Water in the atmosphere

Bjorn Stevens and Sandrine Bony

Much of what we know, and even more of what we don't know, about Earth's climate and its propensity to change is rooted in the interplay between water, air circulation, and temperature.

Every schoolchild learns about the role of the atmosphere in Earth's water cycle. But few get the chance to learn about water's role in determining the properties of the atmosphere. Water determines not only how the Sun's energy is partitioned through the atmosphere and across Earth's surface but also the character of the large-scale circulations, which that energy drives. Recognition of the fundamental influence of water even predates formal scientific thinking. In the Judeo-Christian creation myth, for instance, one of the creator's first tasks, after separating darkness from light, was to separate water from water to create the sky.

That water assumes such a defining place in the sky is remarkable given that it only accounts for 0.25% of the total mass of the atmosphere. That's the equivalent of a liquid layer only 2.5 cm deep, barely enough to make a global puddle, distributed through the atmosphere almost entirely (99.5%) in the form of vapor. By way of comparison, the global ocean, if spread uniformly over Earth's surface, would have an average depth of about 2.8 km. Fresh water on Earth's terrestrial surface—ice sheets, lakes, rivers, wetlands, and soils—is 2000 to 3000 times more abundant than atmospheric water. No matter how you look at it, being suspended in the atmosphere is an exceedingly unlikely state for a water molecule to find itself in; but while in that state, water makes a world of difference.

An absorption virtuoso

Water stands out because of its physical and radiative properties. As figure 1 shows, it is a small molecule with a large appetite for IR radiation. The water molecule is endowed with a plethora of rotational absorption modes, which result from the tumbling of its strong electric dipole around three small and disparate moments of inertia. These modes contribute to a rich set of spectral lines that stretch from the near-IR into the microwave. Some of them arise because the rotational modes ornament three vibrational modes that form the fundamental rotational–vibrational (ro-vibrational) bands. One, $\lambda_2 = 6.3\ \mu$m, is associated with H-O-H bending; the other two, λ_1 and λ_3, associated with symmetric and asymmetric stretching, are located near 2.7 μm and overlap with an overtone of the bending mode.

Bjorn Stevens directs the Max Planck Institute for Meteorology in Hamburg, Germany, and **Sandrine Bony** is a CNRS senior scientist at the Laboratoire de Météorologie Dynamique in Paris.

ERIK CHRISTENSEN, PORKERI

Such a surfeit of absorption lines also endows the water molecule with a potent continuum absorption. Debate continues about the relative contributions to that continuum from the overlap of the wings of different spectral lines and from the transient production of molecular clusters of water dimers during collisions in the vapor (see PHYSICS TODAY, April 2013, page 18). Whatever the cause, the continuum absorption is an important manifestation of water vapor's unusual effectiveness at interacting with radiation throughout the thermal IR.[1] This effectiveness is felt also in the near-IR and makes water vapor the most important absorber of solar radiation in the lower atmosphere. Other molecules found in the atmosphere also have strong, or well-placed, absorption features in the IR. But when it comes to interacting with the full spectrum of IR radiation, including overtones at shorter wavelengths, none approaches the virtuosity of the water molecule. (See the article by Raymond Pierrehumbert, PHYSICS TODAY, January 2011, page 33.)

Condensed water, which atmospheric physicists refer to collectively as hydrometeors, can take various forms, such as the simple crystals, droplets, snowflakes, graupel, and hail illustrated in figure 1. The extent to which they scatter electromagnetic radiation depends on their refractive index and their size relative to the radiation's wavelength: Shorter waves are preferentially scattered, longer waves are absorbed. Because the typical size of cloud droplets (and to a lesser extent cloud ice) is commensurate with the shorter wavelengths of the thermal IR, clouds are effective at absorbing energy at these and longer wavelengths.

They are much less effective in doing so at the 10- to 100-fold shorter wavelengths found in the solar part of the spectrum, where the radiation is instead scattered. As a result, clouds exhibit both a strong greenhouse effect and a strong planetary albedo, despite containing only about 0.5% of the atmosphere's water.[2] As only a small fraction of that water is distributed in large, precipitating hydrometeors such as rain and snow, which have relatively small surface-to-volume ratios, precipitating water is far less important for Earth's radiative budget than water vapor or suspended condensate.

Thermodynamics and phase changes

Because water cycles through vapor and condensate phases in the atmosphere, the laws of thermodynamics place important constraints on the coupling between water vapor and air temperature. Through the Clapeyron equation, the second law of thermodynamics dictates how the saturation vapor pressure e_s depends on temperature T and fundamentally constrains the humidity structure of the atmosphere. The equation can be expressed as

$$d \ln e_s = \frac{\beta}{T} \, d \ln T. \tag{1}$$

The factor β is roughly equal to the ratio of the enthalpy of vaporization—the energy required to transform water from liquid to gas at constant pressure—to the water vapor gas constant. Its value (about 5400 K) expresses the strength of the effect of temperature variations on saturation vapor pressure. Because of water's unusually large enthalpy of vaporization, β is a factor of two or three larger than that of other common condensable vapors such as carbon dioxide, methane, and ammonia.

The large vaporization enthalpy also implies that at typical surface temperatures, e_s approximately doubles for every 10-K rise in temperature. If the vapor pressure e rises above e_s—for instance, as a result of expansional cooling or a radiant loss of

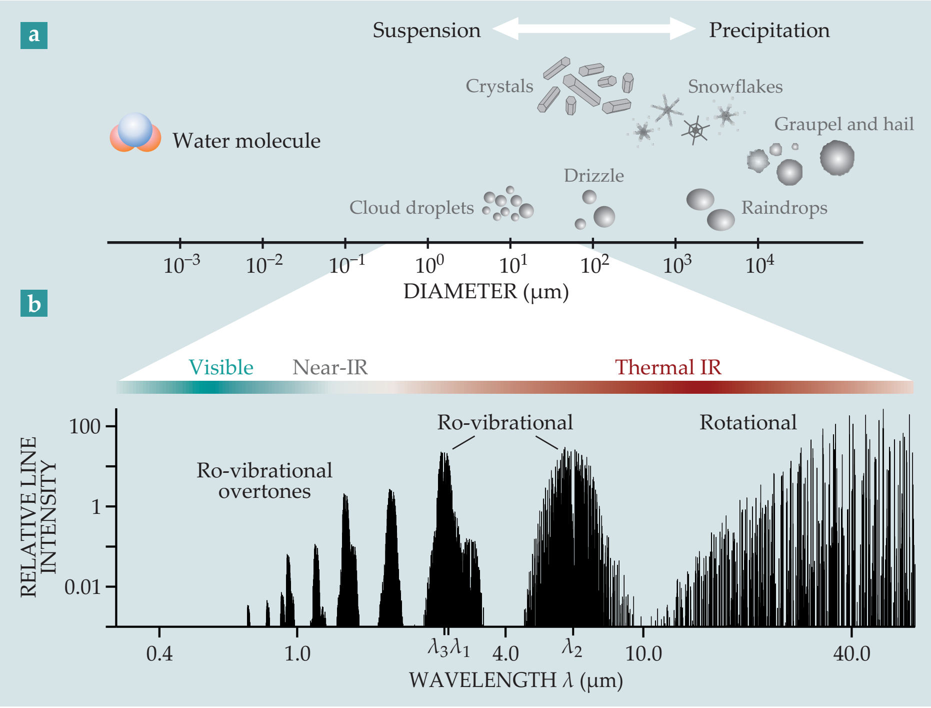

Figure 1. Hydrometeors (a), the condensed forms of water in the atmosphere, come in several sizes. They mostly scatter visible light but absorb over a broad range of the IR. **(b)** The near- and thermal-IR regions of the spectrum excite the molecule and produce its rotational–vibrational (or ro-vibrational) and rotational bands. Specific lines λ_1, λ_2, and λ_3 mark the symmetric stretching mode, bending mode, and asymmetric stretching mode, respectively.

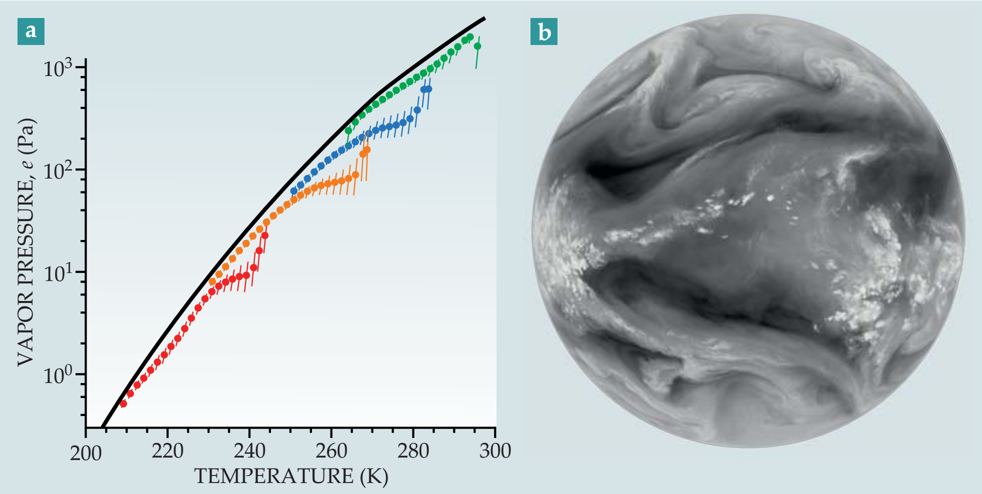

Figure 2. The atmospheric vapor pressure as a function of temperature **(a)** is bounded by the atmosphere's vapor pressure when saturated with water (solid line). The data are shown as the median (dot) and range (error bar) of 228 monthly values at different isobaric levels—900 hPa (green), 700 hPa (blue), 500 hPa (orange), and 300 hPa (red). (Data are provided by the European Centre for Medium-Range Weather Forecasts.) **(b)** Based on satellite measurements of radiation emitted at 6.2 µm, this map reveals the relative humidity of the middle and upper troposphere. It also reveals how closely humidity levels are tied to atmospheric circulation patterns: Pronounced dry regions (dark) show large-scale descending motions of air concentrated in the subtropics. Clouds, regions of complete saturation, are visible as white features. (Image from Meteosat.)

energy—condensate will generally form to maintain the ambient vapor pressure near its saturation value. For short-lived and dilute condensate clouds, the condensate largely follows the atmospheric flow until it evaporates again as the temperature rises or the vapor pressure falls—for instance, through mixing with drier air. For clouds that are sufficiently long-lived or condensate laden, larger hydrometeors such as rain or snow form and rapidly precipitate from the atmosphere.

The first law of thermodynamics dictates how temperature changes for adiabatic displacements of air parcels and thus fundamentally constrains the thermal structure of the atmosphere. Because air has mass, its pressure decreases with altitude. So as a parcel of air ascends, it expands and cools adiabatically; likewise, descending air warms. The adiabatic temperature change that accompanies such vertical displacements is determined by γ^*, the adiabatic temperature lapse rate.

For effectively dry air, $\gamma^* = \gamma_d = g/c_p \approx 10\ \text{K km}^{-1}$, where g is the gravitational acceleration and c_p is the isobaric specific heat capacity of the air. The adiabatic lapse rate of moist air, γ_m, is less than γ_d. The difference, which in today's atmosphere can be more than a factor of two, arises because condensation, and hence latent heating, accompanies adiabatic cooling as an air parcel expands. (See the Quick Study by Dale Durran and Dargan Frierson, PHYSICS TODAY, April 2013, page 74.) The rate of condensation is linked to the rate of temperature change by equation 1, making it straightforward to derive that

$$\gamma_m \equiv -\frac{\partial T}{\partial z} \approx \gamma_d \left[\frac{1+\dfrac{e_s}{p}\left(\dfrac{\beta}{T}\right)}{1+\dfrac{e_s}{p}\dfrac{R_d}{c_p}\left(\dfrac{\beta}{T}\right)^2}\right], \qquad (2)$$

where R_d is the specific gas constant in dry air and p is the ambient pressure. At warm temperatures near Earth's surface, $\gamma_m \approx 4\ \text{K km}^{-1}$, but it approaches γ_d in the cold and desiccated regions higher in the atmosphere.

The strong coupling between water and temperature is evident in figure 2a, which shows that the ambient water vapor pressure e is bounded—over four orders of magnitude—by its saturation value e_s. One can think of the circulation of air and its humidity as also coupled. Whether a parcel of air is relatively moist or dry depends on its history—in particular, the temperature when it was last saturated with water. In the cold of high altitudes, for instance, very little water can be sustained in the vapor phase, making saturated air quite dry. So relatively dry regions of the atmosphere are indicative of descending currents, which adiabatically warm and thereby lower their relative humidity (e/e_s). Such descending currents are preferentially located in the subtropics, which explains why the warm regions at a given isobaric level tend to be relatively dry, as figure 2b bears out. Hence, temperature places an upper but not a lower limit on the atmospheric humidity, and the degree of saturation is strongly connected to atmospheric circulation.

Understanding climate through water

Water's radiative properties determine the magnitude of the greenhouse effect, the planetary albedo, and hence Earth's surface temperature. As we'll see, those properties also determine the strength of the hydrological cycle and influence the thermodynamic structure of the troposphere, the lower 10–15 km of the atmosphere. Taken together, the effects from the presence of water also provide a foundation for understanding broad characteristics of the atmospheric circulation, particularly in the tropics.

A moist atmosphere is largely transparent to visible light but opaque to the IR. The enthalpy—or heat content, roughly—that's radiated to space by atmospheric water must thus be balanced by an input of enthalpy from Earth's surface. But radiative transfer alone does not maintain that balance. In an atmosphere that contains sufficient water, radiative transfer can only maintain the balance if the actual lapse rate γ is large—much larger than γ_m. In that situation, the adiabatic displacement of saturated air would make the air warmer than its surroundings and hence unstable. Vigorous convective currents, either in the form of towering cumulus clouds

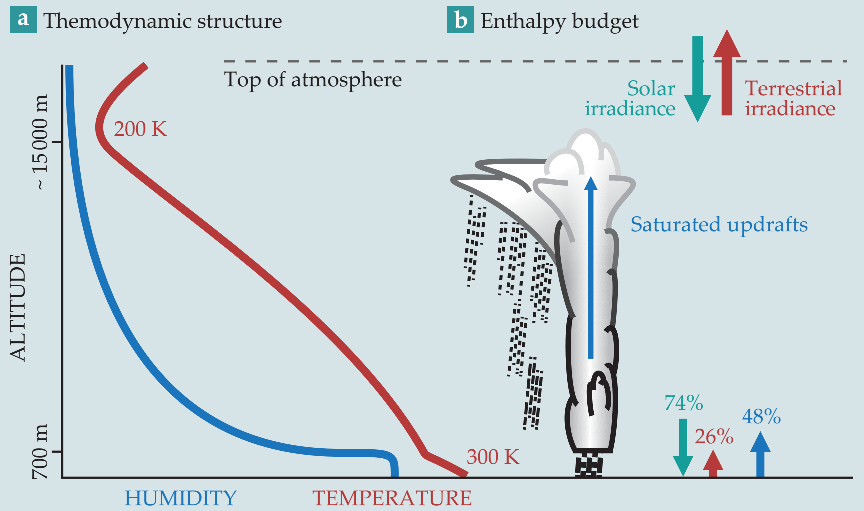

Figure 3. The thermodynamic structure and enthalpy budget of the atmosphere. **(a)** The atmosphere's temperature (red) and its absolute humidity (blue) are closely coupled. **(b)** At the top of the atmosphere solar and terrestrial irradiances balance one another. According to calculations, most (74%) of the incoming solar irradiance reaches the surface, but the net terrestrial irradiance at the surface is only a small fraction (26%) of its value at the top of the atmosphere. The radiative deficit (48%) is balanced by surface turbulent fluxes of enthalpy, arising mostly from evaporation, that transport warm water vapor from the surface to the troposphere, where it cools and condenses.

in the tropics or eddies in the mid latitudes, rapidly respond to the slow destabilization by radiative processes and mix together layers of the troposphere until the air reaches a convectively neutral state, at which $\gamma \approx \gamma_m$.

The convective currents transport water upward to much lower pressure and temperature, where it condenses and precipitates from the atmosphere. Over time, the net condensational heating must balance the radiative cooling (see figure 3). It is in this sense that one can understand that the strength of the hydrological cycle is determined by the rate at which the atmosphere radiatively cools. And as has been appreciated for more than a century, the troposphere is best described as being in a state of radiative convective equilibrium (RCE) rather than radiative equilibrium.[3] Less appreciated is the role water plays, through its interaction with IR radiation, in demanding such a balance.

Thermodynamic constraints accompanying the presence of water help dictate basic features of the atmospheric thermal structure near RCE. Upward-directed convective currents produced by radiative cooling rapidly saturate with water and adopt a temperature profile roughly equal to γ_m. Because $\gamma_m < \gamma_d$, the downward branches of the circulation must stably equilibrate to the thermal structure of the upward branches rather than vice versa. That equilibration is carried out by gravity waves, which act to eliminate density differences (essentially due to temperature) in air along isobaric surfaces. The process is particularly efficient in the tropical troposphere.[4]

Because water vapor pressure decreases as temperature decreases, as shown in figure 2, and temperature decreases with altitude, roughly following γ_m, at some altitude water molecules become so scarce that their net radiative contribution is negligible. That altitude defines the upper limit of the radiatively driven convection layer and the height of the troposphere. The temperature at the top of the troposphere is thus relatively independent of the temperature at the surface, and the tropospheric height adjusts to maintain consistency with the lapse rate γ_m.

The difference between γ_d and γ_m imposes an asymmetry on vertical displacements in the atmosphere. In air that becomes just saturated, upward displacements will be accompanied by an adiabatic temperature change that follows γ_m, while that from downward displacements will follow γ_d. Because the thermal structure of the descending zones of the atmosphere is tied to that of the ascending zones—that is, $\gamma \approx \gamma_m$—the difference between the actual lapse rate and the adiabatic lapse rates in those regions must adjust to balance the radiative cooling. That balance dictates the rate of descent, w, to satisfy $Q \approx w(\gamma_d - \gamma_m)$, where Q is the radiative cooling rate. That equation helps atmospheric physicists understand why a relatively small fraction of the atmosphere contains ascending currents.

The power of the constraints imposed by water is borne out by idealized, but nonetheless quantitative, studies of RCE. Despite a great many simplifications, these studies reproduce the main features of Earth's atmosphere,[5] including its thermodynamic structure and enthalpy budget. They also confirm that water is at least as important an influence on the atmosphere's structure and circulation patterns as are Earth's rotation, the presence of continents, and poleward gradients in solar insolation. According to computer simulations, making water invisible to radiation leads to a very un-Earth-like atmosphere—one for which the surface would be cooler, on average, by more than 20 K and in which convection plays almost no role in the transfer of enthalpy. The atmosphere would find itself in a state of near radiative equilibrium and thus relative stasis. Put simply, if water wasn't radiatively active, there would be no need for rain.

Calculations like those that generated the enthalpy-budget numbers in figure 3, or more exact approaches based on cloud irradiances measured by satellites, can be used to distinguish the contributions of water vapor to the energy budget from those of water condensate (in the form of clouds). Both approaches demonstrate that in a globally averaged sense, clouds act to cool Earth's surface. That result arises nontrivially out of the balance of two large but opposing effects: the cloud-greenhouse effect, whose magnitude can be measured by the difference between the cloud-top temperature and the surface temperature; and the cloud-albedo effect, whose magnitude depends on the amount of incident sunlight and differences between the reflectivities of the cloud and Earth's surface.

During the day, most clouds cool the surface, particularly low-level clouds whose greenhouse effect is small. At night, all clouds warm the surface. The differences between the irradiances from a fully clouded sky and a clear sky can be used to quantify the clouds' opposing effects. At the top of the atmosphere, satellites measure the cooling albedo effect to be about 50 W/m^2, about twice as large as the warming greenhouse effect. On balance, clouds moderate the substantial warming of the surface that arises from the presence of water vapor.[6,7]

The net radiative effects of clouds also have the potential to influence the strength of the hydrological cycle. Although in today's climate, the net radiative heating of the atmosphere attributable to clouds is nearly zero, when averaged globally. Clouds can have a large effect on the radiative heating rate regionally. And in so doing they help shape the structure and pattern of large-scale circulation systems.[8] Whether or not that circulation helps ensure that the clouds' globally averaged radiative effect on atmospheric heating is near zero remains a mystery.

Amplifier of climate change

Because water is so closely coupled to temperature, it cannot drive climate change. Rather, it orchestrates and amplifies the effects of agents capable of acting independently of surface temperature. For instance, as concentrations of long-lived greenhouse gases such as CO_2 increase, so does the IR opacity of the atmosphere, which in turn raises the altitude z^* from which thermal radiation is emitted to space. Such a change—referred to as radiative forcing—produces an imbalance in Earth's radiation budget. The balance can be restored either by increasing the temperature at z^* or by increasing the planetary albedo. In the former case, the constraints on the lapse rate require that the surface temperature must also increase.

Simple calculations show that if only temperature was allowed to change in response to a doubling of atmospheric CO_2, surface temperature would increase by a little more than 1 K. The presence of water amplifies the response in several well-understood ways (see figure 4). In the so-called water-vapor feedback, an increase in temperature supports an increase in the water vapor pressure throughout the troposphere, which further increases the IR opacity of the atmosphere. That, in turn, further increases z^*, which raises temperatures further, and so on.

The lapse rate γ decreases in magnitude with warming because of the temperature dependence of γ_m in equation 2. In the lapse-rate feedback, that decrease reduces the surface temperature change implied by a given change in z^*. Through positive cloud-greenhouse feedback, surface warming produces a greater temperature difference between the surface and the top of the troposphere. It thereby increases the greenhouse potential of clouds to further warm the surface.[9] Finally, in the surface-albedo feedback, surface warming prompts the planetary albedo to decrease as snow and sea ice melt.

The water-vapor and lapse-rate feedbacks have opposing signs and are linked, though the water-vapor feedback is considerably larger in magnitude. Indeed, even after subtracting the lapse-rate feedback, the water-vapor feedback, which accompanies an atmosphere that warms while maintaining a constant relative humidity, is still as large as, or larger than, the cloud-greenhouse and surface-albedo feedbacks put together, as figure 4 shows. The strength of the water-vapor feedback rests on the observation that the relative humidity change little with temperature; that is, the vapor pressure changes closely follow the change in saturation vapor pressure, as figure 2 shows.[3,10,11] Hence, the presence of water in the atmosphere amplifies any external radiative forcing; that fact has been at the centerpiece of our understanding of climate change for more than 30 years.[12]

Amplification of a radiative forcing isn't the only consequence of water's presence. The hydrological cycle can also change in response to those external forcings, and our understanding of water helps explain those changes. Because the radiative cooling of the troposphere is an increasing function of its temperature, a warmer atmosphere cools more rapidly. And because at the top of the atmosphere the IR flux emitted to space must balance the absorbed solar radiation flux, additional atmospheric cooling occurs through a reduction of the net upward thermal irradiance at the surface. To the extent that the increased cooling cannot be offset by increased absorption of solar radiation by water vapor, or by a change in cloudiness, it must be associated with an increase in global precipitation.

Regionally, wind currents will mediate where that increase occurs. If the currents change little compared to the amount of water they transport, wet regions will import more atmospheric moisture and become wetter; conversely, dry regions will export more moisture and become, on average, drier.[10,13] Those circulation effects are governed by equation 1 and combine with the energetic constraints to provide a good first approximation to changes in precipitation,

$$\delta P \approx \delta R + \frac{\beta}{T_{\mathrm{sfc}}}(P - E)\,\delta \ln T_{\mathrm{sfc}}, \qquad (3)$$

where δR is the change in the net cooling rate of the atmosphere, P and E measure precipitation and evaporation (in enthalpy flux units), respectively, and T_{sfc} is the surface temperature. Equation 3 helps explain the precipitation changes predicted by complex models and gives confidence in the assessment that regional changes will accompany global warming.[14] The last term of equation 3 emphasizes how closely precipitation is coupled to changes in atmospheric circulation patterns, which are themselves coupled to water in ways that remain poorly understood.

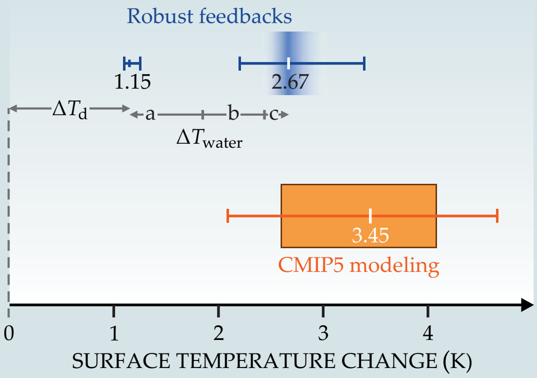

Figure 4. How much does the temperature of Earth's surface change from doubling the carbon dioxide in the atmosphere? Comprehensive climate models from the coupled-model intercomparison project (CMIP5) yield a median estimate of 3.45 K bounded in the orange box by the 25%–75% spread over the total range of values. Shown in blue are the estimates of the temperature change based on well-understood feedback processes. In the absence of the effects of water, one expects a warming ΔT_d of about 1 K. Water in the atmosphere increases that dry response by an additional 1.5 K. The amplification includes contributions from processes described in the text: the combined water-vapor and lapse-rate feedbacks (a), the cloud-greenhouse feedback (b), and the surface-albedo feedback (c).

Cloudy futures

Scientists have good reasons to believe that changes to water in response to a given radiative forcing account for more than half of Earth's average surface temperature change. But estimating the climate sensitivity with greater precision remains difficult. Models that encapsulate the basic properties of water produce a wide range of estimates (see figure 4). Most of the imprecision in climate sensitivity and regional patterns of rainfall changes can be related to a poor understanding of how clouds change in a warming climate[15] and how changing clouds affect atmospheric circulations.[8]

Although clouds have long been recognized as crucial for Earth's radiation budget, only in the past few decades have researchers appreciated that clouds can both warm and cool the atmosphere and the surface. Early models of the climate system, to the extent that they considered clouds at all, assumed that their effect on the enthalpy budget was not a function of the climate state. But RCE models developed in the early 1970s and later general-circulation models demonstrated that clouds exert a marked influence on climate sensitivity.[8,16]

In addition to the idea of a positive feedback associated with changes in the cloud-greenhouse effect, other ideas have begun to emerge as to why cloudiness might depend on the working temperature of the atmosphere. In most cases the ideas stem from the fundamental properties of water. For instance, because the lapse rate of air that remains saturated as it rises is a function of temperature, warmer climates might be characterized by more condensate-laden clouds; the larger optical depth increases the cloud-albedo effect and thereby moderates the warming. In contrast, warming is also expected to be accompanied by increased evaporation, which drives more mixing in the lower atmosphere and may lead to fewer clouds, enhancing warming.[17]

As the singular challenge clouds pose to our understanding of climate and climate change has become better appreciated, research on clouds has intensified. In recent years detailed experimentation and analyses of climate models have demonstrated which cloud regimes and processes are critical to explaining intermodel differences in the projections of future climate.[15] Recent research also shows that clouds directly mediate the response of the atmosphere to an external forcing, and they do so on time scales as short as a few hours.[18] More generally, so strong is the coupling between clouds and circulation systems, from thunderstorms to monsoons, that advancing our understanding of regional climate change rests firmly on advancing our understanding of clouds and cloud processes.

Despite imperfect models, our understanding of the behavior of the climate system is so deeply rooted in the basic physicochemical properties of the water molecule that we can confidently conclude that global warming from anthropogenic emissions of long-lived greenhouse gases poses serious risks. And yet we're hampered by an inability to clearly pin down the pace of that warming and the nature of regional changes the planet is likely to experience. A grasp of both is crucial for adaptation measures. That fact highlights the urgent need to better understand the ways in which water couples to the atmosphere's circulation systems.

We thank Kerry Emanuel, Isaac Held, John Mitchell, and Raymond Pierrehumbert for their extensive and thoughtful comments on an early version of this manuscript, and Aiko Voigt for contributions to our discussion of RCE.

References

1. K. P. Shine, I. V. Ptashnik, G. Rädel, *Surv. Geophys.* **33**, 535 (2012).
2. R. T. Pierrehumbert, *Principles of Planetary Climate*, Cambridge U. Press, New York (2010).
3. S. Manabe, R. T. Wetherald, *J. Atmos. Sci.* **24**, 241 (1967).
4. K. A. Emanuel, J. D. Neelin, C. S. Bretherton, *Q. J. R. Meteorol. Soc.* **120**, 1111 (1994).
5. D. Popke, B. Stevens, A. Voigt, *J. Adv. Model. Earth Syst.* (in press), doi:10.1002/jame.20009.
6. B. Stevens, S. E. Schwartz, *Surv. Geophys.* **33**, 779 (2012).
7. G. L. Stephens et al., *Nat. Geosci.* **5**, 691 (2012).
8. A. Slingo, J. M. Slingo, *Q. J. R. Meteorol. Soc.* **114**, 1027 (1988).
9. D. L. Hartmann, K. Larson, *Geophys. Res. Lett.* **29**, 1951 (2002).
10. J. F. B. Mitchell, C. A. Wilson, W. M. Cunnington, *Q. J. R. Meteorol. Soc.* **113**, 293 (1987).
11. S. C. Sherwood et al., *J. Geophys. Res.* **115**, D09104 (2010).
12. S. Bony et al., in *Climate Science for Serving Society: Research, Modelling, and Prediction Priorities*, G. Asrar, J. W. Hurrell, eds., Springer, Berlin (in press).
13. I. M. Held, B. J. Soden, *J. Climate* **19**, 5686 (2006).
14. S. Bony et al., *Nat. Geosci.* (in press), doi:10.1038/ngeo1799.
15. S. Bony et al., *J. Climate* **19**, 3445 (2006).
16. V. Ramanathan, J. Coakley Jr, *Rev. Geophys. Space Phys.* **16**, 465 (1978).
17. M. Rieck, L. Nuijens, B. Stevens, *J. Atmos. Sci.* **69**, 2538 (2012).
18. J. Gregory, M. Webb, *J. Climate* **21**, 58 (2008). ■