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REGIONAL FORECASTING WITH
GLOBAL ATMOSPHERIC MODELS

THIRD YEAR REPORT

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FOREWORD

This report was prepared by the Applied Research Corporation (ARC), College Station, Texas, under subcontract to Pacific Northwest Laboratory (PNL) as part of a global climate studies task. The task supports site characterization work required for the selection of a potential high-level nuclear waste repository and is part of the Performance Assessment Scientific Support (PASS) Program at PNL. The work is under the overall direction of the Office of Civilian Radioactive Waste Management (OCRWM), U.S. Department of Energy Headquarters, Washington, D.C.

The scope of the report is to present the results of the third year's work on the atmospheric modeling part of the global climate studies task. The development testing of computer models and initial results are discussed. The appendices contain several studies that provide supporting information and guidance to the modeling work and further details on computer model development. Complete documentation of the models, including user information, will be prepared under separate reports and manuals.

The task work involves the services of two other subcontractors: The University of Maine (UM), Orono, Maine (global ice sheet modeling), and the Lamont-Doherty Geological Observatory (LDGO), Palisades, New York (global climate data base development). The studies by UM and LDGO directly support the atmospheric modeling effort and are reported separately by the respective contractor.

SUMMARY

The Applied Research Corporation (ARC) conducted this study for the Pacific Northwest Laboratory (PNL). The purpose of the study is to make regional forecasts of climate change in order to ensure the long-term security of nuclear waste disposal sites.

In this report we review progress toward implementing a plan for making future forecasts. The plan is based on the recognition of a need to develop a hierarchical climate modeling approach using both simple and complex models. The plan also recognizes that before we can predict future climates with confidence, we must demonstrate some skill in predicting past climates.

At the end of Year Three, the following accomplishments can be noted as follows:

- updated the climate plan originally written in May 1988
- wrote a paper on the overall plan and presented it at the International Conference on High Level Nuclear Waste Disposal in Las Vegas in April 1990
- completed development of a nonlinear energy balance model (NEQMOD) and conducted a number of runs for higher levels of CO₂
- completed a sensitivity comparison of a low-resolution general circulation model (LRM) with the higher-resolution National Center for Atmospheric Research (NCAR) model for climate change over the last 18,000 years
- completed adaptation of the radiation code from the NCAR model to the LRM (this is required in order to make CO₂ runs with the LRM)
- analyzed precipitation fluctuations in the southwest U.S. for the last 100 years in order to determine what features of the large-scale circulation are responsible for precipitation extremes at Yucca Mountain
- developed a statistical technique for forecasting future global ice volume fluctuations for the next 100,000 years

- compared forecasts for a future greenhouse warming with past warm periods in order to determine if there are any analogs from past climates
- compared paleoclimate model simulations of climate change over the last 18,000 years with observations in order to assess the level of confidence we have in the models
- completed quality assurance work on NEQMOD and a number of long auxiliary codes.

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1.0 INTRODUCTION

Applied Research Corporation (ARC) is pleased to submit its third year report on "Regional Forecasting With Global Atmospheric Models", Pacific Northwest Laboratory (PNL) Contract No. 017113-A-B1. We have been involved in a number of activities, and this overview will summarize our work and include selected references to earlier work accomplished during the first two years of the project. Detailed technical discussions are included as separate appendices.

Our work has covered ten main categories: 1) updating and reporting on the climate plan and progress to date, 2) comparison of modeled and observed time series in the southwestern U.S. for the last 300,000 years, 3) development and experimentation with a nonlinear energy balance model in order to conduct a preliminary examination of the effects of very high CO₂ levels, 4) modification and testing of a low-resolution general circulation model (LRM) which will be used in Stage II of our project, 5) analysis of precipitation fluctuations in the southwest U.S. over the last 100 years in order to determine which changes in the large-scale circulation cause precipitation extremes at Yucca Mtn., 6) development of a statistical technique to forecast future global ice volume fluctuations over the next 100,000 years, 7) comparison of a future greenhouse warming with past warm periods in order to search for any geological analogs which might help constrain future forecasts, 8) comparison of paleoclimate simulations with observations in order to assess the level of confidence we have in model predictions, 9) development of auxiliary codes, 10) quality assurance testing of a number of codes. Our summary will discuss these activities in the same order.

2.0 UPDATING AND REPORTING ON CLIMATE PLAN

In March 1988 ARC, at the request of PNL, developed a climate plan for the global modeling project that included the contributions of two companion groups at the University of Maine (UM) and Lamont-Doherty Geological Observatory (LDGO). As work progressed it became evident that it was time to update the plan and report on progress to date. The updated plan is presented in Appendix A, and this section will summarize the original plan, modifications, and progress.

As originally discussed the climate plan was to include three parts: 1) a global climate modeling activity at ARC, 2) an ice sheet modeling activity at UM, and 3) development of an empirically based forecast method at LDGO. The UM group was to present ARC with different ice boundary conditions for global climate model runs. Forecasts for the southwest U.S. would then be compared with statistical studies of the range of past climate fluctuations completed at LDGO in order to have two separate and independent methods of assessing the range of future possible climates.

Because of the large array of possible climates in the future, it was determined that a hierarchical climate modeling approach was necessary for the ARC activities. This involved development and testing of a relatively simple class of models (energy balance models, EBMs) which can prove useful for scoping out the range of parameter space that should be tested with more powerful (and more expensive) higher-resolution general circulation models (GCMs). It was furthermore decided to utilize two different levels of GCMs — a low-resolution general circulation model (LRM; Otto-Bliesner et al. 1982) and a version of the NCAR Community Climate Model (CCM; e.g., Ramanathan et al. 1983). As of May 1988 it was proposed to give the output of the

global climate modeling study to a group involved with development and testing of a fine-scale model useful for regional forecasts. However, the details of that transfer were not worked out as of that time.

As two years have passed since the initial development of the plan it was deemed useful to re-examine it, revise it where needed, and include accomplishments to date. This was originally accomplished in December 1989 but has again been updated for the purposes of this report.

A reassessment of the overall plan indicates that the basic idea is still sound and required only minor course corrections. However, there was considerable scientific progress in a number of areas that helped define the plan more clearly. Plus, there was an important development toward identification of a formal group to do the regional-scale forecasting and coordination of activities between the two climate groups.

The updated plan contains the following substantive additions:

- Completion of the energy balance model stage of development. One of the most useful results to emerge from this study is that comparison of modeled and estimated sea-surface temperatures over the last 18,000 years (Figure 1) yielded very good agreement. This result is important because it gives us the hope that we can estimate with some level of confidence future SSTs — a critical boundary condition for future GCM runs. The result may also be important because it suggests that SSTs can be generated with a relatively simple model. This feature is important from a computational viewpoint, because it is much cheaper to run a high-resolution GCM with prescribed (or fixed) SSTs than it is if they have to be calculated.

Global Annual Average Temp Departures (Ocean)

a) Present CO₂ levels

Global Annual Average Temp Departures (Ocean)

b) 15,18 ka CO₂ levels

FIGURE 1. Comparison of EBM and GCM Estimates of Global Annual Average Temperature Departures over the Ocean for the Past 18,000 Years. Comparison using a) the present value of CO₂ concentration in a linear energy balance model; and b) with decreased CO₂ concentrations at 15 and 18 ka. Over the open ocean, the NCAR-GCM values are the estimates by CLIMAP Project Members (1981). [From Hyde et al. 1989]

- Development of a statistical method to forecast future global ice volume fluctuations (this will be discussed more fully in Section 5.0 of this report and Appendix H). This step is important because it helps us identify the range of a future critical boundary condition for our forecast runs.
- Clarification of the number of runs that will have to be made by ARC. These will involve 2X, 4X, and 8X CO₂ runs, a super-glacial, two intervals of intermediate-magnitude glaciation, and two scenarios that will capture some of the more important conjectured features of time-dependent (i.e., transient) climate change due to steadily increasing levels of CO₂ forcing.
- Identification of the boundary conditions and circulation pattern that may be the most likely candidate for maximum precipitation events that could affect the Yucca Mtn. site (see also Appendix G).
- Near-completion of the UM ice sheet modeling activities, which will provide the ice boundary conditions for the ARC runs.
- Development of a strategy at LDGO for making statistical forecasts of future climate change. This method is based on comparing the course of climate change after a peak interglacial period over several different realizations of climate cycles during the last few hundred thousand years, and then forward-projecting the ensemble range of climate patterns.
- Development of a strategy to evaluate uncertainty in climate model precipitation forecasts. The basic idea is that sufficient preliminary work has been done to indicate that past precipitation-minus-evaporation (P-E) changes, in models and nature, do not change by much more than a factor of two

FIGURE 2. Latitudinal Distributions of the Difference Between the Rate of Precipitation and that of Evaporation (P-E) for Different Levels of CO₂ Change. Note that P-E changes for higher CO₂ levels increase by less than a factor of two in zonal averages (regional changes could be larger, however). Units are in centimeters per day. [From Manabe and Bryan 1985]

or three at most, and changes are usually much less (Figure 2). Preliminary groundwater modeling studies suggest that larger changes than that are required to jeopardize the Yucca Mtn. site (Czarnecki 1985). In other words, the perturbation in P-E required to endanger the Yucca Mtn. site may be greater than anything that might occur with a changed climate. However, this conjecture needs to be tested further and conclusions coordinated closely with hydrologists before any final statement can be made on the subject.

FIGURE 3. Time Lines for Research Activities Discussed in Text. Heavier shading indicates completed work. Dashed lines refer to input from one group to another group's project.

- Coordination of activities with regional-scale modeling efforts. Under the direction of Dr. S. Thompson of NCAR, an intensive effort has now been initiated to develop and make regional-scale forecasts from ARC global-scale output. In conjunction with this effort, ARC will also utilize a version of the NCAR CCM developed by Dr. Thompson in making their global-scale estimates.
- Refinement of timetables and deliverables (Figure 3).

In addition to updating the plan, a general description was written up for the International Conference on High Level Nuclear Waste Disposal in Las Vegas in April 1990 (Appendix B).

3.0 APPLICATION OF CLIMATE MODELS TO REGIONAL FORECASTS

As discussed in the research plan developed in Appendix A, our strategy for making regional forecasts involves utilizing a number of different climate models of varying complexity. In the first year we developed a new climate model, a linear energy balance model (LINEBM) that is efficient and has comparable sensitivity to more complex atmospheric general circulation models (GCMs) in terms of seasonal responses to temperature changes over the last 18,000 years (Hyde et al. 1989).

During the second year we applied LINEBM towards making a regional forecast for the southwest U.S. for the last 309,000 years (Appendix C). We chose this time interval because it is the same as a recently published time series of ground water variations for this region (Winograd et al. 1988). This represents the best-dated time series of climate fluctuations beyond about 20,000 BP (years before present) for the Great Basin. In their paper Winograd et al. also questioned the chronology of climate change that has been independently developed by marine geologists (Imbrie et al. 1984), for the Great Basin record significantly led fluctuations in the marine $\delta^{18}\text{O}$ record (Fig. 4).

Because previous work (e.g., Hays et al 1976, Curry and Crowley 1987) has established that sometimes local records (e.g., Great Basin) can have significant phase offsets from the global ice volume record, we investigated the possibility that the Great Basin record could be some combination of global and local forcing -- for example, global ice volume and local forcing from orbital insolation changes (the Milankovitch effect).

We therefore generated a 309,000 year synthetic time series of temperature fluctuations in the southwestern U.S. As a first step, we assumed that groundwater variations are linearly related

FIGURE 4. Time Series of Great Basin $\delta^{18}\text{O}$ Groundwater Variations Compared to the SPECMAP Chronology of $\delta^{18}\text{O}$ and Global Ice Volume (kbp, thousand years before present). [From Appendix C]

to temperature fluctuations. There are some significant differences between the modeled and observed time series (Figure 5). We attribute these differences to one of three processes: 1) the model is inadequate; 2) the model is adequate but we have not included some important variable; 3) the Great Basin time series is in error.

Because several independent lines of research (see Appendix C) support the original Pleistocene chronology, we have tabled further research on this problem until the chronology of the Great Basin time series is clarified.

FIGURE 5. Residual of Modeled and Great Basin Time Series (dashed). [From Appendix C]

4.0 DEVELOPMENT OF A NONLINEAR ENERGY BALANCE MODEL

In line with the hierarchical modeling approach, ARC spent part of Year Three on developing and testing a new nonlinear energy balance model. This model is a follow-up of a linear model developed in Year One of the project. The nonlinearity involves snow and ice feedback, which varies as temperatures vary. The model has been successfully completed and written up (Appendix D). In general there is a favorable comparison between the nonlinear EBM and observations (e.g., Figure 6) and also other GCMs for altered CO₂ values (Tables D.3 and D.4, Appendix D).

One of the most important features we wanted to determine from the model was the effect of very high CO₂ levels on temperatures at Yucca Mtn. and in the eastern Pacific, upstream of the site. Results (Figures 7 & 8) indicate that very high levels of CO₂ raised temperatures at Yucca Mtn. almost 10°C — a negative feedback in terms of a potential threat to the site because it would increase evaporation at the site. However, very high CO₂ levels also significantly increased SSTs upwind from the site. Thus we have the possibility of significantly enhanced moisture transport into the site. Which of these two factors will dominate remains to be determined, but at this stage there is at least a possibility of moisture levels for very high CO₂ levels exceeding the historical range of variations. This possibility will be an important target of future investigations.

FIGURE 6. Comparison of Modeled and Observed Southern Limits of Snow and Pack Ice in the Northern Hemisphere in January (Observations from Foster 1986 and Kutzbach and Guetter 1986). Agreement is generally good, but deviations follow a consistent pattern: the model ice is too far north over land, and too far south over the ocean.

FIGURE 7. Predicted Increases in Temperature for the Yucca Mtn. Site Due to Various Levels of CO₂.

FIGURE 8. Predicted Increases in SST at an Eastern North Pacific Site (35°N, 135°W) Upwind of Yucca Mtn.

5.0 DEVELOPMENT OF LOW-RESOLUTION GCM (LRM)

At the same time we were working on the nonlinear EBM, a parallel line of activities was directed toward improving the LRM (Low-Resolution GCM) so that we could switch over to the improved model without delay after completing the EBM activities. Although the LRM has been in use for almost a decade, several tasks and upgrades were required. In Appendixes E and F we summarize the following aspects of the LRM:

1. development and improvements of the LRM (Appendix E)
2. comparison of the LRM with the CCM in order to test for relative sensitivity of the models. This comparison is discussed in Appendix E. Overall, there is fairly good agreement, although the LRM is slightly less sensitive in some fields than the CCM. The differences in model response may fall within the range of intermodel differences documented for other GCMs (e.g., Grotch 1988).
3. incorporation of a more advanced radiation package into the LRM in order to conduct experiments with higher CO₂ levels. As of July 1990 the CCM radiation package has been patched into the LRM (Appendix F) and we have just started compiling the model and testing it. We hope to have it available for climate simulations by the end of October 1990.

6.0 ANALYSIS OF HISTORICAL PRECIPITATION FLUCTUATIONS IN THE SOUTHWEST U.S.

Since the main worry of our project is the possibility of significantly enhanced precipitation in the vicinity of Yucca Mtn., we conducted a study of historical precipitation fluctuations to determine the magnitude of past precipitation extremes and circulation patterns associated with them (Appendix G). Results indicate that precipitation in southern Nevada correlates with precipitation in southern California, and that the most extreme years have precipitation increases of 100-200% above the mean (cf. Appendix G, Table G.4b). Most of these extreme events are associated with El Niños — vast warming of SSTs in the eastern equatorial Pacific.

The large precipitation increases for extreme events of the last 100 years appear to fall within the range of increased precipitation for both cool periods of the past 10,000 years and the last ice age. These results therefore give us an upper limit to the natural variability of precipitation in the Southwest. If these were the only factors to consider, then a groundwater modeling study suggests such variations may not be enough to endanger the Yucca Mtn. site (Czarnecki 1985; cf. Appendix A, section A.7.3). However, it is possible that future CO₂ changes could cause a climate change that might exceed the natural variations of the past 20,000 years (see Section 2.0 and Appendix D).

Another important conclusion from the analysis of historical records is that most extreme events are associated with El Niños. These events affect Yucca Mtn. in two ways: 1) higher SSTs represent an increased source of moisture availability for storms entering the southwestern U.S., and 2) perturbations in the

atmospheric circulation associated with El Niños sometimes result in enhanced frequency of storms (e.g., Figure 9).

The above two features will significantly aid us in reconstructing a climate scenario for perhaps the most extreme threat to Yucca Mtn. — one involving very high SSTs from a CO₂ warming with an atmospheric circulation pattern that causes frequent incursions of low pressure systems into the southwest U.S. (Figure 9). This knowledge, in turn, allows us to address a problem which is not presently tractable by brute force methods — namely, the fact that future climate change will involve continuing changes in boundary conditions (CO₂) which may cause climate variations not seen in equilibrium runs with, for example, 2X CO₂ (cf. Stouffer et al. 1989). To address the "transient" CO₂ problem requires very long runs with fully coupled ocean-atmosphere models. Unfortunately, neither we nor anyone else presently has the capability to make long runs that are even moderately believable. By identifying possible boundary conditions which could afford the greatest threat to the Yucca Mtn. site, we can therefore erect a scenario from a plausible coupled ocean-atmosphere realization without actually doing the transient simulation. It is entirely conceivable that a brute force run might never produce such a pattern; our approach avoids that possibility and is more consistent with the need for exploring all possible sources of danger to the site.

FIGURE 9. Schematic Upper-Air Circulation Pattern Associated with the 1982-1983 El Niño. This pattern resulted in one of the most extreme precipitation anomalies associated with the Yucca Mtn. area (see Appendix G, Table G.4).
[Figure from Rasmusson and Wallace 1983]

7.0 FORECASTING FUTURE GLOBAL ICE VOLUME

One of the important boundary conditions for future climate forecasts involves extent and distribution of ice. Although the University of Maine is heavily involved in this activity, ARC had an opportunity to explore a new method for global ice volume fluctuations. In conjunction with Prof. Joseph Newton, Head of the Dept. of Statistics at Texas A&M Univ., we have examined two different linear statistical methods of forecasting future ice variations based on past fluctuations (Appendix H). One advantage of this approach is that it allows us to place probability limits on the chances of climate variations of various magnitude. This paper is now in press at the Journal of Time Series Analysis.

An example of one of the forecast methods (e.g., Figure 10) indicates a relatively small chance for an ice volume event during the next ten thousand years (the time interval dominated by the greenhouse effect). There is an increased likelihood of a small glacial interval about 20,000 years in the future, and a 5% chance of a glaciation 60,000-70,000 years in the future that will equal or exceed the glaciation of the last glacial maximum (18,000 years ago). Comparison of the same forecast with a different forecast based on coupling predicted future insolation changes with a simple ice sheet model (Imbrie and Imbrie 1980) again yielded good agreement (Figure 11).

A note of caution is required in interpreting the future forecasts. All forecasts assume that the system is stationary, i.e., that the boundary conditions for the future will be identical to the past. In fact, we know that a future greenhouse warming will violate stationarity and probably result in ice volume increases smaller than here predicted. Nevertheless, the study is useful as an attempt to determine a baseline for future forecasts.

FIGURE 10. Forecasts and 95% Probability Limits Using Autoregressive Model

COMPARISON OF ICE VOLUME FORECASTS

FIGURE 11. Comparison of Future Ice Volume Forecasts from a Statistical Model (labeled AR, cf. Figure 10, and Appendix H) with Ice Volume Forecasts from a Simple Model.

8.0 SEARCH FOR GEOLOGIC ANALOGS FOR A FUTURE GREENHOUSE WARMING

The future greenhouse perturbation threatens to be a major change in climate. Although we are addressing the consequences of such perturbations in an extended modeling effort, it is also useful to examine geologic records to seek "analogs" for a future warming. If found, such analogs may provide constraints on regional climate variations at Yucca Mtn.

We have undertaken a systematic search for geologic analogs to a greenhouse warming (Appendix I) and come to the surprising conclusion that there are probably in fact no past analogs — the future greenhouse perturbation may represent a unique climate realization (at least in the last billion years). This conclusion is based on the fact that the future greenhouse perturbation is occurring at a geologically very rapid rate (Figure 12). Thus, there will be a significant transient (nonequilibrium) element to the nature of future climate response. Geologic records, on the other hand, represent long time-averaged snapshots that presumably reflect equilibrium responses. Also, past changes in orography and geography should result in regional climate patterns significantly different from the future warming.

Although initially reported on in the 1989 annual report, we have added considerable detail to the analysis of the likelihood of future analogs. The paper has now been accepted for publication by the Journal of Climate.

FIGURE 12. Schematic Comparison of Future Greenhouse Warming with Past Changes in Temperature. Note that Pleistocene climate oscillations are idealized and that interglacial temperatures were probably not significantly warmer than present (see Appendix I). Pre-Pleistocene changes are not well fixed in magnitude, but their relative warmth is approximately correct. Maximum warming in the Cretaceous based on estimates by Barron and colleagues (Thompson and Barron 1981, Barron and Washington 1985). Time intervals in between have been scaled accordingly. [From Crowley 1989]

9.0 UTILIZATION OF PALEOCLIMATE DATA TO VALIDATE CLIMATE MODELS

Part of the climate plan (Appendix A) is to conduct "Hindcast" experiments so that they can be validated against geologic data in order to build a level of confidence in the models. This is a continuation of a long term research program conducted by the authors. In separate studies we have conducted extensive comparisons of climate models with observations, especially for the last 18,000 years but also for warm periods of the Cretaceous, when climates were significantly warmer than present (see Appendix J and Crowley and North 1990).

As summarized in Appendix J, there is only a fair-to-good agreement between models and observations during the last 20,000 years (Figure 13). Fortunately, the agreement is better for the southwestern U.S. when there is a large ice sheet present, and there is also some agreement for this region with summer (but not winter) precipitation changes during the early part of the present Holocene Interglacial (Spaulding 1990). These results suggest that, although climate models clearly have increased our understanding of climate change, sufficient uncertainty remains to require enhanced research efforts.

" Paleo-Modeling Scorecard"	
Region and Time	Model-Data Agreement
N. High-Lat. (18k)	Very Good
Low-Lat. (18k)	Poor
Low-Lat. (9k)	Very Good
S. High-Lat. (18k)	Fair
Ocean Circ. (18k)	Insufficient Work

FIGURE 13. Status of Model Data Agreements for Various Time and Space Slices over the Last 20,000 Years. Evaluation represents the author's judgement, but there is considerable support for each rating. [After Crowley 1989]

10.0 AUXILIARY CODE DEVELOPMENT

During the course of our work it has become necessary to spend a considerable amount of time developing auxiliary computer codes to be used in analysis of model output and comparisons with geological data. Appendix K includes a brief description of the following auxiliary codes: 1) a program for spectral analysis (used in Appendix C); 2) a scientific analysis and file management program. This package consists of one main program with 16 subroutines. It accomplishes a number of tasks (e.g., computation of model output statistics, spectral to grid transformations, Fourier analysis). The file management functions involve such tasks as algebraic manipulation of data sets, editing input files, and transformation of files; 3) a utility code for plotting; and 4) a utility code for making polar projections.

11.0 QUALITY ASSURANCE WORK

QA activities proceeded along several fronts (Appendix L). Code management activities involved ten major codes — one, the nonlinear energy balance model (NEQMOD) and the rest numerous, detailed packages to implement model runs and analysis of output. Data management activities grew to archiving and processing of 33 different data bases comprising almost 460 data sets.

Administration activities included indoctrination and training of two new staff personnel, routine internal surveillances, and hosting a biannual PNL QA audit. A new staff person has been hired as a QA assistant to handle expanded QA activities, especially as we move into Year Four of the project and become involved with numerous large-scale model runs. Handling such a large volume of output may require some modification of QA procedures, and in the early part of Year Four of the project we plan to conduct a detailed study assessing QA problems associated with very large computer codes, such as are associated with GCMs.

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APPENDIX A

GLOBAL CLIMATE STUDIES:

SUBTASK PLAN

APPENDIX A

GLOBAL CLIMATE STUDIES:

SUBTASK PLAN

A.1 GLOBAL CLIMATE MODELING

A.1.1 Objectives and General Strategy

Summary of Project Needs

Nuclear waste is radioactive for tens of thousands of years. In order to assure the long-term security of any nuclear waste storage site, it is necessary to consider the possibility that future climate change could alter the hydrologic regime and mobilize the waste. We must therefore be able to predict future changes in the hydrologic regime. Since local changes in moisture availability are related to the large-scale atmospheric planetary circulation, it is necessary to develop global models in order to make regional predictions.

The purpose of this document is to develop a strategy for regional projections based on global climate modeling in order to meet the requirements of the overall climate plan for the Yucca Mountain Site, as laid out in the Site Characterization Plan (Chapter 5 and Chapter 8, Section 8.3.1.5). The strategy is sufficiently general that it should apply to any modifications in the Site Characterization Plan or even changes in site location. In addition to outlining the research plan, we also discuss research papers.

General Strategy

Regional-scale estimates of climate parameters will be based on both dynamical and empirical techniques. Dynamical methods will utilize global climate models for prediction of input

parameters into regional-scale (mesoscale) models. The output of the mesoscale model is used to make site-specific forecasts (e.g., Yucca Mtn.). Since the seasonality of temperature and precipitation is very strong and region-dependent, seasonal climate models are essential in these studies. Empirical techniques will examine past climate fluctuations as a means of estimating probabilities of future climate fluctuations of different magnitudes. The groups involved in the above tasks (Table A.1) have expertise in climate and ice sheet modeling, and paleoclimatology, and are from the Applied Research Corporation (ARC), Texas A&M University (TAMU), the University of Maine (UM), and Lamont-Doherty Geological Observatory (LDGO). The associated group involved in mesoscale modeling is the National Center for Atmospheric Research (NCAR).

Because climate models in their most advanced form are expensive to operate, (e.g., Figure A.1), we will employ a hierarchical approach to climate modeling. This strategy involves developing and conducting numerous model runs of simple climate models in order to explore the range of model behavior. These pilot studies will help identify the most promising candidates for more detailed (and expensive) climate simulations. The latter will form the basis for making regional-scale estimates of relevant climate parameters.

TABLE A.1. Global Climate Modeling Personnel

Institution	Specialty
Applied Research Corporation	
Gerald R. North (P.I.) ^a	Climate Modeling
Thomas J. Crowley	Paleoclimatology
William T. Hyde	Climate Modeling
Kwang-Yul Kim	Climate Modeling
Bette Otto-Bliesner	Climate Modeling
Matthew K. Howard	Climate Modeling
Neil R. Smith	Quality Assurance
University of Maine	
Terence J. Hughes (P.I.)	Ice Sheet Modeling
James R. Fastook	Ice Sheet Modeling
Lamont-Doherty Geological Observatory	
George J. Kukla (P.I.)	Paleoclimatology
^a Texas A&M University	

INCREASING COMPLEXITY ↓		COMPUTING TIME (CRAY EQUIV. HRS)	COST (\$)
	LINEAR EBM	.0007 (- 2.5 SEC)	0.70
	NONLINEAR EBM	.01	10
	LOW-RESOLUTION GCM (PERPETUAL FORCING)	.4	400
	GCM (PERPETUAL FORCING)	3.5	3,500
	GCM (ANNUAL CYCLE)	70	70,000

FIGURE A.1. Need for a Hierarchical Approach to Climate Forecasts. This figure illustrates that computing costs rapidly increase for complex climate models. In order to realistically explore possible future climate scenarios it is necessary to make many runs with simple climate models and from these runs select a subset for runs with more realistic (and expensive) models. Perpetual forcing refers to simulations that incorporate sea-surface temperatures (SSTs) as input and the model is run for January or July only, but for extended (i.e., perpetual) periods. Annual cycle models calculate SSTs. For additional definitions, see section A.8.

Since climate model simulations have inherent uncertainties, it is important to validate the models against independent data sets. A necessary element of the research program will be to explore how climate models respond to past altered boundary conditions, and compare model output with geologic data. In order to conduct these "hindcast" experiments, the models will require various paleoclimate drivers as boundary conditions (ice, sea-surface temperature, CO₂, seasonal insolation forcing). These drivers will be developed as subtasks on the project. Hindcast validation will be mostly against datasets developed by other subtasks on the project. As we are adopting a hierarchical approach to the modeling endeavor, an additional important step will be to validate simple models against more complicated models in order to determine their relative sensitivities to altered boundary conditions.

Once the various climate models have been tested against past data, the models will be operated in a forecast mode, utilizing altered boundary conditions specified by various team members. Results will be compared with empirical projections based on statistical analyses of past climate fluctuations. The combined dynamical and empirical forecasts, together with their uncertainties, will be analyzed and synthesized. The final report will then be provided to a site-specific climate modeling group (NCAR), which will be funded separately from this study. Figure A.2 illustrates in a generalized diagram how the individual tasks relate to each other and to the project goals; Figure A.3 displays the time lines for different activities that will be discussed in detail below; and Table A.2 lists the milestones, deliverables, and institutional responsibilities in the program.

TABLE A.2. Milestones, Deliverables, and Institutional Responsibilities for This Study.

<u>Date</u>	<u>Deliverables and Milestones</u>	<u>Responsible Organization</u>
Sep. 90	University of Maine submit final report on global ice sheet model	UM
Nov 90	University of Maine Ice sheet data to ARC	UM
Feb 91	ARC provide paleoboundary conditions to NCAR for paleosimulations	ARC
May 91	Lamont-Doherty global paleo-climate data base simulations	LD
July 1, 91	Completion of Low Resolution Model studies	ARC
April 92	ARC Genesis runs delivered to NCAR	ARC
Sept 92	ARC submit final report on global climate model	ARC

FIGURE A.2. Flow Diagram Illustrating General Course of Research Activities for Global Climate Modeling.

A.2 ACTIVITY: GLOBAL CLIMATE MODELING (ARC)

A.2.1 Objectives

The objectives of this activity are to develop and/or utilize climate models for the purposes of providing regional-scale estimates of climate parameters to be used in smaller-scale models.

A.2.2 Parameters

The parameters of this activity are seasonal fields of surface temperature, precipitation, evaporation, winds, averaged solar flux absorbed at surface, averaged solar flux absorbed by

FIGURE A.3. Time Lines for Research Activities Discussed in Text. Heavier shading indicates completed work. Dashed lines refer to input from one group to another group's project.

earth and atmosphere, and averaged long-wave flux at top of atmosphere.

A.2.3 Description

The great majority of contributions to climatic variance come from the large spatial and long temporal scales (Figure A.4). To forecast climate at the regional scale requires understanding and forecasting of the global circulation of the atmosphere. As an example of this effect, consider Figure A.5, which illustrates the modeled atmospheric circulation during the last glacial maximum. Note that the huge Laurentide Ice Sheet caused a splitting of the jet stream. The southern axis of the jet tracks across the southwest U.S. and is associated with enhanced rainfall in the model. These results are consistent with observations from the Great Basin. Other examples of remote effects of climate involve El Niños — warming of equatorial Pacific waters which contribute

VARIANCE OF TEMP FOR DIFFERENT SCALES

FIGURE A.4. The Amount of Climate Variance Explained at a Point as a Result of Contributions from Different Space Scales. Increasing spherical harmonic degrees imply smaller space scales (0 = global average; 10 = 1500 km). This figure illustrates that a substantial percentage of local fluctuations are due to processes operating on very large space scales. [Modified from North et al. 1982]

to interannual variations in rainfall in western North America. For example, El Niños may affect the frequency and intensity of winter low pressure systems moving into the region from the Gulf of Alaska (cf. Enzel et al. 1989). In fact, the years of most intense rainfall in Southern Nevada appear to be associated with El Niños (cf. Appendix G).

When used in conjunction with empirical evidence from the past, the strongest tool for forecasting is the numerical climate model, which is currently being developed and used in similar applications by many scientific laboratories around the world. Our first task is to provide a logical framework for examining the general circulation of the atmosphere as forced by radiation balance changes induced by changing orbital configuration, changes in the atmospheric composition, and lower boundary conditions such

FIGURE A.5. Example of how Regional Climates in the Southwest U.S. are Influenced by Planetary-scale Processes. This figure illustrates NCAR climate model simulations in northern midlatitudes for the last glacial maximum (18,000 yr BP, from Kutzbach and Wright 1985). Winds aloft (stippled arrows), surface winds (double-shafted arrows). Location and highest elevation of the North American, Greenland, and European ice sheets are shown with hatched lines. The dashed lines along the southern jet stream enclose the band of increased precipitation. Note that the huge ice sheet caused a bifurcation of the jet stream, with the southerly branch tracking across the northwest U.S., increasing precipitation. Area-average values of departures (18,000 yr BP-minus-present) of temperature (K), precipitation (%), and precipitation-minus-evaporation (%) are shown for seven areas: Alaska (ALA), northwest USA (NE), Europe (EUR), and the Mediterranean (MED). The values are expressed by the nearest 5K or 5% increments that bracket the area average. For example, in areas NW in January 18,000 yr BP the area-average temperature was 6K below present; the table entry is -5/-10. Note that this is not necessarily the range of temperatures over the area. Underlining indicates that the departure from the model control was significant above the 90% level compared to the model's natural variability.

as positioning and thickness of ice sheets and ocean circulation. All of these forcing functions will be referred to as "drivers."

The main contributions to the variance at a site come from scales larger than the regional scale, and these are forced by the drivers, which can be considered external to the atmospheric system. To calculate the large scale atmospheric behavior, these must be given. In some cases we have a good idea how these agents will change. For example, the orbital properties can be calculated very accurately (Berger 1978). On the other hand, the future concentration of greenhouse gases in the atmosphere is more uncertain. Other considerations such as the positioning and thickness of continental ice sheets, which profoundly affect the heat balance of the atmosphere and its circulation, need to be modeled separately and used as inputs to the atmospheric model. Similarly, oceanic currents are likely to change and these also affect the atmospheric circulation. Plausible scenarios for the drivers need to be included in atmospheric simulations. These plausible scenarios can be obtained by simplified models of the driver systems or by sensitivity studies that eventually include worst-case examples.

Since atmospheric climate models in their most complex form are expensive to operate (Figure A.1), it is necessary to consider a hierarchical approach. The strategy is to use a sequence of numerical climate models that are increasingly complex, with lower members being used for exploratory studies and capable of tentative estimates, but higher members being used only for the later most refined estimates, after worst-case boundary conditions have been roughed out. Some estimates of the regional climate state will be forthcoming with each year of work, but the uncertainties will be large in the first years and smaller and at least better understood in the out-years.

Under separate funding ARC has developed a one level energy-balance climate model (EBM, cf. North et al. 1981, 1983) that is

capable of producing the regional seasonal cycle for given drivers and/or boundary conditions (Figure A.5). The resolution of this model is coarse (it resolves areas about 1/4 the size of the U.S.), and it only produces the seasonal cycle of the surface thermal field. The model, termed LINMOD, although incapable of accounting for nonlinear feedbacks, has the ability to reproduce the present seasonal cycle, and track the NCAR Community Climate Model (CCM) through paleoclimate scenarios (Figure A.6). Its very low cost of operation (Figure A.1), suggest that this tool is the first to use in roughing out the response to a range of driver scenarios. The model has been applied to a number of significant paleoclimate problems (e.g. North et al. 1983; Crowley et al. 1986, 1987). During the first 2-1/2 years of the project, modeling activities included development of more advanced low-resolution EBMs to account for limitations in LINMOD and initial use of more complex climate models that produce atmospheric fields such as precipitation, evaporation, and winds.

A.2.4 Development of Energy Balance Model (EBM) with Regionally-Varying Albedo (LINEBM)

Although LINMOD does a respectable job of simulating the seasonal cycle of temperatures, it was not set up to account for regional variations in albedo due to ice sheets and sea ice extent, and it does an unsatisfactory job handling sea ice. For this reason it was necessary to modify LINMOD to include these important features. During the first phase of ARC's contract with PNL, this model, termed LINEBM, has now been developed and tested (see Section A.4.3).

A.2.5 Development of a Nonlinear EBM (NEOMOD)

Experience with LINEBM suggests that if the albedo of snow and ice is specified, the model does a very good job of calculating the seasonal temperature field at the earth's surface (see HINDCASTS-I section). However, the model does not predict

FIGURE A.6. Comparison with Observations of the Modeled Present Seasonal Cycle of Temperature as Simulated with an Energy Balance Model. On the left is given the amplitude of the annual harmonic ($^{\circ}\text{C}$) in the observed annual cycle of surface temperature. Climatological data were smoothed by a spatial filter which preserves features having a length scale larger than about 1800 km. (For further comparisons including phase lags, higher harmonics, and polar projections, see North et al. 1983.) The graphic on the right gives the same data as the graphic on the left, but for the model. [From North et al. 1983]

seasonally varying snow and ice lines, and it is important to consider these variables when simulating the climate for higher levels of carbon dioxide. Experience with these models suggests that inclusion of interactive snow-ice feedback can increase model sensitivity by as much as 30 to 50%. Hence, the inclusion of this kind of mechanism is potentially very significant.

In order to correct the latter defect and increase model capabilities, a new EBM has been developed that calculates snow and sea ice fields dynamically and treats their physics more realistically. This nonlinear model (NEQMOD) represents a significant increase in model complexity (cf. Figure A.1). The return is potentially very high, because of its ability to test the sensitivity of model solutions to certain nonlinear features inherent in modeling snow and sea-ice feedback (cf. Appendix D).

A.2.6 Simulations with Higher-Level Climate Models

The various versions of the EBMs discussed above should do an acceptable job of generating the seasonal patterns of temperature fields, which are important for influencing atmospheric circulation fields. After the ensemble of runs have been generated for a range of different scenarios, a subset of scenarios will be selected as candidates for general circulation model (GCM) runs. Because GCMs are expensive to run, we hope to include an intermediate stage of simulations with a lower-resolution GCM (LRM).

A candidate intermediate model has been produced at the University of Wisconsin (Kutzbach and Otto-Bliesner 1982). This model type incorporates atmospheric flows and even attempts to simulate the hydrological cycle in a crude form. It has about the same horizontal resolution as LINEBM and NEQMOD but has five atmospheric layers in the vertical. This model is the lowest member of the proposed hierarchy that can produce reasonable winds and moisture products. The cost of running this model is also

intermediate between EBMs and GCMs such as the NCAR CCM (Figure A.1).

Since this project was initiated (August, 1987), ARC has engaged Dr. Bette Otto-Bliesner as a subcontractor on the project (October, 1988). Dr. Otto-Bliesner developed the LRM for her Ph.D. thesis and is very well qualified to work on this project. In the past two years, Dr. Otto-Bliesner has been improving and modifying the LRM code in preparation for simulations with altered boundary conditions. She is also validating her model against a more complex general circulation model (Kutzbach and Guetter 1986).

The model being contemplated for higher-resolution studies is a version of the NCAR CCM. Because of its prohibitive cost as indicated in Figure A.1, only limited experimentation can be envisioned for this model class. Hence, candidate runs must be thoroughly planned in advance through experimentation with lower members of the hierarchy. Under separate funding this model is being developed at the National Center for Atmospheric Research (NCAR).

A.3 ACTIVITY: BOUNDARY CONDITIONS FOR CLIMATE MODEL SIMULATIONS OF PAST CLIMATES (ARC and UM)

A.3.1 Objective

The objectives of this activity are to identify the parameters that will be used to force climate model simulations.

A.3.2 Parameters

The parameters for this activity are:

1. seasonal insolation at the top of the atmosphere
2. atmospheric greenhouse gas composition
3. ice volume and placement

4. sea-surface temperature (SST).

A.3.3 Description

Seasonal insolation

The seasonal distribution of energy from the sun drives the atmospheric circulation. Past changes in the insolation field, due to gravitational perturbation of the planets, has significantly modified past climates (Hays et al. 1976). The past and future insolation regime can be calculated to a relatively high degree of accuracy and is available as an acquired code (e.g., Berger 1978).

CO₂

There have been natural variations in the CO₂ content of the atmosphere. Although the mechanisms responsible for such fluctuations are not well understood, the record of atmospheric CO₂ fluctuations is available from ice cores (Barnola et al. 1987) and can be explicitly specified for hindcast experiments. For hindcast experiments on older intervals that do not have direct CO₂ measurements, reasonable estimates can be made based on the similarity of CO₂ and ice volume fluctuations over the last 130,000 years (the detailed ice volume record has been extended to 2.5 million years ago).

Ice Volume and Area

The amount and areal distribution of ice strongly affects the atmospheric circulation and is a necessary boundary condition for hindcast experiments. These parameters are reasonably well known for the last 18,000 years (CLIMAP 1981, Kutzbach and Guetter 1986) and are available as an acquired data set. For hindcast experiments on intervals prior to the last glacial maximum, the global ice volume is well known (Imbrie et al. 1984), but its regional extent less so. However, members of the research team at all three institutions are qualified to develop reasonable

scenarios for ice volume and area. Based on results from Hindcast experiment I (HINDCAST-I, see below), investigators will identify additional time intervals to simulate and provide appropriate ice boundary conditions. Because of uncertainties, more than one set of boundary conditions may have to be stipulated for ice volume and area for different time intervals.

Sea-surface temperatures

SSTs play a key role in forcing the atmospheric circulation, and the ocean is the ultimate source of water vapor for precipitation on land. Reasonable estimates of SSTs are available for the last 18,000 years (CLIMAP 1981). A more selected set of time series is available for older time intervals (e.g., Imbrie et al. 1989). For the latter (HINDCAST-II) experiments (see below), geographic variations in SSTs can be estimated based upon records from selected sites. Alternatively, SSTs can be calculated from a model. Experience gained in the HINDCAST-I experiments may justify this approach (see below). If SSTs are prescribed, team members from ARC and LDGO will assume responsibility.

A.4 ACTIVITY: HINDCAST EXPERIMENTS (ARC)

A.4.1 Objectives

The objectives of this activity are to force climate models with altered boundary conditions and validate results with independent geological data. These experiments will enable us to evaluate levels of confidence for future model forecasts.

A.4.2 Parameters

The parameters of this activity are:

1. boundary conditions of ice area and volume, SSTs, CO₂, and seasonal insolation for the last 18,000 years and other older, selected time intervals

2. model outputs of a) seasonal temperature field for last 18,000 years and b) seasonal temperature and/or precipitation minus evaporation for older, selected time intervals.

A.4.3 Description

Introduction

Model simulations will proceed in three stages, herein designated HINDCAST-I, HINDCAST-II, and HINDCAST-III experiments. The HINDCAST-I experiments are designed to determine how well the hierarchical approach can be applied to solution of the program goals. HINDCAST-I is a key experiment, for it compares EBMs and GCMs at key intervals over the last 18,000 years, a time interval for which the boundary conditions are well known (Figure A.7). An extensive amount of GCM work has already been done for climate simulations of the last 18,000 years (e.g., Gates 1977, Hansen et al. 1984, Kutzbach and Wright 1985, Manabe and Broccoli 1985, Kutzbach and Guetter 1986), and there is reasonably good agreement between models and observations over this time interval (e.g., Crowley and North 1990).

Our first task is to test whether EBM-generated temperature fields are comparable to GCM-generated fields over this interval. If so, we will proceed to the HINDCAST-II experiments and identify other older time intervals for which we will simulate the temperature fields with EBMs. Based on the set of all experiments done from the HINDCAST-II set, we will then select a subset to be run with higher-level models (e.g., low-resolution GCMs) in order to generate regional precipitation fields (HINDCAST-III experiments). Lastly, regional temperature and precipitation fields will be compared to geological data for the purposes of model validation.

HINDCAST-I Experiments

Development, documentation, and reporting of LINEBM has now been completed and results have been published in Journal of

FIGURE A.7. Schematic Diagram of Major Changes Since 18,000 BP in External Forcing [Northern Hemisphere Solar Radiation in June-August (S_{JJA}) and December-February (S_{DJF}) as Percent Difference from Present] and Internal Boundary Conditions: land ice (ICE) as percent of 18,000 yr BP ice volume; global mean-annual sea-surface temperature (SST), including calculated surface temperature over sea ice, as departure from present, °K; excess glacial-age aerosol (AEROSOL), arbitrary scale; and atmospheric CO₂ (CO₂), in ppmv. The arrows correspond to the seven sets of simulation experiments with CCM. [From Kutzbach and Guetter 1986]

Climate (Hyde et al. 1989). Comparison of LINEBM and GCM generated temperature fields for the last 18,000 years are complete and results are very promising. Results support our working hypothesis that simple climate models capture important features of the seasonal cycle of temperatures. For example, comparison of EBM and GCM generated surface temperature fields for the last 18,000 years (Figure A.8) indicates very good agreement in the northern hemisphere (Hyde et al. 1989). Results are not as good in the southern hemisphere (not shown), but the shortcomings are understood and may be correctable with more advanced versions of the models.

The results from the HINDCAST-I experiment justify model experiments in the HINDCAST-II mode (see below). The results also have important implications for specifications of SST fields, a very important boundary condition from GCM simulations. In the GCM (Figure A.8), SSTs were prescribed for CLIMAP (1981) estimates. In the EBM, SSTs were calculated based on the radiation balance, which changed principally due to the higher albedo of the ice sheets. This result suggests that SSTs may be calculated from first principles — a conclusion that has important implications for specification of SST fields for both HINDCAST and FORECAST experiments.

HINDCAST-II Experiments

The second stage of modeling involves time intervals older than 18,000 years (cf. Figures A.9 and A.10 for Pleistocene terms and trends). Two approaches are planned: 1) utilizing EBMs to generate long time series (several hundred thousand years) of temperature that can be compared with other proxy time series in the Great Basin, and 2) LRM experiments for older time periods, such as the last major interglacial-glacial transition (70,000 BP), interglacial climates at 80,000 and 105,000 BP, glacial

Northern Hemisphere July Land Temp Anomalies

Global Annual Average Temp Departures (Ocean)

FIGURE A.8. Comparison of EBM and GCM Simulations over the Last 18,000 Years for (top) Northern Hemisphere Land in July, and (bottom) Global Ocean Temperatures. Note the comparable sensitivities of the models to seasonal changes in forcing. GCM runs utilized prescribed SSTs, whereas they were calculated in the EBM. The good agreement suggests that we may be able to constrain SST values for time intervals where they are not known. [Figure from Hyde et al. 1989, GCM results from Kutzbach and Guetter 1986]

FIGURE A.9. Some key Late Cenozoic Terms, With Hominid Evolutionary Events Included for Reference. Magnetic epoch and event boundaries are from Mankinen and Dalrymple (1979). The curve on the right is a high-resolution $\delta^{18}O$ record for the last 150,000 years from a core in the eastern tropical North Atlantic (from Shackleton 1977). $\delta^{18}O$ stages are from Emiliani (1955); substages are from Shackleton (1969). Boundaries at 11,000 and 128,000 years BP are called terminations I and II (Broecker and van Donk 1970). The interval between these two terminations is known as the last glacial cycle. Evolutionary and migratory events for man (some of which are overlapping) are from Patrusky (1980), Rensberger (1980), Bischoff and Rosenbauer (1981), Guidon and Delebrías (1986), and Valladas et al. (1988). For reference, some stages in man's cultural evolution are Paleolithic [Acheulian (1.5 - ~0.15 Ma), Mousterian (~40 - 120 Ka), Upper Paleolithic (~10 - 40 Ka)], and Neolithic [~5 - 10 Ka]. About 5000 BP the age of metals began in Mesopotamia (based on information from McAlester (1977), Bye et al. (1987), and Bischoff et al. (1988). [After Crowley 1983]

FIGURE A.10. Composite Record of $\delta^{18}\text{O}$ Fluctuations for the Last 2.5 Million Years, based on averaging measurements from three hydraulic piston cores taken by the Ocean Drilling Program. [Based on data in Raymo et al. 1990]

inception interval (115,000 BP), full-interglacial climate at 120,000 BP, and the penultimate glacial maximum (150,000 BP).

Model output of temperature fields for these simulations can be validated from a number of key paleoclimatic regions: lake and groundwater records from the western U.S., selected sites in the world ocean, Vostok and Greenland ice cores, pollen cores from France, Spain, Macedonia, and Columbia, and loess records from China and central Europe. Validation will be done at ARC, with assistance from LDGO.

Some progress has been made on the HINDCAST II-1 experiments. A 309,000 year series of temperature fluctuations in the Great Basin has been generated with LINEBM and compared to a groundwater

FIGURE A.11. Comparison of Modeled Time Series for the Great Basin with Observed Time Series from Winograd et al. 1988 (kbp, thousand years before present). [From Appendix C]

$\delta^{18}\text{O}$ time series from the same region (Figure A.11, cf. Appendix C). Results indicate significant discrepancies between the two time series. There are several possible explanations for the discrepancy: temperature fluctuations are not linearly correlated with groundwater fluctuations, additional variables need to be considered for the modeled time series (e.g., Pacific sea surface temperatures), or the chronology of the Great Basin time series could be in error. Since the original publication of the Winograd et al. report (1988), several studies suggest that the chronology explanation is a distinct possibility (Kukla et al. 1988, van den Bogaard et al. 1989, Li et al. 1989, Petit et al. 1990). For that

reason we have tabled pursuit of this matter until the chronology question is clarified.

HINDCAST-III Experiments

Based on the full array of results from the HINDCAST-II experiments, the research team will select the most promising candidates for high-resolution GCM experiments. This subset will then be run in order to generate temperature and precipitation-minus-evaporation (P-E) fields. The team will then validate results against the above-mentioned long records, with special emphasis on results for the southwestern U.S. Experience with climate model simulations at 18,000 BP (Figure A.5) indicate that the altered circulation resulted in increased precipitation in the southwestern U.S. at 18,000 BP, a result in accord with paleoclimate observations of higher lake levels at this time (Smith and Street-Perrott 1983). This result is especially noteworthy, since most other parts of the earth were dry during the last glacial maximum (Street-Perrott and Harrison 1985). It will be necessary to coordinate the validation part of this activity with other members of the Yucca Mountain research group working on regional-scale reconstructions for the southern Great Basin.

Because of the relatively large number of simulations being examined for both LRM and high-resolution GCM consideration, it may be necessary to take advantage of one of the results of the EBM-GCM comparison. If SSTs can be generated from a lower-resolution model (Figure A.8), then we may consider generating SSTs with the LRM and running the high-resolution GCMs with these prescribed SSTs. Such an approach would result in a substantial savings in computer time.

A.5 ACTIVITY: BOUNDARY CONDITIONS FOR FORECAST EXPERIMENTS (ARC, UM)

A.5.1 Objectives

The objectives of this activity are to identify the parameters that will be used to force climate model simulations of future climate. In addition to scenarios based on various activities described below, we will also conduct a "worst case" simulation for a "super-glacial."

A.5.2 Parameters

The parameters for this activity are boundary conditions for different scenarios:

1. seasonal insolation at the top of the atmosphere
2. atmospheric CO₂ composition
3. ice volume and extent
4. sea-surface temperatures.

A.5.3 Description

General Strategy. A range of different scenarios will be constructed based on probabilistic and numerical methods in order to identify the possible combinations of insolation, CO₂, ice volume and extent, and SSTs that will be used to force a hierarchy of climate models. Because the range, scope, and precision required of the forecast scenarios exceeds anything attempted before in climate projections, explicit commitment to one approach is unwise at the present time. It is more appropriate to propose a provisional approach and leave the matter open to revisions, based on experience gained during the first years of this project.

The initial approaches adopted for the forecast scenarios involves the tentative conclusion from the HINDCAST experiments (Section 4) that if the ice and CO₂ fields can be specified, then

SAMPLE DECISION MATRIX OF FORECAST SCENARIOS FOR FUTURE CLIMATE CHANGE

FIGURE A.12. Sample Decision Matrix for Different Future Scenarios of CO₂ and Ice Extent, which Will Be Used to Calculate Surface Temperature Fields, which Will then Be Stipulated as Boundary Conditions for GCM-generated Precipitation Forecasts. This figure is an example only and is not meant to imply actual probabilities. Solid blocks imply least probable scenarios; open squares set of candidate scenarios; "partially filled" squares subset of most probable scenarios. Additionally, "super-glacial" and "super-interglacial" scenarios will be simulated. These scenarios may fall outside the umbrella of "most probable" events, but will be included for completeness.

SST can be calculated from radiation-balance considerations (Figure A.8b). Since transport feedback between the ocean circulation and sea ice may modify the latter field (Spelman and Manabe 1984), a range of scenarios may need to be considered. Ice sheets will be dealt with separately (see below).

Initially, we will construct a matrix of possible CO₂/Ice scenarios (e.g., Figure A.12), eliminating only the few subsets that are deemed most unlikely on pure physical grounds (e.g., low CO₂/low ice; high CO₂/high ice). From the set of remaining scenarios, four different methods will be utilized to decide on the subset of most probable driver scenarios: 1) statistical forecasts based on past ice volume fluctuations, 2) one-dimensional models of global ice volume based on past fluctuations, 3) 2-D and 3-D ice models based on realistic physics of ice sheets, and 4) analog forecasts of the climate fields based on past interglacial-glacial transitions. In addition, a simulation will be run for a "super-glacial" and a "super-interglacial" — scenarios that may fall outside the "most probable" subset.

Seasonal Insolation

Assuming constant solar output, the seasonal insolation field for the next 200,000 years can be calculated with a relatively high degree of confidence (cf. Berger 1978) and is available as an acquired code.

CO₂

The burning of fossil fuel and release of radiatively important trace species (National Research Council 1982, Dickinson and Cicerone 1986) will result in a significant increase in the greenhouse effect by the beginning of the next century (Figure A.13). The temperature and regional climate changes from such an effect could be quite significant (e.g., Manabe and Wetherald 1986, Schlesinger and Mitchell 1987, Ramanathan 1988). Although

FIGURE A.13. Comparison of Future Greenhouse Warming with the Mean Surface Temperature for the Northern Hemisphere Since 1860. Solid line shows "natural variability" of climate over the last one hundred years. This value is about $\pm 0.5^{\circ}\text{C}$ (Jones et al. 1986). Dotted line shows possible future global mean surface temperature change. Crosshatching shows uncertainties in model calculations. [Figure from Stuiver et al. 1981, adapted from Mitchell 1977 and Kellogg 1977]

there are a number of significant uncertainties with respect to greenhouse models, for the sake of site assessment we have to consider the worst-case projections that in fact the perturbation could be quite severe. To assess just how severe it might be, we have compared future projected warming with past geological warm periods (Crowley 1989, 1990). Indications are that the future warming could be a major perturbation of the environment (Figure A.14) that may be a unique climate realization in earth history.

FIGURE A.14. Schematic Comparison of Future Greenhouse Warming with Past Changes in Temperature over Geological Time Periods. Note that Pleistocene climate oscillations are idealized and that interglacial temperatures were probably not significantly warmer than present. Pre-Pleistocene changes are not well fixed in magnitude, but their relative warmth is approximately correct. Maximum warming in the Cretaceous based on estimates by Barron and colleagues (Thompson and Barron 1981, Barron and Washington 1985). Time intervals in between have been scaled accordingly. [From Crowley 1989]

As the time scale for removal of the CO₂ perturbation is a few thousand years (Sundquist 1985), then the greenhouse effect may likely dominate future climate scenarios over the next few thousand years.

Projecting future greenhouse gas concentrations is either very hard or very easy. It is very hard, almost impossible, if we try to take into account all the economic and human factors that affect future CO₂ outputs. It is relatively easy if we establish a

number of benchmark levels into which all CO₂ scenarios fall — for example, low, medium, high and very high concentration (2X, 4X, 8X present levels), and conduct a suite of sensitivity experiments for each. This approach is motivated by the recognition that it is undesirable to exclude too many scenarios at an early stage. We will proceed in the latter manner.

Ice Volume and Extent

Prediction of ice volume and extent will develop along three separate but complementary paths: probabilistic, simple models of global ice volume, and more complex models for individual ice sheets.

Probabilistic Models. In a stationary system (i.e., causes of variations are independent of time), it is possible to objectively predict future ice volume fluctuations based on the frequency distribution and autocorrelation characteristics for past ice volume fluctuations. This approach utilizes standard statistical techniques to quantify probabilities of occurrence of different ice volumes for the future (e.g., North and Cahalan 1981). Initial results (Newton et al. 1990, cf. Appendix H) indicate that over the next ten thousand years there is a relatively low probability of significant ice volume increases (Figure A.15), but there is a 95% probability that approximately sixty thousand years from now there will be a major ice age.

Global Models of Ice Volume. Numerous attempts have been made to model past ice volume fluctuations (e.g., Imbrie and Imbrie 1980, Hyde and Peltier 1985); which are estimated by the ¹⁸O record of deep-sea microfossils. As ice volume is related to orbital-insolation forcing, whose future effect is known, this approach will examine different models of global ice volume to predict its future course. As an example of work already done on this problem, Imbrie and Imbrie (1980) have estimated that a significant ice growth event will occur 23,000 years from now,

FIGURE A.15. Forecasts and 95% Probability Limits of Future Ice Volume Fluctuations Using an Autoregressive Statistical Model. Solid curve on left is observed proxy ($\delta^{18}\text{O}$) ice data (from Imbrie and Imbrie 1980). Middle curve on right represents ice volume forecast. Upper and lower curves represent 95% probability envelope. Positive values on vertical axis represent times of low global ice volume. [From Newton et al. 1990]

with a larger event around 60,000 years in the future. These calculations are broadly similar to that illustrated for the probabilistic model (Figure A.15, cf. Figure 11).

Numerical Models of Ice Fields. The five main areas of ice in the world are the Arctic and Antarctic sea ice fields, and the Greenland, West Antarctic, and East Antarctic Ice Sheets. As ice affects climate in many ways, it is necessary to identify possible scenarios as future boundary conditions. At this stage, the identification of boundary conditions for forecast experiments

diverge from the similar track for hindcast experiments, because it is most reasonable to treat sea ice fields as outputs from climate model results (e.g., Parkinson and Kellogg 1979, Parkinson and Bindshadler 1984), whereas the ice sheets, which are partly decoupled from the atmosphere/ocean system because of their long response time, can be treated somewhat separately.

The ice-sheet modeling activities for this project have been done at the University of Maine. In order to model the future course of ice sheets, it is necessary to focus on the most energetic part of the ice sheet dynamic system - the basal ice, especially the ice-bed interface (e.g., Hughes 1987). Near the margins of present-day ice sheets in Greenland and Antarctica, basal conditions that promote ice-bed uncoupling tend to develop linearly along certain flowlines, converting them into ice streams, which are fast currents of ice that drain about 90 percent of these ice sheets. The physical processes associated with time variations in ice streams are not understood in detail. Efforts to quantify the physical processes associated with ice stream dynamics center on theoretical studies and numerical modeling of ice stream dynamics. Progress on both fronts depends heavily on data collected from field studies of ice streams, surging mountain glaciers, and calving tide-water glaciers. Field work conducted on ice streams by the University of Maine has been concentrated on Byrd Glacier, the largest Antarctic outlet glacier buttressed by the Ross Ice Shelf, and Jakobshavns Isbrae in Greenland, which has a calving floating terminus and is the fastest-known ice stream. Both these studies are designed to produce the field data that will test theoretical models of ice stream dynamics, and constrain the computer models based on these theoretical concepts. The ice sheet modeling approach is philosophically identical to the climate modeling approach, in that there is a progressive upgrade in the sophistication of global models, as detailed below:

One-dimensional Flowband Model - This model produces essential features of sheet flow, stream flow, and shelf flow from the central ice divide of an ice sheet to its melting or calving margin, and includes a range of boundary conditions that covers the spectrum of known behaviors on existing ice sheets and mountain glaciers. The model has also been used as a check on more complex gridpoint models discussed below.

Two-dimensional Gridpoint Models - These models have been developed with both finite-difference and finite-element solutions. Both models produce the same features for the same range of boundary conditions as produced by the one-dimensional flowband model, but in addition they construct flowbands as output rather than input. The finite-element solution can also vary the density of gridpoints to capture details of bed topography and surface flow for gridpoints in plan view, and can solve the equations of heat flow for gridpoints in profile along a flowband. This includes plotting the advance and retreat of ice sheets through time, and changing temperature profiles along flowbands through time as boundary conditions change. Global ice sheets that are late Quaternary glacial maxima are being reconstructed using these models. For example, we have now simulated the three-dimensional structure of the Laurentide Ice Sheet at the last glacial maximum (Figure A.16).

Time-dependent Gridpoint Finite-element Model - This model will be used to simulate the advance and retreat of ice sheets for future glaciation cycles over the next 100,000 years. Three scenarios will be considered:

In the first scenario, anthropogenic "greenhouse" warming has no significant effect on the polar ice cover. In the model, winter precipitation would increase in high latitudes in step with insolation changes controlled by the 41,000 year axial-tilt cycle, and increase in mid latitudes in step with insolation changes controlled by the 21,000 year axial-precession cycle, in

FIGURE A.16. Example of Three-dimensional Ice Sheet Model Simulation for the Last Glacial Maximum. [From T. Hughes and J. Fastook, unpublished results]

conjunction with a decrease in summer melting. Northern Hemisphere ice sheets would then nucleate in highlands where winter snow did not melt and in shallow marine embayments where sea ice thickened and grounded, with their advance and retreat controlled by both the tilt and precession cycles. The fall and rise of sea level with these advances and retreats would cause the Antarctic Ice Sheet to advance and retreat to the edge of the Antarctic continental shelf. Analogs from the past would be the late Quaternary glaciation cycles.

In the second scenario, anthropogenic "greenhouse" warming melts the perennial polar sea ice. In the model, precipitation over Antarctica would then increase greatly, causing the Antarctic Ice Sheet to advance to the edge of the continental shelf and

retreat to near its present-day position, in step with the 41,000 year high-latitude insolation cycle controlled by axial tilt, which would also control sea-level changes caused by the advance and retreat of high-latitude ice sheets in the Arctic. Northern hemisphere ice sheets would not reach mid-latitudes, where the 21,000 year axial precession cycle would control insolation. Analogs from the past may be the early Quaternary glaciation cycles.

In the third scenario, anthropogenic "greenhouse" warming melts the perennial polar sea ice and disintegrates floating ice shelves that fringe Antarctica. In the model, removal of the ice shelves could unbuttress ice streams that drain large marine subglacial basins in East and West Antarctica, causing a nearly total collapse of the West Antarctic Ice Sheet and a partial collapse of the East Antarctic Ice Sheet over subglacial basins. Global sea level would rise from 6 to perhaps 30 meters. Terrestrial portions of the Antarctic Ice Sheet would remain, and could advance and retreat in step with increased winter precipitation and reduced summer melting controlled by the 41,000 year insolation cycle of axial tilt. Northern Hemisphere ice sheets would not form, because summer temperatures would be too high in the Arctic. Analogs from the past may be the late Tertiary glaciation cycles. These tasks will occupy the 1989-1990 subcontract year.

Sea Surface Temperatures

As stated in the introduction to this section, it is hoped that specification of albedo and CO₂ fields will enable us to calculate SSTs, both hemispheric-averages and geographic distribution.

Extreme Scenarios

The various approaches outlined above may not capture the range of all possible scenarios. To cover this possibility we

will also consider "super-glacial" and "super-interglacial" scenarios.

"Super-glacial" Scenario. An additional "super-glacial" simulation may be run to account for glaciation more severe than the last glacial maximum at 18,000 BP. The most likely past candidate for a super-glacial is ¹⁸O Stage 6 (150,000 BP) — subtropical SSTs were colder than the last glacial maximum (Luz 1973, Thompson and Shackleton 1980, Crowley 1981); deep-water changes were probably larger than the last glacial maximum (Curry and Lohmann 1982, Boyle and Keigwin 1985/1986), and southern hemisphere winds may have been stronger (Thiede 1979). As most of these changes represent extensions of the patterns inferred for 18,000 BP (CLIMAP 1981), the latter interval can be used as an initial condition upon which 150,000 BP changes will be superimposed.

"Super-interglacial" Scenario. Although this scenario might also fall outside the subset of "most probable" scenarios (Figure A.12), simulations will nevertheless be done for the purpose of completeness. Likely candidates are simulations involving 4X and 8X concentrations of CO₂-levels that would be approached due to complete exhaustion of fossil fuel resources (e.g., Keeling and Bacastow 1977) and reduced absorption of emitted gases by the oceans and biosphere. Corresponding atmospheric temperatures could approach those for an ice-free earth (Thompson and Barron 1981), although the ice sheets would probably still be present because of their inertial properties (Crowley 1989).

Transient Scenarios

One of the most difficult problems to address will be the fact that the future range of climate change cannot be entirely captured with equilibrium runs for 2X, 4X, and 8X CO₂. For example, ocean modeling experiments indicate that time-dependent changes in atmospheric CO₂ can result in significantly different

regional patterns than might be expected from a CO₂ doubling (Stouffer et al. 1989, Washington and Meehl 1989, Mikolajewicz et al. 1990). Such patterns may affect regional temperature patterns at Yucca Mtn.

Unfortunately, the combination of the limitation in the present state-of-the-art in coupled ocean-atmosphere models and prohibitive expenses of coupled ocean-atmosphere model transient experiments, makes it impossible to repeat all of the projected forecast experiments in a transient mode. However, it is still possible to address this problem using a "scenario" approach.

"Reduced North Atlantic Deep Water" Scenario. Geologic data (e.g., Boyle and Keigwin 1987) and conjectures and models of future transient climate change (Broecker 1986, Washington and Meehl 1989, Mikolajewicz et al. 1990) suggest that deep water production in the North Atlantic may be reduced for some greenhouse scenarios. Such a response would result in lower SSTs in the subpolar North Atlantic, a pattern that could affect the circulation pattern at Yucca Mtn. through upstream adjustment of the planetary waves. To accommodate this effect, we will conduct one sensitivity experiment with cold SSTs in the North Atlantic but warmer SSTs elsewhere due to a CO₂ warming.

"Most Extreme" Scenario. Transient circulation changes could result in other patterns than a cold North Atlantic. Based on our studies of historical climate fluctuations (cf. Appendix G), it may be possible to erect a set of boundary conditions that would represent the greatest threat to the integrity of the Yucca Mtn. site. For example, studies from both the last 100 years and the Pleistocene (Appendix G, Kutzbach and Wright 1985) suggest that a combination of relatively high SSTs and a circulation pattern with frequent low pressure systems moving in from the Pacific present conditions which cause precipitation extremes in southwest Nevada.

It is possible that future CO₂ changes may cause a variation that exceeds historical variations (100-200%). To test this idea, we will make one run with very high SSTs (from 8XCO₂) and constrain the atmospheric circulation to respond in a manner that delivers frequent low pressure systems to the southwest U.S. Such a pattern may never occur in an actual transient run, but by taking advantage of our understanding of the circulation patterns of greatest concern, we may be able to circumvent our inability to address the problem with a brute force coupled ocean-atmosphere run.

A.6 ACTIVITY: FORECAST EXPERIMENTS (ARC, LDGO)

A.6.1 Objectives

The objectives of this activity are to force climate models with altered boundary conditions in order to develop an ensemble of model projections about future climate change. A subset of model simulations will be chosen for GCM runs. Model output (with uncertainties) will be compared to empirical forecasts of future climate.

A.6.2 Parameters

The parameters of this activity are:

1. boundary conditions of ice, SST, CO₂, and seasonal insolation for the next 100,000 years
2. dynamical model output of seasonal temperature fields from ensemble runs
3. dynamical model output of seasonal climate fields from the ensemble-subset runs
4. empirical model output of climate projections for the next 100,000 years.

A.6.3 Description

Model runs

A matrix of CO₂/Ice scenarios (Figure A.12) will be utilized as boundary conditions for an ensemble of runs with lower-resolution climate models in order to assess the range of possible temperature fields on the earth's surface (FORECAST-I experiments). From the ensemble a subset will be chosen that is deemed either a) most likely or b) most extreme. This subset will be used to identify boundary conditions for precipitation simulations with a low-resolution GCM (FORECAST-II experiments). The entire suite of results will then be evaluated, with the final product (with uncertainties) being dynamical predictions of model fields as input to regional-scale models.

Empirical model forecasts

In parallel with the dynamical forecasts, a suite of empirically-based forecasts will be developed at LDGO. This approach will be based on "reasoning by analogy". For example, during the last one million years the average duration of an interglacial event is about 10,000 years. As the present Holocene interglacial has lasted for about 10,000 years, in a stationary system we might expect a return to cooler climates, with the rate of return driven by changes in the rate of forcing. Empirically-based forecasts of future climate will then be based on ranges of possible climates that result in the 13,000 year interval following deglaciation (cf. Kukla et al. 1981). Likely candidate events are the interval 110,000-120,000 BP (the last interglacial/glacial transition) and 380,000-410,000 BP (a prior interglacial/glacial transition that was associated with smaller and slower changes in orbital forcing, as will occur in the next 20,000 years).

Initial results of the empirical model forecasts are illustrated in Figure A.17. This figure was based on examining

FIGURE A.17. The Distribution of Past Climate States as a Function of Time Elapsed from Terminations. Terminations are defined as the midpoints of sharp environmental changes accompanying major deglaciations. Climate states are measured by a climate index (CI), which is defined by linear interpolation between a full glacial environment with CI = 100 and the full interglacial (Mid-Holocene) environment with CI = 200. The time scale is in thousand years. The time series starts at -13 ka, reflecting the age of the last Termination I, and extends 120,000 years into the future. Thus the origin of the time scale (0) can be taken as a present day analog in the glacial/interglacial climate cycle. Full line shows the arithmetic average of the stacked record. Based on the seven segments of a Macedonian pollen core. [G. Kukla, unpublished]

the range of climate conditions following peak interglacial for each of the past four interglacials. It is evident that a range of possible climates resulted. This information is important as it helps quantify the range in uncertainty in our future forecasts.

A.7 ACTIVITY: SYNTHESIS AND CONCLUSIONS

A.7.1 Objectives

The objectives of this activity are to synthesize the results from dynamical and empirical forecasts and produce a final report specifying boundary conditions (with uncertainties) for regional-scale model forecasts.

A.7.2 Parameters

The parameters of this activity are regional forecasts of various climate parameters.

A.7.3 Description

Comparison of Dynamical and Empirical Forecasts

The entire suite of forecasts for the future will be compiled and evaluated in terms of range and probability of occurrence of different events. Objective techniques (e.g., Campbell and Cranwell 1988) will be used to estimate uncertainties.

Comparison of Results with Estimates of Groundwater Recharge

The final product will include a set of estimates of changes in precipitation-minus-evaporation (P-E) for various different climate scenarios. This information will be used for regional-scale forecasts.

There are some indications that future studies, despite their uncertainties, may provide enough information to reasonably assess the likelihood of site endangerment by climate change. The reasoning is as follows. Most studies of past precipitation

FIGURE A.18. Latitudinal Distributions of the Difference Between the Rate of Precipitation and that of Evaporation (P-E) for Different Levels of CO₂ Change. Note that P-E changes for higher CO₂ levels increase by less than a factor of two in zonal averages (regional changes could be longer, however). Units are in centimeters per day. [From Manabe and Bryan 1985]

changes suggest that precipitation did not change by more than a factor of two (e.g., discussions in Czarnecki 1985, Winograd and Szabo 1988). Evaluation of five different GCM results for doubled CO₂ indicates that P-E changes in winter for this region varied from about -10% to +20% for the southwestern U.S. (Kellogg and Zhou 1988). Experiments with higher levels of CO₂ suggest maximum latitudinally averaged changes of about a factor of two (Figure A.18, Manabe and Bryan 1985).

The above P-E changes are of interest when compared to changes required to endanger the Yucca Mtn. site. Groundwater modeling studies suggest that a 100% increase in precipitation and

a fifteenfold increase in groundwater recharge would raise the water table about 130 m (Czarnecki 1985), significantly less than the 200 m minimum distance between the repository and the present water table, and significantly greater than the ~10 m groundwater fluctuations in this region during the last 30,000 years that were associated with an estimated 100% increase in precipitation (cf. Spaulding et al. 1984, Winograd and Szabo 1988). The recharge values appear to be extreme estimates, which in fact probably reflect precipitation increases >>100%. Although there are a number of uncertainties in all these processes, overall assessment provides a clear focus for future research activities.

Final Product: Forecasts of future climate change

The set of all forecasts, together with team-member assessments of uncertainties, will then be delivered as a final product to research teams that will be making regional-scale forecasts.

A.8 SOME CLIMATE MODELING DEFINITIONS

A GLOBAL CLIMATE MODEL is a mathematical computer model that is capable of simulating the seasonal cycle of certain weather variables starting with the planetary scales of atmospheric behavior and including other scales down to some "resolution". These models must be run long enough on the computer for transient effects to die out. An annual cycle model typically uses a simple 50 - 75 meter deep ocean for the lower boundary constraint on the atmosphere. Simulations with these models take a minimum of 15 model years. An imitation of a part of the seasonal cycle can be had by setting the sun at its January position, for example, and prescribing the sea surface temperature for that month from data. While the verisimilitude of this imitation is a matter of dispute it cuts down computer time by about a factor of 20 and it may be valuable for certain purposes. Global atmospheric models have to

use other boundary conditions as input, such as the placement and thickness of ice sheets, the atmospheric composition and the oceanic structure. The output of a global climate model is the seasonal cycle in each grid box over the earth. We call this the regional climate forecast.

Examples of global models used in this study:

EBM. EBM is a generic term for an "energy balance model", which calculates temperature as a function of the incoming solar radiation, modified by the albedo (reflectivity) of the earth. In our report, we focus strictly on seasonal models, which involve realistic seasonal forcing and response (as opposed to annual-averaged forcing). We also use only two-dimensional models, with the dimensions being equivalent to latitude and longitude so that continental outlines can be resolved (this step is necessary because it can be demonstrated that land-sea distribution represents the principal control of seasonal temperature fluctuations). There are several different types of EBMs that we discuss and employ:

LINMOD. LINMOD is the basic version of the 2D-EBM from which others are derived. It was developed by North et al. (1983) and resolves the seasonal cycle but is a linear model and does not allow for the regionally varying albedo of snow and ice.

LINEBM. LINEBM is a modification of LINMOD that allows for regionally varying but prescribed albedo of snow and ice. This model, which is linear, was developed at ARC, described in Hyde et al. (1989), and tested against general circulation model (GCM) simulations of climate change over the last 18,000 years.

NEQMOD. NEQMOD is a model recently completed under this PNL contract. The abbreviation stands for Nonlinear EQUilibrium MODEL, for it is a nonlinear model that allows for interactions of altered seasonally varying snow and ice fields with the planetary radiation budget. Such a model is important for testing, for

example, the future effect of greenhouse gas changes on climate, for much of the climate system response is nonlinear and involves the snow and ice fields. The "equilibrium" term in the title refers to the fact that there are two different methods of solution for the ice-albedo feedback, one involving spherical harmonic expansions in order to achieve an equilibrium solution, the other involving finite-differencing, which determines a time-marching solution.

GCM. GCM stands for general circulation model. This class of models solves the standard equations governing momentum, moisture, etc. and explicitly computes fields such as circulation, precipitation, etc. It therefore provides much more information than an EBM, but at a much-greater cost and presumably is more realistic. At present our project envisions using two different types of GCMs:

LOW-RESOLUTION GCM (LRM). LRM is, as the name implies, a low-resolution version of a GCM. The model has approximately $15^\circ \times 15^\circ$ resolution in the horizontal, with five levels in the vertical, and a simplified radiation code. All of these features enable the model to be run approximately ten times faster than the high-resolution version. The LRM is an ideal model for testing a range of parameter space values before settling on a few experiments to run with the most comprehensive type of GCM (see below).

HIGH-RESOLUTION GCM. The High-Resolution GCM is, again as the name implies, a high-resolution model, with about a $5^\circ \times 5^\circ$ horizontal resolution and about ten levels in the vertical, and a more complicated radiative code. This model gives the most detailed information about climate change, but is also the most expensive to operate. The model we plan to use later in the project is a version of the Community Climate Model (CCM) from the National Center for Atmospheric Research (NCAR; cf. Ramanathan et al. 1983, Williamson et al. 1987).

MESOSCALE MODEL. Results from global-scale models will be used as input to a finer-mesh regional model being developed at NCAR (Dickinson et al. 1989, Giorji and Bates 1989, Giorji 1990). This model is capable of simulating changes on a spatial scale an order of magnitude finer than the GCM.

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APPENDIX B

MODELING FUTURE CLIMATES
FOR NUCLEAR WASTE DISPOSAL^c

by

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-
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APPENDIX B

MODELING FUTURE CLIMATES FOR NUCLEAR WASTE DISPOSAL

B.1 INTRODUCTION

Long term disposal of high level nuclear wastes requires consideration of climate change on time scales of thousands of years. This kind of research activity is unprecedented in its scope. We have paleoclimate data that we can draw some guidance from. For example, during the last glacial maximum (18,000 BP, before present), conditions were moister in the southwest U.S. (Smith and Street-Perrott 1983, COHMAP 1988). However, we have no assurance that future climates will behave in the same way as past climates on the time scales of tens of thousands of years. In fact, the greenhouse effect could be a significant non-stationary climate perturbation (see below).

The purpose of this report is to outline a plan for making future climate forecasts for the Yucca Mtn. site near Las Vegas, Nevada. The plan is divided into several sections: 1) general structure of the plan, 2) rationale for a hierarchical climate modeling approach, 3) progress to date, and 4) strategy for combining climate results with hydrologic estimates linking groundwater recharge to climate change.

B.2 GENERAL STRATEGY

Regional-scale estimates of climate parameters will be based on both dynamical and empirical techniques. Dynamical methods will utilize global climate models for prediction of input parameters into regional-scale models (regional-scale forecasts

are defined as grid-point output from global climate models). The output is then used as input for higher resolution models designed to make site-specific forecasts (e.g., Yucca Mtn.). Since the seasonality of temperature and precipitation is very strong and region-dependent, seasonal climate models are essential in these studies.

Empirical techniques will also examine past climate fluctuations as a means of estimating probabilities of future climate fluctuations of different magnitudes. The groups involved in the above tasks have expertise in climate and ice sheet modeling, and paleoclimatology, and are from the Applied Research Corporation (ARC), Texas A&M University (TAMU), the University of Maine (UM), and Lamont-Doherty Geological Observatory (LDGO).

Because climate models in their most advanced form are expensive to operate, we will employ a hierarchical approach to climate modeling. This strategy involves developing and conducting numerous model runs of simple climate models in order to explore the range of model behavior. These pilot studies will help identify the most promising candidates for more detailed (and expensive) climate simulations. The latter will form the basis for making regional-scale estimates of relevant climate parameters.

Since climate model simulations have inherent uncertainties, it is important to validate the models against independent data sets. A necessary element of the research program will be to explore how climate models respond to past altered boundary conditions, and compare model output with geologic data. In order to conduct these "hindcast" experiments, the models will require various paleoclimate drivers as boundary conditions (ice, sea-surface temperature, CO₂, seasonal insolation forcing). These drivers will be developed as subtasks on the project. Hindcast validation will be mostly against datasets developed by other groups working on the site selection project. As we are adopting

a hierarchical approach to the modeling endeavor, an additional important step will be to validate simple models against more complicated models in order to determine their relative sensitivities to altered boundary conditions.

Once the various climate models have been tested against past data, the models will be operated in a forecast mode, utilizing altered boundary conditions specified by various team members. Results will be compared with empirical projections based on statistical analyses of past climate fluctuations. The combined dynamical and empirical forecasts, together with their uncertainties, will be analyzed and synthesized. The final report will then be provided to a site-specific climate modeling group, which will be funded separately from this study.

B.3 STRATEGY FOR USING CLIMATE MODELS

The only feasible approach to future forecasts is through the use of global climate simulation models. The reason for this is that the variability of climate is over a continuum of spatial scales, with the great majority of the variance at scales larger than the contiguous United States. This is due partly to the dynamics of the climate system and due to the fact that most of the forcings that cause climate change are also at very large spatial scales. For example, Figure B.1 shows a variance spectrum in spherical harmonic degrees indicating the concentration of variance at smaller degrees (larger spatial scales).

Climate modeling research has blossomed over the last 15 years in response mostly to concerns about the global warming on the decadal time scale. There is a range of climate models starting with simple thermodynamic equations that balance the incoming and outgoing planetary radiation and therefore determine

VARIANCE OF TEMP FOR DIFFERENT SCALES

FIGURE B.1. Schematic Plot of the Variance of Surface Temperature as a Function of Spherical Harmonic Degree Indicating the Amount of Climate Variability at Scales Larger than Degree 10 (Corresponding to about 1800 km in linear distance), the Resolution of many Simple Climate Models. Some general circulation models have resolution at degree 15.

the planetary average surface temperature (North et al. 1981) to the giant models which are used only in global medium range weather forecasting because they do not run much faster than nature (Bengtsson and Simmons 1983). Neither of these extremes are useful for the problem at hand. Fortunately, there is a hierarchy of models (Schneider and Dickinson 1975) in between these and we must make use of the hierarchy concept to put limits on future climates for our practical application.

The main assumption underlying the hierarchy framework is that a sequence of approximations exist starting with models of the planetary scales and leading toward models with finer and finer spatial and temporal resolution. At each stage of the sequence more and more physical detail is included. The general circulation models which include the motion of the air, detailed radiative transfer subroutines and moisture transformation

processes are the ultimate tool for climate modeling. However, even they are as yet imperfect mainly because of uncertainties associated with processes at finer scales than typical numerical grid resolutions used in current climate models (about 500 km horizontally). Finding consistent climate responses to boundary condition changes up and down the model hierarchy tends to increase our confidence in the output of all the models. This robustness of climate response seems to especially hold when the perturbing influences cover large spatial scales.

Our approach has been to develop and select a suitable skeletal hierarchy of climate models and check their responses to hypothesized boundary condition perturbations that seem consistent with future scenarios, with special attention to extremes. First consider the climate models and how they were chosen and tested. We have chosen three models:

1. a two-dimensional seasonal energy balance model (EBM; Hyde et al. 1989) of the surface temperature field with spatial resolution of about 1800 km
2. a low resolution general circulation model (LRM; Otto-Bliesner and Houghton 1986) with the same horizontal resolution as the EBM but with five vertical levels and retention of atmospheric winds and moisture
3. the general circulation model (GCM) developed at the National Center for Atmospheric Research, the Community Climate Model (CCM; Ramanathan et al. 1983), which has 11 vertical levels and horizontal resolution of about 500 km.

The philosophy is to first check the EBM and LRM against the CCM in forced climate change experiments, taking the latter as truth. We have chosen experiments where CCM runs have already been conducted for other purposes, mainly paleoclimate studies. Since the CCM has been widely used in climate change experiments ranging from the last glacial maximum 18,000 years before present

to the doubling of carbon dioxide, it is a good standard by which to test our less expensive alternatives. Of course, it is well known that the half dozen GCMs around the world have minor differences in sensitivities and hence "truth" here has to be taken in the sense that our models are good imitators of the CCM, which may differ slightly from other models. We use the CCM as our standard because it is the most widely used GCM and the most archived data from paleoclimate experiments are available for it. In its defense, while it may not be arguably the best GCM it is certainly a reasonable representative and by far the best documented.

B.3 PROGRESS TO DATE

B.3.1 Some Intermodel Comparison Studies

We have completed (Hyde et al. 1989) a series of tests of the EBM against the CCM consisting of equilibrium seasonal cycles at present, 3000, 6000, 9000, 12,000, 15,000 and 18,000 years ago. The results are quite encouraging (Figure B.2), since we find that when fairly large areas (boxes of order 3000 km) of land or ocean are smoothed over, the two models agree to within 1-2°C for each of the time periods. We conclude that the EBM is an excellent imitator of the CCM in both its simulation of the present climate and in its sensitivity to boundary condition and forcing changes. Unfortunately, the EBM can only calculate the seasonal distribution of surface temperature and this is only part of the description of climate needed in the present application. We are presently conducting the same series of experiments with the LRM.

B.3.2 Identifying Boundary Conditions for Future Scenarios

How are the boundary conditions for future scenarios to be chosen? The three most important factors that can modify climate on time scales of millennia are: 1) geographic variations in orbital insolation receipt at the top of the atmosphere (the

Northern Hemisphere July Land Temp Anomalies

Global Annual Average Temp Departures (Ocean)

FIGURE B.2. Comparison of EBM and GCM Simulations over the Last 18,000 Years for (top) Northern Hemisphere Land in July, and (bottom) Global Ocean Temperatures. Note the comparable sensitivities of the models to seasonal changes in forcing. GCM runs (Kutzbach and Guetter 1986) utilized prescribed SSTs, whereas they were calculated in the EBM (Berger 1978). The good agreement suggests that we may be able to constrain SST values for time intervals where they are not known.

Milankovitch effect), 2) variations in global ice volume, and 3) variations in atmospheric greenhouse gas composition.

To varying degrees of confidence we can estimate future variations in the above boundary conditions. Seasonal variations in geographic insolation can be calculated accurately from celestial mechanics considerations (Berger 1978). On the distribution of glacial ice we can take some guidance from past glaciations. The record of ^{18}O can be used to estimate global ice volume over the last million years. The record is especially good over the last 400,000 years (Imbrie et al. 1984). A power spectrum analysis of the data show strong peaks at the frequencies of orbital element changes (Milankovitch theory) with a background of red noise.

One can use either simple physical models or a linear statistical model to forecast future ice volumes in the next 100,000 years. The advantage of the latter procedure (Newton et al. 1990) is that one can also place error bars on the forecast to find the probability of ice volumes outside a certain envelope (Figure B.3). This study suggests a low probability for a large ice volume increase in the next 10,000 years. For any of the time intervals in the next 100,000 years we can also construct ice sheets on the continents consistent with the volume constraint and largely based upon past placement of these sheets from geological evidence and from physical models of ice sheets. These latter inputs are being provided by the University of Maine.

Greenhouse gas fluctuations are another potential threat to future climates. Although the magnitude and nature of this effect has been a topic of some debate, for the purposes of site assessment we must consider the worst-case possibility. A comparison (Crowley 1989) of the future warming due to the greenhouse effect with past geologic fluctuations (Figure B.4)

FIGURE B.3. Forecasts and 95% Probability Limits of Future Ice Volume Fluctuations Using an Autoregressive Statistical Model. Solid curve on left is observed proxy ($\delta^{18}\text{O}$) ice data (Imbrie and Imbrie 1980). Middle curve on right represents ice volume forecast. Upper and lower curves represent 95% probability envelope. Positive values on vertical axis represent times of low global ice volume (Imbrie and Imbrie 1980). [From Newton et al. 1990]

suggests that the greenhouse effect will likely dominate the earth's climate for the next few thousand years (about 5000-10,000 years are required to remove the gases by buffering with the deep-ocean carbonate reservoir (Sundquist 1985).

The atmospheric composition is to be input to the models taking high values of carbon dioxide as examples. We will try various atmospheric loadings that might lead to climate extremes in the western United States. With the EBM we can now make many runs to look for interesting cases to examine more closely with the more sophisticated models.

FIGURE B.4. Schematic Comparison of Future Greenhouse Warming with Past Changes in Temperature over Geological Time Periods (Crowley 1989). Note that Pleistocene climate oscillations are idealized and that interglacial temperatures were probably not significantly warmer than present. Pre-Pleistocene changes are not well fixed in magnitude, but their relative warmth is approximately correct. Maximum warming in the Cretaceous is based on estimates by Barron and colleagues (Thompson and Barron 1981, Barron and Washington 1985). Time intervals in between have been scaled accordingly.

B.4 PRELIMINARY ASSESSMENT OF CLIMATE PROJECTIONS

There are some indications that future studies, despite their uncertainties, may provide enough information to reasonably assess the likelihood of site endangerment by climate change. The reasoning is as follows. Most studies of precipitation changes during the last 20,000 years suggest that it did not change by much more than a factor of two (e.g., discussions in Spaulding et al. 1984, Czarnecki 1985, Benson and Thompson 1987, Winograd and Szabo 1988). Evaluation of five different GCM results for doubled

FIGURE B.5. Latitudinal Distributions of the Difference Between the Rate of Precipitation and that of Evaporation (P-E) for Different Levels of CO₂ Change (after Manabe and Bryan 1985). Note that P-E changes for higher CO₂ levels increase by less than a factor of two in zonal averages (regional changes could be larger, however). Units are in centimeters per day.

CO₂ indicates that precipitation-minus-evaporation (P-E) changes in winter for this region varied from about -10% to +20% for the southwestern U.S. (Kellogg and Zhao 1988). Experiments with higher levels of CO₂ (Figure B.5) suggest maximum latitudinally averaged changes in P-E of about a factor of two (Manabe and Bryan 1985).

Initial assessment of the above results suggests a strategy for risk assessment at the Yucca Mtn. site, even though there are a number of significant uncertainties. For example, ground-water modeling studies suggest that a 100% increase in precipitation and a fifteenfold increase in ground-water recharge would raise the water table about 130 m (Czarnecki 1985), significantly less than

the 200 m minimum distance between the repository and the present water table, and significantly greater than the ~10 m ground-water fluctuations in this region during the last 30,000 years that were associated with an estimated 100% increase in precipitation (cf. Spaulding et al. 1984, Czarnecki 1985, Winograd and Szabo, 1988). The recharge values appear to be extreme estimates, which in fact probably reflect precipitation increases >100%. Although the calculations are subject to some uncertainty and require considerable refinement, they provide some hope that the recharge values required to endanger the Yucca Mtn. site could significantly exceed the present range of uncertainty in climate model precipitation estimates.

B.5 FUTURE RESEARCH

Our future plans include trying to establish bounds on climate forcings due to solar activity using time series analysis on past solar indices, investigation of the synoptic situations that cause increased precipitation in the Southwest, blending the glacial boundary conditions into our work from the Maine group, and, of course, continuing the experimentation with the climate model hierarchy with its emphasis on climate extremes in the Southwest.

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APPENDIX C

COMPARISON OF SIMULATED AND OBSERVED GEOLOGICAL
TIME SERIES FROM THE SOUTHWESTERN U.S.

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APPENDIX C

COMPARISON OF SIMULATED AND OBSERVED GEOLOGICAL TIME SERIES FROM THE SOUTHWESTERN U.S.

C.1 INTRODUCTION

One of the goals of the Global Climate Modeling project (Appendix A) is to test forecast models by making "hindcast" comparisons between the models and the geologic record, with the assumption being that we cannot make satisfactory forecasts until we demonstrate that we understand past fluctuations. Until recently, our most detailed knowledge of past changes in climate in the Yucca Mountain area were restricted to about the last 20,000 years. Over that time interval there are some impressive agreements between models and observations (Kutzbach and Wright 1985).

Recently, a group from the U.S. Geological Survey has produced a 250,000 year record of ground water variations from a site near Yucca Mountain (Winograd et al. 1988). The chronology of this radiometrically dated series was significantly different (Figure C.1) than the chronology of events established for the global ice volume record by the SPECMAP group (Imbrie et al. 1984). Although Winograd et al. (1988) suggest that the chronology of the ice volume record may be in error, there are two other possible interpretations of their record: 1) the record may represent some combination of global (ice volume) and local forcing that resulted in a phase shift between the Great Basin record and the deep-sea $\delta^{18}\text{O}$ record of global ice volume; 2) there may be unidentified sources of dating error in the Great Basin time series.

FIGURE C.1. Time Series of Great Basin $\delta^{18}\text{O}$ Groundwater Variations (after Winograd et al. 1988) Compared to the SPECMAP Chronology of $\delta^{18}\text{O}$ and Global Ice Volume (from Imbrie et al. 1984).

In this report we discuss results that primarily address the possibility that the Great Basin record may represent some combination of global and local forcing, which produces very large phase differences between local and global records. Such responses are known to occur in some marine records, with large phase leads between local and the global signals that are not unlike the Great Basin/global ice volume couplet (e.g., Curry and Crowley 1987). For example, the carbon record illustrated in Figure C.2 shifts to a more-interglacial pattern at about the same time (140 kbp, thousand years before present) as the Great Basin $\delta^{18}\text{O}$ record (cf. Figure C.1). Yet the deep-sea carbon and oxygen isotope measurements were made on the same specimens. With respect to the offset, there is no question of dating error, as there is a physical separation in the sediment record between the two transitions.

FIGURE C.2. Comparison of Marine $\delta^{18}\text{O}$ and $\delta^{13}\text{C}$ Records from the Equatorial Atlantic (from Curry and Crowley 1987). The equatorial Atlantic record (termed $\Delta\delta^{13}\text{C}$) is a composite based on differences between the surface-water $\delta^{13}\text{C}$ signal and a deep-ocean $\delta^{13}\text{C}$ record (Shackleton and Pisias 1985). Note the phase offsets between the records, especially at 140 and 210 kbp. (Curves represent 3 point smoothing of raw data.)

Could the same explanation apply to the Great Basin series; i.e., is the Great Basin series dated correctly but offset from the global ice volume record due to local processes? Since modeling studies indicate that the Great Basin was probably influenced by global ice volume fluctuations (Kutzbach and Wright 1985), we may specifically hypothesize that the Great Basin record represents some combination of global forcing from ice volume and local forcing from some other processes.

Orbital insolation forcing represents one source of local variance in the Great Basin record. In this report we will examine this possibility by generating a modeled time series that is a linear combination of global (ice volume) and local (orbital

insolation) forcing, and then compare the synthetic with observed time series in order to determine level of agreement. The synthetic time series is generated with an energy balance model that was developed in the first year of this project (Hyde et al. 1989). We assume for the sake of simplicity that ground water variations are linearly related to temperature variations. This assumption has obvious limitations but can be justified if we adopt the position that it is first prudent to examine the system in a linear mode in order to isolate variance attributable to this pathway. After systematically addressing this element, we can then determine more objectively the role of nonlinearities.

C.2 METHODS

The modeled time series was generated by making 206 separate runs with the two-dimensional energy balance model of Hyde et al. (1989). This is a thermodynamic model that calculates geographic variations in temperature as a function of solar forcing, local albedo, and the local heat capacity (high if water, low if land). This model compares favorably with the NCAR GCM runs of Kutzbach and Guetter (1986) but is much faster to run (see Hyde et al. 1989). It is therefore relatively easy to produce a number of simulations for different time intervals in order to generate a time series. Note that this strategy represents an excellent application of the hierarchical modeling approach outlined in the Global Plan (Appendix A), as it is not possible to produce such a time series with more complicated models.

There were two main simulations: 1) generation of a 309,000 year time series (equals maximum age of the Great Basin record) by making 103 separate runs at 3000 year intervals, with local temperature variations due to changes in orbital forcing (Berger 1978); 2) generation of a separate 309,000 year series of temperature changes at Yucca Mountain that were due to changes in global ice volume.

In order to generate the second time series, we started with the SPECMAP ice volume record (Imbrie et al. 1984) for the last 309,000 years. This time series does not give us a geographic pattern of ice volume. We addressed this problem by first determining the temperatures at Yucca Mountain due to known ice extent at 9, 12, 15, and 18 kbp, comparing the ice extent with the SPECMAP ice volume signal over this same interval, and then scaling ice volume fluctuations prior to 18 kbp to the same geographic extent as for the known positions of the last 18,000 years. This approach has limitations, but we consider it useful as a first step in addressing a difficult problem.

After producing the two time series, they were linearly combined. Primary emphasis was placed on winter temperatures, because it is thought that ground water variations occur only during this season — any summer rainfall variations do not result in groundwater recharge because rainfall is almost immediately evaporated. Results were then compared with the digitized time series of Winograd et al. (1988).

The technique used to calculate the smoothed spectral and cross-spectral estimates follows Newton (1988). The data were preprocessed by first interpolating them onto a regularly spaced grid, with either two or three thousand year time steps, and then removing the mean. The spectral estimates were obtained through the direct discrete Fourier transformation of the auto- or cross-covariance functions of the data. Smoothing was done by convolving an averaging kernel with the covariance functions before performing the Fourier transformation. A Parzen averaging kernel was used and the desired degree of smoothing was controlled by varying the width of this function. We found that a width equal to 40 percent of the record length provided the best compromise between leakage reduction and peak resolution.

The width selection was based on the most favorable comparison of our spectrum to the deep-sea $\delta^{18}\text{O}$ spectrum of a previously published record (Imbrie et al. 1984).

C.3 RESULTS

C.3.1 Analysis of the Great Basin Record

Before doing the comparison with the synthetic time series, we analyzed the Great Basin record (Figure C.1) separately to determine whether it had any characteristics of a time series recording both global and local forcing. The spectrum (Figure C.3) has peaks at 114.0 KY (period in thousands of years), 16.3 KY, and 10.9 KY. Depending on choice of time series package, smoothing, and interpolation schemes, these numbers vary somewhat, but there are no indications of 41 KY or 23 KY power commonly found in any of the analyses of the geological time series.

Given the shortness of the time series, the 114 KY peak could be a manifestation of the orbital forcing period around 100 KY (a spectral analysis of the truncated 309 kbp SPECMAP time series resulted in a 124.5 KY peak, whereas an analysis of the entire 780 kbp series shifts the peak to 97.9 KY). Peaks around 10 KY and 15-16 KY have been found in a number of other time series (e.g., Molfino et al. 1984, Pokras and Mix 1987, Crowley and Parkinson 1988, Pisias and Rea 1988), but it is not known whether any of these peaks are in phase with each other. A cross-spectral analysis of the Great Basin and SPECMAP $\delta^{18}\text{O}$ records (Figure C.4) indicates that the records have a high coherence (0.88) only at 114 KY. The Great Basin record leads the deep sea record by almost a quarter phase (about 28,000 years) at this period.

SPECMAP AND GREAT BASIN SMOOTHED SPECTRAL DENSITY

FIGURE C.3. Spectrum of the Great Basin Record and SPECMAP. Tick marks refer to dominant orbital periods (100, 41, 23, 19 KY).

SPECMAP AND GREAT BASIN SMOOTHED CROSS-SPECTRAL DENSITY

FIGURE C.4. Cospectrum of Great Basin and SPECMAP Records. Tick marks refer to dominant orbital periods (100, 41, 23, 19 KY).

The net conclusion from this part of the study is that there is at best only weak support for the hypothesis that the Great Basin record may be some combination of global and local forcing, for its spectral signature is significantly different than other geological time series (i.e., lack of power at 41 KY and 23 KY that usually occurs in these records).

C.3.2 Comparison of Modeled and Observed Time Series

The modeled time series (insolation only and insolation-plus-ice) are shown in Figure C.5, and a comparison with the observed time series is illustrated in Figure C.6. From this latter figure it is evident that there are significant differences between the model and observations. Comparison of the differences between these two time series (Figure C.7) yields a residual which clearly has a pattern to it. If our approach is at all valid, it suggests that the missing "element(s)" (i.e., residual) which may reconcile the two time series involves processes that primarily peak during glacial periods.

In order to obtain further insight into why the models and observations disagree so strongly, we conducted a spectral analysis of the residual and compared it to the original SPECMAP time series (Figure C.8). The residual has spectral peaks at 124.5 KY, 22.6 KY, and 8.6 KY. The cospectrum (Figure C.9) is highly coherent with the SPECMAP record at 124.5 KY (coh=0.95), 22.6 KY (coh=0.91), and 41.5 KY (coh=0.91). Given the shortness of the time series, these are all essentially the three dominant Milankovitch periods. The phase offsets for the different peaks are:

1. 124.5 KY — residual lags by 34.1° (about 12,000 years)
2. 22.6 KY — residual lags by 29.0° (about 2000 years)
3. 41.5 KY — residual lags by 39.2° (about 4500 years).

WINTER YUCCA MOUNTAIN LINEBM TEMPERATURES W/WO ICE

FIGURE C.5. Synthetic Time Series of: (top) Insolation-Only January Temperatures at Yucca Mtn.; (bottom) Insolation-Plus-Ice January Temperatures at Yucca Mtn.

FIGURE C.6. Comparison of Insolation-Plus-Ice Synthetic Time Series with Observed Time Series from Winograd et al. (1988).

FIGURE C.7. Residual of Two Time Series in Figure C.6 Compared to SPECMAP $\delta^{18}\text{O}$ Time Series.

SPECMAP AND
WINTER YUCCA (ICE) - GREAT BASIN
SMOOTHED SPECTRAL DENSITY

FIGURE C.8. Spectra of Residual and SPECMAP Time Series. Tick marks refer to dominant orbital periods (100, 41, 23, 19 KY).

SPECMAP AND
WINTER YUCCA (ICE) - GREAT BASIN
SMOOTHED CROSS-SPECTRAL DENSITY

FIGURE C.9. Cospectrum of Residual and SPECMAP Time Series.
Tick marks refer to dominant orbital periods
(100, 41, 23, 19 KY).

A note of caution is needed before evaluating the high coherence at Milankovitch peaks between the residual and SPECMAP. To a certain extent, the residual contains the SPECMAP signal embedded in it. The high coherence might therefore be a reflection of this embedded signal. We are analyzing the results further to assess this possibility, and the discussion below should keep this caveat in mind.

The cross-spectral analysis between the residual and SPECMAP therefore suggests that there may well be a third element that has to be factored into the equation before we can understand the Great Basin record. Whatever this variable is, it peaks during glacial periods, fluctuates at the dominant Milankovitch periods, and lags global ice volume.

C.4 SUMMARY AND CONCLUSIONS

The above work represents a first attempt to compare a modeled time series with observations of climate fluctuations in the Great Basin during the last 300,000 years. Our approach represents an excellent application of the hierarchical modeling approach to studying past climates. We know of no other method to generate such a time series.

We have generated a time series of winter temperature fluctuations at Yucca Mountain that are due to a linear combination of local and global forcing (i.e., orbital insolation and global ice volume). Comparison of this time series with an observed signal indicates some significant differences. Analysis of the residual suggests there may be an additional factor responsible for moisture variations at Yucca Mountain that we have not considered.

Allowing for the intermediate stage of analysis, we are presented with two conclusions of differing nature: 1) because the spectral signature of the Great Basin record is significantly

different than other geological time series, there may be little justification for attempting a reconciliation between the Great Basin and SPECMAP chronologies; one or the other may be in error. A number of recent independent studies support this possibility (Kukla et al. 1988, van den Bogaard et al. 1989, Li et al. 1989, Petit et al. 1990), 2) if we nevertheless persist in attempting a reconciliation, our ability to achieve it rests on identifying a third factor (Pacific sea surface temperatures?) which might be contributing to rainfall variations at Yucca Mountain. Because of the increasing probability of chronology errors, we have tabled further pursuit of this problem until the chronology picture is clarified.

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APPENDIX D

APPLICATION OF A TWO DIMENSIONAL NONLINEAR
ENERGY BALANCE MODEL TO THE GREENHOUSE PROBLEM

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APPENDIX D

APPLICATION OF A TWO DIMENSIONAL NONLINEAR ENERGY BALANCE MODEL TO THE GREENHOUSE PROBLEM

D.1 INTRODUCTION

As discussed in the Global Climate Studies Subtask Plan (Appendix A, this report) the three most important "drivers" of future climate change will probably be CO₂, ice volume, and "Milankovitch" forcing. Of these three drivers it is expected that CO₂ will dominate on short time scales (10^1 - 10^3) years, while the ice sheet and Milankovitch effects should dominate on medium to long time scales (10^3 - 10^5 years).

While the linear effect of each of these drivers is substantial, and subject to reasonably accurate analysis by LINEBM (which is an acronym for "Linear Energy Balance Model", see Hyde et al. 1989), they can be greatly amplified by nonlinear effects, particularly on regional scales. For example, Broccoli and Manabe (1987) find the effect of a CO₂ perturbation to be relatively uniform in space, except at the sea ice margins where albedo feedback greatly increases the effect. This is the well known "polar amplification" seen in many studies of CO₂ change (perhaps better named "subpolar amplification").

Thus in forecasting future climates there is a need for models which incorporate nonlinear effects. In line with our hierarchical approach to the study of climate, as defined in our Global Climate Studies Subtask Plan, the first such model should be as simple as possible, subject to the constraints that it must include an important nonlinear effect (which we take to be albedo feedback), and must be complex enough to be able to offer regional

forecasts. Such a model can be most efficiently constructed by adding a seasonally varying albedo to LINEBM.

In this appendix we will demonstrate that the nonlinear equilibrium model, NEQMOD, performs at least as well as LINEBM in simulating the current climate and that it has a more realistic sensitivity to external forcing. We shall then apply the model to the problem of the climatic response to anthropogenic CO₂ change.

D.2 MODEL PHYSICS

Unlike its linear predecessors, the nonlinear model determines the surface albedo interactively by a snow feedback process. The energy balance equation is formally the same as in Hyde et al. (1989), except that the coalbedo is now time dependent:

$$C(\vec{r}) \frac{\partial T(\vec{r}, t)}{\partial t} + A + BT - \nabla \cdot [D(\vec{r}) \nabla T] = Q a(\vec{r}, t) S(x, t) \quad (1)$$

where $\vec{r} = (\theta, \phi)$ is a point on the earth's surface

$C(\vec{r})$ = the local heat capacity per unit area

$D(\vec{r})$ = the local diffusion constant

Q = one quarter of the solar constant

$S(x, t)$ = the distribution function for isolation

$x = \sin(\theta)$ and

$a(\vec{r}, t)$ = the local coalbedo.

The coalbedo consists of a quadratic background coalbedo for ice free areas,

$$a(x,t) = a_0 + a_2 P_2(x)$$

where P_2 is the Legendre polynomial of order two, and of snow (a_I) and sea ice (a_{SI}) coalbedos where those surface types are predicted.

We have assumed the transport of heat to be diffusive, and that the infrared emission to space can be parameterized as a linear function of surface temperature, i.e.,

$$IR = A + BT$$

where A and B are determined from satellite observation (Short et al. 1984). As in North et al. (1983) the heat capacity is taken to be a triply valued step function, very large over the ocean, smaller over sea ice, and very small over the land, in the ratio of approximately 120:20:1. Values for model parameters are listed in Table D.1.

The principal difference between the two models is that in the nonlinear model the coalbedo field at time t is determined by a "critical temperature" snow rule:

$$a(\vec{r},t) = \begin{cases} a_0 P_0 + a_2 P_2, & \text{if } T(\vec{r},t) > T_c(\vec{r}); \\ a_I, & \text{otherwise,} \end{cases} \quad (2)$$

where T_c is the temperature below which snow is assumed to exist and a_I is the coalbedo for snow. In essence, the snow rule relates the temperature to the surface type (snow covered or snow free) thereby affecting the absorptivity (coalbedo) of the surface. The $\vec{r}$ dependence of T_c allows us to account for the effect of elevated areas on snow formation. Since the EBM's predicted temperatures are sea level temperatures, at any point on the earth's surface the critical temperature for snow formation is increased by 6.5°

per kilometer of surface height. Thus snow falls earlier, and melts later, over elevated areas such as Tibet or the Rocky mountains — which is an advance in realism over the strictly sea level model.

It may at times be desirable not to use this snow rule. For example, the large thermal inertia of the Greenland and Antarctic ice sheets ensures that they will remain as high albedo surfaces for centuries, even if the global average temperature were to rise dramatically. In simulations of this sort of climate one may invoke an option which fixes the albedo of specified areas regardless of temperature. Thus, even though the model is an equilibrium one, we may want to examine the nonequilibrium climate which is likely to exist over the next millennium (Crowley 1990).

The critical daily average temperatures for the formation of snow and sea ice were taken to be two degrees centigrade lower than their melting points, i.e. -2°C and -4°C , respectively. Clearly, no one temperature can perfectly mimic the complex processes inherent in the creation and melting of snow and sea ice so these temperatures are to some extent free parameters. Subject to the condition that $T_{c(\text{sea ice})}$ must be two degrees lower than $T_{c(\text{snow})}$ (because of the lower freezing temperature of sea water) we have examined possible critical temperatures in the region 0° to -5°C and found that $T_{c(\text{snow})} = -2^{\circ}\text{C}$ gave the best fit to the observations. The snow and sea ice in the model have no persistence — both melt immediately whenever the critical temperature is exceeded. As will be shown later, this approximation causes certain difficulties. Note that albedo feedback is the only nonlinear feature of the model. We do not allow the heat capacity to vary even though the ocean/sea ice border moves.

In the linear model, equation (1) was solved by an eigenfunction expansion in space and time, so that no integration was required. However, since in the nonlinear case the spatial-temporal structure of $\alpha(\vec{r},t)$ is not known a priori, a change in the

method of solution is necessary and there are two ways of doing this. The most direct approach to this problem is to abandon the solution by expansion in time harmonics and solve the model by integrating directly in time, as is done (for example) in all general circulation models. The disadvantage of such a method of solution is that the oceanic component of the model requires fifteen to twenty years to come to equilibrium with an imposed change in forcing or boundary conditions (e.g. a CO₂ change). The advantage of a time marching model is the obverse of this disadvantage — with such a model it is possible to study the transient effects caused by an imposed change.

The experiments described in this report utilize a more subtle approach — solving the nonlinear problem by an iterative method similar to that described in the paper by Lin and North (1990). This method avoids direct time integration and preserves the highly efficient solution by expansion in time harmonics employed in North et al. (1983) For example, despite considerable effort devoted to optimizing a time-marching model, an equilibrium run requires about 600 minutes of VAX 8300 cpu time, as opposed to 40 minutes for an equilibrium cycle by the iterative model.

In this iterative model successive equilibrium temperature fields are computed based on the energy balance equation (1). Via a spectral/grid transformation, two-dimensional arrays of temperature can be constructed at several uniformly distributed times through the year from an initial temperature field given in terms of Fourier-spherical expansion coefficients. The grid points are the Gaussian quadrature points in the meridional direction (which are roots of the Legendre polynomial of order equal to the number of meridional points desired) and are uniform in the zonal direction. A corresponding three-dimensional (space and time) coalbedo field can be computed by applying the snow rule, equation (2). This array is then converted into Fourier-

spherical expansion coefficients of the coalbedo field and a new equilibrium temperature field corresponding to the new coalbedo field follows from equation (1). This procedure is repeated until convergence is achieved in the temperature field, with the global mean temperature serving as the convergence criterion. The number of iterations to convergence depends on the severity of the convergence criterion, but even if a strict condition is imposed (in the experiments described later in this paper iteration was carried on until the global mean temperature changed less than 0.00001 degree between iterations) less than twenty iterations will normally be required. Five iterations will usually converge to within 0.1°C of the solution. Near a bifurcation point the required number of iterations rises.

Treatment of Ocean Heat Transport

It is often the case that a simple model will neglect two competing physical mechanisms, thus producing a reasonable looking result. When a more sophisticated model incorporates only one of these mechanisms, the results will appear worse. For example, LINMOD (a slightly less flexible version of LINEBM) neglects both the thermal isolation of the Antarctic plateau from the ocean and its high albedo. As a result the semi-annual cycle is well modeled. LINEBM, by including the albedo effect, cuts the predicted semi-annual cycle in half, and does not agree so well with the data. A similar problem occurs when moving from LINEBM to NEQMOD. LINEBM neglects both albedo feedback and the effect of western boundary currents such as the Gulf Stream. The effect is to keep the North Atlantic relatively warm. NEQMOD, which includes albedo feedback but not the Gulf Stream, predicts a very cold North Atlantic [indeed, in the last ice age the Gulf Stream did not carry heat as far north (e.g. McIntyre et al. 1976) and the climate was very much as predicted by uncorrected (see below) versions of NEQMOD]. The effect of this imbalance is rather small on a global scale, and an unmodified NEQMOD can be used for paleo-

investigations, but as it seems reasonable to assume that the Gulf stream will be relatively unchanged for the next few millennia, we have attempted to include its effects in the model.

While diffusive energy transport cannot incorporate the localized but intense effects of western boundary currents, one can mimic the physics by adding a purely arbitrary, localized heat source (which we choose to be of Gaussian form) to the right side of equation (1). Figure D.1 is a representation of this heat source, which we have placed in the Norwegian Sea. The intensity of the source is comparable to the vertical heat flux in this region as estimated by Häkkinen and Cavalieri (1989), and Bunker (1976). This modification stops the "runaway albedo feedback" in the North Atlantic and greatly improves the fit of NEQMOD to observed temperatures.

D.3 FIT TO THE CURRENT CLIMATE

D.3.1 Seasonality

Figures D.2 and D.3 show plots of the annual and semi-annual cycles predicted by the linear and nonlinear models. In NEQMOD the annual cycle of temperature is slightly better represented than in LINEBM over Asia, while the increased sea ice heat capacity seems to have worsened it slightly over the Arctic regions of North America. While the semi-annual cycle is somewhat too large over central Asia, this is a great improvement on the gross underestimate of this cycle in earlier models. In particular, the 1°C contour now extends south to Tibet, in much better agreement with data (e.g. Hyde et al. 1989). The representation of the semi-annual cycle is also dramatically improved over North America, but the improvement over Antarctica is only slight. This improved representation comes from the fact that the snowline tends to follow the forcing, thus enhancing the winter cold. The globally and annually averaged sea level

temperature is 15.2°C. This is 0.5 degrees higher than the global temperature of 14.7, but this latter figure is a surface temperature average.

In Figures D.4 and D.5 we compare the model's predicted snow and ice boundary with data while in Table 2 we do the same for snow and sea ice areas. In both cases the dataset was assembled from Zwally et al. (1983), Foster (1986), and Parkinson et al. (1987). As the land/sea and snow/no-snow boundaries are only defined to within $\pm 2.5^\circ$, these figures are only approximate. From Figures D.4b and D.5a, it is apparent that the winter ice and snow line is well reproduced. In the northern hemisphere the model contour is generally too far north except in the Norwegian sea, where our modeled Gulf stream heat source is not sufficient to produce the required meltback, though it does improve the fit to data considerably. In the southern hemisphere the winter fit is also very good. Summer sea ice contours present a greater problem. In the Northern Hemisphere summer the area and general location of sea ice are predicted adequately, but the specific features of sea ice distribution are not well modeled. It has been suggested that the very large meltback of summer sea ice in the Antarctic is due to an increase in the vertical oceanic heat flux (Gordon 1981, cf. Crowley and Parkinson 1988). Without such physics, the model cannot simulate the annual variation in Antarctic sea ice. This area is evidently a problem for other modelers as well. According to Schlesinger and Mitchell (1985) the NCAR model with mixed layer ocean predicts too much Antarctic sea ice, the GISS (Goddard Institute for Space Studies) model too little (despite being too cold over the sea), while the GFDL (Geophysical Fluid Dynamics Laboratory) GCM is too warm in this area, although sea ice extent is not given.

D.3.2 Model Sensitivity

An important measure of a climate model's utility is its sensitivity to imposed change. This is usually measured by the climate sensitivity parameter, β , which is defined by

$$\beta = \bar{T}(1.01Q) - \bar{T}(Q)$$

where Q is the solar constant and $\bar{T}$ is the global annual average temperature. It is important to note that this is not $\frac{Q_0}{100} \frac{\partial \bar{T}}{\partial Q}$

because $\lim_{\Delta Q \rightarrow 0} \frac{\Delta \bar{T}}{\Delta Q}$ may be ill defined.

The linear energy balance models used in North et al. (1983) and Hyde et al. (1989) have a sensitivity of $\beta = 1.14$, while the sensitivity of our nonlinear model is $\beta = 1.52$. The sensitivity of the nonlinear model can be evaluated by comparing it to the sensitivities of other models for altered CO₂ levels. In our model a change in CO₂ concentration is represented by a change in the infrared parameter, A . Values for this change were obtained from a formula due to Kiehl and Dickinson (1987) which was a fit to the output of their radiative-convective model. A doubling of CO₂ levels is approximately the same as a 2% change in the solar constant. General circulation models with interactive clouds and mixed layer oceans respond to a CO₂ doubling with a global average warming of 1.9–5.2°C (Schlesinger 1989, Mitchell et al. 1989). Thus GCMs have an average β of about 2. It is clear from Table D.3 that, except for the result of Manabe and Bryan (1985) for 0.5XCO₂ the EBM sensitivity for changed CO₂ levels is quite comparable to that of GCMs. Thus the inclusion of snow-feedback has raised the EBM's sensitivity into the lower end of the range of GCM sensitivities. This is not a surprising result, considering the very considerable meltbacks in snow and sea ice in most doubled CO₂ GCM experiments (e.g. Manabe and Stouffer 1980).

D.4 RESULTS

Investigation of our NEQMOD results falls naturally into three topics:

1. a comparison of our doubled CO₂ results with the many GCM experiments of CO₂ doubling
2. a comparison of our model results for various CO₂ levels with the similar set of experiments conducted by Manabe and Bryan (1985)
3. examination of the model's response at the Yucca Mountain site.

These will be dealt with in the paragraphs which follow.

The change in global temperature for a doubling of CO₂ predicted by GCM and NEQMOD experiments is presented in Table D.4. It is rather clear from this table that the NEQMOD results are within the ensemble of GCM results for such experiments. The mean of the GCM results is 3.2°C and the standard deviation is 1.0°C, thus placing the NEQMOD result well within one standard deviation of the mean. Most results fall within the range of $3 \pm 1.5^\circ\text{C}$ estimated some years ago (National Research Council 1982). It is remarkable how little the ensemble has changed over the past few years.

Figure D.6 compares NEQMOD's predicted global temperature response to various levels of CO₂ with similar experiments performed by Manabe and Bryan (1985) for an idealized earth using the GFDL GCM. It is clear from this figure that in the range 2XCO₂ to 4XCO₂ the models are in essentially complete agreement. Manabe and Bryan did not perform 6XCO₂ or 10XCO₂ experiments, but their 8XCO₂ experiment predicts 81% of the NEQMOD response.

Figure D.7 shows NEQMOD's predicted summer and winter change in temperature at the Yucca mountain site for CO₂ levels in the

range 1X-10X. The changes at Yucca mountain are quite large for high levels of CO₂, suggesting enhanced evaporation at the site. This effect could be offset by equally large increases in sea-surface temperatures in the eastern North Pacific (Figure D.8). Such increases may well result in enhanced moisture transport into the southwest U.S. (cf. Appendix G). Resolving the relative importance of these opposing effects will be an important research goal during the remaining part of this project.

D.5 CONCLUSIONS

The purpose of this appendix is twofold — to illustrate that on large scales NEQMOD results fall well within the ensemble of GCM results, and to determine what predictions NEQMOD has for the Yucca site.

The first point seems to us to be sufficiently established. NEQMOD's response to doubled CO₂ is within one standard deviation of the mean of the GCM results, and the agreement with Manabe and Bryan's results for yet higher levels of CO₂ is also excellent. NEQMOD is thus established as a good indicator of GCM performance for different CO₂ levels. It can therefore be used to investigate a large number of possible future scenarios cheaply and efficiently. Selected members can then be examined in detail with the LRM or a full GCM.

NEQMOD does not, of itself, predict a strong seasonal component due to albedo feedback in the CO₂ warming at Yucca mountain. Even in the 10XCO₂ case the seasonal component is only one degree, while at 4XCO₂ and below it is negligible. It is possible, however, that further nonlinear processes contained in GCMs will amplify this component.

The pattern of the NEQMOD response at Yucca suggests that GCM experiments should concentrate on the range 2-4XCO₂, with 10XCO₂

examined as an extreme example. There seems to be little to be gained by examining the response at $6-8\times\text{CO}_2$.

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FIGURE D.1. Contour Plot of the Gaussian Heat Source Employed in NEQMOD to Simulate the Effect of the Gulf Stream. Contour Intervals are 25 W/m².

AMPLITUDE OF ANNUAL CYCLE (LINEAR MODEL)

FIGURE D.2a. Amplitude of the Annual Cycle in Surface Temperature Simulated by the Linear Model (in Degrees Centigrade). Contours shown are 1°C and from 2.5°C to 22.5°C at intervals of 2.5°C.

AMPLITUDE OF ANNUAL CYCLE (NONLINEAR MODEL)

FIGURE D.2b. Amplitude of the Annual Cycle in Surface Temperature Simulated by the Nonlinear Model (in Degrees Centigrade). Contours shown are 1°C and from 2.5°C to 22.5°C at intervals of 2.5°C.

FIGURE D.3a. Amplitude of the Semi-annual Cycle in Surface Temperature Simulated by the Linear Model (in Degrees Centigrade). Contours shown are at intervals of 1°C.

AMPLITUDE OF SEMI-ANNUAL CYCLE (NONLINEAR MODEL)

FIGURE D.3b. Amplitude of the Semi-annual Cycle in Surface Temperature Simulated by the Nonlinear Model (in Degrees Centigrade). Contours shown are at intervals of 1°C.

FIGURE D.4a. Modeled and Observed Ice and Snow Boundaries - N.H. July. The heavy solid line is the digitized record, the outer dashed line is the -2°C (snow) contour, and the inner dashed line is the -4°C (sea ice) line. Observed fields are digitized from Foster (1986), and Parkinson et al. (1987).

FIGURE D.4b. Modeled and Observed Ice and Snow Boundaries - N.H. January. The heavy solid line is the digitized record, the outer dashed line is the -2°C (snow) contour, and the inner dashed line is the -4°C (sea ice) line. Observed fields are digitized from Foster (1986), and Parkinson et al. (1987).

FIGURE D.5a. Modeled and Observed Ice and Snow Boundaries - S.H. July. The heavy solid line is the digitized record, the outer dashed line is the -2°C (snow) contour, and the inner dashed line is the -4°C (sea ice) line. Observed fields are digitized from Zwally et al. (1983).

SOUTHERN HEMISPHERE SNOW BOUNDARIES (JANUARY)

FIGURE D.5b. Modeled and Observed Ice and Snow Boundaries - S.H. January. The heavy solid line is the digitized record, the outer dashed line is the -2°C (snow) contour, and the inner dashed line is the -4°C (sea ice) line. Observed fields are digitized from Zwally et al. (1983).

FIGURE D.6. A Comparison of NEQMOD with the GCM Results of Manabe and Bryan (1985). The change in global annual average temperature for a given CO₂ concentration is plotted for each model.

FIGURE D.7. Predicted Increases in Temperature for the Yucca Mtn. Site Due to Various Levels of CO₂.

FIGURE D.8. Predicted Increases in SST at an Eastern North Pacific Site (35°N, 135°W) Upwind of Yucca Mtn.

TABLE D.1. Sources and Values of the Various Model Parameters. Values new to the present study are refinements of those used in Hyde et al. (1989).

<u>Model Parameters</u>		
<u>Symbol</u>	<u>Value</u>	<u>Reference</u>
A	203.3	North and Coakley (1979)
B	2.094	North and Coakley (1979)
C _w	9.7	North and Coakley (1979)
C _I	1.5	present study
C _L	0.08	present study
D ₀	1.05	present study
D ₂	-1.33	North et al. (1983)
D ₄	0.67	North et al. (1983)
a ₀	0.70	present study
a ₂	-0.09	present study
a _I	0.30	North et al. (1983)
a _{SI}	0.50	present study

TABLE D.2. Comparison of EBM-Predicted Snow and Sea Ice Areas (in Millions of Square Kilometers) with Data from Zwally et al. (1983), Foster (1986) and Parkinson et al. (1987). As boundaries between land and ocean are defined only to $\pm 2.5^\circ$, these figures are somewhat approximate.

<u>Areas of Snow and Ice Cover (model/data)</u>		
<u>Season</u>	<u>Land</u>	<u>Ocean</u>
N.H. Jly	4.8/3.4	5.0/6.0
N.H. Jan	41.9/42.0	14.7/10.8
S.H. Jly	13.6/13.7	20.5/13.0
S.H. Jan	13.6/12.2	10.2/3.1

TABLE D.3. Comparison of the EBM's Predicted Temperature Change for Various Levels of CO₂ with GCM Predictions.

<u>CO₂ Comparisons</u>		
ΔCO_2	EBM ΔT	GCM Δ
0.5x	-4.1	-11.8 ^a
-100 ppm	-2.4	-1.2 ^b
2x	2.7	1.9-5.2 ^c
4x	5.8	5.4 ^a
8x	8.8	7.3 ^a

^a Manabe and Bryan 1985.

^b Broccoli and Manabe 1987.

^c based on Schlesinger (1989), and Mitchell et al. (1989).

TABLE D.4. Comparison of the EBM's Predicted Change in Global Annual Average Temperature for a Doubling of CO₂ with Changes Predicted by an Ensemble of GCMs.

<u>Model Responses to Doubled CO₂</u>		
Model	ΔT	Reference
NEQMOD	2.7	This report
CCC	3.5	N. MacFarlane, pers. comm.
OSU	2.8	Schlesinger and Zhao 1989
GFDL	3.2	Manabe and Bryan 1985
GFDL	4.0	Wetherald and Manabe 1986
NCAR	3.5	Washington and Meehl 1989
NCAR	1.6	Washington and Meehl 1989
GISS	4.2	Hansen et al. 1984
UKMO	5.2	Wilson and Mitchell 1987
UKMO	2.6	Mitchell et al. 1989
UKMO	1.9	Mitchell et al. 1989

APPENDIX E

THE LOW-RESOLUTION MODEL (LRM):
A DESCRIPTION AND INTERCOMPARISON

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APPENDIX E

THE LOW-RESOLUTION MODEL (LRM):

A DESCRIPTION AND INTERCOMPARISON

A wide range of forcing configurations, physical processes, and feedback mechanisms must be explored to identify the climatological extrema at Yucca Mountain. General circulation models (GCM) exist which simulate the requisite forcing, physics, and feedbacks but they are unsuitable as exploratory tools due to the large number of runs needed and the high cost per run of these supercomputer class codes.

As described in Appendix A, ARC has adopted a hierarchical approach in which the results of numerous experiments, made with less cpu intensive models, are used to direct the focus of subsequent efforts made with models of greater complexity and larger computational requirements. At one end of the hierarchy are the energy balance models (EBMs) such as LINMOD and NEQMOD. These models require order 10 to 100 minutes of VAX cpu time per run. At the other end are the big GCM's, which may take 10 to 100 Cray cpu hours per simulation. Intermediate in complexity and efficiency is the LRM, a low resolution GCM, requiring order 1 cpu hour per run on a Cray XMP.

In the first part of this appendix we describe the evolution and capabilities of the LRM. This is followed by a comparison of the predicted seasonal temperatures and precipitation changes over the past 18,000 years as generated by the LRM and the NCAR Community Climate Model (CCM).

E.1 MODEL DEVELOPMENT

The model is a global, spectral, primitive equation model based on the Australian spectral model (Bourke 1974, McAvaney et al. 1978). It has five equally spaced sigma levels in the vertical and has a horizontal resolution triangularly truncated at wave number 10. Physical parameterizations for radiation, convection, large-scale condensation, and surface fluxes of momentum, latent energy, and sensible heat are included.

The initial code was written for data initialization of a numerical weather prediction model by G. Branstator of the National Center for Atmospheric Research (NCAR). The model was expanded to allow long-term atmospheric integrations for climate studies by B. Otto-Bliesner (Otto-Bliesner et al. 1982). This model version has been used in many sensitivity studies and is designated LRM0. To improve the model simulation without resorting to the use of higher horizontal or vertical resolution, the parameterizations for orography, snow cover, soil moisture, surface fluxes and condensational processes were refined (designated LRM1, Gallimore et al. 1986). Additionally, the specification of cloud cover for shortwave radiation was adjusted to give planetary albedo values in agreement with satellite data (designated LRM2, Metzger 1984). The current version of the model also uses an algorithm based on earth orbital theory and Milankovitch astronomical theory in the shortwave radiation calculation (Kutzbach and Gallimore 1988).

The atmospheric model can be coupled to a constant or variable depth mixed-layer ocean (designated, respectively, OLRM2 and VOLRM2, Gallimore and Houghton 1987) or a soil hydrologic budget model (designated SLRM2, Gallimore and Kutzbach 1989). A chart summarizing the model development is given in Table E.1. References for the model development and use are included in the References section. Comparisons of the LRM to other models can be found in Kutzbach and Guetter 1984, and Crowley 1988.

E.2 MODEL IMPROVEMENTS

Model code streamlining has resulted in an 18% shortening of the program code. To make the model more "user-friendly", the model code has been rewritten to incorporate all the variations into one version of the model (instead of a version for each variation) with the specification of the desired enhancements requiring simple changes to input parameters. In particular, the model code now allows simulation for any combination of the following cases:

Atmosphere-Ocean Coupling

1. atmospheric model - prescribed climatological sea-surface temperatures and sea ice, present-day or past climatic conditions
2. coupled atmosphere - constant mixed-layer depth (fixed at 50 meters) ocean model with predicted sea ice
3. coupled atmosphere - variable mixed-layer depth (prescribed seasonal and latitudinal variation between 40 meters and 150 meters) ocean model with predicted sea ice.

Surface Hydrology

1. fixed soil moisture varying between 0.035 for dry desert land and 0.25 for wet humid land
2. predicted soil moisture and runoff using hydrologic budget approach.

Orography

1. present-day orography
2. no mountains
3. any other prescribed mountain and/or land-sea configuration.

The new model code when run with a coupled variable-depth mixed-layer ocean and predicted soil hydrology takes 38 minutes of CRAY XMP computer time for a one-year seasonal simulation.

Long-term integrations have verified that the LRM can successfully reproduce much of the observed seasonal and interannual variation of the large-scale features of the present-

day atmospheric circulation and sea-surface temperature distribution (Otto-Bliesner et al. 1982, Otto-Bliesner 1984, Gallimore et al. 1986, Gallimore and Houghton 1987). Consequently, the applications of LRM to a wide range of climate sensitivity problems have been considered. The sensitivity of the atmospheric seasonal cycle to ocean conditions has been investigated (Otto-Bliesner and Houghton 1986). The model has been used extensively for the study of paleoclimates, where the verifying data base is relatively coarse (Kutzbach and Otto-Bliesner 1982, Kutzbach and Gallimore 1988, Gallimore and Kutzbach 1989, Kutzbach and Gallimore 1989). In addition, Kutzbach and Guetter (1984) have shown that the climate sensitivity of the LRM and the NCAR Community Climate Model (CCM) to orbital parameter changes for 9000 years BP are similar for the Asian-African monsoon region. A more extensive comparison of paleoclimatic simulations for 9000 and 18,000 years BP by the LRM and the CCM of the global distributions of temperature and precipitation is currently in progress .

A portfolio of programs that constitute the LRM Processor are available for computing and processing time-averaged results from the history data sets output by the LRM. In particular, these programs generate gridded values and plots of model variables with user-specified time and space averaging and allow comparisons between model simulations. As part of this project, the LRM processing programs have been streamlined and made "user-friendly". These versions are tabulated in Table E.2 and are being documented in the "Users' Guide for the LRM Processors".

Planned improvements to the model include a predictive snowcover/snowmelt package (currently being tested at the University of Wisconsin) and an enhanced radiation package (see Appendix F). Several model documents to aid the user and to satisfy Quality Assurance requirements are being written. These include: 1) "Description of the Low-Resolution General

Circulation Model (LRM)" which details the governing equations, physical parameterizations, and numerical algorithms defining the model, 2) "A User's Guide to the Low-Resolution Model (LRM)" that provides details of the model code and illustrates how to modify and run the model, and 3) "Documentation of the LRM Program Modules" which describes each model subroutine and is designed to be used in conjunction with the code itself.

E.3 INTRODUCTION TO THE LRM/CCM INTERCOMPARISON

A benchmark study has been done to compare the responses of the Low Resolution Model (LRM) and the NCAR Community Climate Model (CCM) to the altered boundary conditions of 18,000 and 9000 years BP (BP = Before Present). Attention is focused on the annual cycle of surface temperature and precipitation computed by the models. Similar comparisons of surface temperatures for the CCM and a linear two-dimensional Energy Balance Model (EBM) can be found in Hyde et al. (1989).

E.4 MODEL SPECIFICATIONS

E.4.1 Boundary Conditions

Both models are run with the earth's orbital parameters and insolation changes obtained from Berger (1978). At 18,000 BP, the solar radiation changes were minimal. At 9000 BP, when perihelion occurred in July, all latitudes experienced increases in radiation in July with the largest increases (10%) at high northern latitudes (Figure E.1). In January, all latitudes experienced decreases in radiation varying from -4% at high southern latitudes to -15% at high northern latitudes. These solar radiation changes resulted in increased seasonality in the Northern Hemisphere and decreased seasonality in the Southern Hemisphere at 9000 BP, compared to present and 18,000 BP (Figure E.2). For the Northern Hemisphere, July (summer) solar radiation was 8% greater at 9000

BP than at present and 18,000 BP, and January (winter) radiation was 8% less. For the Southern Hemisphere, solar radiation in January (summer) was decreased 6% at 9000 BP compared to present and 18,000 BP, and in July (winter) it was increased by 5%. The annual-average radiation change, both global and hemispheric, is near zero for the last 18,000 years.

The CLIMAP February and August data sets (CLIMAP Project Members 1981) interpolated to the appropriate model resolution are used for sea-surface temperature, ice extent, albedo, orography, and land-sea distribution at 18,000 BP. At 9000 BP, present surface boundary conditions are assumed for the LRM simulation, while the CCM simulations also incorporate a small residual ice sheet over North America.

E.4.2 Model Differences

Several model differences are pertinent to understanding the comparisons between simulations.

Resolution

The CCM uses a rhomboidal truncation at wavenumber 15, which gives a horizontal resolution of 4.5° latitude by 7.5° longitude, or approximately 700 km. The LRM uses a triangular truncation at wavenumber 10, which gives a horizontal resolution of 11.6° latitude by 11.25° longitude, or approximately 1300 km. This difference affects the detailed extent and height of the ice sheets, as well as the scale of the regional changes.

Insolation

The LRM uses annual cycle experiments, while because of cost considerations, the CCM uses perpetual month experiments. Zwiers and Boer (1987) note that perpetual month climates are usually more continental than their annual cycle counterparts. The model response to seasonally varying solar insolation anomalies will be exaggerated over the interior of land masses when held fixed for

perpetual month simulations. In addition, the lag of the response of sea-surface temperatures and sea ice extent behind solar radiation changes is artificially shortened in the CCM by assuming the CLIMAP February and August values are valid for the January and July simulations, respectively.

Averaging Interval

The LRM statistics are computed from a simulation with three complete seasonal cycles. The CCM statistics are from perpetual January and July simulations. The CCM annually-averaged results assume a dominant annual cycle for the parameters of concern such that January and July values averaged together can be assumed to give a valid estimate of the annual average. The annual cycle simulations of the LRM allow for contributions of higher frequency variations to the annual average. The semi-annual harmonic has been shown to account for a significant amount of variance for precipitation at tropical latitudes (Hsü and Wallace 1976, Otto-Bliesner and Houghton 1986).

Snow and Ice

The Northern (Southern) Hemisphere snowline is prescribed at 42.2° latitude in January (July) and 68.9° latitude in January (July) in the CCM. The albedo of land and sea ice varies from 0.5 to 0.8 depending on the surface temperature, snowcover, and wavelength and source of the radiation. The equatorward limit of snowcover is set to vary sinusoidally from 42° latitude on the 25th of January (July) to 70° latitude on the 25th of July (January) in the LRM. A parameterization for assigning snowcover to land at high elevations is also included. Over the permanent land ice of Greenland, Antarctica, and the glaciers, a surface albedo of 0.8 is adopted. The albedo of sea ice varies from 0.4 to 0.8 depending on the distance from the seasonal snowline.

E.5 SURFACE TEMPERATURE

E.5.1 Regional Variations

At 18,000 BP (Figure E.3), both models give similar patterns of cooling in July with decreases of 20-30°C confined to the ice sheets over North America and Europe and decreases of 2-5°C over the rest of the Northern Hemisphere land. In the Southern Hemisphere, the CCM predicts much cooler temperatures (20-30°C) at 18,000 BP than present over areas of sea ice and warmer temperatures (2-5°C) over Antarctica. The LRM, on the other hand, predicts cooling of 15-20°C over Southern Hemisphere sea ice and cooling of 0-10°C over Antarctica.

In January, the axis of greatest cooling at 18,000 BP extends from the sea ice regions of the western Atlantic through Europe in the CCM. The maximum cooling in the LRM is shifted westward to over the eastern United States and extending only into the northwest section of Europe. The LRM does not reproduce the CCM warming over Alaska, northern Canada and northeastern Asia. Kutzbach and Guetter (1986) suggest that a split northern jet in the CCM 18,000 BP simulation leads to advective warming of these regions. Similar to July, the CCM still simulates warming over much of Antarctica at 18,000 BP compared to present which is not present in the LRM simulation.

On an annual average (Figure E.4), the LRM predicts less cooling over land areas at 18,000 BP compared to present than the CCM at all but polar latitudes where the CCM simulations suggest warmer conditions at 18,000 BP. Over the oceans, both models give similar cooling at tropical latitudes because of the specified sea-surface temperatures. Greater cooling in the CCM than the LRM occurs at latitudes of new year-round sea ice at 18,000 BP.

The radiational excess in July at 9000 BP leads to increased surface temperatures (2-5°C) over the Northern Hemisphere continents in both the CCM and LRM except over the North American

ice sheet in the CCM and the snowcovered Himalayas in the LRM (Figure E.5). In January, small decreases in the surface temperatures are simulated over most land areas. The increases over northern North America and northeastern Asia are within the range of interannual variability of the models.

E.5.2 Hemispheric and Global Averages

Annual-average global surface temperature simulated by the LRM for 18,000 BP is 2.8°C lower than present (Table E.3). This compares to a surface temperature lowering of 3.9°C in the CCM and 3.5°C in the EBM. Much of the difference between the CCM and the LRM is related to the surface temperature changes over land areas. At 9000 BP, both models predict annual-average surface temperatures near modern values because of compensating seasonal solar forcing.

In the Northern Hemisphere, both models predict substantial decreases in the July surface temperature over land at 18,000 BP (Table E.4), although the LRM response is not as large (-5.6°C versus -7.0°C). At 9000 BP, the LRM simulates surface temperature over land 1.6°C warmer than present, while the CCM gives temperatures 1.2°C warmer than present. The enhanced warming in the LRM is related to the absence of the North American ice sheet in its simulations. Over the oceans, both models give a decrease of July surface temperatures of about 1.5°C at 18,000 BP compared to present with 9000 BP surface temperatures similar to modern.

In January, the Northern Hemisphere surface land temperature decreases by 3.8°C at 18,000 BP in the LRM versus 7.6°C in the CCM. Some, but not all of this difference is a result of the use of perpetual month simulations by the CCM. The 18,000 BP temperature anomalies over the Northern Hemisphere oceans, on the other hand, are -2.3°C in the LRM compared to -1.2°C in the CCM. At 9000 BP, the LRM gives land surface temperature decreases of 1.5°C versus CCM decreases of 0.5°C.

In the Southern Hemisphere, both models give colder ocean temperatures at 18,000 BP than present (Table E.5). Differences between models can be attributed to differences in the prescribed boundary conditions and the predicted surface temperatures over sea ice. Over Southern Hemisphere land areas, both models have cooling in January of $< 1.0^{\circ}\text{C}$ at 18,000 BP and 1.6°C at 9000 BP. The latter is due to the decreased radiative forcing during the Southern Hemisphere summer. In July (southern winter), both models predict a warming of the Southern Hemisphere continents of about 1.0°C at 9000 BP. At 18,000 BP, the CCM predicts a 0.9°C warming in surface temperature over the Southern Hemisphere, while the LRM predicts a 1.0°C cooling.

E.6 PRECIPITATION

E.6.1 Regional Variations

Both models suggest decreased precipitation (0 to -20%) in July over land and ocean areas at 18,000 BP (Table E.6). These decreases over land areas occur at mid-latitudes of both hemispheres (Figure E.6) in regions of lower surface and air temperatures. The CCM and LRM both predict enhanced precipitation over the tropical land areas in July at 18,000 BP in areas where the surrounding oceans are slightly warmer than at present. The polar regions have only small amounts of precipitation resulting in inflated percentage precipitation departures.

In January, the LRM predicts increased precipitation (1 to 3%) in the Northern Hemisphere and decreased precipitation (-6 to -16%) in the Southern Hemisphere at 18,000 BP compared to present. The CCM, on the other hand, predicts decreased precipitation over the Northern Hemisphere land area with increases over the adjacent oceans, with the polarity reversed in the Southern Hemisphere.

At 9000 BP, the LRM and CCM both simulate greatly increased precipitation in July over land areas, especially at low and

middle latitudes in the Northern Hemisphere, and decreased precipitation over the oceans compared to present (Tables E.6 and E.7, and Figure E.6). These changes are related to the enhanced 9000 BP radiational forcing of the sea-to-land monsoonal circulations. In January, with decreased radiational forcing at 9000 BP compared to present, there is less precipitation over land areas and more over ocean areas than at present.

E.6.2 Global Averages

The general trend of precipitation changes in the models is apparent in the global, annual averages (Figure E.7a). Both models have decreased precipitation (about -5%) over both land areas and the oceans at 18,000 BP. At 9000 BP, both models exhibit a slight increase (1 to 3%) in precipitation on an annual average compared to present.

The global, annual precipitation-minus-evaporation (P-E) anomalies for the LRM and CCM at 18,000 BP (Figure E.7b) are positive over the land areas due to decreased precipitation but even greater decreases in evaporation because of cooler temperatures. Over the oceans, the P-E anomalies for 18,000 BP are negative. This suggests an enhanced transport of water vapor from ocean to land areas at 18,000 BP compared to present.

The P-E anomalies at 9000 BP are much smaller due to cancellation of opposite sign seasonal departures.

E.7 YUCCA MOUNTAIN

Variability of the LRM 100-year simulated time series of temperature, precipitation, and P-E at grid point (36.5N, 116W) can be used to assess climatic changes at the Yucca Mountain site (Figure E.8). The mean surface temperature at this grid point in January is 3.2°C with a standard deviation of 2.0°C. Using a two-sided t-test, the bounds for the 10% significance level are -0.1°C

to 6.5°C. That is, there is a 90% probability that the simulated surface temperature for this grid point will fall in this range if the value is from the same sample. The simulated surface temperature for the 3-year 18,000 BP run is 3.4°C, which is not significantly different from the control run.

Similarly, one cannot conclude that the simulated precipitation of .150 cm/day and P-E of .114 cm/day at 18,000 BP are from samples different from the control. Note, that because of large interannual variability in the LRM when considering a single grid point, a long sensitivity simulation is necessary to evaluate anomalies due to changes in the forcing mechanisms.

E.8 CONCLUSIONS

At 18,000 BP when surface boundary forcing is important, the LRM and CCM give similar regional patterns of cooling, but the overall cooling in the LRM is about 1°C less than in the CCM. Corresponding to lower temperatures at 18,000 BP, both models also simulate decreased global precipitation, but evaporative changes result in positive P-E departures over land areas. At 9000 BP when radiational forcing changes predominate, the LRM and CCM have similar patterns and magnitudes of heating in July and cooling in January. The enhanced July monsoonal circulation leads to increased precipitation at low and middle northern latitudes.

Model (LRM) 100-year time series of temperature, precipitation and precipitation-minus-evaporation for a grid point near Yucca Mountain exhibit large variability making conclusions on changes of these parameters for 18,000 BP impossible. In addition, the control years (years 6-8) corresponding to the 18,000 BP sensitivity simulation are years of above average surface temperature and below average precipitation and P-E.

A set of one-year sensitivity studies with the LRM suggest that changes in the snowcover and surface temperature

parameterizations may bring surface temperature departures for 18,000 BP more in line with the CCM results. Longer simulations are needed to fully assess these model changes.

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FIGURE E.1. Solar Radiation Departures (Past Minus Present, in W/m) for July and January as a Function of Latitude and Time (18,000 BP to the present). The numbers in parentheses are the departures from the present expressed in percent. [Kutzbach and Guetter 1986]

FIGURE E.2. Solar Radiation (W/m^2) for July, January, and Annual-Average for the Northern and Southern Hemispheres. Values are for 18,000 BP to the present [from Kutzbach and Guetter 1986].

LRM JULY DIFFERENCE FIELDS (18000 BP - PRESENT)

FIGURE E.3a. LRM July Temperature Anomalies (°C) Between 0 and 18,000 BP Experiments. Contours are ± 0 , 2.5, 5, 10, 20, and 30°C.

CCM JULY DIFFERENCE FIELDS (18000 BP - PRESENT)

FIGURE E.3b. CCM July Temperature Anomalies ($^{\circ}\text{C}$) Between 0 and 18,000 BP Experiments. Contours are ± 0 , 2.5, 5, 10, 20, and 30°C .

LRM JANUARY DIFFERENCE FIELDS (18000 BP - PRESENT)

FIGURE E.3c. LRM January Temperature Anomalies (°C) Between 0 and 18,000 BP Experiments. Contours are ± 0 , 2.5, 5, 10, 20, and 30°C.

CCM JANUARY DIFFERENCE FIELDS (18000 BP - PRESENT)

FIGURE E.3d. CCM January Temperature Anomalies ($^{\circ}\text{C}$) Between 0 and 18,000 BP Experiments. Contours are ± 0 , 2.5, 5, 10, 20, and 30°C .

FIGURE E.4a. Annual, Zonal Mean Temperature Anomalies (°C) Between the 0 and 18,000 BP Experiments for the CCM and LRM over Land.

FIGURE E.4b. Annual, Zonal Mean Temperature Anomalies (°C) Between the 0 and 18,000 BP Experiments for the CCM and LRM over Ocean.

FIGURE E.5a. LRM July Temperature Anomalies ($^{\circ}\text{C}$) Between the 0 and 9000 BP Experiments. Contours are ± 0 , 1, 2.5, 5, 10, 20, and 30 $^{\circ}\text{C}$.

FIGURE E.5b. CCM July Temperature Anomalies ($^{\circ}\text{C}$) Between the 0 and 9000 BP Experiments. Contours are ± 0 , 1, 2.5, 5, 10, 20, and 30°C .

LRM JANUARY DIFFERENCE FIELDS (9000 BP - PRESENT)

FIGURE E.5c. LRM January Temperature Anomalies ($^{\circ}\text{C}$) Between the 0 and 9000 BP Experiments. Contours are ± 0 , 1, 2.5, 5, 10, 20, and 30°C .

CCM JANUARY DIFFERENCE FIELDS (9000 BP - PRESENT)

FIGURE E.5d. CCM January Temperature Anomalies ($^{\circ}\text{C}$) Between 0 and 9000 BP Experiments. Contours are ± 0 , 1, 2.5, 5, 10, 20, and 30°C .

FIGURE E.6a. July, Zonal Mean Precipitation Departures (%) over Land for the CCM and LRM at 18,000 BP Minus Present.

FIGURE E.6b. July, Zonal Mean Precipitation Departures (%) over Land for the CCM and LRM at 9000 BP Minus Present.

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FIGURE E.7a. Global, Annual Precipitation Changes Between the 0 and 9000, and 0 and 18,000 BP Experiments for the CCM and LRM. Precipitation departures (%) over land (top) and ocean (bottom).

FIGURE E.7b. Global, Annual P-E Changes Between the 0 and 9000, and 0 and 18,000 BP Experiments for the CCM and LRM. P-E changes (cm/day) over land (top) and ocean (bottom).

YUCCA MOUNTAIN JANUARY SFC TEMP

FIGURE E.8a. LRM 100-Year Simulated Time Series of Temperature ($^{\circ}\text{C}$), at Yucca Mtn. (grid point 36.5N, 116W). The mean is represented by solid horizontal line and bounds for the 10% significance level by dashed horizontal lines.

FIGURE E.8b. LRM 100-Year Simulated Time Series of (top) Precipitation (cm/day) and (bottom) P-E (cm/day) at Yucca Mtn. (grid point 36.5N, 116W). The mean is represented by solid horizontal line and bounds for the 10% significance level by dashed horizontal lines.

TABLE E.1. Low Resolution Model (LRM) Overview.

MODEL	PUBLICATIONS
LRM0	(Basic model) Otto-Bliesner 1980: Model description & basic validation Gamache 1981: First and second moment statistics Nase 1981: Cloud sensitivity and radiation Kutzbach 1981; 1983a,b: Paleoclimate application Kutzbach and Otto-Bliesner 1982: Paleoclimate application Otto-Bliesner et al. 1982: Model description & basic validation Holopainen and Otto-Bliesner 1982: Diffusion effects Otto-Bliesner 1984: Energetics Ni et. al. 1985; 1986; 1987a,b: Mountain effects
LRM1	(Improved parameterizations) Gallimore et al. 1986: Model modifications & validation
OLRM1	(First version with constant-depth mixed layer ocean) Metzger 1984: Model description & validation
LRM2	(New cloud specification) Otto-Bliesner & Houghton 1986: SST role in seasonal cycle
OLRM2	(Constant-depth mixed layer ocean) Metzger 1984: Model description & validation Gallimore & Houghton 1987: Ocean-atmosphere energy exchange Kutzbach & Gallimore 1988: Paleoclimate application ^a
SLRM2	(Interactive soil moisture) Gallimore & Kutzbach 1989: Paleoclimate application ^a
SOLRM2	(Interactive soil moisture & constant-depth mixed layer) Kutzbach & Gallimore 1989: Paleoclimate application ^a
VOLRM2	(Variable-depth mixed layer ocean)

^a Celestial longitudes used for short-wave radiation calculation.

TABLE E.2. Low Resolution Model (LRM) Processing Programs.

File	Function	Stored Location	Input	Output
Programs:				
GRDPMON	To convert history output data set to pressure level/grid format and desired units and to generate time averages for individual years.	/OTTOBLI/GRDPMON/ GRDPMON.IFT	History output file from model run	Time-average file
AVGSTD	To compute multiyear time averages and standard deviations for spatial fields.	/OTTOBLI/AVGSTD/ AVGSTD.IFT	Individual year time-average files from GRDPMON	Multi-year average file & std dev files
ZLAREA	To compute multiyear time averages and standard deviations for zonal and selected areal averages.			
POSTPROC	To provide printout of spatial, zonal, and areal averages of time-averaged variables and microfilm plots of spatial data.	/OTTOBLI/POSTPROC/ POSTPROC.FTN	Individual year file from GRDPMON or multi-year average & std dev files from AVGSTD	Print-out or Micro-fiche
CONRECSM	To generate microfilm or microfiche contour plots of variables--modified NCAR graphics subroutine used by POSTPROC	/OTTOBLI/POSTPROC/ CONRECSM.FTN	-	-
Data:				
DATALB	Model albedo data used by POSTPROC to specify land-ocean configuration.	/OTTOBLI/POSTPROC/ DATALB.DAT	-	-

TABLE E.3. Global, Annual Average Surface Temperature Changes (°C) from Present at 18,000 and 9000 BP for the LRM and CCM.

	18,000 BP minus Present	9000 BP minus Present
a) Land plus ocean		
LRM	-2.8	0.0
CCM	-3.9	-0.1
b) Land only		
LRM	-3.4	0.0
CCM	-5.1	-0.2
c) Ocean only		
LRM	-1.9	0.0
CCM	-2.4	0.0

TABLE E.4. Northern Hemisphere Surface Temperature Changes (°C) from Present at 18,000 and 9000 BP for the LRM and CCM.

	18,000 BP minus Present	9000 BP minus Present
a) July, land only		
LRM	-5.6	1.6
CCM	-7.0	1.2
b) July, ocean only		
LRM	-1.7	0.0
CCM	-1.4	0.1
c) January, land only		
LRM	-3.8	-1.5
CCM	-7.6	-0.5
d) January, ocean only		
LRM	-2.3	0.0
CCM	-1.2	-0.2

TABLE E.5. Southern Hemisphere Surface Temperature Changes (°C) from Present at 18,000 and 9000 BP for the LRM and CCM.

	18,000 BP minus Present	9000 BP minus Present
a) July, land only		
LRM	-1.0	0.8
CCM	0.9	0.9
b) July, ocean only		
LRM	-2.6	0.0
CCM	-4.3	0.0
c) January, land only		
LRM	-0.1	-1.6
CCM	-0.7	-1.6
d) January, ocean only		
LRM	-1.3	0.0
CCM	-1.8	-0.1

TABLE E.6. July Precipitation Departures (%) from Present at 18,000 and 9000 BP for the LRM and CCM; NH (Northern Hemisphere), SH (Southern Hemisphere).

	18,000 BP minus Present	9000 BP minus Present
a) NH, land only		
LRM	-3%	+23%
CCM	-12%	+18%
b) SH, land only		
LRM	-8%	+6%
CCM	-2%	+4%
c) NH, ocean only		
LRM	-13%	-13%
CCM	-5%	-4%
d) SH, ocean only		
LRM	-6%	-1%
CCM	-3%	-2%

TABLE E.7. January Precipitation Departures (%) from Present at 18,000 and 9000 BP for the LRM and CCM.

	18,000 BP minus Present	9000 BP minus Present
a) NH, land only		
LRM	+3%	-1%
CCM	-9%	-1%
b) SH, land only		
LRM	-16%	-3%
CCM	+3%	-21%
c) NH, ocean only		
LRM	+1%	+8%
CCM	+2%	+2%
d) SH, ocean only		
LRM	-6%	+4%
CCM	-2%	+8%

APPENDIX F

RADIATION/CLOUD PARAMETERIZATIONS
AND THE LRM/CCM HYBRID MODEL

by

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APPENDIX F

RADIATION/CLOUD PARAMETERIZATIONS AND THE LRM/CCM HYBRID MODEL

F.1 INTRODUCTION

Radiation is the only significant source of energy for the earth's climate system (atmosphere-ocean-biosphere) and is also the ultimate sink of energy. On the space and time scales of climatic fluctuations, many scenarios of climatic change are related to radiative mechanisms, such as the greenhouse effect and Milankovitch orbital theory. The proper representation of radiative processes in climate models is therefore important.

F.2 BRIEF DESCRIPTION OF PRESENT RADIATIVE CODE IN LRM

The original design of the Low-Resolution Model (LRM, see Appendix E) was based on two considerations. First, it needed to be complex enough to simulate the seasonal change of the present-day atmospheric circulation on global and large scales. Second, it had to be sufficiently efficient that numerous annual cycle simulations could be made without undue computing demands and cost. The radiation and cloud parameterizations incorporated into the model are designed to balance these two criteria. The radiative heating and cooling rates and cloud cover for the model are computed using simple, noninteractive schemes based on mean climatological data.

Solar-terrestrial radiation can be broken into two parts: solar or shortwave radiation with wavelengths less than 4 micrometers and terrestrial or longwave radiation with wavelengths greater than or equal to 4 micrometers.

The shortwave radiation absorbed by the model atmosphere is a function of latitude and time of year. The diurnal sum of incident radiation is apportioned equally throughout the day. The atmospheric albedo is calculated using effective cloud amounts derived from satellite data of planetary albedo. The absorptivity of the atmospheric gases and the apportionment of this absorbed radiation among the model layers are implicitly included by prescribing these parameters from climatological data with no seasonal variation. The surface albedo is defined as a boundary condition.

The longwave radiative cooling rates are derived from model temperatures and empirical longwave radiative emissivities calculated from climatological monthly mean humidities, cloudiness, and cooling rates. The effects of the atmospheric gases on the longwave radiation are implicitly included in these parameters. Cloud amounts are prescribed and are noninteractive.

The current radiation parameterization and cloud prescription are appropriate for simulations and sensitivity studies of the present-day climate and for past and future climate scenarios where no change in the atmospheric composition and cloud cover is a good approximation (see Appendix E for a discussion of the applications of the LRM to climate sensitivity problems). On the other hand, the current version of LRM cannot be easily used for investigating climate scenarios where either the atmospheric composition is different (for example, increasing atmospheric CO₂ content) or the cloud distribution is expected to change (see Ramanathan et al. 1989, for a discussion of the importance of cloud-climate feedback).

F.3 REQUIRED MODIFICATIONS

A necessary improvement to the model to meet the goals of this project is to incorporate a more sophisticated radiation

parameterization scheme and interactive clouds. The approach chosen is to adapt the radiation and cloud parameterizations of the NCAR Community Climate Model (CCM) to the LRM. The CCM has been shown to reproduce many key features of the present climate (Pitcher et al. 1983, Ramanathan et al. 1983) and gives reasonable results when used to simulate past and future climate scenarios (Barron and Washington 1984, COHMAP 1988, Washington and Meehl 1984, Covey et al. 1985). In addition, both the LRM and CCM are global, spectral, primitive equation models based on the same governing equations and using similar numerical algorithms, although the CCM has higher horizontal and vertical resolution than the LRM.

The NCAR CCM radiation and cloud routines are documented by Kiehl et al. (1987). The clear-sky radiative calculation is separated into two parts, shortwave and longwave. In addition to Rayleigh scattering, the shortwave component explicitly includes absorption by O_3 , H_2O , CO_2 , and O_2 . The longwave component considers absorption and emission by CO_2 , O_3 , and H_2O . The surface albedo includes solar zenith-angle dependence. Clouds are formed in the model interactively and may be of the convective or non-convective type. All levels permit clouds except the lowest one and the top two. If the model-simulated relative humidity exceeds a saturation value, clouds are formed. The fractional cloud cover depends on whether the clouds are convective or non-convective, and the clouds are assumed to be randomly overlapped in the vertical. Multiple reflections between cloud layers and between the clouds and the surface are included as is a zenith angle dependence for cloud albedo.

F.4 CONSTRUCTION OF THE LRM/CCM1 HYBRID MODEL

As it is, the LRM achieves a good balance between economy, resolution, and physical sophistication. However, to conduct LRM experiments with altered CO_2 levels it is necessary to alter the

LRM radiation package. The parameterizations of radiation, and clouds used in the CCM1, Ramanathan et al.(1983), are superior to the LRM and include the treatment of atmospheric gases. We have elected to graft the CCM1 radiation and cloud routines on to the LRM, thus forming a new hybrid model. The construction of the hybrid is detailed in the following sections.

F.5 THE LRM CODE

From a programmers point of view, the LRM program is approximately 5800 lines of FORTRAN code. It was originally coded for execution on the CYBER-750. Then as now, high speed random access memory (RAM) was limited and large programs relied on special programming techniques to enable them to run in limited space. The standard technique is to cycle values between RAM and the lower speed mass storage devices, i.e. a harddisk or a solid-state disk. Although the LRM is not a large program by present day standards, it was at the time and is coded accordingly. The LRM field variables are maintained at run time in a block of storage on a solid-state disk. These variables are not used directly but rather they are copied from the central area into working arrays. The working arrays are used in the computations, then the computed quantities are used to update the central storage area. The storage area is accessed by calling the system specific routines, LCMRD and LCMWT. LCMRD transfers data from the block storage into a local variable. LCMWT transfers data from local storage back into the block storage. The subroutines take, as arguments, an offset pointer into the block storage, the number of words to transfer, and the name of the local variable used in the transfer. The pointers are initialized in subroutine RESLCM and made available to the various subroutines via the FORTRAN common block LCMLOC. The local variables are almost always working arrays. The LRM also provides the subroutines which convert from grid to spectral or spectral to grid representations.

The modifications to the LRM are minimal. We simply eliminate calls to those original subroutines which handle physical processes and make a call, once per model day, to the binding routine RADPAT described below. RADPAT will use the LCMRD and LCMWT mechanism to fetch or store the LRM data.

F.6 THE NCAR COMMUNITY CLIMATE MODEL (CCM1)

The NCAR Community Climate Model has been evolving since 1981 when the basic adiabatic model was acquired from the European Centre for Medium Range Weather Forecasts in Reading, England. After the first suite of changes the code was frozen into the version CCM0B. Further modification, primarily in the radiation, cloud, and hydrological sections gave rise to the version designated CCM1. It is these radiation and cloud routines that we are borrowing for use with the LRM. A full description of CCM1 can be found in Williamson et al. (1987). The physical basis, algorithmic structure, and coding particulars for the radiation and cloud routines are documented in detail in Kiehl et al. (1987). In its present configuration, the CCM1 program uses a rhomboidal truncation of order 15. The resulting grid has 40 latitude points and 48 longitude points, or approximately 4.5° by 7.5°. The model has 12 vertical levels and uses timesteps of about 30 minutes. The program is fairly large. The radiation and cloud routines alone contain approximately 8700 lines of FORTRAN code.

The LRM and the CCM1 are similar in many respects. Both are global, spectral, multi-level, primitive equation models and both compute the physics terms in grid space. While the CCM1 does not use the same method of accessing the field arrays as the LRM, the relationship between the LRM and its field data and the relationship between RADCODE (the subroutines that make up the radiation and cloud routines of the CCM1 model) and its field data are analogous.

Like the LRM, RADCODE does not use the field data directly but rather it makes a copy and uses that in the computations instead. The field data are stored in the FORTRAN common array BUF. Access to a particular field is facilitated with a pointer mechanism. RADCODE maintains its own working storage array F which is also indexed with a system of pointers. Using the two sets of pointers the RADCODE copies data back and forth between the main model array BUF and the local storage arrays F.

In the CCM1, the RADCODE routines are invoked via a single call to subroutine PHYS which in turn calls the various routines that make up the parameterization (Figure F.1). Subroutine PHYS is called once for each latitude and thus RADCODE computes values along a line of longitude, one latitude at a time. The pointers used to index the F array never change but those used to access the BUF array must be augmented based on the latitude index.

F.7 RADPAT

Subroutine RADPAT (radiation patch) is approximately 2000 lines of FORTRAN code. It performs several tasks related to exchanging field data and computed quantities between the LRM main program and the RADCODE subroutines. To the LRM, RADPAT appears to be a subroutine which returns heating tendencies and surface fluxes once a day. To the RADCODE routines, RADPAT appears to be a main program which is providing input data sets and requesting some computations be done for each latitude. This is shown schematically in Figure F.2.

RADPAT accesses the block storage area maintained by the LRM by using the routines LCMRD and LCMWT and the pointers in common block LCMLLOC. RADPAT allocates and maintains a BUF array similar to that of the CCM1. Using a set of locally defined pointers RADPAT packs data in the order that RADCODE expects to

FIGURE F.1. Calling Structure for the Radiation and Cloud Routines from the CCM1 Model. The entire structure is invoked by a single call to subroutine PHYS. The routines are called in left to right order.

FIGURE F.2. Schematic Diagram of the Relationship Between the Low Resolution Model (LRM), the Solid State Disk, the Grafting Routine RADPAT and the Radiation and Cloud Routines of the CCM1 (RADCODE).

find it. RADPAT also provides unit conversions, spectral-to-grid, grid-to-spectral transformations and data reordering when needed.

Early in the conversion process the assumption was that the resolution of the RADCODE would be reduced to match that of the LRM. It turns out that there is considerably more involved than simply changing a few parameter values. The same is true of the LRM which is also "tuned" for its current resolution. Our solution is to interpolate between the two resolutions using 2-dimensional splines. Efficient single purpose coding will keep the additional overhead to a minimum.

F.8 INPUT DATA SETS

Physical constants, such as gravity, CO₂ concentration and the like, are initialized with assignment statements in subroutine RADINI. Most of the field data RADCODE uses are passed from the main routines via the buffering system. The rest are read directly from disk by RADCODE routines. These data include:

Long wave absorptivities. These are the total longwave absorptivities due to water vapor, carbon dioxide and ozone. These numbers were computed by RADCODE previously during the last full longwave calculation. Since RADCODE does not store data between calls the main buffers are used to hold these data until needed.

Cloud fractions. These are cloud coverage fractions at each layer interface multiplied by the cloud emissivity and are computed internally. (They are computed in the cloud routines and used in the radiation routines.)

Pressure on layer interfaces. Pressure on the sigma surfaces which are the layer interfaces are computed externally by main routines.

Water Vapor mixing ratio. Water vapor for each layer is computed in the main program and passed to RADCODE.

Surface and layer temperatures. Temperatures at the surface (ground) and at the layer midpoints are computed externally and passed to RADCODE.

Surface cover flags. Information about the surface type, land, ocean, snow cover, ice etc. is read from disk files accessed in the main program. These covers are updated in the main routines as well. This information is used in RADCODE to set surface cover flags.

Ozone mixing ratios. Zonally averaged ozone mixing ratios at twenty-six levels are read from disk by RADCODE routines. Appropriate vertical interpolation and monthly averaging is done in routines internal to RADCODE.

Strong and weak angle albedos. Albedo is introduced into the model via five disk files. There is a strong and weak zenith angle dependence for each of infrared and visible wavelengths. The fifth file contains a specification for each grid point indicating the fractional dependence on strong or weak angle.

F.9 OUTPUT DATA SETS

Longwave absorptivities. Total longwave absorptivities due to water vapor, carbon dioxide, and ozone are computed during full longwave computations only. The values are output to the main model buffers where they are stored for future reuse by RADCODE.

Net Longwave heating. The net heating rate in each layer due to longwave absorption is output by RADCODE for use in the LRM to compute temperature tendencies.

Net shortwave heating. The net heating rate in each layer due to short wave absorption is output by RADCODE for use in the LRM to compute temperature tendencies.

Surface Flux. Total longwave and shortwave downward flux at the ground is output by RADCODE for use in the LRM to compute surface temperatures.

F.10 FUTURE PLANS

Recently the hybrid model was ported from the XMP to the new NCAR YMP. Preliminary work put the code in the form required by the pre-processing and code maintenance routines there. The model compiles and is currently in the run-time debugging phase. Early testing of the hybrid will involve validation runs, comparisons with previous runs of the LRM, and any other cases deemed appropriate for acceptance testing. Following acceptance, runs with altered CO₂ concentration will begin.

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APPENDIX G

ESTIMATES OF PRECIPITATION POTENTIAL IN SOUTHERN NEVADA
BASED ON HISTORICAL PRECIPITATION RECORDS

by

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APPENDIX G

ESTIMATES OF PRECIPITATION POTENTIAL IN SOUTHERN NEVADA BASED ON HISTORICAL PRECIPITATION RECORDS

G.1 INTRODUCTION

Natural mechanisms of interannual variability of precipitation at the Yucca Mtn. site are of interest to the efforts of climate modeling projections for site characterization. Several large-scale atmospheric phenomena which affect regional climatologies are known to display a distinctive interannual variability and thus, it might be presumed, drive significant interannual variability of regional climate statistics. Schonher and Nicholson (1989) relate California precipitation variability to occurrences of El Niño/Southern Oscillation. Blasing and Fritts (1975) suggest a link between Pacific Coast precipitation and a particular circulation pattern, later identified by McGuirk and Douglas (1988) as particular to Sudden Stratospheric Warming (SSW). Enzel et al. (1989) present evidence suggesting that significant rainfall enhancements occurred over the Mojave Desert at approximately 400 and 3600 years ago (cf. Stine 1990). These two intervals represent cool periods of the last few thousand years (Crowley and North 1990). At these times even shallow lakes existed in the Mojave Desert. Enzel et al. further link these wetter periods to modifications in the large scale atmospheric circulation over the eastern Pacific Ocean and the southwestern U.S.

To what extent has the Yucca Mtn. precipitation record been affected by various atmospheric phenomena? Does Nevada precipitation variability reflect the West Coast variability associated with large-scale atmospheric mechanisms? In this paper

we conduct an examination of historic precipitation records of southern Nevada and the region of Yucca Mtn. for evidences of several atmospheric mechanisms which have been proposed to exert control over the interannual variability of southwest U.S. regional precipitation.

Several assumptions are made. First, the seasonal scale climate of Yucca Mtn. is assumed to be represented adequately by stations surrounding it within limited distances. These proxies for Yucca Mtn. will be adequate only if their precipitation signatures are all the same. In mountainous terrain, small distances can lead to large changes in microclimate. However, it is assumed that, even in this arid mountainous region, large scale forcing mechanisms (Enzel et al. discussed storm tracks; this paper considers others) will lead to commensurate annual scale precipitation modifications. A second assumption is that, at least what has happened in the historical past, is possible for the future; this hypothesis has been described in detail by Bryson (1975). This hypothesis gains stronger credence if atmospheric circulation mechanisms are related to observed climate shifts or anomalies.

Finally, it is assumed that for major climate shifts, the historical climate record (which exists for approximately 100 years at the stations considered) may not be representative. In other words, the climate of the past 100 years, and its particular suite of forcings, is a sampling from only one of the possible climate regimes, and projections from this record encompassing shifts to other regimes of forcings is not feasible.

The methodology for this paper is a three-step process: 1) Link the records for stations surrounding Yucca Mtn. to other, longer or higher quality records; 2) Relate both sets of records to known climate varying processes. The first step includes establishing a quantitative link between the precipitation at the

FIGURE F.1. Station Location

several stations used in this study. The second step includes a quantitative assessment of both the strength and the reliability of the link between precipitation and mechanism; 3) conclude by examining the implications of past changes for the future.

G.2 DATA

Although precipitation records are available for a number of microclimates in the vicinity of Yucca Mtn. (e.g., the Nevada Test Site), they are all of short duration. A number of precipitation records for sites in southern Nevada and California extend back to 1900 and even earlier. The stations chosen for this study are Santa Barbara and San Diego, in southern California; Carson City, Austin and McGill in central Nevada; and Beatty, Desert National Wildlife Refuge (Desert NWR) and Las Vegas. The last three are in the region of Yucca Mtn., southern Nevada. Figure G.1 displays the relative positions of the stations and the Yucca Mtn. site.

The data for San Diego and Santa Barbara are from the work of McGuirk (1982) and represent monthly precipitation from Bulletin (1926) and Climate Data Summaries of States. These records are continuous, having had missing data interpolated from the average of the five preceding and following years to provide a conservative correction with respect to any climate signal. Two advantages of the California records are the analysis and careful checking of the data quality of the coastal stations (McGuirk 1982) and, especially, the record length at San Diego (123 years).

The Nevada data collected for this study are from the same sources. However, for the Nevada data, missing monthly totals were not replaced, and annual (and cool season) totals for years without the full complement of monthly data were considered unavailable.

Not a single original record was continuous for its entire length. The most significant hiatus occurred at Las Vegas where the station location changed in 1938; since the original and moved records overlapped, it was possible to correct the older Las Vegas record — 1895 through 1937 — to an equivalent record more representative of conditions at the relocated site.

This study used annual and six-month cool season totals of precipitation almost exclusively. The six-month cool season is January, February, and March of the year cited plus October, November, and December of the prior year. Monthly data from the record were summed, and disparate record annual totals were assumed to be clerical.

Table G.1a summarizes the length of record for the sites chosen and located in Figure G.1. Table values are annual and cool season, separated by "/". (Note that the annual and cool season record lengths can be different; e.g., the six months of data centered on January 1st of a year may be available for computing a total, while the full 12 months of the same year are

TABLE G.1a. Station Precipitation Record Lengths and Descriptive Statistics. In each column: annual / cool season values. Cool season is October through March.

Station	Record Size	Record Mean (inches)	Sample Std. Dev.
Santa Barbara	105 / 104	17.66 / 15.60	7.84 / 7.21
San Diego	123 / 122	9.74 / 8.47	4.05 / 3.64
Austin	96 / 95	12.82 / 6.84	3.64 / 2.04
Beatty	69 / 67	5.02 / 3.21	2.51 / 1.59
Carson City	94 / 94	10.96 / 8.61	4.12 / 3.87
Las Vegas	83 / 83	4.32 / 2.35	2.03 / 1.45
McGill	99 / 97	9.59 / 4.44	3.04 / 1.76
Desert NWR ^a	47 / 47	4.33 / 2.46	2.19 / 1.35

^a Desert National Wildlife Reserve

not, and conversely.) Also presented in Table G.1a are station descriptive statistics — annual and cool season means and standard deviations. Cool season means were used to examine winter circulation features (at some, but not all stations, the precipitation record is dominated by winter processes). The record lengths vary from 123 years at San Diego to only 47 years at Desert NWR, due east of Yucca Mtn. Nevada annual means range from about 13 in. to the north (Austin) to less than 5 in. southeast of Yucca Mtn. (Las Vegas and Desert NWR); the standard deviations are typically 30-50% of the means. At least for the Nevada stations, the cool season precipitation is about 50% of the annual means, but somewhat more variable (30-80%) than the southern California stations.

G.4 THE LOCAL CLIMATE

The target site of Yucca Mtn. in southern Nevada is shown in Figure G.1. The primary moisture sources for the location are the Pacific Ocean, with trajectories from a relatively low region of the Sierras to the southwest of the Mojave desert, and from the Gulf of California northeastward up the Imperial Valley and north across the Mojave Desert. These moisture sources are restricted by very high mountains to the west (Sierra Nevadas and the Sierra de Juarez), smaller local ranges to the west (Spring and Providence Mountain Ranges, of which Yucca Mtn. is one peak), to the east by the vast expanse of the Rockies, and to the south by the San Bernardino Mountains. The relative heights of the mountains and the narrow range of moisture-bearing wind directions result in an arid climate.

The westward and southward mountain restrictions represent an upstream moisture sink for the Yucca region. However, southern California, on the upstream side of the mountain barriers is also a dry climate. An implication is that the climates of both regions are controlled to some extent by the same large-scale atmospheric processes.

G.5 STATION CORRELATIONS

Tables G.1b and 1c summarize the cross correlations between the observations at each station; Table G.1b is based on annual totals and Table G.1c on 6-month cool season totals. It was anticipated that the observations for at least the Nevada stations would be highly correlated. This assumption was true for the three southern stations, which correlated highly (0.64-0.85). However, the southern stations correlated poorly (0.04-0.41) with the more northern stations; in fact, the northern stations correlated equally poorly with each other as well (0.22-0.41).

TABLE G.1b. Cross Correlation Matrix Between Stations for Annual Season Precipitation.

	Desert NWR	McGill	Las Vegas	Carson City	Beatty	Austin	San Diego	Santa Barbara
Santa Barbara	.65	.30	.60	.46	.77	.24	.70	1
San Diego	.72	.23	.69	.29	.78	.22	1	
Austin	.40	.41	.33	.28	.44	1		
Beatty	.78	.36	.66	.20	1			
Carson City	.09	.26	.10	1				
Las Vegas	.82	.32	1					
McGill	.41	1						
Desert NWR	1							

These variations were not anticipated based on previous comparisons of regional precipitation throughout the southwestern U.S. (McGuirk 1982, Stidd 1967). The former, based on precipitation correlations for California coastal stations suggests typical correlation magnitudes of about 0.6-0.8 for stations separated by distances typical of those separating the Nevada stations, and given similar climate controls. This latter assumption obviously is not met in the arid desert and mountains of Nevada. Stidd (1967) demonstrated that the annual cycle of Nevada precipitation is controlled essentially by three features: topography, which will not introduce an interannual effect; winter Pacific storm tracks, which dominate the first eigenmode of precipitation; and, the "southwestern monsoon", which is described by the second and third eigenmodes. The Nevada stations should vary together in the Pacific winter control because all stations, except McGill, have high amplitudes in Stidd's leading eigen mode (72% of the variance). The southern and northern stations correlate poorly in eigen mode amplitudes in the "monsoonal" eigen functions (totaling 21% of the variance); thus, the correlation of

TABLE G.1c. Cross Correlation Matrix Between Stations for Cool Season Precipitation.

	Desert NWR	McGill	Las Vegas	Carson City	Beatty	Austin	San Diego	Santa Barbara
Santa Barbara	.57	.16	.49	.50	.57	.24	.70	1
San Diego	.66	.18	.67	.23	.72	.27	1	
Austin	.41	.27	.20	.29	.33	1		
Beatty	.73	.14	.64	.12	1			
Carson City	.04	.22	.32	1				
Las Vegas	.85	.11	1					
McGill	.33	1						
Desert NWR	1							

summer precipitation among all the Nevada stations should be considerably weaker.

This anticipation was determined not to be the case, however. It matters little whether annual totals or only 6-month cool season totals are used. At nearly all stations, the cool season cross correlations are actually smaller (but not statistically so) than correlations based on annual totals.

What is observed is that, for annual totals, precipitation at the three southern Nevada stations surrounding Yucca Mtn. correlate more highly with precipitation at San Diego (0.69-0.77) and Santa Barbara (0.60-0.67) than with the northern Nevada stations. Presumably the correlation is higher at San Diego because the mountains are lower between the Nevada stations and San Diego than the mountains toward Santa Barbara. The most important conclusion of these tables is that about half of the variance of precipitation in southern Nevada can be determined from a knowledge of precipitation of southern California coastal stations. This conclusion is true for both annual totals and cool season (winter-dominated) totals.

G.4 INTERANNUAL VARIABILITY OF PRECIPITATION

G.4.3 El Niño Variation

Schonher and Nicholson (1989) relate California precipitation to the occurrence of ENSO (El Niño/Southern Oscillation; these events are normally linked and referred to as ENSO, with El Niño associated with the ocean part of the event and Southern Oscillation to the atmospheric part). The most intense ENSOs (warmest water east of the dateline) are invariably wet throughout California; other classifications of ENSO provide more diverse results (even one ENSO year of drought). Their results include only 11 ENSOs which occurred after 1950.

Use of Quinn et al's (1987) ENSO climatology and the present longer precipitation records allows a more definitive correlation between ENSO and precipitation, at least for the more localized southern California and Nevada region. Rasmusson (1984) used Quinn's data [which later was documented by Quinn et al. (1987)] to classify ENSO intensity on a 1-4 scale, 4 being most intense. Table G.2a and 2b describe annual and cool season precipitation, respectively, for all available events of the Quinn et al. (1987) record. The first three columns give the means for all ENSOs (left), the strongest two ENSO classifications — 3 and 4 (center), and only the most intense ENSOs — 4 (right). The numbers in parentheses give the number of events in the sample. The fourth column lists climatological information for the entire precipitation record at each station.

Average annual ENSO precipitation is higher than mean precipitation at all stations. Cool season precipitation is poorly related to ENSO occurrence; although category 4 ENSOs have wetter cool

TABLE G.2a. Annual Precipitation Averaged for ENSO Events (see text for ENSO intensity). Precipitation in inches; numbers in parentheses give sample size.

	All ENSO Events	Moderate(3&4) ENSO Events	Intense(4) ENSO Events	Annual Climatic Mean
Santa Barbara	18.94 (45)	19.42 (31)	22.75 (13)	17.66 (105)
San Diego	10.83 (53)	10.88 (32)	12.22 (14)	9.74 (123)
Austin	13.33 (37)	13.07 (26)	13.98 (12)	12.82 (96)
Beatty	5.49 (29)	5.95 (17)	6.94 (9)	5.02 (69)
Carson City	11.55 (38)	10.83 (26)	12.94 (11)	10.96 (94)
Las Vegas	4.56 (32)	4.94 (20)	5.21 (11)	4.32 (83)
McGill	9.96 (39)	10.12 (26)	10.50 (12)	9.59 (99)
Desert NWR	5.18 (18)	5.76 (10)	6.12 (7)	4.33 (47)

TABLE G.2b. As in Table G.2a, Except for 6-month Cool Season.

	All ENSO Events	Moderate(3&4) ENSO Events	Intense(4) ENSO Events	Annual Climatic Mean
Santa Barbara	15.83 (44)	15.73 (30)	17.44 (13)	15.06 (104)
San Diego	8.38 (52)	8.35 (32)	9.05 (14)	8.47 (122)
Austin	7.02 (36)	6.67 (25)	7.44 (11)	6.84 (95)
Beatty	3.04 (28)	3.00 (16)	3.57 (9)	3.21 (67)
Carson City	8.15 (38)	7.16 (27)	8.39 (12)	4.12 (94)
Las Vegas	2.18 (33)	2.18 (21)	2.59 (11)	2.35 (83)
McGill	4.37 (38)	4.34 (25)	4.31 (12)	4.44 (97)
Desert NWR	2.63 (18)	2.61 (10)	3.02 (7)	2.46 (47)

seasons, McGill is drier than climatology. For the annual totals, the category 4 ENSO means all meet or exceed the 5% significance level, based on a normal distribution and using a simple t-test. The confidence band for the annual mean precipitation for all

ENSOs lies between 5 and 10% for all the stations. With only two exceptions (Austin and Carson City for category 3 and 4 only), increasing ENSO intensity is associated with increasing annual precipitation. The results are more mixed for cool season precipitation, except that the most intense ENSOs, category 4, always average wetter than all ENSOs together. Generally, the variation among the ENSO events, as measured by the standard deviation among ENSO events, is larger than the climatological standard deviations; the variance is typically 40% greater, or more, for ENSO classifications; however, based on a chi-square distribution (conservative for precipitation), only San Diego variance meets an F-test for significantly different variances at the 5% level. Thus, even though the precipitation seems to vary more among ENSO years than among years chosen randomly from climatology, this increased variability is not large in a statistical sense.

In summary, ENSOs are wetter and slightly more variable than average throughout southern California and Nevada. The stronger the ENSO event, the stronger the precipitation signal. Although the signal can be detected in cool season precipitation, both the length of the averaging period and its sequencing with the onset of ENSO effect lead to a much stronger response in annual precipitation totals. The signal is stronger on the California coast (San Diego and Santa Barbara) than near the Yucca region. In Nevada, the signal is larger and more persistent in the south, around the Yucca region, than it is in north central Nevada.

G.4.2 Southern Oscillation Variation

When Darwin sea level pressure is high, the SO index is low, by convention, and is termed a "low/wet" phase, with the low/wet referring to sea level pressure/precipitation in the central Pacific. Precipitation totals at the eight stations were averaged for years with positive Darwin pressure anomalies and for years with negative pressure anomalies (Darwin SLP anomalies from

TABLE G.3. Average Annual Precipitation Associated with a Southern Oscillation Indicator (Darwin sea level pressure). See text for explanation.

Annual Precipitation for Darwin SLP Anomaly Years

Station	All + Anomalies	All - Anomalies	+ ≥1mb	- ≥1mb
San Diego	11.39	9.36	15.03	8.23
Santa Barbara	19.58	18.19	25.09	16.54
Austin	13.69	11.43	14.08	12.32
Beatty	5.62	4.33	6.60	4.80
Carson City	11.42	11.36	11.48	13.80
Las Vegas	4.94	4.27	5.29	4.61
McGill	9.94	9.13	10.96	8.28
Desert NWR	4.53	3.74	5.62	3.40

Trenberth and Shea 1987). Two pressure thresholds were used:
 1) positive or negative sea level pressure anomalies; and,
 2) positive or negative anomalies exceeding 1 millibar (mb). If 9 months or more exceeded the monthly average pressure threshold at Darwin, the year was termed a positive or negative Darwin anomaly year. The results are summarized in Table G.3, according to the sign of the anomaly.

Precipitation is systematically greater for years when the Darwin sea level pressure anomaly is positive. The results seem to depend on amplitude: The precipitation differences between positive and negative SO phase are greater for anomalies in excess of 1 mb than for those years when the anomalies are required to be only greater than, or less than, zero. Again, when the Darwin pressure is high, central Pacific pressure is normally low, and anomalous convection normally appears in the central equatorial Pacific. Apparently, this low/wet relation extends to California and Nevada.

For strong SO events (anomalies greater than 1 mb), the peak to peak amplitude of the precipitation variation (between opposite SO phases) ranges between about 7 or 8 in. on the California coast to about 0.5 to 2 in. in Nevada; for all SO events the differences range between about 2 to 0.75 in. The sample is highly restrictive for strong events and somewhat restrictive for all anomalous years; it is difficult to find many years (January through December, the period of an annual precipitation record) when the pressure anomaly exceeded 1 mb for at least nine months. These averages in Table G.3 are based on 6-10 years for strong anomalies (varying slightly with station), except at Desert NWR where data existed for only 3 positive anomaly years and 5 negative anomaly years; for the less restrictive set (all positive anomalies or all negative anomalies) the sample size varied between 25-30 years for each set of anomalies. The number of years are different for each station because of both different record lengths and because of the occasional gaps of missing data. The statistical significance of the difference of the means for strong positive and negative SO events ranges between 1% for the California stations to 10% for most of the Nevada stations; note that Carson City is the only station drier during the low SO phase (and then, for only the strong anomaly case). The statistical significance for all anomalies is weak, ranging between the 5% level at San Diego, Austin and Beatty, to no convincing difference between positive and negative SO phases at Carson City and Desert NWR (primarily because of the shortness of the record). Because of the associated physical mechanism — the fluctuation of the Walker circulation — the precipitation differences are likely to be real. The signal, however, is small in a statistical sense, due both to record length and to the relative size of each station standard deviation of annual precipitation.

G.4.4 Extreme Values

How common is it to find extremely wet years near Yucca Mtn. or along the California coast? The statistical distribution of precipitation in dry climates is usually skewed right, with about 60% of the years drier than the mean, offset by a small number of excessively wet years. Only 6 years (out of the 716 station-years - that is, the summation of the records at the six Nevada stations plus San Diego and Santa Barbara) experienced positive annual precipitation anomalies in excess of three standard deviations; there was only one year in which precipitation exceeded four standard deviations. These are almost exactly the number of outliers expected from a normally distributed data set. Most of the large precipitation events are associated with ENSO. Table G.4a summarizes all observations in excess of two standard deviations. Over two thirds of the large anomalies occurred during ENSO; over 50% with the strongest ENSO classification. The periods 1941 and 1982-3 dominate Table G.4a, although Quinn et al. (1987) note that 1884 may have been among the wettest ENSOs ever observed; only the San Diego and Santa Barbara records extended back to this event in this calculation.

The extremely wet non-ENSO years were in no case the result of a single wet month. Six of these seven annual totals were made up of at least four extremely wet months (greater than 1.0 in.); Las Vegas had only three such months, but its mean precipitation is only 4.32 in. Except for Beatty in 1978, the wet months of these non-ENSO years were scattered throughout the year; at Beatty, the heavy precipitation was restricted to January through April. The excessively wet months occurred during the normal wet seasons at every station without exception.

Thus, most of the wettest years occur systematically during ENSO; however, over the duration of the records, these wet events occur no more frequently than would be expected of a random precipitation record with a normal distribution. Extremely wet

TABLE G.4a. Summary of Statistical Outliers (wet anomalies exceeding 2 standard deviations) for Annual Precipitation Totals Within the Total Record. Hyphen indicates periods before records became available, or missing data. Underlined values are intense ENSOs (4); wiggly underlined values are moderate ENSOs (3); letter N next to the date indicates no ENSO.

Year	SB	SD	Aus	Bea	Car	LV	McG	DNWR
1884	<u>38.8</u>	<u>27.6</u>	-	-	-	-	-	-
1891			<u>21.1</u>	-		-	<u>18.0</u>	-
N 1894				-		-	16.7	-
N 1897				-			17.2	-
N 1909	43.2		-	-	-			-
1914				<u>10.2</u>	-			-
1918					-	<u>8.4</u>	<u>16.2</u>	-
N 1931						8.4		-
1941	<u>41.5</u>	<u>24.9</u>		<u>11.9</u>		<u>10.7</u>		<u>8.7</u>
1943		17.7		<u>10.0</u>				
N 1950					24.2			
1965		<u>19.0</u>						
N 1971			20.7					
1975	-	-	<u>20.3</u>					
N 1978	-	-		10.8				
1982	-	-			<u>19.2</u>			
1983	-	-	<u>22.4</u>	<u>11.5</u>	<u>21.8</u>			
N 1984	-	-		-			-	10.4

years are not the result of a single catastrophic precipitation event in a single month, but are characterized normally by above average wetness over many months scattered throughout the year.

Table G.4b compares the increase of precipitation during extreme years with the average precipitation at each site. In

TABLE G.4b. Statistical Outliers as Percentage of Station Means.
 As in Table G.4a, but with precipitation converted to percentage of station means shown in Table G.1.
 Hyphen indicates periods before records became available, or missing data. Underlined values are intense ENSOs (4); wiggly underlined are moderate ENSOs (3); letter N next to the date indicates no ENSO.

Year	SB	SD	Aus	Bea	Car	LV	McG	DNWR
1884	<u>220%</u>	<u>283%</u>	-	-	-	-	-	-
1891			<u>165%</u>	-		-	<u>188%</u>	-
N 1894				-		-	174%	-
N 1897				-			179%	-
N 1909	245%		-	-	-			-
1914				<u>203%</u>	-			-
1918					-	<u>194%</u>	<u>169%</u>	-
N 1931						194%		-
1941	<u>235%</u>	<u>256%</u>		<u>237%</u>		<u>248%</u>		<u>201%</u>
1943		182%		<u>199%</u>				
N 1950					221%			
1965		<u>195%</u>						
N 1971			161%					
1975	-	-	<u>158%</u>					
N 1978	-	-		215%				
1982	-	-			<u>175%</u>			
1983	-	-	<u>175%</u>	<u>229%</u>	<u>199%</u>			
N 1984	-	-		-			-	240%

most cases extreme events result in a 100-200% increase over the annual mean. These results may represent an upper limit to historical variations in precipitation, as the same numbers have been obtained for centennial-scale climate fluctuations of the last few thousand years (Enzel et al. 1989, Stine 1990) and even

ice-age changes during the last glacial maximum (e.g., Benson and Thompson 1987). Thus, past changes in lake level may represent a more frequent occurrence of the types of events which represent extreme cases at the present time.

G.4.1 Winter SSW Variations

Little skill is available for forecasting yearly precipitation anomalies into the future. The best that can be done is to examine the statistical frequency of certain kinds of precipitation patterns and then to relate these patterns to the local station precipitation at the Yucca Mtn. region. McGuirk and Douglas (1988) showed that Sudden Stratospheric Warming (SSW) can affect precipitation anomalies for periods of up to two months; it may be that these anomalies are of sufficient magnitude to affect the entire year, or at least the six-month cool season. One problem for the present study is that Nevada is the weakest responding state to SSW, at least in temperature. Additionally, Diaz and Fulbright (1981), in an EOF decomposition of U.S. precipitation, distribute Nevada winter precipitation almost uniformly across three of the first four eigen functions. This distribution suggests that the factors controlling Nevada precipitation are not a simple structure.

Table G.5 presents the comparison of annual, cool season and winter precipitation over Nevada and southern California for normal climatology and 16 periods of SSW (SSW periods from Labitzke 1982, McInturff 1978). At San Diego and around Yucca Mtn., SSW precipitation totals are smaller for winter and annual totals. No signal at all exists for the cool season, which contains precipitation from three or four months which occur before the SSW event. None of the SSW vs. climatology differences are statistically significant. If a physical difference exists in precipitation during SSW events, it does not lead to a meaningful difference in precipitation at Yucca Mtn. To the extent that the Pacific circulation pattern described by Fritts et al. (1979) and

TABLE G.5. Precipitation Means for Stratospheric Sudden Warming Periods.

Station	<u>Full Record / SSW Years</u>		
	Annual	Cool Season	JFM
San Diego	9.74/9.01	8.47/9.19	5.18/4.48
Austin	12.82/13.48	6.84/7.60	3.92/4.02
Beatty	5.02/4.94	3.24/3.16	2.00/1.91
Carson City	10.96/11.07	8.61/8.95	4.92/4.88
Las Vegas	4.16/3.79	2.43/2.51	1.37/1.28
McGill	9.59/7.89	4.44/3.39	2.38/1.36
Desert NWR	4.33/4.16	2.46/2.67	1.29/1.24
SSW Years: 1952, '58, '63, '68, '70, '71, '73, '74, '75, '77, '79 '81, '84, '85, '86.			

discussed by Enzel et al. (1989) is associated with the major circulation disturbances of SSW, this event does not seem to affect precipitation consistently near Yucca Mtn.

G.5 CONCLUSIONS AND DISCUSSION

This paper relates precipitation fluctuations to specific atmospheric circulation changes and climate altering mechanisms, with specific application to precipitation change at Yucca Mtn. north of the Mojave Desert. The analysis included stations around Yucca Mtn., stations farther north in central Nevada, and stations along the southern California Coast. To the extent that fluctuations in the precipitation records are related to large scale weather or climate fluctuations, the possibilities of climate change can be estimated. As climate forcing becomes more extreme, inferences from the historical record will become less valid.

The most significant enhancer of precipitation at Yucca Mtn. documented in this study is El Niño. Annual precipitation increases near Yucca Mtn. by about 50% during strong El Niño. The majority of statistical outliers in annual precipitation, about 70%, are associated with ENSO. Individual strong ENSO show annual local anomalies of precipitation of 200%. If the climate shifted to a regime in which these strong ENSO occurred with much more regularity, the precipitation regime of the southwestern U.S. would shift dramatically. Quinn et al. (1987) report that current ENSO frequency is about one every 3.8 years over the last several centuries. If all of these events became intense ENSOs, or if the frequency of occurrence increased (a practical upper limit is probably one every two years in association with the SO), a major climate shift could occur. If such a dramatic change of ENSO forcing is possible, a near doubling of average annual precipitation at Yucca Mtn. would not seem unrealistic. The 1941 event resulted in 10-12 in. of annual precipitation across southern Nevada.

At the shortest scales (SO), the fluctuation in annual precipitation for southern Nevada stations varies between 0.7 and 2.2 in. for the strongest SO events. This fluctuation could amount to precipitation increases of about 50% during strong, low/wet phases of the SO. SO should not lead to any significant increase in average annual precipitation unless the effect became much larger; if the signal became large enough, eventually the dry signal would saturate (a station cannot get any less precipitation than no precipitation). This effect would be modest, at best, and is not highly realistic.

Wintertime precipitation enhancement, or decrease, due to stratospheric warming events, is not likely at Yucca Mtn., even though this mechanism is strong in other parts of the U.S. Enzel et al. (1989) have presented evidence that this may be an important mechanism over longer time scales. At best, estimates

from McGuirk and Douglas's (1988) analysis would limit this effect to about twice the monthly precipitation for the one or two months over which the phenomenon acts. Although a related, similar phenomenon (atmospheric blocking) can act longer, its ebb and flow is not now predictable. Even if this mechanism could operate (and this paper suggests that SSW-related blocking has not been significant over the record of study), it might occasionally add an inch or so of precipitation to the annual total.

In conclusion, this summary is not a prediction. It is an assessment of what is possible, based on what has been observed over approximately the last century. The possibility of forcing changes, or of different mechanisms altogether, has not been assessed. It is an assessment of how wet it can get in southern Nevada, based on occurrences of the last century.

This study is useful for the assessment of future climate change at Yucca Mtn. because it helps us identify the boundary conditions most favorable for increasing moisture transport into the region. Evidently, a combination of warm sea-surface temperatures with an atmospheric circulation pattern characterized by frequent low pressure systems moving in from the Pacific along the southern west coast (ENSO) represents an optimal configuration for enhancing precipitation at Yucca Mtn. Scenarios for future climate change should take account of these observations in assessing future likelihood of even more extreme climate perturbations.

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APPENDIX H

FORECASTING OF GLOBAL ICE VOLUME

by

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APPENDIX H

FORECASTING GLOBAL ICE VOLUME

H.1 INTRODUCTION

Over the past two million years there have been numerous glacial-interglacial climate fluctuations which have a time scale of 10^3 - 10^4 years. Recently it has become necessary to consider how future climate might fluctuate on the same time scales. Such inquiries have been stimulated by the need to identify "safe" burial sites for radioactive wastes, which may remain hot for tens of thousands of years. It is also of interest to inquire whether a future greenhouse warming could be offset by a new ice age. In this paper we discuss methodologies for developing forecasts that are based on two different time series models.

In Figure H.1 we display the time series of ^{18}O levels as measured from deep sea cores. The series consists of 220 points which correspond to time intervals of 2000 years. Note that the most recent observations are at the right of the horizontal axis. The levels of ^{18}O are used as proxies for total global ice volume (see Imbrie et al. 1984). Thus in forecasting this series we will be describing the future behavior of ice volume. Note that the values of the series are included in Table H.1.

Traditional analyses of these data have been based upon the assumption that global ice volume is determined to a large extent by the earth's orbital element changes (Hays et al. 1976, Imbrie et al. 1984), e.g. the eccentricity changes with a period of about 100,000 years, the obliquity (tilt of the axis) at about 41,000 years, and the perihelion precesses through the seasons at 22,000 years. Thus a statistical model consisting of the sum of sinusoids of these periods (and possibly others) plus correlated

noise has been thought to be adequate for describing the data. We will refer to this model as the strict periodicity model. In Section H.3, we will produce forecasts and probability limits based upon such a model.

In Section H.2, we investigate the validity of the strict periodicity model and show that a model based upon disturbed periodicities (i.e. cycles that vary in frequency and amplitude) appears to describe well the spectrum of the data. We will use this disturbed periodicities model in Section H.3 to produce forecasts as well.

H.2 SEARCHING FOR PERIODICITIES IN THE ICE DATA

In this section we consider the spectral analysis of the ice volume data using both the power spectrum (periodogram) and maximum entropy (autoregressive) spectral analysis. Let $y(1), \dots, y(n)$ denote the values of the series. Then the power spectrum of the y 's is defined to be the function

$$\hat{f}(\omega) = \frac{1}{n} \left| \sum_{t=1}^n y(t) e^{2\pi i(t-1)\omega} \right|^2 = \frac{1}{n} (a^2(\omega) + b^2(\omega)), \quad (1)$$

where $\omega \in [0, 1]$, and ω is interpreted as the number of cycles per sampling interval. In the case of the ice data for example, $\omega = 1/50$ refers to a cycle of length 100,000 years as the sampling interval is 2000 years. The power spectrum provides a decomposition of the time series and its sample variance

$$\hat{\sigma}^2 = \frac{1}{n} \sum_{t=1}^n (y(t) - \bar{y})^2 \quad (2)$$

into sinusoids (and variability due to them) as it can be shown for n even that (see Newton 1988, p. 16)

$$y(t) - \bar{y} = a(1/2)(-1)^{t-1} + 2 \sum_{k=2}^{n/2} [a(\omega_k) \cos 2\pi(t-1)\omega_k + b(\omega_k) \sin 2\pi(t-1)\omega_k] \quad (3)$$

$$\text{and} \quad \hat{\sigma}^2 = \frac{1}{n} \left[2 \sum_{k=2}^{n/2} \hat{f}(\omega_k) + \hat{f}(1/2) \right], \quad (4)$$

where $\omega_k = (k-1)/n$.

In Figure H.2, we display the power spectrum of the data. The periods corresponding to the three largest peaks are 102,400, 41,000, and 23,280 years, while the fourth peak (which is very small) is at 19,140 years. These periods all correspond to periods of orbital forcing (see Berger 1978). If we regress the ice data on a constant and sines and cosines of these four periods, we obtain a multiple correlation coefficient of $R^2=0.7134$ indicating that about 71% of the variability in the data is explained by these four sinusoids. In Figure H.3, we give the power spectrum of the residuals from the regression model. Notice that there is still a large peak in the region of the first peak in the original data. Thinking that we had improperly determined the period of the original peak, we repeated our analysis for a variety of periods, but found in each case that there was still a peak in the residual power spectrum. This led us to consider the possibility that rather than containing a strict periodicity, the ice data has a disturbed periodicity in the neighborhood of

~102,000 years. We investigated this possibility using maximum entropy spectral estimation.

The Maximum Entropy Method

The maximum entropy method can be described in a variety of ways. From the statistical point of view (see Newton 1988, p. 247 for example) it consists in finding a linear filter that when applied to a time series results in white noise output. This can be viewed as determining the order p and coefficients $\alpha(1), \dots, \alpha(p)$ of an autoregressive (AR) process X satisfying

$$X(t) + \alpha(1)X(t-1) + \dots + \alpha(p)X(t-p) = \varepsilon(t) \quad (5)$$

for a white noise series ε . The process is determined so that its properties agree as well as possible with those of the observed data $y(1), \dots, y(n)$. These properties include the autocorrelation function $\rho(u) = \text{Corr}(X(t), X(t+u))$ and spectral density function

$$f(\omega) = \sigma_X^2 \sum_{v=-\infty}^{\infty} \rho(v) e^{-2\pi i v \omega} = \sigma_\varepsilon^2 \frac{1}{\left| \sum_{j=0}^p \alpha(j) e^{2\pi i j \omega} \right|^2} \quad (6)$$

where $\omega \in [0, 1]$, and σ_X^2 and σ_ε^2 are the variances of $X(t)$ and $\varepsilon(t)$ respectively.

In Figure H.4 we display the autoregressive spectral density estimator for order $p = 36$, which was the lowest order for which the estimator had clear evidence of three peaks. In an attempt to further identify possible periodicities in the data, we can

investigate the zeros of

$$\hat{g}(z) = 1 + \sum_{j=0}^{\hat{p}} \hat{\alpha}(j) z^j \quad (7)$$

From its form, it is clear that peaks in the AR spectral density are caused by having complex pairs of zeros of $\hat{g}$ be close to one in modulus (see Gray and Woodward (1986) for more information on using the zeros of $\hat{g}$ for spectral estimation). These are often called poles of the filter in the engineering literature. Thus we can find the roots of the degree-36 polynomial $\hat{g}$, and investigate those that are close to one in modulus.

In Table H.2 we list the real and imaginary parts of the 36 zeros of $\hat{g}$ as well as the moduli of the zeros. Notice that the first two zeros are real, while the four pairs of roots numbered (29,30), (33,34), (31,32), and (27,28) are closest to one in modulus in that order. We can determine the frequency of the peaks for each of these four pairs by 1) constructing an AR(2) filter from the zeros, 2) calculating the spectral density for this AR(2) process, and 3) using the method described by Newton and Pagano (1983) for identifying peak frequencies in an autoregressive spectral density. It is important to point out that this method is not identifying strict periodicities in data, but rather the disturbed periodicities we referred to above. Notice that a sampled cosine curve $X(t) = 2\cos(2\pi(t-1)/k)$ repeats every k points. It can be modeled by the deterministic difference equation

$$X(t) - 2 \cos(2\pi/k) X(t-1) + X(t-2) = 0, \quad t \geq 2. \quad (8)$$

Further, (see Newton 1988, p. 156, for example) the process

$$X(t) + \alpha(1)X(t-1) + \alpha(2)X(t-2) = \varepsilon(t) \quad (9)$$

will have a disturbed periodicity of period γ if $\alpha(2) \neq 1$ and $\alpha(1) = -2\cos 2\pi/\gamma$.

In Table H.3, we give the coefficients and periods of the peaks for each of the four AR(2) processes. Notice that they do tend to agree with the periods in the sinusoidal regression model.

H.3 FORECASTING

In this section we consider using the two models derived in Section G.2 to produce forecasts for the ice data. We begin with the strict periodicity model. For a regression model

$$y = X\beta + \varepsilon \quad (10)$$

having autoregressive errors ε , it can be shown (see Goldberger 1962) that the minimum mean square error linear predictor $\hat{y}(n+h)$ of $y(n+h)$ given the data $y(1), \dots, y(n)$ is given by

$$\hat{y}(n+h) = x^T \hat{\beta} + w^T V^{-1} e, \quad (11)$$

where x^T contains the values of the independent variables for the value of y being forecast, e are the residuals from the regression, the matrix V is the $(n \times n)$ autocovariance matrix of the

autoregressive process, $\hat{\beta} = (X^T V^{-1} X)^{-1} X^T V^{-1} y$, and

$$w = \text{Cov} \left(\varepsilon(n+h), \begin{bmatrix} \varepsilon(1) \\ \vdots \\ \varepsilon(n) \end{bmatrix} \right). \quad (12)$$

Further, the forecast error variance can be shown to be

$$\begin{aligned} \text{Var}(\hat{y}(n+h) - y(n+h)) &= \sigma_\eta^2 + x^T (X^T V^{-1} X)^{-1} x - 2x^T (X^T V^{-1} X)^{-1} X^T V^{-1} w \\ &\quad - w^T [V^{-1} - V^{-1} X (X^T V^{-1} X)^{-1} X^T V^{-1}] w \\ &= \sigma_\eta^2 + x^T (X^T V^{-1} X)^{-1} x - 2x^T \hat{\beta}_{w|X} - \text{SSR}(w|X), \end{aligned} \quad (13)$$

where σ_η^2 is the error variance for the autoregressive process, while $\hat{\beta}_{w|X}$ and $\text{SSR}(w|X)$ are the least squares coefficients and sum of squares of residuals if we were to perform a weighted least squares regression of w on X .

In Figure H.5, we display the graph of the last 50 values of the ice data as well as forecasts and 95% probability limits for the next 50 values based on the strict periodicity model with periods 102,400, 40,960, 23,280, and 19,140 years with the error series modeled as an AR(7) process. This error model matches the peak in the spectrum in the residual series. The values of the forecasts and limits are given in Table H.4.

For an AR(p) time series X , it can be shown (see Newton 1988, p. 131) that the minimum mean square error linear forecasts of future values of the series are obtained using the defining difference equation with the ε 's set to 0, while

$$\text{Var}(X(n+h) - \hat{X}(n+h)) = \sigma_\varepsilon^2 \sum_{j=0}^{h-1} \gamma_j^2, \quad (14)$$

where the γ 's are the coefficients of the polynomial $1/\hat{g}(z)$. In Figure H.6, we display the graph of forecasts based on the AR(36) model discussed in Section H.2, while the values are also given in Table H.4.

Finally, in Figure H.7 we superimpose graphs of the forecasts based on the two methods described above. The forecasts are rather similar and have some similarity to earlier forecasts based on a simple climate model (Imbrie and Imbrie 1980).

H.4 CONCLUSION

The two models considered in this paper both appear to adequately describe the behavior of the observed data. The forecasts obtained by the two models are very similar in appearance. It would appear from a statistical point of view that there is at least one disturbed periodicity in the data, although it does not seem possible to conclude this definitively based solely on the data. Finally, from our forecasts, it can be seen that there is only a small chance that there will be an ice volume in the next 100,000 years that is equal in magnitude to that during the last ice age (10,000-20,000 years before the present), which was in turn one of the largest glacial events of the last two million years. However, there is a significantly larger probability of a large ice volume event approximately 60,000 thousand years in the future.

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FIGURE H.1. 220 Values of ^{18}O Measured at 2000 Year Intervals. The most recent values are at the right of the horizontal axis. Values at the top of the graph represent times of low global ice volume.

FIGURE H.2. The Power Spectrum for the Ice Data.

FIGURE H.3. The Power Spectrum for the Residuals from the Regression Model.

FIGURE H.4. The Maximum Entropy Spectral Estimator for Order 36 for the Ice Data.

FIGURE H.5. Forecasts and 95% Probability Limits Using Regression Model.

FIGURE H.6. Forecasts and 95% Probability Limits Using Autoregressive Model.

FIGURE H.7. Forecasts for the Two Models. Those based on the sinusoidal model are given by the solid curve while those for the AR(36) model are given by the dashed curve.

TABLE H.1. The Ice Volume Data. If read across, the data correspond to the plot from left to right as given in Figure H.1.

1.34	-1.49	-1.50	-1.17	-.69	-.25	.10	.34	.53	.68
.79	.91	1.08	1.27	1.46	1.64	1.77	1.77	1.66	1.51
1.35	1.19	1.07	.98	.90	.82	.73	.61	.46	.30
.15	.03	.07	.25	.43	.50	.38	.18	-.03	-.28
-.53	-.74	-.86	-.88	-.86	-.86	-.94	-1.07	-1.12	-.75
.04	.84	1.40	1.72	1.79	1.65	1.45	1.23	1.03	.91
.95	1.06	1.19	1.31	1.34	1.20	.93	.61	.30	.05
-.05	-.02	.05	.15	.31	.46	.47	.35	.19	.01
-.21	-.42	-.60	-.77	-.89	-.85	-.67	-.55	-.50	-.43
-.37	-.40	-.52	-.66	-.78	-.70	-.30	.20	.65	1.03
1.18	1.02	.68	.31	.02	-.12	.03	.31	.64	1.01
1.27	1.40	1.35	1.23	1.12	1.41	1.00	.88	.88	.99
1.17	1.62	1.62	1.31	.88	.42	-.11	-.62	-.83	-.71
-.47	-.27	-.18	-.12	-.11	-.15	-.25	-.40	-.60	-.85
-1.07	-1.30	-1.45	-1.58	-1.57	-1.41	-1.26	-1.26	-1.33	-1.32
-1.28	-1.35	-1.33	-1.05	-.51	.26	1.19	1.89	2.12	1.98
1.72	1.53	1.19	.73	.50	.51	.59	.69	.80	.96
.91	.80	.71	.52	.46	.47	.45	.48	.77	.98
.88	.69	.53	.24	-.22	-.66	-.93	-1.00	-.89	-.68
-.50	-.41	-.37	-.38	-.50	-.56	-.59	-.67	-.77	-.85
-.94	-.96	-.94	-.94	-.96	-1.02	-1.14	-1.38	-1.65	-1.78
-1.81	-1.58	-1.01	-.29	.44	1.02	1.41	1.74	1.91	2.09

TABLE H.2. The Roots and Their Moduli for the AR(36) Polynomial
Fitted to the Ice Volume Data.

Real Part of Polynomial Roots

1	-1.764	-1.201	-1.084	-1.084	-1.014	-1.014	-.958	-.958	-.728	-.728
11	-.592	-.592	-.426	-.426	-.205	-.205	-.018	-.018	.157	.157
21	.348	.348	.509	.509	.710	.710	.803	.803	.869	.869
131	.971	.971	1.002	1.002	1.122	1.122				

Imaginary Part of Polynomial Roots

1	.000	.000	-.151	.151	.374	-.374	-.613	.613	-.761	.761
11	.911	-.911	.968	-.968	1.017	-1.017	1.077	-1.077	-1.074	1.074
21	-.988	.988	.907	-.907	-.768	.768	.654	-.654	-.512	.512
31	.302	-.302	-.125	.125	-.204	.204				

Moduli of Roots

1	1.764	1.201	1.094	1.094	1.081	1.081	1.138	1.138	1.054	1.054
11	1.086	1.086	1.058	1.058	1.037	1.037	1.077	1.077	1.086	1.086
21	1.048	1.048	1.040	1.040	1.046	1.046	1.036	1.036	1.008	1.008
32	1.017	1.017	1.009	1.009	1.140	1.140				

TABLE H.3. Coefficients and Periods for Spectral Peaks for Each of Four AR(2) Processes Corresponding to the Pairs of Roots in Table H.2 that Are Closest to One in Moduli.

Pair	Modulus	$\alpha(1)$	$\alpha(2)$	Peak Period
(29, 30)	1.008	-1.709	.9838	22,616
(33, 34)	1.009	-1.966	.9816	101,642
(31, 32)	1.017	-1.878	.9674	40,167
(27, 28)	1.036	-1.498	.9322	18,417

TABLE H.4a. Values for Regression Model Forecast

Regression Forecasts										
1	2.07	2.02	1.96	1.92	1.83	1.67	1.38	.98	.50	.02
11	-.36	-.60	-.66	-.58	-.41	-.21	-.05	.03	.02	-.06
21	-.19	-.34	-.50	-.63	-.75	-.84	-.89	-.89	-.86	-.79
31	-.71	-.63	-.56	-.49	-.41	-.28	-.06	.24	.63	1.04
41	1.41	1.67	1.74	1.62	1.35	1.00	.68	.46	.39	.48
Lower Probability Limits										
1	1.79	1.51	1.24	1.04	.84	.61	.27	-.16	-.67	-1.16
11	-1.56	-1.80	-1.87	-1.80	-1.63	-1.43	-1.27	-1.19	-1.20	-1.28
21	-1.41	-1.56	-1.71	-1.85	-1.97	-2.05	-2.10	-2.11	-2.07	-2.01
31	-1.92	-1.84	-1.77	-1.70	-1.62	-1.49	-1.28	-.97	-.59	-.17
41	.20	.45	.53	.41	.14	-.21	-.53	-.76	-.82	-.74
Upper Probability Limits										
1	2.36	2.53	2.67	2.79	2.82	2.73	2.49	2.12	1.66	1.20
11	.83	.60	.54	.63	.81	1.01	1.17	1.25	1.24	1.16
21	1.03	.88	.72	.58	.46	.38	.33	.32	.35	.42
31	.50	.58	.65	.72	.80	.94	1.15	1.45	1.84	2.25
41	2.62	2.88	2.95	2.84	2.56	2.22	1.89	1.67	1.60	1.69

TABLE H.4b. Values for Autoregressive Model Forecast

Autoregressive Forecasts

1	2.21	2.25	2.23	2.24	2.20	2.14	2.01	1.84	1.62	1.34
11	1.06	.76	.46	.24	.12	.11	.18	.29	.39	.41
21	.33	.18	-.05	-.36	-.67	-.99	-1.25	-1.43	-1.49	-1.47
31	-1.42	-1.37	-1.33	-1.31	-1.29	-1.24	-1.14	-.97	-.71	-.38
41	-.03	.30	.57	.76	.85	.87	.84	.77	.70	.66

Lower Probability Limits

1	2.02	1.83	1.58	1.39	1.20	1.03	.84	.64	.40	.12
11	-.17	-.47	-.78	-1.00	-1.12	-1.14	-1.07	-.97	-.90	-.90
21	-1.00	-1.18	-1.43	-1.74	-2.06	-2.38	-2.65	-2.82	-2.89	-2.87
31	-2.83	-2.80	-2.77	-2.78	-2.79	-2.75	-2.66	-2.50	-2.24	-1.91
41	-1.56	-1.23	-.96	-.77	-.68	-.66	-.71	-.79	-.88	-.95

Upper Probability Limits

1	2.40	2.67	2.87	3.08	3.20	3.24	3.17	3.04	2.84	2.57
11	2.28	1.99	1.70	1.48	1.36	1.35	1.43	1.56	1.68	1.72
21	1.67	1.54	1.32	1.02	.71	.40	.14	-.03	-.09	-.07
31	-.01	.06	.12	.16	.20	.27	.38	.56	.82	1.15
41	1.50	1.83	2.10	2.29	2.39	2.41	2.38	2.33	2.28	2.27

APPENDIX I

ARE THERE ANY SATISFACTORY GEOLOGIC ANALOGS
FOR A FUTURE GREENHOUSE WARMING?

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APPENDIX I

ARE THERE ANY SATISFACTORY GEOLOGIC ANALOGS FOR A FUTURE GREENHOUSE WARMING?

I.1 INTRODUCTION

The future greenhouse warming threatens to be a climate change of very large magnitude, with CO₂ doubling estimates for global temperature increases ranging from 1.8-5.2°C (Mitchell et al. 1989, Schlesinger 1989). Other radiatively important trace gases should also have a significant climate effect (e.g., Dickinson and Cicerone 1986). In the quest to determine what climate might be like in the future, scientists have sometimes suggested or cited past warm periods as a frame of reference or even possible "analogs" for our future warming (e.g., Kellogg 1977, Webb and Wigley 1985, Hansen and Lebedeff 1987, Budyko and Sedunov 1988, Zubakov and Borzenkova 1988). Regional climates for these warm periods have been thought of as snapshot views as to how regional climate may vary in the future.

Although the above suggestion seems eminently reasonable, advances in our understanding of past and present climates have uncovered some significant problems in continued use of the term "analog" (cf. Schneider 1984). In this paper I review the time periods most often cited as analogs and argue that, although paleoclimate studies can contribute much valuable insight into mechanisms of climate change, continued efforts to identify warm time periods as analogs rest upon often unstated assumptions that are probably not valid. I therefore suggest that the future greenhouse warming may represent a unique climate realization in earth history. My conclusions are based on analysis of the time intervals most frequently cited as analogs: 1) the early Holocene

(6000 BP, years before present) and last interglacial (sometimes known as the Eemian, 120,000 BP); and 2) pre-Pleistocene warm periods — the Pliocene (about 3-4 Ma, million years ago), Eocene (~50 Ma) or mid-Cretaceous (~100 Ma). There are serious shortcomings in invoking any of these time periods as geologic analogs.

I.2 COMPARISONS WITH PLEISTOCENE WARM PERIODS

I.2.1 Early Holocene (6000-9000 BP)

The geologic literature is replete with discussions that the early part of the present interglacial and also the past interglacial were warmer than the present. However, extensive studies of the early Holocene (COHMAP 1988) suggest that the warmth may have been primarily seasonal in nature over much of the northern hemisphere land masses, that times of greater warmth were not always globally synchronous, and that much of the warmth can be explained by variations in the seasonal cycle of insolation at the top of the atmosphere — the Milankovitch effect (Figure I.1). Modeling studies (Kutzbach and Guetter 1986, COHMAP 1988) indicate that changes in the seasonal cycle of insolation can cause significant seasonal changes of land temperature (1-3°C) in continental interiors (e.g., Figure I.2).

Statements about early Holocene warmth sometimes assume that conditions were uniformly warm globally. However, the geographic extent of early Holocene warmth needs to be better documented. Save for a few regions, sea-surface temperatures (SST) do not seem to be significantly warmer than the present. There is clear evidence for temperatures 1.0-2.0°C warmer in the high latitudes of the southern hemisphere (Hays et al. 1976, Jouzel et al. 1987, Clapperton et al. 1989), but these changes could also be due to Milankovitch forcing. Although it is sometimes assumed that orbital forcing is out-of-phase between the hemispheres, this

statement is true only for precessional forcing (~20,000 year period) during times of low eccentricity, such as the last 80,000 years [during time of high eccentricity, precession response could theoretically dominate obliquity (tilt) response at high southern latitudes; cf. Short et al. 1990]. Since forcing at the obliquity period (41,000 years) is in-phase with the northern hemisphere, synchronous warming in the high-latitudes of the southern hemisphere warming may reflect this process. This point is illustrated in Figure I.3 by means of a simple climate model that has been used to calculate the surface temperature response to changes in orbital forcing at 80°S.

Climate model studies indicate that the same orbital changes that caused summer warming in the northern hemisphere were associated with winter cooling in these regions and also a 0.1-0.3°C mean annual cooling in tropical SSTs (Kutzbach and Gallimore 1988, Hyde et al. 1989). The winter cooling has been only marginally documented, perhaps because many proxy climate indices may be biased toward recording a summer signal. The tropical cooling is below the detection limits of geologic data, but because of the area involved it can affect estimates of global mean annual temperatures. Figure I.4 illustrates global mean temperatures over the last 18,000 years as calculated from two different modeling studies (Kutzbach and Guetter 1986, Hyde et al. 1989). Both studies show that even though Antarctic mean annual and northern hemisphere summer temperatures were warmer at 9000 BP, the global mean temperature was within 0.1°C of the present value.

In order to get higher global mean temperatures at 6000-9000 BP, the planetary energy balance would have to be modified, perhaps due to higher CO₂ levels, greatly reduced ice cover, or changes in clouds. Ice cores do not provide any strong evidence for large changes in either of the first two variables (e.g. Figure I.5), and the fluctuations of the latter are so

problematical that we cannot say anything with confidence. Unless there are some other means to change the planetary energy balance, it therefore does not seem that the global average annual temperature of the early Holocene was significantly different from the present.

To summarize, much of the Holocene temperature response may be explainable in terms of orbital forcing variations that in general had little effect on mean annual temperatures. Where increases in temperature were mean annual, changes could still reflect a localized circulation response to seasonal forcing. We may carry this analysis one more step to identify another potential problem with assumed greater warmth in the early Holocene: if global mean temperatures were warmer at 9000 BP, we have no explanation for the warmth. That is, the climate system would be even more sensitive to forcing than indicated by models. Finally, even if the warmth were mean-annual, it might still not be justifiable to use the interval as an analogy, for the seasonal cycle of insolation was clearly different from the present.

I.2.2 Last Interglacial (120,000 BP)

The above reasoning applies equally well to the last interglacial (cf. MacCracken et al. 1990). Seasonal changes in orbital forcing were greater than in the early Holocene, so climate models simulate larger summer warming in continental interiors (greater than 4°C over a large part of Eurasia; cf. Figure I.6). There also seems to have been some significant mean annual warming along the coast of northwest Europe (see summary in CLIMAP 1984) and Antarctica (Figure I.7). However, there is some evidence that these northern and southern hemisphere warming events were not synchronous (CLIMAP 1984). For example, southern hemisphere warming seems to have peaked very early in the interglacial, while in some areas North Atlantic warming was significantly later (Figure I.8).

SSTs exert a strong influence on global average temperatures. Taken as a whole SSTs for the last interglacial (Figure I.9) were not significantly different from the present (CLIMAP 1984) and there are even a few regions (eastern boundary currents of North Atlantic and North Pacific) where last interglacial SSTs seemed to have been below present values (Crowley 1981, Muhs and Kyser 1987). Regional changes in mean-annual SSTs may reflect changes in the ocean circulation. For example, changes in the thermohaline circulation (cf. Manabe and Stouffer 1988) can alter heat exchange between hemispheres, for at present export of cold North Atlantic Deep Water (NADW) across the equator is associated with import of approximately 1.0×10^{15} W of warm water from the South Atlantic to the North Atlantic (e.g., Hastenrath 1980). Since NADW production appears to have been less than the present during the last interglacial (Boyle and Keigwin 1985/1986), heat storage in the southern hemisphere should have increased. The above process might well apply to explaining some regional differences in SSTs between the present and past interglacials.

In some cases, annual changes in SSTs could also result from Milankovitch forcing. For example, in high latitudes increased summer insolation results in net mean annual increases in insolation. GCM studies suggest that such changes reduce sea ice cover, leading to mean annual increases in SSTs (Kutzbach and Gallimore 1988, Mitchell et al. 1988). Changes in cloudiness might also increase midlatitude SSTs (Mitchell et al. 1988). Finally, intense seasonal heating over land can lead to greater subsidence over the ocean (Prell and Kutzbach 1987), possibly causing a spinup of the wind-driven ocean circulation. Such a pattern was inferred for the North Atlantic gyre during the last interglacial (Crowley 1981) and could explain both increased transport of Gulf Stream waters into the Norwegian Sea (cf. Kellogg 1980) and greater equatorward transport along the eastern boundary current. This response would cause net warming and

cooling in these regions, respectively, thereby possibly explaining some of the regional differences in last interglacial temperature response described above.

Utilizing the same energy balance climate model that agrees so well with the NCAR model at 9000 BP (Figure I.4), calculated global mean temperatures at 125,000 BP are only 0.2°C warmer than the present. This conclusion assumes no significant changes in CO₂ — an assumption consistent with ice core studies indicating that last interglacial CO₂ levels (Figure I.7) were comparable to the Holocene (280-290 ppm), except for a brief interval (1000-2000 years) when they were slightly higher (~300 ppm; Barnola et al. 1987) but still ~45 ppm less than present values. In order to find times when it was significantly warmer than the present, we may therefore have to turn to the pre-Pleistocene (next section).

I.3 COMPARISONS WITH PRE-PLEISTOCENE CLIMATES

It is often stated in paleoclimate reviews (e.g., Crowley and North 1990) that climates were warmer during much of the 100 million year time period preceding the Pleistocene ice ages. This conclusion is based on evidence for displacement of fauna and flora into high latitudes and from ¹⁸O measurements of benthic (bottom-dwelling) foraminifera, which show a steady decline through time (Figure I.10). The decline has been interpreted in terms of decreasing bottom water temperatures. Although some of the above climate changes have also been interpreted in terms of seasonal and geographic redistributions of heat (e.g., Shackleton and Boersma 1981, Crowley et al. 1986), most discussions conclude that global mean annual temperatures were probably higher (an important caveat to this conclusion is that our knowledge of past tropical SST changes, which strongly influence global mean temperatures, needs to be better constrained). If we accept the above conclusion, we can use Figure I.11 to project back in time when we might find a suitable time period to examine as an analog.

I.3.1 General Considerations

Even if some pre-Pleistocene time intervals were warmer than the present, they still do not represent satisfactory analogs for future warming. This conclusion is based on two main lines of reasoning:

First, the time intervals being analyzed or proposed as "greenhouse analogs" presumably represent long time-averaged equilibrium climates with much reduced ice cover, whereas the future greenhouse perturbation will likely be out-of-equilibrium for a significant duration of its total effect (thousands of years). For example, polar ice sheets will still be present, at least during the initial stages of warming (e.g., first millennium). The non-equilibrium nature of the future climate reflects the fact that the atmosphere and mixed-layer ocean are being perturbed at a very high rate ($\sim 2-4^{\circ}\text{C}/\text{century}$). Yet the deep oceans and ice sheets, with response times one to three orders of magnitude longer than the atmosphere, are changing more slowly (e.g., Stouffer et al. 1989). The response time of the East Antarctic ice sheet is so long (perhaps tens of thousands of years, cf. Whillans 1981) that this ice sheet may only start responding to the greenhouse warming after the CO_2 perturbation has been washed out of the system by buffering with the deep ocean carbonate reservoir.

Because the future climate will be out of equilibrium with the greenhouse gas radiative perturbation, which itself will be continuously changing, we will therefore almost assuredly have the bizarre combination of warm atmosphere/mixed layer ocean and polar ice sheets — i.e., a non-glacial atmosphere with glacial-age polar regions (because of its intermediate response time, the deep ocean should be somewhere in between in its adaptation to changing boundary conditions). There may be only one time period when polar ice caps and higher CO_2 levels may have co-existed — the Late Ordovician, 450 million years ago (cf. Crowley et al. 1987).

Even here the sequence of events, time scale considerations, adequacy of geologic deposits, and continental positions are so different from the present that we cannot feel secure in citing this time as an analog.

Secondly, there were significant differences in geography that raise serious questions about the appropriateness of past warmer climates as analogs for the future. During the last 100 million years there has been a general drift of land into the high latitudes of the northern hemisphere and changes in continental areas (e.g., Figure I.12) which should have had a significant effect on atmospheric temperatures (Barron 1985, Crowley et al. 1986). Opening and closing of "oceanic gateways" has also probably significantly affected ocean heat transport. For example, ocean GCM calculations (Figure I.13) indicate that an open Central American isthmus may have greatly reduced poleward ocean heat transport and thermohaline overturn in the North Atlantic (Maier-Reimer et al. 1990). Finally, there has been uplift of major mountain ranges such as the Alps, Himalayas, and Rocky Mountains in the time interval under consideration (Ruddiman et al. 1989). Uplift of the mountains could have a significant effect on the atmospheric circulation (Kutzbach et al. 1989, cf. Mullen 1989). When all of these geographic effects are considered together, it is evident that, even if global temperatures were warmer, the regional climate patterns may have been significantly different than what we might experience in the future.

I.3.2 Application of General Arguments to Some pre-Pleistocene Warm Intervals

The above general statements can be applied to the most commonly cited warm intervals of the last 100 million — the Early Pliocene (~3-5 Ma), the Early Eocene (~50 Ma), and the mid-Cretaceous (~100 Ma). The Middle Miocene (~15-17 Ma) is another warm interval, but we have less information on global syntheses

from both land and sea for this time than for some of the other three time intervals.

Early Pliocene (~3-5 Ma)

The Early Pliocene warm interval preceded slightly the rapid onset of midlatitude northern hemisphere glaciation at about 2.5 Ma. Various geologic evidence indicates that conditions were perhaps at least seasonally 10-20°C warmer than the present in high northern latitudes (cf. summaries in Zubakov and Borzenkova 1988, and Raymo et al. 1990), Greenland and Arctic ice cover were probably reduced, and circum-Antarctic SSTs perhaps 6-10°C greater than present (Cieselski and Weaver 1974). Sea level may have been at least 30 m higher (cf. Haq et al. 1987) — a result consistent with inferred decreases in both Greenland and Antarctic ice cover (e.g., Webb et al. 1984, Funder et al. 1985)

Despite its relative closeness to the present, the Early Pliocene would differ from a future greenhouse warming in the following respects: 1) reduced ice sheet cover would not occur in the near future, 2) there may have been substantial uplifts in western North America and the Tibetan Plateau in the last few million years (Ruddiman et al. 1989), 3) the central American isthmus was just on the verge of closing (cf. Maier-Reimer et al. 1990), so there may have been some differences in ocean circulation response between the Pliocene and future warm periods (cf. Figure I.13). Thus, the criteria laid out as critiques for pre-Pleistocene analogs has been met.

Early Eocene (~50 Ma)

This interval is a particularly interesting time from an observational viewpoint. Temperatures were substantially warmer in high latitudes of both hemispheres (see summary in Crowley and North 1990) and bottom waters were ~15°C — results incompatible with extensive high latitude ice cover. Sea level was at least 75 m greater than the present, flooding a number of low-lying

regions, especially in North America and Europe. In addition, there was an extensive seaway south of Europe and Asia — the Tethys Sea (cf. Figure I.12). Perhaps the most astonishing feature of Early Eocene climates is the oxygen isotope evidence for nearly uniform values in low- and high-latitudes, and in the deep ocean (Zachos et al. 1990). If the oxygen isotope values can be translated into temperatures (a translation that requires independent validation), then latitudinal and depth temperature gradients in the ocean were only about one-fifth of the present; i.e., the Early Eocene ocean may have been nearly isothermal! There are some indications that CO₂ levels may have been greater (2-4X?) during this time (e.g., Lasaga et al. 1985, Owen and Rea 1985, Berger and Spitzzy 1988).

As with the Pliocene, a number of significant differences from the present mitigates the use of the Eocene as a future greenhouse analog. Again, we have the absence of high latitude ice, higher sea level, significantly lower orography (Ruddiman et al. 1989), and presence of an open central American isthmus (cf. Figure I.13). There are also some indications that the deep-water circulation may have been different in the Eocene (Barron and Sloan 1990), in that bottom water may have primarily formed in the subtropics due to excess evaporation. This conjectured bottom water, termed "warm, saline bottom water" (WSBW, Brass et al. 1982), would form in much the same way that Mediterranean Intermediate Water forms today (initial density of this water is greater than Antarctic Bottom Water; however mixing with ambient water masses rapidly reduces its density). Eocene WSBW may have formed as a result of some combination of higher CO₂ levels increasing evaporation in the subtropics and an enlarged Tethys creating more favorable sites for bottom water formation, especially in expanded shelf areas (cf. Figure I.12). Although formation of extensive WSBW needs more empirical support, it is clear that Eocene oceanographic conditions present some

fascinating opportunities for modeling (cf. Barron and Sloan 1990). However, it is unlikely that an Eocene oceanographic analog will be repeated in the future, because the existence of high-latitude ice should maintain deep-water formation in these regions. Less extensive shelf areas in the subtropics might also mitigate production of a future WSBW.

Mid-Cretaceous (~100 Ma)

The mid-Cretaceous is another time interval that has been the focus of many studies (e.g., Barron et al. 1981, Barron and Washington 1982, 1985). This was probably the warmest time interval of the last 100 million years, with little if any evidence for high-latitude ice. The general consensus is that CO₂ levels several times higher than the present (5-7x?) were responsible for Cretaceous warmth (Berner et al. 1983, Barron and Washington 1985). In the future such levels could be approached if we burned all available fossil fuels and decreased the rate of CO₂ sequestering in the ocean.

As with the other warm intervals, the general criticisms outlined for the pre-Pleistocene are applicable to the mid-Cretaceous: lack of extensive (or any) high latitude ice in the Cretaceous would be substantially different from the situation encountered during the early stages of a greenhouse warming. The significantly different geography (Figure I.12) would present boundary conditions for orographic and ocean forcing greatly different than what would occur in the future. Although it is important to better understand the origin and regional patterns of Cretaceous warmth, these conditions do not justify its use as an analog.

I.4 SUMMARY AND CONCLUSIONS

I.4.1 Summary

Examination of the paleoclimate record indicates that there may not be any satisfactory geological analogs for a future greenhouse warming. In terms of global mean temperature, Pleistocene interglacials may not have been significantly warmer than the present. If pre-Pleistocene intervals were warmer, changes in geography would almost certainly have created regional climates different than what might occur in the future. Finally, because future temperatures may be increasing at a very high rate (2-4°C/century), we will have the unique combination of warm atmospheres and polar ice sheets — a condition very different from the pre-Pleistocene warm periods.

I.4.2 "Time Period Analogs" vs. "Process Analogs"

It is important to note that, although there may be no individual geological time periods that are true analogs for future warming, paleoclimate studies may still give valuable insight into certain processes that may operate in the future. For example, Oerlemans (1982) has conjectured that a future warming might be associated with increased snow accumulation on Antarctica as a result of enhanced water vapor levels in the warmer atmosphere. Such a response is consistent with ice core observations on Antarctica, which demonstrate that Holocene snow accumulation rates were greater than during the last glacial maximum (Lorius et al. 1985). Likewise, conjectures about abrupt changes in future climate, due to instabilities in the ocean-atmosphere system (Broecker et al. 1985, Broecker 1987), receive some support from the geologic record (e.g., Dansgaard et al. 1984, Broecker and Denton 1989). Finally, geologic time periods provide an excellent opportunity to validate climate models against an independent set of boundary conditions. In the future, experiments with pre-Pleistocene time periods may provide very

sturdy tests of critical parameterizations in coupled ocean-atmosphere models (e.g., flux adjustments between the atmosphere and ocean — cf. Sausen et al. 1988).

I.4.3 Conclusions

The above arguments therefore suggest that there may not be any time periods that are satisfactory geologic analogs for a future greenhouse warming. The warming may represent a unique climate realization in earth history. From a geologic perspective, we will be sailing into the future on uncharted waters. Thus, although paleoclimate data can be very valuable in their own right and can be used to validate atmosphere and ocean models, we should be careful about overemphasizing their usefulness when attempting to construct "snapshot" scenarios for future warming.

I.5 ACKNOWLEDGMENTS

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FIGURE I.1. Solar Radiation Departures (Past-Minus-Present, in W/m^2) for July and January as a Function of Latitude and Time (18,000 BP to Present). The numbers in parentheses are the departures from present expressed in percent. [From Kutzbach and Guetter 1986]

FIGURE I.2. Effect of Altered Radiation Field at 9000 BP on Surface Temperatures. Figure illustrates July temperature differences (9000 BP minus present) as calculated by an energy balance model. July 9000 BP was warmer than the present in northern midlatitudes. Similar results are obtained from GCM runs (cf. Kutzbach and Guetter 1986). [From Hyde et al. 1989]

FIGURE I.3. Simulation of Mean Annual Temperature Changes in the High Latitudes of the Southern Hemisphere (80°S) over the Last 30,000 Years. Calculation is from a two-dimensional energy balance model and refers to changes due only to orbital insolation forcing (cf. Short et al. 1990). Note that the magnitude of the temperature range is probably less than would occur in more complicated models (cf. Short et al. 1990), that the calculated temperatures do not include the effect of topography, and that the temperature peak in the early Holocene is in phase with Northern Hemisphere warming and consistent with ice core measurements from the same location (Jouzel et al. 1987).

Global Annual Average Temperature Departures

FIGURE I.4. Comparison of Energy Balance Model and NCAR General Circulation Model Simulations of Global Mean Annual Temperature Departures from 18,000 to the Present. Even though seasonal temperature changes at 9000 BP can be significantly different than the present (Figure I.2) global average temperatures at 9000 BP are essentially the same as present. The same is true for the last interglacial (=cross; 128,000 BP was chosen because this is the time of maximum departures of solar insolation forcing - cf. Figure I.1). [9000 BP results from Hyde et al. 1989, 128,000 BP results from this study]

FIGURE I.5. Atmospheric CO₂ Concentration in Byrd (Antarctica) Ice Core. Early Holocene levels (290 ppm) are only about 10 ppm higher than pre-industrial levels (cf. Friedli et al. 1986). [From Neftel et al. 1988]

FIGURE I.6. Departures (Experiment Minus Control) of Surface Temperature (T), Winds (Arrows), and Precipitation (P) for the Indian Ocean Sector for July 126,000 BP, Illustrating the Effect of Increased Summer Insolation on T and P. [After Prell and Kutzbach 1987]

S. HEMISPHERE CLIMATE RECORDS

FIGURE I.7. Comparison of High Latitude Southern Hemisphere Records for the Present and Last Interglacial with the Ice Core CO_2 Record from Vostok, Antarctica. Both the Southern Ocean SST and ice core temperature record indicate warmth greater than the present at the last interglacial. However, the warmth was apparently not synchronous with some of the best documented examples of warmth in the northern hemisphere (cf. Figure I.8). Note that CO_2 values are comparable for both the present and past interglacial. [SST record from Martinson et al. 1987, ice core deuterium record from Jouzel et al. 1987, CO_2 record from Barnola et al. 1987, ice core chronology plotted according to revisions of Petit et al. 1990]

FIGURE I.8. Comparison of the Timing of Peak Warmth During the Last Interglacial in the North Atlantic and Southern Hemisphere. North Atlantic record from Ruddiman and McIntyre (1981), southern hemisphere record from Martinson et al. (1987). Peak warmth does not occur synchronously in the two cores, a response that may reflect redistributions of heat by the ocean-atmosphere system. Note that this separation in timing may not be as dramatic in all records.

120 K FEBRUARY Δ SST

FIGURE I.9. Difference Between Modern Sea-Surface Temperature and Estimated February Sea-Surface Temperature (in $^{\circ}\text{C}$) at the Last Interglaciation 120,000 Years Ago. Negative values mean that the last-interglacial ocean was colder than today. Most SST values are similar to present; any spatially coherent increases in one region are more or less offset by equivalent decreases in other areas. [From CLIMAP 1984]

FIGURE I.10. Oxygen Isotope Record over the Last 100 Million Years. The record has been interpreted primarily to indicate past changes in deep water temperature. It is assumed that very warm deep waters are incompatible with extensive ice cover in polar regions, a conclusion consistent with other geological data from high latitudes (cf. Crowley and North 1990). [After Douglas and Woodruff 1981]

FIGURE I.11. Schematic Comparison of Future Greenhouse Warming with Past Changes in Temperature. Note that Pleistocene climate oscillations are idealized and that interglacial temperatures were probably not significantly warmer than present (see text). Pre-Pleistocene changes are not well fixed in magnitude, but their relative warmth is approximately correct. Maximum warming in the Cretaceous based on estimates by Barron and colleagues (Thompson and Barron 1981, Barron and Washington 1985). Time intervals in between have been scaled accordingly. [From Crowley 1989]

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FIGURE I.12. Paleogeographic Reconstruction at 100 Ma. Light areas on continents indicates regions flooded by shallow seas (maximum depth 100-200 m). [After Barron et al. 1980]

FIGURE I.13. Example of How Altered Solid-Earth Boundary Conditions May Have Resulted in Regional Climate Patterns Significantly Different from the Present During Pre-Pleistocene Warm Periods. Comparison of poleward ocean heat transport in the Atlantic for an ocean GCM simulation with present boundary conditions and with an open central American isthmus. An open isthmus results in an almost-complete collapse of the North Atlantic thermohaline cell. [After Maier-Reimer et al. 1990]

APPENDIX J

UTILIZATION OF PALEOCLIMATE RESULTS TO
VALIDATE PROJECTIONS OF A FUTURE GREENHOUSE WARMING

by

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APPENDIX J

UTILIZATION OF PALEOCLIMATE RESULTS TO VALIDATE PROJECTIONS OF A FUTURE GREENHOUSE WARMING

J.1 INTRODUCTION

A future greenhouse warming threatens to be a climate change of significant magnitude. In order to test climate projections, modelers have sometimes turned to the paleoclimate record for answers to a number of critical questions concerning greenhouse studies. In this paper I describe the state-of-the-art of paleoclimate studies with respect to two important climate problems: validation of climate models and assessment of the role of CO₂ as a significant climate feedback. Because of the scope of the problems being reviewed, all of the above questions cannot be answered in excruciating detail. The reader is referred to Crowley and North (1990) for a more detailed discussion of many of the observations and conclusions reviewed in this paper.

J.2 VALIDATION OF CLIMATE MODELS

The same climate models that have been used for making greenhouse projections have also been used for paleoclimate simulations. Comparison of model results with paleoclimate observations therefore allows us to establish some level of confidence in the models. In this study I will focus on the time intervals most examined (the last 20,000 years), but include a few comments on the mid-Cretaceous (90-120 Ma, million years ago) interval of inferred higher CO₂ levels.

Agreements

For the last 20,000 years, a detailed comparison of models and observations is summarized in Crowley and North (1990). The net conclusions (Figure J.1) of this comparison are as follows. In the high-latitude Northern Hemisphere at the last glacial maximum (18,000 BP) there is overall a very good agreement between models and observations. The giant Laurentide Ice Sheet exerted a very strong influence on regional circulation patterns over much of the area north of about 30°N. Regional comparisons of models and observations are often very favorable (Kutzbach and Wright 1985, Manabe and Broccoli 1985a, COHMAP 1988). One exception to this pattern involves some indications that the models are not as cold in winter over North America as suggested by the observations.

Another outstanding agreement between models and observations involves fluctuations of the African-Asian monsoon during the early part of the present interglacial period (~5000-10,000 BP). Calculations indicate that increased summer insolation resulted in enhanced heating on land, which in turn caused increased precipitation in the monsoon belt. Comparison of model-generated precipitation-minus-evaporation (P-E) changes and observations indicate in general very good agreement (Kutzbach and Street-Perrott 1985, COHMAP 1988).

Disagreements

A less satisfactory agreement involves climate change in the high latitudes of the Southern Hemisphere at 18,000 BP. Initial calculations (Manabe and Broccoli 1985b) suggested that the direct effect of the Northern Hemisphere ice sheets was quite small, as the temperature perturbation from the ice sheet radiates to space over a length scale of ~1500 km (cf. North 1984). Inclusion of lower ice-age CO₂ levels resulted in a much better agreement between models and observations, especially with respect to sea

ice changes around Antarctica (Manabe and Broccoli 1985a). Comparison of CO₂ and ice core temperature fluctuations over Antarctica further indicate a fairly good agreement (Genthon et al. 1987).

There are some problems with the apparently satisfactory convergence of models and data in the high latitudes of the Southern Hemisphere. Comparisons of CO₂ changes with glacial-interglacial transitions indicates that CO₂ lags Southern Hemisphere temperature changes (Figure J.2). This lag suggests that either CO₂ is not an important feedback or that there are other agents contributing to climate change in this area. Possibilities include changes in atmospheric methane, interhemispheric deep ocean heat transport, orbital forcing, or atmospheric dust levels (Crowley and Parkinson 1988, Harvey 1988, Chapellaz et al. 1990, cf. Petit et al. 1990). As an example of how these other mechanisms could modify the CO₂ temperature relation in Figure J.2, I have plotted the combined effect of CO₂, interhemispheric deep ocean heat transport, and orbital forcing (weighted 3:1:1) in Figure J.3. The choice of weights is arbitrary, for the purpose at this stage is to merely illustrate a point, not rigorously determine relative importance. Notice that the CO₂-temperature offsets are not nearly so prominent in this comparison, despite the fact that CO₂ is still accounting for 60% of the variation in the synthetic curve. The effect of dust (not considered here) is probably relatively small, for absolute concentrations are almost an order of magnitude higher in the Northern than in the Southern Hemispheres (Thompson and Mosley-Thompson, 1981). Furthermore, the radiative effect of the dust may have been overestimated by almost an order of magnitude (Anderson and Charlson 1990).

The apparently good agreement between models and observations over Antarctica at 18,000 BP (Jouzel et al. 1989) may also be misleading, because the result is dependent on stipulation of a

500-m thicker ice sheet over Antarctica by CLIMAP (1981). Utilization of a near-dry adiabatic lapse rate of $\sim 8^{\circ}\text{C}/\text{km}$ suggests that a 4°C temperature cooling, nearly one-half of the ice core range, is directly attributable to this elevation effect. However, more recent glaciologic studies over Antarctica (e.g., Denton 1985) no longer support inferences of a thicker ice sheet. Thus, we can explain only about one-third of the observed temperature change over East Antarctica at 18,000 BP.

An even more vexing area of model-data disagreement involves climate change in the tropics at 18,000 BP (Webster and Streeten 1978, Rind and Peteet 1985). Because CLIMAP (1981) tropical SSTs are relatively stable, models do not show much differences between control and perturbed runs. However, observations consistently indicate that tropical snowlines were lowered by about 1 km; data less consistently suggest that lowland precipitation was decreased (see Crowley and North 1990). Repeated efforts to validate the CLIMAP SST estimates with independent methods continue to support their main conclusion (Prell 1985, Brassell et al. 1986, Broecker 1986, Anderson et al. 1989).

It is possible that some of the lowland model P-E/data discrepancies may reflect processes not being incorporated into climate models. For example, lower CO_2 levels could alter a significant vegetation response, thereby giving a false impression of drier conditions. Biosphere feedback may also be important (e.g., Lean and Warrilow 1989). However, the upland model/data discrepancies are more difficult to explain. Geologists (e.g., van der Hammen 1985) have sometimes suggested changes in lapse rate, but this suggestion is usually considered anathema to climate modelers (Webster and Streeten 1978, Rind and Peteet 1985). Furthermore, (very) scattered tropical lower elevation (1-2 km) temperature estimates suggest decreases of $4\text{--}6^{\circ}\text{C}$ in these regions (cf. Liu and Colinvaux 1985, Colinvaux 1989, Bonnefille et al., 1990) — a result not consistent with stable SSTs offshore.

The above picture is further complicated by comparisons of upland and lowland temperature changes over Europe at 9000 BP, in which observations suggest a larger temperature response than over lowlands (Porter and Orombelli 1985, Huntley and Prentice 1988). Geologic indications of an increased temperature sensitivity vs. height are consistent with results from the 18,000 BP tropics and again diverge from model predictions (COHMAP 1988). The 9000 BP data provide additional support for the altered-lapse rate scenario, and perhaps it is prudent to re-examine this process in climate models.

Summary of Model-Data Comparisons

Direct comparisons of models and data for the last 20,000 years yield a mixed picture. We have two very good agreements and two cases where there are significant discrepancies. Trying to determine an "average rating" is not satisfactory, for none of the test cases give a middle-of-the-road response. A further consideration involves the fact that where models and data diverge, the discrepancy is not necessarily a condemnation of the model. Data are also open to reinterpretation and sometimes a model may not include certain processes that need to be considered.

Two other factors need to be considered when weighing the relative significance of the cases where there is a very good agreement between models and data. For the northern high latitudes at 18,000 BP, the very good agreement is impressive. However, it is not necessarily a justification for giving great credence to forecasts for greenhouse studies, because the Laurentide Ice Sheet was so huge that it essentially forced the atmosphere into a relatively narrow range of possible responses. The situation will be quite different for a future greenhouse perturbation, where there will be no massive localized forcing.

In the case of the second very good agreement between models and data (the monsoon at 9000 BP), there is also reason for caution in extrapolating successes to greenhouse forecasts. Intercomparisons (Crowley et al. 1986) of linear and nonlinear model responses to orbital forcing (Figure J.4) and additional GCM sensitivity experiments (Oglesby and Park 1989) yield the surprising result that in continental interiors the initial thermal response to orbital insolation forcing on seasonal time scales is very nearly linear (subsequent precipitation processes are more nonlinear, however). This result is also consistent with observations: if one were to compare the spectrum of observed temperature response in Eurasia to solar forcing at the top of the atmosphere, the amplitude of the annual and semi-annual harmonics are nearly identical for forcing and response, and there is very little variance in higher harmonics of the temperature field. If the temperature response were highly nonlinear, one would expect significantly greater variance in the higher harmonics of this field.

The above results, which are not completely understood from a theoretical viewpoint, imply a significant limitation to extrapolating seasonal model behavior to changes in mean-annual forcing, such as occurs with greenhouse gases. For example, it has sometimes been stated that because climate models simulate the present annual cycle so well, and since this cycle is larger than glacial-interglacial changes, then we can use this success as one measure of confidence in greenhouse predictions (Schneider 1989). The paleoclimate results suggest that the seasonal model example is sampling the system over the linear domain of its response. This result is a nice example of how paleoclimate studies have enhanced our understanding of how the climate system operates.

With the above considerations in mind I think we can say that at present we have overall only a fair agreement between models and data for the last 20,000 years. Going further back in time,

modeling studies suggest that CO₂ is perhaps the most important forcing factor influencing long-term climate change with mid-Cretaceous (90-120 Ma) levels perhaps 5-7X present levels (Berger and Spitzzy 1988, Berner 1990). Climate models would predict that such concentrations would produce tropical SSTs perhaps 2-4°C warmer than present and increased subtropical drought (Manabe and Bryan 1985, Hansen et al. 1988, Schlesinger 1989). Such changes are not observed in the record (Crowley and North 1990, Horrell 1990). It is possible that some of these discrepancies may reflect inadequacies in estimating paleo-SSTs (Horrell 1990) or changes due to continental drift (Barron and Washington 1985); nevertheless there is a possibility of a very significant discrepancy for this time interval that warrants more attention.

J.3 ROLE OF CO₂ AS A CLIMATE FEEDBACK

Critiques of greenhouse predictions seems to be almost a monthly occurrence. Given the importance of the problem, is there anything we can say about the role of CO₂ as a significant climate feedback based on observations and modeling of the geologic record? Despite the fact that our understanding of past changes is incomplete, I think it is fair to summarize the present state of knowledge by concluding that, although other factors are contributing to the long-term change of climate on 10³-10⁶ yr time scales, modeling studies exploring the importance of these mechanisms generally yield smaller changes than observed in the geologic record. Changing CO₂ levels, supported by both observations and geochemical calculations (e.g., Berner et al. 1983, Barnola et al. 1987), result in a closer correspondence between models and data (Crowley and North 1990). However, some significant problems still remain. Thus, from a geologic perspective we can say that CO₂ appears to play a very important role in both warming and cooling climates, and although the precise sensitivity of climate change to CO₂ levels still remains

to be more precisely determined, geologic studies suggest that the sensitivity is sufficiently large (approximately the same as the present generation of climate models) to justify concerns about future atmospheric perturbations due to greenhouse gas changes.

Geologic studies also provide us with other perspectives on the greenhouse problem. For example, the reaction of many greenhouse sceptics to the CO₂ predictions seems to somehow assume that there are any number of self-correcting features in the climate system that prevent large-scale climate change. Yet the geologic record states the exact opposite — that the system changes, sometimes greatly, and that its changes are larger than we would expect from modeling studies; in other words the system may be more sensitive than our present ability to model it.

The above changes lead me to conclude that, from a geologic perspective, there is a likelihood that the greenhouse forecasts are approximately correct. Temperatures will probably rise in the future, perhaps reaching levels not achieved for tens of millions of years (Figure J.5). Yet it is not possible to cite these past time intervals as analogs for a future warming (Crowley 1990), as future changes will be very rapid and involve a significant non-equilibrium component, whereas past warm periods represent long time-averaged equilibrium snapshots of warmer climates. Large changes in orography and ocean circulation also occurred in the past (e.g., Kutzbach et al. 1989, Maier-Reimer et al. 1990), thereby further making regional comparisons by analogy invalid. Thus, study of the geologic record yields the startling conclusion that the future greenhouse warming may be a unique climate realization in earth history.

J.4 CONCLUSIONS

1. Although there has been some success in modeling the climates of the last 20,000 years, there are significant areas of model/data disagreement. Where there is good agreement, it is felt that the model successes cannot at present be used as a gauge of the reliability of models for future greenhouse predictions. There may also be some significant discrepancies between models and observations for inferred mid-Cretaceous time of higher CO₂.
2. Modeling studies suggest that past CO₂ changes have played a critical role in the long-term evolution of climate; from this perspective geologic studies support the conclusion that the future greenhouse perturbation represents a significant climate change.
3. If the climate does change in the future, it is likely to result in climate state for which there may be no geologic analog; i.e., it may be unique climate realization in earth history.

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" Paleo-Modeling Scorecard"	
Region and Time	Model-Data Agreement
N. High-Lat. (18k)	Very Good
Low-Lat. (18k)	Poor
Low-Lat. (9k)	Very Good
S. High-Lat. (18k)	Fair
Ocean Circ. (18k)	Insufficient Work

FIGURE J.1. Status of Model Data Agreements for Various Time and Space Slices over the Last 20,000 Years. Evaluation represents the author's judgment, but there is considerable support for each rating. [After Crowley 1989]

S. HEMISPHERE CLIMATE RECORDS

FIGURE J.2. Comparison of High Latitude Southern Hemisphere Records for the Present and Last Interglacial with the Ice Core CO_2 Record from Vostok, Antarctica. Both the Southern Ocean SST and ice core temperature records indicate warmth greater than the present at the last interglacial. Note that CO_2 changes lag temperature in the Vostok ice core. SST record from Martinson et al. 1987, ice core deuterium record from Jouzel et al. 1987, CO_2 record from Barnola et al. 1987, ice core chronology plotted according to revisions of Petit et al. 1990. [From Crowley and North 1990]

VOSTOK DHO VS. COMPOSITE CO₂, NADW, TILT CURVE

FIGURE J.3. Example of How Additional Mechanisms Can Partially Reconcile the Southern Hemisphere CO₂-Temperature Offsets in Figure J.2. Figure illustrates the combined effect due to CO₂, changes in North Atlantic Deep Water (NADW) production, and changes in local orbital forcing over Antarctica (the latter from Short et al. 1990). Changes in NADW production affects interhemispheric exchange of heat (e.g., Hastenrath 1980, Crowley and Parkinson 1988), time variations in NADW based on equatorial Atlantic core analyzed in Mix and Fairbanks 1985. CO₂, NADW, and tilt variations have been weighted at 3:1:1. These ratios give approximately the best fit to the Vostok temperature record (solid line). However, the physical justification of the relative weighting requires more examination. Thus, this figure is for illustrative purposes only.

FIGURE J.4. Comparison of Energy Balance and General Circulation Model Responses to Increased Summer Insolation at 9000 BP. The 4°C contours indicate the difference between the model response for 9000 years ago and the present (the earlier time period is warmer). The energy balance model is from North et al. (1983); the low-resolution model and the NCAR CCM are from Kutzbach and Otto-Bliesner (1982) and Kutzbach and Guetter (1984), respectively. Note that all models have comparable sensitivity. Since the energy balance model is linear, this agreement suggests that the seasonal GCM response on land is also linear. [From Crowley et al. 1986]

FIGURE J.5. Schematic Comparison of Future Greenhouse Warming with Past Changes in Temperature. Note that Pleistocene climate oscillations are idealized and that interglacial temperatures were probably not significantly warmer than present. Pre-Pleistocene changes are not well fixed in magnitude, but their relative warmth is approximately correct. Maximum warming in the Cretaceous based on estimates by Barron and colleagues (Thompson and Barron 1981, Barron and Washington 1985). Time intervals in between have been scaled accordingly. [From Crowley 1989]

APPENDIX K

AUXILIARY CODES DEVELOPED IN SUPPORT OF
GLOBAL CLIMATE MODELING ACTIVITIES

by

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APPENDIX K

AUXILIARY CODES DEVELOPED IN SUPPORT OF GLOBAL CLIMATE MODELING ACTIVITIES

K.1 SCIENTIFIC CODE SATS

Program SATS is a scientific computer code conducting spectral analysis of univariate or bivariate data sets. The statistical algorithms of the code are based on the book Timeslab: A Time Series Analysis Laboratory, by H. Joseph Newton (Newton 1988). This program is written in Fortran 77 and runs on a VAX/VMS operating system. The system should include the NCAR (National Center for Atmospheric Research, Boulder, Colorado) graphics package.

The main diagnostics of the program are in the form of plots. Diagnostic plots (optional) are:

1. plots of data
2. periodograms
3. correlograms
4. smoothed power spectra
5. AR power spectra
6. coherence and phase (cospectrum)

In addition to these plots, two formatted files are created. One file contains the data sets for the plots above; the other file contains the frequency and value of the dominant spectral peaks.

This code was used to conduct some of the research reported in the Appendix C.

K.2 SCIENTIFIC ANALYSIS AND FILE MANAGEMENT PACKAGE SCANFM

Program SCANFM (SCientific ANalysis and File Management) is a scientific analysis and file management package consisting of 1 main program and 16 subroutines. The primary scientific functions of SCANFM are:

1. computation of global or zonal mean of global data sets
2. modal analysis of spherical expansion coefficients
3. simple statistical analysis
4. 2-dimensional spline interpolation
5. Fourier analysis
6. spectral to grid transformation.

The file management functions of SCANFM are:

1. algebraic manipulation of data sets
2. transformation of files
3. editing input files
4. handling spectral expansion files.

Although scientific analysis and file management are two distinct functions, they are assembled into one code so that SCANFM has a desirable flexibility and portability. Two important design considerations were the following:

1. Scientific analysis routines should be able to process data sets of foreign origin as well as domestic.
2. Pre-processing (manipulation) of data sets often precedes a scientific analysis of them.

This package is written in Fortran 77 and is designed to run under VAX/VMS operating systems. No graphic diagnostics are generated. This package has been extensively used to generate results in Appendix D and Hyde et al. (1989).

K.3 UTILITY CODE XYPLOT

This is a fairly general line-drawing utility code. A broad spectrum of data sets can be processed with this utility program. The code is written in Fortran 77 and runs on a VAX/VMS system. The NCAR graphic package should be supported by the system.

K.4 UTILITY CODE PROJECT

This is a utility program which draws contours on a stereographic or an orthographic projection with the center at a pole. Continental boundaries are optional. This code is an azimuthal-projection version of the ARC utility code CONTOUR. The code is written in Fortran 77 and runs on a VAX/VMS operating system. The NCAR graphic package should be supported by the system.

K.5 REFERENCES

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APPENDIX L

ARC QUALITY ASSURANCE PROGRAM

by

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APPENDIX L

ARC QUALITY ASSURANCE PROGRAM

L.1 INTRODUCTION

ARC has now completed two full years under Phase II, Technical Work. For the Quality Assurance (QA) Program, this has meant almost exclusive activity under the Software Management Procedures. In this appendix we summarize QA activities under the third year of our project and also review activities for the second year.

L.2 YEAR TWO REVIEW

In Year Two, the QA program was very productive, turning over to PNL close to 2200 pages of supporting documentation. However, experience highlighted need for improvement or modification of the program in some areas, with a view toward future project work with the very large General Circulation Models (GCM).

L.2.1 Software Activities

The end of the second contract year had seen a total of 25 software programs with complete QA procedural documentation. Eighteen were classified as scientific, 3 were utility codes, and 3 were system-supported and maintained packages (see Table L.1). There was also a significant increase in data sets under configuration management, up to 230 in number.

Code Management

Quality Assurance documentation was completed early in the year on a contour mapping code, CONTOUR, which is instrumental in producing contour maps of global data. The program generates contour maps suitable for reporting to PNL and for publication in

the refereed scientific literature, but it has played a vital part in representing the results of model development and verification testing.

The base Energy Balance Model (EBM), LINMOD, was documented and verified. As an acquired code, the model was run to demonstrate that the acquired configuration was the same as that used to generate the published results. First efforts failed, and the discovery was made that the published results chosen for replication employed a data set of boundary conditions not provided with the model. Successful results followed acquisition of the correct data set. Although the model has been well documented in the scientific literature, and the published results have received the scrutiny of the refereed literature, the staff conducted its own robust verification test. Verification of the model involved a demonstration that model results were in fact a solution to the energy balance equation. The test entailed a restricted, though characteristic, exercise for the model which could be reproduced with hand calculations. LINMOD successfully passed these verification tests. Verification of new versions of the model could then be simplified to the replication of appropriate results from the first version.

The new model version, LINEBM, was the first such exercise. The primary difference between the two models, and the purpose for generating the new version, was the incorporation of a coalbedo field that varies in the meridional as well as zonal direction (i.e., is two-dimensional). Other changes were made for the sake of clarity of organization, efficiency of computations, and generality of function, all of which resulted in externalizing

TABLE L.1. QA-Processed Codes for Year Two

<u>Software Name/Version</u>	<u>Software Classification</u>	<u>Configuration Control Log File Category</u>
CONTOUR/1.0	Utility	2010.1
CONTOUR/1.1	Utility	2010.1
LINMOD/1.0	Scientific	2010.2
SUN/1.0	Scientific	2010.3
WORLD/1.0	Scientific	2010.4
LINEBM/1.0	Scientific	2010.5
PREP/1.0	Utility	2010.6
QUAD/1.0	Utility	2010.7
EXSMTH/1.0	Scientific	2010.8
COEF/1.0	Scientific	2010.9
NUMBER/1.0	Scientific	2010.10
PRELIM/1.0	Scientific	2010.11
EVLU/1.0	Scientific	2010.12
GRID/1.0	Scientific	2010.13
GRID/2.0	Scientific	2010.13
DIFF/1.0	Scientific	2010.14
DIFF/2.0	Scientific	2010.14
TZONL2/1.0	Scientific	2010.15
TZONL2/2.0	Scientific	2010.15
DZAV/1.0	Scientific	2010.16
RMS/1.0	Scientific	2010.17
COR/1.0	Scientific	2010.18
GINT/1.0	Scientific	2010.19
NCAR Graphics Library	System	
IMSL Library	System	
VAX/VMS KERMIT/3.2	System	

some functions to separate auxiliary software. The mathematical problem and numerical strategy of LINEBM are the same as LINMOD, with the few exceptions of numerical method confined to the auxiliary programs. Consequently the new auxiliary software were subjected to the procedural internal development, review, and testing appropriate to new modeling efforts. The auxiliaries and the core program LINEBM, as a cohesive model package, were then verified as a new version of LINMOD in the manner mentioned above.

The bulk of results reported in Hyde et al. (1989) represent analyses of LINEBM simulations carried out by seven diagnostic

software routines developed internally for that purpose. These routines were developed, reviewed, and tested according to the appropriate procedures. Three of these codes were further modified under procedure to conduct the same analysis on the acquired GCM results discussed in the report. For the results therein, a total of ten codes beyond the EBM were processed under QA procedures.

Reported Results

Software activities extend to documentation of applications of the controlled codes that culminate in reported results. When a climate model and/or other codes are executed to produce reported results, the activity is recorded in an Application Package which includes statements of the purpose of the executions, a log of the code applications, a description of all input data with their justification, and an overall execution summary. A follow-on report is generated which identifies the specific Application Packages corresponding to the reported data, and this report is submitted as a research record. For example, for the first year annual summary results, a report is filed linking the reported data to Applications Packages of some 223 pages in total.

Data Management

Data Base Control entailed management of 230 total data sets under the categories:

- the acquired data of NCAR GCM paleosimulations
- compiled climatology of current global surface temperature
- baseline test input data and test results from code verification procedures
- digitized model boundary conditions for heat capacity, coalbedo, land/sea geography, and ice sheet elevations
- archived code application results and graphics.

L.2.2 Administrative Activities

Administrative activities for the second year were primarily routine maintenance, in nature. Activities included purchases, surveillances, sponsoring audits, a minor procedure change, and indoctrination and training of a new staff member:

- One printer, purchased under contract funds, was documented as to purchase requisition approval and item inspection upon delivery.
- Internal surveillances were conducted following a revised understanding of scheduling, and findings were minor and were immediately correctable without effect on reported results.
- ARC hosted a PNL QA audit in November 1988. Two findings were noted but were not considered adverse by the PNL Lead Auditor. The requested corrective action responses were submitted to PNL within the required time limit. Audit corrective action included a request for change of a QA Program procedure, with the change subsequently approved by PNL QA. Additionally, ARC was found not to have generated a written plan for bimonthly internal surveillances. Cause for the discrepancy was a misinterpretation of the QA procedure wording, but a consensus understanding was arrived at during the audit, resulting in a much smoother and easier method of conducting the surveillances.
- A new member was brought on board the project in January 1989. Standard indoctrination and training was conducted for the new member.

L.3 YEAR THREE SUMMARY

The ARC Quality Assurance (QA) Program matured in the third year, gaining more experience in the application of developed and acquired software. Further experience was gained with software

verification tasks, and significant QA Program needs were brought to light that will affect project goals in both the near and long term. Software activities were in a transitional phase, from climate model development to diagnostic utility development and processing of externally developed GCM output. Several new data sets were acquired, and with a view to increasing data management activity, a QA assistant was hired to take over the data management and other duties.

L.3.1 Software Activities

Code Management

Software development slowed by comparison to Year Two activities, at least regarding the number of original codes developed in house (26 last year, 10 this year). The last climate model development (NEQMOD) envisioned for the project was completed. However, there were several very useful and sophisticated diagnostic codes written for analysis of model and time series output. Table L.2 lists the software processed through QA procedures this year, along with classification and a brief description of function. It is useful to consider what these software involve.

This year saw a winding down of climate model coding with the culmination of Energy Balance Model (EBM) development, as the hierarchical modeling approach (Appendix A) progresses into the more complex General Circulation Model (GCM) activities of Year Four. The final EBM development step was completed with the modification of the linear EBM, LINEBM, to incorporate non-linearity in an albedo feedback process — tested and renamed NEQMOD. Two codes, QWORLD and PLMQ, were developed in parallel with NEQMOD to handle computations associated with the non-linear model features.

TABLE L.2. QA-Processed Codes for Year Three

Code/Version	Classification	Function
QWORLD/1.0	Scientific	NEQMOD Aux.: interpolate LINEBM geography to Gaussian grid.
PLMQ/1.0	Scientific	NEQMOD Aux.: generate norm. assoc. Legendre functions
NEQMOD/1.0	Scientific	Non-linear version of LINEBM global energy balance model
SCANFM/1.0	Scientific	File/data manipulation, management, editing; interpolation, Fourier analysis, global data & simple statistics
SATS/1.0	Scientific	Series statistics, spectral analysis, and Auto-Regressive modeling
CONTOUR/1.2	Utility	Improvement on global data contouring
PROJECT/1.0	Utility	Global data contouring in polar projection
CO2/1.0	Scientific	Computation of IR back-radiation due to given CO ₂ .
KALEIDAGRAPH/2.0	System	Graphic data plotting
TIMESLAB/1988	System	Statistics package

Several software were developed to handle diagnostic analysis and processing of data sets, both modeling products, and acquired data. SCANFM is a collection of ten routines which conduct scientific analysis of both foreign and domestic origin data sets, and the pre-processing (manipulation) of data sets which often precede a scientific analysis. Features include data file editing and manipulation, array transformations, data set algebra, simple statistics, global array averaging specific to project needs, Fourier analysis, 2D spline interpolation, and modal analysis of spherical expansions. The testing was quite involved, as each feature was exercised individually.

SATS is a coding of some of the excellent features of the commercially available TIMESLAB (Newton 1988) time series analysis package for an IBM Personal Computer. SATS was developed to provide these time series tools for the VAX computers, where most of our computing is conducted. Testing was extensive, though simplified, in that requirements essentially were to reproduce computations published in the user manual for TIMESLAB.

CO2 is a relatively short routine which codes equations for generating estimates of back-scattered infrared radiation received at the Earth's surface due to a given level of atmospheric CO₂ concentration (Kiehl and Dickinson 1987). The output was used for adjusting model radiation parameters for the study reported in Appendix D.

CONTOUR, version 1.2, is an update of the previously developed contouring code to provide enhanced ease of use. PROJECT is a 2D contouring code developed especially for polar projections of global arrays. PROJECT was used extensively during development of NEQMOD, providing indispensable analysis of observed and model-generated polar ice extents.

KALEIDAGRAPH is a commercially available science-oriented graphics program for the Apple Macintosh, with sophisticated report-quality graphing capabilities, curve fitting routines, statistics, and programmable data manipulation tools. The program has generated many high-quality graphics presentations for this report and those submitted to professional journals.

Data Management

Data management this year expanded considerably with the archival and processing of 33 different data bases comprising almost 460 data sets. The data consist of archives of observed global parameters, modeled output, model boundary conditions, time series of observed and modeled parameters, and software baseline and test results.

In the near future, data base management activity could possibly double, with processing, documentation, and archival of the enormous GCM output. The added QA staff should enable this program to adequately handle the increased duties.

L.3.2 Administrative Activities

Administrative activities included indoctrination and training of new staff personnel, routine internal surveillances, and hosting bi-annual PNL QA Audits.

A new scientist and a QA assistant were processed through the indoctrination and training procedure. The scientist was trained in Software Management QA procedures, while the QA assistant was introduced to both Administrative and Software Management procedures. The scientist was well-versed in fluid-dynamics modeling, so that project background and model documentation reading assignments were sufficient for technical training. The QA assistant is undergoing a more extended training, involving hands-on QA duties under tutelage of the QA Manager. Primary duties are Data Base Steward, but responsibilities are to include assisting code custodian and technical staff with standard QA documentations, and other duties as familiarity with the project proceeds.

ARC internal surveillances this year revealed minor deficiencies which were primarily backlogs of routine documentations such as logs of software and data management duties. Deficiencies were corrected at the time of surveillance, and there was no effect on reported results.

ARC hosted one PNL semi-annual QA Audit this year, with no findings or observations. The second PNL QA Audit was postponed to the next contract year due to rescheduling at PNL.

L.3.3 Evaluation

As identified in the internal Semi-annual QA Program Assessments, the QA program implementation is satisfactory. Adequacy of the program is good, overall, but some procedure revision will be needed for future project needs. As an example, the minimum requirements for code verification are written for relatively simple numerical methods. These requirements are not relevant to the extremely large and complex GCMs which fall under the future plans of the project. Changes to the procedures may involve more robust and generalized requirements which are applicable to both simple diagnostic tools and GCMs.

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APPENDIX M

CLIMATE MODELING DEFINITIONS

APPENDIX M

CLIMATE MODELING DEFINITIONS

M.1 INTRODUCTION

A GLOBAL CLIMATE MODEL is a mathematical computer model that is capable of simulating the seasonal cycle of certain weather variables starting with the planetary scales of atmospheric behavior and including other scales down to some "resolution". These models must be run long enough on the computer for transient effects to die out. An annual cycle model typically uses a simple 50 - 75 meter deep ocean for the lower boundary constraint on the atmosphere. Simulations with these models take a minimum of 15 model years. An imitation of a part of the seasonal cycle can be had by setting the sun at its January position, for example, and prescribing the sea surface temperature for that month from data. While the verisimilitude of this imitation is a matter of dispute it cuts down computer time by about a factor of 20 and it may be valuable for certain purposes. Global atmospheric models have to use other boundary conditions as input, such as the placement and thickness of ice sheets, the atmospheric composition and the oceanic structure. The output of a global climate model is the seasonal cycle in each grid box over the earth. We call this the regional climate forecast.

Examples of global models used in this study:

EBM. EBM is a generic term for an "energy balance model", which calculates temperature as a function of the incoming solar radiation, modified by the albedo (reflectivity) of the earth. In our report, we focus strictly on seasonal models, which involve realistic seasonal forcing and response (as opposed to annual-averaged forcing). We also use only two-dimensional models, with

the dimensions being equivalent to latitude and longitude so that continental outlines can be resolved (this step is necessary because it can be demonstrated that land-sea distribution represents the principal control of seasonal temperature fluctuations). There are several different types of EBMs that we discuss and employ:

LINMOD. LINMOD is the basic version of the 2D-EBM from which others are derived. It was developed by North et al. (1983) and resolves the seasonal cycle but is a linear model and does not allow for the regionally varying albedo of snow and ice.

LINEBM. LINEBM is a modification of LINMOD that allows for regionally varying but prescribed albedo of snow and ice. This model, which is linear, was developed at ARC, described in Hyde et al. (1989), and tested against general circulation model (GCM) simulations of climate change over the last 18,000 years.

NEOMOD. NEOMOD is a model recently completed under this PNL contract (Appendix D). The abbreviation stands for Nonlinear Equilibrium MODEL, for it is a nonlinear model that allows for interactions of altered seasonally varying snow and ice fields with the planetary radiation budget. Such a model is important for testing, for example, the future effect of greenhouse gas changes on climate, for much of the climate system response is nonlinear and involves the snow and ice fields. The "equilibrium" term in the title refers to the fact that there are two different methods of solution for the ice-albedo feedback, one involving spherical harmonic expansions in order to achieve an equilibrium solution, the other involving finite-differencing, which determines a time-marching solution.

GCM. GCM stands for general circulation model. This class of models solves the standard equations governing momentum, moisture, etc. and explicitly computes fields such as circulation, precipitation, etc. It therefore provides much more information

than an EBM, but at a much-greater cost and presumably is more realistic. At present our project envisions using two different types of GCMs:

LOW-RESOLUTION GCM (LRM). LRM is, as the name implies, a low-resolution version of a GCM. The model has approximately $15^{\circ} \times 15^{\circ}$ resolution in the horizontal, with five levels in the vertical, and a simplified radiation code (Otto-Bliesner et al. 1982). All of these features enable the model to be run approximately ten times faster than the high-resolution version. The LRM is an ideal model for testing a range of parameter space values before settling on a few experiments to run with the most comprehensive type of GCM (see below).

HIGH-RESOLUTION GCM. The High-Resolution GCM is, again as the name implies, a high-resolution model, with about a $5^{\circ} \times 5^{\circ}$ horizontal resolution and about ten levels in the vertical, and a more complicated radiative code. This model gives the most detailed information about climate change, but is also the most expensive to operate. The model we plan to use later in the project is a version of the Community Climate Model (CCM) from the National Center for Atmospheric Research (NCAR; cf. Ramanathan et al. 1983, Williamson et al. 1987).

MESOSCALE MODEL. Results from global-scale models will be used as input to a finer-mesh regional model being developed at NCAR (Dickinson et al. 1989, Giorgi and Bates 1989, Giorgi 1990). This model is capable of simulating changes on a spatial scale an order of magnitude finer than the GCM and will be used to make the best estimates of precipitation variations at Yucca Mtn.

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