouds until they become large enough to fall and grow by coalescence of droplets or other ice crystals encountered in their fall. Similar differences in water vapor occur for salty drops, and the vapor pressure over a droplet also depends on the curvature (radius) of the drop. Thus, particle size distribution in clouds and in spray over the ocean are always changing due to exchange of water vapor. For drops to become large enough to rain out, a coalescence-type growth process must typically be at work, since growth by condensation is rather slow.

Sources of Data

Very few direct measurements of the flux of water vapor are available over the ocean at any one time. The mean quantities ($\bar{U}, \overline{q_a}$, SST) needed to evaluate the bulk formula are reported regularly from voluntary observing ships (VOS) and from a few moored buoys. However, most of such buoys do not measure surface humidity, only a small number in the North Atlantic and tropical Pacific Oceans do so. The VOS observations are confined to shipping lanes, which leaves a huge void in the information available from the Southern Hemisphere. Alternative estimates of surface humidity and the water vapor flux include satellite methods and the surface fluxes produced in global numerical models, in particular, the re-analysis projects of the US Weather Service's National Center for Environmental Prediction (NCEP) and the European Center for Medium Range Weather Forecasts (ECMWF). The satellite method has large statistical uncertainty and, thus, requires weekly to monthly averages for obtaining reasonable accuracy ($\pm 30\,\mathrm{W\,m^{-2}}$ and $\pm 15\,\mathrm{W\,m^{-2}}$ for the weekly and monthly latent heat flux). Therefore, these data are most useful for climatological estimates and for checking the numerical models' results.

Estimation of Evaporation by Satellite Data

The estimation of evaporation/latent heat flux from the ocean using satellite data also relies on the bulk formula. The computation of latent heat flux by the bulk aerodynamic method requires SST, wind speed ($\bar{U}_{10_N}$), and humidity at a level within the surface layer ($\overline{q_a}$), as seen in eqn [3]. Therefore, evaluation of the three variables from space is required. Over the ocean, $\bar{U}_{10_N}$ and SST have been directly retrieved from satellite data, but $\overline{q_a}$ has not. A method of estimating $\overline{q_a}$ and latent heat flux from the ocean using microwave radiometer data from satellites was proposed in the 1980s. It is based on an empirical relation between the integrated water vapor W (measured by spaceborne microwave radiometers) and $\overline{q_a}$ on a monthly timescale. The physical rationale is that the vertical distribution of water vapor through the whole depth of the atmosphere is coherent for periods longer than a week. The relation does not work well at synoptic and shorter timescales and also fails in some regions during summer. Modification of this method by including additional geophysical parameters has been proposed with some overall improvement, but the inherent limitation is the lack of information about the vertical distribution of q near the surface.

Two possible improvements in E retrieval include obtaining information on the vertical structures of humidity distribution and deriving a direct relation between E and the brightness temperatures (T_B) measured by a radiometer. Recent developments provide an algorithm for direct retrieval of boundary layer water vapor from radiances observed by the Special Sensor Microwave/Imager (SSM/I) on operational satellites in the Defense Meteorological Satellite Program since 1987. This sensor has four frequencies, 19.35, 22, 37, and 85.5 GHz, all except the 22 GHz operated at both horizontal and vertical polarizations. The 22 GHz channel at vertical polarization is in the center of a weak water vapor absorption line without saturation, even at high atmospheric humidity. The measurements are only possible over the oceans, because the oceans act as a relatively uniform reflecting background. Over land, the signals from the ground overwhelm the water vapor information.

Because all the three geophysical parameters, $\bar{U}_{10_N}$, W, and SST, can be retrieved from the radiances at the frequencies measured by the older microwave radiometer, launched in 1978 and operating to 1985 – the Scanning Multichannel Microwave Radiometer (SMMR) on Nimbus-7 (similar to SSM/I, but with 10.6 and 6.6 GHz channels as well, and no 85 GHz channels) – the feasibility of retrieving E directly from the measured radiances was also demonstrated. SMMR measures at 10 channels, but only six channels were identified as significantly useful in estimating E. SSM/I, the operational microwave radiometer that followed SMMR, lacks the low-frequency channels which are sensitive to SST, making direct retrieval of E from T_B unfeasible. The microwave imager (TMI) on the Tropical Rainfall Measuring Mission (TRMM), launched in 1998, includes low-frequency measurements sensitive to SST and could, therefore, allow direct estimates of evaporation rates. **Figure 3** gives an example of global monthly mean values of humidity obtained solely with satellite data from SSM/I.

To calculate q_s, gridded data of sea surface temperature can also be used, such as those provided operationally by the US National Weather Service based on infrared observations from the Advanced Very High Resolution Radiometer (AVHRR) on operational polar-orbiting satellites. The exact coincident timing is not so important for SST, since SST varies slowly due to the large heat capacity of water, and this method can only provide useful accuracies when averages are taken over 5 days to a week. Wind speed is best obtained from scatterometers, rather than from the microwave radiometer, in regions of heavy cloud or rain, since scatterometers (which are active radars) penetrate clouds more effectively. Scatterometers have been launched in recent times by the European Space Agency (ESA) and the US National Aeronautic and Space Administration (NASA) (the European Remote Sensing Satellites 1 and 2 in 1991 and 1995, the NASA scatterometer, NSCAT, on a Japanese short-lived satellite in 1996, and the QuikSCAT satellite in 1999).

Future Directions and Conclusions

Evaporation has been measured only up to wind speeds of $18\,\mathrm{m\,s^{-1}}$. The models appear to converge on the importance of the role of sea spray in evaporation, indicating that its significance grows beyond about $20\,\mathrm{m\,s^{-1}}$. However, the source function of spray droplets as a function of wind speed or wave breaking has not been measured, nor are techniques for measuring evaporation in the presence of droplets well–developed, whether for rain or sea spray. Since evaporation and the latent heat play such important roles in tropical cyclones and many other weather phenomena, as well as in oceanic circulation, there is great motivation for getting this important energy and mass flux term right. The bulk model is likely to be the main method used for estimating evaporation for some time to come. Development of more direct satellite methods and validating them should be an objective for climatological purposes. Progress in the past 30 years has brought the estimate of evaporation on a global scale to useful accuracy.

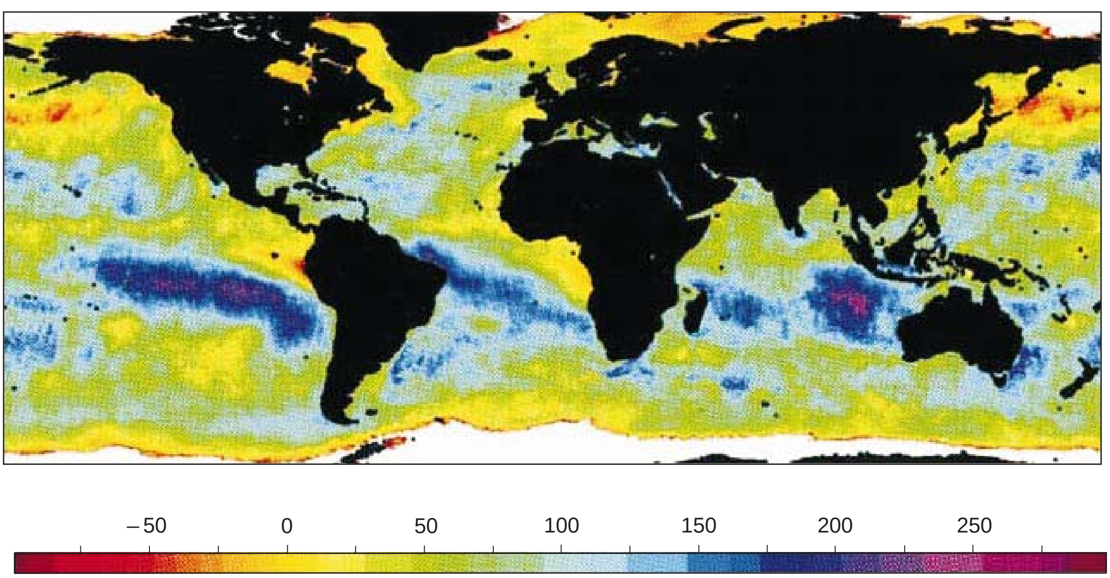

Figure 3 Global distribution of monthly mean latent heat flux in $\mathrm{W\,m^{-2}}$ for September 1987. (Reproduced with permission from Schulz *et al.*, 1997.)

See Also

IR Radiometers. Satellite Remote Sensing of Sea Surface Temperatures. Sensors for Mean Meteorology.

Further Reading

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EXOTIC SPECIES, INTRODUCTION OF

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Introduction

Exotic species, often referred to as alien, nonnative, nonindigenous, or introduced species, are those that occur in areas outside of their natural geographic range. Vagrant species, those that appear from time to time beyond their normal range, may often be confused with exotic species. Because marine science evolved following periods of human exploration and worldwide trade, there are species that may have become introduced, whose identity as either native or exotic species remains unclear. These are referred to as cryptogenic. The full contribution of exotic species among native assemblages remains, and probably will continue to remain unknown, but will add to the diversity of an area. There are no documented accounts of an introduced species resulting in the extinction of native species in marine habitats as has occurred in freshwater systems. Nevertheless, exotic species can result in habitat modifications that may reduce native species abundance and restructure communities. The greatest numbers of exotic species are inadvertently distributed by shipping either attached to the hull or carried in the large volumes of ballast water. Introductions may also be deliberate. The dependence for food in developing countries and expansion of luxury food products in the developed world, has led to increases in food production by cultivation of aquatic plants, invertebrates and fishes. Many native species do not perform as well as the desired features of some introduced organisms now in widespread cultivation, e.g., the Pacific oyster and Atlantic salmon. Unfortunately, production of