Lecture 1: Introduction and Overview

1.1 Scope of the course

This course, Climate Dynamics, concerns itself with understanding how Earth's climate operates. With "climate", we mean, very loosely, some time-average of all the physical properties of our environment, especially of atmosphere and oceans. For several reasons, we keep the definition of climate vague. First, a rigorous definition of the appropriate time-averaging is difficult; second, there is no widespread agreement on whether some components of the Earth system, particularly the oceans, are part of climate or "merely" influence climate. In principle, we use an inclusive perspective: If some natural process influences climate, it is part of climate. That said, we quickly back off and admit that this course can only cover a limited range of processes that are important for climate. From our own interests and expertise, a bias towards physical processes in atmosphere and, particularly, oceans arises naturally.

The second crucial characteristic of this course is that it focuses on dynamics, While we will make liberal use of observations to learn the fundamental empirical facts about Earth's climate, the emphasis is not on a comprehensive description. Rather, this course aims to convey a thorough understanding of the processes that shape climate. To this end, we will develop a host of simple, conceptual models that teach us how the most fundamental processes in climate work. Thus, this is a "theory" course, but we will not use the label "Theory of Climate" since that would suggest a more comprehensive understanding than actually exists.

This distinction brings us to the third crucial characteristic of this course. We will routinely touch on the limits of our (that is, the research community's) understanding of how climate works. This course contains some material based on very recent research results, possibly even unpublished ones. It has been inspired by, and has in turn inspired, our own research, and we expect this mutual fertilisation of research and teaching to continue.

1.2 Temperature distribution on Earth

Figure 1.1 (Fig. 7.4 from Peixoto and Oort 1992) shows a global map of the most elementary climate variable, surface air temperature for winter and summer, respectively. We see the familiar features, high temperatures at low latitudes and low temperatures at high latitudes, the out-of-phase seasonal change between Northern and Southern Hemispheres, and the smaller annual cycle over the ocean, relative to land (for example, the Atlantic is relatively warm in January and relatively cold in July.) Owing to the paucity of landmasses in the Southern Hemisphere (SH), there is generally much less zonal variation in the SH. The seasonal stratification of Fig. 1.1 somewhat masks another important but less appreciated property of surface air temperature, brought out more clearly by Fig. 1.2. It shows the annual-mean temperature (based on model-aided data analyses by the European Centre for Medium-Range Weather Forecasting, ECMWF). Temperatures over and east of the North Atlantic, north of 40˚N, are considerably higher than their counterparts in the Pacific. This is a crucial point to which we shall return repeatedly. Figure 1.2 also

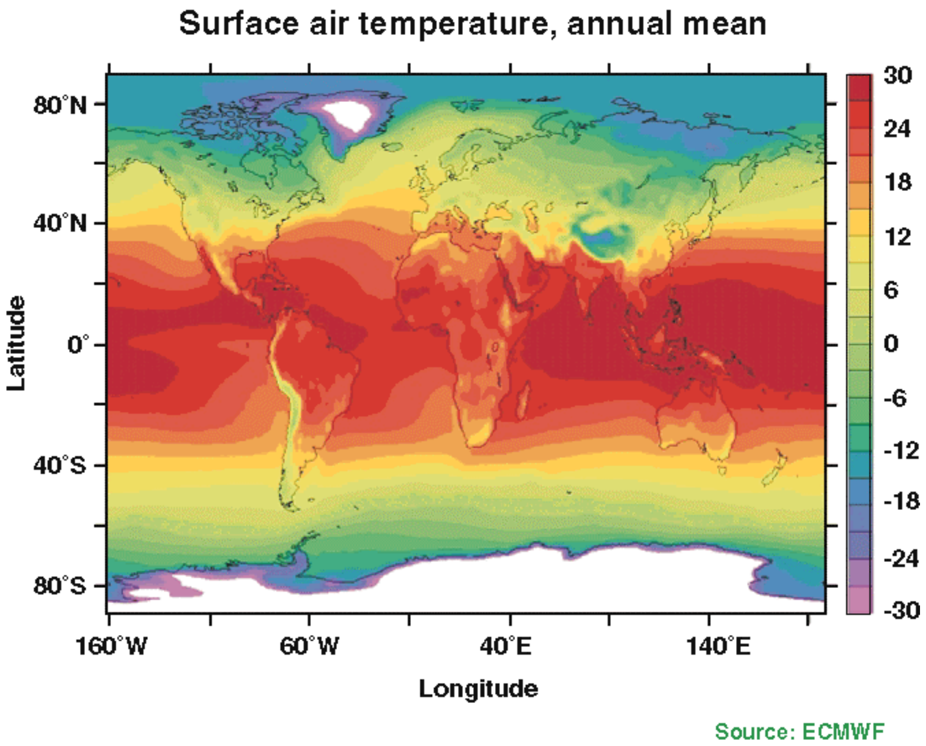

Figure 1.2: Annual-mean surface air temperature.

informs us that the annual-mean meridional temperature contrast, “pole-to-equator”, is of the order of 40˚C.

What determines the magnitude and distribution of this very basic climate variable, surface air temperature? The complete answer is immensely complex, as indicated by the sketch in Fig. 1.3a (Fig. 1.1 from Crowley and North 1991). It shows interactions between space, atmosphere, ocean, sea ice, land ice, land biota, and the solid Earth. Fortunately for us, a good guiding principle exists for selecting a manageable subset of components and exchanges. Figure 1.3b (Fig. 1.2 from Crowley and North 1991) shows the timescales on which, roughly, various processes operate. While leaving out all chemical exchanges, it correctly indicates that we can neglect solid-Earth processes if we restrict ourselves to timescales shorter than “orbital cycles” (referring to Earth’s orbit). The problem of the long-term evolution of climate, especially why apparently it has been moderate enough on Earth to permit the presence of liquid water for billions of years, is fascinating but beyond the scope of this course. The book by Kump et al. 1999 gives an immensely readable account of this. We will here concern ourselves with timescales of roughly years to tens of millennia; most notably, we will not consider in any great detail the annual cycle, nor tropical climate variability. The latter, dominated by the El Niño – Southern Oscillation phenomenon, is omitted essentially because of time constraints.

At a more manageable level, the surface air temperature is determined by the distribution of incoming solar radiation. Figure 1.4 (Fig. 1.2 from Trenberth et al. 1996, itself the Overview Chapter of the 1995 report for the Intergovernmental Panel on Climate Change, IPCC 1996) shows a sketch of the annual-mean conditions. Since the intensity of radiation received at the surface depends on the incidence angle, and the high-latitude surface experiences more oblique incidence (the sun appears lower in the sky), the low latitudes receive more solar radiation per area than is lost to space by longwave (infrared) radiation. The opposite is true for the high latitudes. To maintain steady state, energy (heat) must be transported by the fluid envelope of the Earth (ocean and atmosphere), from low to high latitudes.

In more detail than in Fig. 1.4, Fig. 1.5a (Fig. 1.1 from Gill 1982) shows that the difference between the radiation received at the top of the atmosphere, and that absorbed at the surface, is roughly independent of latitude. But the curve showing radiation emitted from the ground is flatter, as a function of latitude, leading to an energy surplus at low latitudes and a deficit at high latitudes; the difference must be balanced by energy export out of low latitudes.

After the introduction and a descriptive overview of atmospheric structure and transport processes, the first major concentration of this course is on “simple climate models”. These explore how relatively simple concepts involving radiative energy balance and meridional transports help to explain the gross meridional and vertical distributions of temperature in the atmosphere. Moreover, these models suggest that Earth’s climate might be perturbed such that its entire surface could freeze over – a scenario that might explain aspects of Earth’s distant past (“Snowball Earth”, Harland 1964, Kirschvink 1992, Hoffman et al. 1998).

1.3 Ocean circulation and climate

We will then shift our attention to the ocean. Figure 1.5b (from VonderHaar and Oort 1973) shows how the meridional heat transport is distributed between atmosphere and ocean. At low latitudes, the ocean dominates, at high latitudes, the atmosphere. But both media have a sizeable influence at almost all latitudes. The numbers shown in Fig. 1.5b are still quite uncertain, but the qualitative picture has been confirmed repeatedly. This is remarkable since prior to 1973, it was widely believed (and sometimes “proved”) that the ocean could not contribute significantly to energy transport in the climate system.

Figure 1.6 (Figs. 13.17 and 13.18 from Peixoto and Oort 1992) first confirms that ocean and atmosphere also have similar energy transports in the Southern Hemisphere (SH). More interestingly still, it shows that the three major oceans have vastly different heat transports. The Pacific, similarly to the global ocean, is nearly antisymmetric about the equator. In contrast, the Indian Ocean has purely southward heat transport, essentially lacking the high northern latitudes to which to export the energy received in the tropics. The most remarkable heat transport occurs in the Atlantic: It is northward everywhere; in particular, it is “up-gradient”, from cold to warm, in the South Atlantic. Not apparent from the figure but firmly established now, is that the North Atlantic heat transport is also more vigorous than that in the North Pacific, although the Atlantic is much narrower. As a consequence, sea surface temperature (SST) is higher by 3˚C to 5˚C in the North Atlantic, compared to the North Pacific (Fig. 1.7, from Levitus 1982). This, in turn, leads to the higher surface air temperatures shown in Fig. 1.2.

Why does the Atlantic exhibit such an anomalous heat transport? This question will be with us for much of the course. A proximate cause lies in the pattern of the global oceanic "thermohaline circulation" (THC, driven by heat and water exchanges with the atmosphere). Figure 1.8 (after Ganachaud and Wunsch 2000) gives a summary picture of the zonally integrated, global ocean circulation, based on modern hydrographic measurements. Its salient features are the large transport around Antarctica, strong transport around Australia, and a vigorous overturning circulation in the Atlantic, with sinking in the north, northward near-surface transport everywhere, and southward deep transport. No such pattern exists in the Pacific; in particular, there is no deep sinking in the North Pacific.

In our quest for ultimate causes, we are led to ask: "Why is no deep water formed in the North Pacific?" (Warren 1983). Again we will be making use of very simple models, such as the ocean "box models" depicted in Fig. 1.9 (from Marotzke 2000, and based in turn on Stommel 1961 and Rooth 1982). We will see that under identical assumed atmospheric forcing, the THC can exhibit multiple stable states, among them ones that are asymmetric with respect to the equator. This suggests that the Atlantic's anomalous behaviour has its ultimate cause in the ocean dynamics themselves and perhaps some historical "accident", picking one of several possible modes. As a run-up to these simple models, we will first cover the consequences of a different aspect of the ocean's role in climate, that of its large heat capacity. Then, we will develop the concepts necessary to understand why the box models are sensible representations of the ocean's thermohaline circulation (THC).

The purely oceanic box models cannot represent how the THC influences atmospheric properties or, more generally, they do not reflect that the THC is part of a coupled ocean-atmosphere phenomenon. To obtain insight into the large-scale air-sea interaction, simple coupled models must be developed, such as shown in Fig. 1.10 (from Marotzke 1996). We will go through the construction and steady-state solutions of this model, and then use it as a prototype for a general discussion of analysing feedbacks. This unearths some interesting fundamental points, as reflected by the two, conflicting, statements shown in Fig. 1.11, made by two distinguished climate scientists.

i) Coupling two individually stable systems leads to increased instability because the coupled system has greater freedom.

ii) A complex system tends to be more stable than a simple one.

Fig. 1.12: Which, if any, is it to be? Click here to find out more, and to run a simple model investigating this.

Here, we will focus on how various aspects of ocean-atmosphere coupling introduce a suite of feedbacks affecting the THC, in particular how air-sea interaction influences the stability of the THC (Fig. 1.12).

1.4 Biogeochemistry and Ice Ages

The anthropogenic increase in atmospheric CO_2 content drives much of the current interest in climate change. This issue is epitomised by Fig. 1.13b (Fig. 3.2 from Houghton 1997), the “Keeling curve”, named after Charles Keeling, who started the measurement series on Mauna Loa. Overlain on a seasonal cycle of roughly 5 parts per million volume (ppmv) is a steady upward trend, which is even more striking when compared to (indirect) measurements of pre-1958 atmospheric CO_2 concentration. Figure 1.14 (Fig. 3.1 from Houghton 1997) gives a first idea of the complexity of the question of where the anthropogenic CO_2 goes. Human emissions are a relatively small part in a system of carbon fluxes characterised by huge and nearly completely compensating exchanges. While a comprehensive treatment is beyond the scope of this course, we will touch on the most crucial aspects of ocean carbon chemistry and its influence on atmospheric CO_2 and climate. But the carbon cycle is important not only for future climate evolution. Figure 1.15 (Fig. 3 from Petit et al. 1999) shows that over the last four major glacial cycles, temperature over Antarctica and CO_2 content varied synchronously, within observational uncertainties. The question naturally arises to what extent glacial-interglacial CO_2 variations played an active role in these massive climate shifts.

During glacial times, climate exhibited extraordinarily large swings over very short times, possibly only a few decades. Figure 1.16 (Fig. 1 from Dansgaard et al. 1993) shows results from an ice core from Greenland, with large, abrupt warming events (since dubbed Dansgaard-Öschger events) and the most famous abrupt climate event, the Younger Dryas cooling at the end of the last Ice Age (marked "a" in Fig. 1.16). These climate events were not merely a regional phenomenon but at least hemispheric in extent. Figure 1.17 (Fig. 2 from Schulz et al. 1998) demonstrates that the Arabian Sea exhibited climate swings in lockstep with Greenland (shown is a measure of biological productivity, assumed to indicated the strength of the monsoons).

What causes these swings? We do not know for sure, but the Atlantic THC is strongly implicated. A case can be made that the North Atlantic is the "centre of action", with the strongest signals, so it makes sense to search there for the underlying cause. For example, the estimate of sea surface temperature at the height of the last glacial maximum (Fig. 1.18; CLIMAP Project Members 1981) shows the greatest differences in the northwestern North Atlantic, suggesting that North Atlantic and North Pacific were more similar then in their THC patterns. The question is, then, whether the THC can undergo abrupt changes such that climate changes rapidly, at least on a regional scale. To investigate this question, considerably more complex models must be employed.

1.5 THC studies using general circulation models

Figure 1.19, after Marotzke and Willebrand 1991, shows the four qualitatively different steady states found in an idealised general circulation model (GCM) of the global THC: Strong THC in the North Atlantic but not in the North Pacific, the reverse case, both oceans with strong sinking in the north, or both oceans with strong sinking in the south. Multiple THC equilibria were also found in a coupled climate model (Fig. 1.20, from Manabe and Stouffer 1988); with drastic consequences for air temperatures over and near the North Atlantic (Fig. 1.21). We will explain what these figures are based on, and how these "numerical experiments" were conducted.

We will finish off by addressing how climate might change in the future – more precisely, what the likely consequences will be of the anthropogenic CO_2 increase. Apart from the expected impacts (general warming, changes in rainfall, soil moisture, etc.), a particularly disturbing aspect is shown in Fig. 1.22 from the third IPCC assessment (after Cubasch et al. 2001). The figure shows the changes in strength of the Atlantic THC as simulated by about a dozen comprehensive climate models. Most of the simulations begin by the middle of the 19th century.

Almost all models, with two notable exceptions, "predict" that by the end of the 20th century, the Atlantic THC begins to decrease considerably. The question

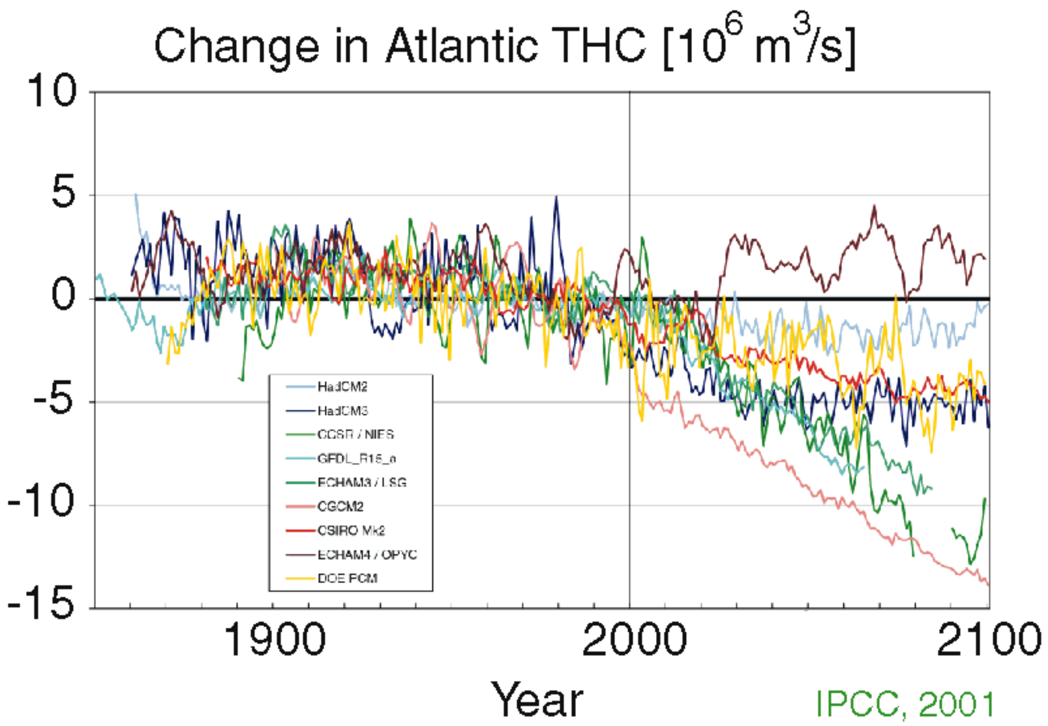

Fig. 1.22 (after Cubasch et al. 2001): Simulations of the Atlantic THC strength under CO_2 increase, relative to the means of the years 1961-1990.

arises what this means for European (and perhaps global?) climate, and whether it indicates an irreversible transition to an ocean circulation without deep sinking in the North Atlantic. We do not know the answers to these questions. We do not even know whether the Atlantic THC has weakened over the last 10 or so years, largely because the observational capabilities have not been in place. But the UK Natural Environment Research Council (NERC) has made a major commitment to fund "Rapid Climate Change and the THC" as a priority area, and the chances are good that a continuous system observing the THC strength will be put in place (further information).

1.6 Broader issues

Apart from giving a fairly rigorous introduction into Climate Dynamics, especially its oceanic aspects, this course pursues some more general goals. One of them, that of leading students up to the cutting edge of research, has been mentioned already. One consequence is that, more so than in "standard" undergraduate courses, the instructors cannot be expected to have all the answers. This is so, in part, because the cutting "edge" of research is really a topologically complex landscape, and no single individual (or any two individuals) can have a truly comprehensive overview of recent research in an area as broad as climate. More fundamentally, even where we do claim expert knowledge, some answers simply aren't there. We hope that the majority of students find this open-endedness stimulating rather than disconcerting.

A second goal is that the students who took this course should be able to deal with climate debates as handled in the popular press. Figure. 1.23 shows how the possible weakening of the Atlantic THC was dealt with in the US magazine "Atlantic Monthly" in January 1998 (Calvin 1998). The article is very interesting and contains much that is well researched and explained, but it also contains some serious errors and misconceptions. The students of this course should be able to spot and, ideally, refute them. This is all the more true for the second example (Fig. 1.24, Johnson 1997). On the title page of EOS, the weekly newsletter for the members of the American Geophysical Union, it was demanded that a dam be built across the Strait of

Gibraltar (largely blocking off the entire Mediterranean), lest the next Ice Age was surely to come.

The third broader goal of this course concerns the conceptual models that figure so prominently here. The standard way of assessing the quality of theories or models is to compare them to observational data. Agreement with data constitutes strong support for a theory. But the simple models we use here are often woefully inadequate to stand the test against observations; it is trivial to “falsify” them. What, then, is their value, and when is a simple model “good”? We do not have a general answer to this epistemologically very deep question. But we hope that through practice (here: revisiting the same phenomenon both with simple and complex models) we can instil into our students a sense that, when the explanations appear robust, use of simple models is indeed a valuable aid towards understanding (see Figs. 1.25 and 1.26), and Oreskes et al. 1994).

References

Calvin, W. H., 1998: The great climate flip-flop. *The American Monthly*, **281,** 47-64.

CLIMAP Project Members, 1981: Seasonal reconstruction of the earth's surface at the last glacial maximum. *Geological Society of America Map Chart Series*, **MC-36**.

Crowley, T. J., and G. R. North, 1991: *Paleoclimatology*. Oxford University Press, 339 pp.

Cubasch, U., G. A. Meehl, G. J. Boer, R. J. Stouffer, M. Dix, A. Noda, C. A. Senior, S. Raper, and K. S. Yap, 2001: Projections of future climate change. *Climate Change 2001: The Scientific Basis. Contribution of Working Group I to the Third Assessment Report of the Intergovernmental Panel on Climate Change*, J. T. Houghton, Y. Ding, D. J. Griggs, M. Noguer, P. J. v. d. Linden, X. Dai, K. Maskell, and C. A. Johnson, Eds., Cambridge University Press, 525-582.

Dansgaard, W., S. J. Johnsen, H. B. Clausen, D. Dahl-Jensen, N. S. Gundestrup, C. U. Hammer, C. S. Hvidberg, J. P. Steffensen, A. E. Sveinbjörnsdottir, J. Jouzel, and G. Bond, 1993: Evidence for general instability of past climate from a 250-kyr ice-core record. *Nature*, **364,** 218-220.

Ganachaud, A., and C. Wunsch, 2000: Improved estimates of global ocean circulation, heat transport and mixing from hydrographic data. *Nature*, **408,** 453-457.

Gill, A. E., 1982: *Atmosphere-Ocean Dynamics*. Academic Press, 662 pp.

Harland, W. B., 1964: Critical evidence for a great infra-Cambrian glaciation. *Geologische Rundschau*, **54,** 45-61.

Hoffman, P. F., A. J. Kaufman, G. P. Halverson, and D. P. Schrag, 1998: A Neoproterozoic snowball Earth. *Science*, **281,** 1342-1346.

Houghton, J. T., 1997: *Global Warming: the complete briefing (2nd ed.)*. Cambridge University Press,, 251 pp.

IPCC, 1996*: Climate Change 1995: The Science of Climate Change. Contribution of Working Group I to the Second Assessment Report of the Intergovernmental Panel on Climate Change*, J. T. Houghton, L. G. Meira Filho, B. A. Callander, N. Harris, A. Kattenberg, and K. Maskell, Eds., Cambridge University Press, 572 pp.

Johnson, R. G., 1997: Climate control requires a dam at the Strait of Gibraltar. *Eos Transactions AGU*, **78,** 277, 280-281.

Kirschvink, J. L., 1992: Late Proterozoic low-latitude global glaciation: the Snowball Earth. *The Proterozoic Biosphere*, J. W. Schopf, and C. Klein, Eds., Cambridge University Press, 51-52.

Kump, L. R., J. F. Kasting, and R. G. Crane, 1999: *The Earth System*. Prentice Hall, 351 pp.

Levitus, S., 1982: *Climatological Atlas of the World Ocean. NOAA Prof. Paper No. 13*, U.S. Government Printing Office, 173 pp.

Manabe, S., and R. J. Stouffer, 1988: Two stable equilibria of a coupled ocean atmosphere model. *J. Clim.*, **1,** 841-866.

Marotzke, J., 1996: Analysis of thermohaline feedbacks. *Decadal Climate Variability: Dynamics and Predictability*, D. L. T. Anderson, and J. Willebrand, Eds., Springer-Verlag, 333-378.

Marotzke, J., 2000: Abrupt climate change and thermohaline circulation: Mechanisms and predictability. *Proceedings National Academy of Sciences (USA)*, **97,** 1347-1350.

Marotzke, J., and J. Willebrand, 1991: Multiple equilibria of the global thermohaline circulation. *J. Phys. Oceanogr.*, **21,** 1372-1385.

Oreskes, N., K. Shrader-Frechette, and K. Belitz, 1994: Verification, validation, and confirmation of numerical models in the Earth sciences. *Science*, **263,** 641-646.

Peixoto, J. P., and A. H. Oort, 1992: *Physics of Climate*. American Institute of Physics, 520 pp.

Petit, J. R., J. Jouzel, D. Raynaud, N. I. Barkov, J.-M. Barnola, I. Basile, M. Bender, J. Chappellaz, M. Davis, G. Delaygue, M. Delmotte, V. M. Kotlyakov, M. Legrand, V. Y. Lipenkov, C. Lorius, L. Pepin, C. Ritz, E. Saltzman, and M. Stievenard, 1999: Climate and atmospheric history of the past 420,000 years from the Vostok ice core, Antarctica. *Nature*, **399,** 429-436.

Rooth, C., 1982: Hydrology and ocean circulation. *Prog. Oceanogr.*, **11,** 131-149.

Schulz, H., U. v. Rad, and H. Erlenkeuser, 1998: Correlation between Arabian Sea and Greenland climate oscillations of the past 110,000 years. *Nature*, **393,** 54-57.

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

Trenberth, K. E., J. T. Houghton, and L. G. Meira Filho, 1996: The Climate System: an overview. *Climate Change 1995: The Science of Climate Change. Contribution of Working Group I to the Second Assessment Report of the Intergovernmental Panel on Climate Change*, J. T. Houghton, L. G. Meira Filho, B. A. Callander, N. Harris, A. Kattenberg, and K. Maskell, Eds., Cambridge University Press, 51-64.

VonderHaar, T. H., and A. H. Oort, 1973: New estimate of annual poleward energy transport by Northern Hemisphere oceans. *J. Phys. Oceanogr.*, **2,** 169-172.

Warren, B. A., 1983: Why is no deep water formed in the North Pacific? *Journal of Marine Research*, **41,** 327-347.