Climate Future in a Warming World: Lessons from the Ice Ages

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Abstract- **Global warming is real and has been with us for at least two decades. Questions arise regarding the response of the ocean to greenhouse forcing, including expectations for changes in ocean circulation, in uptake of excess carbon dioxide, and in upwelling activity. The large climate variations of the ice ages, within the last million years, offer the opportunity to study responses of the ocean to climate change. A histogram of sealevel positions for the last 700,000 years (based on a new d^{18}O stratigraphy here compiled) shows that the present is near the margin of the range of fluctuations, with only 6 percent of positions indicating a warmer climate. Thus, the future will be largely outside of experience with regard to fluctuations of the recent geologic past. The same is true for greenhouse forcing. Our inability to explain sudden climate change in the past, including the rapid rise of carbon dioxide during deglaciation, and differences in ocean productivity between glacial and interglacial conditions, demonstrates a lack of understanding that makes predictions suspect. This is the lesson from ice age studies.**

I. INTRODUCTION

Global warming is a reality and has been with us for at least two decades. A number of pressing question arise in the context, such as: How will the ocean circulation respond to further warming? Will central gyres expand? Will the deep circulation become less vigorous? How will such change affect the uptake of carbon dioxide by the deep ocean from the atmosphere? What will be the effect on the ocean's ecosystems? Will there be more fish or fewer? What about marine mammals and sea birds?

At this time it is exceedingly difficult to address such question from first principles; our understanding is insufficient. The past certainly offers useful insights, but we must be aware that this approach also has its drawbacks. For once, the future has no analog in the past. We do not understand the carbon cycle, as shown by an inability to explain the rise in carbon dioxide at the end of the last glacial (and all previous terminations). We do not understand ocean productivity, as shown by the recent discussions involving the maintenance of high nutrient concentrations in surface waters over large regions in the Pacific and in the Circumpolar Current. And we do not understand system stability, as shown by the discussions surrounding the high rate of deglaciation, and the origin of a cold phase during the transition of glacial to postglacial time.

Despite these basic deficiencies in comprehension, we readily gain an appreciation for surprises which may be ahead by studying the record.

II. THE FUTURE HAS NO ANALOG IN THE PAST

One important ingredient of climate change readily emerges even from a qualitative consideration of the evidence: the large role of positive feedback from albedo. The record tells us that the largest changes on the surface of Earth took place in the Arctic and subarctic regions. This region is also the one where albedo changes are greatest, and where pronounced future changes are expected. But we cannot simply assign to future conditions a probable position on the range of climate conditions in the last million years. Only during 6 percent of the time was the climate warmer than today. To show this, a revised oxygen isotope standard ("OJplus03") is offered here back to 0.838 Ma, based on data given in Berger et al. (1996), and complemented by data in Bassinot et al. (1994) and in Mix et al. (1995) (Figure 1). Stacking procedure and results are given in Table 1. The resulting histogram yields the 6 percent value for conditions warmer than today (Figure 2). (I have not used the usually employed SPECMAP scale because it is unreliable for the older part of the record, since Imbrie et al. 1984 assumed the then current age of 730 kyr for the Brunhes-Matuyama boundary, an age that is too low by 7 to 8 percent.)

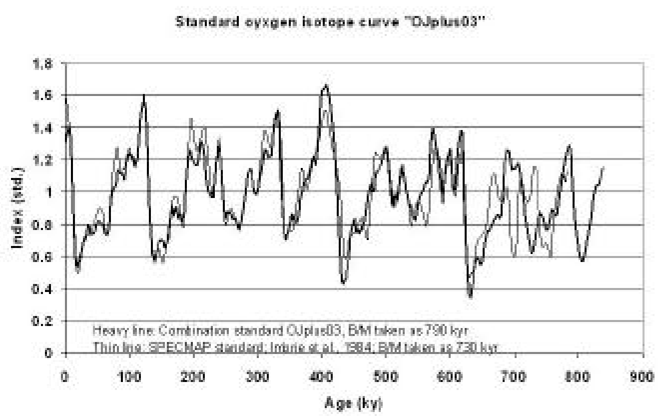

Fig. 1. Standard oxygen isotope curve "Ojplus03" (data in Table 1) compared with SPECMAP curve (Imbrie et al., 1984).

In addition, there is no past analog for the present carbon dioxide content of the atmosphere, as far as that can be determined, for the last 800,000 years. For the last 400,000 years this is demonstrated by the gas content in Antarctic ice (Petit et al., 1999). For the previous 400,000 years the range is probably not much larger, based on extrapolation using correlations with deep-sea stratigraphy (Berger et al., 1996).

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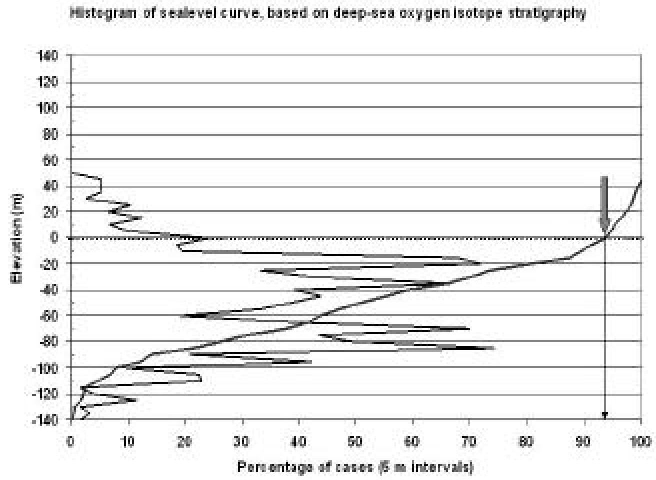

Fig. 2. Histogram of "OJplus03" oxygen isotope curve, transposed into sealevel variation by assuming a typical glacial-postglacial range of 110 m. Note that values higher than today only occurred 6 percent of the time.

III. CARBON CYCLE

An enormous literature exists about how the ocean controls the concentration of carbon dioxide in the atmosphere. A great number of communicating carbon reservoirs are involved. The estimates of fluxes between these reservoirs are tagged with considerable error. Since the atmosphere is a relatively small reservoir and is intimately linked to a large one – the ocean – the uncertainties concerning the expected atmospheric level, for different conditions, are large. We now know, thanks to the ice-core results, that the range of CO_2 values is rather well defined. There is a lower boundary near 180 ppm and an upper one near 300 ppm. The rise from lower values in glacial periods (around 190 ppm) to higher values in interglacials (around 290 ppm) is remarkably rapid and follows closely on the temperature changes seen in the ice (Figure 3). These, in turn, are highly correlated with the oxygen isotope record (Berger et al., 1996). No generally accepted combination of mechanisms has been proposed, so far, to account for the change of 100 ppm seem. In addition, there is the equivalent of another 40 ppm or so that is hidden in the growth of forests and associated soil carbon pools.

It is unlikely that the mix of mechanisms responsible for providing all this carbon to the atmosphere, at the time of deglaciation, will soon emerge. However, there is a message here: The system reacts to warming and sealevel rise with an increase of atmospheric carbon dioxide (Petit et al., 1999). Many models assume some sort of equilibrium between the ocean and the atmosphere, in regard to carbon dioxide. Clearly, it is unlikely that equilibrium prevails during deglaciation and rapid rise of sea level. On the contrary, transient phenomena, such as cessation of deep circulation and destruction of nitrate at depth must be considered.

The lesson is that equilibrium thinking is unlikely to contribute to an appreciation of what the future might bring. In any case, our inability to explain the past is telling.

IV. OCEAN PRODUCTIVITY

The role of iron in controlling the productivity of large parts of the ocean is the topic of lively discussions in biological oceanography, ever since the late John Martin and colleagues urged iron deficiency to explain high nutrient content of surface waters covering vast regions of the Pacific (Martin and Fitzwater, 1988). Some have maintained that, rather than nutrient deficiencies, it is rapid grazing that prevents phytoplankton from depleting the surface waters. Obviously, these are not minor differences in opinion. If one cannot agree which is more important, nutrient control or grazing, the system is not well understood.

Geologists studying the record of productivity have come up against a paradox, especially well illustrated in Pliocene-Pleistocene sediments off Namibia, whereby

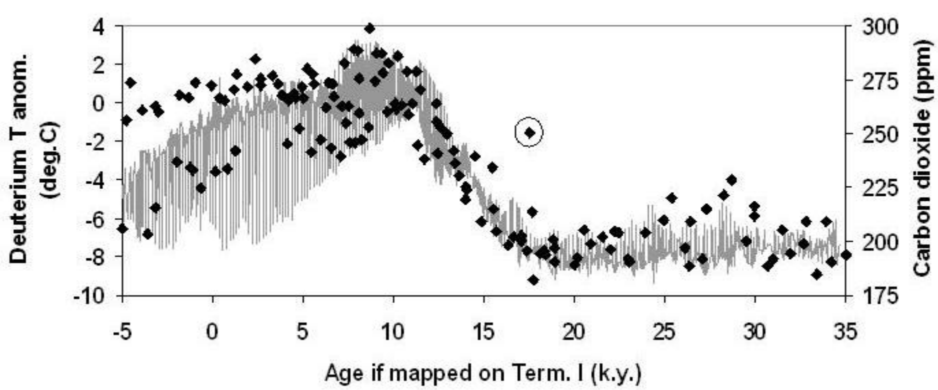

Fig. 3. Deuterium and carbon dioxide data from Vostok ice core (Petit et al., 1999) mapped on Termination I to show regularity of the course of deglaciation. Circled carbon dioxide value: highly unusual.

increased upwelling stimulated by the intensified winds of glacial periods resulted in a diminished supply of diatom debris to the sea floor (Berger and Wefer, 2002). Traditionally, it had been assumed that the abundance of diatoms is a good indicator of paleoproductivity. Indeed, a drastic drop in productivity of the California Current, well documented in the CalCOFI survey data (McGowan et al., 1996), was initially proposed from noting a strong decrease in diatom deposition in the Santa Barbara Basin, where anaerobic conditions permit preservation of a year-by-year record (Lange et al., 1990). However, global changes in nutrient content of the deep sea must be considered as an equal partner with the rate of upwelling, when attempting to reconstruct ocean productivity.

As concerns the likely effect of warming on ocean productivity, it is apparently closely tied to the response of the trade wind system. In the record, in any case, strong trade winds during glacial times produce increased upwelling both in coastal regions and along the equator. On the whole, trade winds are a response to the planetary temperature gradient. As this gradient weakens (because of the warming of high latitudes) we should expect a corresponding weakening of the trade winds. Predators that depend on high food density (such as marine mammals and sea birds) would then suffer from a general decrease in the focusing of ocean productivity. Fisheries depending on herring-like fish that strain productive waters also will suffer, beyond the damage done by overfishing.

Fig. 4. Seabirds, because of their high metabolism, depend on high productivity of the sea, and are vulnerable to a drop in productivity from warming.

V. SUDDEN CLIMATE CHANGE

The largest abrupt change in recent times, geologically speaking, is the transition from the last glacial maximum to postglacial conditions in the early Holocene. Much has been written about the possible mechanisms driving this change. Some invoke changes in the circulation of the ocean, others call on collapse of ice sheets. The truth, perhaps not surprisingly, may be in a compromise that allows a number of mechanisms to interact, in a feedback system.

There is no question that marine-based ice is more vulnerable to destruction than ice safely anchored on solid dry ground. As the land sinks under the ice masses piling up during a glacial period, more and more of the ice becomes marine-based. Also, erosion of shelves on time scales of hundreds of thousands of years is important in this process of destabilization. Large regions once deeply covered by ice are now under water, from Hudson Bay on northward into the Canadian archipelago. Thus, geography itself reminds us of the likely importance of this factor, in the deglaciation process. Deep fjords enter far into the land once covered by the Scandinavian ice shield, and the Baltic Sea is a perfect analog for Hudson Bay.

The point is that a relatively small rise in sea level, by removing support from marine-based ice, can trigger substantial surging, and also open access for seawater to the base of previously grounded ice. Whenever such thresholds are involved, prediction is made virtually impossible, except perhaps in a general statistical sense.

It is noteworthy that deep fjords enter into Greenland, serving potentially as venues for heat carried landward by seawater. The deeply eroded shelves around Antarctica, and the fact that much of the ice in West Antarctica is marine-based, invite speculation about the stability of these ice sheets. In fact, there has been considerable discussion about possible contributions of Antarctic ice to high sealevel stands in the past.

Another source of potential instability has recently become the focus of much discussion: methane ice deposits on the continental slope. By some estimates these deposits are extensive. They are vulnerable to warming at the upper depth limit of their distribution. A large-scale release of methane (such as postulated for the more distant past; Dickens et al., 1995) could presumably generate unprecedented greenhouse warming (Haq, 1998). We do know that the climate system reacts to warming with an increase in the methane content of the atmosphere, as seen in the deglaciation periods, within the Antarctic ice (Petit et al., 1999). But the mechanisms are as yet unclear.

VI. CONCLUSION

Considerable doubt arises with regard to the reliability of forecasts, be they favorable, neutral, or unfavorable. As long as we cannot identify the mechanisms responsible for known past changes in carbon dioxide content of the atmosphere, or those responsible for changes in productivity, or for abrupt climate change, our understanding is demonstrably insufficient for forecasting. On the other hand, geologic experience is valuable in providing a range of possible scenarios for the future especially scenarios that include unexpected sudden response to gradual forcing.

Table 1. Oxygen isotope stratigraphy "Ojplus03" for the last 838,000 years

Standardized values for the deep-sea oxygen isotope record, in 2 kyr intervals, starting with zero. The standard was generated as follows: The isotope curve of Mix et al. (1995) for ODP Site 849 was used as a guide for long-term change. The isotope curve of Berger et al. (1996) (based on numerous cores from Ontong Java Plateau) was given twice the weight for short-term change (100-kyr scale). The isotope curve of Bassinot et al. (1994) (a core from the Indian Ocean) was given the same weight as ODP 849 for short-term change. The sequence of calculations was as follows: (1) recalculate the ODP 849 curve for average = 1 and std = 0.25; (2) calculate the average and std for a running window of 100 kyr for the ODP 849 curve; (3) for each position, reset the Indian-Ocean curve for the 100-kyr average and std of ODP 849; (4) for each position, reset the Ontong-Java stack for the 100-kyr average and std of ODP 849; (5) add the ODP 849 standardized curve (1), plus the Indian-Ocean curve (3), plus twice the Ontong-Java stack (4), and re-standardize for average = 1 and std = 0.25. From 700 kyr back, only the Ontong Java stack is used. Sea level is obtained by assuming a range for the glacial-postglacial contrast; here taken as 110 m (to take account of missing extremes).

**1.31, 1.36, 1.38, 1.41, 1.35, 1.13, 0.85, 0.64, 0.55, 0.53, 0.57, 0.60, 0.63, 0.67,
0.70, 0.70, 0.74, 0.79, 0.78, 0.75, 0.74, 0.74, 0.75, 0.76, 0.79, 0.81, 0.82, 0.82,
0.80, 0.79, 0.77, 0.74, 0.73, 0.75, 0.83, 0.93, 1.00, 1.02, 1.03, 1.06, 1.12, 1.14,
1.12, 1.12, 1.11, 1.08, 1.10, 1.16, 1.20, 1.24, 1.24, 1.23, 1.23, 1.21, 1.17, 1.19,
1.20, 1.31, 1.45, 1.51, 1.55, 1.56, 1.54, 1.40, 1.12, 0.85, 0.75, 0.62, 0.61, 0.56,
0.61, 0.64, 0.67, 0.69, 0.70, 0.70, 0.69, 0.67, 0.66, 0.69, 0.71, 0.81, 0.89, 0.92,
0.86, 0.89, 0.91, 0.84, 0.87, 0.85, 0.82, 0.86, 0.87, 0.98, 1.11, 1.19, 1.25, 1.27,
1.24, 1.21, 1.19, 1.17, 1.16, 1.17, 1.22, 1.29, 1.32, 1.28, 1.22, 1.13, 1.05, 1.01,
0.99, 0.97, 0.98, 0.97, 1.09, 1.19, 1.23, 1.29, 1.30, 1.21, 1.08, 0.94, 0.83, 0.85,
0.88, 0.88, 0.87, 0.86, 0.84, 0.83, 0.84, 0.82, 0.79, 0.78, 0.80, 0.84, 0.89, 0.95,
1.01, 1.08, 1.12, 1.14, 1.14, 1.12, 1.07, 1.01, 0.99, 0.99, 1.03, 1.06, 1.10, 1.15,
1.21, 1.26, 1.27, 1.24, 1.22, 1.23, 1.23, 1.31, 1.40, 1.46, 1.49, 1.49, 1.39, 1.22,
1.03, 0.86, 0.77, 0.71, 0.71, 0.74, 0.77, 0.80, 0.86, 0.83, 0.86, 0.82, 0.81, 0.84,
0.90, 0.96, 0.99, 1.04, 1.08, 1.08, 1.08, 1.10, 1.13, 1.17, 1.21, 1.23, 1.20, 1.17,
1.22, 1.33, 1.46, 1.56, 1.63, 1.64, 1.65, 1.67, 1.66, 1.61, 1.56, 1.48, 1.39, 1.31,
1.21, 1.06, 0.88, 0.69, 0.52, 0.45, 0.43, 0.46, 0.46, 0.53, 0.55, 0.59, 0.72, 0.77,
0.80, 0.84, 0.78, 0.75, 0.77, 0.75, 0.76, 0.79, 0.80, 0.83, 0.90, 0.96, 0.99, 1.03,
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0.98, 0.93, 0.91, 0.92, 0.86, 0.84, 0.83, 0.83, 0.88, 0.93, 0.97, 0.99, 0.99, 1.02,
1.05, 1.06, 1.07, 1.11, 1.23, 1.35, 1.39, 1.35, 1.29, 1.27, 1.23, 1.19, 1.10, 0.97,
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0.86, 0.86, 0.84, 0.85, 0.95, 1.09, 1.18, 1.23, 1.26, 1.25, 1.25, 1.21, 1.14, 1.15,
1.14, 1.16, 1.17, 1.17, 1.14, 1.08, 1.00, 0.93, 0.86, 0.80, 0.75, 0.69, 0.65, 0.62,
0.64, 0.68, 0.74, 0.79, 0.83, 0.85, 0.87, 0.86, 0.84, 0.81, 0.78, 0.76, 0.77, 0.82,
0.86, 0.89, 0.89, 0.87, 0.86, 0.87, 0.92, 0.98, 1.05, 1.11, 1.16, 1.21, 1.24, 1.27,
1.29, 1.29, 1.23, 1.11, 0.97, 0.84, 0.75, 0.70, 0.65, 0.60, 0.57, 0.57, 0.60, 0.64,
0.69, 0.73, 0.77, 0.83, 0.89, 0.96, 1.01, 1.04, 1.05, 1.05, 1.06, 1.10, 1.13, 1.15**

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