DETECTION OF GREENHOUSE-GAS-INDUCED CLIMATIC CHANGE

GRANT No. DE-FG02-86ER60397-A014

Final Report covering the period 1 March 1995 - 31 May 1998

P.D. Jones and T.M.L. Wigley

26 May 1998

CLIMATIC RESEARCH UNIT
School of Environmental Sciences
University of East Anglia
Norwich, NR4 7TJ
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CONTENTS

1. PROJECT SPECIFICATIONS

Title: Detection of Greenhouse-Gas-Induced Climatic Change

Area: Office of Health and Environmental Research, Global Change Program.

Principal Investigators: P.D. Jones and T.M.L. Wigley

Organization: Climatic Research Unit, University of East Anglia (UEA).

Budget: 1995/96 192K, 1996/97 200K, 1997/98 200K.

Objective: To assemble and analyse instrumental climate data and to develop and apply climate models as a basis for (1) detecting greenhouse-gas-induced climatic change, and (2) validation of General Circulation Models.

Product: In addition to changes due to variations in anthropogenic forcing, including greenhouse gas and aerosol concentration changes, the global climate system exhibits a high degree of internally-generated and externally-forced natural variability. To detect the anthropogenic effect, its signal must be isolated from the "noise" of this natural climatic variability. High quality, spatially extensive data bases are required to define the noise and its spatial characteristics. To facilitate this, available land and marine data bases will be updated and expanded. The data will be analysed to determine the potential effects on climate of greenhouse gas and aerosol concentration changes and other factors. Analyses will be guided by a variety of models, from simple energy balance climate models to coupled atmosphere-ocean General Circulation Models. These analyses are oriented towards obtaining early evidence of anthropogenic climatic change that would lead either to confirmation, rejection or modification of model projections, and towards the statistical validation of General Circulation Model control runs and perturbation experiments.

Approach: Global surface climate data bases will be expanded and updated using the extensive resources available to the Climatic Research Unit. Data analyses will focus on the use and development of appropriate statistical techniques for signal detection and pattern recognition. Interpretations will be guided by appropriate climate models, and models and model analyses developed especially for this project.

Deliverables: Estimates of global- and hemispheric-mean near-surface temperature, based on land and marine data will be made on a monthly basis. Gridded data sets of temperature, precipitation and pressure, produced during earlier projects will be updated and made available through CDIAC. These will be specifically useful for model validation purposes.

Program Coordination:

- o Estimates of greenhouse-gas concentration changes over the last 200 years.
- o Carbon cycle modeling (Scripps, IOS, Univ. New Hampshire, GFDL, and others).
- o Model estimates of the climatic response to external forcing agents including SO_2 (LLNL, NCAR, GFDL and others - including MPI and UK Hadley Centre).
- o Model estimates of the regional and global climatic response to anthropogenic forcing (all modeling groups).
- o Global precipitation analyses (Univ. Massachusetts).
- o General Circulation Model validation studies (LLNL-PCMDI, SUNY).
- o Climate data compilation and dissemination (CDIAC).

2. INTRODUCTION

The aims of the U.S. Department of Energy's Global Change Program (Environmental Sciences Division) are to improve assessments of anthropogenic climatic change and to define and reduce uncertainties through selected research. Four major questions can be identified.

(1) What are the regional and seasonal details of the expected climatic changes?

(2) How rapidly will these changes occur?

(3) How and when will the climatic effects of greenhouse gases and aerosols be first detected?

(4) Natural variability - what are the relationships between anthropogenic climatic change and changes caused by other external and internal factors?

The present project addresses all of these questions.

Many of the diverse facets of greenhouse-gas- and aerosol-related climate research can be grouped under three interlinked subject areas:

(a) Modeling. This involves the development, validation and use of climate models of different types to estimate the details of climatic change due to changing greenhouse gas and aerosol concentrations. Transient response aspects (i.e., modeling the time-dependent response to realistic time-dependent changes in forcing) are considered to be particularly important.

(b) First Detection. The most direct form of model validation is to be able to identify, in the observational record, the model-predicted, evolving signal of anthropogenic climatic change. This is the detection problem. Detection research includes better defining the anthropogenic signal, the signals of climatic change resulting from other forcing factors and the characteristics of natural climatic variability. Such information is central to determining how and when human influences on climate can be detected with a high level of confidence.

(c) Supporting Data. The compilation and homogenization of past instrumental and paleoclimatic data is essential to support activities in areas (a) and (b). Past data are required to elucidate the mechanisms and causes of climatic change, to define the range of past variations, to document possible analogs for future greenhouse-gas-induced climatic change, to estimate the sensitivity of the climate system to external forcing, and to aid in model development and validation.

The main research areas covered by this proposal are (b), First Detection and (c) Supporting Data. The project will also include work under area (a), Modeling: specifically, analysis of climate forcing factors, the development and refinement of transient response climate models, and the use of instrumental data in validating General Circulation Models (GCMs).

3. OUTLINE OF THE PROPOSED RESEARCH

We propose to continue the research work carried out in previous contracts in four main areas:

A. Global climate data. Updating, improvement and analysis of our global (land and marine) temperature data set.

B. Multivariate detection methods. The further development and use of multivariate techniques for the detection of both greenhouse-gas-induced and aerosol-related climatic change.

C. Transient response studies. The use of both simple and more complex transient-response climate models in order to throw further light on the natural variability of the climate system and the possible effects of aerosol-related forcing.

D. GCM validation. Validation of General Circulation Models using a variety of test statistics.

The way these items contribute to the major questions addressed by the Department of Energy's Global Change Program (see Section 2 above) is summarized in the following Table.

Project elements / Major questions	Global climate data (A)	Multivariate detection methods (B)	Transient response studies (C)	GCM validation (D)
Details of future changes (1)				X
Rapidity of future changes (2)			X	
First detection of effects (3)	X	X	X	X
Natural variability (4)	X	X	X	

Work on all items will be carried out continuously throughout the project's three-year duration.

4. PUBLICATIONS ARISING TO DATE

Summary

Over the 3-year period 73 scientific papers were produced that were either fully or partially supported by the project. These papers are listed below under the main aims of the project. The numbers of the references are keyed to the principal section of the progress report to which they relate.

Report Section	Year 1	Year 2	Year 3	Total
A1	1, 5-10, 12-14, 16-17	25-28, 31	47, 50-51, 53-4, 72	23
A2		29	52, 53	3
B	4, 18-22	37-38	62, 73	10
C1	3, 15, 24	36	69	5
C2	2		67	2
C3		45	65, 66	3
C4		35		1
C5				-
C6			61	1
C7		32-34	56, 60	5
D	11, 23		57-9	5
Miscellaneous		42-44, 46	48, 64, 68, 70	8
Contributions to IPCC		30, 39-41	49, 63, 71	7

Publications produced during year 1 (1995/96) of present project (a few publications that were in press from the previous project are included during year 1)

1. Bradley, R.S. and Jones, P.D., 1995: Recent developments in studies of climate since A.D. 1500. In, *Climate Since A.D. 1500 2nd Edition* (R.S. Bradley and P.D. Jones, Eds), Routledge, London, 666-679.

2. Enting, I.G., Wigley, T.M.L. and Heimann, M., 1994: Future emissions and concentrations of carbon dioxide: key ocean/atmosphere/land analyses. CSIRO Division of Atmospheric Research Technical Paper No. 31, 118pp.

3. Hulme, M., Raper, S.C.B., Wigley, T.M.L., 1995: An Integrated Framework to Address Climate Change (ESCAPE) and Further Developments of the Global and Regional Climate Modules (MAGICC). *Energy Policy* **23**, 347-355.

4. Jones, P.D., 1994: Towards detection of the enhanced greenhouse gas effect. In, *Climate Variations in Europe* (R. Heino, Ed.), Silmu, Helsinki, Finland, 69-70.

5. Jones, P.D., 1994: Decadal-to-century timescale variability of regional and hemispheric scale temperature. In, *Climatic Variations in Europe* (R. Heino, Ed.), Silmu, Helsinki, Finland, 136-140.

6. Jones, P.D., 1994: Decadal-to-century timescale variability of regional and hemispheric scale temperature. In, *Global Climate Change: Science, Policy and Mitigation Strategies* (C.V. Mathai and G. Stensland, Eds), Air and Waste Management Association, Pittsburgh, 98-101.

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10. Jones, P.D., Briffa, K.R. and Schweingruber, F.H., 1995: Tree-ring evidence of the widespread effects of explosive volcanic eruptions. *Geophysical Research Letters* **22**, 1333-1336.

11. Jones, P.D., Hulme, M., Briffa, K.R., Jones, C.G., Mitchell, J.F.B. and Murphy, J.M., 1996: Summer moisture availability over Europe in the Hadley Centre General Circulation Model based on the Palmer Drought Severity Index. *International Journal of Climatology* **16**, 155-172.

12. Jones, P.D., Wigley, T.M.L. and Gregory, J.M., 1994: England and Wales area-average precipitation amount. In, *Trends '93: A compendium of Data of Global Change* (T.A. Boden, D.P. Kaiser, R.J. Sepanski and F.W. Stoss, Eds), ORNL/CDIAC-65. Carbon Dioxide Information Analysis Center, Oak Ridge National Laboratory, Oak Ridge, Tennessee, 975-983.

13. Jones, P.D., Wigley, T.M.L. and Briffa, K.R., 1994: Global and hemispheric temperature anomalies - land and marine instrumental records. In *Trends '93: A compendium of Data of Global Change* (T.A. Boden, D.P. Kaiser, R.J. Sepanski and F.W. Stoss, Eds), ORNL-CDIAC-65. Carbon Dioxide Information Analysis Center, Oak Ridge National Laboratory, Oak Ridge, Tennessee, 603-608.

14. Kelly, P.M., Pengqun, J. and Jones, P.D., 1996: The spatial temperature response to large volcanic eruptions. *International Journal of Climatology* **16**, 537-550.

15. Palutikof, J.P. and Wigley, T.M.L., 1995: Developing climate change scenarios for the Mediterranean region. In *Climate Change and the Mediterranean,* Volume **2**, (L. Jeftic, J.C. Pernetta and S. Keckes, Eds), Edward Arnold, London, U.K., 27-56 .

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17. Salinger, M.J., Basher, R.E., Fitzharris, B.B., Hay, J.E., Jones, P.D., MacVeigh, J.P. and Schmidely-Leleu, I., 1995: Climate trends in the southwest Pacific. *International Journal of Climatology* **15**, 285-302.

18. Santer, B.D., Mikolajewicz, U., Brüggemann, W., Cubasch, U., Hasselmann, K., Höck, H., Maier-Reimer, E. and Wigley, T.M.L., 1994: Ocean variability and its influence on the detectability of greenhouse warming signals. *PCMDI Report No. 14,* Lawrence Livermore National Laboratory, Livermore, California, 73pp.

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23. Srinivasan, G., Hulme, M., Jones, C.G., Jones, P.D. and Osborn, T.J., 1995: An evaluation of the spatial and interannual variability of tropical precipitation as simulated by GCMs. In *Proceedings of the First International Atmospheric Model Intercomparison Project (AMIP) Scientific Conference*, 193-198 .

24. Wigley, T.M.L. and Raper, S.C.B., 1995: An heuristic model for sea level rise due to the melting of small glaciers *Geophysical Research Letters* **22**, 2749-2752.

Publications produced during year 2 (1996/97) of present project

25. Barnett, T.P., Santer, B.D., Jones, P.D., Bradley, R.S. and Briffa, K.R., 1996: Estimates of the low-frequency natural variability in near-surface air temperature. *The Holocene* **6**, 255-263.

26. Jones, P.D., 1996: A la recherche du temps perdu (News and Views). *Nature*, **381**, 375-376.

27. Jones, P.D. and Briffa, K.R., 1996: What can the instrumental record tell us about longer timescale paleoclimatic reconstructions? In *Climatic Variations and Forcing Mechanisms of the last 2000 years* (P.D. Jones, R.S. Bradley and J. Jouzel, Eds), Springer-Verlag, Berlin, 625-644.

28. Jones, P.D. and Hulme, M., 1996: Calculating regional climatic time series for temperature and precipitation: methods and illustrations. *International Journal of Climatology* **16**, 361-377.

29. Karoly, D.J., Hope, P. and Jones, P.D., 1996: Decadal variations of the Southern Hemisphere circulation. *International Journal of Climatology* **16**, 723-738.

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31 Kelly, P.M. and Jones, P.D., 1996: Removal of the El Niño-Southern Oscillation signal from the gridded surface air temperature data set. *Journal of Geophysical Research* **101**, 19013-19022.

32. Osborn, T.J., 1995: Internally-generated variability in some ocean models on decadal to millennial timescales. PhD thesis, University of East Anglia, 650pp.

33. Osborn, T.J., 1996: Are model simulations of oceanic variability relevant to the real world? *Climate Monitor* **24**, 23-28.

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35. Raper, S.C.B. and Cubasch, U., 1996: Emulation of the results from a coupled general circulation model using a simple climate model. *Geophysical Research Letters* **23**, 1107-1110.

36. Raper, S., Viner, D., Hulme, M. and Barrow, E., 1997: Global warming and the British Isles. In *Climates of the British Isles: present, past and future* (M. Hulme and E. Barrow, Eds), Routledge, London, 326-339 .

37. Santer, B.D., Taylor, K.E., Wigley, T.M.L., Jones, P.D., Karoly, D.J., Mitchell, J.F.B., Oort, A.H., Penner, J.E., Ramaswamy, V., Schwarzkopf, M.D., Stouffer, R.J. and Tett, S., 1995: *A search for human influences on the thermal structure of the atmosphere*. Program for Climate Model Diagnostics and Intercomparison (PCMDI) Report No. 27, Lawrence Livermore National Laboratory, Livermore, California, 26pp.

38. Santer, B.D., Taylor, K.E., Wigley, T.M.L., Johns, T.C., Jones, P.D., Karoly, D.J., Mitchell, J.F.B., Oort, A.H., Penner, J.E., Ramaswamy, V., Schwarzkopf, M.D., Stouffer, R.J. and Tett, S., 1996: A search for human influences on the thermal structure of the atmosphere. *Nature* **382**, 39-46.

39. Santer, B.D., Wigley, T.M.L., Barnett, T.P. and Anyamba, E., 1996: Detection of climate change and attribution of causes. In *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, New York, N.Y., 407-443.

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41. Warrick, R.A., Le Provost, C., Meier, M., Oerlemans, J. and Woodworth, P. (contributing author list includes Raper, S.C.B., Wigley, T.M.L. and 17 others), 1996: Changes in sea level. In *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, New York, N.Y., 358-405.

42. Wigley, T.M.L., 1995: A successful prediction? *Nature (News and Views)* **376**, 463-464.

43. Wigley, T.M.L., 1996: Tip of the iceberg. *Nature* (Scientific Correspondence) **382**, 574.

44. Wigley, T.M.L., 1996: IPCC report, chapter and verse. *Nature* (Scientific Correspondence) **383**, 214.

45. Wigley, T.M.L., Richels, R. and Edmonds, J.A., 1996: Economic and environmental choices in the stabilization of atmospheric CO_2 concentrations. *Nature* **379**, 240-243.

46. Wigley, T.M.L., Santer, B.D., Mitchell, J.F.B. and Charlson, R.J., 1996: Climate change report. *Science* (Correspondence) **271**, 1481-1482.

Publications produced during year 3 (1997/98) of the present project (The list for this year includes some from year 2 that have still not been published)

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48. Flannery, B.P., Kheshgi, H., Marland, G., MacCracken, M.C., Komiyami, H., Broecker, W., Ishatani, H., Rosenberg, N., Steinberg, M., Wigley, T.M.L. and Morantine, M., 1997: Geoengineering climate. In *Engineering response to global climate change* (R.G. Watts, Ed.) CRC Press, Lewis Publishers, Boca Raton, 379-427.

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50. Jones, P.D., 1997: Hemispheric and Global temperatures, 1851-1996, Part 1. *Climate Monitor* **25**, 20-30.

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

52. Jones, P.D., Jónsson, T. and Wheeler, D., 1997: Extension to the North Atlantic Oscillation using early instrumental pressure observations from Gibraltar and SW Iceland. *International Journal of Climatology* **17**, 1433-1450..

53. Jones, P.D., Osborn, T.J., Wigley, T.M.L., Kelly, P.M. and Santer, B.D., 1997: Comparisons between the MSU2R temperature record and the surface record over 1979 to 1996: real differences or potential discontinuities? *Journal of Geophysical Research* **102**, 30135-30145.

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57. Osborn, T.J., 1997: Areal and point precipitation intensity changes as simulated by climate models. *Geophysical Research Letters* **24**, 2829-2832.

58. Osborn, T.J. and Hulme, M., 1997: Development of a relationship between station and grid-box rainday frequencies for climate model validation. *Journal of Climate* **10**, 1885-1908.

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60. Osborn, T.J., 1998: The vertical component of isopycnal diffusion and the dianeutral component of horizontal diffusion. *Journal of Physical Oceanography* **28**, 485-494.

61. Penner, J.E., Wigley, T.M.L., Jaumann, P., Santer, B.D. and Taylor, K.E., 1997: Anthropogenic aerosols and climate change: a method for calibrating forcing. In *Assessing Climate Change - The Story of the Model Evaluation Consortium for Climate Assessment* (W. Howe and A. Henderson-Sellers, Eds), Gordon and Breach Science Publishers, Amsterdam, The Netherlands, 91-111.

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63. Schimel, D.S., Grubb, M., Joos, F., Kaufmann, R.K., Moss, R., Ogana, W., Richels, R. and Wigley, T.M.L., 1997: Stabilisation of atmospheric greenhouse gases: physical, biological and socioeconomic implications. IPCC Technical Paper 3. Intergovernmental Panel on Climate Change, Geneva, 52pp.

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67. Wigley, T.M.L., 1998: The importance of climate feedbacks in determining future methane concentrations. *Climatic Change* (in press).

68. Wigley, T.M.L., 1998: Stumbling at the start of a marathon. *Earth*, April 1998, 54-55.

69. Wigley, T.M.L., 1998: The Kyoto Protocol: CO_2, CH_4, temperature and sea level implications. *Geophysical Research Letters* (in press).

70. Wigley, T.M.L. and Santer, B.D., 1997: Up with IPCC climate report; down with per capita energy use. *Physics Today* (Letters), Nov. 1997, 90-91.

71. Wigley, T.M.L., Jain, A.K., Joos, F., Nyenzi, B.S. and Shukla, P.R., 1997: *Implications of proposed CO_2 emissions limitations.* IPCC Technical Paper 4, Intergovernmental Panel on Climate Change, Geneva, 41pp.

72. Wigley, T.M.L., Jones, P.D. and Raper, S.C.B., 1997: The observed global warming record: what does it tell us? *Proceedings of the National Academy of Sciences* **94**, 8314-8320.

73. Wigley, T.M.L., Jaumann, P.J., Santer, B.D. and Taylor, K.E., 1998: Relative detectability of greenhouse-gas and aerosol climate change signals. *Climate Dynamics* (in press).

5. PROGRESS

A. Global Climate Data

A1. Temperature

Gridded Analyses

Two sets of gridded temperature anomalies based on land stations continue to be produced. The first updates the original analysis (Jones *et al.*, 1986a,b, and Jones, 1988a) and interpolates temperature values to a 5° latitude by 10° longitude grid, with a reference or base period of 1951-70. This dataset has been combined with sea surface temperature anomalies over marine regions into 5° x 5° boxes with the reference period adjusted to that of the marine data (1950-79). The primary reference for this data set is Jones and Briffa (1992). Unfortunately the marine data are no longer available with a base period of 1950-79 so, in future, only the land component, referenced to 1951-70 will continue to be produced.

The second and more recent analysis (Jones, 1994) averages the land station temperature anomalies, with respect to the new World Meteorological Office reference period of 1961-90, directly into the same 5° x 5° boxes as the marine data. The second dataset (Jones, 1994) has been combined with the Hadley Centre's (U.K. Meteorological Office) analysis of sea surface temperatures (Folland and Parker, 1995). To achieve this, the sea surface temperature anomalies were adjusted to the new base period of 1961-90 (Parker *et al.*, 1995).

The second analysis has been extensively used by the Intergovernmental Panel on Climate Change (IPCC) in its second assessment report (see the chapter on climate observations, Nicholls *et al.*, 1996). Both Principal Investigators (PIs) and Dr Sarah Raper have been involved in the IPCC second assessment as lead and contributing authors and reviewers in the area of instrumental data, modelling, detection and radiative forcing.

Updates of both of our land-only datasets have been made available to the Carbon Dioxide Information Analysis Center (CDIAC) in Oak Ridge, Tennessee and the National Center for Atmospheric Research (NCAR) in Boulder, Colorado. The earlier (1950-79 reference period) combined analysis is also available from CDIAC and NCAR. Many more scientists use the hemispheric and global time series published in a number of places (TRENDS 93: Boden *et al.*, 1994). All datasets are also available from the Climatic Research Unit *via* our server (http://www.cru.uea.ac.uk). The 1961-90 combined analysis used extensively by IPCC is also available on a CD (Jackson *et al.*, 1996). Our observational database is discussed in more detail in Wigley *et al.* (1997 - reproduced here as Appendix A).

Recent Trends

The warmth of 1994 and 1995 continued until the end of November 1995, when a marked cooling affected the Northern Hemisphere. Despite the cool December in 1995, the average temperature for the calendar year of 1995 made it the then warmest year in both temperature analyses, 0.39°C above the 1961-90 level (0.04°C above 1990). The surface warmth of that year has also been discussed by Hansen *et al.* (1996).

The corresponding value for 1996 was 0.22°C above the 1961-90 level, markedly cooler than 1995, but warmer than 1992/3 when dust and aerosols from Mt. Pinatubo had their greatest effects. Cooling in 1996 was partly due to a dramatic switch in the North Atlantic Oscillation (NAO) during winter (95/96) from the positive phase which had dominated from the mid-1980s up to 1994/5, to the most negative value yet recorded in 1995/6 (see Figure 3 later). Negative phase NAO winters tend to be cool winters even on a hemispheric basis (see Hurrell 1996).

The NAO returned to a near normal value in the Northern Hemisphere (NH) winter (96/97) and this, combined with the large El Niño/Southern Oscillation (ENSO) warm event raised temperatures dramatically during the second half of the year, made 1997 the warmest year globally, 0.43°C above the 1961-90 base period.

Figure 1 shows hemispheric and global annual temperatures based on the combined land and marine temperatures. The last 10 years of the global series (1988-1997) contain 10 of the 13 warmest years of the series. The only warm years not in this period are 1983 (7th), 1987 (8th) and 1944 (10th). Table 1 gives the values for the 10 years and their ranks. The average value for the 10 years is 0.27°C above the 1961-90 average, a dramatic rise in temperature above the level of the 1970s and

Table 1: Annual global estimates from land and marine data temperature (°C with respect to 1961-90).

Year	Global Value	Rank (Warmest = 1)
1988	0.25	6
1989	0.18	12
1990	0.35	3
1991	0.29	4
1992	0.15	13
1993	0.19	11
1994	0.26	5
1995	0.39	2
1996	0.22	9
1997	0.43	1

early 1980s (see Figure 1). The breakdown of the curve into the two hemispheric series indicates a greater rise in the NH (average for the same 10 years of 0.30°C) than the Southern Hemisphere (SH: 0.23°C). Variability from year-to-year over this time period is also greater in the NH, but this feature is not new, having occurred since the beginning of the record in 1856.

The first few months of 1998 are warmer than the same months in 1997. If the warmth continues, which is to be expected from the still occurring but weakening warm ENSO event, we estimate that 1998 will be warmer still.

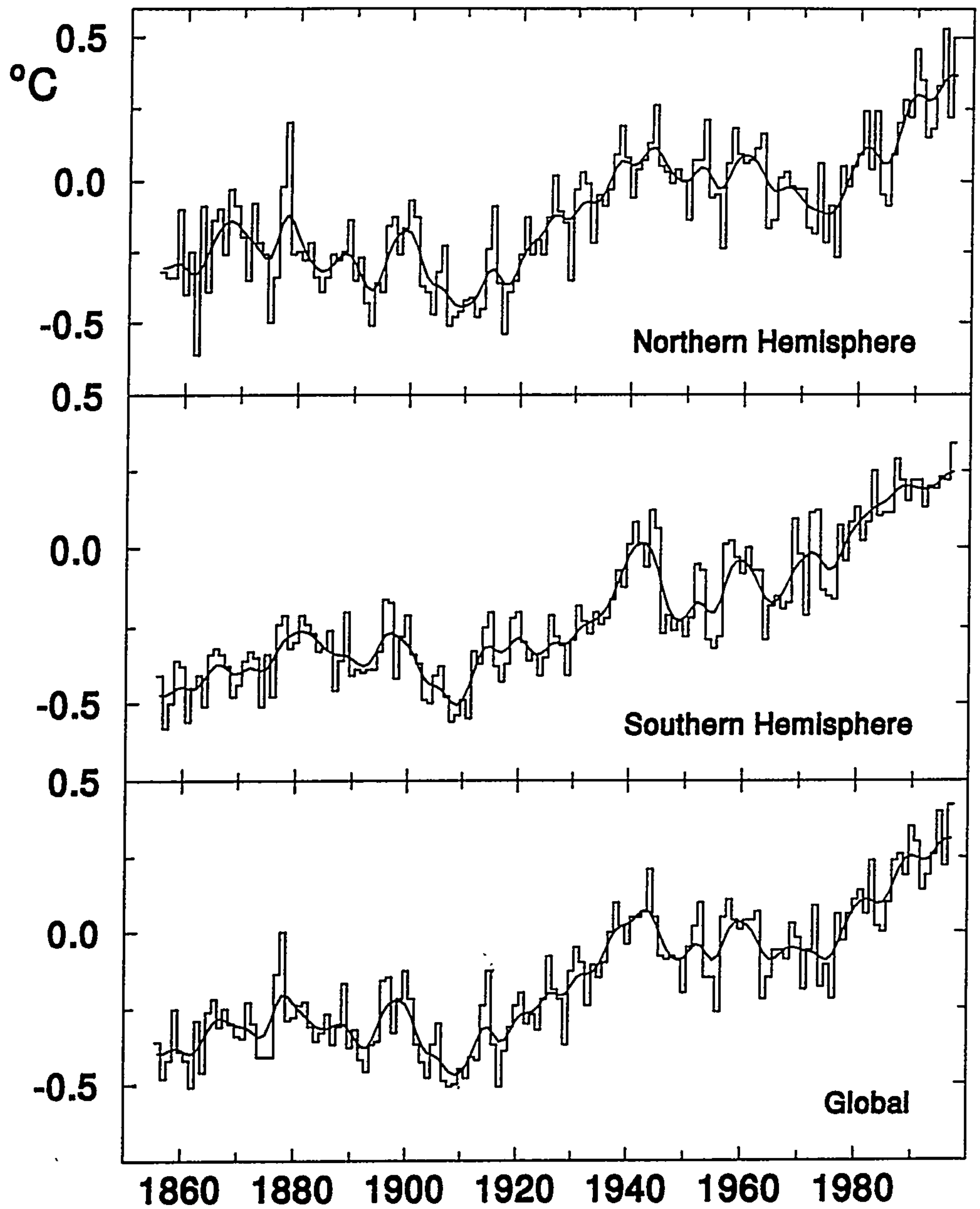


Figure 1: Average annual hemispheric and global temperatures since 1856. The smooth line is a decadal filter fit through the data.

Surface temperature comparisons with lower tropospheric temperature estimates

Considerable interest in the popular scientific media has centred on the series of average lower tropospheric temperatures developed from the microwave sounding units (MSUs) on NOAA series satellites. The MSU data extend from 1979 to the present. This interest has been partly fuelled by the surface warmth in 1995 and the growing warmth of the surface relative to MSU2R (Spencer and Christy, 1992). 1995 was clearly warm at the surface, but the MSU2R series (which mainly measures the average temperature in the atmospheric column from the surface up to roughly 250 mb) indicates it was average for the 1979-1996 period. There are a number of reasons why the two records differ and these have been extensively addressed (most recently by IPCC in Nicholls *et al.*, 1996, and also by Hurrell and Trenberth, 1996).

We have extensively analysed the two datasets on a 5° x 5° grid box basis and a paper has been published by the *Journal of Geophysical Research* (Jones *et al.*, 1997a - reproduced here as Appendix B). The comparison reveals differences in the course of temperature trends over the 1979-96 period in the two time series. Surface data warms relative to MSU2R by 0.19°C per decade over this period, with much of the change occurring as apparent jumps in the difference series during 1981 and 1991. The differences either reflect problems in one or both of the surface or MSU2R records or, if both records are correct, as seems most likely, a significant change in lapse rates in the lower part of the atmosphere on a global scale especially after 1991.

Both the MSU2R and the surface records can be independently checked with other datasets : the MSU2R with radiosonde measurements (Parker *et al.*, 1997), while the independent land and marine components of the surface record mutually support each other (e.g., Jones and Briffa, 1992 and Nicholls *et al.*, 1996). The recent divergence of the trends of the surface and the MSU2R is all the more perplexing given that the trends of surface and radiosonde data agree almost exactly over longer time intervals (e.g., 1965-96; Parker *et al.*, 1997). The differences have been used by greenhouse-sceptics to cast doubt on the longer-term warming evident in the surface record (see Figure 1), but if the differences are real, climatologists should be seeking the reasons rather than discussing whether one record is better than the other.

Other Temperature Analyses

We have developed a technique to extract the ENSO-related signal in surface temperature on a grid-box basis (Kelly and Jones, 1996). Up until this time, this had only been attempted on a hemispheric or global basis (Jones, 1988b; Jones 1995a, and Hurrell, 1996). The spatial removal was achieved by relating the Southern Oscillation Index (SOI) to the first six Principal Components (PCs) of surface air temperatures. The SOI was found to be significantly correlated with PC1 and PC3 at different lags. The whole of PC3 and the regression-based SOI-related part of PC1 was removed. It was hoped that this removal of the ENSO signal from the observed surface temperature dataset might aid in detection exercises (e.g. Santer *et al.*, 1995). This has been attempted but has resulted in no significant improvement, the decadal smoothing used in Santer *et al.* (1995) effectively removing the ENSO influence as well (see Section B on Multivariate Detection).

The other major piece of work in this section during year 3 of this project has been the development of a method for estimating the uncertainty of observed, regional, hemispheric and global-mean temperature series (Jones *et al.*, 1997b). The method assesses the sampling error at the grid box level. The sampling error depends upon three parameters; the number of site records within each grid box; the average inter-record correlation between these sites and the temporal variability of each grid box temperature time series. Various methods are used to estimate the parameters and these are tested with both the observed data and 1000-year long General Circulation Model (GCM) control runs.

The parameters used in the method imply that the errors depend on the timescale of interest, reducing as the timescale increases. Large-scale (hemispheric and global) standard errors (expressed as variances) can be computed by taking the areal average of the grid box standard errors (expressed as variances) divided by the number of effective or independent samples over the surface of the Earth. For observed global data as annual means, the 95% confidence interval on the interannual timescale for years since 1951 is ±0.11°C. Prior to 1900 the confidence interval is ±0.18°C. Equivalent values on the decadal timescale are ±0.10°C (1951-1995) and ±0.16°C (1851-1900). Figure 2 shows the hemispheric and global temperature series on the decadal timescale with appropriate 95% confidence intervals. The hemispheric, global and 5°x5° grid box error fields have been made available to any interested scientists over the Climatic Research Unit's web page. They should prove particularly useful in detection studies (see e.g., Hegerl *et al.*, 1996).

All the methods devised were assessed using temporal variability and spatial coherence statistics of surface temperature, computed from observations and from long model control run data. Comparison of these statistics can be considered as model validation and the results are summarized in section D of this report.

We are continuing to collect maximum and minimum temperature time series from locations in many regions of the world. Many of these data have been received on disk although many other sites require the data to be coded from this material. This work is therefore still ongoing for continued assessment of differential rates of warming between day and night temperatures (e.g. Karl *et al.*, 1993, Jones 1994 and Jones 1995b). Jones has been involved in an analysis of all available maximum and minimum temperature data, which is described in Easterling *et al.* (1997). This updated analysis indicates that the differential rise of minimum temperatures, compared to maximums, is continuing to occur during the early 1990s and is apparent in all new regions where quality data of this type can be located.

The Global Climate Observing System (GCOS) is now up and running and Jones has been involved in attempts to assess the adequacy of the surface temperature database (see e.g. Jones 1995a). Jones initially produced a list of 800+ stations as a starting point for the designation of a GCOS reference network. A later meeting organized by the World Meteorological Organization (WMO) in Norwich in February 1996 initiated the development of an algorithm to provide a numeric basis for the station selection. This work has just been completed (Peterson *et al.*, 1997) and the list has been circulated to all countries by WMO. Responses from the countries are beginning to be received and a second meeting to assess these and possibly decide upon alternatives for GCOS stations was held in De Bilt, The Netherlands during June 1997. The network has been modified slightly following national preferences. Most countries have accepted the revised network and it is hoped that the network will remain the same for many decades.

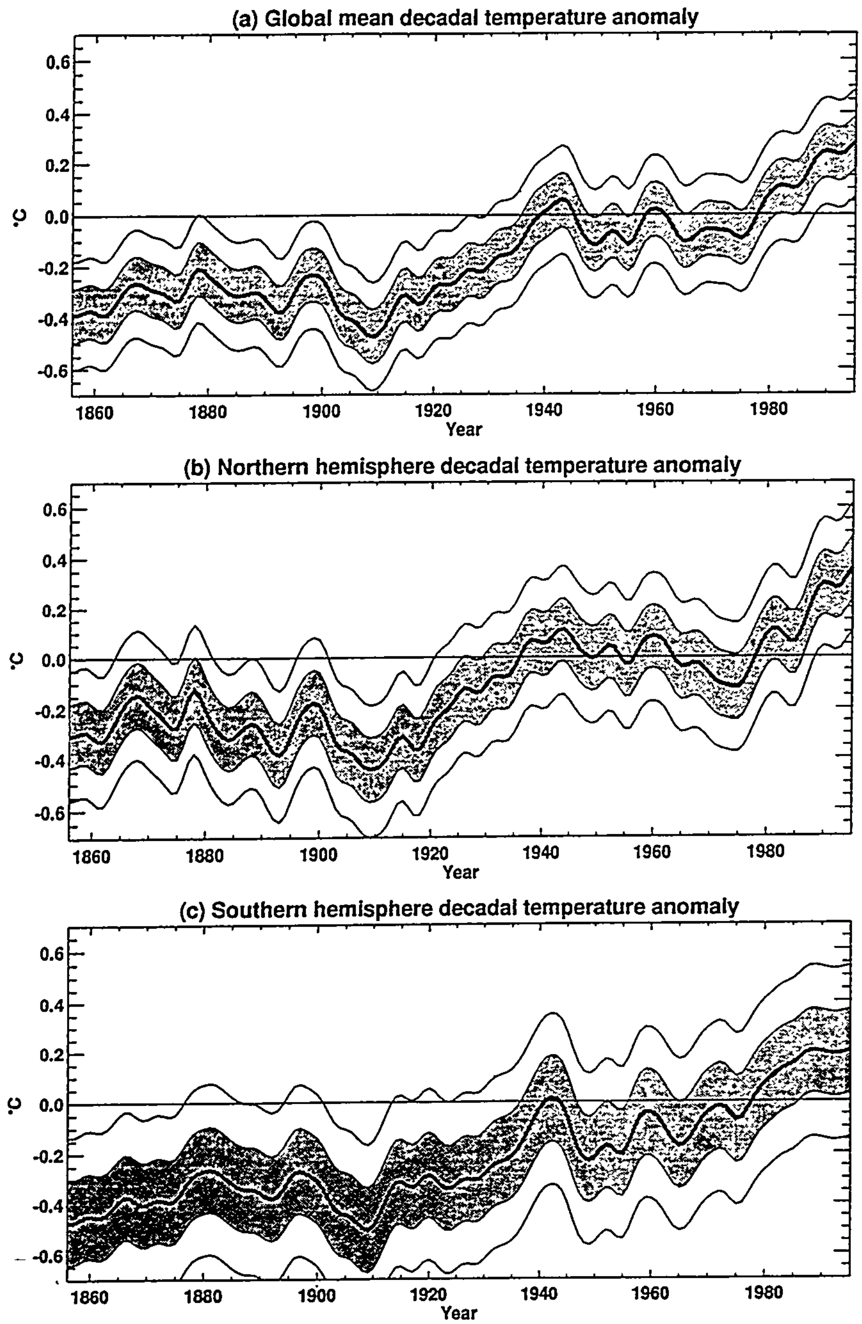


Figure 2: Decadal timescale surface temperature record for (a) global; (b) Northern Hemisphere; and (c) Southern Hemisphere means, with ±1 standard error (shaded) and ±2 standard errors (thin lines) indicated. Anomalies (°C) relative to 1961-90 mean.

A2. Mean-sea-level pressure (MSLP)

We have made available to the Division of Atmospheric Research, CSIRO, Aspendale, Australia and the Hadley Centre of the UK Meteorological Office all the monthly mean pressure series (both station and gridded MSLP) that we have collected and analysed during the last 10 years of Department of Energy support. These data have been reanalysed and published as an Atlas and CD (Allan *et al.*, 1996).

We have also continued to collect and update our series of MSLP data in the Eurasian and North American sectors and in the Antarctic. No new analyses of this material have been made. Two pieces of work in other areas have, however, been completed.

The first represents an extension to a North Atlantic Oscillation (NAO) series using early pressure series located in Gibraltar, southern Spain and Reykjavik in Iceland. Both series can be extended back to 1821. The homogeneity of both series presents challenges as prior to about 1850 there are few records near either site for comparison. The homogeneous pressure and the extended NAO series are now published (Jones *et al.*, 1997c); work performed in conjunction with a colleague in Iceland and one in Sunderland, UK who had undertaken earlier work in southern Spain.

Figure 3 shows the extended NAO series for the winter (November - March) produced using the Gibraltar and Reykjavik pressure records. The longer 1822-1996 record highlights the recent trend towards more positive NAO values since the mid-1970s. The period has been shown in this analysis to be very unusual in the 170-year context of this record. The second most positive NAO value in the winter series occurred in 1994/5 with the following year (1995/6) witnessing the most negative value yet recorded. This dramatic change in the circulation of the North Atlantic led to the much cooler temperatures (particularly over much of Eurasia) in early 1996 compared to the warmth of 1995 (see Figure 1). Stronger westerlies in the North Atlantic, especially since the mid-1980s, have also led to greater precipitation amounts being recorded over much of northwestern Europe during the winter half year. The enhanced precipitation totals have been particularly evident in regions of high orography (e.g., western Scotland and western Norway; Jones and Conway, 1997). The switch in the NAO for the last two winters (1995/6 and 1996/7) has led to reduced precipitation totals in the first winter and enhanced precipitation in the second in southwestern Europe (Spain, Portugal and Morocco), breaking longstanding drought conditions in this region. Large scale changes of this type, in precipitation as well as temperature patterns in Europe during winters, are in accord with the work of Hurrell (1995).

The extended NAO series, and our gridded datasets of MSLP, surface temperature and terrestrial precipitation, have been used to evaluate the Hadley Centre's coupled climate model (HADCM2) in terms of its simulation of the North Atlantic Oscillation during a 1400-year control integration. The results of this evaluation are described in section D of this report, but it is worth mentioning here that the recently observed trend towards positive NAO index values (Figure 3) cannot be explained by either 'natural' internally-generated variability or variations driven by

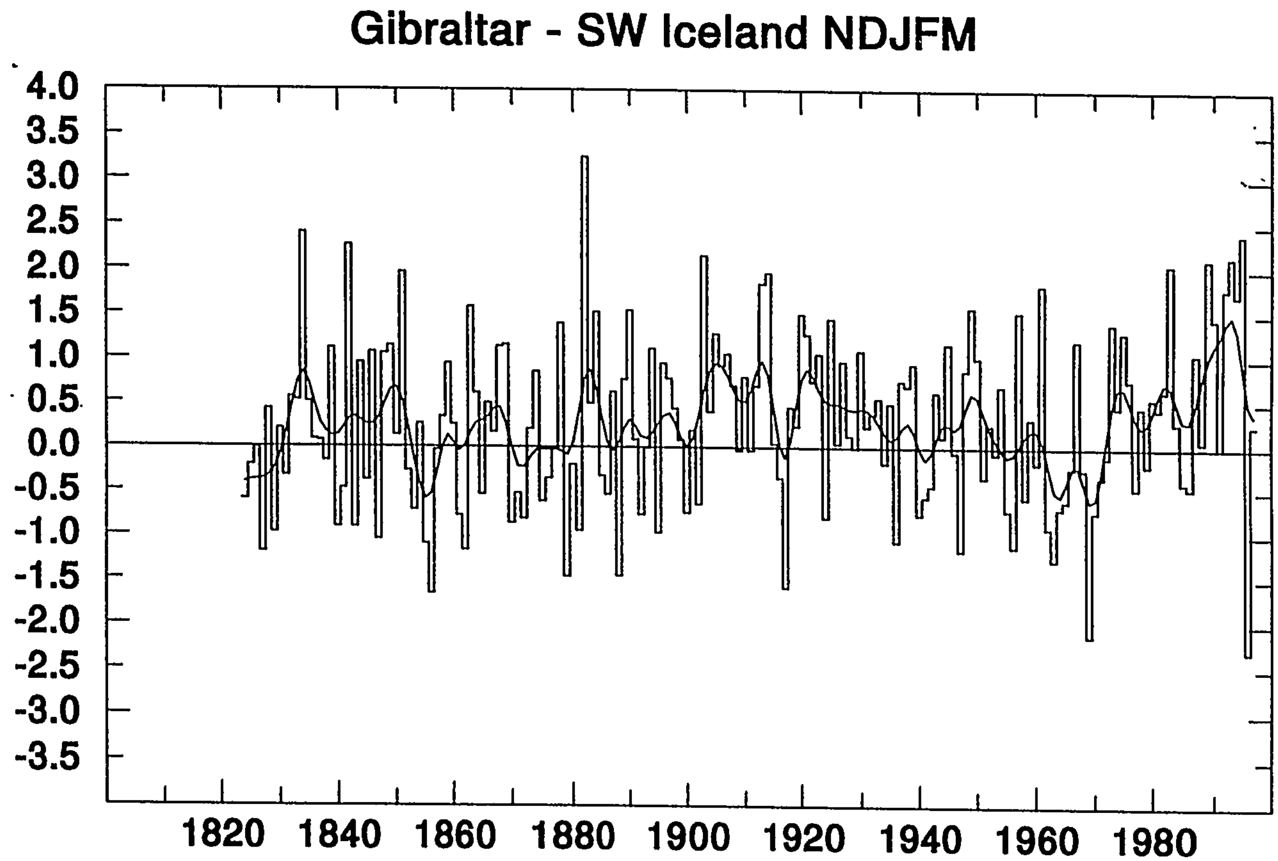


Figure 3: The North Atlantic Oscillation for the November to March period based on the normalized pressure at Gibraltar minus the normalized pressure at Reykjavik (SW. Iceland). Winters are dated by the January and normalization was performed over the 1951-80 period.

increased greenhouse gas or sulphate aerosol forcing, according to the HADCM2 simulations at least.

The second piece of work presents an extension of the Southern Oscillation Index (SOI) using early pressure data from Jakarta, Indonesia and Tahiti (Können *et al.*, 1998). Figure 4 illustrates this work by comparing SOIs produced using Jakarta and using Tahiti and Darwin. As with the NAO there has been a recent tendency towards departures of the same sign - in the case of the SOI to more negative values and associated increased frequency of El Niño events. The rarity of this feature, particularly the prolonged 1990-5 El Niño/Southern Oscillation (ENSO) event, has been assessed by Trenberth and Hoar (1996) based on Darwin pressure data which extends back to 1882. The new data uncovered for Jakarta and Tahiti will enable assessments like Trenberth and Hoar's to be made using much longer records.

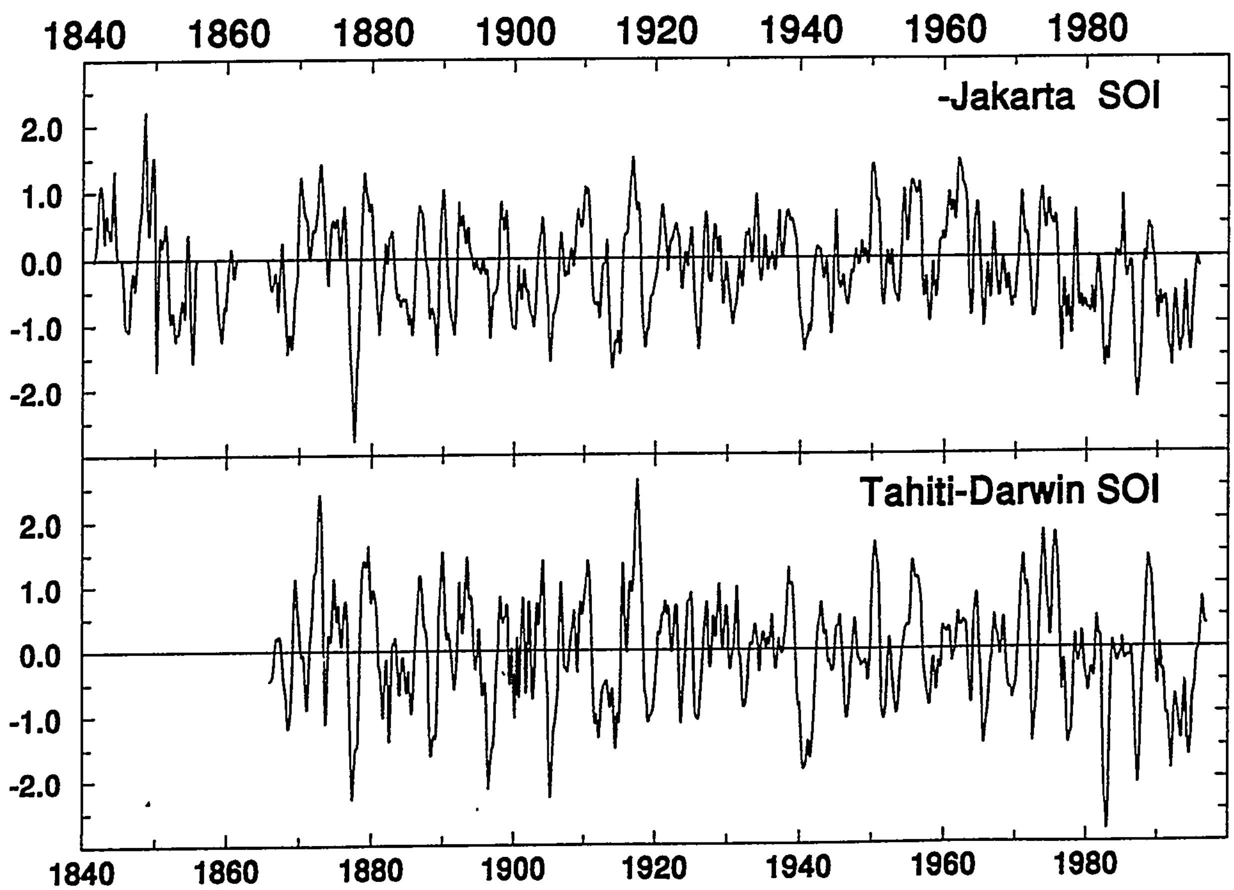


Figure 4: Time series of the Jakarta SOI series for 1841-1995 (top) compared to the normal SOI (Tahiti-Darwin) for 1866-1997 (bottom). Both curves show monthly values after filtering with a 12 month Gaussian filter.

B. Multivariate Detection Methods

The major pieces of work under this category are Wigley *et al.* (1998) and Santer *et al.* (1996b). Both papers extend the Santer *et al.* (1995) analysis, which was summarised in the last report. The Santer *et al.* (1995) work, in turn, is an application of the pattern correlation method we developed in earlier work (Santer *et al.*, 1993) to the equilibrium GCM results from Taylor and Penner (1994) for the CO_2-alone, aerosol-alone and combined CO_2-aerosol experiments. In the 1995 paper we stated that there was a statistically significant trend in the detection statistic R(t) for the combined CO_2-aerosol forcing during the JJA and SON seasons for horizontal (latitude-longitude) temperature change patterns. As we stated in the previous progress report this paper has come in for much scrutiny by the scientific community. It has a prominent place in the most recent IPCC report (Santer *et al.*, 1996a).

The new work by Wigley *et al.* (1998) complements the 1995 paper by using synthetic observed data to study the expected behaviour of the pattern correlation statistic, R(t). The main conclusions are:-

- even with a perfectly known signal and near present day surface temperature data coverage, the expected maximum value of R(t) to date is only about 0.5.
- R(t) values have no direct relationship to the strength of the signal.
- R(t) values need not increase linearly with time. They may even decrease for short periods even when the signal is increasing.
- Individual realizations of R(t), such as in the real world, may show substantial inter-decadal variability.
- The combined CO_2-aerosol pattern should be easier to detect than the CO_2-alone or the aerosol-alone patterns. Detectability would be improved if the past history of aerosol forcing were known better.
- The use of a simple low pass temporal filter to smooth the data enhances detection by reducing noise levels.

All of these items address aspects of R(t) that occur in analyses of the real-world observed temperature data. They explain why these results look the way they do, and they counter criticisms of the R(t) method and its results that have been raised by the greenhouse "skeptics".

An equally significant paper is that recently published in Nature (Santer *et al.*, 1996b). Both PIs on this project have been extensively involved in this work. The paper extends the work of Santer *et al.* (1995) using the same (and additional) model data but applying the detection method to the temperature structure of the atmosphere. The Santer *et al.* (1996b) paper has recently received two awards: the Outstanding Scientific Paper Award from the Environmental Research Laboratories of NOAA for

1997 and the Norbert Gerbier MUMM prize from the World Meteorological Organization.

The main conclusion of the work is summarized by the opening paragraph in *Nature* 382, p. 39.

> 'The observed spatial patterns of temperature change in the free atmosphere from 1963 to 1987 are similar to those predicted by state-of-the-art climate models incorporating various combinations of changes in carbon dioxide, anthropogenic sulphate aerosol and stratospheric ozone concentrations. The degree of pattern similarity between models and observations increases through this period. It is likely that this trend is partially due to human activities, although many uncertainties remain, particularly relating to estimates of natural variability.'

These papers (Santer *et al.*, 1995, 1996b) and related work carried out elsewhere (e.g. Hegerl *et al.*, 1996; Tett *et al.*, 1996), all of which have been in preparation for some time, were of major significance in leading IPCC to conclude in its 1996 report (Houghton *et al.*, 1996, p. 4) that 'The balance of evidence suggests a discernible human influence on global climate'.

C. Transient Response Studies

C1. Future global-mean temperature and sea-level changes, including probabilistic assessments

We have continued to adapt and change the gas cycle/climate model in the light of scientific progress as described below and in sections C4 and C5. It was not therefore appropriate to carry out any probabilistic assessments in the current reporting period (beyond what we reported on in an earlier Progress Report, 1994).

Improved Ice Melt Models

Building on previous work (Wigley and Raper, 1995: Raper et al., 1996a) and towards refining our estimates of future ice melt we have coupled the mass-balance degree-day model of Braithwaite and Zhang (submitted) to a new simple geometric glacier model. This model can be used to compare the response to climate change of individual glaciers of different size and shape, which presently exist under different climate conditions.
The geometric model equations involve scaling factors for relating glacier area to volume and width to length. Such scaling factors have recently been investigated using world glacier inventory data (Bahr *et al.*, 1997; Bahr, 1997). In the coupled model, the change in volume is calculated as the product of the net mass-balance averaged over the glacier and the glacier area. The time evolution of glacier area and width and length are determined from

$$(S_t/S_0) = (V_t/V_0)^n \quad (1)$$

and

$$(W_t/W_0) = (L_t/L_0)^m \quad (2)$$

where S_t, V_t, W_t and L_t are the glacier area, volume, mean width and mean length respectively, at time t. The zero subscripts represent the present or reference values. Equation (1) also determines the evolution of the mean glacier depth. The equations (1) and (2) are linked because the area is the product of the length and width. Finally, to apply the model to a particular glacier, it is necessary to define the reference altitude distribution of glacier area based on the observed data. The scaling parameters, n and m, determine the changes in the altitude distribution of the glacier area for any change in volume. This information is subsequently used in the calculation of the area average net mass-balance at the next time step.

Thus the geometric model determines the altitude distribution of the glacier area for any volume. For any particular volume, the shape of the glacier is the same irrespective of whether the glacier is in advance or retreat. Sensitivity studies show that when perturbed with, for example, a 1°C warming, the model can reproduce the differing response of glaciers of different shape under different climate conditions as expected from the glaciological literature. As a case study, we have applied the coupled model to Hintereisferner, Austrian Alps, using climate data from a nearby station (Fig. 5). Subsequently this model will be applied by region over all the worlds glaciated regions. This will entail assumptions about the distribution of glacier altitudes, shape and size in the various regions.

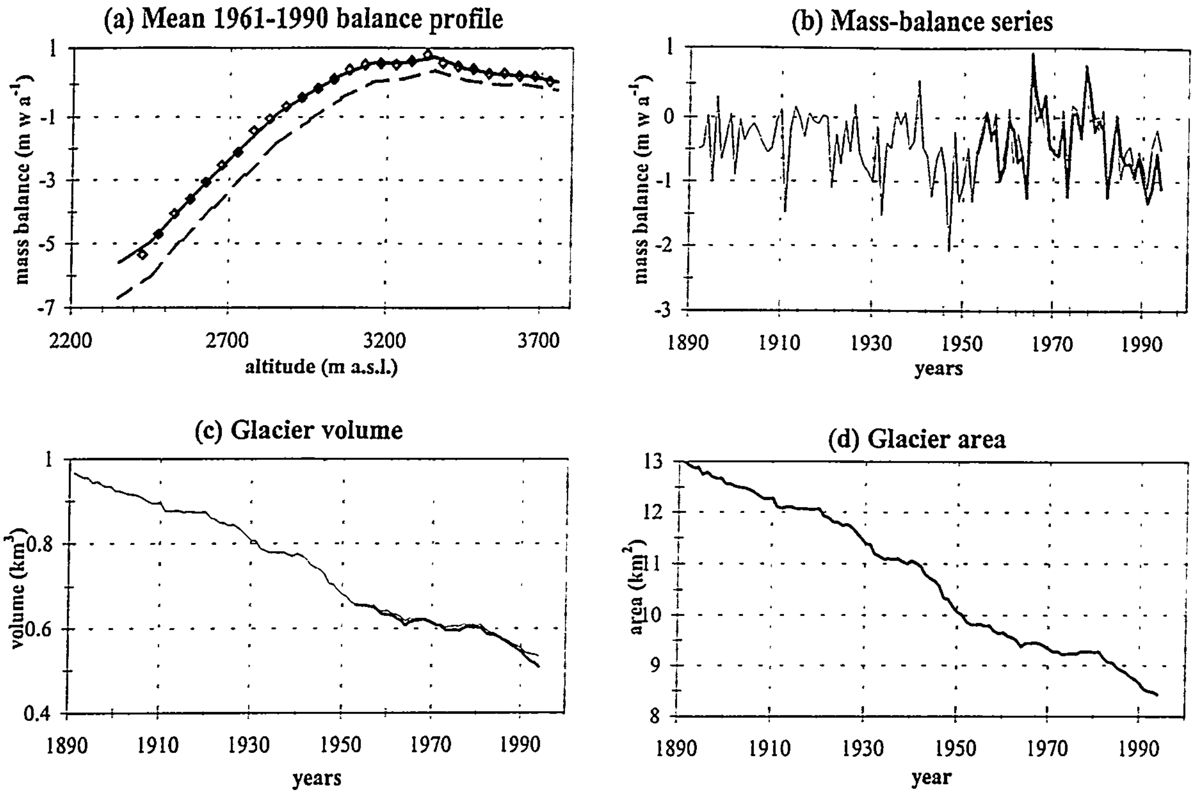


Fig 5: Observed and modelled mass-balance, volume and area for Hintereisferner, Austrian Alps: (a) altitude distribution of observed (diamonds) and modelled (solid line) mean mass-balance over 1961-1990, and modelled sensitivity to a 1^{o}C warming (dashed), (b) annual mass-balance versus time; observed, 1953-1995 (thick), and modelled, 1891-1995 (thin), (c) volume versus time; based on observed mass-balance (thick), and modelled (thin), (d) modelled area versus time.

The response time of the Greenland and Antarctic ice sheets is long compared to the time scale considered here, so at present, the areas of these ice sheets are assumed to be constant and their volume change with time is related directly to their mass balance. The mass balance is divided into two components. The first represents gain or loss of ice due to any initial disequilibrium state of the ice sheet and has units mm/year sea level equivalent. For Antarctica, we have previously (for example, in our recent calculations of sea level rise for the IPCC Second Assessment Report; Warrick *et al.*, 1996) used a value of 0.1mm/yr for the initial disequilibrium state. Huybrechts (1992), however, gives a value for the mean over the last 100 years as 0.39mm/yr. The use of 0.39mm/yr improves the modelled estimate of past sea level rise (see Section C5). The second component remains unchanged.

C2. Gas cycle/climate modelling

We have published an important paper under this heading in collaboration with two economists, Richard Richels (EPRI) and Joe Edmonds (Battelle, PNL); Wigley *et al.* (1996 - given in full in Appendix C). This work considers the issue of CO_2 concentration stabilization and proposes alternative pathways towards stabilization than those originally devised (by us, in collaboration with Ian Enting of CSIRO, Australia) for IPCC (see Enting *et al.*, 1994, for details). Using the inverse version of our carbon cycle model, the WRE (as this work has come to be known) pathways allow CO_2 emissions to follow an existing policies (or "Business as Usual", BAU) pathway for a decade or more, instead of requiring an immediate reduction below BAU (as required by the earlier IPCC pathways).

In our work, we constrain CO_2 emissions to follow BAU initially, as an idealization of a slow departure from BAU. On economic grounds, there are reasons both to expect and to prefer a slow departure from BAU. We do not, however, advocate any delay in responding to the need to reduce, eventually, CO_2 emissions substantially below the BAU trajectory.

WRE also consider the global-mean temperature and sea level consequences of stabilization, and the differential consequences for different pathways towards the same stabilization level. The differentials are small (much smaller in relative terms than the emissions differentials) and are critically dependent on what one assumes for future SO_2 emissions and aerosol forcing. Interpreting these differentials in terms of differentials in averted impacts (i.e., in the "benefits" side of a cost-benefit analysis) is exceedingly difficult, because impacts assessments require regionally-specific climate change information that is currently unreliable at best or unavailable at worst. Because of this, the WRE paper has spawned a rapid growth in studies of the cost side of the equation - i.e., the costs of reducing CO_2 emissions. The WRE work has a prominent place in the latest IPCC reports (Schimel *et al.*, 1996; Schimel *et al.*, 1997). Full details of the carbon cycle model calculations are given in Wigley (1998a).

Some additional gas cycle modelling work has been carried out on methane, investigating the potential feedback on methane concentration change associated with global warming. A simple temperature-dependent term has been added to the natural methane emissions component in our coupled gas-cycle/climate model, so that, as the globe warms, these natural emissions change. The enhanced methane increase accruing by this method is then interpreted in terms of its differential effects of global-mean temperature and sea level. The effects are found to be very small, even using unrealistically large magnitudes for the emissions feedback. This work is to be published in the journal *Climatic Change* (Wigley, 1998b).

A number of other publications have recently arisen from the above work. We note above that the main IPCC sea level calculations (Warrick *et al.*, 1996) were carried out using the model framework developed as part of this project. In addition, the standard IPCC global-mean temperature projections (Kattenberg *et al.*, 1996) were produced by us using this model framework. In spite of advances in more sophisticated coupled ocean/atmosphere GCMs, the use of models like ours is still necessary; mainly because

- no single O/AGCM can investigate key uncertainties associated with the climate sensitivity and other model-specific factors

- while different O/AGCMs have different sensitivities, large differences between models preclude using them to assess sensitivity uncertainties in an internally consistent way

- computing constraints make it impossible to examine the full spectrum of future emissions possibilities with O/AGCMs.

These IPCC analyses were extended in IPCC Technical Paper 4 (TP4 - Wigley *et al.*, 1997), which provides an assessment of the efficacy of the Kyoto Protocol. This work was further extended in Wigley (1997). While also dealing with the complications of the Kyoto Protocol, this work was not constrained by IPCC publication guidelines, which meant that publications more recent than the IPCC SAR (Houghton *et al.*, 1996) could be used. Wigley (1997) provides a more scientifically up-to-date analysis than that given in TP4, and considers issues that TP4 could not because of political constraints (such as the role of developing countries in meeting the ultimate stabilization goal of the United Nations Framework Convention on Climate Change). Both of these papers were published prior to the Kyoto meeting. The implications of the agreed Protocol are described in Wigley (1998c).

C3. Inverse climate and sea level modelling

We have deferred further work on inverse modelling pending the inclusion of recent and planned improvements in the models. Some inverse carbon cycle model results are included in items covered under C2. In addition, inverse methane calculations are given in Wigley (1998c) in order to determine the reduction in emissions of methane required to meet the Kyoto Protocol target and address the issue of how to determine "CO_2-equivalent" emissions.

C4. Comparison between UD model and O/AGCM results

In our progress report for FY94/95, we reported on a comparison of the results of our upwelling-diffusion (UD) model with those of the GFDL O/AGCM. The models appear to agree well over a wide range of forcings for both surface temperature change and thermal expansion (Kattenberg *et al.*, 1996, Fig 6.13 p.312, Fig 6.17 p.318). Results were also reported of a comparison between the UD model and the Hamburg ECHAM1/LSG O/AGCM results (Raper and Cubasch, 1996). In that comparison the thermal expansion in the former exceeds that of the latter by about 50%. In FY95/96 and in FY96/97, we extended this work by undertaking a comparison between the UD model and some experiments performed with the Hadley Centre O/AGCM (HADCM2) (Mitchell *et al.*, 1995a). Some of these results are described below.

During the course of our comparison studies it became apparent that the value of the radiative forcing for a doubling of CO_2 concentration that we used in our UD model for the IPCC Second Assessment Report, was too big in comparison with, for example, that calculated from the HADCM2 model, after allowing for stratospheric adjustment. This value, denoted ΔQ_{2x}, appears in the calculation of the radiative forcing from CO_2 concentrations as follows

$$\Delta Q(t) = \Delta Q_{2x} \ln(C(t)/C_0)/\ln(2). \qquad (3)$$

We used $4.367 Wm^{-2}$ as recommended by Shine *et al.* (1990) and used subsequently in *all* IPCC calculations, but the preferred value now is about $3.5Wm^{-2}$ (this is the HADCM2 value, but it accords well with other recent GCM results: see Harvey *et al.*, 1997). The corresponding values of $\Delta Q_{2x}/\ln(2)$ are $6.3Wm^{-2}$ and $5.0Wm^{-2}$ respectively. Figure 6 shows the effect of using $\Delta Q_{2x}/\ln(2)=5.0$ rather than the original 6.3 for two forcing scenarios IS95a and IS95c (Kattenberg *et al.*, 1996) and using a climate sensitivity, ΔT_{2x}, of 2.5°C. To put the results in context we also show results for ΔT_{2x}=2.0°C and 1.5°C and note that the effect of the revised $\Delta Q_{2x}/\ln(2)$ is much smaller than the effect of uncertainty in ΔT_{2x}.

First, we note that in percentage terms the effect of the revised $\Delta Q_{2x}/\ln(2)$ on the temperature change, ΔT, is larger over 1765 to 1990 than thereafter. Thus, the consequences for empirical ΔT_{2x} estimates are that they must be larger than those given in Wigley *et al.* (1997 - Appendix A). In terms of future projections, if one were to 'tune' ΔT_{2x} to the past, this would have a much greater effect on future ΔT than the direct ΔQ_{2x} effect shown in Figure 6.

It is clear from Figure 6 that the magnitude of the revised $\Delta Q_{2x}/\ln(2)$ effect depends on the forcing scenario. For changes over 1990-2100, we summarise in Table 2, the effect on forcing, temperature change and sea level of using $\Delta Q_{2x}/\ln(2)=5.0Wm^{-2}$ versus $6.3Wm^{-2}$ for a range of IPCC scenarios. We note that the effects are proportionally much less for temperature and sea level than for forcing. There are a number of factors which bring about this result. For CO_2 forcing alone, the UD model equilibrium warming is independent of $\Delta Q_{2x}/\ln(2)$, but the transient response is affected because the response time is slower for lower $\Delta Q_{2x}/\ln(2)$. Including other greenhouse gases whose net forcing is positive, the effect is to offset the response time effect associated with CO_2 forcing, since their relative forcing is greater for lower $\Delta Q_{2x}/\ln(2)$. For the negative aerosol forcing the opposite is true, so that the proportional effect is smaller for scenarios with constant aerosol forcing beyond 1990 compared to those with increasing aerosols. These factors largely explain why the effect of $\Delta Q_{2x}/\ln(2)$ is so large prior to 1990; i.e., it is mainly because of the strong role that aerosol forcing has prior to 1990 relative to green-house-gases.

Some conclusions that can be drawn from the Table are that the effect of the revised $\Delta Q_{2x}/\ln(2)$ value is greater for the larger forcing scenario IS95e (which has the highest CO_2 to other-gases forcing ratio) and for higher ΔT_{2x}. The ΔT_{2x} effect arises because the influence of a change in response time is larger for more rapid responses. This is also a factor in explaining why the effect of $\Delta Q_{2x}/\ln(2)$ is greatest with IS92e.

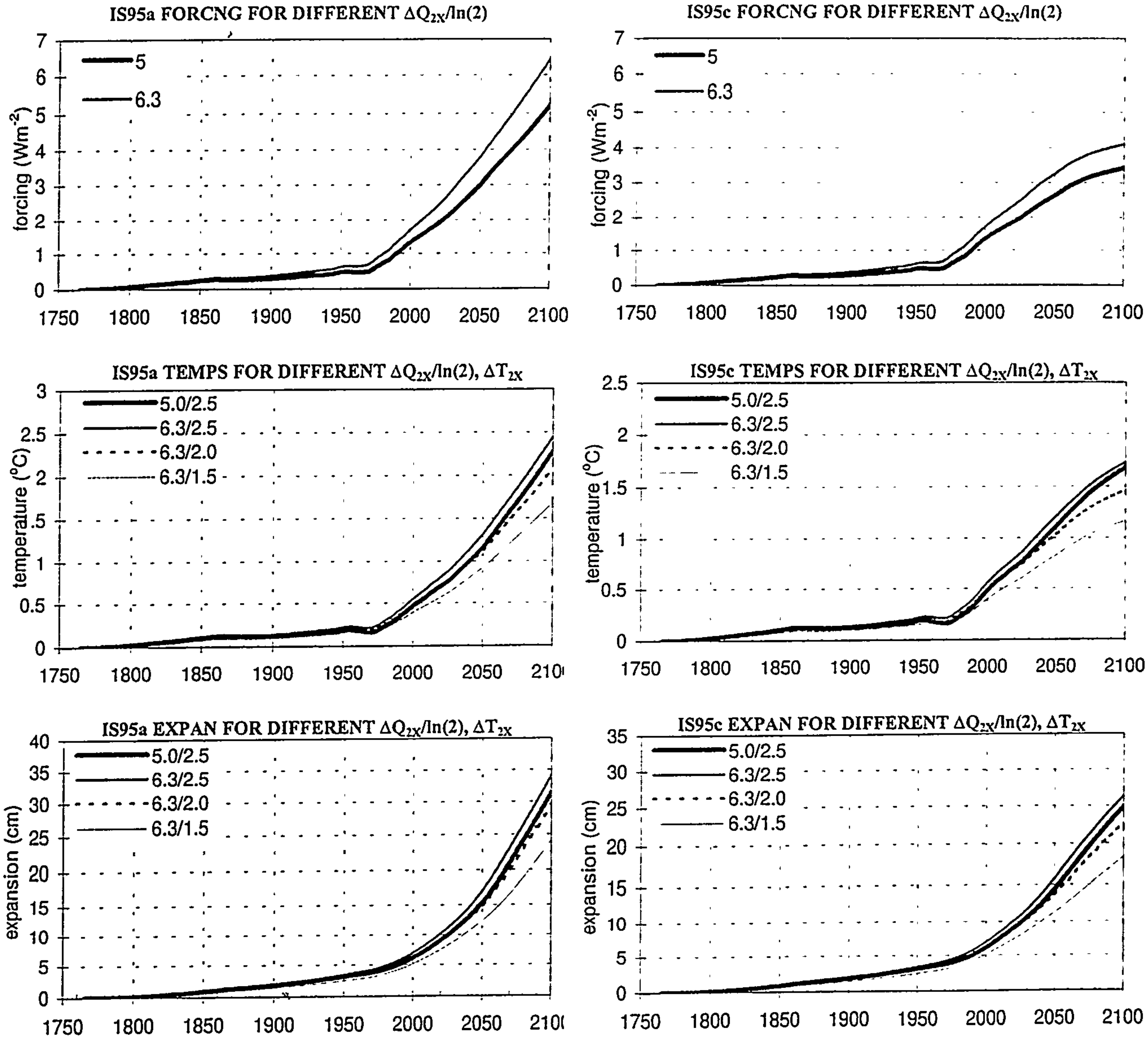


Figure 6: Changes in radiative forcing (top panel), global mean temperature (mid panel) and sea level (bottom panel) for $\Delta Q_{2x}/\ln(2) = 6.3$ and $5.0 Wm^{-2}$ for different ΔT_{2x}. Left hand plots are for IS95a and right hand plots are for IS95c.

Table 2: The effect in terms of percentages of using $\Delta Q_{2x}/\ln(2)=6.3Wm^{-2}$ versus $5.0Wm^{-2}$ on forcing, temperature change and sea level change for a range of forcing scenarios and climate sensitivities (ESO_2 = Emissions of SO_2).

Scenario	forcing results	Temperature change results for $\Delta T_{2x}=1.5°C$	$\Delta T_{2x}=2.5°C$	$\Delta T_{2x}=4.5°C$
IS92a	21.3%	2.2%	4.9%	8.7%
IS92a (const. ESO_2)	18.5%	-1.2%	1.2%	4.7%
IS92c	15.4%	-4.5%	-1.8%	2.1%
IS92c (const. ESO_2)	17.0%	-3.4%	-1.0%	2.8%
IS92e	24.1%	5.7%	8.6%	12.5%
IS92e (const. ESO_2)	19.4%	-0.2%	2.3%	5.7%

Scenario	as above	Sea Level change results for $\Delta T_{2x}=1.5°C$	$\Delta T_{2x}=2.5°C$	$\Delta T_{2x}=4.5°C$
IS92a		5.8%	7.3%	8.2%
IS92a (const. ESO_2)		1.6%	3.6%	5.0%
IS92c		1.4%	4.0%	5.3%
IS92c (const. ESO_2)		1.4%	3.8%	5.1%
IS92e		8.7%	9.8%	10.4%
IS92e (const. ESO_2)		2.2%	3.9%	5.3%

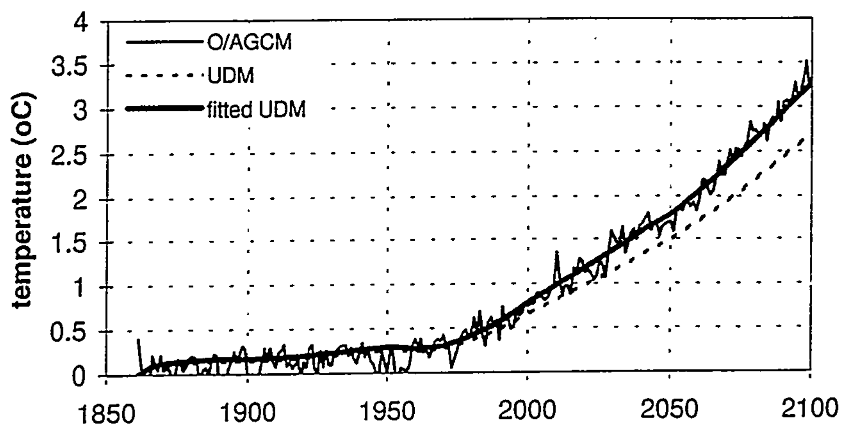
Since the appropriate value of the radiative forcing for a CO_2 doubling in the HADCM2 model is $3.471Wm^{-2}$, this value has been used in the UD model for the comparison experiments. For several forcing experiments we considered two sets of results for the UD model. In the first set, with the exception of the radiative forcing for a CO_2 doubling, the UD model parameter values are the central estimate values used by Raper *et al.* (1996b) and by the IPCC (Kattenberg *et al.*, 1996) – labelled the "UDM" case. In particular, the climate sensitivity, ΔT_{2x}, is 2.5°C (a value apparently consistent with the O/AGCM) and the land/ocean climate sensitivity ratio is 1.3 (see below). For the second set of results, the UD model has been modified by inclusion of a sea-ice parameter and the value of two other parameters has been adjusted to accord with the HADCM2 results; these results are labelled "fitted UDM".

First, we discuss the original "UDM" results compared with the HADCM2 results as shown for the SUL experiment in Figure 7. Note, first, the good agreement between the temperature results for the UD model and the O/AGCM over 1860 to 1995. With this forcing and sensitivity, both model results are also apparently in good agreement with the observed warming.

For the future, the two sets of model results (UDM and O/AGCM) start to diverge, so that by the end of the next century the UD model results give smaller global mean temperature changes and larger thermal expansion than the O/AGCM. Two reasons for this have been identified and can be corrected. First, in the O/AGCM the ocean surface warms faster than the mixed layer because of sea ice retreat. We have now incorporated a simple sea ice parameter into the UD model whereby the surface temperature change is a fixed factor of the mixed layer change (a suitable value for this factor is 1.2). (See Section C5 for the derivation of the sea ice parameterization).

The second reason for the divergence of the results is associated with changes in the upwelling rate. Transient weakening of the thermohaline circulation slows the global mean surface warming and increases the heat flux into the deep ocean as shown by Wigley and Raper (1987) and, with a 2-D model, by Harvey (1994). In the UD model simulations given in Raper *et al.* (1996b), the upwelling rate is assumed to decline linearly with mixed layer temperature increase until a near-zero value is reached. The mixed layer temperature change at which near zero upwelling is reached is a free parameter in the model, assumed in Raper *et al.* (1996b) to be 7°C based on results from the Hamburg model ECHAM1/LSG SCENA (Cubasch *et al.*, 1992) experiment (and in broad agreement with the GFDL results). The 7°C value is also fairly consistent with the HADCM2 1% CO_2 experiment in which the optimum value turns out to be 5.6°C. However, in the HADCM2 GHG and SUL (Mitchell *et al.*, 1995a,b) experiments the decline in the upwelling rate is much smaller and the appropriate value for this parameter is around 20°C. Thus, the assumed upwelling-rate/temperature relationship used in Raper *et al.* (1996b) and by the IPCC (Kattenberg *et al.*, 1996), may not be applicable to all future forcing cases. If UD model and O/AGCM upwelling-rate changes differ, so too will their temperature and expansion results.

As well as reproducing the global mean results, it is also important for the UD model to reproduce well the different warming rates of the land and ocean. For the future, O/AGCMs generally show greater warming over the land than over the ocean. It is, however, not yet possible to verify this affect with the observed record (Nicholls *et al.*, 1996, Fig. 3.3d). As reported in FY94/95, in Raper *et al.* (1996b) the greater warming over the land was represented in the UD model by using a differential land/ocean climate sensitivity (the ratio used was 1.3) (see C5 for derivation). To achieve a good fit to the HADCM2 results a value of the ratio of the land/ocean climate sensitivity of about 1.6 is required. Using this value, together with the sea ice parameter described above and in section C5, and a value of 20°C for the mixed layer temperature change at which the upwelling rate declines to zero, the UD model results give a very good simulation of the global mean and land/ocean O/AGCM results as shown for the global mean in Figure 7. These parameter values also give a very good fit to the HADCM2 GHG experiment results. Thus, with a judicious (but far from arbitrary) choice of parameter values, it is possible to replicate these O/AGCM results with the UD model.



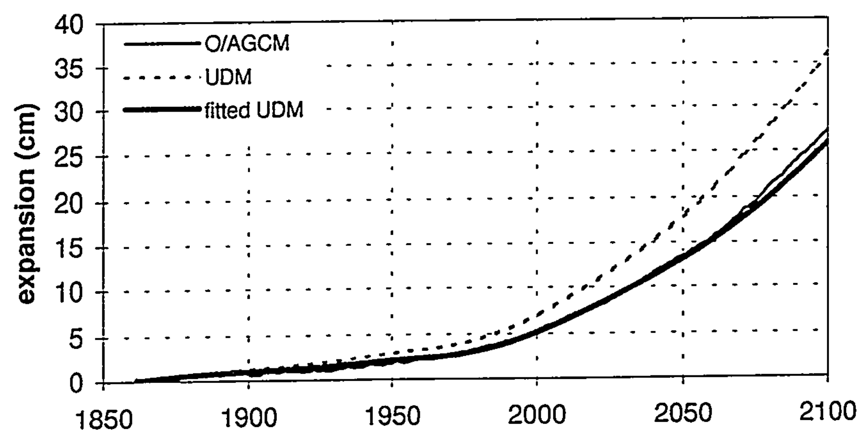


Figure 7: Comparison of results between the Hadley centre O/AGCM, HADCM2 SUL experiment (Mitchell *et al.*, 1995a) and the UD model. In the latter, the appropriate forcing is divided between the four land/ocean boxes. The two sets of UD model (UDM) results are with parameter settings i) as given in Raper *et al.* (1996b) and ii) amended to fit the O/AGCM results (see text).

In view of the recent emphasis on CO_2 concentration stabilisation scenarios (e.g. Wigley *et al.*, 1996; Schimel *et al.*, 1997), it is important to consider model results beyond 2100 and the temperature and sea level commitments implicit in these scenarios. For this purpose we have compared the HADCM2 experiment for which the forcing is a 1% increase in CO_2 concentration until doubling followed by constant $2xCO_2$ concentration. We have looked at the results for 900 years of this experiment (see Figure 8, grey). Using our UD model with the sea ice parameter included and ΔT_{2x}=2.5°C, good agreement for surface temperature is achieved until year 300; but the UD model thermal expansion starts to diverge towards lower values from about year 150.

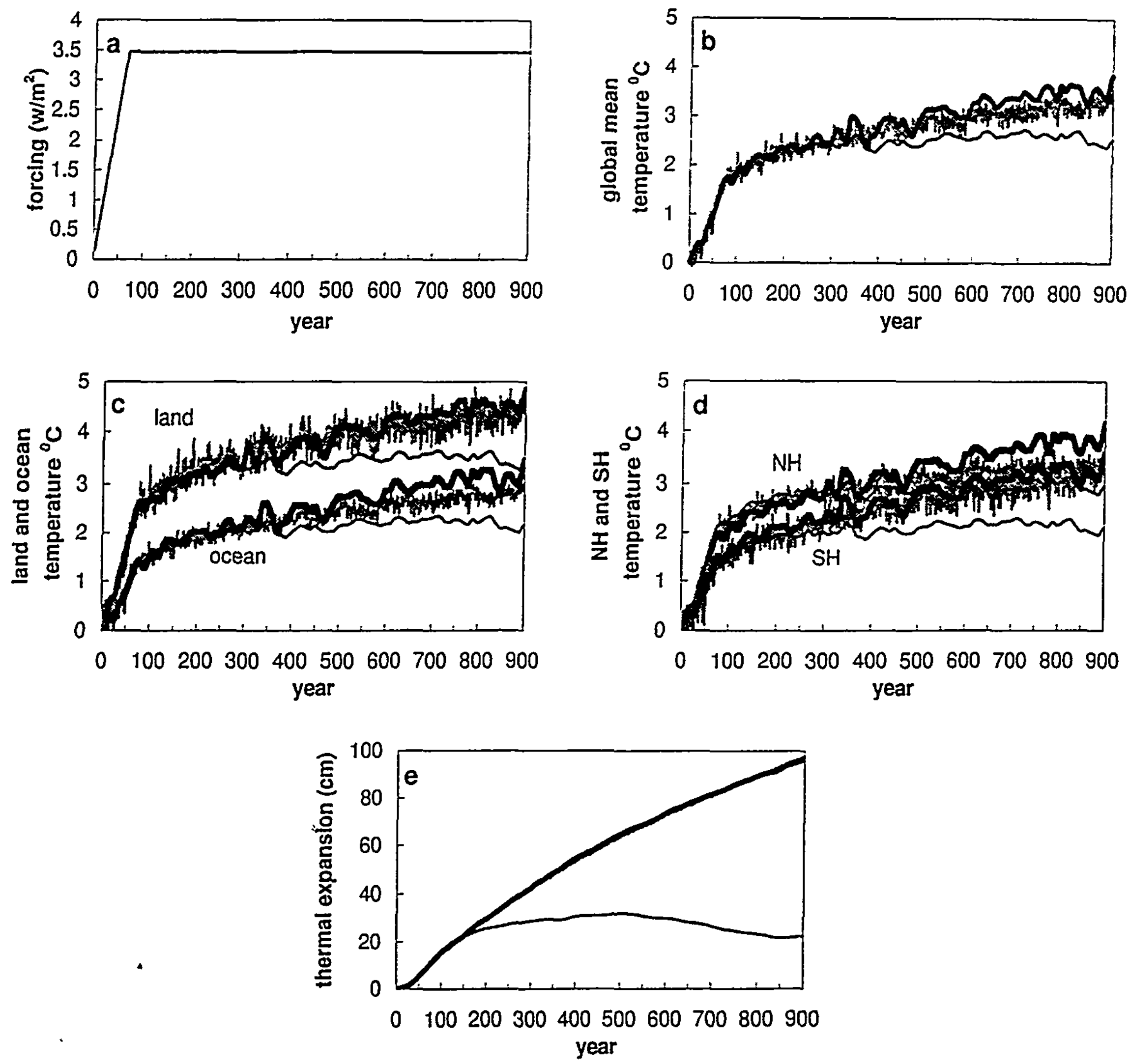


Figure 8: Comparison of results of the HADCM2 O/AGCM (grey) with those of the upwelling diffusion energy balance model with i) sea ice parameter (thin black) and ii) also with time dependant climate sensitivity and effective vertical diffusivities (thick black). Both the latter use time varying upwelling rate as obtained from the O/AGCM. Note the very much lower expansion of the former UD model results.

Unlike the UD model result, the O/AGCM thermal expansion shows no sign of stabilising even after 900 years. We have investigated the reasons for the difference in the behaviour of the models. First we have investigated the behaviour of the O/AGCM climate sensitivity. Assuming the heat capacity of the land is zero, the effective feedback parameter λ, can be estimated from the following equation

$$f_O\, Cd\Delta T_M/dt = \Delta Q - \lambda\Delta T_G - f_O\, \Delta F \qquad (4)$$

where $f_O Cd\Delta T_M/dt$ is the heat flux into the mixed layer and $f_O\Delta F$ is the heat flux into the deeper ocean, f_O being the ocean fraction. The resulting decade mean estimate of the O/AGCM climate sensitivity for the $2xCO_2$ experiment is shown in Figure 9. The increase of the effective climate sensitivity through the experiment can explain the continued surface warming in the O/AGCM compared to the UD model with a fixed climate sensitivity of 2.5^0C. The value of the O/AGCM λ's effective climate sensitivity rises to about 3.8^0C by year 900.

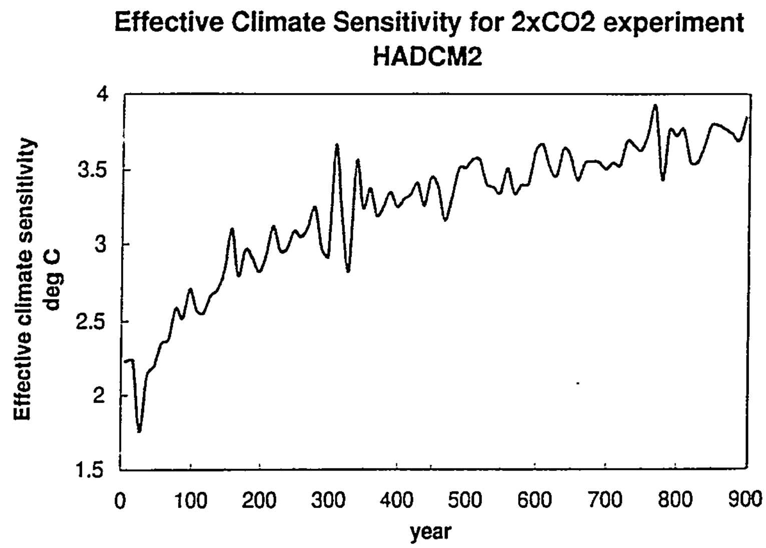


Figure 9: Effective climate sensitivity of the HADCM2 $2xCO_2$ experiment as estimated from decadal mean temperature profiles according to equation (4).

However, this alone does not explain the large thermal expansion result with the O/AGCM. To investigate this further we have calculated the time evolution of the effective diffusivity profiles for the O/AGCM results. In order to reproduce the O/AGCM model thermal expansion result we find that the profile of diffusivities in the UD model must have a maximum value at about 2000m and that the values at around this depth in the ocean must increase with time. With these diagnosed time evolving values of the climate sensitivity and diffusivity the UD model is able to reproduce well the O/AGCM results (thick line in Figure 8). The slight overestimate in the surface temperatures towards the end of the run are probably caused by the disappearance of the sea ice, so that by this time our sea-ice parameterisation becomes invalid. This comparison study highlights the uncertainty in the thermal expansion commitment with, for example, this CO_2 concentration doubling experiment. The realism of this O/AGCM result is unknown. It is, therefore, important to analyse the results of other O/AGCMs.

C5. Generalization of the UD model

In order to facilitate comparisons between our UD model and coupled O/AGCM results (see Section C4.), we have had to implement two features originally built in to the model: time variable upwelling rate and land/ocean differential climate sensitivity. These aspects are important features of O/AGCMs (see, e.g., Murphy, 1995; Manabe and Stouffer, 1993). They must be considered in any comparison, and in any application of the UD model to the production of future climate change projections. The variable upwelling rate option has been used previously (Wigley and Raper, 1987; Jones *et al.*, 1987). However, the differential sensitivity code has not previously been used. The derivation of the differential sensitivity code and the sea ice parameter is described below, followed by a discussion of the Antarctic initial disequilibrium parameter.

Derivation of land/ocean differential climate sensitivity code

The way we originally coded this aspect was in terms of the feedback parameters in the energy balance equations - i.e., each box in the model has a $\lambda_i\Delta T$ term representing the feedback, with the various λ_i specifiable independently. Independent specification of λ_i is, however, very inconvenient. In the present situation, λ_L (land) and λ_O (ocean) need to be input to the model, but the information usually available from GCMs is the *global* sensitivity (ΔT_{2x}) and the corresponding equilibrium warming differential between land and ocean ($R = \Delta T_{eq,L}/\Delta T_{eq,O}$). We have therefore derived analytical relationships between the equivalent pairs (λ_L, λ_O) and (ΔT_{2x},R), which allows the latter to be input directly. The method used is as follows:

The aim is to specify the ratio of equilibrium temperature changes over land versus ocean and the global-mean sensitivity, and calculate λ_O and λ_L for direct input into the UD model.

If changes in upwelling rate are ignored, the energy balance equation for box i (where i = 1, 2, 3, 4 corresponds to NH ocean (NO), NH land (NL), SH ocean (SO) and SH land (SL)) is of the form

$$f_i(C\, d\Delta T_i/dt + \lambda_i\Delta T_i) = f_i\Delta Q_i - f_i\Delta F_i + k(\Delta T_j - \Delta T_i) + k_H(\Delta T_k - \Delta T_i) \qquad (5)$$

Here, f_i is the fraction of the area in box i (sum over i equal to 1), $C = \rho ch$ is the heat capacity and ΔF_i if the change in flux to the deep ocean (both zero for land boxes) and the k terms specify the inter-box exchanges. Subscript H on k identifies the interhemisphere exchange term.

For the present calculations, we only need the steady state result, which is

$$f_i\lambda_i\Delta T_i = f_i\Delta Q_i + k(\Delta T_j - \Delta T_i) + k_H(\Delta T_k - \Delta T_1) \qquad (6)$$

This also holds for the variable upwelling rate case. In this case, ΔF_i is not zero, but it is compensated for by the term involving Δw. If this didn't happen, the global-mean steady-state temperature would be inconsistent with $\lambda\Delta T = \Delta Q$.

We now write equ. (6) in full, assuming the ΔQ_i are the same in each box and noting $f_{NO} + f_{SO} = f_O$ and $f_{NL} + f_{SL} = f_L$:

$$f_{NO}\lambda_O\Delta T_{NO} = f_{NO}\Delta Q + k(\Delta T_{NL} - \Delta T_{NO}) + k_H(\Delta T_{SO} - \Delta T_{NO}) \quad (7)$$
$$f_{NL}\lambda_L\Delta T_{NL} = f_{NL}\Delta Q + k(\Delta T_{NO} - \Delta T_{NL}) \quad (8)$$
$$f_{SO}\lambda_O\Delta T_{SO} = f_{SO}\Delta Q + k(\Delta T_{SL} - \Delta T_{SO}) + k_H(\Delta T_{NO} - \Delta T_{SO}) \quad (9)$$
$$f_{SL}\lambda_L\Delta T_{SL} = f_{SL}\Delta Q + k(\Delta T_{SO} - \Delta T_{SL}) \quad (10)$$

Adding these gives

$$\lambda_O f_O\Delta T_O + \lambda_L f_L\Delta T_L = \Delta Q = \lambda\Delta T = \lambda(f_O\Delta T_O + f_L\Delta T_L) \quad (11)$$

where $f_O = f_{NO} + f_{SO}$, etc., and all temperatures are equilibrium values.

Now denote the ratio $\Delta T_L/\Delta T_O$ by R. Equ. (11) becomes

$$\lambda_O f_O + \lambda_L f_L R = \lambda(f_O + f_L R)$$

Hence

$$\lambda_L = \{\lambda(f_O + f_L R) - \lambda_O f_O\}/f_L R \quad (12)$$

Here both R and λ ($= \Delta Q_{2X}/T_{2X}$) are known. Hence, if λ_O were given, λ_L could be calculated.

The problem now is to determine a consistent value of λ_O. An efficient way to do this is iteratively. First, a trial value is chosen. From this, equ. (12) gives λ_L and then ΔT_{NO}, ΔT_{NL}, ΔT_{SO} and ΔT_{SL} can be calculated from equs. (7)-(10). From these, the land/ocean temperature ratio can be calculated and compared with the input value. Iteration on λ_O until the output and input R values are the same leads to a fully consistent set of values for all quantities.

To solve equs. (7)-(10), we write these in matrix form:

$$\underline{\underline{C}}\,\underline{\Delta T} = \underline{Q} \quad (13)$$

where

$[\Delta T]^T = [\Delta T_{NO}, T_{NL}, \Delta T_{SO}, \Delta T_{SL}]$ (SuperscriptT=transpose)
$[Q]^T = f_{NO}\Delta Q, f_{NL}\Delta Q, f_{SO}\Delta Q, f_{SL}\Delta Q]$

and

$$\underline{\underline{C}} = \begin{bmatrix} f_{NO}\lambda_O + k + k_H & -k & -k_H & 0 \\ -k & f_{NL}\lambda_L + k & 0 & 0 \\ -k_H & 0 & f_{SO}\lambda_O + k + k_H & -k \\ 0 & 0 & -k & f_{SL}\lambda_L + k \end{bmatrix}$$

(Superscript T denotes the transpose.) The solution to equ. (13) is

$$\underline{\Delta T} = \underline{\underline{C}}^{-1}\, \underline{Q} \tag{14}$$

Since $\underline{\underline{C}}$ is symmetric with some convenient zero elements, its inverse is easy to calculate.

The above calculation algorithm has been inserted into our model code. We have tested the algorithm with a stand-alone program, inserting the λ_O and λ_L values directly into the original code, and verified that the correct asymptotic values for the equilibrium global-mean temperature change and land/ocean temperature ratio are obtained.

Derivation of sea ice parameter

As discussed in C4, comparisons between our UD model and O/AGCM results have demonstrated the need for a representation of the effect of sea ice changes in the UD model. For the parameterization of sea ice we derive a relationship between the change in the air temperature over the ocean, ΔT_A, and the change in the mixed layer temperature, ΔT_M, as follows. First, we may write for unit area,

$$T_A = A_I\, T_{AI} + A_O\, T_{AO}, \tag{15}$$

where T_A is the air temperature over the ocean, A_I and A_O are the fractional areas of the sea ice and open water respectively, and T_{AI} and T_{AO} are the air temperatures over the sea ice and open water.

Here, the total area of the ocean is constant so that, $A_I = 1 - A_O$, and $\Delta A_I = -\Delta A_O$. A change in the air temperature over the ocean can then be written (assuming products of anomalies to be negligibly small) as,

$$\Delta T_A = \Delta A_I\, T_{AI} + A_I\, a\, \Delta T_{AO} + \Delta A_O\, T_{AO} + A_O\, \Delta T_{AO} \tag{16}$$

where we have assumed that a change in the air temperature over sea ice is directly proportional to a change in the air temperature over the open ocean (i.e., $\Delta T_{AI} = a\, \Delta T_{AO}$). If we assume that the change in the ice area depends on the change in the ocean temperature, and also that a change in the air temperature over open-water is equal to a change in the mixed layer temperature (i.e., ΔT_{AO}=ΔT_M) then,

$$\Delta A_I = -\alpha \, \Delta T_M,$$

so that (6) becomes,

$$\Delta T_A = \alpha \, \Delta T_M \, (T_{AO} - T_{AI}) + \Delta T_{AO} \, (a \, (1-A_O) + A_O)$$

$$= \Delta T_M \, \{ \, \alpha \, (T_{AO} - T_{AI}) + (a \, (1-A_O) + A_O) \, \} \tag{17}$$

Thus, with the above stated assumptions, the air temperature change over the ocean can regarded as a fixed multiple of the mixed layer temperature change. We have examined data from several O/AGCM experiments (see C4) and find that the assumption of a fixed proportionality is a good approximation and that a value of about 1.2 is appropriate (i.e., $\Delta T_A=1.2\Delta T_M$).

To allow for the sea ice changes in the UD model equations, the mixed layer temperature change in each hemisphere must be multiplied by the proportionality constant, denoted CICE, wherever the air temperature over the ocean is required. This occurs in three contexts: (i) in the feedback terms over the ocean (involving λ_0 CICE ΔT_O), (ii) in the land-ocean exchange terms (involving K(CICE $\Delta T_O - \Delta T_L$), where K is the exchange coefficient), and (iii) in the calculation of the air temperature over the ocean for output.

Effect of new parameters on modelled past sea level rise

Using the new parameter value for the initial disequilibrium state of the Antarctic ice sheet (see section C1), together with the new sea ice parameter and a value for the ratio of the land/ocean climate sensitivity of 1.5 (previously 1.3, the value of 1.6 which fits the HADCM2 results may be too high for fitting other O/AGCM results), but retaining for the time being a value of 7^0C for the surface temperature at which the upwelling velocity is effectively zero, the range of modelled sea level rise to 1990 is 7-17cm (previously 2-19cm) for a 1880-1990 warming of 0.5°C, with a central estimate of around 12.5cm (previously 10cm). These results may be compared with (and are in better agreement with) the observationally based estimate of 10-25cm (Warrick *et al.*, 1996).

C6. Sulphate aerosol effects

We have indefinitely deferred work on forcing an ocean general circulation model with differential hemispheric warming (induced by sulphate aerosol cooling of the Northern Hemisphere). Now that experiments with coupled ocean/atmosphere GCMs are becoming available (e.g., Mitchell *et al.*, 1995a, Meehl *et al.*, 1996, Haywood *et al.*, 1997) the need for such simplified sensitivity experiments is less pressing than when this proposal was first written.

C7. Low-frequency climate variability

In previous proposals and progress reports, we have outlined a hierarchy of coupled model studies aimed at investigating low-frequency internally-generated climate variability. An ocean GCM (OGCM) is included in all levels of the hierarchy (since the oceans are known to be of primary importance on decadal and century timescales), while the representation of the atmosphere increases in complexity at each step up the hierarchy. The first steps of this hierarchy (the ocean coupled to a time-invariant atmosphere, with or without realistic air-sea flux anomalies applied to the ocean) have previously been studied for the Hamburg LSG OGCM by others (e.g., Mikolajewicz and Maier-Reimer, 1990) and by us. Our work has improved overall understanding of the propagating modes that generate the 300-year period variability that this model exhibits, and we have tested the sensitivity of the variability to changes in some of the model's physical and numerical schemes. This work has now been written up in a Ph.D. thesis (Osborn, 1995) and three papers (Osborn, 1996a,b; 1997 - reproduced here as Appendix D).

The third step of the hierarchy involved coupling the LSG OGCM to an empirical atmosphere model [using principal-component-based statistical relationships to compute patterns of air temperature and fresh-water flux for any given pattern of sea surface temperature (SST) anomalies generated by the OGCM]. The statistical relationships were derived from an extended AMIP AGCM simulation (Gates, 1992), and the construction of the model and its coupling to the OGCM are described in Osborn (1995).

Earlier speculation suggested that the 300-year oscillation was a natural mode determined by the characteristics of the Atlantic basin. In conjunction with the work of Pierce *et* al. (1996), we have shown (Osborn, 1995, 1997) that the 300-year period variability of the LSG OGCM is caused by an oscillation between a state with strong convection in the Southern Ocean and a state with little convection there due to the formation of a surface fresh-water (and hence, low density) cap. An example is given in Figure 10a, where the state of the Southern Ocean is characterised by the strength of the Antarctic Circumpolar Current (ACC). When *fresh-water flux feedback* is allowed *via* the empirical atmosphere model, the variability is slightly enhanced (Figure 10c). This is evident by the extended periods of low ACC strength (years 4415 to 4490 and 5920 to 5995 in Figure 10c), caused by a slightly enhanced fresh-water flux (generated by colder SSTs) maintaining the fresh-water cap. When the *air temperature is allowed to vary* instead, also *via* the empirical atmosphere model, the internal variability is greatly reduced (Figure 10d). This reduction is caused by a weakening of the feedback process that generates the variability. When the fresh-water cap reduces convection, the surface should cool (thereby enhancing convection once more) but, when the air temperature is prescribed (and fixed), it is unable to do so by very much. The variable air temperature allows temperature changes to offset some of the effect of salinity changes on the vertical density structure, thus weakening the overall variability (Osborn, 1995).

Our demonstration of the importance of air temperature variations in the generation/suppression of internal ocean variability re-inforces other work (e.g., Mikolajewicz and Maier-Reimer, 1994; Pierce *et al.*, 1996) with this OGCM coupled to simple atmosphere models. The difference here is that the variability is only partially suppressed when the air temperature is allowed to vary, giving a continuum of variability "states" rather than the discrete end-member cases obtained in previous work.

Our study of multi-century variability of the Southern Ocean thermohaline circulation (THC) in the LSG OGCM is now complete. Since first being identified (Mikolajewicz and Maier-Reimer, 1990) this mode of THC variability has been the subject of much research, since, if it were realistic, it would be the largest oscillation in the climate system. We have identified similarities between this mode and those in another Hamburg OGCM (Osborn, 1997) and have presented explanations for differences between the LSG behaviour and GFDL OGCM simulations (Osborn, 1996b). Allowing air temperature to vary weakens the processes that generate this mode, and hence weakens the mode itself. With a sufficiently strong freshwater flux input, however, the mode can still exist, and we suggest (Osborn, 1997) that the poor ice-ocean haline coupling in the LSG OGCM results in an unrealistically strong input of freshwater to the model's southern ocean. When the coupling is improved, the required freshwater input is reduced and the 300-year mode all but disappears (Figure 10b). In carrying out this investigation we have obtained results that can be applied to improving coupled models and explaining the causes of coupled model climate drift (Osborn, 1997).

This work is of considerable significance, because it suggests that thermohaline circulation variability obtained in other GCM experiments may be qualitatively correct, but quantitatively overestimated. There are, of course, other reasons for expecting that modelled THC variations may be too high - see, e.g., McDougall *et al.* (1996) - and it is interesting that our analyses of the Hadley Centre model results (see Section C4) also point to lower THC variability compared with the GFDL and Hamburg models. THC sensitivity to climate change is still an issue of great importance, though (see, e.g., Rahmstorf, 1995; Manabe and Stouffer, 1993).

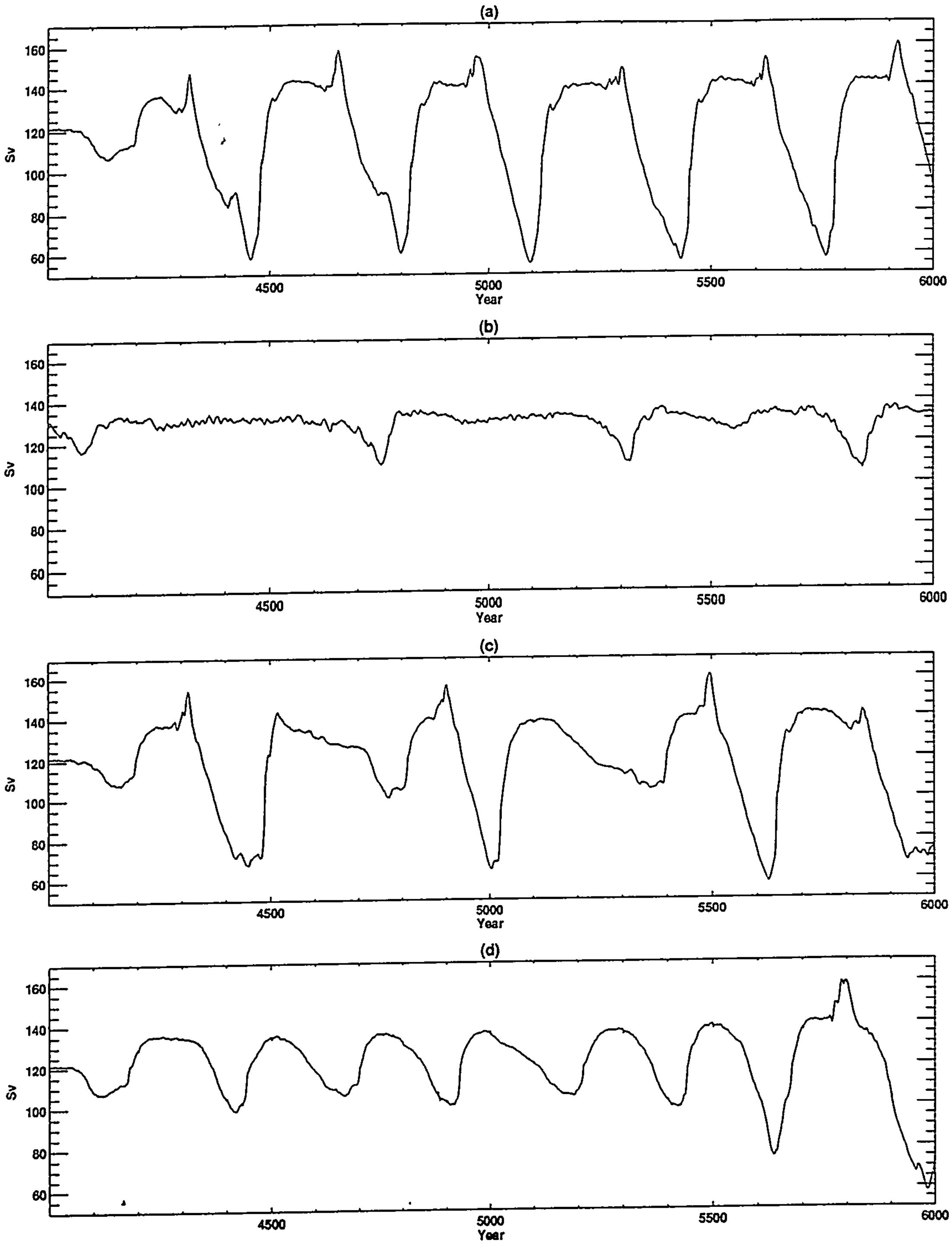


Figure 10: The strength of the Antarctic Circumpolar Current (Sv) from integrations with the LSG OGCM coupled to (a) a time-invariant atmosphere, except for the addition of stochastic fresh-water flux noise; (b) as (a), but with an improved ice-ocean coupling; (c) an empirical atmosphere model with fresh-water fluxes controlled by model SST but prescribed air temperature; (d) an empirical atmosphere model with air temperature controlled by model SST but prescribed fresh-water fluxes.

D. Model Validation

D1. Assessment of interannual and interdecadal variability in GCMs

Two main pieces of analysis have been undertaken in this area. The first (Jones *et al.*, 1997b) had the primary aim of assessing errors in hemispheric and global temperature series (see section A). To estimate two of the parameters required for the error assessment (the average inter-record correlation between stations within a grid box and the effective number of independent samples over the surface of the Earth) we computed temperature correlation decay lengths (Briffa and Jones, 1993, and Jones and Briffa, 1996), fitting an equation of the form

$$r = e^{-x/x_0} \tag{18}$$

to the grid box temperatures. Here, r is the correlation between pairs of grid-box temperatures, x the distance between boxes and x_0 the characteristic decay length. A subsidiary part of this work applied the methodology to three long (800-1000 year) control integrations of GCMs. These were the 1000-year control run of the Geophysical Fluid Dynamics Laboratory (GFDL) model (Manabe and Stouffer, 1996); the 1000-year control run of the latest Hadley Centre model (HADCM2; Mitchell *et al.*, 1995a,b) and the last 800 years of the 1260-year control run of the Hamburg model (ECHAM1/LSG; von Storch 1994). The results are all reported in Jones *et al.* (1997b).

The comparisons assess the ability of each model (in the absence of external forcing) to reproduce the decay of temperature correlation with distance evident in the real world, and how this changes with timescale. (Note that this is not strictly a like-with-like comparison, because the real world climate is clearly affected by both natural and anthropogenic external forcings, whereas the models are not.) All three models exhibit lower mid-latitude correlation decay lengths than in the observed data, with the difference being greater (and extending to higher latitudes) the longer the timescale.

There are two explanations for the above results. First, they could reflect a systematic error in the models. If so, then, from a validation perspective, this would reduce our confidence in using the models to estimate variability on decadal-to-century timescales where the observed data record is too short to give reliable values. The second explanation (as noted above) is that they arise because the model integrations only contain internally-generated variability (generally through the oceanic component) whereas the real world is a combination of internally and externally generated variability. External forcings that may be important include solar, volcanic and anthropogenic components. If the responses to these forcings are highly spatially coherent they will tend to raise the correlation decay lengths when superimposed on the natural or internal variability. The increase will tend to be larger for longer timescales as the response to such signals will be stronger as the timescale increases.

We tested this possibility in two ways. First, we recalculated the correlation decay lengths using the HADCM2 model (Mitchell *et al.*, 1995a) with historical forcing using estimates of greenhouse gas and sulphate aerosol effects. For the second approach, we attempted to subtract the greenhouse gas and sulphate aerosol response from the observed data. The validity of both methods depends on how well the signal component can be modelled. In the first case, this is determined directly by the HADCM2 model. In the second approach, we used a model-based estimate of the signal and its time evolution (Wigley *et al.*, 1998) tuned to give a global-mean warming equal to that observed. This was removed and the residuals were analysed.

The results of this exercise show that the 'global warming' signal (either what is removed from the observations or added to the model during the perturbed integration) is responsible for part of the differences in the correlation decay lengths. The historically-forced HADCM2 integration produces greater correlation decay lengths in the Southern Hemisphere, the increase being greater on the interdecadal timescale. The improvement in the Northern Hemisphere is less, except for an increase at high latitudes on the decadal timescale.

The assessment of the structure of spatial and temporal variability in the control runs of GCMs is important, as it is this variability that is used to assess the significance of model-produced patterns of greenhouse gas and sulphate aerosol responses in the observed data (see Section B and Santer *et al.*, 1995). The only direct means of assessing model-estimated variability on century timescales is with paleoclimatic data. We have explored this area (see e.g. Barnett *et al.*, 1996 and Briffa *et al.*, 1996); but we note that there are still serious problems associated with uncertainties in and lack of representativeness of paleoclimate data, and with the fact that these data combine the effects of internal and external factors, while model data represent only the former.

Our second major analysis involved the evaluation of the HADCM2 simulation of the North Atlantic Oscillation (NAO) during a 1400-yr control integration and during ensembles of integrations perturbed by increased greenhouse gases and sulphate aerosol concentrations (Mitchell et al., 1995a,b). The focus was on the winter (December to March) season, when the NAO and its influence are at their strongest. Spatially, the simulation is found to be very realistic (Figure 11), in terms of the changing storm track, sea level pressure (Figure 11a,b), precipitation and temperature (Figure 11c,d) patterns associated with different phases of the oscillation. Temporally, too, the simulation evaluates well against the observed record (e.g., the Gibraltar minus Iceland pressure difference that we have extended back to the early nineteenth century - see Figure 3 and Jones *et al.*, 1997c), in terms of the interannual and interdecadal variability levels, *provided* the most recent observed changes from the low index 1960s to the high index 1990s (Figure 3) are ignored.

We have shown that this recent observed change is outside the range of natural internally-generated variability that the HADCM2 control integration exhibits. This suggests a possible anthropogenic influence on the atmospheric circulation. Comparison with the perturbed integrations of HADCM2 fails to resolve the issue; the greenhouse-gas-forced ensemble shows no increase in the NAO index from the late 1960s to the early 1990s, although when sulphate aerosols are also included a weak increase is induced over that period. Extending the integrations into the next century shows a decreased pressure gradient between Gibraltar and Iceland in all perturbed integrations, that is statistically significant when the ensemble means are considered.

Despite the generally positive evaluation of the HADCM2's NAO, it has failed to resolve the uncertainty surrounding the recent changes in the NAO index and their link with winter warming of the Northern Hemisphere (see, e.g., Hurrell, 1996). Either the observed changes have been driven by some other external forcing (e.g., solar or volcanic forcing) or the HADCM2 simulation is unrealistic in its levels of natural variability or its response to anthropogenic forcing.

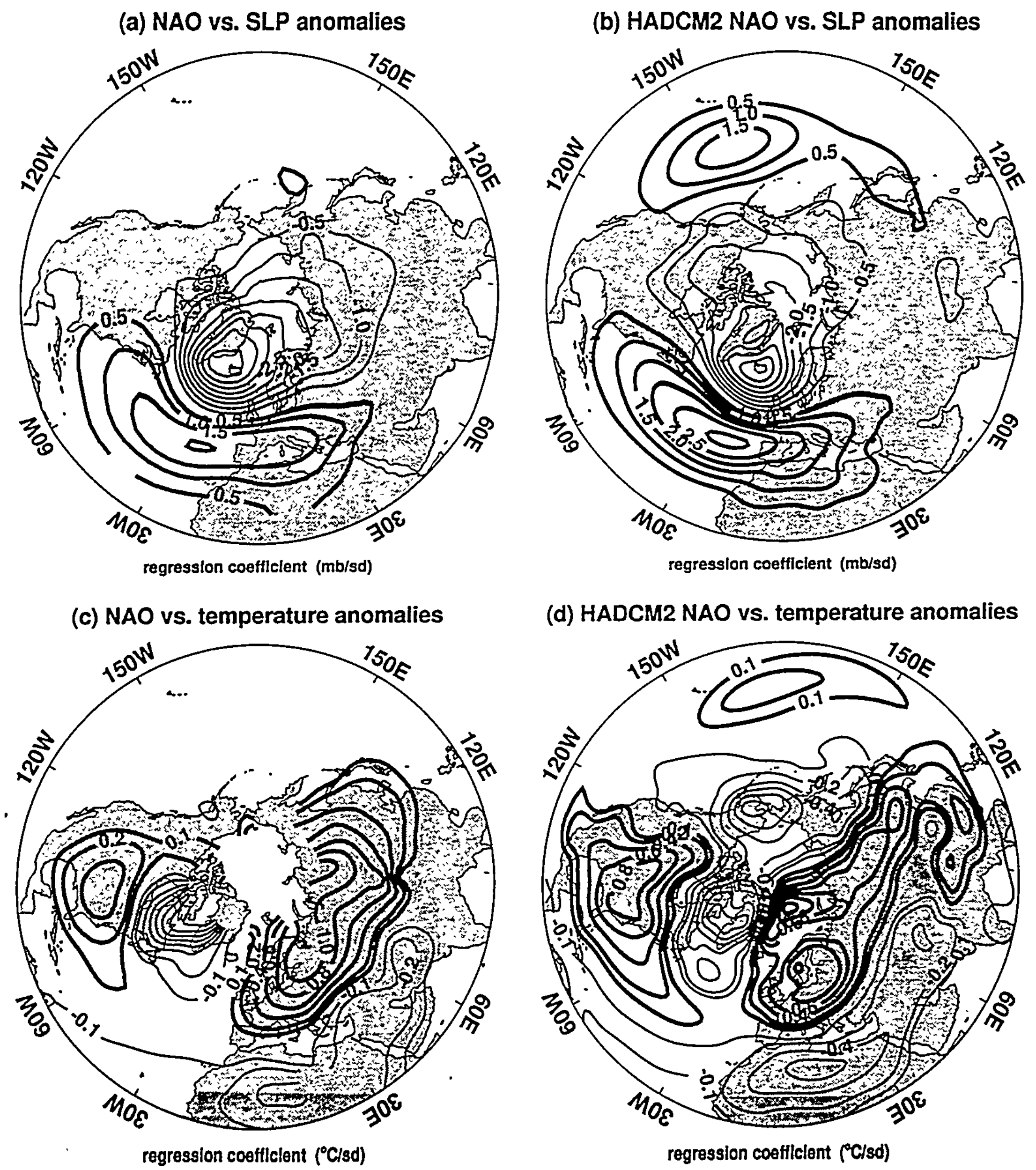


Figure 11: Regression coefficients between **(a,b)** local sea level pressure (mb per sd) and **(c,d)** local near-surface temperature (K per sd) and the normalised NAO index for December to March. Observed results are shown in **(a,c)** and HADCM2 results are shown in **(b,d)**. Positive isolines are thick; isoline intervals are 0.5 mb per sd in (a,b); isoline values are ±(0.1, 0.2, 0.4, 0.6, 0.8, 1 and 1.2) K per sd in (c,d).

D2. Validation of interdaily variability

A major area of our model validation work has been an intercomparison of the reproducibility of precipitation and its daily variability by a range of GCMs. This work is registered as AMIP (Atmospheric Model Intercomparison Project; Gates, 1992) diagnostic subproject 21 ("Surface monthly and daily timescale climatologies and regional climate anómalies") and we have utilised the simulations of twelve models completed for the AMIP project. These models comprise all those that have provided model output on a daily timescale, and include many that have been used for climatological purposes (e.g., BMRC, CSIRO, GFDL, UKMO). The methodology used for the comparison is described in Osborn and Hulme (1997), and its application to model validation has also been completed (Osborn and Hulme, 1998 - reproduced here as Appendix E).

The validation of mean precipitation, on a seasonal basis, is straightforward. The models have been compared with the means obtained from the gridded European climatology of Hulme et al. (1995), and indicate that most simulate the area-averaged mean precipitation well (with a smaller error in winter than in summer). The *pattern* of mean precipitation is simulated better in summer than in winter by ten out of the twelve models. On the basis of both the area-averaged precipitation and its spatial pattern, it appears that the GFDL and UKMO models are superior to the other ten for the European region.

The validation of the daily variability of precipitation is important for the application of model results to climate change impacts studies, but it provides greater difficulties in generating a suitable observed dataset to compare the models with. The daily standard deviation, the number of days with precipitation and the mean intensity on those days all depend on the number of observed stations available for producing an area-mean. The latter number is an essential quantity when comparison with the grid-box GCM output is performed. We have developed relationships (Osborn and Hulme, 1997) that allow the standard deviation and rainday frequency of the mean of many stations to be estimated from the statistics of individual stations and a parameter that describes the spatial character of precipitation events. For the case of standard deviations, the latter parameter is the mean inter-station correlation of pairs of stations within a grid-box. For the case of rainday frequencies, it is the mean inter-station probability that a precipitation event will affect a pair of stations within a grid-box. These relationships allow a more reliable estimate to be made of the daily statistics of grid-box mean precipitation from a limited number of observational stations than could be made otherwise. This is because the key statistical parameters can be reliably estimated from a relatively small number of stations.

These relationships have been used to produce datasets for model validation, using 170 stations over part of western Europe (covering the UK, France, Switzerland, Italy, Benelux and Germany). For the case of raindays, Figure 12 (taken from Osborn and Hulme, 1998) shows the results of the comparison between the twelve model simulations and the validation datasets for winter and summer. Eleven of the models produce too many raindays in winter, whereas in summer five have too many raindays, three too few and four are within the uncertainty level. The pattern of wet-day frequencies over western Europe is stronger (hence better captured) in summer (Figure

12b). Combining pattern and area-averaged values, the root-mean-squared error (Figure 12c) between simulations and observations suggests that higher resolution models perform better (particularly so if the two models developed primarily for numerical weather prediction - DERF and ECMWF - are ignored).

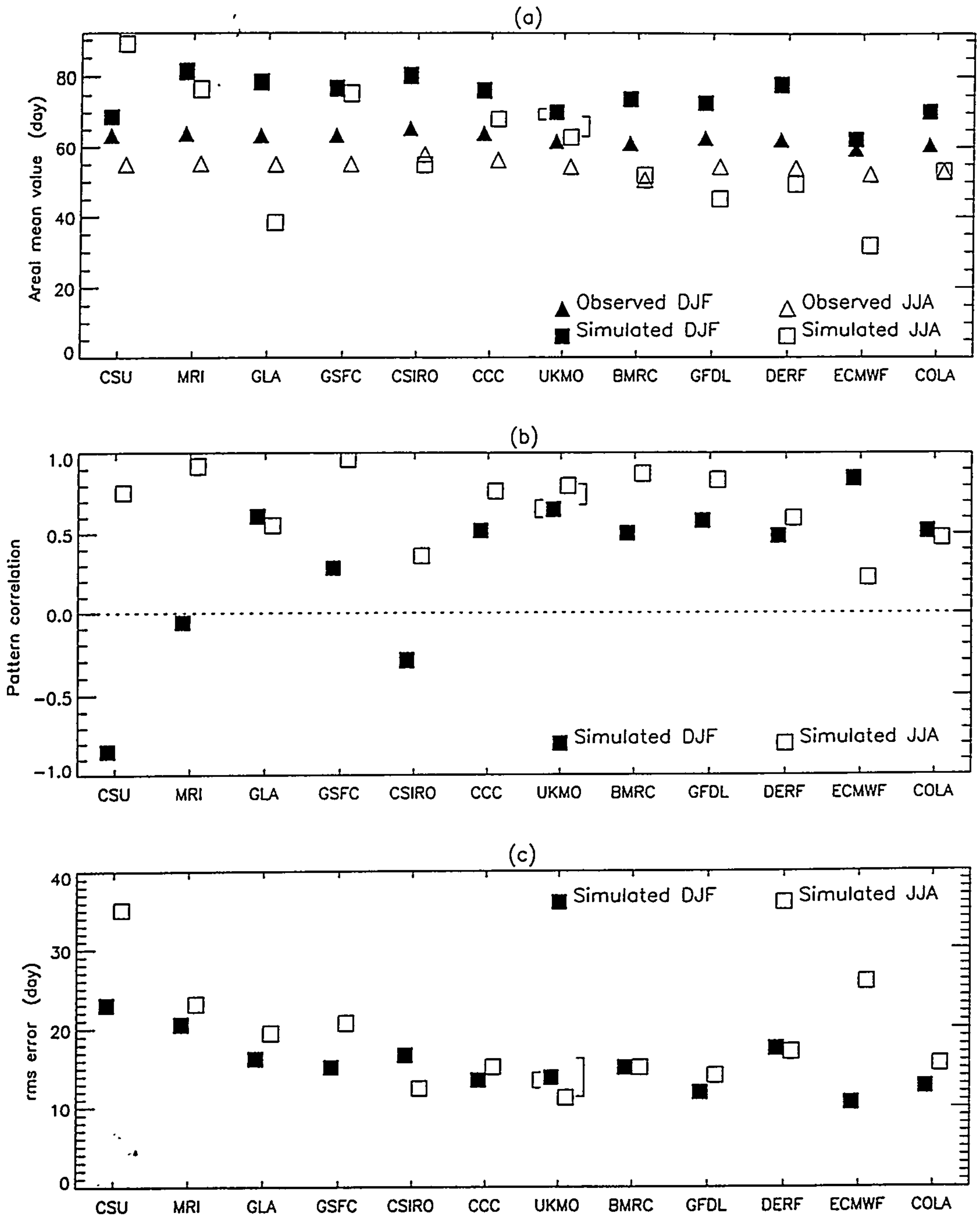


Figure 12: Validation statistics of summer (open symbols) and winter (filled symbols) wet-day frequencies over western Europe : a) area-averaged frequencies observed (triangles) and simulated (squares) ; b) pattern correlations and c) root-mean-squared model errors. Models are shown in order of increasing horizontal resolution (acronyms defined in Appendix E). The uncertainty arising from the short simulations available is indicated for the UKMO model by the square bracket.

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