5.4: Coupled Models and Climate Projections

Peter R. Gent
National Center for Atmospheric Research,
Boulder, Colorado, USA.
gent@ucar.edu

5.4.1. Formulation of Coupled Models

Coupled climate models consist of atmosphere, land and sea ice components, as well as the ocean component. The atmosphere component is similar to the ocean in that it solves the primitive equations of motion, usually with hydrostatic balance, the continuity equation and predictive equations for temperature and specific humidity. The vertical coordinate is usually pressure, or terrain-following, or a combination of both. A comprehensive review of atmosphere components can be found in Randall et al. (2007). The land surface component usually solves equations in the vertical direction for heat and water that include many complex interactions at the land surface and soil layers below. It also calculates water runoff, which is then routed by a realistic horizontal pattern of river basins to create the river runoff field that is sent to the ocean component. The sea ice component solves a complicated equation for the ice rheology and thermodynamic equations for the heat balance of the sea ice. When the ice melts, fresh water is sent to the ocean component, and the ocean receives a brine rejection flux when the sea ice is formed.

The formulation of the ocean component of coupled models has been described in detail earlier in Chapter 5.1. In that chapter it is conjectured, and I concur, that ocean model simulations depend much more strongly on the parameterizations of mixing and the effects of unresolved scales than on details of the numerical discretization of the equations (Chassignet et al., 1996). The parameterization of vertical mixing has been discussed earlier in Chapter 3.3, and the parameterization of unresolved lateral transport in Chapter 3.4.

The first climate model that used realistic geometry was assembled by Syukuro Manabe, Kirk Bryan and co-workers at the Geophysical Fluid Dynamics Laboratory (GFDL), and the results were published in two landmark papers; Manabe et al. (1975) and Bryan et al. (1975). The horizontal grid spacing was 5° x 5°, and there were nine vertical levels in the atmosphere component and five levels in the ocean component. Since then, the components have become much more sophisticated and complex, and the horizontal resolution has increased to about 1° x 1° or finer in present day climate models that use about 30-60 vertical levels in the atmosphere and ocean components.

5.4.2. Flux Adjustments

Through most of the 1990s, present day control runs of all climate models would quite quickly drift away from the realistic initial conditions with which they were initialized. This drift was usually corrected by the use of flux adjustments described in Sausen et al. (1988), which could be calculated in two different ways. The heat and fresh water fluxes between the atmosphere and ocean components were diagnosed when the atmosphere component was run using observed sea surface temperatures (SSTs) and the ocean component was run using observed wind stresses and atmosphere surface variables. These flux diagnoses were quite different, and their differences were the flux adjustments. Alternatively, the coupled model was run with strong relaxation back to observations of SST and sea surface salinity, and these relaxation terms were used as the flux adjustments. The flux adjustments were added to the heat and fresh water fluxes between the atmosphere and ocean in a present day coupled control run at each time step. This method corrected most of the model drift, but was very unsatisfactory because flux adjustments are completely unphysical, mask deficiencies in the atmosphere and ocean components that require their use, and likely give an unrealistic response to large perturbations of the climate system.

The first model that could maintain the present day temperature climate in a control run without the use of flux adjustments was the Community Climate System Model, version 1 (CCSM1). A 300 year control simulation that showed very little drift in surface temperature was run during the second half of 1996, and documented in Boville & Gent (1998). The reason for this success was refinements of the atmosphere, and especially the ocean component, so that the surface heat flux produced by the two components in stand-alone runs was close enough to be compatible. However, the ocean salinity did drift because CCSM1 did not have a realistic river runoff component. A history of this success is contained in a recent review by Gent (2011). Quite quickly, the climate centers in Australia and the United Kingdom implemented two of the new ocean parameterizations from the CCSM1, and were also able to run their models with much reduced, or without, flux adjustments (Hirst et al., 1996, Gordon et al., 2000).

Now, almost all climate models are run without using flux adjustments. A consequence is that current climate models cannot reproduce the present day atmosphere surface temperature and SSTs quite as well as when flux adjustments were used. However, this small detraction is far outweighed by the fact that climate models no longer use the unphysical flux adjustments. For example, in some models the flux adjustment transported nearly as much heat northwards in the North Atlantic Ocean as did the ocean component. This can be seen in Figure 2.3.1 of northward heat transport in the Hadley Centre models with and without flux adjustments shown in Section 2.3.3 of the first edition of this book by Wood & Bryan (2001).

5.4.3. Control Runs

A control run is an integration of a climate model in which the solar input at the top of the atmosphere component has a repeating annual cycle, and all the other forcings have a repeating annual cycle or are kept constant in time. The forcings can include the atmospheric concentrations of carbon dioxide (CO_2) and other greenhouse gases, and the distributions of several aerosols that affect the atmosphere radiation balance. Quite often a control run is made using forcings from present day conditions. This has the advantage that the simulation can be compared to a whole range of observations that have been made in the recent past. However, it is well known that the present-day climate is not in equilibrium because important forcings, namely the greenhouse gas concentrations and aerosol emissions, have been increasing quite rapidly. Therefore, an alternative control run is made for 'pre-industrial' conditions, when it is assumed that the climate was in equilibrium. Pre-industrial is often chosen to be 1850, which is not truly before the industrial revolution. However, 1850 is chosen as a compromise between only a very small increase in CO_2 over truly pre-industrial levels, and reducing the length of climate model runs between pre-industrial times and the present day.

Choosing the initial conditions to start an 1850 control run is not straightforward because the ocean, sea ice and land states for this period are unknown. The atmosphere initial condition is not as important because it is forgotten within a few weeks of the start of the run, so a state from an atmosphere model run forced by observed SSTs is usually chosen. Present day states for the sea ice are usually chosen, although the 1850 sea ice volume is thought to be larger, so that sea ice should grow during the control run. It is important not to put far too much sea ice into the initial condition, because when it melts quickly, it will suppress convection in the ocean component which can take a few hundred years to overcome. It is also important that the water content of the soil in the land component is initialized realistically, because the timescale to change the water content of the deepest soil level can also be a few hundred years. The traditional way to initialize the ocean component is to use a climatology from the late 20th century; Levitus et al. (1998) and Steele et al. (2001). This initial condition has a larger heat content than in 1850, so that the climate model would be expected to lose heat initially from the ocean component and the whole system, although this does not happen in all climate models as some have the ocean gaining heat in 1850 control runs (e.g. Griffies et al. 2011).

The SST acts as a negative feedback on the atmosphere-to-ocean heat flux, so it would be expected that the atmosphere-to-ocean heat flux, and consequently the top of atmosphere heat budget, would converge towards zero after a few decades of an 1850 control run. However, in practice this seldom happens because the ocean heat loss or gain occurs in the deeper ocean and not in the surface layers. The climate model heat loss or gain is often almost constant in time and is difficult to get very close to zero. Most climate modeling groups aim to get the globally-

averaged heat imbalance less than 0.1 W/m², usually by tuning a cloud parameter or two in the atmosphere component (e.g. Gent et al. 2011). Even so, this rate of ocean heat loss or gain sustained over an integration of 1000 years means the ocean volume average temperature decreases or increases by 0.2°C compared to the 3.7°C value in the Levitus et al. (1998) data used to initialize ocean components. Many well constructed climate models conserve water quite well, even though not to machine accuracy. The volume average ocean salinity can vary because of changes in the volume of sea ice and soil water content over the course of the integration. However, the volume average ocean salinity changes only marginally even over a 1000 year control run if water is almost conserved.

One disadvantage of a pre-industrial control run compared to a present-day control run is that there are very few observations relevant to that period. However, an 1850 control run does reveal the internal variability of the climate model given repeating annual cycle forcing, which is important to document. Variability occurs on all timescales from the diurnal to centennial, and this variability can be compared to late 20th century observations. For example, the largest interannual signal in the climate system is the El Nino/Southern Oscillation (ENSO) phenomena. This has been a difficult signal for climate models to simulate well. However, there has been significant progress in recent years. For example, Wittenberg (2009) looks at ENSO in a 2000 year control run of the GFDL CM2.1 climate model, and shows there is decadal and centennial variability in both the frequency and amplitude of the signal. Figure 1 shows the frequency of NINO3 area (90°W-150°W, 5°S-5°N) monthly SST anomalies from observations and the CM2.1 when the run is divided into 20, 100, and 500 year segments. The amplitude has a lot of variability over the 20 year segments, and still significant variability over the 100 year segments of the control run. A period in the run can be found where the model ENSO matches quite well the observed record from 1956 to 2010. The ENSO signal in the recently completed CCSM4 model is significantly improved over earlier model versions

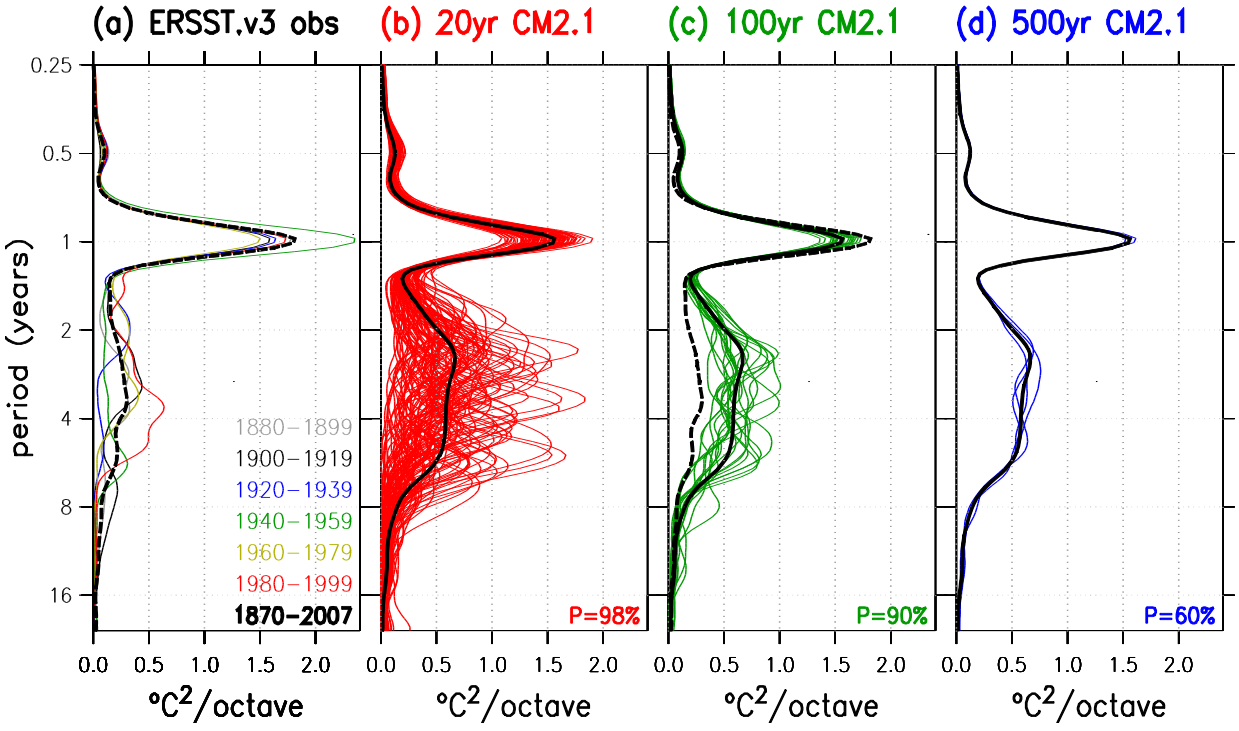

Figure 1. Power spectra of NINO3 SSTs as a function of the period in octaves of the annual cycle. (a) Spectra for six 20 year epochs (solid) and one 138 year epoch (dashed and repeated in (c)) from the ERSST observations, and spectra from the CM2.1 pre-industrial control run for (b) 20 year epochs, (c) 100 year epochs and (d) 500 year epochs, where the thick black solid line is the spectrum from the full 2000 year run. From Wittenberg (2009).

(Neale et al., 2008), and the above comments also apply to ENSO in the 1300 year long CCSM4 1850 control run (Deser et al., 2012). The other purpose of an 1850 control run is to provide initial conditions for an ensemble of integrations of the model from 1850 to the present day, which are often called 20th century runs.

5.4.4. 20th Century Runs

Time series over 1850 to the present day of four or five types of input variables are needed to force 20th century runs. They are the atmospheric concentrations of CO_2 and other greenhouse gases, the level of solar output, the levels of natural and human-made aerosols, the amount of atmospheric aerosols from volcanic eruptions, and some models use historical changes in land use in the land component. The volcanic aerosol level is determined from the observed levels of aerosols from recent eruptions, such as El Chicon in 1982 and Pinatubo in 1991, and then scaled by the size of eruptions earlier in the period, such as the large Krakatoa eruption in 1883. Usually, an ensemble of these 20th century runs is made with each climate model, where the initial conditions are taken from different times in the 1850 control run. These initial conditions are chosen after a few hundred years of the control run, so that the modeled climate system, including the upper kilometer or so of the ocean, has had time to come into equilibrium. If a climate model is to be useful, then its 20th century runs must reproduce well many of the observed changes in the earth's climate over the last 150 years. Most of these comparisons with observations will use the last 50 years of these runs, which is when virtually all of the observations were made.

The best measured quantities going back to 1850 are surface temperature over land and SST. These can be combined into a globally-averaged surface temperature with only a small uncertainty, so that it is well known that the earth's surface temperature has increased by about 0.7°C over the time 1850 to 2000. Figure 2 shows both the observations and the ensemble of five 20th century simulations using the GFDL CM2.1 model. The CM2.1 ensemble mean reproduces the observations rather well. There are an enormous number of other quantities that can be

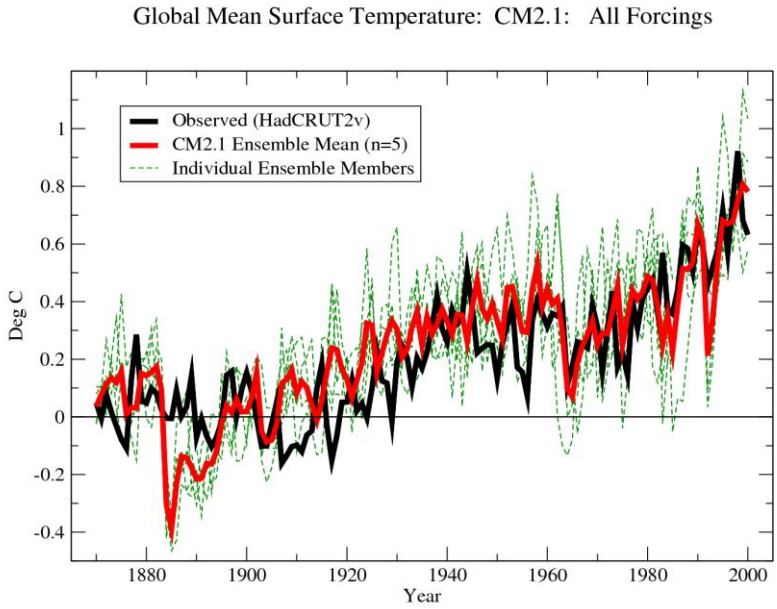

Figure 2. Globally-averaged surface air temperature over 1870 to 2000 from the HadCRUT2v observations and both the individual members and the ensemble mean from 20th century runs using the GFDL CM2.1 model. From Knutson et al. (2006).

compared to observations, and choices have to be made about which are compared because they are most important. For example, satellite observations of Arctic sea ice between 1979 and 2012 have shown a very significant decrease in the ice area during September, which is the month when the ice area is a minimum. If the climate model is to be used to project the future state of Arctic sea ice, including when the Arctic Ocean might be virtually ice free in September, then the model must reproduce the observations well within the ensemble of 20th century runs. Figure 3 shows the September Arctic sea ice area from satellite observations (dotted line), and the ensemble mean and spread of 20th century runs using the CCSM version 4 (solid line and shading) through 2005, which is when the 20th century runs end. The CCSM4 reproduces the observed decline of September sea ice quite well.

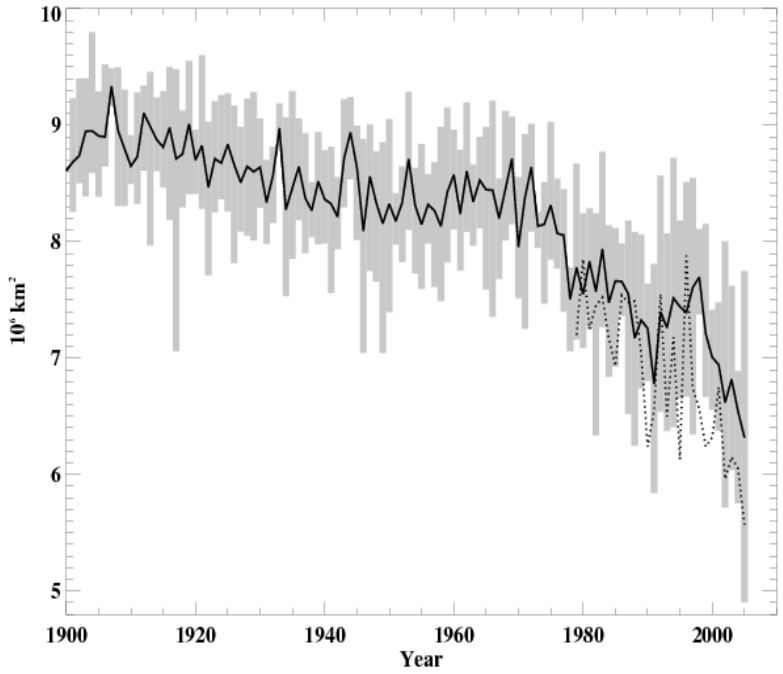

Figure 3. Area of Arctic sea ice in September from satellite observations (dotted line) and the ensemble mean (solid line) and spread (gray shading) from six 20th century runs using the CCSM4. From Gent et al. (2011).

Another use of 20th century runs is to determine the cause of the earth's warming that has increased markedly since the mid 1970s. Additional runs are made where all the natural forcings are retained and the levels of the greenhouse gases and man-made aerosols are kept at their pre-industrial values, and the converse of this run where just the anthropogenic forcings are allowed to vary. Figure 4 shows the globally-averaged surface temperature from ensembles of these two runs using the GFDL CM2.1 model. The conclusion is that the all natural forcings runs cannot explain the accelerated warming since 1975, whereas the anthropogenic forcings runs explain the observed temperature rise within the internal variability of the model. The confidence with which this result can be stated depends on the amplitude of the model internal variability and the length of time since the model surface temperature started to increase markedly. The increase in the model surface temperature over the last 35 years is now above two standard deviations of most models' internal variability. The 4th Assessment Report of the Intergovernmental Panel on Climate Change (IPCC), (Solomon et al., 2007), concluded that it is now 'very likely' that the recent warming is due to anthropogenic causes. The level of certainty about this conclusion has increased in successive IPCC reports as, over time, the size of the observed temperature increase has become considerably larger compared to climate models' internal variability. It is important to note here that there is considerable uncertainty about the parameterization of clouds and the interaction of aerosols and clouds in the atmosphere component of climate models. A detailed discussion of these topics can be found in Chapters 1 and 2 of Solomon et al. (2007).

5.4.5 Future Projections

In order for a climate model to run projections, the future levels need to be estimated of the same four or five types of input variables that were required for 20th century runs. The future level of solar output is usually chosen to be constant, although a very small 11 year cycle is sometimes imposed. Future volcanic eruptions are unknown, so a constant background level of volcanic aerosols is usually used. The future levels of natural aerosols are often chosen to be constant, along with a gradual reduction of man-made aerosols later in the 21st century. Finally, a scenario for the future atmosphere concentrations of CO_2 and other greenhouse gases has to be chosen. There is much debate and uncertainty about whether future emissions of CO_2 will remain at their present levels, accelerate over the next few decades, or reduce in the second half of the 21st century as the world turns to alternate energy sources. Because of this uncertainty, climate models have been run with a wide variety of scenarios for the future

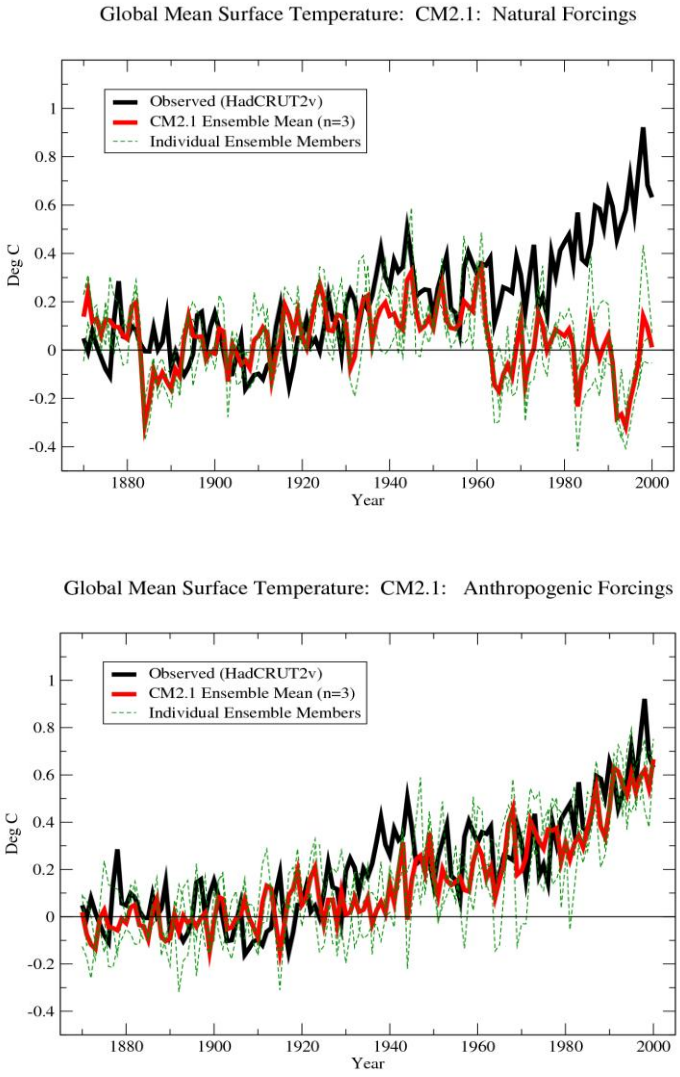

Figure 4. Globally-averaged surface temperature 1870-2000 from observations, and ensembles of (a) all natural forcing runs, and (b) anthropogenic forcings runs using the GFDL CM2.1. From Knutson et al. (2006).

concentrations of CO_2. These concentrations are usually based on a fixed percentage of the CO_2 emissions remaining in the atmosphere, or are the output of integrated assessment models which use CO_2 emissions as input. This approach ignores the possible feedbacks from the land and ocean components, which may take up less or more of the emitted CO_2 in the future compared to the past. If a climate model is to predict the future CO_2 concentrations given the emissions, then it must have an interactive carbon cycle that can predict the future levels of CO_2 uptake by the land and ocean components (see Chapter 6.4).

An informative future projection is obtained if the concentrations of CO_2 and other greenhouse gases are kept constant at their present day values. This is called a 'commitment' run, because it shows the future changes that we have committed to by raising the atmosphere CO_2 concentration to its present level. Assuming no future changes in the forcings, most climate models simulate an increase in the globally-averaged surface temperature of between 0.2° and 0.4°C over the next 100 years, see Figure 5b. This increase depends on the model's equilibrium climate sensitivity, sensitivity to aerosols, and the rate of ocean heat uptake. The equilibrium climate sensitivity is defined as the surface temperature increase due to a doubling of the CO_2 concentration when the ocean component is a simple mixed layer model. Using a mixed layer model ensures an equilibrium response after about 30 years, rather than the order 3000 years if the full depth ocean component is used, (Stouffer 2004, Danabasoglu and Gent 2009). It is interesting to note that the future temperature rise is very small after about 50 years of a commitment run. This is in sharp contrast to the future heat uptake by the ocean component and the resulting sea level rise due to thermal expansion, which increase almost linearly with time over the entire duration of commitment runs, see Figure 5c. As stated above, the time for the full ocean to reach equilibrium is on the order of 3000 years, so that the ocean heat content will not start to asymptote away from the linear increase for well over 1000 years. A similar, extremely long timescale is also relevant for glaciers and ice caps to come into equilibrium with the temperature increase that we have committed to. The conclusion is that, even if future temperature rises are stabilized by reductions in CO_2 emissions, the sea level rise we have committed to will continue for the next 1000 years. This extremely long timescale for sea level rise is frequently not appreciated; more discussion is in Chapter 6.1.

Figure 5a shows the future CO_2 concentrations from three different scenarios and Figure 5b shows the resulting globally-averaged surface temperatures. As is to be expected, the faster the rise in greenhouse gases, the faster the rise in surface temperature. The rate of temperature increase varies across climate models depending upon their equilibrium climate sensitivity, sensitivity to aerosols, on the rate of heat uptake by their ocean components, and several other factors. Some future projections have been run where the CO_2 concentration reaches a maximum and is then kept constant. The temperature increases quickly, but then only increases slowly when CO_2 is constant, and so do other important quantities such as the rate of decrease of Arctic sea ice. However, as shown in Figure 5c, ocean heat uptake, and the resulting rise in sea level due to thermal expansion, do not level off; the rate of heat uptake only decreases slightly once the CO_2 concentrations are kept constant because of the very long ocean adjustment timescale.

5.4.6 North Atlantic Meridional Overturning Circulation

The Gulf Stream transports warm water northwards in the North Atlantic. Most of this water returns to the south in the gyre circulation in the upper ocean, but some is returned southwards in a vertical circulation. Deep water is formed in the Labrador and Greenland/Iceland/Norwegian Seas and is then returned south in the deep western boundary current along the west side of the North Atlantic. This meridional overturning circulation (MOC) is important because it carries a majority of the heat that is transported towards the Arctic by all the oceans in the Northern Hemisphere. This heat transport modulates the climate of Western Europe compared to other land masses at the same latitude. An equivalent, large-scale vertical overturning circulation does not occur in the North Pacific Ocean. An important question is how future increases in CO_2 will affect modes of climate variability, such as the North Atlantic MOC?

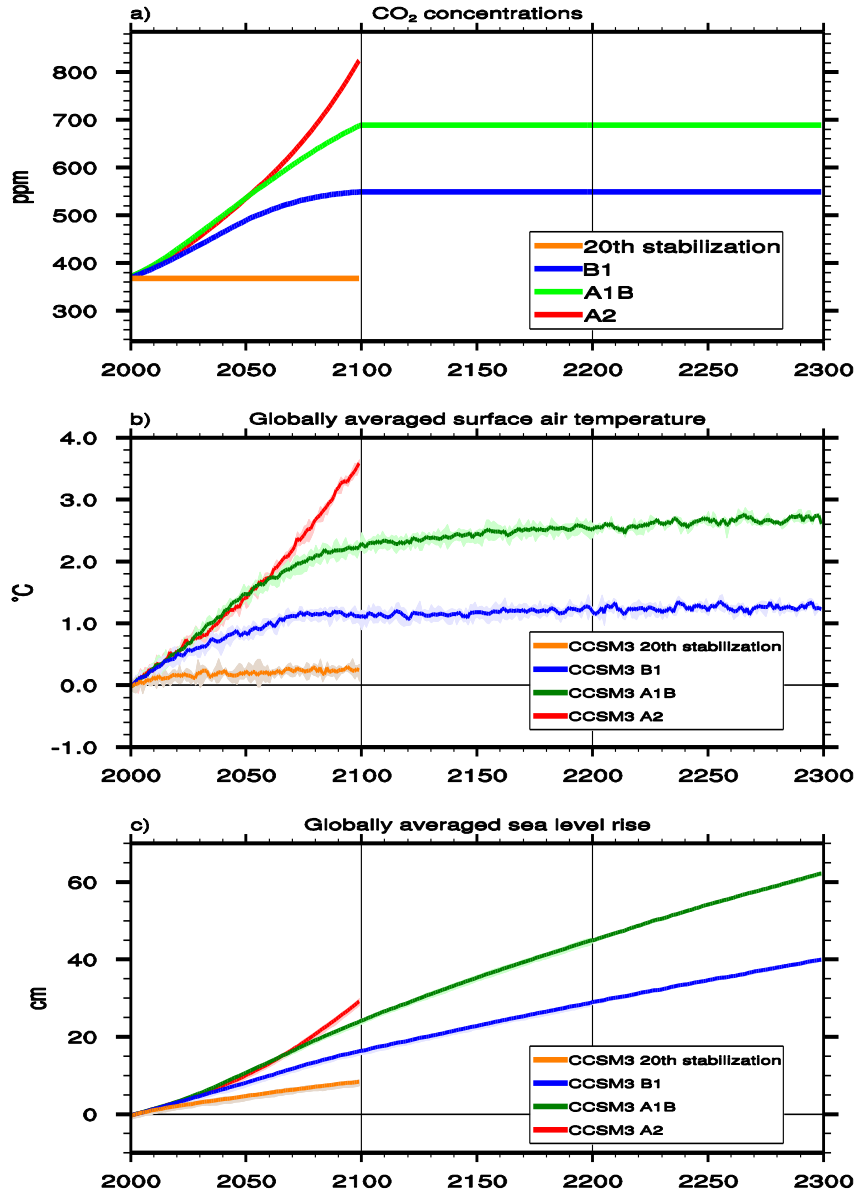

Figure 5. a) CO_2 concentrations from four future scenarios, b) the resulting globally-averaged surface temperature increase, and c) the globally-averaged sea level rise due to thermal expansion from runs using the CCSM3. From Meehl et al. (2006).

In models, the MOC is characterized by the streamfunction formed by the meridional and vertical velocities when the continuity equation is integrated zonally across the ocean basins. A typical Atlantic Ocean MOC streamfunction from the CCSM4 is shown in Figure 6. It shows water moving northwards in the upper kilometer, sinking down to between 2 and 4 km between 60° and 65°N, and then returning southwards. Note that a large fraction of this return flow crosses the equator into the Southern Hemisphere. Below this deep return flow, the negative contours show a weak northward flow near the ocean floor, which is the model representation of Antarctic Bottom Water flowing northwards. The magnitude of the North Atlantic MOC is virtually always taken to be the maximum value of the overturning streamfunction in Sverdrups (Sv), which is usually at a depth of about 1km. Note that this maximum value cannot be directly measured from observations, although estimates can be made both from observations at particular latitudes (Cunningham et al., 2007) and using assimilation models of ocean circulation (Wunsch & Heimbach, 2006).

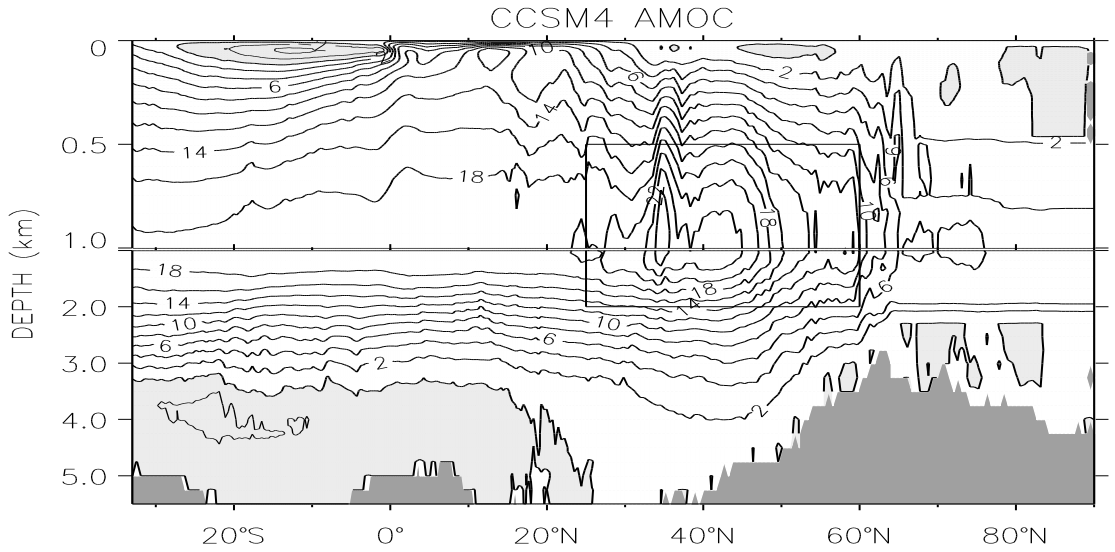

Figure 6 North Atlantic meridional overturning streamfunction in Sv from the CCSM4. From Danabasoglu et al. (2012).

Figure 7 shows the maximum North Atlantic MOC values from the pre-industrial control runs of the climate models CCSM3 and CCSM4 (from the box shown on Figure 6). It shows that the maximum value takes on the order of 400-500 years to adjust to the model climatology from the initial condition. This is one reason why the initial conditions for 20th century runs described earlier should be taken after 500 years or longer of the control runs. Figure 7 also shows a large difference in the internal variability of the North Atlantic MOC between the two versions of the model. CCSM3 has a fairly regular, large amplitude oscillation with a period of about 20 years, although its amplitude decreases after 500 years, whereas CCSM4 shows much smaller variability with no regular oscillation. This change is mostly caused by the implementation of a new overflow parameterization in CCSM4, which is described in Danabasoglu et al. (2010). This parameterization improves the representation of the Denmark Strait and Greenland/Iceland/Scotland overflows, which results in North Atlantic Deep Water reaching down to 4km near 60°N, and is therefore more realistic. This highlights the fact that the representation of the North Atlantic MOC in the ocean components of climate models is rather sensitive to several of the parameterizations used. This

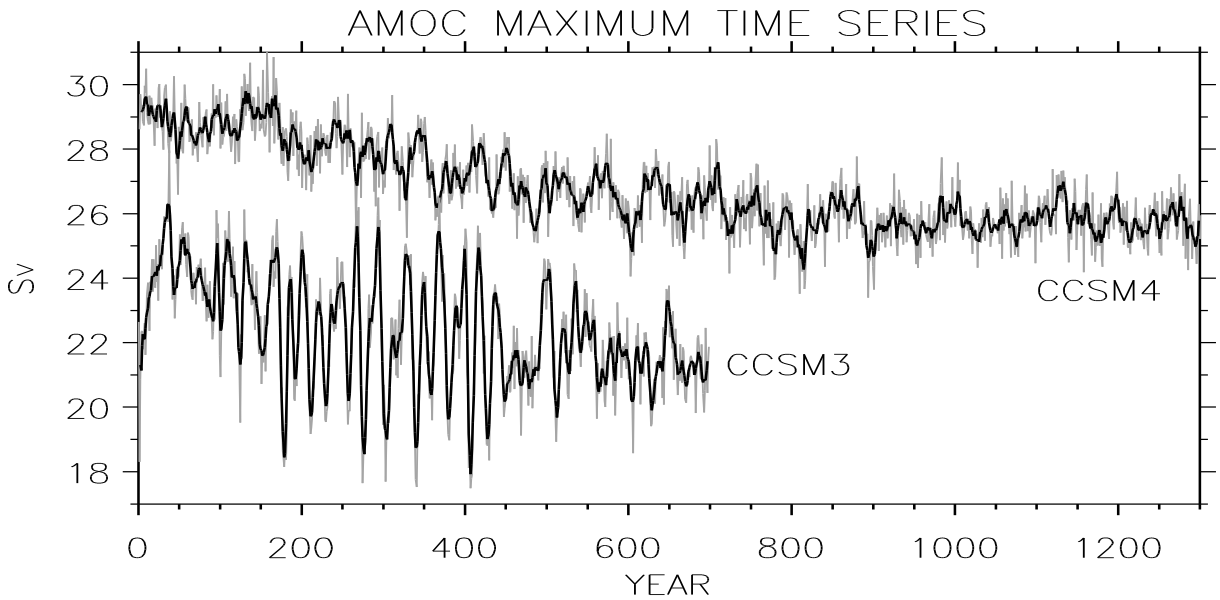

Figure 7. Time series of the maximum value in Sv of the North Atlantic meridional overturning circulation from the pre industrial control runs of the CCSM3 and CCSM4. From Danabasoglu et al. (2012).

results in both the mean maximum value and the internal variability being quite different across all the climate models used to make future climate projections. Much more discussion of the North Atlantic MOC is in Chapter 5.6.

It has long been thought that future climate change could reduce the magnitude of the North Atlantic MOC, (Manabe et al., 1991). Both heating of the upper ocean and an increase in surface fresh water from melting sea ice and land ice, especially around Greenland, would reduce the density of the near surface water. This would increase the stability of the upper ocean compared to the deeper ocean, and hence has the potential to reduce the rate of deep water formation in the high latitude North Atlantic. There has also been much speculation that, if the MOC does reduce significantly, then Europe's climate could become colder in the future, rather than becoming warmer. The IPCC 4th Assessment Report covered the most recent work on this subject in Chapter 10.3.4 of Solomon et al. (2007). The results show that the MOC does get weaker in the future in all 19 climate models assessed, and the rate of weakening increases as the future rate of CO_2 rise increases in the forcing scenarios used. In addition, there is a range of weakening rates in the MOC across climate models that are forced with the same scenario CO_2 increase. The range of MOC decreases are bounded by no change at all in the commitment run where the CO_2 remains constant at the present day value, and a decrease of about 50% in the most sensitive models using a scenario where the CO_2 concentration has doubled from its 2000 value by the end of the 21st century. Some of these future projections have been run beyond 2100, where the CO_2 concentration is then kept constant at the 2100 value. In all 'modern' models, and for all the scenarios used for CO_2 concentration at 2100, the MOC begins to strengthen once the CO_2 concentration is kept constant (Meehl et al. 2006, Solomon et al. 2007). Again, the rate of recovery depends on the particular model and the future scenarios used. 'Modern' here means full climate models using a horizontal resolution of 1^o or finer, no flux adjustments, and modern physics in the ocean component including the much more physically realistic diffusion of heat and salt along sloping isopycnal surfaces rather than along horizontal surfaces. In addition, for all climate models and all future CO_2 scenarios used in the IPCC 4th Assessment Report, the radiative warming effect of the increased greenhouse gases over Europe is larger than the cooling effect of a reduction in the MOC. So, current climate models do not support a future cooling of Europe's climate, (Solomon et al. 2007).

It has frequently been suggested that the North Atlantic MOC could 'collapse' as a result of future increases in CO_2 concentration, and then remain at a very small magnitude for a long time into the future. Manabe and Stouffer (1993) showed that when the CO_2 concentration was quadrupled after 140 years and then held constant, the North Atlantic MOC did become very small, and then stayed very small for the duration of the 500 year integration. However, a later paper (Stouffer and Manabe 2003) shows that the MOC did recover after about 1500 years, and then stayed near to its initial value before CO_2 was increased for the remainder of the 5000 year run. The reason was that the warming near the surface diffused down slowly over the 1500 years, so that the upper 2-3 km of ocean became less stratified, and deep water formation started again. Thus, the MOC recovery time was set by the diffusive timescale for heat to reach the ocean mid-depths. When the same run up to 4xCO_2 was made with more 'modern' climate models, then the MOC did start to recover soon after the CO_2 was held constant in both the HadCM3 (Wood et al. 2003) and the CCSM3 (Bryan et al. 2006b). Therefore, I believe that the Manabe and Stouffer (1993) result is influenced by the very coarse ocean resolution of about 4^o in the horizontal and 12 vertical levels, horizontal mixing of heat and salt, and the large flux adjustments of heat and fresh water. Mikolajewicz et al. (2007) also showed that the North Atlantic MOC can become small and stay small when the increase in CO_2 is a factor of 5, and sometimes when it is a factor of 3. However, the horizontal resolution of their model is very coarse at 5.6^o, and it has a very large flux adjustment of fresh water in the North Atlantic. The question then arises: Is there a final CO_2 concentration where the MOC becomes very small and subsequently does not recover in 'modern' climate models? Very recently, the Representative Concentration Pathway 8.5 (RCP8.5) has been used to force the CCSM4 between 2005 and 2300 from the end of a 20th century run from 1850 to 2005. The CO_2 starts at 285 ppm in 1850, is 385 ppm in 2000, and rises to near 1962 ppm by 2250 before leveling off. This run has already been discussed in Chapter 1.1. The globally-averaged surface temperature and the North Atlantic MOC index from Meehl et al. (2012) are shown in Figure 1.1.13. The MOC index starts at ~25 Sv in 1850, reduces to ~8 Sv in 2250, and remains near this value as the CO_2 forcing becomes almost constant. The average surface temperature in the

Arctic Ocean has risen by more than 20° C by 2300 in this RCP8.5 run, and there is no sea ice in the Arctic all year round. The deep water formation in the Labrador Sea, Greenland-Iceland-Norwegian Seas, and Arctic Ocean is completely shut off by 2200 because of the large rise in SSTs and rapid freshening in surface salinity caused by the sea ice melt. The maximum of the MOC streamfunction north of 40°N, shown in Figure 8, has fallen to ~3 Sv by 2200, so that the index of ~8 Sv shown in Figure 1.1.13 is associated with the subtropical gyre near 20°N at 500 m depth, rather than the MOC itself. This run has been continued out to 2500 with constant CO_2 forcing, and the maximum MOC value north of 40°N remains <3 Sv throughout to 2500. If this run were to be continued further, I think that the MOC would remain small for several hundred years, and a recovery would probably be on the long diffusive timescale of heat to reach the ocean mid-depths, as in the 5000 year run of Stouffer and Manabe (2003). Thus, the CCSM4 does show that its MOC can be switched off quasi-permanently when it is forced by a CO_2 rise of a factor of almost 7 between 1850 and 2250. I suspect that other 'modern' climate models display similar behavior when the forcing is so strong that all the Arctic sea ice melts, and deep water formation completely shuts off.

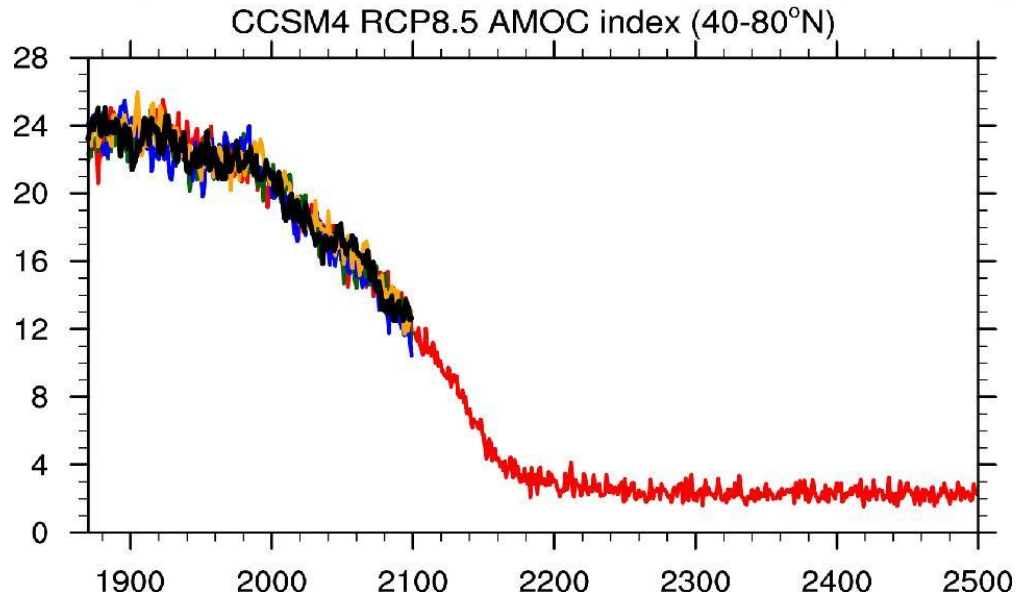

Figure 8. Maximum value of the North Atlantic MOC streamfunction north of 40°N in an ensemble of CCSM4 RCP8.5 runs, and in one extension from 2100 to 2500: The CO_2 value is constant after 2300. Courtesy of A. Hu.

Stommel (1961) showed that in a simple box model the MOC could possess multiple equilibria, which is two different stable solutions with the MOC either strong or weak, given identical forcing of the model. In addition, Marotzke and Willebrand (1991), Hughes and Weaver (1994) and many more recent studies have found multiple equilibria of the MOC in global ocean models that use very coarse resolution of about 4°, diffusion of heat and salt along horizontal surfaces, and are forced by mixed boundary conditions of restoring to an atmospheric temperature, but an imposed flux of fresh water. More recently, Marsh et al. (2004) and Sijp et al. (2012) have found MOC multiple equilibria in Intermediate Complexity coupled models that use a full ocean component with coarse resolution, but a considerably simplified atmosphere component. The last of these results is despite the fact that, using the same model, Sijp et al. (2006) find that the stability of North Atlantic deep water formation to imposed fresh water inputs is significantly increased when diffusion of heat and salt along horizontal surfaces is replaced by diffusion along sloping isopycnal surfaces. The reason is that vertical exchange can then be accomplished by the diffusion, whereas it has to be accomplished by convection when the diffusion is along horizontal surfaces. Manabe and Stouffer (1988, 1999) did find multiple equilibria of the North Atlantic MOC in a full climate model. However, I believe this result is a consequence of the very coarse ocean resolution, horizontal mixing of heat and salt, and the very large flux adjustment of fresh water. I know of no evidence of multiple equilibria of the North Atlantic MOC in

full, 'modern' climate models, so there are two possible explanations. Either multiple equilibria of the MOC exist in full climate models, but have not yet been found because of computational time constraints, or they do not exist when all the ocean-atmosphere feedbacks are working, which is not the case in Intermediate Complexity climate models. I favor the second explanation, but am willing to be proved wrong.

5.4.7. El Nino/Southern Oscillation

ENSO is the largest and best observed interannual signal in the Earth's climate system. However, it has proved rather difficult to simulate well in climate models. Most models have difficulty reproducing the mean precipitation pattern in the tropical Pacific Ocean, because they have too much rain south of the equator in the western ocean that reaches too far into the central Pacific. In addition, the frequency of model ENSO variability in several models is too short, and is not the broad-band maximum between 3 and 7 years seen in observations. Much work over recent years has resulted in good ENSO simulations in a small number of climate models, such as the Hadley Centre HadCM3 (Collins et al., 2001), the GFDL CM2.1 (Wittenberg, 2009), and the CCSM4 (Deser et al., 2012). These improvements have resulted mainly from refinements to the deep convection parameterizations in the atmosphere components of these models, and from increased resolution in both the atmosphere and ocean; Guilyardi et al. (2009). Figure 9 shows the correlation of monthly mean NINO3 SST anomalies with global SST anomalies from observations, and the CCSM versions 3 and 4. The improvements in the CCSM4 are the much wider region of positive correlation in the eastern Pacific Ocean that reaches into the subtropics, and the horseshoe pattern of

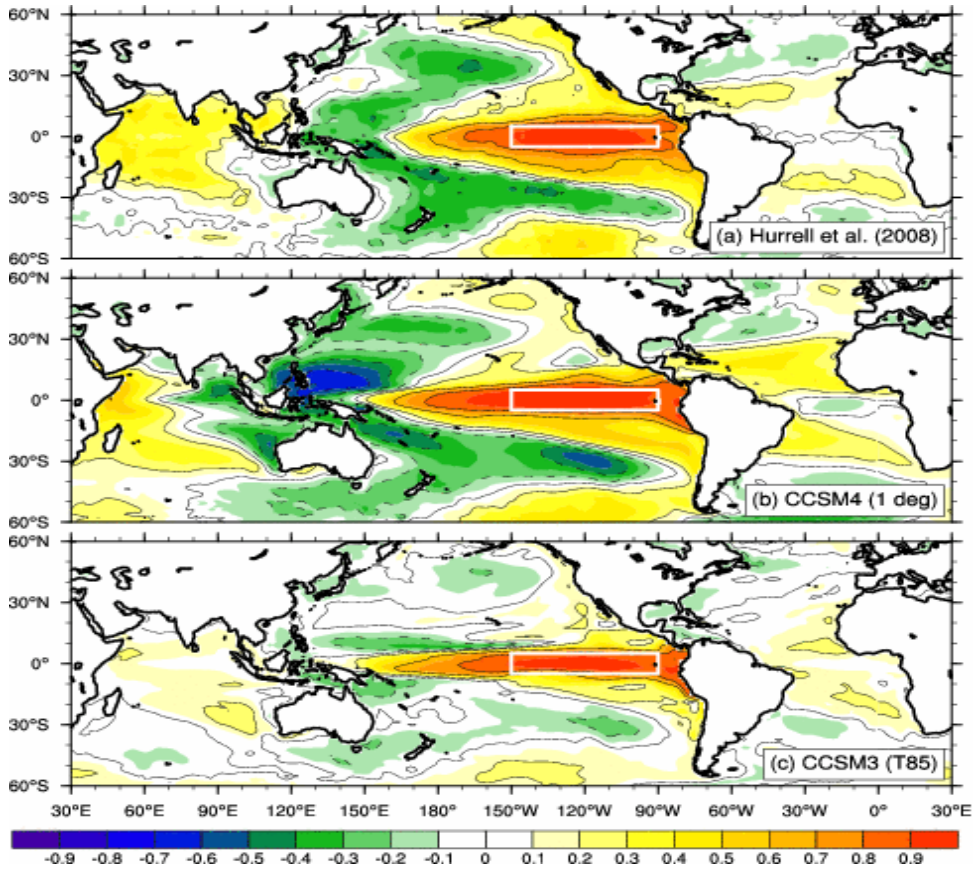

Figure 9. Correlation of monthly mean NINO3 (defined by the white box) SST anomalies with global SST anomalies from: a) Observations, b) the CCSM version 4, and c) the CCSM version 3. From Gent et al. (2011).

negative correlation in the west Pacific Ocean that extends into the central Pacific Ocean at mid-latitudes. Degradations are that the negative correlation is now much stronger than observations in the western Pacific Ocean, and it reaches into the east tropical Indian Ocean, probably because the ENSO amplitude is too large. These improvements result from changes to the atmosphere component deep convection scheme; see Neale et al., (2008).

The models that now simulate ENSO well show large very low-frequency variability in its amplitude on decadal and centennial timescales. Wittenberg (2009) shows the Nino3 SST anomaly from a 2000 year control run using 1860 conditions from the GFDL CM2.1 model, see Figure 1. The very low-frequency variability is large, and is reminiscent of the ENSO simulations from the earlier, much simpler coupled model of Zebiak & Cane (1987). A frequently asked question is how ENSO will change in the future. However, Wittenberg (2009) and Stevenson et al. (2012) show that the very low-frequency ENSO variability in the CM2.1 and CCSM4 control runs is so large, that runs of at least 500 years would be required in order to detect a statistically significant change. Therefore, these models suggest that ENSO variability in the 21st century will be large, but will not be statistically different than the intrinsic variability of ENSO if the climate were stationary and in equilibrium. Much more discussion of ENSO and its prediction is in Chapter 5.5.

5.4.8. Uses of Climate Models

Sections 5.4.3 through 5.4.5 describe the use of climate models to simulate a control run when the forcing is stationary, the earth's climate from 1850 to the present time, and to make future climate projections out to 2100 and beyond. In addition to the variables and phenomena described above, these runs make future projections of the sea level rise due to the warming of the oceans and the regional changes in sea level due to changes in ocean circulation. There are two other factors that change sea level; the first is the sea level rise due to melting of ice on land, including mountain glaciers and the Greenland and Antarctic ice sheets (see Chapter 6.1). There is strong evidence from GRACE satellite data of increasing ice melt off Greenland, Antarctica and other land glaciers over the last decade (Velicogna, 2009, Rignot et al., 2011), and during this time it is estimated that this fresh water source has contributed an equal amount to sea level rise as the thermosteric rise due to warming of the oceans. This contribution to sea level rise can be included in a climate model, if it has components that represent land glaciers and the two large ice caps. Several climate models are now in the process of including these two components. The final contributor to sea level changes is the height of the earth's crust, which is affected by melting of the Greenland ice sheet, changes in the Earth's geoid, and glacial isostatic adjustment (Douglas et al., 2001). It is very unlikely these effects will be included in climate models, but they can be estimated using glacial rebound models (Kopp et al., 2010).

If a climate model has an interactive carbon cycle, then it must have a biogeochemistry module in its ocean component. Several climate models now do have an interactive carbon cycle, which means they predict the concentration of CO_2 in the atmosphere given a scenario of emissions, rather than taking the atmospheric concentration as an input. In this case, the model will make future projections of the oceanic carbon cycle, which can be used as a projection for future ocean acidification, and for how this affects the future state of marine ecosystems (see Chapters 6.4 and 6.5).

Another very important use of the control and 20th century climate model runs is to explore the mechanisms and processes of climate variability on all time scales. For time scales from the diurnal, through the seasonal cycle and out to the several year time scale of ENSO, the model mechanisms can be compared to our knowledge from ocean observations. However, ocean variability on decadal and longer time scales is not well documented by observations, so our knowledge of the mechanisms driving ocean decadal variability comes mostly from models. Another use of climate models is to document how ocean variability, such as the North Atlantic MOC described in Section 5.4.6, affects the atmosphere and sea ice. In general, climate models are used to assess how processes in one component affect all the other components of the climate system, which cannot be determined by runs of the individual components alone.

Another use of climate models is to run simulations for many different paleo-climates of the past. These range from the last glacial maximum 21,000 years ago when the orbital parameters were different (e.g. Otto-Bliesner et al.,

2006), to deep time experiments many million years ago, when the locations of the continents were very different (e.g. Kiehl & Shields, 2005).

Over the past 10 years or more, climate models have been used to make ENSO and seasonal forecasts out to six months or a year. The difference between a forecast and a future projection is that the climate system needs to be initialized to the current climate state in order to make a forecast, whereas it is not initialized to observations when making a projection. Making a seasonal forecast is similar to making a weather forecast for the next few days, except that in the climate system on seasonal time scales it is just as important to initialize the upper ocean state as well as the atmospheric state. For an ENSO forecast, it is only important to initialize the upper part of the tropical Pacific Ocean (Rosati et al., 1997), and seasonal forecasts require the upper ocean to be initialized globally. More information on these ENSO and seasonal forecasts is in Chapter 5.5. A very recent development is the use of climate models to make decadal forecasts of the future climate. For a decadal forecast, the deeper parts of all the oceans need to be initialized, not just the upper ocean, and it may also be important to initialize the current state of sea ice in the Arctic and Antarctic, and the levels of soil moisture in the land component, although this is as yet unproven. Initializing the full depth ocean component is a major new challenge because it is a new aspect of ocean science, and much more information and details of decadal forecasts are in Chapter 5.6.

Obtaining an estimate of the current ocean state entails assimilating ocean data into a run of an ocean model that is driven by the best estimate of atmosphere forcing over the past few years. This has been done much more frequently with atmosphere models to produce 'reanalyses' of the past state of the atmosphere. It is now done routinely by several groups using ocean models, and some ocean 'reanalyses' have been produced recently (see Chapter 5.3). The accuracy of these 'reanalyses' has increased over the past few years (Carton & Santorelli, 2008), not only because of the experience obtained, but mostly because since about 2003 there are many more observations of temperature and salinity down to 2 km depth that have been obtained using ARGO floats (Roemmich et al., 2009). However, whether these 'reanalyses' are accurate enough to initialize the ocean component so that climate models can make useful decadal forecasts is a research question still to be explored in the future.

5.4.9. Limitations of Climate Models

The standard future projections of climate change made with the current models mostly use a horizontal resolution of around 1° or a little finer in all the components, and quite coarse resolution in the vertical. Almost certainly, the largest limitation of 1° resolution is that several very important processes must be parameterized because they cannot be resolved. Probably the best known of these processes is clouds in the atmosphere component. It has been known since the first climate models were assembled that the radiative properties, including the greenhouse effect, depend strongly on the various cloud parameterizations that have been used in different models (Cess et al., 1989). The various cloud schemes, and the parameterization of the interaction between aerosols and clouds, also dictate to a large degree the model equilibrium climate sensitivity, which is the globally averaged surface temperature increase in response to a doubling of the carbon dioxide concentration. Despite almost thirty years of research, the range of model equilibrium climate sensitivities remains about a factor of two.

In the ocean component, a resolution of 1° means that the viscosity required has to be much higher than desired or used in higher resolution models, so that the intrinsic variability is much smaller than in the real ocean. Also, a resolution of 1^{o} only begins to resolve the first Rossby radius of deformation near the equator, which is helped by the finer meridional resolution which is often used around the equator. Therefore, the effects of mesoscale eddies have to be parameterized. Most climate models use the Gent & McWilliams (1990) (GM) eddy parameterization, but there are significant differences in how it is implemented near the ocean surface, in the value of the coefficient chosen, and whether the coefficient is a constant or varies with position. It has been shown that a resolution of 1/10° is necessary to resolve the eddy effects so that a model's sea surface height variability matches that from satellite observations (Bryan et al., 2006a). Thus, it will still be over a decade before eddy-resolving ocean components will be used for standard climate projections. However, some climate model control runs with the ocean and sea ice components using 1/10° resolution have recently been completed, (McClean et al. 2011, Kirtman et al. 2012). In general, these high resolution integrations have a worse climate in many respects compared to the coarse resolution

runs because we have less experience in how to choose the best parameter values. Chapter 3.4 shows how these high resolution integrations can be used to evaluate the GM parameterization.

Another important limitation of climate models is that they lack components that may be important for simulating certain potentially large future changes in the climate system. Good examples have already been mentioned, such as an active carbon cycle and a component for the Greenland and Antarctic ice sheets. Without these components, a climate model cannot address the possibility that the ocean will take up a smaller fraction of the CO_2 output in the future, which will leave a larger fraction in the atmosphere, and the future sea level rise due to water melting from the ice sheets. In addition, there are several more possibilities of severe, possibly abrupt, climate changes with a very small probability, but with very large consequences. Examples are the release of large amounts of methane from ocean clathrates (Archer, 2007) and the possible fast breakup of the West Antarctic Ice Sheet (Bamber et al., 2009). These are difficult to simulate accurately and to assess quantitatively the possibilities that they will occur.

5.4.10. Cutting Edge Issues

The numerical discretization of the depth coordinate ocean components that are used in most climate models are quite old. They are based on latitude and longitude grids, although nearly all models now transpose the North Pole into a nearby land mass. There are also grids that have two poles in northern land masses, so that the resolution of the Arctic Ocean can be more comparable to the resolution in the rest of the global oceans. However, over the past ten years there has been much work on new grids that are much more uniform over the globe than latitude and longitude grids. These have mostly been developed with the atmosphere component in mind, and have been designed to work efficiently on the tens of thousands of processors that make up modern supercomputers (Staniforth & Thuburn, 2012). These grids can be made to vary in resolution from one part of the globe to another quite routinely. Therefore, the prospect is that grids with variable resolution will soon be available for global ocean components. Where to enhance the resolution and where to keep it coarse will then have to be decided, but enhanced resolution at the equator, along coasts and to resolve important narrow passages is now in prospect. The hope is that a variable grid would have many fewer grid points than a global 1/10° resolution grid, but would give comparable results.

Also over the past ten years, it has been well documented that depth coordinate models contain some cross-isopycnal mixing due to the numerics (Griffies et al., 2000, Ilicak et al., 2012). This is a problem if the numerical mixing is comparable to the very small level of cross-isopycnal mixing imposed in the deeper ocean. One way to avoid this is to use an isopycnal model where the vertical coordinate is potential density (Bleck, 2002, Hallberg, 2000). This type of model can be run stably with no cross-isopycnal mixing at all. Some climate models use an isopycnal ocean component (Furevik et al., 2003, Sun & Bleck, 2006, Dunne et al., 2012), but still the large majority of climate models use a depth coordinate ocean component. One reason is familiarity, but benefits of depth coordinates are the ease of simulating the upper mixed layer and that they keep uniform vertical resolution throughout the global ocean. Maintaining realistic deep overflows has also been a traditional problem using depth coordinates, but this can be improved by incorporating an explicit overflow parameterization (Danabasoglu et al., 2010). Isopycnal coordinate models represent deep overflows quite well, but their traditional drawbacks are the difficulty of representing the mixed layer, where vertical density gradients are small, and poor resolution in the high-latitude oceans, where the top-to-bottom density gradient is very small. I think that both types of models should continue to be developed, so that the positives and negatives of the two coordinate systems can be further compared and contrasted. A third vertical coordinate system used in coastal ocean models is sigma, or terrain-following, coordinates, which automatically have very thin grid cells in the shallow regions of the ocean. This is an advantage when studying coastal currents and processes, but is a real disadvantage in the climate context. The reason is that the allowable time step due to vertical advection in the shallow ocean is much smaller than that using the other vertical coordinates, so that the ocean model takes much more computer time to run.

As important as upgrading the numerical aspects of ocean climate components is further development of all the parameterizations used in the model. The vertical mixing scheme in the momentum and tracer equations not only sets the mixed layer and thermocline depths, but also dictates the amount of cross-isopycnal mixing in the deeper

ocean. It should include contributions from many sources such as internal wave breaking, tidal mixing, possibly mesoscale eddies and other processes (see Chapter 3.3). It also needs to work well across a range of vertical resolutions, because the vertical grid varies considerably with depth in most models. How to specify the coefficient in the GM eddy parameterization as a function of position is the subject of ongoing work (Eden et al., 2009), as well as how to transition the GM scheme into horizontal tracer mixing in the mixed layer (Danabasoglu et al., 2008, Ferrari et al. 2010), (see Chapter 3.4). The prospect of variable resolution grids discussed above leads to the requirement that ocean component parameterizations work well over a large range of scales. A good example is the GM parameterization for the effects of eddies on the mean flow. How should the coefficient be specified in a grid that varies between 1° and 1/10°? The same question needs to be asked of the horizontal viscosity scheme.

A final cutting edge issue is which processes and interactions are missing from current ocean components that could be important in future climate change? One example is the interaction of the ocean with ice shelves. This is very important because the breakup of small ice shelves has already been observed in Greenland and the Antarctic Peninsula. There is some potential for much larger Antarctic ice shelves to break up into the ocean. Current ocean components do not include the coastal and estuarine environments. These are areas of strong interactions between physical ocean properties and the carbon cycle. It is also where nutrients are injected from river outflows and most ocean biology occurs. These are potentially important aspects of the climate system, which are discussed in Chapters 6.3, 6.4 and 6.5.

Acknowledgments

I would like to thank my NCAR colleagues: Gokhan Danabasoglu for Figures 6 and 7, Marika Holland for Figure 3, Jerry Meehl for Figure 5, and Rich Neale for Figure 9. Aixue Hu ran the continuation of the CCSM4 RCP8.5 run out to 2500, and provided Figure 8. Thanks is also due to several colleagues at GFDL: Andrew Wittenberg for Figure 1, and Tom Knutson for Figures 2 and 4. Steve Griffies, Ron Stouffer, and two other reviewers provided very thorough reviews of this chapter, which contributed to a much improved final version. Thomas Stocker provided some comments, and his views on AMOC bifurcations and the possibility of multiple equilibria in full climate models.

References

Archer, D., (2007). Methane hydrate stability and anthropogenic climate change. *J. Geophys. Res.-Biogeo.*, 4, 521-544.

Bamber, G. L., R. E. Riva, B. L. Vermeersen, & A. M. LeBrocq, (2009). Reassessment of the potential sea-level rise from a collapse of the West Antarctic ice sheet. *Science*, 324, 901-903.

Bleck, R., (2002). An oceanic general circulation model framed in hybrid isopycnic-Cartesian coordinates. *Ocean Modelling*, 37, 55-88.

Boville, B. A. & P. R. Gent, (1998). The NCAR Climate System Model, version one. *J. Climate*, 11, 1115-1130.

Bryan, F.O., M.W. Hecht, & R.D. Smith, (2006a). Resolution convergence and sensitivity studies with North Atlantic circulation models. Part I: The western boundary current system. *Ocean Modelling*, 16, 141-159.

Bryan, F. O., G. Danabasoglu, N. Nakashiki, Y. Yoshida, D-H. Kim, J. Tsutsui, & S. C. Doney, (2006b). Response of the North Atlantic thermohaline circulation and ventilation to increasing carbon dioxide in CCSM3. *J. Climate*, 19, 2382-2397.

Bryan, K., S. Manabe, & R. C. Pacanowski, (1975). A global ocean-atmosphere climate model. Part II. The oceanic circulation. *J. Phys. Oceanogr.*, 5, 30-46.

Carton, J. A. & A. Santorelli, (2008). Global decadal upper-ocean heat content as viewed in nine analyses. *J. Climate*, 21, 6015-6035.

Cess, R. D., & Coauthors (1989). Interpretation of cloud-climate feedback as produced by 14 atmospheric general circulation models. *Science*, 245, 513-516.

Chassignet, E. P., L. T. Smith, R. Bleck, & F. O. Bryan, (1996). A model comparison: Numerical simulations of the north and equatorial Atlantic Oceanic circulation in depth and isopycnic coordinates. *J. Phys. Oceanogr.*, 26, 1849–1867.

Collins, M., S. F. Tett, & C. Cooper, (2001). The internal climate variability of HadCM3, a version of the Hadley Centre coupled model without flux adjustments. *Clim. Dyn.*, 17, 61-81.

Cunningham, S. A. & Coauthors, (2007). Temporal variability of the Atlantic meridional overturning circulation at 26.5^{o} N. *Science*, 317, 935-938.

Danabasoglu, G., R. Ferrari, & J. C. McWilliams, (2008). Sensitivity of an ocean general circulation model to a parameterization of near-surface eddy fluxes. *J. Climate*, 21, 1192-1208.

Danabasoglu, G. & P. R. Gent, (2009). Equilibrium climate sensitivity: Is it accurate to use a slab ocean model? *J. Climate*, 22, 2494-2499.

Danabasoglu, G., W. G. Large, & B. P. Briegleb, (2010). Climate impacts of parameterized Nordic Sea overflows. *J. Geophys. Res.* 115, C11005, doi:10.1029/2010JC006243.

Danabasoglu, G., S. G. Yeager, Y. O. Kwon, J. J. Tribbia, A. S. Phillips, & J. W. Hurrell, (2012). Variability of the Atlantic meridional overturning circulation in CCSM4. *J. Climate*, 25, 5153–5172.

Deser, C., A. S. Phillips, R. A. Tomas, Y. Okumura, M. A. Alexander, A. Capotondi, J. D. Scott, Y. O. Kwon, & M. Ohba, (2012). ENSO and Pacific decadal variability in Community Climate System Model version 4. *J. Climate*, 25, 2622-2651.

Douglas, B. C., M. S. Kearney, & S. P. Leatherman, (2001). Sea level rise: History and consequences. International Geophysics Series 75, Academic Press.

Dunne, J. P. & Coauthors, (2012). GFDL's ESM2 global coupled climate-carbon Earth system model. Part I: Physical formulation and baseline simulation characteristics. *J. Climate*, 25, 6646-6665.

Eden, C., M. Jochum, & G. Danabasoglu, (2009). Effect of different closures for thickness diffusivity. *Ocean Modelling*, 26, 47-59.

Ferrari, R., S. M. Griffies, A. J. Nurser, & G. K. Vallis, (2010). A boundary-value problem for the parameterized mesoscale eddy transport. *Ocean Modelling*, 32, 143-156.

Furevik, T., M. Bentsen, H. Drange, I. K. Kindem, N. G. Kvamstø, & A. Sorteberg, (2003). Description and evaluation of the Bergen climate model: ARPEGE coupled with MICOM. *Clim. Dyn.*, 21, 27-51.

Gent, P. R. & J. C. McWilliams, (1990). Isopycnal mixing in ocean circulation models. *J. Phys. Oceanogr.*, 20, 150-155.

Gent, P. R., (2011). The Gent-McWilliams parameterization: 20/20 hindsight. *Ocean Modelling*, 39, 2-9.

Gent, P. R. & Coauthors, (2011). The Community Climate Model version 4. *J. Climate*, 24, 4973-4991.

585 Gordon, C., C. Cooper, C. A. Senior, H. Banks, J. M. Gregory, T. C. Johns, J. F. Mitchell, & R. A. Wood, (2000).
586 The simulation of SST, sea ice extents and ocean heat transports in a version of the Hadley Centre coupled model
587 without flux adjustments. *Clim. Dyn.*, 16, 147-168.

588 Griffies, S. M., R. C. Pacanowski, & R. W. Hallberg, (2000). Spurious diapycnal mixing associated with advection
589 in a z-coordinate ocean model. *Mon. Wea. Rev.*, 128, 538–564.

590 Griffies, S. M. & Coauthors (2011). The GFDL CM3 coupled climate model: Characteristics of the ocean and sea
591 ice simulations. *J. Climate*, 24, 3520-3544.

592 Guilyardi, E. & Coauthors (2009). Understanding El Nino in ocean-atmosphere general circulation models:
593 Progress and challenges. *Bull. Amer. Met. Soc.*, 90, 325-340.

594 Hallberg, R. W., (2000). Time integration of diapycnal diffusion and Richardson number-dependent mixing in
595 isopycnal coordinate ocean models. *Mon. Wea. Rev.*, 128, 1402-1419.

596 Hirst, A. C., H. B. Gordon, & S. P. O'Farrell, (1996). Global warming in a coupled climate model including oceanic
597 eddy-induced advection. *Geophys. Res. Lett.*, 23, 3361-3364.

598 Hughes, T. M. C. & A. J. Weaver, (1994). Multiple equilibria of an asymmetric two-basin ocean model. *J. Phys.*
599 *Oceanogr.*, 24, 619-637.

600 Ilıcak, M., A. J. Adcroft, S. M. Griffies, & R. W. Hallberg. (2012). Spurious dianeutral mixing and the role of
601 momentum closure. *Ocean Modelling*, 45, 37-58.

602 Kiehl, J. T. & C. A. Shields (2005). Climate simulation of the latest Permian: Implications for mass extinction.
603 *Geology*, 33, 757-760.

604 Kirtman, B. P. & Coauthors, (2012). Impact of ocean model resolution on CCSM climate simulations. *Clim. Dyn.*,
605 39, 1303-1328.

606 Kopp, R. E., J.X. Mitrovica, S.M. Griffies, J. Yin, C.C. Hay, & R.J. Stouffer, 2010: The impact of Greenland melt
607 on local sea levels: a partially coupled analysis of dynamic and static equilibrium effects in idealized water-
608 hosing experiments. *Climatic Change*, 103, 619-625.

609 Knutson, T. R., & Coauthors, (2006). Assessment of twentieth-century regional surface temperature trends using
610 the GFDL CM2 coupled models. *J. Climate*, 19, 1624–1651.

611 Levitus, S., T. Boyer, M. Conkright, D. Johnson, T. O'Brien, J. Antonov, C. Stephens, & R. Gelfeld, (1998).
612 *Introduction*, Vol. 1, *World Ocean Database 1998*, NOAA Atlas NESDIS 18.

613 Manabe, S., K. Bryan, & M. J. Spelman, (1975). A global ocean-atmosphere climate model. Part I. The atmospheric
614 circulation. *J. Phys. Oceanogr.*, 5, 3-29.

615 Manabe, S. & R. J. Stouffer, (1988). Two stable equilibria of a coupled ocean-atmosphere model. *J. Climate*, 1,
616 841-866.

617 Manabe, S., R. J. Stouffer, M. J. Spelman & K. Bryan, (1991). Transient responses of a coupled ocean-atmosphere
618 model to gradual changes of atmospheric CO_2. Part I: Annual mean response. *J. Climate*, 4, 785-818.

619 Manabe, S. & R. Stouffer, (1993). Century-scale effects of increased atmospheric CO_2 on the ocean-atmosphere
620 system. *Nature*, 364, 215-218.

621 Manabe, S. & R. J. Stouffer, (1999). Are two modes of thermohaline circulation stable? *Tellus A*, 51, 400-411.

622 Marotzke, J. & J. Willebrand, (1991). Multiple equilibria of the global thermohaline circulation. *J. Phys.*
623 *Oceanogr.*, 21, 1372-1385.

Marsh, R. & E. J. Coauthors (2004). Bistability of the thermohaline circulation identified through comprehensive 2-parameter sweeps of an efficeient climate model. *Climate Dynamics*, 23, 761-777.

McClean, J. L. & Coauthors, (2011). A prototype two-decade fully-coupled fine-resolution CCSM simulation. *Ocean Modelling*, 39, 10-30.

Meehl, G. A. & Coauthors, (2006). Climate change projections for the twenty-first century and climate change commitment in the CCSM3. *J. Climate*, 19, 2597-2616.

Meehl, G. A. & Coauthors, (2012). Climate system response to external forcings and climate change projections in CCSM4. *J. Climate*, 25, 3661-3683.

Mikolajewicz, U., M. Groger, E. Maier-Reimer, G. Schurgers, M. Vizcaino & A. Winguth, (2007). Long-term effects of anthropogenic CO_2 emissions simulated with a complex earth system model. *Clim. Dyn.*, 28, 599-633.

Neale, R. B., J. H. Richter, & M. Jochum, (2008). The impact of convection on ENSO: From a delayed oscillator to a series of events. *J. Climate*, 21, 5904-5924.

Otto-Bliesner, B. L., E. C. Brady, G. Clauzet, R. Tomas, S. Levis, & Z. Kothavala, (2006). Last glacial maximum and holocene climate in CCSM3. *J. Climate*, 19, 2526-2544.

Randall, D. A., & Coauthors, (2007). Climate models and their evaluation. In *Climate Change 2007: The Physical Science Basis. Contribution of Working Group I to the fourth Assessment Report of the IPCC.* Cambridge University Press.

Rignot, E., I. Velicogna, M. R. Van den Broeke, A. Monaghan & J. Lenaerts, (2011). Acceleration of the contribution of the Greenland and Antarctic ice sheets to sea level rise. *Geophys. Res. Lett.*, 38, L05503, doi:10.1029/2011GL046583.

Roemmich, D., G. C. Johnson, S. Riser, R. Davis, J. Gilson, W. B. Owens, S. L. G. Owens, C. Schmid, & M. Ignaszewski, (2009). The Argo Program: Observing the global oceans with profiling floats. *Oceanography*, 22, 24-33.

Rosati, A., K. Miyakoda, & R. Gudgel, (1997). The impact of ocean initial conditions on ENSO forecasting with a coupled model. *Mon. Wea. Rev.*, 125, 754-772.

Sausen, R., R. K. Barthels, & K. Hasselmann, (1988). Coupled ocean-atmosphere models with flux correction. *Clim. Dyn.*, 2, 154-163.

Sijp, W. P., M. Bates, & M. H. England, (2006). Can isopycnal mixing control the stability of the thermohaline circulation in ocean climate models? *J. Climate*, 19, 5637-5651.

Sijp, W. P., M. H. England & J. M. Gregory, (2012). Precise calculations of the existence of multiple AMOC equilibria in coupled climate models. Part I: Equilibrium states. *J. Climate*, 25, 282-298.

Solomon, S., D. Qin, M. Manning, Z. Chen, M. Marquis, K.B. Averyt, M. Tignor & H.L. Miller, (2007). *Climate Change 2007: The Physical Science Basis. Contribution of Working Group I to the Fourth Assessment Report of the IPCC.* Cambridge University Press.

Staniforth, A., & J. Thuburn, (2012). Horizontal grids for global weather and climate prediction models: a review. *Q. J. R. Meteorol. Soc.*, 138, 1-26.

Steele, M., R. Morley, & W. Ermold, (2001). PHC: A global ocean hydrography with a high-quality Arctic Ocean. *J. Climate*, 14, 2079-2087.

Stevenson, S., B. Fox-Kemper, M. Jochum, R. Neale, C. Deser, & G. Meehl, (2012). Will there be a significant change to El Nino in the 21st century? *J. Climate*, 25, 2129-2145.

Stommel, H., (1961). Thermohaline convection with two stable regimes of flow. *Tellus*, 13, 224-230.

Stouffer, R. J. & S. Manabe, (2003). Equilibrium response of thermohaline circulation to large changes in atmospheric CO_2 concentration. *Clim. Dyn.*, 20, 759-773.

Stouffer, R. J., (2004). Time scales of climate response. *J. Climate*, 17, 209-217.

Sun, S., & R. Bleck, 2006: Multi-century simulations with the coupled GISS-HYCOM climate model: Control experiments. *Clim. Dyn.*, 26, 407-428.

Velicogna, I., (2009). Increasing rates of ice mass loss from the Greenland and Antarctic ice sheets revealed by GRACE. *Geophys. Res. Lett.*, 36, L19503, doi:10.1029/2009GL040222.

Wittenberg, A. T., (2009). Are historical records sufficient to constrain ENSO simulations? *Geophys. Res. Lett.*, 36, L12702. doi:10.1029/2009GL038710.

Wood, R. A. & F. O. Bryan, (2001). Coupled ocean-atmosphere models. In *Ocean Circulation and Climate*, G. Siedler, J. Church, J. Gould Eds., International Geophysics Series 77, Academic Press.

Wood, R. A., M. Vellinga, & R. Thorpe, (2003). Global warming and thermohaline circulation stability. *Phil. Trans. R. Soc. Lond. A*, 361, 1961-1975.

Wunsch, C. & P. Heimbach, (2006). Estimated decadal changes in the North Atlantic meridional overturning circulation and heat flux 1993-2004. *J. Phys. Oceanogr.*, 36, 2012–2024.

Zebiak, S. E. & M. A. Cane, (1987). A model El-Nino-Southern Oscillation. *Mon. Wea. Rev.*, 115, 2262-2278.

Abstract

Coupled climate models consist of atmosphere, ocean, land and sea ice components. Most climate models now do not need to use flux adjustments to maintain the present day climate in a control run, when the forcings have a repeating annual cycle or are constant in time. A control run must simulate well known important large-scale phenomena, such as the El Nino/Southern Oscillation and the North Atlantic overturning circulation. Climate models are used to simulate the climate of the 20th Century, and to make projections of the future climate. The uses and limitations of climate models are then described, and several cutting edge issues are discussed.

Keywords

Coupled Models

North Atlantic overturning circulation

El Nino/Southern Oscillation

Control Runs

20th Century Runs

Future Climate Projections

Limitations of Climate Models

Cutting Edge Issues