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Committed warming and its implications for climate change

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Abstract. Time lags between changes in radiative forcing and the resulting simulated climate responses are investigated in a set of transient climate change experiments. Both surface air temperature (SAT) and soil moisture responses are examined. Results suggest that if the radiative forcing is held fixed at today's levels, the global mean SAT will rise an additional 1.0K before equilibrating. This unrealized warming commitment is larger than the 0.6K warming observed since 1900. The coupled atmosphere-ocean GCM's transient SAT response for the year 2000 is estimated to be similar to its equilibration response to 1980 radiative forcings - a lag of ~20 years. Both the time lag and the warming commitment are projected to increase in the future, and depend on the model's climate sensitivity, oceanic heat uptake, and the forcing scenario. These results imply that much of the warming due to current greenhouse gas levels is yet to be realized.

Introduction

The earth's climate system is not currently equilibrated with the present day radiative forcing, nor is it likely that complete equilibration of all parts of the climate system has occurred in the past. Given that the radiative forcing varies in time (due to natural and human-induced reasons) and that components of the climate system such as the deep ocean exhibit multi-century or longer response times, it seems inevitable that the climate system is continually going through some adjustment processes. Developing a better understanding of the magnitude of the disequilibria and the adjustment time scales involved is important for the study of both past and potential future climate changes.

Using a global model of the atmosphere with parameterized oceanic heat transports, *Hansen et al.* [1984] obtained a warming of 4.2K when the model was allowed to equilibrate to a doubling of atmospheric CO_2. Then, employing a 1-D radiative-convective-equilibrium model calibrated to yield the same "climate sensitivity" (*i.e.*, the equilibrium response of the global mean surface air temperature to doubled CO_2), they examined the disequilibrium that developed between rising greenhouse gas levels and the global mean surface air temperature (SAT). They estimated that the SAT would rise an additional 1.0K if greenhouse gas levels were to remain steady in time. Later studies using relatively simple climate models (*e.g.*, energy balance atmospheres coupled to diffusive ocean models) also found considerable lags in the climate response to radiative forcing changes [*Mintzer,* 1987; *Wigley and Raper*, 1993] - lags attributed to the large thermal inertia of the oceans.

The difference between the realized warming at a given time and the warming of climate that would occur if the climate had an infinitely long time to adjust to that radiative forcing (*i.e.*, the gap between the equilibrium and realized temperature change for a given forcing) is referred to here as the "warming commitment." Two main factors that determine the magnitude of the committed warming are the amount of oceanic heat uptake and the climate sensitivity. The efficiency with which the deep ocean mixes with the upper ocean affects oceanic heat uptake. In a coupled model having a dynamical ocean, ocean heat uptake is influenced by ocean circulation patterns, the static stability of the ocean, the ocean model's subgrid scale mixing parameterizations [*Weaver and Wiebe*, 1999], and other factors. A model's climate sensitivity depends upon the nature of the feedbacks present in the model, and upon the model's state before CO_2 doubling.

Here we seek to build upon the earlier work of Hansen, Wigley and co-workers by analyzing a coupled atmosphere-ocean general circulation model (AOGCM) and by comparing the AOGCM's results with those produced by a less costly model in which an atmospheric GCM is coupled to a non-dynamic mixed-layer ocean. The AOGCM used here can be considered a more complex and sophisticated representation of the earth's climate system than the models employed in the previously cited studies. We re-examine the present day warming commitment, its future changes and other committed climate responses. Use of the AOGCM allows the climatic commitment of a wide range of variables to be examined, providing potentially useful input to policy makers contemplating strategies to deal with future climate changes.

Model Description and Experimental Design

A coupled atmosphere-ocean general circulation model and an atmosphere-mixed layer ocean model (ML model) are both used in this study. The coupled AOGCM is constructed by combining an atmospheric general circulation model with a dynamical model of the world ocean [*Manabe et al.,* 1991]. The ML model combines the same atmospheric model with a simple model of the ocean's mixed layer whose thickness is taken to be 50 meters [*Manabe et al.,* 1991]. Both models include simple land surface and sea ice formulations.

The atmospheric component is a spectral model rhomboidally truncated at wave number 15 (R15) with a transform grid of 7.5^o longitude by 4.5^o latitude and nine vertical levels. The model incorporates global geography, the seasonal cycle of insolation (but no diurnal variation), and heat and water budgets at the continental surface. Land surface hydrologic processes are simulated using the so-called "bucket" method described in *Manabe* [1969].

The dynamical ocean component of the AOGCM is a variant of the Bryan/Cox ocean GCM [*Bryan and Lewis*, 1979]. Its horizontal resolution is 3.75^o longitude by 4.5^o latitude with 12 vertical levels. In this version, a component of the subgrid scale mixing of potential temperature and salinity is oriented along density surfaces [*Tziperman and Bryan*, 1993]. The ML model has a relatively simple thermodynamic sea ice model. The AOGCM uses the same ice thermodynamics while allowing the sea ice to be advected by the simulated surface ocean currents.

Because the ML model utilizes a simple thermally conducting oceanic layer, it requires markedly less computer time to integrate one model year than is the case for the AOGCM. The ML model

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also approaches equilibrium more quickly than the AOGCM because the ML model lacks the long response times associated with the deep ocean. The ML model assumes no changes occur in the oceanic heat transport as the climate is altered. On the other hand, the more costly AOGCM allows an evaluation of large-scale, time-dependent changes in the oceanic heat transports and heat storage, and the effect of those changes upon the climate.

Estimates of variations in past and future radiative forcing were applied to the two models. The combined forcing of greenhouse trace gases is prorated to an equivalent CO_2 concentration. Historical reconstructions of equivalent CO_2 levels are prescribed for years 1765 to 1990 [*Mitchell et al.*, 1995], followed by a 1% per year increase (compounded) for years 1990 to 2065. The direct radiative forcing of sulfate aerosols is simulated by increasing the surface reflectance of solar radiation according to the method of *Mitchell et al.* [1995]. Temporal and spatial variations in the sulfate aerosols follow the IS92a scenario of the IPCC-1992 report [*Leggett et al.*, 1992] and are applied as in *Haywood et al.* [1997].

The two coupled climate models are used to conduct three different types of climate change experiments: (a) A transient fully coupled AOGCM experiment ("TAOGCM") [*Haywood et al.*, 1997] in which time varying radiative forcings are imposed as described above; (b) Same transient forcings as (a) except that the ML model is used (referred to here as the "transient mixed-layer" or "TML" model experiments); (c) The same as (b) except that the ML model is integrated to equilibrium for six separate time periods with the total radiative forcing held fixed at values corresponding to the years 1980, 2000, 2020, 2040, 2050 and 2060 (referred to as the "equilibrium ML" or "EML" experiments). The responses to forcing changes are determined by comparison with control integrations (either AOGCM or ML) in which the equivalent CO_2 and sulfate aerosol levels are held fixed at 1765 levels.

Comparing results from these three types of experiments will provide a measure of the ocean's influence on the simulated climate change. Differences between the results obtained from the EML model experiments and those produced by the TAOGCM integration can be thought of as estimates of the committed climate response. The transient AOGCM and equilibrated ML model responses differ primarily because of the larger heat capacity of the AOGCM's global ocean. Differences between the TML and EML results represent the influence of only the top 50 m of the ocean's water mass. As will be shown later, the TML vs. EML differences are fairly minor, so that with a small error, the TML can also be thought of as a surrogate for an EML result at any point in time.

The experimental design assumes that the equilibrium response of the ML model is similar to that of the AOGCM. Historically, inexpensive ML models have been used to estimate the climate sensitivity of more costly AOGCMs (see for example, *Kattenberg et al.* [1996]). For a doubling of atmospheric CO_2, the climate sensitivity of the ML model is 3.7K and the climate sensitivity of the AOGCM is estimated to be 4.5K [*Stouffer and Manabe*, 1999].

That the two models' climate sensitivities are not identical is partially attributable to the fact that the AOGCM produces a cooler control (pre-CO_2 increase) state with more snow and sea ice, thereby allowing more positive ice and snow albedo feedbacks to occur as the climate warms [*Spelman and Manabe*, 1984]. The climate drift in the AOGCM contributes ~0.2K to the difference in the climate sensitivities of the EML and AOGCM (an EML integration using the SSTs and sea ice distribution from the end of the coupled integration has a 3.9K global SAT temperature increase for a doubling of CO_2).

Physical processes found only in the AOGCM (*e.g.*, advection of sea ice and variations of meridional ocean heat transport) also

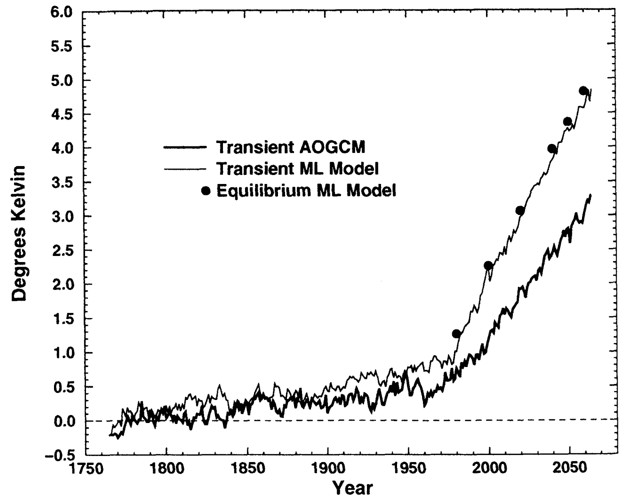

Figure 1. Model-simulated global mean surface air temperature anomalies (K). Thick and thin lines represent time series for the coupled TAOGCM and TML experiments respectively. Large dots indicate EML model results.

can contribute to feedback mechanisms that influence climate sensitivity. For example, experiments are underway to investigate the role that changes in the oceanic heat transport have in altering the climate sensitivity of the AOGCM vs. the ML model (a ~0.1 petawatt reduction in the AOGCM's North Atlantic's maximum poleward heat transport occurs when going from the control state to one with doubled CO_2).

Mindful of the prohibitive cost of performing numerous lengthy equilibrium experiments with the AOGCM, the differences between the models' climate sensitivities are considered small enough to allow the EML (and TML) results to be used as a proxy for the AOGCM's equilibrium response.

Simulated Climate Responses

The ML model experiments need to be integrated only 10-15 years with constant radiative forcing to reach a steady state. The small differences between the TML and EML results are evidence of this (Figure 1). A similar degree of equilibration takes thousands of years to achieve in the AOGCM [*Stouffer and Manabe*, 1999]. That the TAOGCM's projected surface air temperature considerably lags that simulated in the TML experiment (Figure 1) is a manifestation of the difference in the models' equilibration time scales. Note that the amount of warming seen at year 2000 in the TAOGCM simulation was realized about 20 years earlier in the TML experiment. Since the TAOGCM responds more slowly to the rapid increase in radiative forcing, the TML continues to warm more rapidly than the TAOGCM and this lag increases with time, approaching 35 years by 2060. Because the ML model equilibrates relatively quickly, the EML experiments' six SAT responses are only slightly greater than that of the TML simulation (Figure 1).

At year 2000, the TML model's SAT response is approximately 0.8K greater than that for the TAOGCM, whereas the response of the EML model is only about 0.2K greater than that for the TML model. So, for the year 2000, the total "committed warming" (the difference between the EML and TAOGCM results) is estimated to be ~1.0K. This is larger than the observed warming that has taken place since 1900 (approximately 0.6K, *Jones et al.* [1997]).

The time lags and committed warming disequilibria arise due to the thermal inertia of the AOGCM's ocean circulation model,

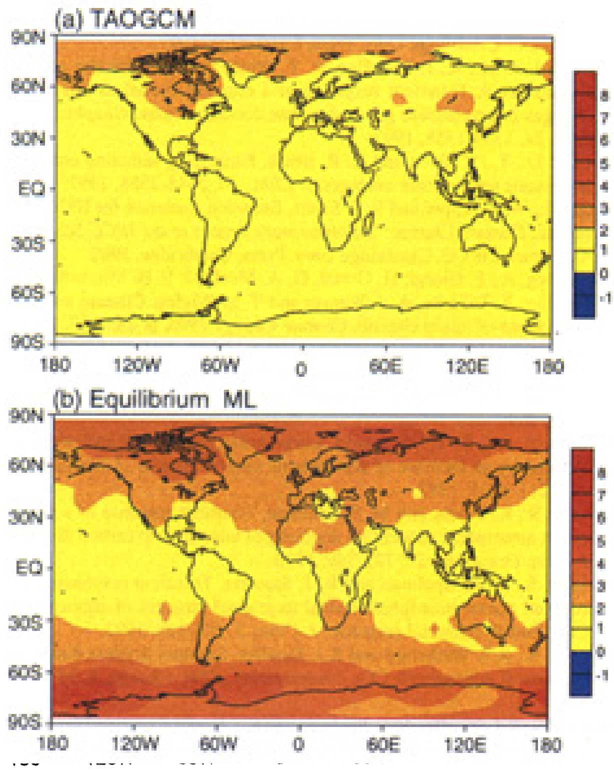

Plate 1. Geographical distribution of the annual mean surface air temperature response as simulated in (a) the TAOGCM experiment at year 2000, and (b) the ML model equilibrated to year 2000 radiative forcings..

which has a much larger heat capacity than a 50 meter oceanic mixed-layer. More specifically, the differences are caused primarily by the smaller response of the Southern Ocean and the extreme northern portion of the Atlantic Ocean in the TAOGCM experiment (Plate 1). This feature has been noted in earlier studies [*Manabe et al.*, 1990; *Manabe et al.*, 1991, *Kattenburg et al.*, 1996] and results from the large effective heat capacity due to the deep oceanic mixing simulated in these regions.

The global mean SAT responses of all of the numerical experiments are considerably greater at year 2060 than at year 2000. As the climate warms, the committed warming increases from about 1.0K at year 2000 to nearly 2.0K at year 2060. The realized warming increases from 0.6K at year 2000 to 3.0K at year 2060.

The preceding results also have important implications regarding land surface hydrology. In general, the patterns of soil moisture change for the TAOGCM and EML simulations are quite similar to one another (Plate 2). Both models show reductions of summertime soil moisture over central North America and southern Europe as well as an increase over India; features noted in earlier studies [*Manabe et al.*, 1981; *Manabe and Wetherald,* 1987; *Mitchell and Warrilow,* 1987; *Manabe et al.*, 1992]. However, changes in the TAOGCM summertime soil moisture are smaller than those simulated in the EML model. The EML model's SAT response is greater than the TAOGCM's (Plate 1), which contributes toward more evaporation over the EML model's land surface, thereby lowering the soil moisture values. This enhanced land evaporation is more than enough to offset the corresponding precipitation rate increase as the hydrologic cycle becomes more intense in the EML model than in the TAOGCM.

Discussion

Results of this study suggest that the climate system's transient SAT response lags the present day radiative forcing by approximately 20 years, leading to a present day warming commitment of about 1.0K. This result is quite similar to the estimate obtained by *Hansen et al.* [1984], who employed a less sophisticated model. *Hansen et al.* utilized a mixed-layer model with prescribed horizontal oceanic diffusion and an energy balance atmosphere model, whereas we use a coupled AOGCM and a ML model to estimate the AOGCM's equilibrium changes. That the results presented here are consistent with those obtained with simpler models in the past may be of interest to those involved with climate change science and policy. Our study also includes an evaluation of soil moisture changes.

The slow penetration of the heat anomaly into the ocean leads to the warming commitment obtained in the current study, and also implies a very large commitment to future sea level changes. *Stouffer and Manabe* [1999] found that only a small fraction of the equilibrium sea level rise was realized by the time of CO_2 doubling in their AOGCM experiments, even when the rate of CO_2 increase was very slow (280 years to CO_2 doubling). The time scale for global mean sea level to completely equilibrate with a constant radiative forcing is on the order of a few thousand years.

Studies performed with simple climate models [*Wigley and Raper*, 1993] indicate that lags in climatic responses depend on the efficiency of the diffusion of the heat perturbation into the ocean. In the coupled AOGCM used here, only part of the excess greenhouse heat stored in the ocean is transported from the surface to the interior of the ocean by motions explicitly resolved by the model grid. Some of the downward penetration of the greenhouse signal

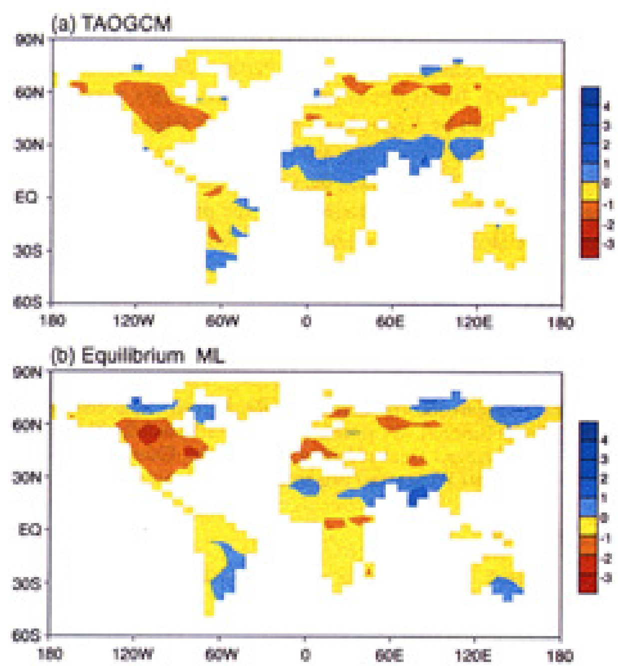

Plate 2. Geographical distribution of the difference in June-July-August soil moisture as simulated in (a) the TAOGCM experiment at year 2050 and (b) the ML model equilibrated to year 2050 radiative forcings. For clarity, the results in (a) are averaged over a 30-year period centered around year 2050 for an ensemble of 9 AOGCM integrations [*Dixon and Lazante,* 1999]. For (b), EML model averages are taken over 100 model years. Units are in cm.

results from parameterized subgrid scale oceanic processes and air-sea heat exchange which are not entirely understood or well quantified [*Weaver and Wiebe*, 1999; *Wiebe and Weaver*, 1999]. In a TAOGCM experiment, the characteristics of the ocean model's heat reservoir vary as the stability-dependent vertical mixing and three-dimensional circulation vary in time and space.

The magnitude of the warming commitment at any given time depends in part on the climate sensitivity of the model. Models with smaller climate sensitivities have tended to have smaller warming commitments [*Hansen et al.* 1984; *Wigley and Santer,* 1993; *Harvey et al.,* 1997]. The AOGCM used here has a climate sensitivity of ~4.5K, which lies in the upper part of the range of 1.5-4.5K estimated by the IPCC [*IPCC*, 1995]. We also note that the magnitude of the committed warming which occurs after the greenhouse gas concentrations are stabilized is a function of the growth rate in these concentrations [*Stouffer and Manabe*, 1999]. The faster the rate of increase, the larger the warming commitment.

There are large uncertainties in the magnitude of the radiative forcing of sulfate aerosols as well as other forcings which are not included in this study. However, we note that when the AOGCM is forced with both historical estimates of the sulfate aerosol and greenhouse gas forcings, the observed warming of the 20th century is simulated reasonably well [*Haywood et al.*, 1997]. Therefore, it is reasonable to speculate that the current study could provide a viable estimate of the present day magnitude of the committed global warming. Since our AOGCM's climate sensitivity is greater than that of the ML model used to estimate the equilibrium responses, the warming commitment magnitudes presented here may be an underestimate of the values that would have been realized if it was feasible to run the AOGCM to equilibrium.

Finally, it should be emphasized that the warming commitment discussion presented above describes the climate response for a constant radiative forcing of the earth. This requires greenhouse gas concentrations to remain constant over time. The protocol reached by the Third Conference of Parties which met in Kyoto, Japan in 1997, limits emissions to approximately present day levels. As shown by *Wigley et al.* [1997] and *Wigley* [1998], this level of emissions does not provide a stable greenhouse gas concentration by 2100. In fact, the CO_2 concentration continues to rise considerably beyond doubling of the pre-industrial values. Much more stringent controls on greenhouse gas emissions will be required to stabilize the greenhouse gas concentrations before 2100. Only after this stabilization occurs will the climate begin to come into equilibrium with that forcing, and only then will the total warming and associated climate changes be realized.

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References

Bryan, K., and L. J. Lewis, A water mass model of the world ocean, *J. Geophys. Res.*, *84*, 2503-2517, 1979.

Dixon, K. W., and J. R. Lanzante, Global mean surface air temperature and North Atlantic overturning in a suite of coupled GCM climate change experiments. *Geophys. Res. Lett.*, *26*, 1885-1888, 1999.

Hansen, J., A. Lacis, D. Rind, G. Russel, P. Stone, I. Fung, R. Ruedy and J. Lerner, Climate sensitivity: Analysis of feedback mechanisms, *Climate Processes and Climate Sensitivity, Geophys. Mono.*, *29*, 130-163, Amer. Geophys. Union, 1984.

Harvey, D., J. Gregory, M. Hoffert, A. Jain, M. Lal, R. Leemans, S. Raper, T. Wigley and J. de Wolde, An introduction to simple climate models used in the IPCC second assessment report, *IPCC Technical Paper II*, 47 pp., IPCC, Cambridge Univ. Press, Cambridge, 1997.

Haywood, J. M., R. J. Stouffer, R. T. Wetherald, S. Manabe and V. Ramaswamy, Transient response of a coupled model to estimated changes in greenhouse gas and sulfate concentrations, *Geophys. Res. Lett.*, *24*, 1335-1338, 1997.

Jones, P. D., T. J. Osborn and K. R. Briffa, Estimating sampling errors in large-scale temperature averages, *J. Clim.*, *10*, 2548-2568, 1997.

Legget, J., W. J. Pepper and R. J. Swart, Emission scenarios for IPCC: An update, *Climate Change: Supplementary Report to the IPCC Scientific Assessment*, IPCC, Cambridge Univ. Press, Cambridge, 1992.

Kattenburg, A., F. Giorgi, H. Grassl, G. A. Meehl, J. F. B. Mitchell, R. J. Stouffer, T. Tokioka, A. J. Weaver and T. M Wigley, Climate models-projections of future climate, *Climate Change 1995*, p. 285-357, IPCC, Cambridge Univ. Press, Cambridge, 1996.

Manabe, S. Climate and ocean circulation. Part I: The atmospheric circulation and the hydrology of the earth's surface, *Mon. Wea. Rev.*, 97, 739-774, 1969.

Manabe, S. and R. T. Wetherald, Large-scale changes of soil wetness induced by an increase of atmospheric carbon dioxide, *J. Atmos. Sci.*, *44*, 1211-1235, 1987.

Manabe, S., K. Bryan, and M. S. Spelman, Transient response of a global ocean-atmosphere model to a doubling of atmospheric carbon dioxide, *J. Phys. Oceanogr.*, *20*, 722-749, 1990.

Manabe, S., M. S. Spelman and R. J. Stouffer, Transient responses of a coupled ocean-atmosphere model to gradual changes of atmospheric CO_2, Part II: Seasonal response, *J. Clim.*, *5*, 105-126, 1992.

Manabe, S., R.T. Wetherald and R.J. Stouffer, Summer dryness due to an increase of atmospheric CO_2 concentration, *Clim. Change*, *3*, 347-386, 1981.

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

Mintzer, I. M. A matter of degrees: The potential for controlling the greenhouse effect, *Research Report No. 5*, World Resources Institute, Washington, D. C. , 60 pp, 1987.

Mitchell, J. F. B., T. C. Johns, J. M. Gregory and S. B. F. Tett, Climate response to increasing levels of greenhouse gases and sulfate aerosols, *Nature*, *376*, 501-504, 1995.

Mitchell, J. F. B., and D. A. Warrilow, Summer dryness in northern mid-latitudes due to increased CO_2, *Nature*, *330*, 238-240, 1987.

Spelman, M.J., and S. Manabe, Influence of oceanic heat transport upon the sensitivity of a model climate, *J. Geophys. Res.*, 89(C1), 571-586, 1984.

Stouffer, R. J. and S., Manabe, Response of a coupled ocean-atmosphere model to increasing atmospheric carbon dioxide: Sensitivity to the rate of increase, *J. Climate*, *12*, 2224-2237, 1999.

Tziperman, E., and K. Bryan, Estimating global air-sea fluxes from surface properties and from climatological flux data using an oceanic general circulation model. *J. Geophys. Res.*, 98, 22,629-22,644, 1993

Weaver, A. J. and E. C. Wiebe, On the sensitivity of projected oceanic thermal expansion to the parameterization of sub-grid scale ocean mixing, *Geophys. Res. Lett.*, *26*, 3461-3464, 1999.

Wiebe, E. C. and A. J. Weaver, On the sensitivity of global warming experiments to the parameterization of sub-grid scale ocean mixing, *Clim. Dyn.*, *15*, 875-893, 1999.

Wigley, T. M. L., The Kyoto Protocol: CO_2, CH_4 and climate implications, *Geophys. Res. Lett.*, *25*, 2285-2288, 1998.

Wigley, T. M. L. and S. C. B. Raper, Future changes in global-mean temperature and sea level. *Climate and Sea Level Change, Projections and Implications*, pp. 111-133, Cambridge Univ. Press, Cambridge, 1993.

Wigley, T. M. L. and B. D Santer, Future climate of the Gulf/Caribbean basin from atmospheric general circulation models, *Climate Change in the Intra-Americas Sea*, pp. 31-54, Edward Arnold, New York, 1993.

Wigley, T. M., L, A. Jain, F. Joos, P. R. Shukla and B. S. Nyenzi, *Implications of Proposed CO_2 Emission Limitations: IPCC Technical Paper 4*, 41 pp., IPCC, Geneva, 1997.

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