OUR WARMING PLANET

TOPICS IN CLIMATE CHANGE

Climate Lecture 3

Building a Climate Model

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Climate, or the average of day-to-day weather, can be very different at various points on Earth. The local climate in the Arabian Desert is hot and dry, while that in the Amazon River basin is hot and humid with frequent rain. In upstate New York, the climate changes from being warm in the summer with sporadic rain to cold in the winter with sporadic snow. Hawaii, on the other hand, has a pleasant climate all year long. However, the day-to-day weather at all of these locations is much more variable. There can be dry days in the Amazon jungle, and rainy days in the Arabian Desert. There are some days in winter that are warmer than some days in summer. For further contrast, daylight in Antarctica lasts up to six months at a time with freezing cold day-in day-out. Can a climate model be built that can reproduce all of this complex behavior?

Model Variables

Climate models require that the climate system be described in terms of physical variables such as temperature, pressure, clouds, wind, and solar radiation, variables that change with time and interact with each other and with the evolving properties of the land, ocean, and atmosphere. Of these, water is the most active, ubiquitous, and most significant substance in the entire climate system, appearing in and transforming between solid, liquid, and vapor forms. It is the principal source of the energy released in destructive storms, as well as being the indispensable enabler in sustaining an efficiently functioning biosphere. There are also tiny amounts of other gases such as ozone and carbon dioxide distributed throughout the atmosphere. It is ozone that keeps the biosphere from being destroyed by extreme ultra-violet radiation. Meanwhile carbon dioxide acts to maintain the strength of the terrestrial greenhouse effect at its present temperate level.

All of these different climate system variables are specified numerically to assure a quantitative mathematical description of the state of the land, ocean and atmosphere at all points on the Earth, at all heights in the atmosphere and at all depths of the ocean, and for every arbitrary time of day. This imposes an enormous amount of information handling, not to mention keeping track of the interactions that take place between the different climate system variables, and requires very large and fast computers to handle the mathematical models that have been specifically designed to simulate the basic physical structure and manner of operation of the climate system. There are basically three different types of climate variables: *prescribed*, *prognostic*, and *diagnostic*.

Prescribed variables include the basic parameters that define the terrestrial climate system such as the radius and total mass of the Earth, the Earth's orbital parameters, positioning of the Sun and Moon, atmospheric composition, and land-ocean distribution and topography. There are also a number of numerical coefficients that have been empirically derived, and can be effectively grouped with the physical constants that are an integral part of the model physics. Prescribed variables can interact with all other climate components, but they themselves are not transformed by these interactions. Simpler climate models may include ozone with its seasonal variability, vegetation albedo, and aerosols, or use specified ocean surface temperature. In a Q-flux climate model, seasonal ocean heat transports are preselected from separate offline calculations. In such Q-flux simulations, ocean currents and salinity stay fixed, but the prescribed heat transports are applied to ocean heat content, so that ocean temperature and sea-ice act as prognostic variables.

Prognostic variables are the basic variables in terms of which the climate system components interact with each other, undergo change, and evolve with time. They are continuously defined at each three-dimensional model grid-box, and include temperature, dry-air mass, water substance, wind components, and changes in clouds, snow/ice cover, and subsurface reservoirs. In coupled atmosphere-ocean climate models, ocean mass, salt, heat content, and currents are all prognostic variables. For some simulations, tracer quantities for chemistry, aerosols, or water isotopes may be prognostic. Scalar quantities are defined at the grid-box center and fluxes at grid-box edges, but velocity locations for atmosphere, ocean, and sea-ice depend on the differencing scheme used (of which there are several). At each model time step, the horizontal and vertical fluxes of mass, momentum, and energy of the three-dimensional grid-boxes are acted upon according to the established laws of physics, thermodynamics, and fluid mechanics, all ensuring that mass, momentum and energy are globally conserved at every time step (Hansen *et al.*, 1983).

Diagnostic variables are continually being derived from prognostic variables. Thus, pressure at any level is computed from the mass of air, ice and water above that level multiplied by gravity. Other examples are: surface albedo, which includes the effects of changing snow accumulation; also surface temperature, which is determined from the net balance of energy contributions from changes in radiative heating and cooling, sensible heat, cooling by evaporation, and net heating or cooling by falling precipitation. Potential, kinetic, latent, and radiative energy components, angular momentum, vertical wind, are diagnostic, although they eventually affect the prognostic variables. Other diagnostic variables are used only for model output, including global maps of reflected solar radiation and emitted thermal radiation that arise from radiative transfer modeling. Such maps serve to compare model performance against observational data. Model diagnostics also provide illustrative explanations of the physical functioning of the climate system.

Serving as the central nervous system of the climate model, are the *fundamental equations* that describe the atmosphere and the physical interactions between the different constituents of the climate system. These insure that the conservation of mass, momentum, and energy are strictly enforced at all times and at all points of the climate system. The *equation of motion* is written for a rotating frame of reference with a fixed rotation rate. It accurately represents the gravitational, pressure-gradient, and Coriolis force control over the dynamic motions of the atmosphere. The *thermodynamic equation* describes the work performed upon a unit mass in compressing it, and includes heating and cooling by all other processes. Except for topography, the Earth is treated as a sphere with the atmosphere being thin relative to the planetary radius. The vertical component of the equation of motion is replaced with the *hydrostatic assumption* to filter out vertical sound waves, and permit longer time steps. The Earth's *orbital parameters* are model specifications that enable paleo-climate simulations. They define the precise amount of solar radiation incident on each model grid column at each time step, in accord with the local time of day and season.

GISS Climate Model Beginnings

Climate modeling at the Goddard Institute for Space Studies (GISS) began in the mid-1970s. Of note was the IBM 360/model-95 computer that occupied half of the second floor of the GISS building and had 5 megabytes of then impressive internal memory. David Rind, a knowledgeable meteorologist, was one of the key developers of the GISS climate model. Using a coarse 8□x10□ latitude-longitude grid with 9 vertical layers, the model employed an alternating dynamics/source term 60-minute time-marching strategy with shorter dynamics time steps that utilized a leap-frog formulation. At the roughly 1000 km spatial resolution, most storm systems went unresolved. Climate system variables needed to be specified at some 7500 grid-boxes in order to numerically describe the pressure and temperature structure of the atmosphere. Global distributions of winds, clouds, and water vapor required the use of numerical *parameterizations* to represent the essence of the unresolved sub-grid physical interactions in terms of their grid-box-mean values.

It is noteworthy that one year's worth of model-simulated time contains hundreds of millions of pressure, temperature, wind, cloud, and humidity data points. In 1970's climate modeling, it was common practice for the instantaneous prognostic and diagnostic variables to be loaded onto *history tapes* with a daily time-sampling frequency. Then, at some later date, statistical analyses would be performed to extract the climate relevant information from the accumulated data as if the model-generated data constituted a detailed collection of real meteorological measurements.

This cumbersome modeling approach was a distinct impediment to rapid model development and assessment of GCM performance, but it was resolved with the implementation of comprehensive *online model diagnostics* in the GISS climate model. The numerical information required to formulate model diagnostics is always readily available as the model runs its course. Assembling and packaging this information online was the GISS climate modeling innovation that enabled literally hundreds of model simulations to be rapidly conducted and evaluated, screening and testing the model physics and parameterizations for validity, accuracy, and reliability. The earlier approach of analyzing the model-generated history tapes would have required much more time to achieve the same model development objectives. Because climate model simulations calculate the full repertoire of weather fluctuations, model diagnostics can be programmed to also tabulate the frequency of occurrence as well as the severity of extreme weather events such as prolonged heat waves, droughts, and onsets of damaging frosts. Thus, online diagnostics can provide many options for rapidly assessing and evaluating climate model output. However, with the advent of ever more powerful computers and readily available mass storage, the "history tape" approach has renewed important applications for inter-model comparisons and other statistical analyses.

The overarching principle in constructing a successful climate model is achieving explicit and detailed conservation of mass, energy, momentum, and water substance. It is taken for granted that explicit conservation of all of these quantities takes place in real world climate, even though the observational verification of that is well beyond measurement capabilities. As the ensuing corollary, in order for any climate model to be considered credible, its model diagnostics must first demonstrate precise conservation of mass, energy, momentum, etc., even though some other model diagnostics might suggest good agreement with observations.

Climate vs. Weather Models

In essence, a climate model is intrinsically a weather model. The key difference is that a weather model operates in a mode of what is called an *initial value problem* in physics. This means that in weather forecasting models, to the extent feasible, pressure, temperature, wind, and humidity fields are all *initialized* with assimilated data from current meteorological observations. As the weather model is then stepped forward in time, it is actively predicting future weather.

The operational strategy in running a climate model is different from that of a weather model. Here, with the same mathematical model operated as a climate model, we are actually working to solve what is called a *boundary value problem* in physics. The computational procedure may appear similar to that of a weather model, since the climate model is also started from an initial state, and then time-marched forward into future time. The key difference is that in a weather model the initial conditions define the sought-after output, whereas in a climate model, the initial atmospheric state is of no particular relevance. The objective of a climate model is to determine the "*climatic mean*" of weather variability over a specified time interval (month, season, year, decade or century), and study how the climate system responds to the *radiative forcings* that act to shift the existing average state of the climate system away from its previously established equilibrium to a new quasi-equilibrium point. It is this "shift" in the average climate state that defines the climate system response to the applied radiative forcing, and it is independent of the initial state of the model. Small sustained radiative forcing changes have a cumulative effect that is magnified by the climate system feedback effects. They must all be modeled accurately.

Computer limitations and the model spatial and time-stepping resolution come into play when simulating unresolved physical processes on a sub-grid scale. For example, physical processes that operate on sub-grid scale cannot be rigorously evaluated, and instead must be replaced by *parameterizations* that are only approximately correct. Also, in both climate models and the real world, the climate system is striving towards its energy balance equilibrium. However, neither the climate model nor the real world are able to approach the equilibrium point in small enough incremental steps so as to reach equilibrium smoothly. Because of this characteristic response of the climate system's physical processes (condensation, cloud formation, precipitation) to the disequilibrium forcing are so strong, they systematically overshoot the equilibrium point, thus producing the familiar quasi-chaotic behavior (or *natural variability*) of terrestrial weather.

Thus, in a way, initializing the weather model with current meteorological data, is an attempt to co-align the otherwise unrelated space-time phasing of the quasi-chaotic behavior of the model with that of the real world. A limitation of weather models is that they accumulate errors arising from their sub-grid parameterizations and unresolved physical process, resulting in a substantial decline in forecast accuracy beyond lead times of only a few days. Climate models, however, are not attempting to solve an initial value problem; their focus is on how realistic their modeling is of the external radiative forcings and of the corresponding climate feedback response.

GISS Model Evolution

To be considered trustworthy for analyzing climate change, the model must first demonstrate that it can successfully reproduce the basic seasonal and geographic variability of current climate – based more on realistic model physics, and less on process parameterizations that might have been tuned for current climate conditions. To this end, as computers have become more powerful and much faster, the GISS climate model has evolved to the point that many prescribed variables have now been reconstituted as prognostic variables. In particular, the space-time variability and distribution of ozone and aerosols are now being calculated online using atmospheric chemistry calculations in conjunction with interactive dynamical transport. Prescribed vegetation properties have also been transformed into prognostic variables with model vegetation cover responding to cumulative changes in precipitation and temperature. Nevertheless, it still remains convenient to utilize prescribed variables, such as the atmospheric change in carbon dioxide, which is known accurately, without incurring the added computational burden of modeling the carbon cycle, when that is not the specific point of interest.

More importantly, the ocean heat transport (Q-flux) is no longer prescribed, having been made fully interactive in what is called a *coupled atmosphere-ocean model*. Like the atmosphere, the ocean is a fluid subject to the same gravity, Coriolis force, and basic laws of fluid mechanics. The principal differences are that the ocean is a thousand times denser and more massive than the atmosphere, and much less compressible. Space-time changes in temperature and salinity are the driving factors that affect ocean circulation, constrained by the ocean-bottom topography. The ocean, with its enormous heat capacity, interacts strongly with the atmosphere by exchanging heat radiatively, and by means of sensible and latent heat. Since the ocean absorbs nearly half of the incident solar radiation, mostly in the tropical regions, this requires a redistribution of heat toward the polar regions. A key aspect of the atmosphere-ocean interaction is the wind influence on ocean surface currents, and its effect on heat redistribution within the ocean mixed layer.

As a result, present-day climate models are invariably coupled atmosphere-ocean models. This is because the question being posed to climate models is: how does the increase in greenhouse gases affect both global and regional climate? The direct global warming due to the increasing greenhouse gases is easy enough to calculate. It is more difficult to determine the precise rate of heat uptake by the ocean, which determines the speed at which the surface temperature warms. As an added complication, the ocean is responsible for the climate system's *natural variability*, which is characterized by radiatively unforced fluctuations occurring on inter-annual and decadal time scales such as the El Niño and La Niña events, or the Pacific Decadal Oscillations.

Current State of the Art

Present-day climate models with their growing realism have become indispensible research tools (Schmidt *et al.*, 2006; 2014). Reliable prediction of future climate change will continue to be the principal objective. This includes understanding how the climate will change on a regional basis, including the impact of global warming on regional precipitation patterns, fresh water resources, sea level rise, severity of storms, and ecological stability. Other areas of active research include reconstructing climate on geological time scales from when Earth was an ice covered snowball 650 million years ago, to global tropical conditions that prevailed during the Cretaceous period. Also of great interest is modeling the regional climate change since the last ice age, in particular what sustained the lush vegetation that existed 6000 years ago in what is now the Sahara desert.

The recent discovery of the existence of thousands of extra-terrestrial planets has opened a new area of investigation where a general-purpose climate model is needed to assess and evaluate the prospects of habitability over a wide range of orbital parameters, parent star type, and planetary mass, size, and atmospheric composition. This has renewed interest for inter-planetary climate comparisons with neighbor planets Mars and Venus, which have atmospheres composed almost entirely of carbon dioxide. Mars, with an atmosphere that is only about 1% as massive as that of the Earth, has a surface temperature scarcely 5°K warmer than direct solar heating can support, while the surface temperature on Venus is 500°K warmer than the absorbed solar energy would imply. Exploring how planetary climate responds to extreme changes in radiative forcing is one example of the types of research studies that can be investigated with a climate model. Other examples include the study of isotope fractionation used in calibrating the temperature dependent precipitation rate of water vapor for deuterium and oxygen-18 isotopes to determine changes in global temperature from geological ice core records (Schmidt *et al.*, 2007).

The current emphasis in climate model development is to upgrade the ocean model component. With its large heat capacity, and as the principal instigator of the natural variability component in the ongoing climate record, modeling this behavior is a major challenge. A significant part of the problem is that extensive measurements of ocean temperature, salinity, and circulation have not been available, as compared to the observational scrutiny of the atmosphere. The impact of lunar and solar tides on ocean circulation has been successfully included. Work continues to upgrade the model treatment of sea-ice formation, degradation, and transport by ocean currents. A better performing ocean model is key to developing a reliable climate forecast model.

Significant progress has been made to upgrade the capability of the numerical methodology used in model calculations, such as replacing the traditional latitude-longitude grid with a cube-sphere grid (Putman and Lin, 2009), and replacing the two horizontal velocity components with three non-independent symmetric components (Russell and Rind, 2015). Interactive computations are also being revised with interactive carbon cycle chemistry modeling having been recently added to the expanding climate modeling capabilities (Romanou and Romanski, 2014).

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Slide 1

Our Warming Planet: Lectures in Climate Change

A Compendium of Lectures on Special Topics
in Climate Dynamics and Climate Change

A Tribute to David Rind
for Outstanding Climate Science Achievements

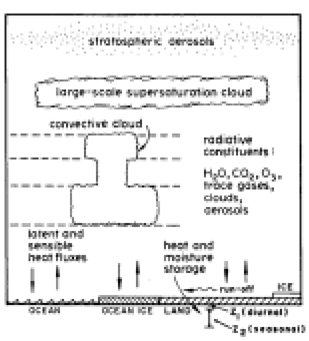
Building a Climate Model

Gary Russell
NASA Goddard Institute for Space Studies

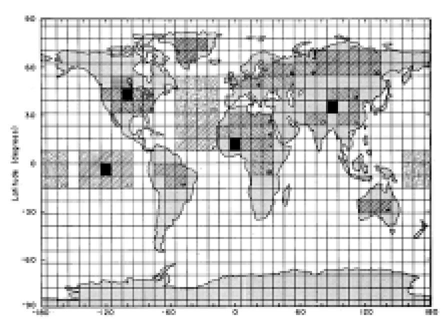
Imperial College Press
2016

Slide 2

GISS General Circulation Climate Model: The Early (1983) Version



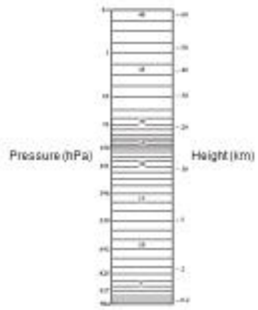
Schematic structure of the terrestrial climate system as depicted in a 1983 version of the GISS climate model.

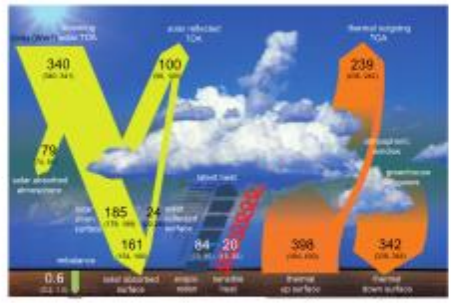


The model's horizontal resolution and surface topography were described on a 8° × 10° latitude-longitude grid.

Slide 3

GISS General Circulation Climate Model: ModelE2 (2014) Version





Radiative, thermodynamic and dynamic processes are included in a model which is broken up into various vertical levels in the atmosphere.

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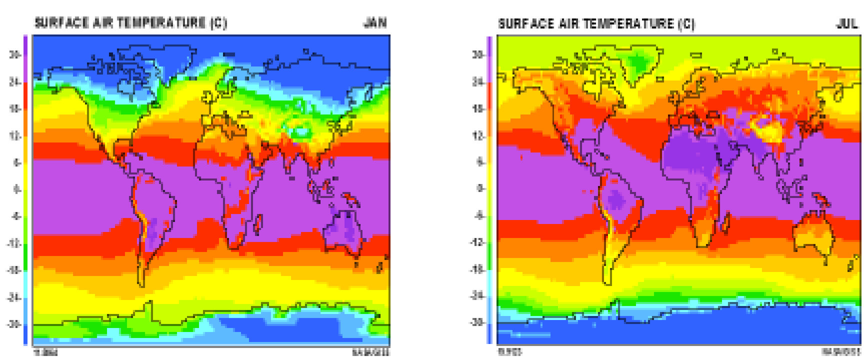
Slide 4

Surface Air Temperature : The Most Significant Global Climate Indicator



Seasonal variability of the surface temperature, between January (left) and July (right).

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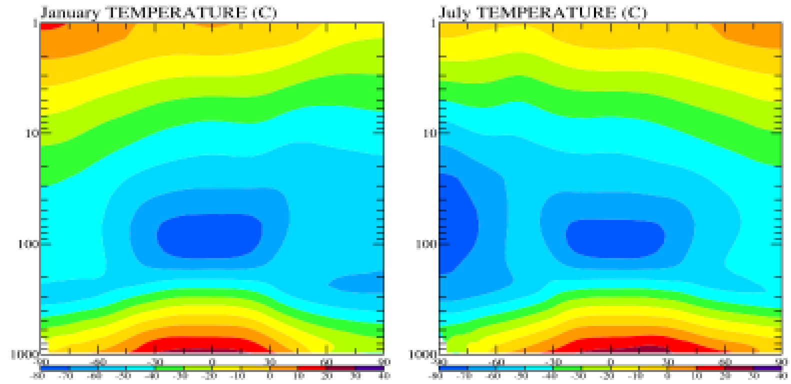
Slide 5

Seasonal Air Temperature as Function of Latitude and Pressure (mb)



Atmospheric temperature decreases steadily with height, from the surface until the minimum temperature tropopause which ranges in altitude from 12 to 20 km. Above the tropopause, temperature increases with height until the stratopause near 50 km, then (beyond the graph) decreasing until the mesopause near 80 km, and finally increasing again into the thermosphere.

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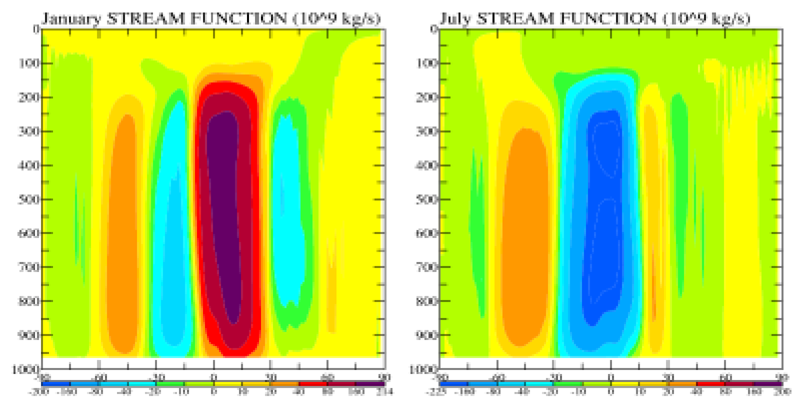
Slide 6

Seasonal Variability of the Atmospheric Stream Function



The chief feature of the atmospheric streamfunction, the Hadley Cell, changes its circulation from clockwise (in the plane of the figure) in January to counterclockwise in July.

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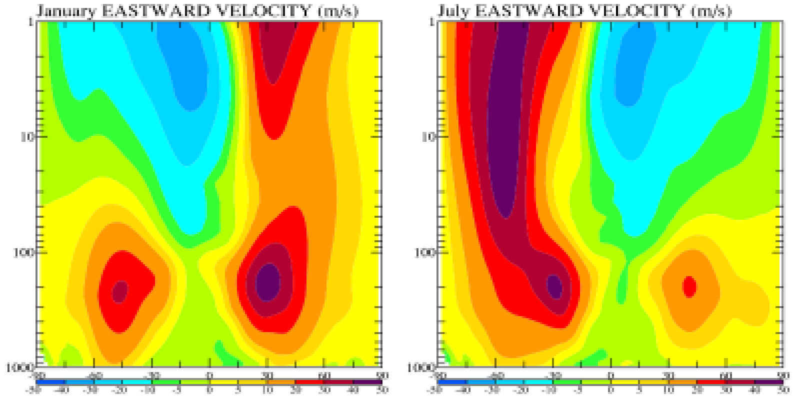
Seasonal Variability of the Zonal Wind Velocity



Conservation of angular momentum causes surface winds from the Hadley cell heading toward the equator to be redirected toward the west; this causes the tropical trade winds. Similarly, pole-ward heading surface winds from the Ferrel cell are redirected eastward. Angular momentum discrepancies between tropics and mid-latitudes also drive the mid-latitude jets.

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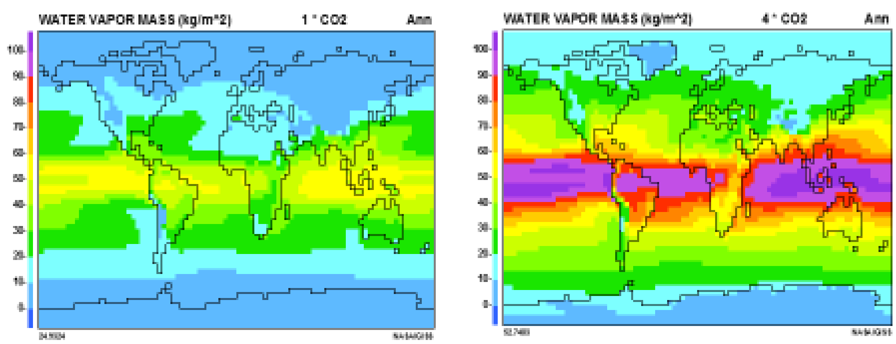
Slide 8

Water Vapor Mass (kg/m^2), 1950 CO_2 Concentration and 4 times CO_2



The amount of water vapor that air can hold is an exponential function of temperature, increasing as temperature increases. Thus, as the planet warms, the atmosphere can, and will, retain more water vapor.

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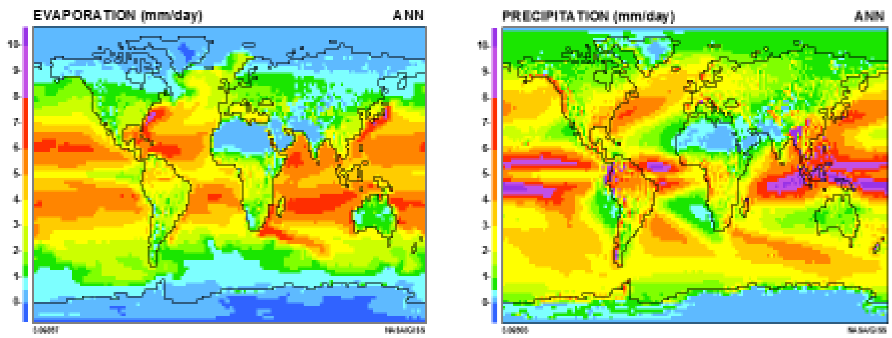
Slide 9

Evaporation and Precipitation, global annual average is 3 (mm/day)



Evaporation is greater in the tropics than over high latitudes, and greater over oceans or lakes than over ground or ice surfaces.

Precipitation occurs when moist air cools. The amount of water vapor that the air can hold is inversely proportional to pressure, and is strongly governed by the exponential temperature dependence as set by the Clausius-Clapeyron formula.

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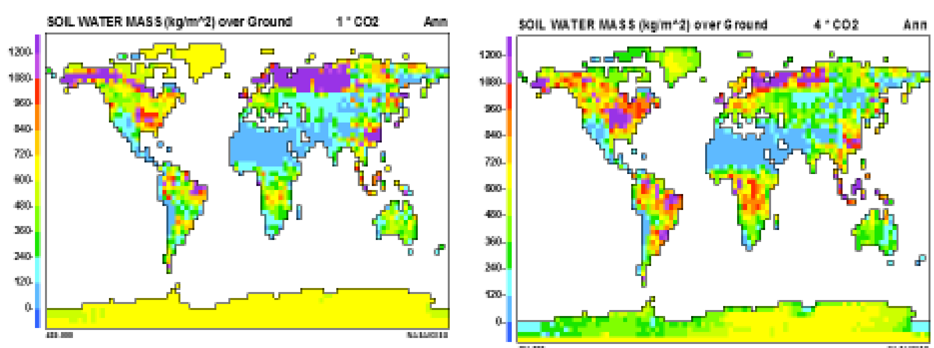
Slide 10

Soil Water Mass (kg/m^2), 1950 CO_2 Concentration and 4 times CO_2



With severe climate change, the distribution of ground water on Earth will be altered.

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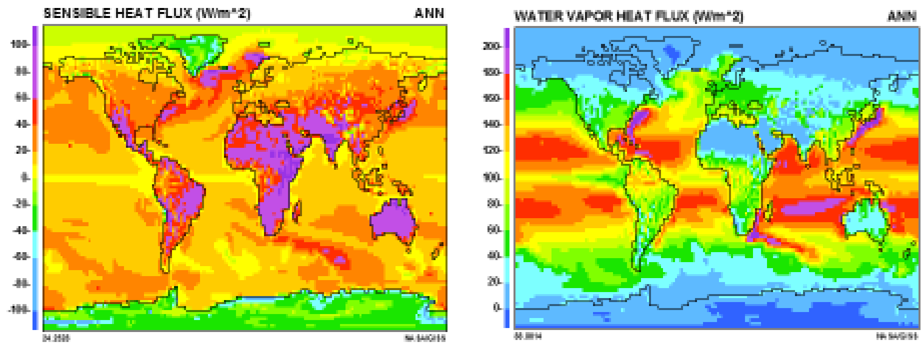
Slide 11

Sensible and Latent Heat Fluxes from Surface to Atmosphere



Fluxes of sensible heat (larger over land) and latent heat (larger over the oceans) help distribute heat from the surface into the atmosphere.

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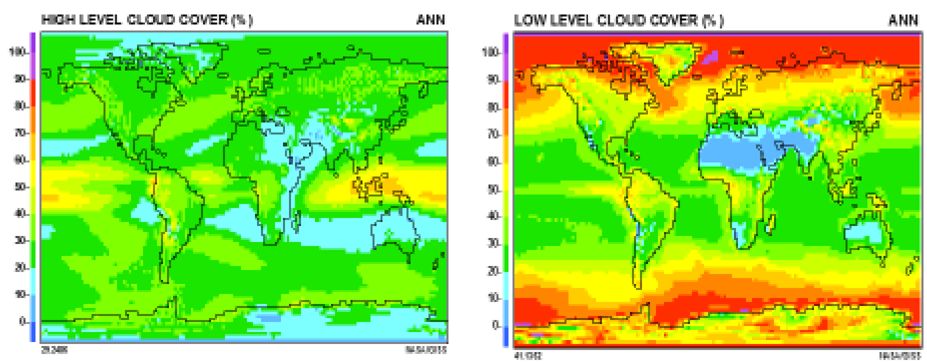
Slide 12

High and Low Cloud Global Annual-mean Distribution



Clouds are important regulators of the Earth's global energy balance. During the day, clouds reflect sun-light, thus cooling the planet. Clouds at night-time keep the planet warmer by reducing the escape of thermal radiation to space.

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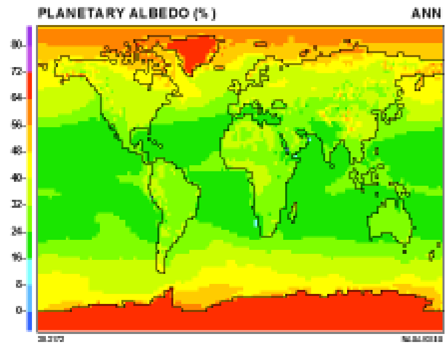
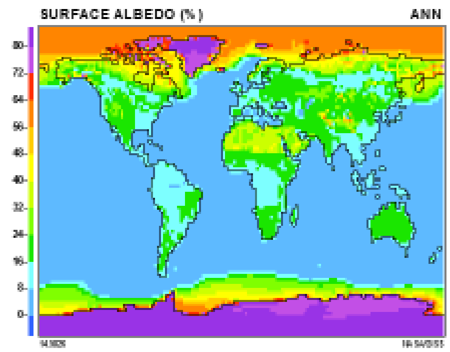
Slide 13

Planetary (Top of Atmosphere) and Surface Albedoes





Albedo is the ratio of the (spectrally integrated) reflected sun-light divided by incident sun-light.

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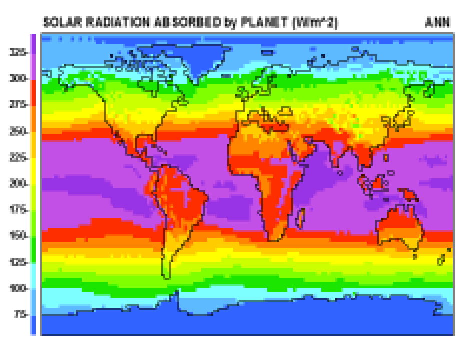
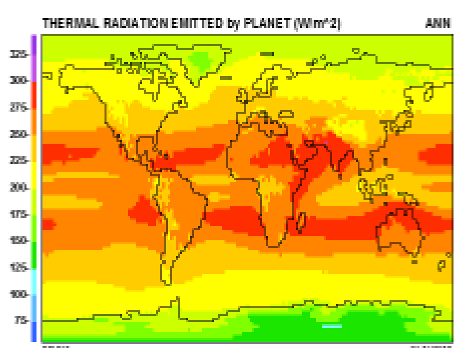
Slide 14

Solar Radiation Absorbed and Thermal Radiation Emitted by Planet





The Earth absorbs shortwave energy from the sun (left) and emits longwave energy back to space (right). When they are equal, the earth's temperature is in equilibrium.

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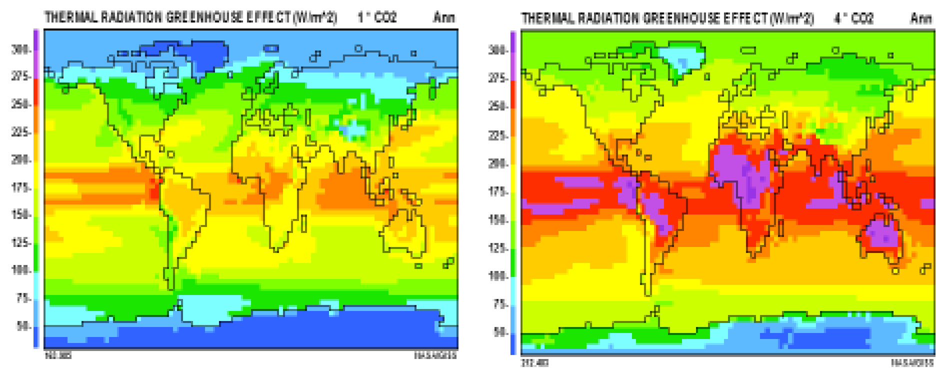
Slide 15

Atmospheric Greenhouse Effect, 1950 CO_2 Concentration and 4 times CO_2



The greenhouse effect (W/m[2]) is defined as the difference between the upward thermal radiation emitted from the surface minus thermal radiation emitted at top-of-atmosphere out to space. The strength of the greenhouse effect is a measure of the heat trapping efficiency of spectral absorption by greenhouse gases and by longwave cloud opacity.

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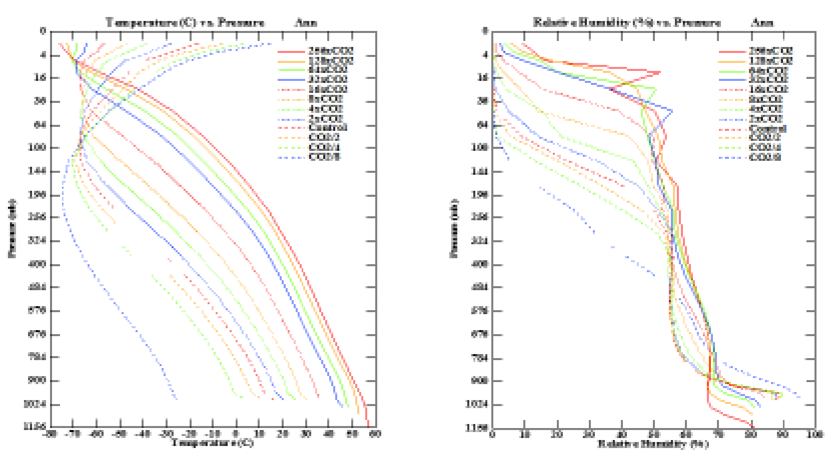
Slide 16

Climate Change Experiments: Temperature and Relative Humidity



The climate of Earth is strongly controlled by the amount of atmospheric CO_2.

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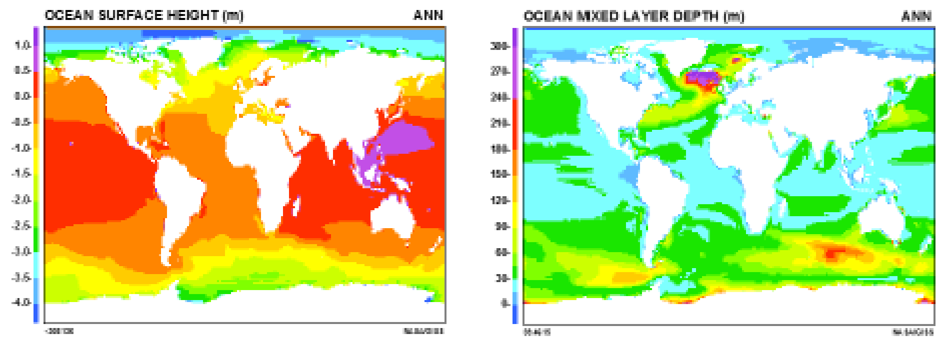
Ocean Surface Height (m) and Ocean Mixed Layer Depth (m)



The reference geoid for the ocean surface is a level where the surface pressure gradient force is zero. This includes effects due to variations of both gravity and of the Earth's rotation.

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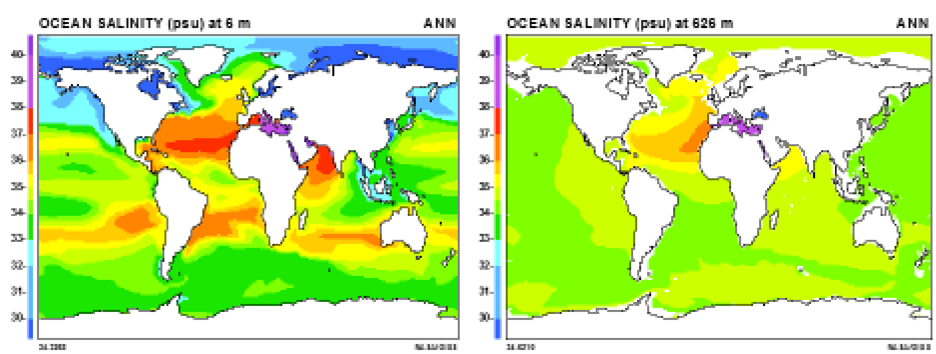
Ocean Salinity at the Surface and at Ocean Depth of 626 (m)



Salinity, because it directly affects the density of sea water, along with the water temperature are the principal are the principal controlling factors that affect the global circulation of the ocean, and thus the transport of heat energy pole-ward from the tropics.

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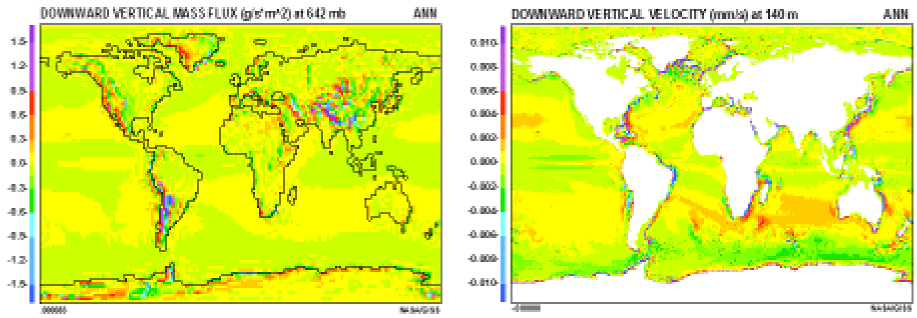
Downward Mass Flux in Atmosphere and Downward Velocity in Ocean



Vertical velocities in both the atmosphere (left) and the ocean (right) are subject to numerical instabilities in the model, especially along mountain ranges and coastlines.

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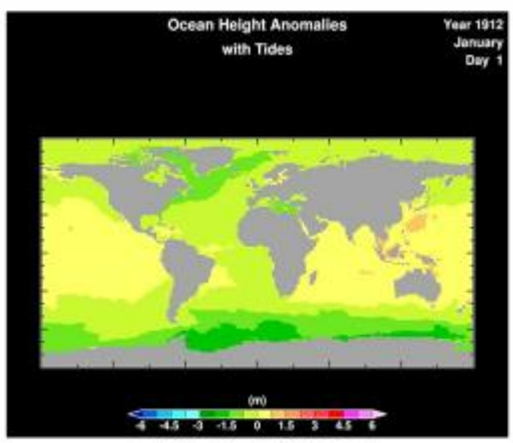
Slide 20

Realistic Lunar and Solar Tides that Accelerate Horizontal Ocean Velocities

Click play on the video to see tidal shift over the month of January 1912.



The side of the Earth facing the Moon is more strongly pulled by the Moon's gravity and the opposite side is pulled more weakly.

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Slide Notes

Slide 2 Schematic structure of the terrestrial climate system (left panel) as depicted in a 1983 version of the GISS climate model. Atmospheric composition included the principal radiatively active gases: water vapor, carbon dioxide, ozone; as well as the minor absorbers such as methane, nitrous oxide, and the major CFCs. Also included were stratospheric and tropospheric aerosols, and multi-layered water and ice clouds. Solar irradiance was programmed with explicit diurnal and seasonal variability. The model surface used 4 horizontal surface types: ocean or lake, ocean-ice, land, and land-ice, and 9 dynamically active vertical layers up to 10 mb, topped off by 3 radiative equilibrium layers. The model included latent, sensible and precipitation heat exchange between the surface and the atmosphere; liquid water run-off and snow accumulation; and full accounting of water and heat stored in atmosphere, ocean, ground, and lake reservoirs.

The model's horizontal resolution and surface topography were described on a $8° \times 10°$ latitude-longitude grid (right panel). Special geographic areas (labeled 1 to 23) were selected for collecting regional on-line monthly diagnostics for more focused analysis. The four grid-boxes (shown in black) were designated to collect hourly weather-type information to assess and evaluate model variability.

Slide 3 The standard version of GISS ModelE2 coupled atmosphere-ocean climate model (*Schmidt et al.*, 2014) operates on a $2° \times 2.5°$ latitude-longitude grid, but can work at higher or lower resolution, if required. It utilizes a vertical resolution of 40 layers with the model top at 0.1 hPa, with adjustable layer spacing (left panel) to provide higher vertical resolution in the more critical near-surface and tropopause regions of the atmosphere. ModelE2 includes all significant radiative, thermodynamic and dynamic physical processes and interactions, as schematically depicted in the right panel (*IPCC-AR5*, *Fig.* 2.11). Numerous model improvements since 1983 include a dynamical ocean, river flow, cloud physics, stratospheric and tropospheric aerosols, and accurate radiative transfer calculations. Non-standard versions of ModelE2 include interactive chemistry of aerosols and ozone, dynamic vegetation, a carbon cycle, and stable water isotopes with fractionation – all important aspects of the increasing physical realism in climate model development. ModelE2 is an indispensible research tool to study the physical processes of the climate system, to reconstruct historical and geological climate changes, and to understand and predict the climate of our ever evolving planet.

Slide 4 Seasonal variability of the surface temperature, TS, between January (left) and July (right). TS best characterizes the prevailing habitability of the biosphere; animal and plant life is strongly constrained by TS even if it lives in the ocean. On a daily basis, TS can undergo large variations due to solar heating by day, and the effects of storms and weather patterns. In general, the Sun heats the ground which in turn heats the atmosphere via sensible, latent, and thermal radiation fluxes. TS is established as the balance between these competing fluxes and is buffered by the large heat capacity of the ocean or the smaller capacities of the land and atmosphere. Note that the tropical regions remain warm all year long. The oceans undergo relatively small changes in temperature, while continental areas have large seasonal changes. The global mean TS is 4 °C higher in July than in January, even though the July solar irradiance is 6.5% less in January.

Slide 5 Atmospheric temperature decreases steadily with height, from the surface until the minimum temperature tropopause which ranges in altitude from 12 to 20 km. Above the tropopause, temperature increases with height until the stratopause near 50 km, then (beyond the graph) decreasing until the mesopause near 80 km, and finally increasing again into the exosphere. Radiative cooling is the principal cause for temperature decreasing with height. Were it not for ultra-violet absorbing ozone concentrated in the stratosphere, temperature would decrease monotonically toward space. Radiatively, the larger thermal opacity, the steeper the temperature gradient. Dry convective instability limits the temperature gradient to be no steeper than about 10 °C/km, but latent heat released by condensing water vapor in the troposphere limits the most-convective temperature gradient to approximately 5-6 °C/km. Tropospheric temperatures are driven by insolation with convective and advective redistribution of energy. Stratospheric temperatures are close to radiative equilibrium, balanced by solar ozone heating and thermal CO_2 cooling.

Slide 6 At any latitude and pressure, the mass stream function is computed as the northward mass flux integrated over longitude and from top of atmosphere down to the pressure level. Negative (counter-clockwise flow on the graph) and positive values of the stream function closest to the equator are described as branches of the Hadley cell. The rising warm air between the branches indicate position of the ITCZ (inter-tropical convergence zone) which varies on either side of the equator and is driven by the seasonal variation of maximum insolation. The ITCZ is where the largest rain storms occur, when rising moist air condenses below the tropopause; this causes the summer Asian monsoons. Just outside branches of the Hadley cell are the weaker Ferrel cells, in which air flows in the opposite direction from the adjacent Hadley cell branch. Sinking dry air between the Hadley and Ferrel cells tends to cause arid and dessert conditions on the surface. Outside the Ferrel cells are the polar cells which are weaker still.

Slide 7 Conservation of angular momentum causes surface winds from the Hadley cell heading toward the equator to be redirected toward the west; this causes the tropical trade winds. Similarly, poleward heading surface winds from the Ferrel cell are redirected eastward. Angular momentum discrepancies between tropics and mid-latitudes also drive the mid-latitude jets. The right-hand plot above is an average of ten Januarys; the jets meander, but are much more confined on an individual day. A property of fluid dynamics is that to transport fluid through a medium, it is most efficient to transport the fluid by means of a narrow jet. The stratospheric zonal wind velocity reverses sign with season. However, the tropospheric jet streams maintain their west to east direction, but are substantially stronger during the winter season. The logarithmic vertical coordinate in the plot is pressure (mb).

Slide 8 The amount of water vapor that air can hold is an exponential function of temperature, increasing as temperature increases. This dependence is governed by the Clausius-Clapeyron formula such that for every increase in temperature by 10 °C, the water vapor holding capacity by the atmosphere doubles. Thus, as the planet warms, the atmosphere can, and will, retain more water vapor. Water vapor is an important greenhouse gas, and is radiatively a stronger contributor than CO_2. This makes water vapor a very important climate feedback effect in response to the imposed forcing of CO_2. In the 4 time CO_2 scenario illustrated above, water vapor contribution to the total greenhouse warming effect far exceeds the CO_2 contribution.

Slide 9 Evaporation is greater in the tropics than over high latitudes, and greater over oceans or lakes than over ground or ice surfaces. Some of the water on continents runs into rivers and lakes and eventually into the ocean as opposed to evaporating. Vaporization of liquid water or ice cools the ground and the latent energy of water vapor escapes into the atmosphere; this occurs more readily when the air is warm or dry. Moist air is transported upward into the atmosphere and also toward the poles and to drier continents. Precipitation occurs when moist air cools. The amount of water vapor that the air can hold is inversely proportional to pressure, and is strongly governed by the exponential temperature dependence as set by the Clausius-Clapeyron formula.

Slide 10 With severe climate change, the distribution of ground water on Earth will be altered. Some lakes may dry up while other may expand. The present climate model predicts that with 4 times CO_2, the eastern half of the United States will become wetter and the western half will be drier; the Amazon and African rain forests will be wetter, but Greenland and Antarctica will be drier. Soil moisture is determined largely by the balance between precipitation and evaporation. In general, the Clausius-Clapeyron formula states that in a warmer climate the atmosphere can hold more water vapor, implying the likelihood of increased precipitation. But a warmer climate also increases the efficiency of evaporation. Shifting patterns in the atmospheric circulation also play an important role.

Slide 11 Sensible heat flux is proportional to the difference between the ground temperature and the surface air temperature. Greater wind speed increases this flux and greater atmospheric stability decreases it. Greater instability leads to convection cells and eventually to turbulence in which sinking cold air encounters the ground that increases the sensible heat flux. Water vapor heat flux includes both the sensible and latent heat content of the vaporized ground water and ice that escapes into the atmosphere. Negative water vapor flux indicates that dew exceeds evaporation; moist air condenses onto the cold ground, warming it because latent heat is released. The sensible and latent heat fluxes are important mechanisms, along with radiation, in establishing the vertical as well as the horizontal temperature structure of the atmosphere.

Slide 12 Clouds are minute particles of liquid water or ice suspended in the air. Clouds are important regulators of the Earth's global energy balance. During the day, clouds reflect sun-light, thus cooling the planet. Clouds at night-time keep the planet warmer by reducing the escape of thermal radiation to space. The high clouds are typically cirrus clouds made up of ice particles. They reflect some sun-light, but their greater affect is to reduce the outward going thermal radiation by absorbing some of it and reemitting some of it back downward toward to warm the surface. For the optically thicker low clouds, reflection of solar radiation is more important. Since thermal emission depends on the cloud temperature and the temperature of low clouds is closer to that of the ground than for cirrus clouds, low clouds are less able to reduce the outgoing thermal radiation.

Slide 13 Albedo is the ratio of the (spectrally integrated) reflected sun-light divided by incident sun-light. The global albedo cannot computed by area-weighting the local albedo. Instead, both the numerator and denominator need to be area-weighted separately, and then the ratio will be an accurate representation of the global albedo. The main difference between planetary (top-of-atmosphere) albedo and ground albedo is the intervening large influence of the atmosphere. Clouds, aerosols, and Rayleigh scattering by air molecules are the principal atmospheric contributors to the planetary albedo. After large volcanic eruptions, sulfuric acid particles in the stratosphere can also reflect sun-light to increase the planetary albedo and thus cool the Earth. For the most part, the atmosphere acts to increase the albedo, although in some locations like Greenland and Antarctica, the surface albedo is higher than the planetary albedo.

Slide 14 The global annual insolation on the Earth is about 340 W/m^2. About 100 W/m^2 is reflected back to space leaving 240 W/m^2 is absorbed by the Earth. Of this, about 68 W/m^2 is absorbed by the atmosphere and 172 W/m^2 by the ground surface. There is an additional 1 W/m^2 globally that is received by the ground from warm rain. To maintain global energy balance equilibrium, 173 W/m^2 must be extracted from the surface by sensible, latent, and thermal radiation heat fluxes. Most of the solar radiation is absorbed in the tropical regions, with progressively less solar radiation absorbed at higher latitudes. By comparison, thermal radiation is more uniformly distributed with latitude. This implies that there is atmospheric and ocean transport of heat horizontally, moving heat energy from the tropics to the poles. For the Earth as a whole to maintain energy equilibrium, 240 W/m^2 must be emitted out to space via thermal radiation.

Slide 15 The greenhouse effect (W/m^2) is defined as the difference between the upward thermal radiation emitted from the surface minus thermal radiation emitted at top-of-atmosphere out to space. The strength of the greenhouse effect is a measure of the heat trapping efficiency of spectral absorption by greenhouse gases and by longwave cloud opacity. In the present model, contributions to the greenhouse effect are roughly 55% by water vapor, 22% by clouds, 18% by CO_2, and 5% by other minor (non-condensing) greenhouse gases such as methane, nitrous oxide, and ozone. Though water vapor and clouds are by far the largest contributors to the greenhouse effect, their contributions are feedback effects in response to changing temperature. The Clausius-Clapeyron formula causes water vapor to increase more rapidly than the applied radiative forcing due to the increase in CO_2, and other non-condensing greenhouse gases.

Slide 16 The greenhouse effect (W/m^2) is defined as the difference between the upward thermal radiation emitted from the surface minus thermal radiation emitted at top-of-atmosphere out to space. The strength of the greenhouse effect is a measure of the heat trapping efficiency of spectral absorption by greenhouse gases and by longwave cloud opacity. In the present model, contributions to the greenhouse effect are roughly 55% by water vapor, 22% by clouds, 18% by CO_2, and 5% by other minor (non-condensing) greenhouse gases such as methane, nitrous oxide, and ozone. Though water vapor and clouds are by far the largest contributors to the greenhouse effect, their contributions are feedback effects in response to changing temperature. The Clausius-Clapeyron formula causes water vapor to increase more rapidly than the applied radiative forcing due to the increase in CO_2, and other non-condensing greenhouse gases.

Slide 17 The reference geoid for the ocean surface is a level where the surface pressure gradient force is zero. This includes effects due to variations of both gravity and of the Earth's rotation. The ocean surface height differs (left-hand panel) from this geoid because of wind stresses and local steric expansion. The reductions in stability of the water column and momentum stresses at the top cause mixing in the upper ocean layers, from the ocean surface down to the mixed layer depth. There is a deficiency in the present ocean model in that there is insufficient mixing in the Labrador Sea, and too much mixing in the Greenland Sea (right-hand panel). These deficiencies arise from the complex interactions of the ocean circulation with sea ice formation, melting, and transport, and with changing fresh water inputs and radiative environment. Improved ocean modeling is a key factor in improved climate model reliability.

Slide 18 Surface salinity is low in the Arctic Ocean because of the significant fresh water river input and low evaporation. Salinity is also low adjacent to California because the ocean temperature there is cold which reduces evaporation. On the other hand, surface air crossing the Sahara and Arabian deserts is very dry.

When this air encounters the ocean, it causes enhanced evaporation resulting in greater salinity. High salinity water, escaping at lower depths from the Mediterranean Sea at Gibraltar, spreads all the way across the Atlantic Ocean. Salinity, because it directly affects the density of sea water, along with the water temperature are the principal are the principal controlling factors that affect the global circulation of the ocean, and thus the transport of heat energy pole-ward from the tropics.

Slide 19 In both the atmosphere and ocean, horizontal velocity is prognostic whereas vertical velocity is computed diagnostically from the convergence of horizontal mass fluxes. This computation is sensitive to numerical instabilities and checker-boarding which is reduced by numerical filters, although it is still noticeable in the atmosphere. Most of the large vertical velocities occur near mountainous regions in the atmosphere (left panel) and next to coast lines in the ocean (right panel). Both water vapor and latent heat energy, which are generated at the surface by evaporation, must first be transported vertically before they are dispersed horizontally by atmospheric winds. In the oceans, near-surface and deep water circulation patterns are quite different, and must be bridged in key locations by vertical velocities.

Slide 20 The side of the Earth facing the Moon is more strongly pulled by the Moon's gravity and the opposite side is pulled more weakly. These three dimensional variations of gravity, by the Moon and Sun, cause the Earth's tides. The maximum horizontal tidal acceleration is about 10^{-7} (m/s^2). Over three hours, the change in velocity can be as large as 1 (mm/s) at time-varying locations. Tides were ignored when the video starts, ocean surface height was relatively calm, but then tides are invoked. Notice the counter-clockwise motion in the North Atlantic. (Click on power-point figure to activate lunar tide movie.)