HADLEY CIRCULATION

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Introduction

The so-called 'Hadley circulation' is perhaps the earliest attempt to account for the global-scale distribution of winds in the Earth's atmosphere in terms of basic physical processes. Halley in 1685 and Hadley in 1735 both proposed that the 'Trade Winds' that blow toward the Equator at low latitudes could be understood as the lower branch of an axially symmetric convection cell driven by the temperature difference between the Equator and poles of the Earth. Their ideas were ahead of their time, especially as there was then no prospect of determining winds at upper levels of the atmosphere and thus verifying their hypothesis. When routine upper-air observations became available in the mid-twentieth century, the ideas of Halley and Hadley were essentially confirmed. Today, the term 'Hadley circulation' refers to the thermally driven meridional overturning motions in the low-latitude troposphere.

Figure 1 is the traditional, and somewhat oversimplified, schematic view of the global atmospheric circulation that will be found in many elementary text books. It divides the Earth into a set of climate zones, with the Trade Wind regime confined to the tropics. The Trade Winds are simply the low-level part of the overturning 'Hadley circulations', with ascent near the Equator, descent in the subtropics and a poleward return flow at upper levels. The more disturbed midlatitudes are characterized by generally westerly winds, with irregular growing and decaying eddies, the cyclonic and anticyclonic weather systems generated by baroclinic instabilities. When averaged around entire latitude circles, this turbulent midlatitude flow averages to a weak 'Ferrel circulation', in which warmer air at lower latitudes sinks and colder air at high latitude rises. There is some evidence of a very weak 'polar cell' at high latitudes.

The energy that drives the Hadley circulation comes from the conversion of heat energy to mechanical energy in the tropical atmosphere: the Hadley circulation is a prototypical example of a thermodynamic 'heat engine'. Such heat engines are ultimately responsible for maintaining all motions in the atmosphere against the dissipative effects of friction. The operation of the atmospheric heat engine is shown in **Figure 2**, which is a classic thermodynamic diagram in

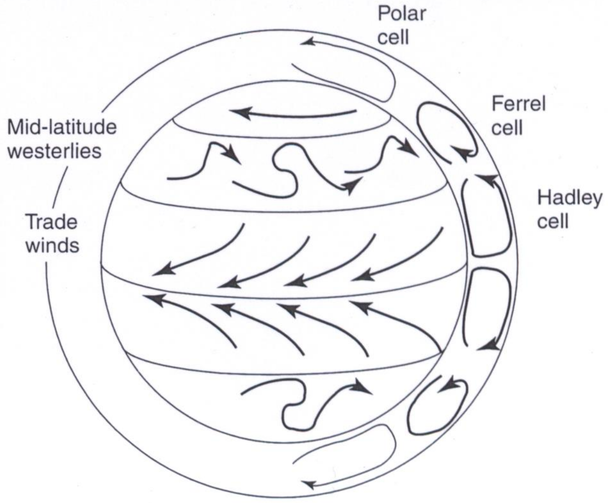

Figure 1 A schematic view of the mean circulation of the troposphere. The arrows on the globe show the winds near the Earth's surface. The circulations at the side show the zonal mean circulation cells at various latitudes.

which temperature is plotted against specific entropy. The thermodynamic state of an air parcel – that is, its temperature, pressure, density and so on – are represented by a point on the thermodynamic diagram, and any change of its thermodynamic state by a curve on the diagram. The area under the process curve is

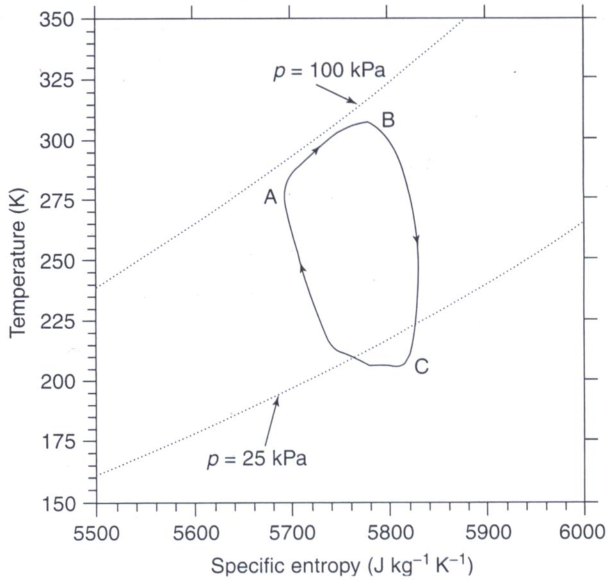

Figure 2 A schematic thermodynamic diagram for the Hadley circulation.

proportional to the heat energy entering an air parcel. The diagram also shows two different lines of constant atmospheric pressure, one near the Earth's surface and one in the tropical upper troposphere. Near the surface, air flows toward the Equator, along the segment marked AB, gaining heat from the surface (this may be in the form of direct or sensible heat, or in the form of latent heat as water evaporates into the air). Near the Equator, it rises almost adiabatically (that is, with little heat entering or leaving the air) along the segment BC. It then moves poleward along segment CA, cooling (that is, losing entropy) by emitting infrared radiation to space and descending. Eventually it returns to its original location A. During this cyclic process, more heat is added to the air along AB than is removed along CA. The excess heat is converted to mechanical energy associated with the circulation of the tropical air.

The condition for such an energy conversion to take place is that heat should on average be added at higher pressure than it is removed. Equivalently, one can say that air must rise on average when it is warmer, and descend when it is cooler. A circulation with these properties is called a 'thermally direct circulation'. A thermally indirect circulation, in contrast, must be driven by a source of mechanical energy; a refrigerator cycle is an example of such a thermally indirect circulation. In the schematic diagram of **Figure 2**, the Hadley circulation is thermally direct, and therefore generates mechanical energy. In contrast, the Ferrel circulation of midlatitudes is thermally indirect and consumes mechanical energy.

The observed annual mean meridional circulation is shown in **Figure 3**. The contours are parallel to the northward and upward winds averaged around latitude circles and in time. The contour values have been scaled to have units of kg s^{-1}. They may be thought of as denoting the mass flux across a line from the edge of the plot to that point. The most striking feature is the strong rising motion near the Equator, and sinking motion at latitudes of about 25° N and S, defining two overturning cells, the 'Hadley cells', one in each hemisphere. However, the actual winds associated with these circulations are not particularly strong: they barely exceed 5 m s^{-1}. The Hadley cells are thermally direct. Weaker, thermally indirect Ferrel cells are seen at higher latitudes. Considerably weaker thermally direct polar cells are seen at high latitudes.

The diagram also reveals that there is a close relationship between the westerly component of the wind, shown by the shading, and the meridional flow. The westerly component is much stronger, with values up to 40 m s^{-1}. These maximum winds, the so-called 'subtropical jet', are found in the upper troposphere, just where the circulations associated with the Hadley cells meet those associated with the Ferrel cells. There is also a close relationship between the zonal winds and the temperature fields: they are linked, to a very good approximation by the thermal wind relationship, which can be written as eqn [1].

$$\frac{\partial[\bar{u}]}{\partial p} = \frac{R}{pf}\frac{\partial[\bar{T}]}{\partial y} \qquad [1]$$

That is, a strong vertical wind shear is associated with a strong poleward temperature gradient. In the deep tropics where the Coriolis parameter f is small, this relationship indicates that the temperature gradients must be small, whatever the wind field. But in the subtropics and midlatitudes, the increasing westerly wind with height is associated with the fall of temperature toward the poles.

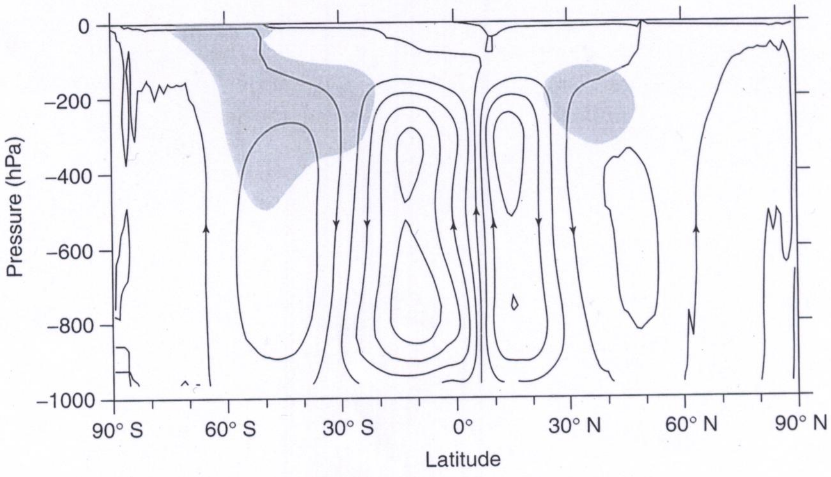

Figure 3 The annual mean meridional streamfunction. Contour interval 2×10^9 kg s^{-1}. Shading shows zonal winds greater than 20 m s^{-1}. Based on an analysis of 20 years of ECMWF analyses.

The Held–Hou Model

An elegant model due to Held and Hou gives considerable insight into the Hadley circulation and the factors that determine its extent. **Figure 4** illustrates the model. The atmosphere is divided into two layers. The lower layer is affected by friction at the ground, and flow within it is supposed to be generally small. Friction is effectively zero in the upper layer and so at this level rings of air conserve their angular momentum as they move poleward. Assuming that such rings start at the Equator with zero motion relative to the solid Earth, the wind at higher latitudes in the upper layer is given by eqn [2], where Ω is the rotation rate of the Earth, a is the radius of the Earth, and y is the distance from the equator, proportional to the latitude.

$$U_M = \frac{\Omega}{a} y^2 \qquad [2]$$

Using the principle of thermal wind balance in the form of eqn [3], the formula for U_M can be used to predict the variation of temperature with latitude, $\theta_M(y)$.

$$\frac{\partial u}{\partial z} = \frac{ga}{2\Omega y T} \frac{\partial T}{\partial y} \qquad [3]$$

This is to be compared with the hypothetical 'radiative equilibrium' temperature distribution $\theta_E(y)$ of an atmosphere that is not permitted to circulate. Where the actual temperature is less than radiative equilibrium there is net heating, and vice versa. In a steady state, this heating and cooling should exactly balance in the Hadley circulation and this requirement fixes the meridional extent and strength of the Hadley circulation.

Figure 5 illustrates a graphical solution of the Held–Hou model. The actual temperature varies very little with latitude in the tropics but drops rapidly in the subtropics and mid-latitudes. The radiative equilibrium temperature has a maximum at the Equator. The temperature on the equator is set by requiring that there be no net heating of air parcels as they circulate, that is, that the two shaded areas must be equal. The poleward limit of the Hadley circulation is at the latitude where these curves cross for the second time. A formula for the distance of the poleward edge of the Hadley cell from the equator results from this solution (eqn [4]).

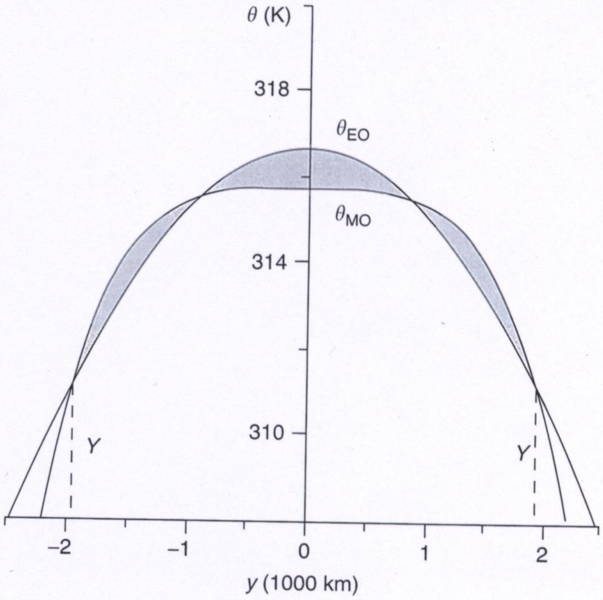

Figure 5 Solution of the Held–Hou model.

$$Y = \sqrt{\frac{5gH\,\Delta\theta}{3\Omega^2\theta_0}} \qquad [4]$$

This formula suggests a value for Y of about 2500 km, in remarkably good agreement with observations considering the simplicity of the model.

The model can be elaborated. For example, the vertical motions, proportional to the heating in the regions of ascent and descent, can be estimated. The model predicts a vertical circulation that is rather weaker than that observed. The effect of latent heat release in cumulonimbus clouds, which leads to intensified but narrower regions of ascent, and broad regions of descent, can be represented. But the basic physics, which predicts that the Hadley circulations are confined to 2500 km or so of the equator, remains relevant.

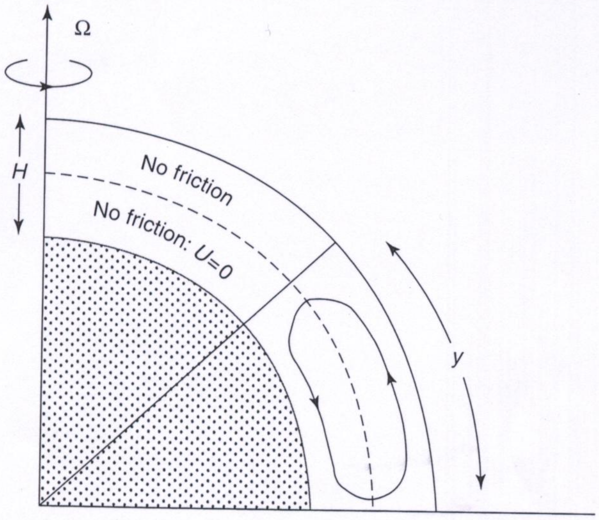

Figure 4 The configuration of the Held–Hou model.

Seasonal Effects

The annual mean circulation shown in **Figure 3** is in fact the average of two quite different circulation

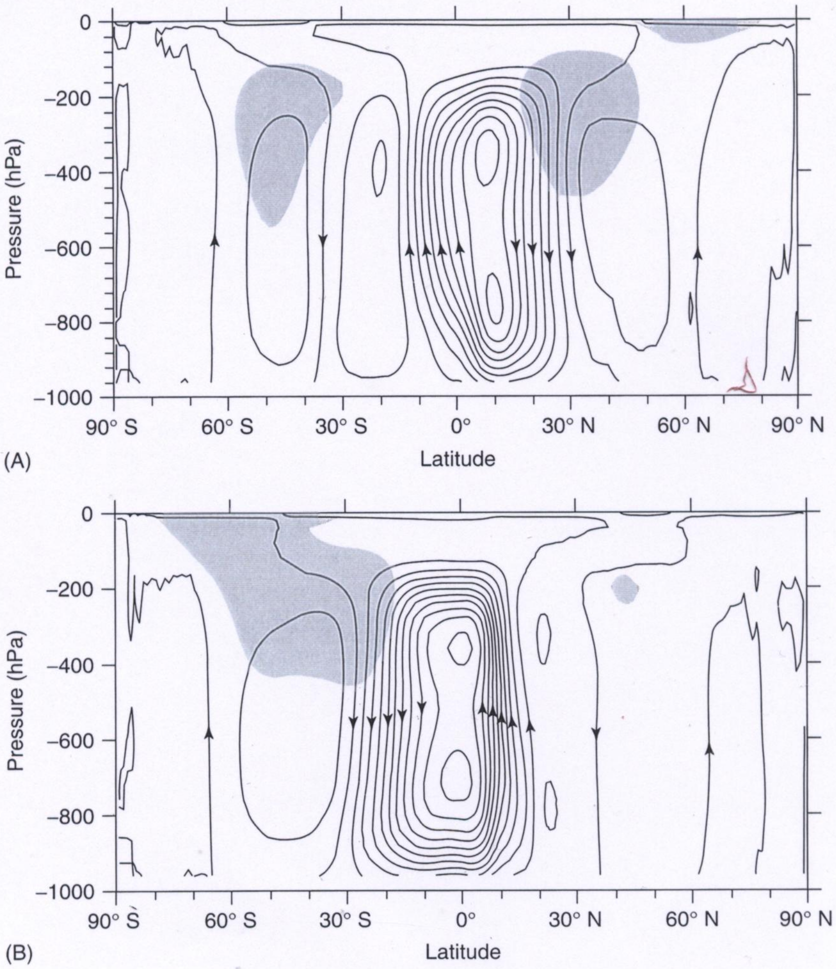

Figure 6 The mean meridional circulation for (A) the December–January–February season and (B) the June–July–August season. Other details are as for **Figure 3**.

regimes that persist around the solstices. **Figure 6** shows the circulation for the mean Northern Hemisphere winter and summer seasons. In both cases, there is a single strong thermally direct Hadley cell with rising motion in the summer hemisphere and descent in the winter hemisphere. Weaker, thermally indirect Ferrel cells are seen at middle latitudes in both hemispheres, but there is little or no sign of a Hadley cell in the summer hemisphere. Looking at the mean circulation for shorter periods reveals that the transition between a circulation like that of **Figure 6A** and one like that of **Figure 6B** is quite abrupt. At most times, there is just a single tropical Hadley cell whose circulation links the two hemispheres: at some point in the spring and autumn its direction of circulation switches abruptly as the temperature maximum crosses the Equator.

The Held–Hou model can be adapted to the situation where the heating is not symmetric about the Equator. Assume that the maximum radiative equilibrium temperature is no longer at the Equator, but at some latitude y_0. As well as the latitude of the northern and southern edges of the Hadley cells, the latitude y_c of the streamline that divides circulation into the summer and winter hemispheres, and which is not the same as y_0, must be determined. The algebra is more complicated, but the steps in the argument are just the same as for symmetric Hadley cell described in the previous section.

Figure 7 shows the results. For even small y_0, the summer cell shrinks drastically and the winter cell intensifies. Almost all the circulation is associated with ascent in the summer hemisphere and with descent in the winter. The strength of the circulation is indicated by the area between the temperature curves and the radiation equilibrium curve. For y_0 of only 500 km, the winter cell has intensified by a factor of about 10 compared to the symmetric case, while the summer cell has weakened by a similar factor. The winter cell is therefore some 100 times as intense as the summer cell. Such a highly nonlinear response to the latitude of the heating maximum means that the annual mean

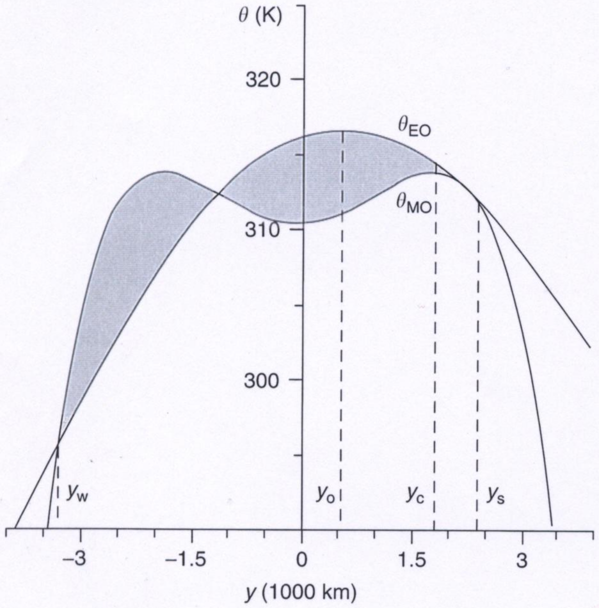

Figure 7 As **Figure 5**, but for a situation in which the heating maximum is located away from the Equator. y_c is the latitude of maximum radiative equilibrium temperature, y_c is the latitude dividing the winter and summer Hadley circulation cells, y_w and y_s designate the limits of the winter and summer Hadley cells, respectively.

meridional circulation is much more intense than the circulation derived from the annual mean heating. This is a particularly pointed example of the problem of 'nonlinear averaging', which is ubiquitous in the study of climate. This result also reconciles the weak circulations of the Held–Hou model with the stronger observed circulation: we should interpret the annual mean circulation as the average of the two solsticial circulations, not as the response to the annual mean thermal forcing.

A Lagrangian View

The diagrams of the meridional circulation shown so far have all been based on so-called 'Eulerian averages'. That is, the winds have been averaged at fixed points in space to produce the time-mean, zonal-mean circulation. At all points in space, the winds and temperatures fluctuate to some degree as weather systems pass across the observing site. An alternative is to follow individual elements of fluid as they move around in the atmosphere, and average their properties to define a mean circulation. Such a mean is called the 'Lagrangian mean', and in many ways is a much preferable way to describe the circulation. For example, the laws of physics applied to the atmosphere all refer to the properties of discrete, identifiable lumps of fluid. However, the Lagrangian mean is very difficult to calculate in practice, not least because individual elements of fluid rapidly become distorted and eventually thoroughly mixed with neighboring elements.

An approximation to the Lagrangian meridional mean circulation can easily be calculated, and is shown in **Figure 8**. In constructing this diagram, the wind data were averaged not on surfaces of constant pressure (as in **Figures 2–7**) but on surfaces of constant 'potential temperature'. The potential temperature of an air parcel generally remains more or less constant for periods of less than a few days. It follows that surfaces of constant potential temperature move up and down in response to the movement of the air. Averaging on potential temperature surfaces is equivalent, to the degree that potential temperature is indeed conserved, to taking the Lagrangian average.

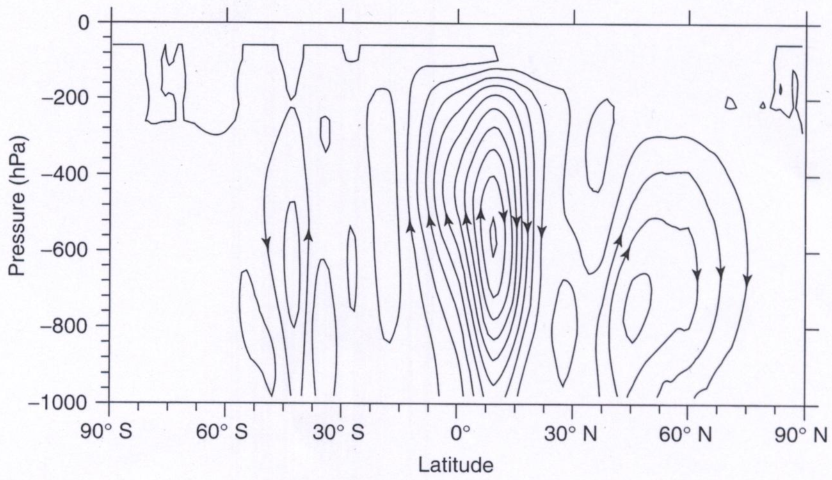

Figure 8 The mean meridional circulation for the December–January–February season, but with the data zonally averaged on surfaces of constant potential temperature rather than on surfaces of constant pressure. Other details are as for **Figure 3**.

Figure 8 differs dramatically from the corresponding Eulerian mean circulation shown in **Figure 6B**. The tropical Hadley cell is still present, but the mid-latitude, thermally indirect Ferrel is more or less eradicated. Instead, a thermally direct circulation extends all the way from the tropics to the pole in the winter hemisphere. The original picture of the global circulation suggested by Halley and Hadley is largely vindicated if one views the circulation in Lagrangian terms.

The thermally indirect Ferrel cell actually transports heat against the temperature gradient, from high latitudes to low latitudes. At the same time, fluctuations in the flow (often termed 'eddies') more than compensate by transporting heat down the temperature gradient, from low latitudes to high. In fact, the partitioning of the flow into mean and eddy parts is arbitrary. The Lagrangian circulation, dominated by thermally direct circulations at nearly all latitudes, is a more natural and less arbitrary description. In the Lagrangian view, thermally direct Hadley circulations dominate the large-scale tropospheric circulation. The Lagrangian view is remarkably like that originally suggested by Halley and Hadley.

See also

Boundary Layers: Convective Boundary Layer, **Coriolis Force**. **General Circulation:** Energy Cycle. **Lagrangian Dynamics**.

Further Reading

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HAIL AND HAILSTORMS

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Introduction

Hailstones are balls of ice that typically fall from cumulonimbus clouds. By convention, they must be greater than 5 mm in diameter but their composition, size, and shape are variable. The largest hailstones can have longest dimensions of 15 cm or more.

Hailstone amounts are also highly variable, but generally the largest hailstones and heaviest hailfalls are from the most severe storms; that is, storms with the strongest updrafts, tallest cloud tops, and largest size. Thus hail is correlated with tornadoes, and also with lightning, though many storms produce lightning with no hail at the ground. Hail is not as well correlated with flooding, which often results from long-lasting and slow-moving precipitation systems that do not produce hail and may not have exceptionally strong updrafts.

Hailstones include various amounts of air bubbles, often in layers that indicate growth stages, but when larger than about 2 cm in diameter their densities are usually within 5% of that of solid ice, $0.91\,\mathrm{g\,cm^{-3}}$. However, hail may be slushy, containing significant amounts of liquid water, and, especially at small sizes, the air content may be great enough that the hail is soft. Soft hail is distinguished from graupel (accumulations of rime on snow particles or small frozen water drops) only by size, and since nearly all hail falls through a thick layer of air above the freezing point before reaching the ground, soft hail is often slushy, because of melting, when it falls. Much rainfall from cumulonimbus clouds in temperate climates is melted graupel and small hail.

Fundamental Concepts of Hail Formation

Hail forms by the accretion of water droplets onto ice particles falling through supercooled cloud. The basic elements needed for understanding the principles of hail formation are as follows.

The Updraft and its Consequences

Humid air rising in the cores of cumulus clouds cools as it rises. The cooling causes the condensation and