

## ATMOSPHERIC WATER

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### Summary

This article is divided into three main parts: water sources; atmospheric circulation and air masses; and anthropogenic enhancement and global and local climate change. The structure of the atmosphere and sources of water are described, and general definitions are provided for atmospheric processes. In the section ‘Atmospheric Circulation and Air Masses,’ consideration is given to scales of motion and the structure of air masses. The final section examines the role of feedback in the climate system and, in particular, it shows that feedback of a rise in water vapor temperature is the process that determines that an increase in global surface temperatures will lead to enhanced evaporation of water vapor from the oceans. Because water vapor is the most important of the greenhouse gases, an increase in its atmospheric concentration will amplify climate warming. It should be pointed out that the Earth’s environment, and especially the atmosphere, has always been subject to change, but that changes are now occurring at a far more rapid rate than at any previously known time in history.

The final part of the article describes current local climate changes, using the Caspian Sea as an example. These results are particularly interesting because they clearly show that local climate changes are already taking place, but they also demonstrate that

modern global climate models (GCM) may be successfully used in the prediction of local as well as global climate changes.

The article is intended for hydrometeorologists, including under- and post-graduate students, as well as specialists from different fields of science and technology, who are engaged in the design and operation of various technical systems in the atmosphere, as well as the assessment of the conditions in which those systems can operate.

## 1. Introduction

The atmosphere is a gaseous envelope surrounding the Earth, held by gravity, having its maximum density just above the solid surface and becoming gradually thinner with distance from the ground, until it finally becomes indistinguishable from the interplanetary gas.

There is no distinct upper limit of the atmosphere. Departing from the surface of the Earth, there may be defined various regions with widely different properties showing a great variety of physical and chemical phenomena (Iribarne J.V., Cho H.R., 1980). It is known that atmospheric mass changes with distance from the Earth. 90% of the mass is contained within the first 20 km (top at 100 mb level), and 99.9% of the mass is contained within the first 50km (top at 1mb level). The atmosphere has a practically constant mass, which is about  $5.157 \cdot 10^{15}$  T (compared with the mass of the Earth being  $5.98 \cdot 10^{21}$ T). The atmosphere is divided vertically into several layers, which are classified according to their temperature distribution, chemical composition, electron density, etc. The classification based on temperature distribution is particularly important.

Starting from the ground and extending up to a certain height, the temperature normally decreases at a rate of 5 to 7 degrees per km. This varies with time and place, and occasionally there even occur shallow layers within which the temperature increases with height: the so-called inversions. The region under consideration is called the troposphere and it is the seat of the weather phenomena that affect the ground. It is also, for obvious reasons, the best-known region and it contains about 4/5 of the total air mass. Its upper limit is defined by a sudden change in the temperature trend, often appearing as a discontinuity in the curve. The temperature stops decreasing more or less suddenly and remains constant or starts increasing slightly. This limit is called the tropopause and its height, also depending on time and place (being higher at the Equator than at the Poles), can vary between 7 and 17 km. The temperature in the tropopause in middle latitudes is  $-50$  to  $-55$  °C.

The next region shows a gradual increase of temperature, reaching a maximum of around 0 °C at 50 km. The region is called the stratosphere and its upper limit (at the temperature maximum) is the stratopause. The temperature then drops again through the region called the mesosphere, to a minimum of  $-100$  °C defining the mesopause, at about 85 km. From the mesopause on the temperature increases steadily, and this region is called the thermosphere. The temperature there reaches high values and then remains constant; at 500 km it may reach values between 400 and 2000 °C, depending on the

time of day, degree of solar activity and latitude; the diurnal variation is 500-800 °C, with a minimum near sunrise and maximum at about 2 p.m.

The physical reasons for this peculiar distribution of temperature in the atmosphere are related to the absorption of radiation. Reactions of ionization and dissociation occur in the upper levels, producing the high temperatures of the thermosphere. The maximum at the stratopause is associated with the presence of ozone. The ground layer is again normally at a maximum due to the absorption at the surface of a large fraction of the solar radiation that reaches that level.

The atmosphere contains about  $1.3 \cdot 10^4 \text{ km}^3$  (2.53 cm conditionally sedimentary layer) or 0.0001% of the total natural water sources on the Earth (Atmosphere Handbook, 1991). Despite a relatively small moisture content in the atmosphere, it is the only source of fresh water regeneration in nature (through evaporation) and the main source of water reserve replenishment (through precipitation). Total evaporation from the ocean surface and continents amounts to  $577,000 \text{ km}^3$  per year. It consumes on the average  $88 \text{ Wt/m}^2$  of heat, which amounts to more than a third of the solar energy supply of the Earth. Atmospheric water, transforming from one state into another, participates in water circulation in nature. Indeed, the entire  $577,000 \text{ km}^3$  of water in the atmosphere precipitates to the Earth. Meridional water vapor transfer is a significant peculiarity of water circulation.

## 2. Water Sources

Water appears in the atmosphere mainly due to its evaporation from the ocean surface and the Earth. Evaporation immediately from green plant surfaces (transpiration) and ice surfaces (sublimation) also increases  $\text{H}_2\text{O}$  content in the atmosphere. Water vapor has an average lifetime of about ten days in the atmosphere and can move thousands of kilometers before condensation occurs.

Moisture transfer is characterized by two principal processes: exchange among latitude bands (meridional moisture exchange) and exchange between land and sea. Water evaporated from subtropical ocean waters is transported to the equator by the dominant surface winds of the Hadley circulation. Near the equator, the surface winds in the northern and southern hemispheres converge in the intertropical convergence zone (ITCZ), thereby causing strong vertical convection and heavy precipitation. As a consequence, the tropical areas remove more water from the atmosphere than they supply. The land – sea moisture exchanges are in balance even though the sea loses more water by evaporation than it receives in precipitation and the land receives more by precipitation than it loses by evaporation. The balance is maintained by river and groundwater flows to the oceans.

When unsaturated air is cooled, it eventually reaches the point of saturation. At that point (the dew point), any further cooling results in the deposition of water vapor onto convenient surfaces, generally being the very small atmospheric particles that serve as condensation nuclei. The hydrometeors that are formed by this condensation process (a hydrometeor is any small condensed form of water falling through or suspended in the

air) grow by further water vapor accretion. When sizes on the order of 10-20 $\mu$ m are reached, the available water vapor is generally depleted and a stable cloud is formed.

An elementary fact of physics is that energy changes are involved in the transformation of water from one state to another. In the case of evaporation, this energy (the heat of vaporization) is 2260 Jkg<sup>-1</sup> at 25 °C. The energy is primarily derived from solar energy retained in water vapor and known as latent heat. It is released to the atmosphere upon condensation and is a very important part of the atmospheric heat balance. Because of the release of latent heat, clouds are warmer than the surrounding cloud-free air, and their buoyancy is enhanced, as may be noticed when one flies through clouds in an airplane. Just as energy is required for water to evaporate, so it is given up when the water vapor condenses. To maintain the heat balance, a layer of water averaging 100 cm per year is evaporated from the surface and subsequently condensed in the atmosphere to produce precipitation.

Condensation occurs when air, being saturated with water vapor, becomes more moistened, and when a relative moisture amounts to 100%. Condensation can be in the form of dew or, if temperature descends below 0 °C, rime.

Condensation in the atmosphere occurs on condensation nuclei, which are small particles (dust, salt crystals, fire smoke, etc.). Water vapor condensation in the atmosphere results in the formation of clouds. These are classified according to the form and state (water drops or ice crystals) into types, kinds and varieties. Moisture is discharged out of the atmosphere due to precipitation in the form of rain, snow, hail, drizzle, sleet, etc. Three main types of precipitation are singled out. Orographic precipitation is caused by vertical air movement arising when flowing about uneven relief in the land. Precipitation from extra-tropic cyclones is related to the migration of low-pressure zones in the middle and high latitudes, and air mass precipitation is caused by perturbations within homogeneous air masses. Air mass precipitation is characteristic for continental mean latitudes in summer.

Fog is the accumulation of the smallest water drops or ice crystals decreasing visibility near the surface air layer. There are fogs of cooling and evaporation. Radiation fog can be formed when there are no clouds and due to the cooling of the land surface by long-wave heat radiation at night. Advection fog can be formed due to the horizontal movement of a relatively warm air mass over a colder subjacent land surface. Evaporation fogs are observed over the water surface and at the interface between two air masses with different temperatures and moisture contents (frontal fog).

### **3. Atmospheric Circulation and Air Masses**

#### **3.3. The Scales of Motion**

Generally speaking, the motion of the atmosphere is governed by the equation of motion, the continuity equation, and the laws of thermodynamics (Graedel T.E., Crutzen P.J., 1993). Circulation in the atmosphere can be considered as a particular solution in the governing equations. These equations are all nonlinear, partial differential equations. Their solutions under a given set of physical and boundary

conditions often cannot be easily obtained. In order to understand the physical and dynamic nature of atmospheric circulation, it is frequently necessary to first classify, according to a certain criterion, the various types of circulation systems observed in the atmosphere. A particular type of circulation can then be isolated and its dynamical nature be examined with the aid of the dynamic equations. Once a better understanding of the nature of the circulation systems is obtained, certain approximations can be introduced so as to make the equations easier to solve. The motion systems that occur in the atmosphere may be classified in various ways. One method of classification that has proven to be very useful is based on the time and distance scales by which the motion systems may be recognized. Atmospheric motion is often composed of a spectrum of circulation systems of different time and length scales. The time scale is usually related to its length scale, the larger the length scale, the longer its time scale. The largest circulation systems of the atmosphere have a length scale comparable to the diameter of the Earth itself. The smallest circulations have a length scale comparable to the mean free path of the individual molecules.

According to the length scale, the spectrum of atmospheric motions can be divided into the planetary-scale motions, synoptic-scale motions, meso-scale motions, and small-scale motions. The boundaries between these subdivisions are not well defined, for the spectrum of atmospheric motions is continuous. Nevertheless, the motion systems in each subdivision have their own distinct dynamic features. Different approximations can be introduced into the dynamical equations for motion systems in each subdivision. This classification has been a very useful conceptual tool in the study of atmospheric dynamics. Planetary-scale motions include circulation systems with a horizontal scale comparable to the dimensions of the Earth. The mean global circulations and the features that vary with zone and are of a length scale comparable to the major continents and oceans are examples of motion systems of this type.

Synoptic-scale motions have horizontal scales smaller than those of planetary-scale motions, yet large enough to be resolved by a conventional observation network. The spacing of stations in such a network is on the order of several hundred kilometers. Most circulation systems responsible for day-to-day weather changes are systems of this category. Motion systems with a horizontal scale on the order of  $10^1 - 10^2$  kilometers are called mesoscale motion systems. Mountain lee waves, thunderstorms, squall lines and hurricanes are a few examples of motion systems of this type. Circulations with horizontal dimensions smaller than those of mesoscale motions are called small-scale motions. Small cumulus clouds and convective and mechanical turbulent eddies near the surface of the Earth are typical examples. Motion systems of this type play a very important role in the dynamics of the lowest kilometer of atmosphere. There are many other ways of classifying atmospheric motions. Depending on the degree of regularity, a flow field may be called either a laminar flow or a turbulent flow. Some air motions are induced mainly by the pressure gradient force, while others are induced by the buoyancy force. Air motions can also be generated by some instability mechanism. Some of the instability processes are mostly thermal in nature, while others are mainly mechanical in nature, such as the shear instability often observed very close to the Earth's surface.

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### Biographical Sketch

**Gennady N. Panin** graduated from the Geographical Faculty of Moscow State University in 1967. He received his Candidate of Sciences degree (equivalent to Ph. D.) in evaporation of the oceans in 1973 at the Institute of Oceanology, Russian Academy of Sciences and the Doctor of Sciences degree in air-sea interaction in 1986 at the Institute of Water Problems, Russian Academy of Sciences. Since 1989 he has

held a position as head of the Laboratory of Evaporation and Humidity Transport at the Institute of Water Problems, Russian Academy of Sciences in Moscow. His research interests are in the fields of surface fluxes; climate change; parametrization of surface fluxes; the theory of interaction between the water/land surface and the atmosphere. His main publications are in the leading international journals on physics of atmosphere and oceans, and water resources. He has had 127 publications including four books:

- 1) Panin G.N.- " Heat- and mass exchange between the water and the atmosphere in the nature."- 1985. Nauka, Moscow, 206 p.(in Russian),
- 2) Panin G.N.-" Evaporation and heat exchange of the Caspian Sea."- 1987. Nauka, Moscow, 89 p. (in Russian).
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During the past ten years he has held a number of short- and long-term research visiting positions in various European universities and scientific institutions.