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# *Early Paleogene oceans and climate: A fully coupled modeling approach using the NCAR CCSM*

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## ABSTRACT

**Attempts to understand early Paleogene climate using models have provided insights, but have consistently had difficulty explaining essential climatic features. A main limitation of those studies has been the lack of interaction between dynamical ocean and atmospheric models. After reviewing previous model studies and quantitatively defining the "low gradient problem," we present results from the first fully coupled general circulation model simulations in equilibrium with reconstructed early Paleogene topography, bathymetry, vegetation, and $pCO_2$ (560 ppm). Although our results should be understood as preliminary, predictions of both Paleogene sea surface temperatures and salinities are, for the first time, made without assuming an arbitrary value of the oceanic or atmospheric heat transport. Model-predicted bottom water temperatures are 7 °C warmer than modern. Deep convection occurs in the North Atlantic and Tethys, forming warm (12–15 °C), salty water masses (35.5 ppt). Through most of the Southern Ocean, vigorous wind-driven upwelling and a stable water column caused by low salinities (~32 ppt) produce a temperature inversion that may explain proxy interpretations of "warm salty deep water" formation in the tropics. Estimates of seawater oxygen isotopic ratios reveal substantially different patterns than usually predicted. Ocean heat transport is little-changed from modern values and warming of tropical temperatures (by ~3 °C) is about half that occurring in high latitudes. The mean annual temperature and seasonality patterns produced by the simulation are biased toward colder and more seasonal values than those reconstructed from temperature proxies.**

## INTRODUCTION

Climate proxies depict the early Paleogene as having warm (6–12 °C) deep ocean and polar temperatures, year-round above-freezing temperatures in continental interiors, near-modern (23–27 °C) tropical sea surface temperatures (SSTs) (Zachos et al., 2001; Zachos et al., 1994; Greenwood and Wing, 1995; Markwick, 1998; Wilf, 2000; Crowley and Zachos, 2000; Bralower et al., 1995), and modern-to-higher greenhouse gas concentrations (e.g., $pCO_2$ of 400–4000 ppm) (Retallack, 2001; Royer et al., 2001; Pearson and Palmer, 2000). This combination of small temperature gradients (the "low gradient problem"), reduced continental seasonality (the "equable climate problem"), and modern or higher greenhouse gas concentrations

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poses some of the most puzzling and important problems in the study of climate (Barron, 1987; Valdes, 2000a).

It is possible that existing early Paleogene tropical SST reconstructions are biased toward cool values, and that an upward revision of those values might help resolve this debate (Huber and Sloan, 2000; Pearson et al., 2001). But, barring that, the one commonly advocated theory that might appear to solve these problems relies on oceanic heat transport and greenhouse gas concentrations much greater than present values (Rind and Chandler, 1991; Barron et al., 1993). Consequently, attention has focused on the challenging problem of modeling the paleoocean circulations and, especially, on placing bounds on ocean heat transport for past greenhouse climates (Schmidt and Mysak, 1996; Lyle, 1997; Brady et al., 1998; Bice et al., 2000a).

The advancement of our understanding of climate requires theories and models that are independent of the data, in order to avoid circularity, and to make predictions testable against those proxy data. Previous attempts to address the question of early Paleogene ocean heat transport and temperature gradients have been limited in their ability to properly represent interaction between ocean circulations and the atmosphere (e.g., Sloan and Rea, 1996; Bice et al., 2000a). In this study, we couple dynamical atmospheric, ocean, and sea-ice component models to achieve a greater level of realism (and a decreased degree of circularity) than is possible in a framework in which climate components can not interact with each other. The coupled simulations are performed with higher-than-modern $pCO_2$, 560 ppm (at the lower-to-mid range of proxy estimates), and boundary conditions appropriate for the early Paleogene. This study is a continuation and expansion of a study described in Huber and Sloan (2001).

In the following section, we summarize the major modeling methods used previously and the relevant results of early Paleogene studies performed with those models. In the third section we outline a method, predicated on the results of previous uncoupled studies, for performing fully coupled simulations of past climates. The fourth section addresses important questions about early Paleogene climates that have not been explored previously in a fully coupled and self-consistent framework. Those questions include: (1) Were ocean currents weaker or stronger in an early Paleogene greenhouse climate? (2) Is the ocean salinity distribution different than usually assumed? (3) What were the locations and watermass properties of deep water formation? (4) Was meridional oceanic heat transport greater than modern in an early Paleogene simulation that is both fully coupled and includes an equilibrated deep ocean circulation? (5) Can the small temperature gradients evidenced in the climate proxy record be maintained in a fully coupled model? We discuss the implications and limitations of this study in the fifth section. This is followed by a summary of the results.

## A SUMMARY OF PREVIOUS MODELING STUDIES

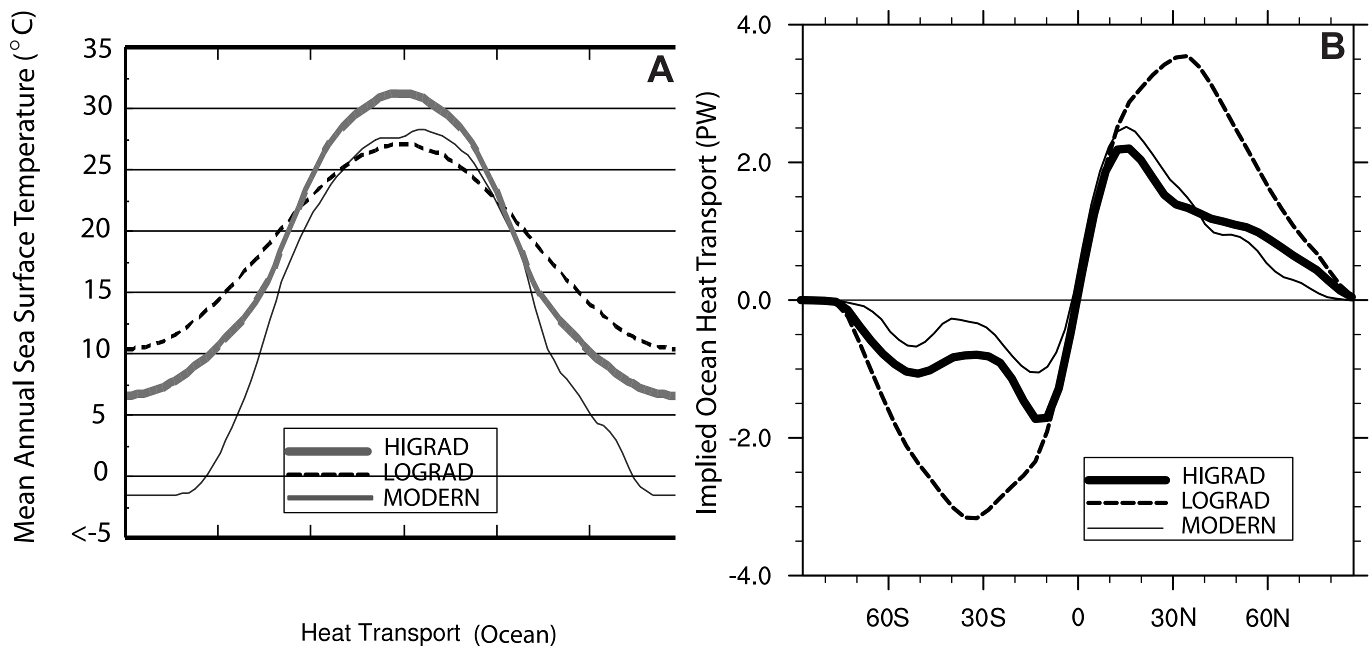
Three types of dynamical numerical model have been used to address the questions posed by early Paleogene climate: (1) atmospheric general circulation models (AGCMs) driven by fixed SST distributions; (2) AGCMs coupled to mixed-layer, "slab" oceans, with specified mixed-layer thickness and meridional heat fluxes or diffusivities; and (3) ocean (O) GCMs usually driven by SST, precipitation minus evaporation (*P–E*), and wind fields taken from the former types of experiments (cf. DeConto et al., 2000). We review results from these studies because they provide the motivation for the use of a comprehensive general circulation model simulations (CGCMs) in this study.

### *Type 1 models*

In studies of type 1, the SST distribution is specified (Fig. 1), and the object is to explore what atmospheric features (e.g., rainfall, wind, outgoing longwave radiation, and terrestrial temperature patterns), exist in equilibrium with such a distribution, and the sensitivity of these features to other factors (e.g., topography). The key physical assumption in these experiments is that the thermal inertia of the ocean is so much larger than that of the atmosphere that the ocean temperatures may be considered constant. By changing the SST distribution that drives the model, assumptions about the role of ocean heat transport also can be tested because the ocean heat transport *implied* by a fixed SST distribution is calculable as a residual from an AGCM simulation (Fig. 1).

In other work by the authors, we have found that high-latitude SST changes lead to small (relative to the data-model bias) responses over high-latitude landmasses, mostly within 1–2 model cells of the ocean, excepting western Europe and northeast North America where larger responses are produced (Huber and Sloan, 1999a; Sloan et al., 2001). Increasing tropical SSTs by 5 °C from 26 to 31 °C leads to extratropical winter warming (up to 5 °C), but the model-predicted sensitivity is not substantial enough to explain year-round above freezing temperatures (Huber, 2001). When tropical SSTs are decreased and high-latitude SSTs are increased at the same time, the total effect is a nearly linear combination of the separate forcings (Huber and Sloan, 1999a, 1999b; Huber, 2001). These low SST gradients, consistent with the traditional interpretation of the geologic record, have little net impact on winter temperatures in extratropical continental interiors. Even when extremely warm SSTs are imposed in high latitudes, below-freezing continental interior winter temperatures occur over most of North America, Asia, and Antarctica (Huber and Sloan, 1999a, 1999b; Sloan et al., 2001). The SST sensitivity results suggest that extratropical oceans with specified warm SSTs do not solve the equable climate problem, unless tropical SSTs were significantly warmer than 31 °C.

When SSTs are specified, $pCO_2$ may be changed as a sensitivity parameter although the results must be interpreted cautiously in light of the fact that the top-of-the-atmosphere radiative balance may not be consistent between experiments. In practice, changes in $pCO_2$ from 560 to 1120 ppm have small impacts on the radiative balance and a small effect on land surface



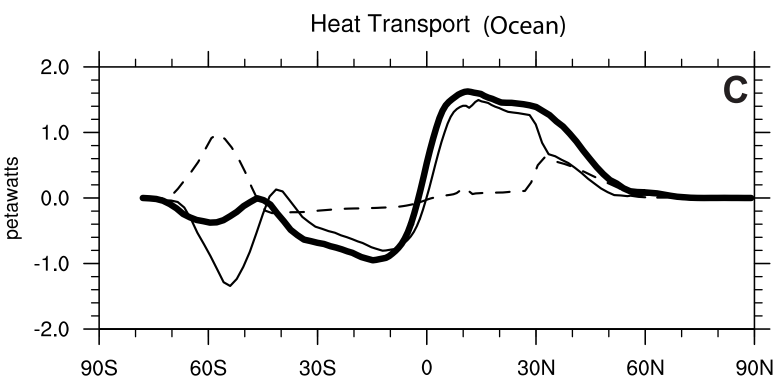




Figure 1. A: Zonal annual mean sea surface temperature (SST). The approximate range of early-to-mid Eocene SSTs from proxies including possible biases are spanned by the HIGRAD and LOGRAD distributions (estimates from Zachos et al., 1994; Crowley and Zachos, 2000). The degraded modern model predicted SSTs are shown with a thin line. B: Implied residual ocean heat transport calculated from the SST distributions (as described in the text). C: The global (thick line), Pacific and Indian Ocean (thin line), and Atlantic (dashed line) values of poleward ocean heat transport (positive northwards) are shown.

temperatures. Temperatures of some regions increase by ~2 °C for each doubling of $pCO_2$ within the range of 280 to1120 ppm (Huber and Sloan, 1999b). The relatively small response of extratropical continental interior winter temperatures has not been sensitive to AGCM formulation. For example, the mismatch between model result and proxies was only marginally improved when CCM3.6 was used instead of the older models GENESIS versions 1 and 2 (cf. Sloan et al., 2001). This same mismatch is further confirmed by Valdes (2000b), using both the UKMO and UGAMP AGCMs.

Type 1 experiments are fundamentally limited because they do not incorporate two-way interaction with the ocean. But, we can conclude from them that specification of warm extratropical SSTs—essentially assuming greater ocean heat transport and infinite ocean heat capacity—does little by itself to warm winter temperatures in continental interiors. Therefore, *if the ocean has a direct role in maintaining warm winter temperatures, it is only through coupled (two-way) interaction with the atmosphere.*

If tropical SST proxies are not biased to too cool values, then either (1) the ocean is not the key to the equable climate problem, or (2) some aspect of coupling a realistic ocean to the atmosphere is the key to solving the equable climate problem.

An initial approach to the second option is to couple the atmosphere to a simplified ocean.

### *Type 2 models*

In studies of type 2, SSTs are predicted, so it is possible to test SST sensitivity to $pCO_2$ (or other parameter) variations. It is explicitly assumed the ocean can be represented as a fixed-depth mixed layer, i.e., that the important ocean thermal inertia is in the upper 50 m, and ocean poleward heat transport is at specified levels (usually modern or higher). With early Paleogene boundary conditions, and near-modern $pCO_2$ and ocean heat transport values, these studies produce zonally averaged SSTs and meridional SST gradients nearly identical to modern. With higher $pCO_2$, tropical SSTs increase (to ~32 °C for 1000 ppm) without changing meridional gradients substantially. Sea ice forms seasonally in the Northern and Southern Hemispheres even at 1000 ppm (Shellito et al., personal commun., 2001). Studies that demonstrate substantial high-latitude amplification of temperature response to increases in $pCO_2$, or other forcings, generally do so because of the nonlinearity introduced by crossing a threshold from extensive sea-ice cover to little or no sea-ice cover (e.g., Sloan and Rea, 1996; Otto-Bliesner and Upchurch, 1997; Peters and Sloan, 2000). Proxy data imply (but do not rule out) that it was unlikely that sea-ice was present at all during past, warm "greenhouse" climates (e.g., Markwick, 1998), therefore this sensitivity may not be representative of actual warm Paleogene climate systems.

In the discussion of the importance of ocean heat transport, a distinction must be made between the potential warming effect of an individual current on a specific coastal region (e.g., <0.1 PW forcing) and the level of ocean heat transport change necessary to maintain relatively cool tropics (e.g., >1.0 PW). The former requires the penetration of a current to high latitudes in order to explain the polar warmth indicated by each isolated, coastal paleoclimate proxy locality. The latter has a major impact on tropical temperatures, which we have argued are the key factor for understanding winter warmth within continental interiors. The important effect of specified ocean heat transport increases in type 2 experiments is that it allows for the existence of higher $pCO_2$ (raising temperatures globally) without the production of overly warm tropical SSTs (Barron et al., 1993; Sloan et al., 1995).

***Type 3 models***

Early Paleogene uncoupled OGCM experiments (type 3) have successfully shown that the ocean circulation is sensitive to freshwater distributions (Bice et al., 1997) and land-sea distributions (Bice et al., 2000a). However, while an uncoupled OGCM modeling framework is useful for testing the sensitivity of ocean flow patterns to forcing, it may not be well suited for investigating some other important questions, such as poleward heat transport.

When the SST distribution is fixed, the implicit assumption is that the atmosphere has an infinite heat capacity relative to the ocean, opposite to the assumption in type 1 experiments. In a steady state and zonal average sense, the surface temperature gradient must map into the vertical temperature gradient, thus bottom water temperatures in such experiments cannot be cooler than the coldest surface value. In type 1 experiments, neither the coldest temperatures (upper atmospheric), nor terrestrial temperatures are fixed, so winds, heat transport, and surface land temperatures can be independently verified. In a fixed SST OGCM experiment, however, it is not clear how to explore the factors maintaining temperature distributions and heat fluxes. This highlights one fundamental difference between the ocean and the atmosphere—the atmosphere has two "free" boundaries, one at the top and one at the bottom, whereas the ocean has the top as its only "free" boundary. If this "free" boundary is specified as in type 3 studies, the problem becomes overdetermined. As this "fixed" assumption is relaxed, by using long restoring times, so-called "bulk forcing" (Brady et al., 1998; Doney et al., 1998), or other methods (Nong et al., 2000), the top boundary specification better approximates a real atmosphere, and, we argue, better estimates ocean heat transport.

An additional difficulty is presented by the definition of poleward heat transport itself. In the real atmosphere-ocean system, there are two ways to calculate poleward ocean heat transport, and if the measurements (or model) are perfect, the methods should yield identical results. The direct method involves integrating the product of poleward and equatorward ocean velocities and temperatures throughout the ocean to calculate the integral of the ocean heat transport divergence. Schematically, $F \propto V\Delta T$, where F is the poleward heat flux, V is the meridional overturning rate, and $\Delta T$, is the vertical temperature gradient. The other, indirect, method for calculating ocean heat transport involves calculating the net energy flux through the *surface* of the ocean and integrating meridionally to obtain the so-called implied ocean heat transport (Peixoto and Oort, 1992; Huber and Sloan, 1999a; Trenberth and Caron, 2001). In steady state, the heat lost or gained through the sea surface (the net sensible, latent heat fluxes and incoming and outgoing radiation) at the ocean surface must be balanced exactly by heat transported by ocean currents. The direct and indirect methods should give identical results in any system, because they are mathematically identical (via Gauss's Divergence Theorem), and in coupled models without "flux corrections" this requirement is nearly exactly met (Hack, 1998). In uncoupled OGCM sensitivity studies, there is no mechanism to guarantee that the implied and direct ocean heat transports agree—frequently, they do not. This inconsistency is important for ocean gateway change studies of ocean heat transport.

In studies of the effects of ocean gateway changes on ocean heat transport it has been common to use the same (or nearly the same) surface forcing (i.e. SSTs, winds, surface moisture flux) throughout the experiments, as land-sea distributions and gateways are changed, and then to compare differences in ocean heat transport (e.g., Mikolajewicz et al., 1993; Bice et al., 2000a). Using the direct method, substantial changes in ocean heat transport may occur as ocean currents are altered by differences in deep water formation patterns and bathymetry. The implied ocean heat transport, which is just as valid a measure of ocean heat transport, can not change since the fields on which it depends have not been allowed to change (or the change is very small).

This is a logical and physical inconsistency that obscures the interpretation of these experiments, because the surface climate state used to drive the ocean model is incompatible with the surface climate that the simulated ocean would create if the boundary conditions were free to evolve (as preliminarily explored by Mikolajewicz et al., 1993, and reexamined in Mikolajewicz and Crowley, 1997). A minimum requirement for realism in any ocean model simulation is that it be a possible solution in a coupled system, because the real atmosphere and ocean are coupled. Thus, the ocean heat transport produced by the circulation should not be substantially out of equilibrium with the heat transport implied by the surface forcing. The results of Bice et al. (2000a) for the early Paleogene are out of equilibrium by >2 PW, which is important for understanding the comparison of our results to that study below.

Another problem with uncoupled OGCM experiments is that, in the absence of salinity proxies, the surface freshwater balance must be arbitrarily specified, usually from AGCM results. A consequence is that one of the two major variables for determining ocean density gradients is unconstrained, either by proxies, or by the physics of ocean-atmosphere exchanges in which interactions are two-way. A further possible pitfall in such uncoupled studies, is an inconsistency between the ocean model circulation and the noninteractive moisture balance forcing that drives it.

As an example, consider a steady state, zonally averaged overturning circulation in equilibrium with a fixed SST and surface salinity distribution in which sinking occurs at high latitudes in cooler regions and fills the deep ocean with cold water. If evaporation is increased in a warmer (e.g., subtropical) region so that deep water formation occurs in this region, then the deep ocean will fill with warmer water. The volume average ocean temperature will therefore increase, although there has been no change in the heat flux at any boundaries. As an artifact of the assumed infinite atmospheric thermal inertia in the top boundary condition, heat has been added to the ocean without removing heat from any real source. In this way, conservation of energy (of the climate system) can be violated if proper account is not taken of the true fluxes into and out of the system.

Studies of modern climate using uncoupled OGCMs avoid these difficulties because the forcing at the top boundary is well known and the goal is to understand the ocean flows in equilibrium with this distribution—these flows are also largely known and thus the models are testable. However, in the Paleogene, constraints on ocean overturning rates, zonal temperature gradients, heat transport, current strengths or directions are weak because of poor coverage, small gradients in carbon and oxygen isotopes, and the absence of a proxy for salinity. Consequently, there are few means to provide an independent check on the predictions of uncoupled OGCMs, and no guarantee of energetic self-consistency in the simulations. An OGCM-produced circulation, which would collapse if coupled to an atmosphere, cannot reveal much about any real past climate. Therefore, regardless of whether a fully coupled model is used, a minimum requirement of OGCM experiments should be careful attention to the compatibility of fluxes passed through the ocean surface between the ocean and atmosphere.

### *A definition of the low gradient problem*

Attention to the fluxes between the atmosphere and ocean allow us to make a quantitative definition of the low gradient problem (cool tropics paradox). The core problem is that atmospheric models generate weak poleward heat transport when driven by small meridional temperature gradients. To maintain the small gradients, the ocean is required to make up the remaining heat transport. In other words, the ocean heat transport implied by the indirect method increases to levels much greater than modern. However, as the imposed SST gradient decreases, the ocean heat flux (F) is expected to decrease quickly with decreasing temperature gradients, because $F \propto V\Delta T$. Ocean heat transports calculated from the direct and indirect methods therefore diverge strongly as the imposed SST gradient decreases, and ocean heat transport greater than modern appears to be both required and ruled out by small temperature gradients.

There is a range, however, of smaller-than-modern temperature gradients that do not imply ocean heat transport increases over modern values. The HIGRAD SST distribution (described above and Figure 1) with an equator-to-pole SST difference, $\Delta T$, of 24 °C requires near-modern values of F, whereas the LOGRAD distribution with a $\Delta T$ of 15 °C requires an F ~2–3 times as large (Fig. 1). In other words, warm polar temperatures of 10 °C and tropical temperatures of 34 °C can exist in equilibrium with near-modern ocean transport values, but if the tropical temperatures are 25 °C, the climate is solidly in the "paradox" regime (D'Hondt and Arthur, 1996).

## THIS MODELING STUDY

### *General issues*

The most obvious missing physics of the experimental frameworks reviewed above is the lack of two-way interactive coupling of complete ocean and atmosphere models—and the best current explanation for the maintenance of these key early Paleogene climatic features is atmosphere-ocean interaction. In a fully coupled approach, assumptions based on modern day are not explicitly incorporated into the simulation (usually in type 2 experiments) and circularity is avoided because temperature proxy interpretations are not used to drive the experiment (usually in type 1 and 3 experiments). Furthermore, energetic consistency is built into the modeling framework. CGCM paleoclimate modeling (type 4) is a new and rapidly evolving field, so the simulations presented here should be considered as first attempts. Further refinement and validation will be required in future work. We describe here the three main challenges. The first lies in the ability of the CGCM to achieve and maintain a steady state climate for distant past conditions. The second is a product of uncertainty in initial and boundary conditions. The third difficulty is that the deep ocean has a characteristic time scale of two thousand years, a length of time over which it is computationally infeasible to integrate a CGCM. In this section, we describe the difficulties in performing CGCM experiments of the distant past, summarize a method that succeeds in resolving these difficulties, and outline the experiments performed.

### *The CCSM and climate drift*

Due to deficiencies in model formulation, most CGCMs require so-called "flux corrections" to maintain a stable steady state climate for modern conditions. Without these corrections, the modeled climate drifts toward unrealistic values of important climate parameters. These models are not convergent on a realistic climate state even for perfectly known initial and boundary conditions and consequently cannot be used for paleoclimate modeling in the deep past. It is therefore critical to use a CGCM that does not require "flux corrections" for modern day when attempting to model the early Paleogene.

We simulate early Paleogene climate with the Community Climate System Model (CCSM) developed at the National Center for Atmospheric Research (NCAR). CCSM (formerly CSM) version 1 (v.1) is described and validated for modern conditions in Boville and Gent (1998) and references therein. The results presented here are produced with an updated version 1.4, described for modern studies in Boville et al. (2001) and Blackmon et al. (2001). The atmospheric and land models are the Community Climate Model (CCM) 3.6 and the Land Surface Model (LSM) 1.2 at T31 (~3.75° by 3.75°) resolution. These are coupled to NCAR's Ocean Model (NCOM) 1.5 and the Community Sea Ice Model (CSIM) 2.2. The latter models are on a stretched grid with 0.9° equatorial and 1.8° high-latitude meridional grid spacing; zonal resolution is 3.6°. Components are connected through the Flux Coupler v. 4.0.5. As described in the references above, the major differences from CCSM v.1 are (1) better resolved tropical ocean circulations and thermocline structure (2) improvements in conservation properties of Flux Coupler interpolation between component models, and (3) an error in the sea-ice in v.1 has been fixed, leading to an order of magnitude decrease in the already weak model "drift."

The inclusion of state of the art treatments of boundary layer (KPP) and isopycnal eddy mixing (GM) in the ocean are notable features of this model (Boville and Gent, 1998). The model also predicts accurately modern El Niño–Southern Oscillation phenomenon (Otto-Bliesner and Brady, 2001), the Australian–East Asian monsoon, the North Atlantic Oscillation (B. Otto-Bliesner, 2001, personal commun.) and their remote climate responses (Meehl and Arblaster, 1998). CCSM also produces a response to future global warming well in the range produced by other models, with equilibrium sensitivity of ~2.1 °C for a doubling of $pCO_2$ (Dai et al., 2001; Meehl et al., 2000). Thus, the ocean-atmosphere interactions and their terrestrial impacts that are important in modern climate are well reproduced in this model, indicating that it is realistically sensitive to external forcing and internal variability. CCSM has also been used with paleoclimate settings in a series of Last Glacial Maximum (Liu et al., 2002; Shin et al., 2002) and Cretaceous simulations (Otto-Bliesner et al., 2002).

### *Initial and boundary conditions and the effects of uncertainty*

***Process and importance of degradation.*** Early Paleogene climate proxies that might be used as initial conditions for a modeling experiment are sparsely distributed in time and space, have inherent uncertainties associated with them, and may not constrain quantities that are important for climate modeling (e.g., salinity). There is also a good deal of uncertainty in boundary conditions, such as topographic or bathymetric reconstructions. Our knowledge of these boundary and initial conditions can never be as great for the early Paleogene as they are for today. Therefore we restrict this investigation to questions that can be answered with the level of detail available for the early Paleogene. We assess which questions can be answered with a process we call "degradation."

Approximation connotes that the essential elements of a system have been captured despite simplification. Given that the boundary and initial conditions are only roughly known and without universal agreement on the essential elements of the climate problem, our approximations are unavoidably wrong to a degree. We prefer the term degradation, because it better describes the process applied to validate this method for coupled model spin-up. Degradation provides one (minimum) estimate of the errors introduced into model predictions by misrepresenting multiple aspects of the system. Although the CCSM is currently one of the best described and validated CGCMs (in terms of number of recent publications), a "perfect" model for modern day might not be perfect for the correct physical reasons (instead, tuning might be responsible), nor nearly as accurate with less accurate boundary conditions. The process of degradation includes changing both the boundary conditions in the modern control case and some "detuning" of the model (used in both modern and paleo-cases). The modern-day bathymetry was also degraded as described below.

In the land surface model (LSM), modern soil sand/silt/clay fractions were set to the same global average values used in Sloan and Rea (1996) and three elements of tuning were turned off in modern and Eocene cases. These tunings make Arabian desert sand darker than average sand, and make snow that falls on land-ice south of 60°S, and north of 40°N between 45 and 165° east longitude (Siberia) more reflective. If the exact color of sand in a desert or the precise albedo of snow or the presence of small islands (like Iceland) actually play a large role in climate, then: (1) this detuning and degrading procedure should show it, and (2) it would raise serious questions about our ability to model past climates.

This degradation has been applied to a modern-day case in a manner analogous to that in which the paleoclimate cases are likely to be degraded. In so doing, many possibly important parameters have been changed at once. A weakness of this decision is that comparing the degraded modern case to standard modern case results is not a single parameter sensitivity study. Single-parameter sensitivity studies have been performed that indicate the effects are small enough to be well within observational error for the early Paleogene. As described below, and in Huber (2001), the degraded modern simulation quickly reaches a near-modern state that is close to modern observations. The strength of degradation is that by changing so many things at once, we more closely match the kind of errors that are likely to be introduced in paleoclimate work—since it is unlikely that only one feature is incorrectly represented. Despite its limitations, degradation addresses the question of whether even the "perfect" model can converge on a realistic modern climate state from poor knowledge of initial and boundary conditions.

***Creation of early Paleogene and "degraded" modern boundary conditions.*** The vegetation and topographic distributions for early Paleogene conditions are described in Sewall et al. (2000). The creation of a bathymetric data set for the early Paleogene began with land-sea distributions created by Sewall et al. (2000). Those land sea distributions were then remapped from a 2° × 2° distribution to the stretched grid of the ocean model. Slight modifications were made to ensure proper resolution of ocean gateways and narrow embayments. Isochron maps created by Royer et al. (1992) were then used to identify the paleolocations of Chron 25 and other late Paleocene–early Eocene isochrons, in order to constrain the location of midocean ridges during the early Paleogene. This information was supplemented by other tectonic reconstruction data in the Pacific (Lonsdale, 1998), Tethys (Dercourt et al., 1986; Zacher and Lupu, 1999) and the Indian Ocean (Zachos et al., 1992; Wilson and Moss, 1999). The locations of passive and active margins consistent with our tectonic reconstruction were mapped. Then, using a modern analogue approach, 6 depth classes appropriate for these tectonic regimes were derived from the modern ETOPO5 data set (NOAA, 1988).

These depth estimates were then spot checked with paleodepth estimates of specific ODP and DSDP cores based on faunal assemblage estimates (Zachos et al., 1992, 1993; Corfield and Norris, 1996; Miller et al., 1987), and found to be in good agreement. The depth of the Panamanian sill was picked somewhat arbitrarily (at 2500–3000 m) given the considerations that there is no evidence of a constriction of flow during the Paleogene, but in the late Paleocene, there is evidence of land faunal migration from S. America into N. America (Marshall et al., 1997). The bathymetric data set was smoothed using a 5 point smoother and then rechecked against the corederived paleodepth estimates mentioned above.

To evaluate the potentially large error introduced in the method of deriving paleobathymetry, analogous filters were applied to modern-day bathymetry. Modern bathymetry was discretized into the same 6 classes; seamounts and islands, including Iceland and Japan were removed, and the bathymetry was passed through a 5 point smoother, in addition to the detuning described above. These Eocene and degraded modern bathymetries are shown in Figure 2. In the simulations, the viscosity/diffusivity values are set at extremely low values (similar to those in Otto-Bliesner and Brady, 2001), which should allow the simulation to provide a more accurate representation of tropical current systems and gateways.

***Initial conditions.*** Early Paleogene simulations were performed with two different initial conditions with identical boundary conditions. The LOGRAD distribution was created from the SST record of Zachos et al. (1994) for the early Eocene. The Zachos et al. (1994) record was used as a basis for the zonally constant meridional SST values in LOGRAD. The coldest temperature in that data set was used as a first estimate of the deep ocean temperature, and SSTs were matched to this deep ocean value with a smooth fit. Salinity was initially set to a global average value of 34.7 ppt. The ocean heat transport implied by this distribution is several times modern (Fig. 1). The HIGRAD distribution was chosen assuming that the interpretation of tropical SSTs were biased toward too cold values in the Zachos et al. (1994) reconstruction and so tropical SSTs were made warmer and polar SSTs were made cooler to ensure the same global average temperature. An alternative interpretation of the HIGRAD distribution is that it is a better match to late middle Eocene or Paleocene values. The ocean heat transport implied by the HIGRAD distribution is near modern values. Deep ocean temperatures were specified in the same manner as in the LOGRAD distribution. Initial ocean conditions for the degraded present case are zonally averaged temperature and salinity taken from Levitus and Boyer (1994) and Levitus et al. (1994) and a state of rest.

The boundary conditions for the early Paleogene cases are described above and include $pCO_2$ of 560 ppm. The different Paleogene cases produce nearly identical surface properties, so we describe the case that began with HIGRAD initial conditions in this study. We discuss the results of only one in depth, because the two case results are similar, though some discussion of the second experiment is included in the Discussion and Summary sections. We have also performed several degraded modern control cases to explore the sensitivity of boundary and initial conditions, and spin-up procedure as described above and in Huber and Sloan (2001) and Huber (2001). The modern control case includes degraded modern boundary condition described above and a preindustrial value of $pCO_2$, 280 ppm.

### *Deep ocean spin-up procedure*

A fundamental difficulty with performing paleoclimate CGCM experiments is integrating the entire system, including the deep ocean, to quasi-steady state. It is not currently feasible with existing computational resources to attempt the ideal solution, which is to integrate several 3000 year simulations with a CGCM. Therefore, an iterative accelerated spin-up technique was developed to allow several thousand years of deep ocean simulation to be carried out in only several hundred years of surface ocean computation.

The procedure draws on two existing procedures: iteration of coupled and uncoupled modeling steps, and deep ocean acceleration. The iteration technique has been used and validated in a somewhat different form than that described here by Kutzbach and Liu (1997), Liu et al. (1999), Vavrus et al. (2000), Shin et al. (2002), and Otto-Bliesner et al. (2002), but their techniques are somewhat different, and the results of different techniques have yet to be quantitatively compared. The combination of uncoupled and accelerated ocean model integrations and fully

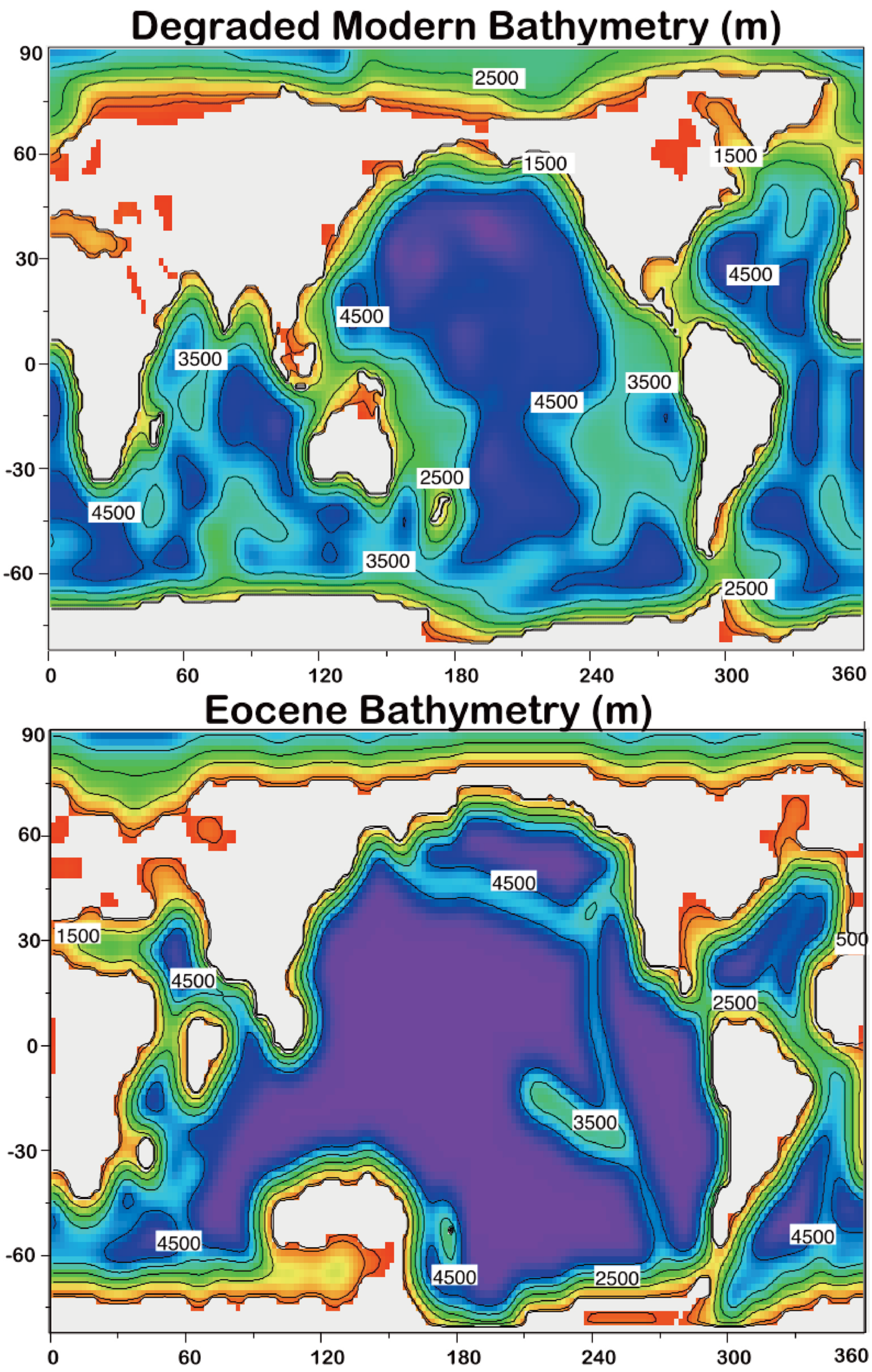



Figure 2. Degraded modern (top) and Eocene (bottom) bathymetry at 2° × 2° resolution. Contour intervals are 1000 m below 1500 m and 500 m above.

coupled unaccelerated integrations is quite common and is the usual way that CCSM integrations are spun up for modern day (see Boville and Gent, 1998; Bryan, 1998), but the process has not been systematically iterated, as it is in this study. The exact spin-up procedure is described in Huber and Sloan (2001).

SST biases in the modern (degraded) control case are <2 °C over most of the ocean, and ocean heat transport values are within the error for existing modern-day estimates (Ganachaud and Wunsch, 2000; Trenberth and Caron, 2001). The major temperature errors are in the stratus cloud regions at the eastern edges of the tropics (~+2 °C) and in the Ross Sea (~+8 °C). The control simulation's major weakness is too-weak formation of Antarctic Bottom water and a sluggish Antarctic Circumpolar current (ACC). CCSM v.1, when integrated for long periods, tends toward overly strong Antarctic Bottom water formation and ACC transport as described in Bryan (1998). Sensitivity studies (not shown) demonstrate that the Ross Sea temperature bias, weak ACC, and weak Antarctic Bottom Water formation are due to sea-ice and albedo feedbacks and spin-up procedures. Sea-ice was not likely to have been present in the warm polar early Paleogene climate and does not play an important role in our results described below, so this sensitivity minimally affects our early Paleogene results. The changes in bathymetry associated with degradation are also important for changes in the ACC transport, appearing to account for about a 25 Sv reduction (~1/5 of the total transport). This is in keeping with current understanding of the unique dynamics of the ACC, and is probably not likely to be a dominant source of uncertainty in our Paleogene simulations. The results of those studies show, in agreement with many previous studies, that in a suitably high resolution and inviscid ocean model the details of bathymetry have small effects on most large-scale ocean currents.

The total length of the early Paleogene integration in the case described here is 2200 deep water years (DWYs), and there was little significant difference between successive iterations, thus indicating convergence (not shown). As described in Huber and Sloan (2001), the iterative method proves to be robust and leads to an equilibrated coupled simulation that reproduces present climate well in the degraded modern case, and a stable and convergent solution in the early Paleogene cases. Extensions of the early Paleogene experiments to a total length of 6000 DWYs, completed during the revisions of the manuscript, produce substantially the same results as those described here, further demonstrating the results' robustness (not shown). In order to cover the results in sufficient depth we focus only on results in Atlantic and Indian Ocean regions; results from the Pacific are covered in other work (e.g., Huber, 2002). Means of the last 20 yr are discussed below.

## RESULTS

The global average surface temperature difference between the degraded and the early Paleogene case is 5.1 °C. This indicates a strong sensitivity of mean temperatures to the changes in boundary conditions (e.g., vegetation, topography), given that the effect of higher $pCO_2$ alone should be ~2.1 °C. Global and volume average ocean temperature in the early Paleogene simulation is 8.1 °C, and the global average deep ocean temperatures are ≥ 6 °C, demonstrating that realistic boundary conditions and 560 ppm $pCO_2$ are sufficient to reproduce generally warm deep ocean temperatures, albeit at the lower range of proxy estimates (e.g., late-middle Eocene, Zachos et al., 2001). There is very little "drift" in the simulation: The temperature trend for the last 90 yr is <.024 °C/100 yr and salinity 0.0003 ppt/100 yr. These values represent quasi-steady state values and the trends are extremely small even relative to modern CGCM integrations (e.g., Blackmon et al., 2001) (the salinity trend is statistically indistinguishable from zero).

Meridional overturning rates in the North Atlantic are steady at ~10 Sv, half that of the control simulation, but there is active deep convection at 55°N latitude with near-surface water penetrating to ~3000 m (Fig. 3). In the upper ocean (500 m) this water mass has a temperature of 13 °C and a salinity of 35.56 ppt (Fig. 4). This "young" North Atlantic Deep Water does not dominate the abyssal Atlantic because significant quantities of this water mass flow into the Pacific at 1–2 km depth through the Panamanian Seaway. The temperature of this NADW at depth is 9.5 °C and the salinity is 35.4 ppt (Fig. 4). In the South Atlantic, vigorous upwelling in the subpolar gyres leads to watermass transformation at the surface in high latitudes, where heat loss and freshening create a watermass analogous to Antarctic Intermediate Water (T = 3–6 °C, S = 33.0–33.75 ppt) (Figs. 3 and 4). Some of this water is exported into the North Atlantic at ~1000 m, but most of it mixes with warmer and saltier water from the North Atlantic to form a deep Antarctic water mass (T = 6 °C, S = 33.75 ppt) (Fig. 4).

The processes operating in Tethys and the Proto-Indian Ocean are not substantially different from those in the Atlantic. Deep convection occurs in the northern portion of Tethys (45°–50°N) (Fig. 3). Seasonal cooling and latent heat fluxes from offshore winds lead to an upper ocean Tethyan watermass with a temperature of ~15 °C and a salinity of 35.5 ppt (Fig. 4). This water flows southwards at depths of 1000–2000 m with a temperature of 10 °C and salinity of 35.5 ppt (Figs. 3 and 4). In the Southern Hemisphere the process and pattern of deep and intermediate water formation is similar to that described for the Atlantic.

Meridional ocean heat transport is similar to modern values (Fig. 1C). The majority of poleward heat transport by the oceans occurs in the Pacific. Atlantic heat transport is neglible in the Northern Hemisphere and directed equatorward in the Southern Hemisphere (as it is today). Peak heat transport values are higher in the Northern Hemisphere (1.6 PW) than in the Southern (0.8), but the distribution is more hemispherically balanced than in the present-day control run. Figures 4A–E present map views of annual mean near-surface or "mixed layer" properties. We present these vertically averaged values (upper two levels), rather than model surface temperatures both as a crude representation of the likely depth habits of foraminifera (e.g., Schmidt, 1999a; Poulsen et al., 1999; Bice et al., 2000b) and to increase the robustness of the model result interpretation.

### *Mixed-layer temperature, salinity, and velocities*

Far North Atlantic temperature values are ~15 °C and Southern Ocean temperatures are predominantly 1–3 °C (Fig. 4A). Tropical mixed-layer temperatures are above 29 °C with peak values of 31 °C. Salinity reaches very low values (31.5 ppt)

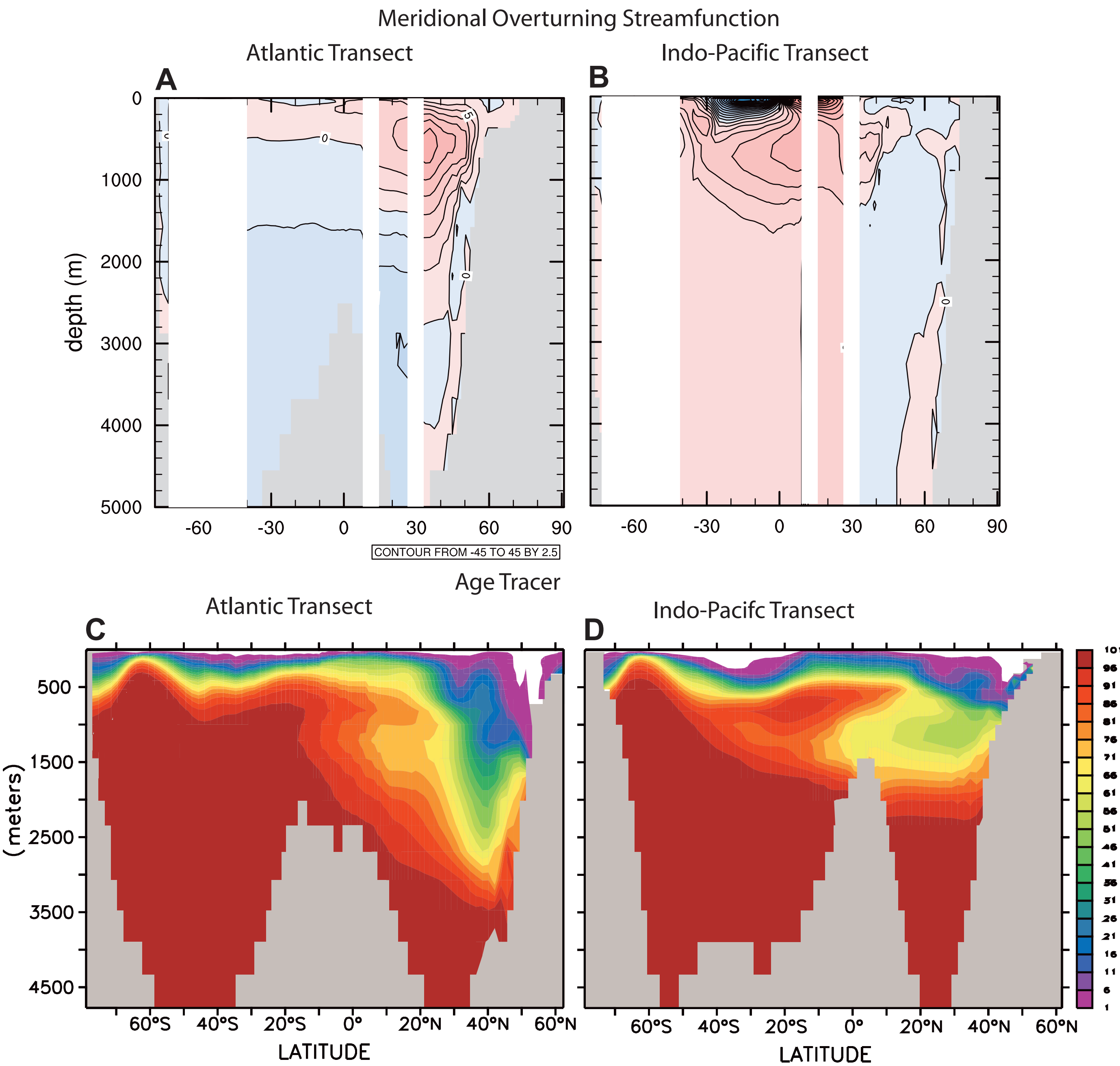



Figure 3. A: Zonal, annual average meridional overturning streamfunction (in Sv) in the Atlantic (A) and Pacific (B). Red indicates clockwise overturning, blue indicates counterclockwise. On the bottom an "ideal age" tracer used as a CFC-like qualitative indicator of the ventilation and transport of water masses in the Atlantic (C) and Pacific (D). The oldest value is 100 and represents values that have not been in contact with the surface within 100 yr, whereas younger values indicate that some or all of the water has been in contact with the surface within the period indicated by the color bar.

in the Southern Ocean and off the west coast of India (Fig. 4B). A low-salinity tongue is advected by currents northwestwards from India, into Tethys and thence into the Atlantic. The rest of Tethys is dominated by an anticyclonic gyre circulation that generates an eastward-directed current along its northern margin. The Benguela Current system advects cool, fresh water (17 °C, 32.5 ppt) northwestward to Brazil. Subtropical gyre evaporation zones have high salinities (>36.5), but are not significantly different from modern values. Strong northward cross-equatorial transport of warm, salty water is generated in this upper layer dominated by Ekman dynamics. A vigorous Gulf Stream-analogue transports warm (18 °C), salty (35.5 ppt) water north

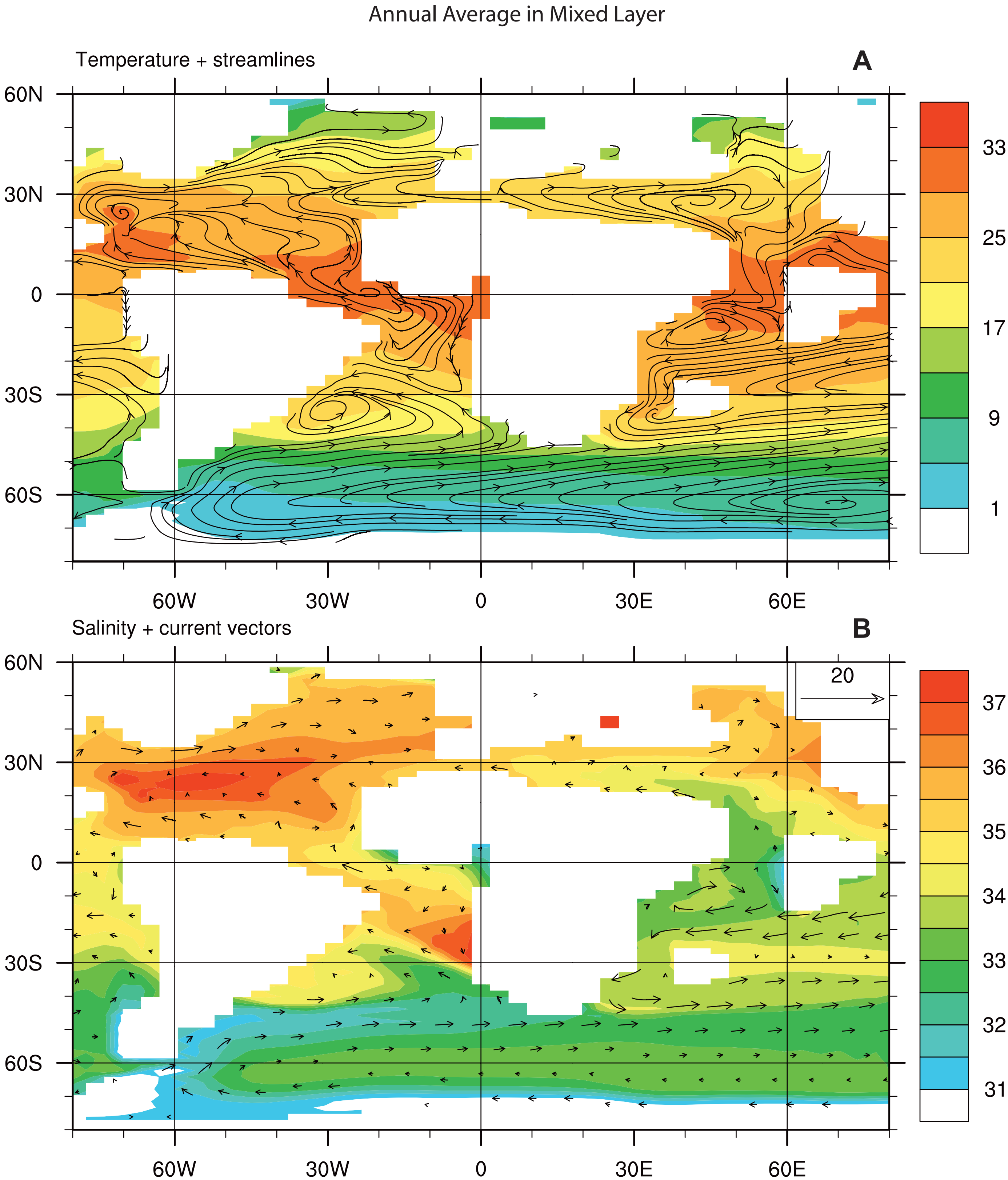



Figure 4 *(on this and following three pages).* Tracer properties and velocities at different levels. In the isotopic calculations we use a correction of –0.98‰ for ice-free conditions and an additional offset of –0.22‰ to convert from VSMOW to VPDB. Values are indicated by the color map and vectors located on the figures.

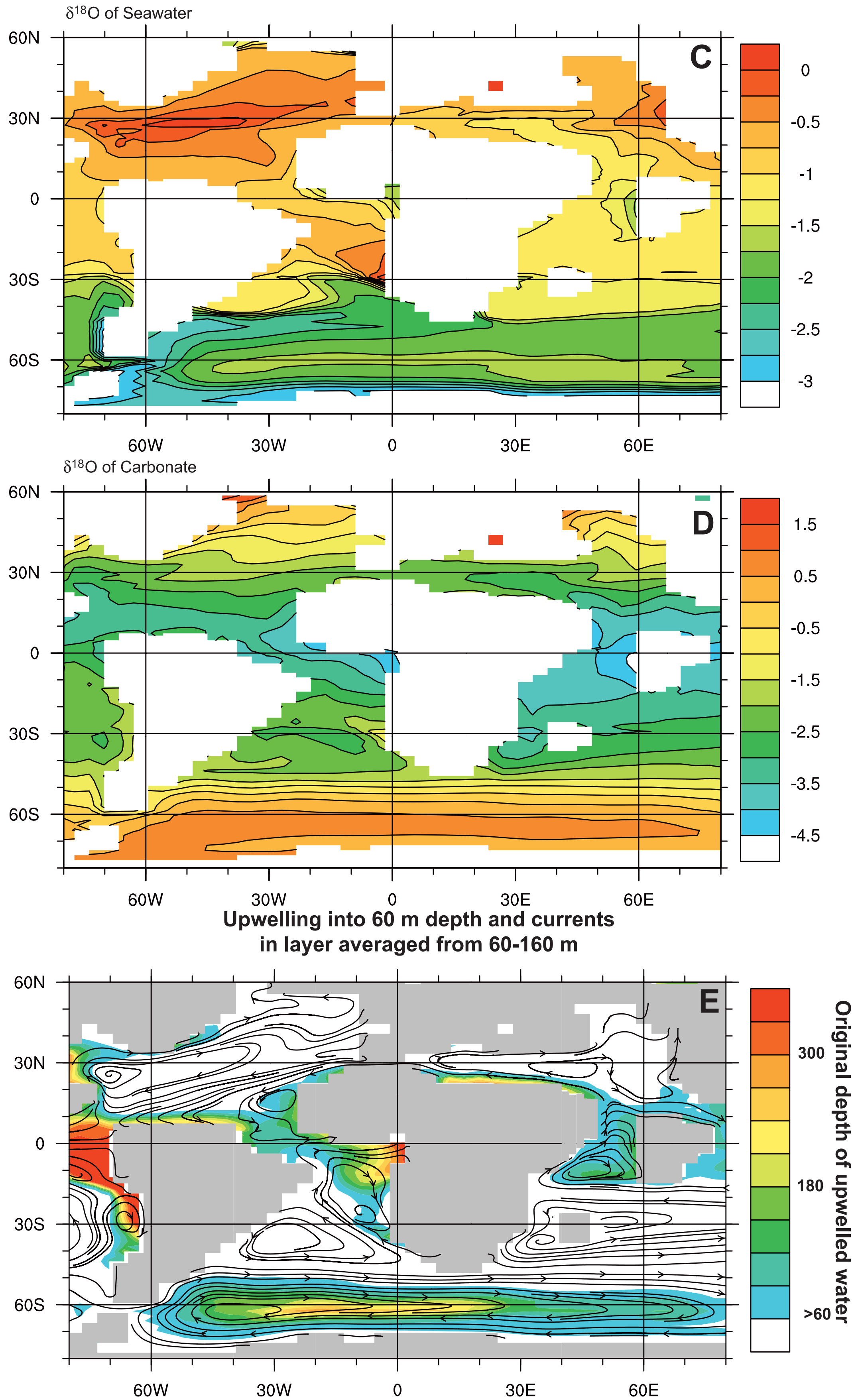



Figure 4. (*continued*)

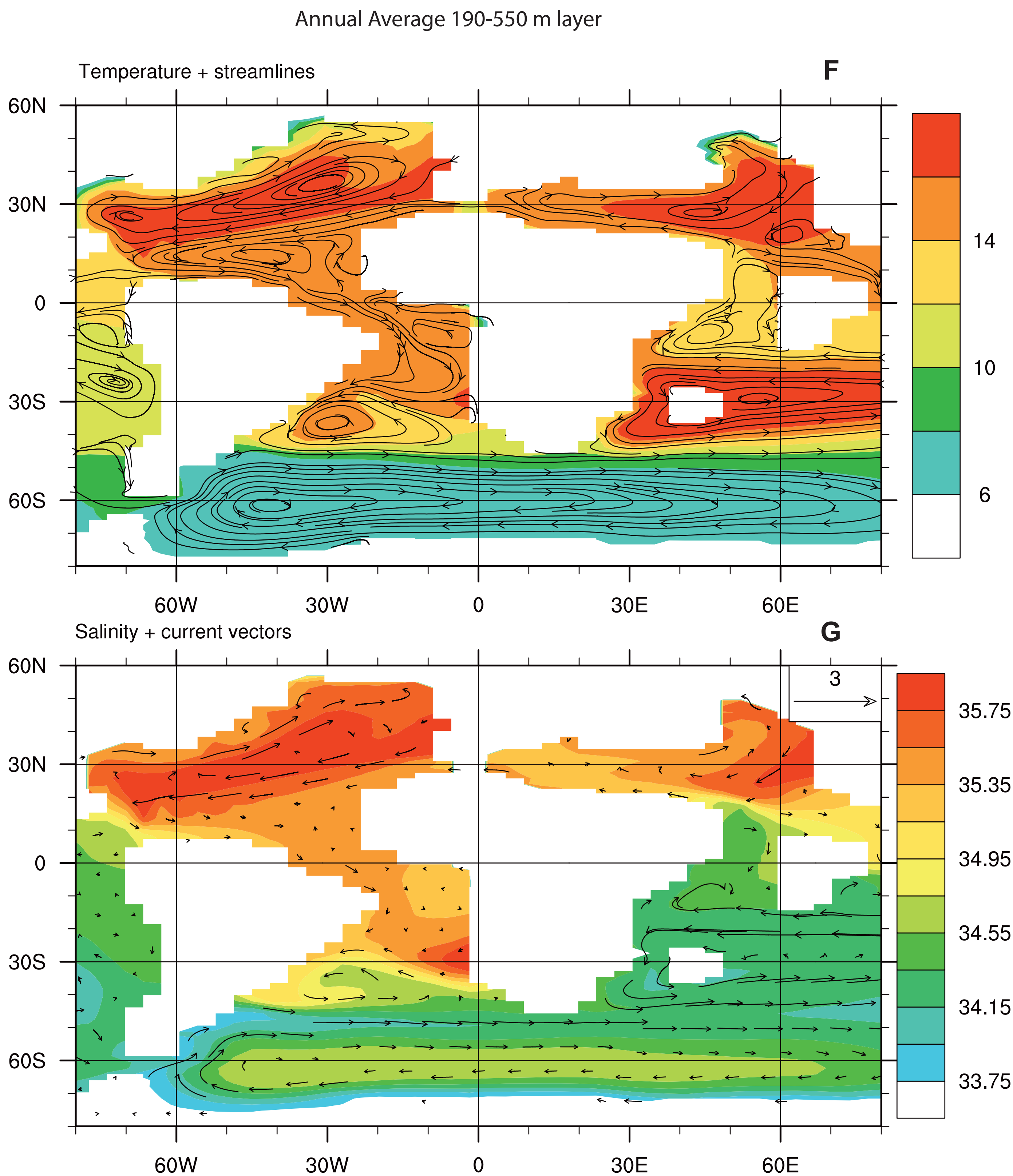



Figure 4. (*continued*)

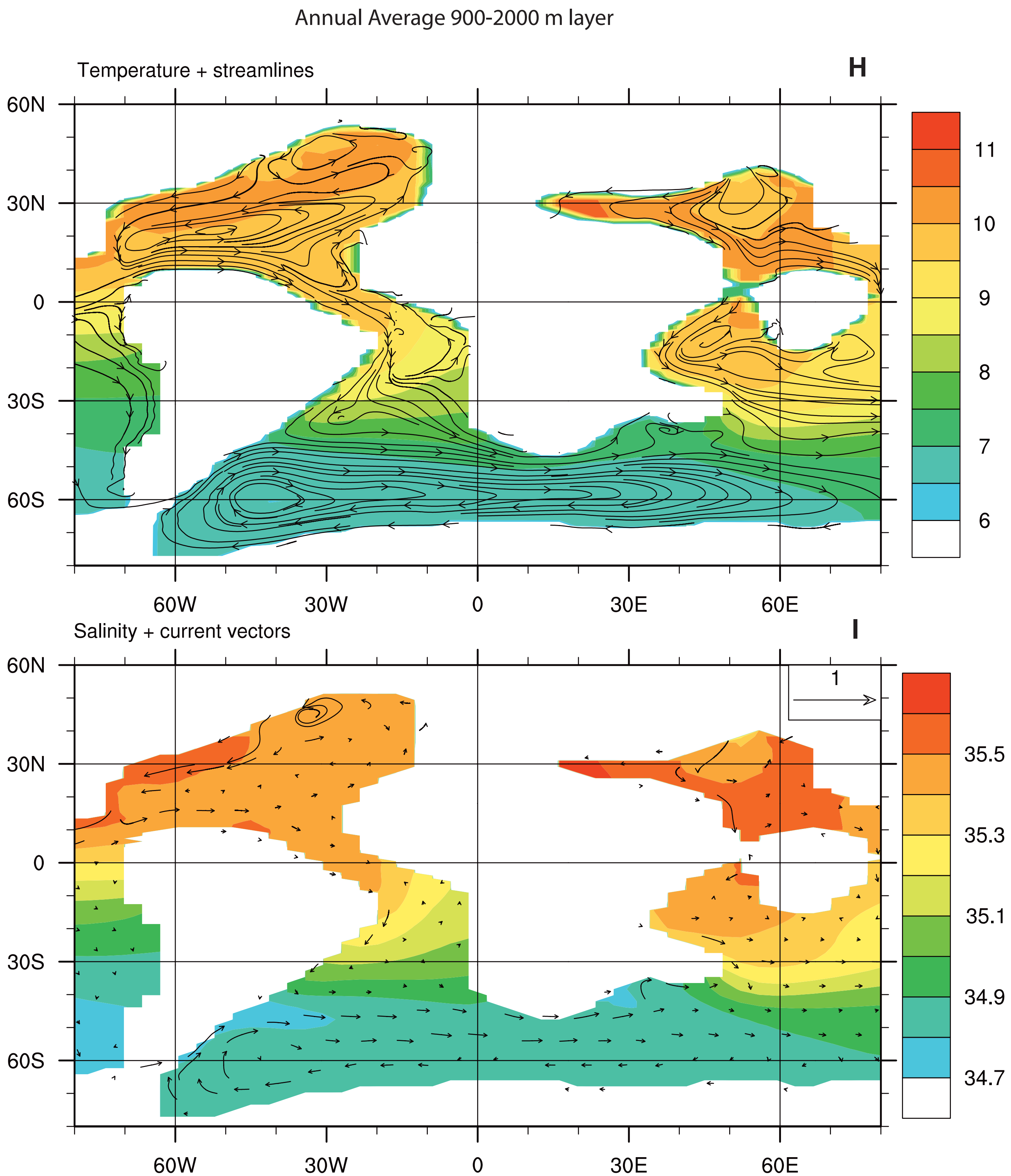



Figure 4. (*continued*)

and eastwards. A subpolar gyre, a Labrador Current-analogue, transports cooler, fresher water southwards along the east coast of North America (this feature is seen more clearly at other depths [below]).

### *Oxygen isotopes*

In Figure 4C, an estimate of seawater $\delta^{18}O$ ($\delta_w$ hereafter) is plotted. The $\delta_w$ distribution is calculated from the empirical relation between seawater salinity, and $\delta_w$ gradients proposed by Fairbanks et al. (1997) for the equatorial region (within 30° of the equator), and the relationship of Broecker (1989) in the extratropics. The strengths and weaknesses of this general technique are discussed in Schmidt (1999b) and Bice et al. (2000b). One improvement of our experimental framework over others is the use of a CGCM, so salinity is predicted from two-way ocean-atmosphere interactions, although we can not avoid the uncertainties introduced by features not well represented by the model, e.g., freshwater continental runoff (Bice et al., 1997). Values in the Southern Ocean are notably more negative than previous estimates (Zachos et al., 1994; Bice et al., 2000b), and Northern Hemisphere values more positive. A point-wise comparison is done below.

To facilitate direct comparison of our results with foraminiferal calcium carbonate $\delta^{18}O$ values ($\delta_c$ hereafter), $\delta_c$ is explicitly calculated from model output in the following way. The calculated $\delta_w$ and model predicted temperatures (averaged through the mixed layer) and the commonly used Erez and Luz (1983) relationship were used to produce our estimate for $\delta_c$ (similar to Schmidt, 1999a). In general the results are more negative (~2‰) than most proxies in the tropics and more positive (1‰) in the high latitudes. Depth habit of the planktonic foraminifera is a key area of uncertainty in any comparison because predicted values of $\delta_c$ are >1‰ more positive at 50 m depth in the tropics than at the surface (similar results were obtained by Poulsen et al., 1999). Some of the tropical results are in agreement with the new estimates of Pearson et al. (2001).

### *Upwelling*

Another important property that can be predicted by the model is the "age" of water upwelled into the mixed layer, which may be compared with previous independent model-derived estimates of upwelling and with upwelling and productivity proxies. In Figure 4E, we show the depth from which water is upwelled into 60 m depth. Greater depths are indicative of more vigorous upwelling and greater nutrient input into the mixed layer, assuming the deep ocean was a net supplier of nutrients. The coastal Gulf of Mexico, Blake Nose, the North Coast of Africa and South America, India, the Benguela Current system and Niger experience significant upwelling on an annually averaged basis. Most of the Southern Ocean ~60°S also is vigorously upwelled. Subtropical gyres including the Brazil Current system, the North Atlantic gyre, and the Agulhas-South Equatorial Current complex experience downwelling as expected. These results are in agreement with the completely independent upwelling estimates of Huber and Sloan (2000), and Handoh et al. (1999) and generally in agreement with proxies, summarized in Huber and Sloan (2000) and Sloan and Huber (2001).

### *Upper ocean*

In the above-thermocline upper ocean (an average from 190 m to 550 m), Sverdrup dynamics dominate the velocity distribution with wind-driven geostrophic gyres in evidence (Figs. 4F and 4G). Sharp fronts are evident between the Southern Ocean subpolar gyre (6 °C, 34.5 ppt) and the subtropical gyres (Brazil Current 13 °C, 34.75 ppt and Agulhas 16 °C, 34.35 ppt). It is noteworthy that there is a sharp gradient in temperature and salinity in the South Atlantic at the subpolar front, but there is no sharp gradient in salinity at the subpolar front in the southern Indian Ocean. Transport along the eastern margin of South America is southward across the equator, bringing a signal of warm (14 °C, 35.4 ppt) and "young" water (some of the water has been at the surface within 100 yr) into the South Atlantic. In the North Atlantic, a strong Gulf Stream flows north and east, but it is noteworthy that the front between this current and the Labrador Current analogue extends down to the location of ODP cores from Blake Nose. This suggests that substantial variability (>4 °C or 1‰ $\delta^{18}O$) may occur at this site due to slight changes in the position of the Gulf Stream. This may explain some of the variability noted by Wade et al. (2000) at Blake Nose.

### *Deep ocean*

The majority of "deep" ocean proxy data are from paleodepths <3000 m and ocean properties become more homogenous with depth, so we average vertically between 900 and 2000 m (Figs. 4H and 4I). The hemispheric asymmetry in temperature and salinity, with cooler, fresher values in the Southern Hemisphere, is apparent at this depth, although the contrasts are weaker. There is southward transport of warm, salty water into the Southern Hemisphere to balance the northward transport in the upper layers. Temperature decreases with depth in the North Atlantic, but in the South Atlantic, deeper values are significantly warmer than upper ocean and surface values. This demonstrates that the sort of "inversion" in temperature seen in Southern Ocean sites (e.g., 689 and 690 during the PETM) can be explained by warm deep water formation at high latitudes rather than at low latitudes in contradiction to the Proteus Ocean hypothesis by Kennett and Stott (1990).

### *Mean annual temperature range*

In the near surface ocean, the seasonal cycle of temperature is an important factor for understanding climate and the interpretation of climate proxies (e.g., Andreasson and Schmitz, 1998). The seasonal temperature cycle also provides a connection between ocean processes and the equable climate problem, so the mean annual temperature range (MATR) over both land and sea is shown in Figure 5. MATR over land and ocean is similar in our early Paleogene and modern control case with some exceptions. Large decreases in the magnitude of MATR occur over land in North Africa, Siberia, Antarctica, and the Arctic Ocean. Increases in MATR occur in Australia and central Canada. MATR over the oceans is virtually unchanged from near-modern values.

### *Pointwise comparison with data*

Mean annual temperatures (MAT) predicted by the model are compared to data in Figure 6. Spatially collocated values for mean annual SST predicted by the model and those produced from the Zachos et al. (1994) $\delta_c$ compilation have similar values except in regions poleward of 40° latitude (Fig. 6A). The model underpredicts SSTs in high latitudes by 5–10 °C. When the model-predicted $\delta_w$ distribution is used instead of the empirical function of latitude proposed by Zachos et al. (1994), estimated SST values decrease by 2–3 °C in the Southern Ocean and by 5–10 °C in the Southern Hemisphere midlatitude oceans, whereas there is little difference in the Northern Hemisphere (Fig. 6B). Although inclusion of the model-predicted $\delta_w$ distribution improves (by 20%) the model-data mismatch at low temperatures, it weakens the correlation at temperatures between 20 and 25 °C (Fig. 6C).

A similar bias is seen in the comparison with terrestrial MAT values (Fig. 6D). The model accurately captures the MAT in one compilation of values from Wyoming and Baja California. Other compilations of MAT in Wyoming, including the high-resolution records of Wing et al. (2000) and Wilf (2000), show substantial variability, from near the model-predicted value to significantly higher values. These warmer values fall along a line that is characteristic of other (low-resolution) terrestrial MAT estimates from the early Eocene. In general, compared to these warm estimates, the model is offset uniformly by 10 °C.

## DISCUSSION AND IMPLICATIONS

### *Comparison to results of previous coupled models*

The basic patterns and processes produced in this study are not qualitatively different from those that appear in the modern day, and the early Paleogene results are generally in agreement with previous coupled model studies. Since this is the first fully coupled model study of the early Paleogene, there are some meaningful comparisons we can make with other coupled model studies.

The study by Saravanan and McWilliams (1995) included a simplified 2-layer atmosphere with rudimentary treatments of moist processes, clouds, and radiation coupled to a low-resolution "sector" ocean model, i.e., a zonally averaged ocean that includes an ad hoc treatment of important features such as western boundary currents, and no treatment of sea-ice. Although not focused on past warm climates, with modern $pCO_2$ concentrations, this study produced three possible ocean circulation modes, and indicated a near-constant total poleward heat transport (atmosphere + ocean) and small changes in temperature gradients ($\Delta T = 28–34$ °C), with higher $pCO_2$. Using a different kind of simplified model, Schmidt and Mysak (1996) conducted a study of past warm climates within a framework in which simple "sector" oceans were connected in an idealized ocean basin configuration. This ocean model was coupled to an atmospheric model in which atmospheric heat fluxes were constrained to maintain a steady climate in equilibrium with the initial (specified) SST distribution. Because this framework assumes the existence of whatever atmospheric fluxes are necessary for steady state, this technique cannot discern whether a climate state with small temperature gradients is possible. Nor can greenhouse gas concentrations be meaningfully changed, because as Schmidt and Mysak acknowledge, "the atmospheric model can be considered to take account of whatever $CO_2$ level is necessary to maintain the specified initial climate." Instead this technique allows the estimation of the ocean and atmospheric heat fluxes necessary to maintain different temperature profiles, and the deep ocean temperature and overturning circulations in equilibrium with those values. The existence of Indian Ocean sourced salty deep water produced in the Schmidt and Mysak (1996) study is consistent with the Tethyan watermass formed in our study.

Bush and Philander (1997) performed a fully coupled simulation of the past warm climate of the Cretaceous that produced surface temperature gradients similar to those of our study, but that simulation was only integrated for 35 yr as opposed to 2200 yr in this study. Otto-Bliesner et al. (2002) have performed fully coupled simulations of Cretaceous climate also with NCAR's CCSM, but the spin-up technique and boundary conditions, including $pCO_2$, are different than in our study (their simulation is for 100 Ma). Some key features of our results and those of Otto-Bliesner et al. (2002) are similar: Their deep ocean temperatures of 8–10 °C (for 1120 ppm $pCO_2$) compare favorably with our results (given our lower $pCO_2$); warm deep water forms in the extratropics in both experiments, the meridional temperature gradients produced are also similar (compare with Huber and Sloan, 2001). Importantly, their simulation shows extremely vigorous Southern Ocean deep water formation (>20Sv), and with it, somewhat increased heat transport in the Southern Hemisphere (relative to the values in our simulation), although in general their heat transport values and ours agree.

To our knowledge no simulation with coupled dynamical models has produced a low-gradient climate, i.e. a $\Delta T$ as low as 15 °C. The $\Delta T$ and ocean heat transport produced in this study is well within the range expected from past coupled work.

### *Overturning, deep water formation and multiple equilibria*

Averaged over all basins, the meridional overturning circulation is weaker than modern (see Huber and Sloan, 2001), but

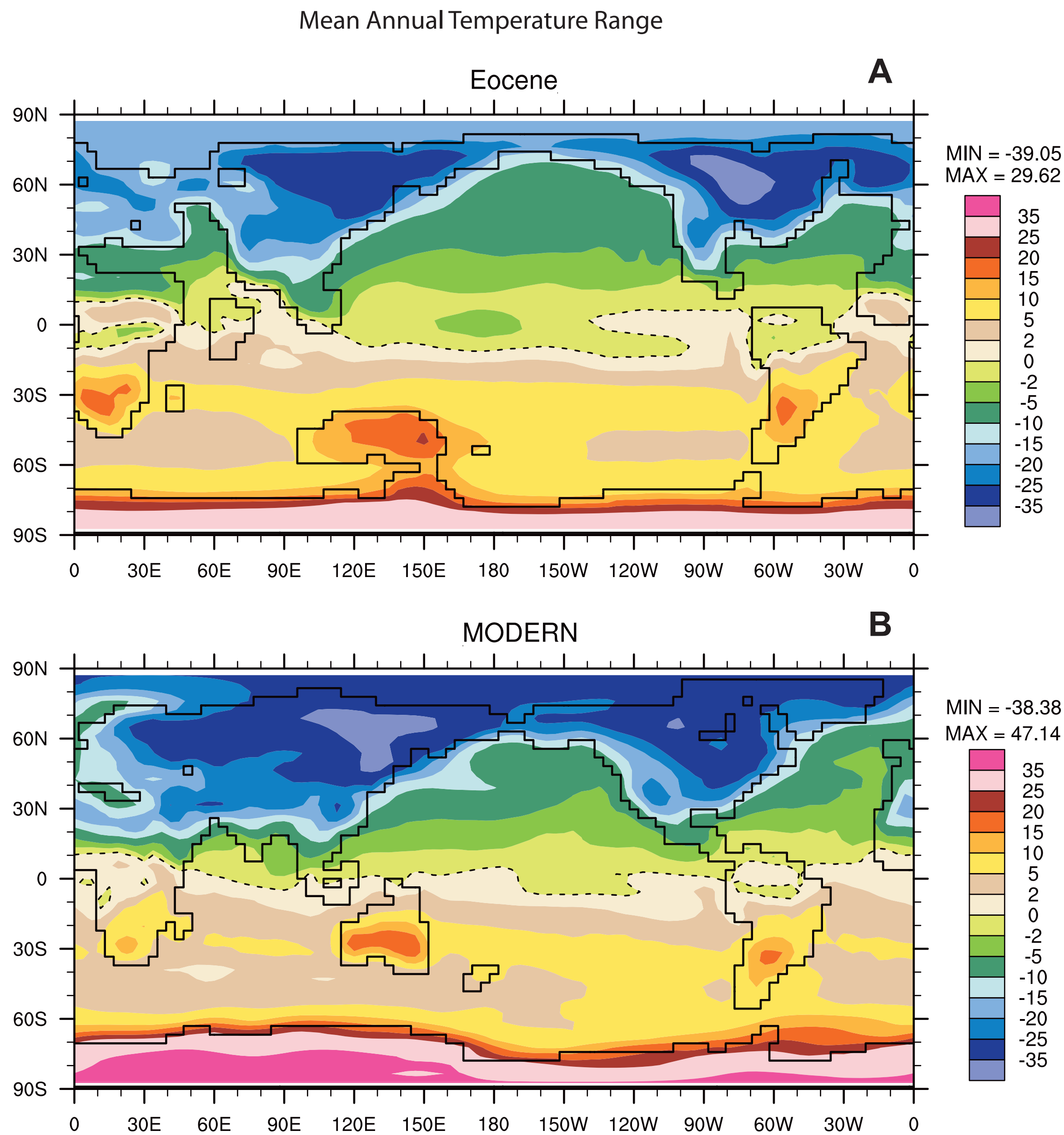



Figure 5. Mean annual temperature range for the early Paleogene (A) and the modern "control" case (B). Dashed line is the zero line. The absolute value of these fields is the relevant climate parameter.

within the individual basins considered here, the ocean velocities and overturning circulations are not dramatically reduced compared to modern (compare with Gent et al., 1998). Deep convection occurs in this simulation in the North Atlantic and Tethys (similar to Barron and Petersen, 1991), and these warm, salty water masses penetrate and mix into the Southern Ocean water masses that are cooler and fresher. In the simulation focused on here, the Southern Ocean is extremely fresh and deep ocean convection is suppressed (although see below for the second simulation).

The net effect is a noticeable hemispheric asymmetry in tracer properties and flow patterns. This bears some resemblance to the "bipolar" ocean circulation proposed by Haupt and Seidov (2001), with a cool overturning cell in one hemisphere and a warm cell in the other. The actual flow is more complex, and one impact of the complex flow patterns is the import of warm, salty NADW into the Southern Ocean, which leads to warmer temperatures at depth in the South Atlantic. In general, the distribution provides an explanation for the isotopic pattern noted in the North Atlantic by Charisi and Schmitz (1996) and

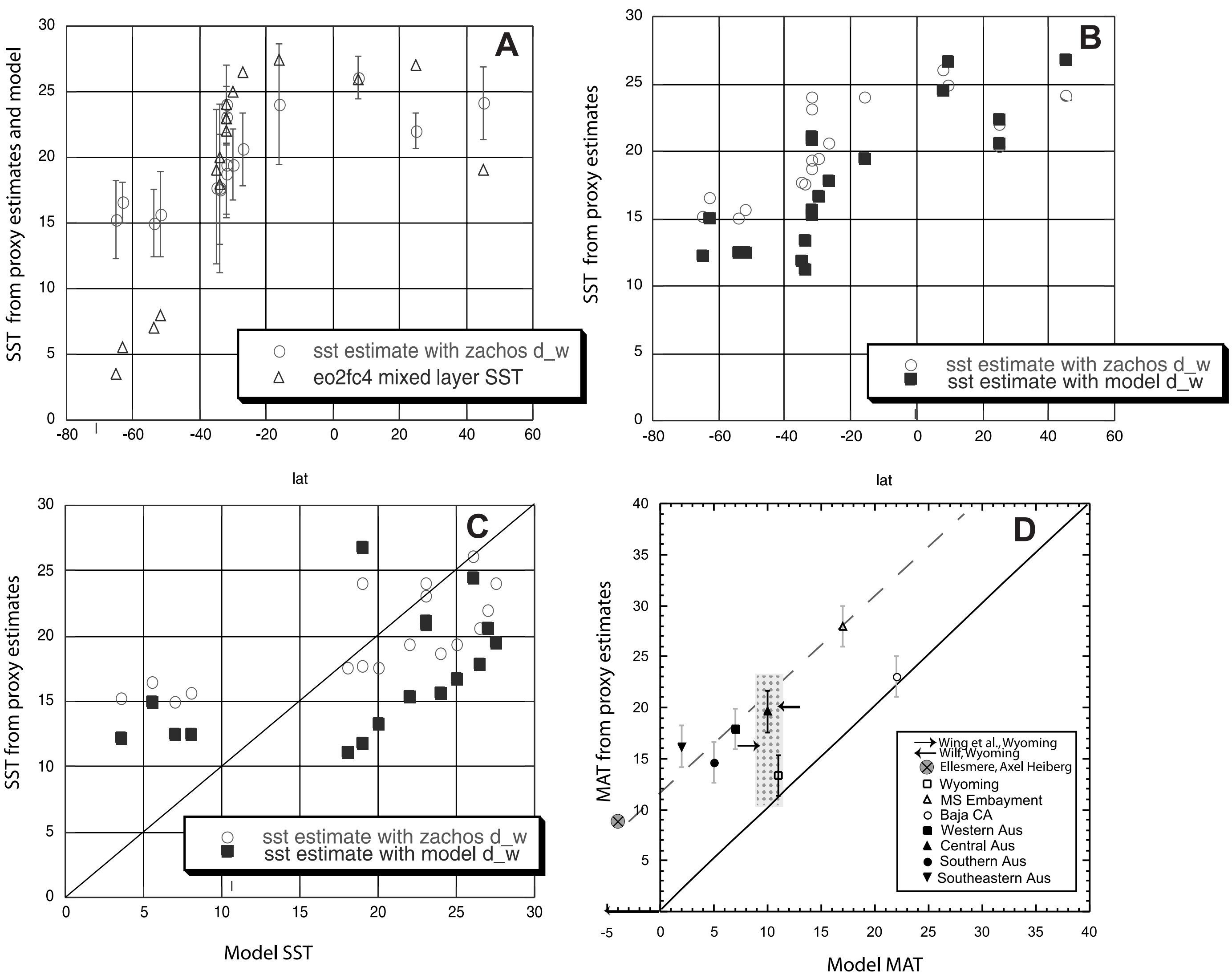



Figure 6. In A the compilation of Zachos et al. (1994) is used to estimate sea surface temperature (SST; circles) and compared with model predictions (triangles). A comparison of SST estimates with the Zachos et al. (1994) latitude dependent calculation of the isotopic composition of seawater and with those recalculated using the salinity produced in this study, from empirical relationships between salinity and isotopic composition of seawater (described in text) is shown in B. The magnitude of the difference between these two estimates is used to create the error bars in A, thus the error indicated is only that introduced by different assumptions about seawater oxygen isotopic gradients. C: Crossplot of model predicted temperatures against proxy estimates of temperature incorporating the Zachos et al. (1994) isotopic composition of seawater (circles) and those predicted in this study. A perfect model would produce values along the black diagonal. In D, mean annual temperature (MAT) from terrestrial proxies are plotted against MAT from the model. The data are from compilations by Sloan and Barron (1992), Greenwood and Wing (1995), Wilf (2000), Wing et al. (2000). The grey stippled zone represents the range of data in the latter two high-resolution studies and the arrows represent the means of those studies.

Corfield and Norris (1996) and in the South Atlantic by Kennett and Stott (1990).

There is some evidence from "aging" gradients of carbon isotopes that deep water tended to form in the Southern Ocean (e.g., Miller et al., 1987; Corfield and Norris, 1996). Because of the potential for multiple equilibria our results do not rule out a switch in modes whereby with the same initial conditions the simulation might evolve to a state with sinking in the Southern Ocean. In fact, the case beginning with LOGRAD initial conditions (not discussed here in detail) produces deep convection in the Southern Ocean, in addition to the NADW and Tethyan sources. The climatic differences, including ocean heat transport and surface temperature, between the two simulations is relatively small with <1 °C temperature change over most of the Earth and <0.1 PW over most of the oceans. The changes in deep water formation may, however, be important in explaining isotopic records. Existing carbon isotope reconstructions are very sparsely distributed, tending to be underrepresented in the North Atlantic; the model-predicted flow patterns are complex, indicating that the simple water mass trajectories assumed in these reconstructions may be questioned. For example, the model results indicate that, at some times in the early Paleogene, the deep central Pacific may have been bathed in "young" water that recently flowed in from the North Atlantic, thus aging gradients between the two regions might be erased by short circuits in the ocean circulation.

### *Ocean heat transport*

Neither the details of deep water formation properties, nor any of the changes to boundary conditions, prove to be critical to understanding how much heat the ocean transports, given the remarkable consistency of the ocean heat transport produced in these simulations. Heat transport produced by the early Paleogene coupled model is not very different than the heat transport implied by the initial HIGRAD SST distribution, but it is somewhat reduced in high latitudes, which is consistent with the cooling there. The other simulation, which we did not describe, began with the LOGRAD SST distribution and rapidly converged on nearly the same SST pattern described in this study, with a nearly identical pattern of ocean heat transport, despite a different meridional overturning structure. Thus, the very vigorous ocean heat transport implied by the LOGRAD distribution is not maintainable in a fully coupled model with early Paleogene conditions—instead, near-modern values are robustly predicted. This implies that previous studies that used type 2 models, which assume near-modern ocean heat transport, are valid in that assumption. This means that results from those previous simulations remain valid and the type 2 experimental framework may still be applied to good advantage in future attempts to understand early Paleogene climate.

It also raises serious questions about studies of type 3 models that produce vastly different ocean heat transports than implied by the atmospheric model output with which they are driven. Our Southern Hemisphere ocean heat transport results are significantly different in magnitude, although not in direction, from those produced for the early Paleogene by Bice et al. (2000a). In fact, none of the fully coupled dynamical models described above produce ocean heat transports of the magnitude found in the Bice et al. (2000a) study. As argued earlier, we believe that the results of their study are most likely due to the SST boundary condition, because it assumes that the atmosphere is an infinite source of heat and it ignores negative feedbacks (i.e. up to 40 $W/m^2/°K$) imposed by the atmosphere (Rahmstorf and Willebrand, 1995; Nong et al., 2000). This is confirmed in an OGCM study by Najjar et al. (2002), which, using bathymetry and other conditions nearly identical to those in the Bice et al. (2000a) study, predicts values of Southern Ocean heat transport in agreement with those here. The main improvement of the Najjar et al. (2002) study is the implementation of a less artificial way of imposing the SST boundary condition. The agreement between our study and that of Najjar et al. (2002) is further indication that the results presented here are robust, which suggests that caution should be used in the interpretation of links between gateway changes and large ocean heat transport changes.

### *The low gradient and equable climate problems*

In equilibrium with early Paleogene boundary conditions and with greenhouse gas concentrations at the low end of proxy estimates, deep ocean temperatures are 7 °C warmer than in the degraded modern control case. Tropical SSTs are ~3 °C warmer. Thus, ΔT is 3–4 °C lower in the early Paleogene simulation, but temperatures in high latitudes are still too cold (by 5–10 °C) and tropical temperatures (by 0–5 °C) too warm relative to existing proxies. The salinity distribution produced by the simulation suggests that existing proxy interpretations may overestimate high-latitude SSTs by 3–5 °C in the Southern Ocean, but underestimate them in the North Atlantic so the net effect is probably small.

Greenhouse gas concentrations in the upper range predicted from some proxies, e.g., Pearson and Palmer (2001), may alleviate the cold bias at high latitudes, but only at the expense of further raising tropical SSTs. The results of Pearson et al. (2001) are encouraging in that regard, but it is not clear whether the upward revision of tropical SSTs suggested from their results is sufficient to allow a resolution of the low gradient and equable climate problems. Our results disprove the speculation by Bice et al. (2000a) that a 2 °C warming above present tropical-subtropical SST values might solve the problem of polar warmth, via latent heat transport feedbacks. Our simulation, which includes latent heat transport feedbacks, produces a 3 °C tropical warming, but does not solve the equable climate problem or the problem of polar warmth.

Other variables need to be constrained to more precisely define the data-model mismatch in the tropics, because possible sampling biases introduce a large degree of uncertainty in tropical SST estimates. Our simulations indicate that if planktonic foraminifera record tropical temperatures from >40 m depth, rather than 6 m, then the low gradient problem disappears (in keeping with the results of Poulsen et al., 1999). This depth habit may be difficult to reconcile with evidence for the presence of photosymbionts associated with many foraminifera in existing reconstructions. Alternatively, our results indicate that a sampling bias toward warm season values in high latitudes and cold (upwelling) season temperatures in low latitudes could also partially resolve the low gradient problem. Warm terrestrial extratropical winter temperatures tend to rule out the existence of a large high-latitude bias in planktonic foraminiferal SST estimates, but the cool bias may still exist in the tropics.

### *Caveats*

Because we have not constrained certain climate properties, e.g., SSTs, to be near values close to those reconstructed from proxies or made explicit assumptions based on modern relationships (with the exception of the isotope calculations), the simulations should be useful as an independent prediction for comparison with proxy interpretations. This also means that some of the predictions may not be as close to proxy interpretations as those from simulations that were more constrained. This is evident when comparing the model results of Bice et al. (2000b) to those here.

An important source of uncertainty in model interpretation is the treatment of runoff, as pointed out by Bice et al. (1997). In this version of CCSM, runoff is collected over all land points and distributed in a weighted fashion over the ocean, with the weighting dependent on the precipitation amount. In other words, CCSM puts most runoff where rainfall is greatest, tending to increase spatial gradients in salinity, but otherwise tending to mask the geographically local effects of runoff. Sensitivity to this will be tested in future work, but we note that the degraded modern case, which begins with zonally constant salinity distributions, produced reasonable salinity distributions and deep water formation properties (i.e., in the North Atlantic not the North Pacific) with the same runoff treatment.

Another source of uncertainty is sensitivity to initial conditions and the technique for spin-up. Two early Paleogene simulations have been performed, with very different initial conditions, that both reached very similar upper ocean properties, current structures, temperature gradients, and poleward flux characteristics, so we think this is not an important source of uncertainty. In work in progress we are experimenting with other techniques for accelerating the coupled model to an equilibrated deep ocean, with good preliminary agreement with the results here. The success of the degraded modern case (Huber and Sloan, 2001) gives us further confidence in the methods.

Vertical diffusion in the ocean is currently a controversial and important issue in understanding heat transport and the meridional overturning. Changes in oceanic vertical diffusion have been proposed by Lyle (1997) to explain past warm climates. The treatment of vertical diffusion in CCSM includes a formulation that depends on vertical density gradients (Richardson number dependence) capturing the essence of the Lyle mechanism, but we find no evidence of important consequences from this dependency. Nevertheless, it is too early to rule out changes in diffusion due to changes in tidal dissipation or bathymetric roughness (e.g., Egbert and Ray, 2001) as an important contributor to past warm climates.

### *Implications*

These results strongly suggest that the ocean circulation is not the solution to the low gradient or equable climate problems. Nor are atmospheric feedbacks, at least as currently exist in this comprehensive model. Our coupled simulations lend credence to the oft-noted compensation of atmospheric and oceanic fluxes, and the near constancy of meridional temperature gradients suggested by climate theory (cf. Saravanan and McWilliams, 1995). The fact that the model is in keeping with these latter characterizations—but at odds with some early Paleogene climate proxies—lends support to the notion that there is an important component of climate that we still do not properly understand, or that the proxy data are being interpreted incorrectly. If the model is essentially correct, then high greenhouse concentrations (e.g., $p\mathrm{CO_2}$ >1000 ppm) may solve the equable climate problem, but only if tropical SSTs are warmer than most current reconstructions. It remains to be seen how large the discrepancy between SSTs predicted by the coupled model, and those from proxy reconstructions, will be under higher greenhouse gas concentrations.

## SUMMARY

Results from previous work described here demonstrate that (1) changes to land surface characteristics have important consequences at the local and regional scales, but not at large scales; (2) specification of warm polar SSTs does little, in and of itself, to warm continental interiors during the winter; (3) midlatitude continental interior temperatures are at least as sensitive—if not more sensitive—to changes in tropical SSTs as to extratropical SSTs.

Results from the simulations in this study demonstrate that:

1. In this model, a transition occurs between an equator-to-pole SST gradient of 24 and 15 °C, at which point the temperature gradient becomes a major climate problem.

2. This early Paleogene CGCM simulation produces continental interior temperatures that are unrealistically cold, showing that a complete and interactive treatment of the ocean is not, by itself, a resolution to the equable climate problem. Model-predicted mean tropical temperatures are closer to those of Pearson et al. (2001) than to those of Zachos et al. (1994).

3. Although zonally averaged overturning is reduced in the early Paleogene simulation relative to modern, the overturning and current systems within individual basins can be as vigorous as modern.

4. Deep waters can be relatively warm (7 °C) and these form at high latitudes. Deep convection of warm (12–15 °C) and saline (35.5 ppt) water occurs in the Northern Hemisphere. The Southern Ocean is relatively fresh (~32 ppt) and surface temperatures are cooler (1–4 °C). In the two simulations described here, one undergoes deep convection in the Southern Hemisphere and the other does not, indicating there may be two quasi-stable states of the early Paleogene system.

5. In all cases, the simulation produces temperature gradients and ocean heat transport within 30% of modern, which suggests that it is unlikely that increased meridional heat transport was responsible for small temperature gradients in the early Paleogene.

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# Early Paleogene oceans and climate: A fully coupled modeling approach using the NCAR CCSM

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Notes