Figure 1.1 Top: The incoming solar radiation impinges on a disk of area *a*2 but is on average spread out over a sphere of area 4*a*2. Bottom: Variation of incoming solar radiation with latitude. A given amount of radiation is spread over a larger area at high latitudes than at low latitudes, so the intensity of the radiation is diminished, and thus high latitudes are colder than low latitudes.

Figure 1.2 Earth’s orbit around the sun and the march of the seasons. Earth’s axis of rotation is at an angle with respect to the axis of rotation of Earth around the sun. The Northern Hemisphere’s summer and the Southern Hemisphere’s winter result when the North Pole points toward the sun, and the opposite season occurs six months later. The eccentricity is much exaggerated in the figure.

Figure 1.3 The energy budget of Earth’s atmosphere, showing the average solar and longwave radiative fluxes per unit area and the convective flux from the surface to the atmosphere. Adapted from Kiehl and Trenberth 1997. xxx

Figure 1.4 An idealized two-level energy-balance model. The surface and the atmosphere are each characterized by a single temperature, *Ts* and *Ta*. The atmosphere absorbs most of the infrared radiation emitted by the surface, but it is transparent to solar radiation.

Figure 2.1 Schematic of the configuration of the oceans and continents over the past 225 million years, since the breakup of the supercontinent Pangaea. Source: Adapted from USGS (<http://pubs.usgs.gov/gip/dynamic/historical.html>). xxx Change Pangaea to Pangea xxx

Figure 2.2 The annual average temperature at the ocean surface, in degrees centigrade. Adapted from World Ocean Atlas, 2009 of the National Oceanic and Atmospheric Administration (http://www.nodc.noaa.gov/OC5/WOA09/pr\_woa09.html).

Figure 2.3 A schematic of the main surface currents of the world’s oceans. The panel at the left shows the zonally averaged zonal (i.e., east–west) surface winds.

Figure 2.4 The zonally averaged density in the Atlantic Ocean. Note the break in the vertical scale at 1,000 m.3 The actual density is approximately 1000 plus the numbers shown, in kg/m3.

Figure 2.5 Schematic of the vertical structure of the ocean, emphasizing the mixed layer. In the mixed layer, typically 50–100 m deep, turbulence and convection act to keep the temperature relatively uniform in the vertical. Below this layer, temperature changes over a depth of a few hundred meters, in the *thermocline,* before becoming almost uniform at depth, in the *abyss*.

xxx Adapted from Marshall and Plumb 2007.

Figure 2.6 Top: An artist’s impression of the global ocean circulation, sometimes called the “conveyor belt.” Bottom: The sea-surface height in the Atlantic on October 15, 2008, indicating the presence of the Gulf Stream and mesoscale eddies.

Figure 3.1 A missile launched from the North Pole toward Africa. Earth rotates beneath the missile, and the missile lands in South America. From the point of view of an observer on Earth, the missile has been deflected to its right, and the force causing that deflection is the Coriolis force.

Figure 3.2 The component of Earth’s rotation in the local vertical direction varies with latitude () like . Its value is at the North Pole, zero at the equator, and at the South Pole. The Coriolis parameter *f* is given by . xxx Check symbols are consistent with figure.

Figure 3.3 A slab (dark shading) floating within a fluid, with *x* and *z* the horizontal and vertical directions, respectively. The force to the right is just the difference of the pressure forces between the right and left surfaces of the slab, and so proportional to Thus, the net force is proportional to the pressure gradientwithin the fluid.

Figure 3.4 An idealized Ekman spiral in the Northern Hemisphere.

Figure 3.5 A body moving in a circle is constantly changing its direction and so accelerating. The acceleration is directed toward the center of the circle and has magnitude *v*2*/r*, where *v* is the speed of the body and *r* is the radius of the circle.

Figure 3.6 An astronaut orbiting Earth. Panel a views the motion in a stationary frame of reference, in which Earth’s gravitational force provides the centripetal force that causes the astronaut to orbit Earth. Panel b views the situation from the astronaut’s frame of reference, in which the gravitational force is exactly balanced by the centrifugal force and the astronaut feels weightless.

Figure 4.1 An idealized gyre circulation in a rectangular ocean basin in the Northern Hemisphere, showing the subtropical gyre (lower, typically extending from about 15° N to 45° N), the subpolar gyre (upper), and the intense western boundary currents on the left.

Figure 4.2 Production of gyres by winds. The winds blowing as shown induce a converging Ekman flow, causing the sea level to increase in the center, thus giving rise to a pressure gradient. This gradient in turn induces a geostrophic flow around the gyre, in the same sense as the winds themselves.

Figure 4.3 Two schematics of a subtropical gyre. The left panel shows the basic response of the circulation to the winds shown, and the right panel shows the gyres in the presence of differential rotation, with western intensification.

Figure 4.4 The production of a western boundary current. Schematic of the torques (namely, the spin-inducing forces: the wind, W; Coriolis, C; and friction, F) acting on parcels of water in the ocean interior (center) and western and eastern boundary layers (left and right), in a Northern Hemisphere subtropical gyre. In the interior, friction is small and the torques balance if the flow (denoted V) is southward. If the northward return flow is in the west, then a balance can be achieved between friction and Coriolis forces, as shown. If the northward return flow is in the east, no balance can be achieved.

Figure 4.5 If parcel A is displaced northward, then its clockwise spin increases, causing the northward displacement of parcels that are to the west of A. A similar phenomenon occurs if parcel B is displaced south. Thus, the initial pattern of displacement propagates westward.

Figure 4.6 Schema of the two main components of the MOC. Top: The mixing-maintained circulation. Dense water at high latitudes sinks and moves equatorward, displacing warmer, lighter water. The cold, deep water is slowly warmed by diffusive heat transfer (mixing) from the surface in mid- and low latitudes, enabling it to rise and maintain a circulation. Bottom: Winds over the Antarctic Circumpolar Current (outlined by dashed lines) pump water northward, and this pumping enables deep water to rise and maintain the circulation. In the absence of both wind and mixing, the abyss would fill up with the densest available water and the circulation would cease.

Figure 4.7 Schematic of the flow in the Antarctic Circumpolar Current (ACC). The wind predominantly blows in a zonal direction around the Antarctic continent, generating an Ekman flow toward the north and a net loss of water from the channel. The water returns at depth, generating a deep overturning circulation, as illustrated in figure 4.8. xxx change to ‘as illustrated in the bottom panel of figure 4.6’

Figure 4.8 Schematic of the meridional overturning circulation, most applicable to the Atlantic Ocean (D.P. indicates the Drake Passage, the narrowest part of the ACC). The arrows indicate water flow, and dashed lines signify water crossing constant-density surfaces, made possible by mixing. The upper shaded area is the warm water sphere, including the subtropical thermocline and mixed layer, and the lower shaded region is Antarctic Bottom Water. The bulk of the unshaded region in between is North Atlantic Deep Water.

Figure 5.1 The seasonal cycle of temperature (°C) in San Francisco and New York. For each city, we plot the average low temperature and the average high temperature for each month. Note the much bigger range in New York and the maximum earlier in the year, in July rather than September.

Figure 5.2 Amplitude and lag of the annual cycle in the Northern and Southern hemispheres, as a function of latitude. The lag is the time, in days, from the maximum solar insolation to the maximum temperature. Source: Trenberth, 1983.

Figure 5.3 a: xxx Meridional xxx heat transport in the total atmosphere–ocean system (solid line), in the ocean (dashed line), and in the atmosphere (dotted line). b: Oceanic heat transport, subdivided into the various basins. Source: Trenberth and Caron, 2004.

Figure 6.1 A schematic of the two phases of the NAO. The left panel shows the positive phase of the NAO, with storms tracking northward bringing mild but wet weather to northern Europe. The right panel shows the negative phase, with storms tracking southward bringing wet weather to southern Europe and colder weather to northern Europe.

Figure 6.2 The NAO index from 1860 to 2005. The index is based on the normalized difference in average winter surface pressure between Lisbon, Portugal, and Stykkishólmur, Iceland. The heavy solid line shows the index after it has been smoothed to remove fluctuations of less than four years.

Figure 6.3 The sea-surface temperature in December of a normal (i.e., non–El Niño) year (December 1996, top panel); in a strong El Niño year (December 1997, middle panel); and their difference (bottom panel). A normal year is characterized by a cold tongue of water in the eastern tropical Pacific, which disappears in El Niño years.3

Figure 6.4 Top: A time series of the sea-surface temperature (SST) in the eastern equatorial Pacific region (specifically, in the so-called Niño 3 region). The spiky curve shows the annual means, and the dots represent Decembers. The smoother curve shows the SST after the application of a 20-year low-pass filter, and the top of the gray bar is the 1876–1975 mean. Bottom: A similar plot for the negative of the Southern Oscillation Index (SOI), the anomalous pressure difference between Tahiti and Darwin. Particularly large El Niño events can be seen in 1877–78, 1982–83, and 1997–98.4

Figure 6.5 Schema of the atmosphere and ocean at the equator in the Pacific, during La Niña/normal conditions (top) and El Niño conditions (bottom). La Niña conditions are similar to normal conditions but with a still steeper-sloping thermocline and maximum SSTs a little further west.

Figure 7.1 The instrumental record of global average surface temperatures from 1880 to 2009, relative to the mean temperature from 1951 to 1980.

Figure 7.2 Top: Lower troposphere temperature as measured by various satellites and by radiosondes; the gray shading indicates the spread between all measurements. Bottom: Surface temperature records from NOAA, NASA, and UKMO, with gray shading again indicating the spread. Records are monthly means, smoothed with a seven-month running mean filter, and are relative to 1979–1997 mean. xxx Adapted from Solomon *et al.*,2007.

Figure 7.3 Reconstructed temperatures of the past 1,300 years. The solid curve extending from about 1850 to 2000 shows a reconstruction from the instrumental record. The gray region shows the overlap between various temperature reconstructions using proxy data; darker shades indicate more overlap and may be regarded as a measure of confidence in the reconstructions. The series are smoothed to remove fluctuations of periods shorter than 30 years, and the temperatures represent anomalies (in °C) from the 1961–1990 mean. Source: Adapted from Solomon *et al.*,2007.

Figure 7.4 The levels of CO2 measured at Mauna Loa Observatory in Hawaii from 1958 to 2010 in parts per million by volume.

Figure 7.5 Schematic of temperature profiles before and after the addition of greenhouse gases. The total outgoing longwave radiation must remain the same because this radiation balances the incoming solar radiation, and so the emissions temperature, *Te*, stays the same. However, the emissions height must increase (from *Z*1 to *Z*2) because of the increased absorptivity of the atmosphere. Hence, if the temperature gradient in the vertical remains similar, the surface temperature must increase.

Figure 7.6 Top: The global heat content for the upper 700 m of the ocean (solid black line, with shading indicating uncertainty) and upper 100 m of the ocean (dotted line, with thin solid lines indicating uncertainty). Bottom: Increase in sea level as estimated from direct measurements (black line and shading) and by combining the contributing components (dotted line and thin solid lines). The time series are all relative to 1961 and smoothed with a three-year running average. Source: Adapted from Domingues *et al.*,2008.

Figure 7.7 Sea ice cover for the Northern Hemisphere from satellite data. Perennial ice excludes seasonal ice cover, and multi-year ice accounts only for ice that has existed for more than one season. Source: Adapted from Comiso, 2002, and Comiso *et al.*,2008.

Figure 7.8 A simple two-box model of the ocean, with a mixed layer at a temperature *Tm* and a deep ocean layer at a temperature *Td*, and exchanges of heat between the components as shown.

Figure 7.9 Schema of a CO2-temperature scenario. Carbon dioxide levels increase from 1900 to 2100 (period B) before leveling off (period C) because of controls on emissions. Temperature increases rapidly in period B, then more slowly in period C. At the end of period C (the year 2300 in the figure), anthropogenic emissions go to zero, and the level of CO2 slowly diminishes through periods D and E back to levels close to, but probably a little above, the preindustrial period. In period D, temperature stays roughly constant for centuries before it too eventually falls back to near pre-industrial levels in period E. Almost all plausible scenarios can be adapted from this plot by changing 2100 and 2300 to other dates and calibrating the *y*-axis. xxx Please change ‘Almost all’ to ‘Many’