

GEOLOGIC RECORD OF SEA-LEVEL AND ITS RELEVANCE TO GLOBAL CHANGE FORECASTS

*An Earth System History Initiative
of the Global Change Research Program*

by

Richard G. Fairbanks and Gregory S. Mountain

with contribution from

James Wright

(all at Lamont-Doherty Geological Observatory)

ABSTRACT

Changes in sea level have produced world-wide changes in shallow-water sedimentary and carbonate deposits that are recorded throughout the geologic record. There were numerous mechanisms responsible for these sea-level changes, including ocean temperature changes, variable storage of water on continents, and changes in the volume of the ocean basins. The magnitudes, rates, and frequencies of sea-level change have been estimated from analysis of marine oxygen isotopic records, diagenetic patterns within atolls, and changes in depositional pattern and their seismic expression along continental margins and within interior basins. At present, the limited accuracy of these estimates cannot isolate the impact of the many mechanisms of eustatic change. Improvements are needed in collecting appropriate geologic and geophysical information, understanding sedimentary basin development, and integrating efforts among the many researchers investigating the history of sea-level change.

The largest component affecting the sea-level record of the last century is believed to be isostatic change, primarily due to continuing crustal rebound near formerly glaciated areas. The eustatic component of the global sea-level monitoring network is estimated using statistical and numerical methods. The numerical solution requires an accurate deglaciation history of the continental ice sheets, and the best available deglaciation reconstruction is based on the sea-level record from submerged coral reefs combined with meltwater plumes measured by the $\delta^{18}\text{O}$ of marine carbonates and dated by AMS ^{14}C techniques.

The record of glacio-eustatic sea level is perhaps the most fundamental monitor of large-scale climate change and the mechanism of sea level change is the most important in global warming scenarios. There is a long and detailed record of glacio-eustatic sea-level change monitored by the ^{18}O concentration in marine carbonates. Evidence from marine oxygen isotopic records and emergent coral reefs indicates that sea level during the last interglacial (~122,000 years BP) was approximately 6 ± 2 meters above the present level. There is debate whether melting of the Greenland or the Antarctic ice cap contributed this excess water to the

ocean. Although climate forcing during the last interglacial was unlike that predicted for the next century, the origin and rate of the extra 6 meters of sea level rise of the last interglacial transgression are of interest in assessing future meltwater sources and possible rates of discharge. It has been proposed by several investigators that the West Antarctic ice sheet, the only marine-based ice sheet today, may have surged during the last interglacial, thus producing the 6 meter higher sea level. Recent ice core analyses suggest that Greenland may have been largely deglaciated during the last interglacial.

Assessing the magnitude of the CO₂ amplifier factor is central to international climate research. The thermodynamic effects of 2X CO₂ are well known; however, the amplifier factor resulting from fast climate feedbacks ranges from 1 to 4. The natural changes in CO₂ vary from 180 to 270 ppm and offer a natural experiment to evaluate this CO₂ amplifier factor. Radiocarbon dating of the ice core CO₂ records can be compared with high resolution sea-level records to evaluate the likelihood of an amplifier factor of 1 versus 4.

INTRODUCTION

The importance of studying sea-level changes recorded in the geologic record has been the subject of many national and international panels charged with prioritizing Global Change research initiatives. Recent documents concerning this issue include the National Research Council's volume on Sea Level Change (1990), the Ocean Studies Board working group report on Continental Margins (NCR Report, 1989), the Ocean Drilling Program COSOD II Report (1987), the NSF-USSAC report on The Role of ODP Drilling in the Investigation of Global Sea Level (1990), and the Chapman Conference volume on Long-Term Sea Level Changes (1991). Each of these reports identified the fundamental and complexly interwoven role of sea-level change in the global Earth system. Furthermore, several discussed the geologic evidence of large and small sea-level oscillations at various frequencies, as well as abrupt shifts that indicate the ocean/climate system may have more than one stable mode of operation. While these abrupt shifts often are assumed to be glacio-eustatic in origin, the specific mechanism responsible for such non-linearities in the Earth system remains a mystery. Few other geologic processes have a more far-reaching impact on the Earth system, and to advance our understanding of eustatic change, research must be conducted on the many geologic time scales that it affects.

The objectives of the U.S. Global Change Research Program are to: (1) document the Earth system on a global scale; (2) understand the many types of processes (geologic, among others) that influence this system; and (3) predict future Earth system behavior. In the following discussion we review the causes of eustatic change, describe how a more detailed understanding can be developed, and outline research directions that will refine the accuracy of eustatic change forecasts.

BACKGROUND

The geologic record contains extensive evidence of sea-level fluctuations whose magnitudes and rates of change approach those of the late Pleistocene. The latter sea-level changes have been attributed to the growth and decay of polar ice caps. Ironically, it is generally agreed (although not unanimously) that this glacial mechanism was lacking for much of the Earth's history prior to 35 Ma, despite the stratigraphic evidence for an uninterrupted history of oscillatory sea-level changes.

Three general classes of processes are known to contribute to eustatic fluctuations. These are changes in the: (1) volume of sea water; (2) mass of sea water; and (3) volume of the ocean basins. The different rates and magnitudes of change for these processes provide an opportunity to separate their effects in the geologic record. In the following discussion, the magnitudes are expressed as meters of global sea-level effect. It should be recognized that these estimates of vertical sea-level change depend on the hypsometric curve that one adopts. This curve can be thought of as the "shape of the container", and as the elevation of the ocean surface is calculated

farther back in time, it becomes less certain that the sides of the "container" have maintained their present shape. Nonetheless, these following estimates are made using uniform hypsometry to provide the basis of comparison between the various mechanisms of eustatic change.

Changing the Volume of Sea Water

The density of ocean water is a function of its temperature. As it heats up it expands, spilling out of its "container" and flooding onto the continents. Warming the upper 500 m of the world's ocean by 10°C would result in a global rise of roughly 1 m; a similar warming of the deep ocean (a much larger reservoir) would raise sea level several times this amount. While the shallow heating could be achieved in several hundred years, it has been estimated that it would take several thousand years to heat the deep sea by the same amount. A warming of 10°C is a reasonable upper limit to the amount that the entire ocean has been warmed at various times during the last 120 Ma. Hence, this "steric" process has probably accounted for a total of less than 10 m of eustatic fall since the mid-Cretaceous.

Changing the Mass of Sea Water

Significant changes in the mass of the world's sea water can be achieved by changing the partition of water between the continents and the ocean stored as either fresh water (lakes and groundwater) or as continental ice. The total amount of liquid water found both on and within the continents represents roughly 1% of the world's total supply; if entirely released to the ocean, it would raise sea level by about 35 m (Woods, 1984). However, 99% of this freshwater is underground (Robin, 1987), and relatively little of it would drain into the ocean during a global lowstand. Glacial growth and decay is a far more significant process for rapidly changing the amount of water in the ocean. The potential for eustatic change by this process is measured in tens of meters on time scales of thousands of years. At present, 2% of the Earth's water is locked up in ice, and if melted would raise sea level by about 70 m. Alpine glaciers constitute about 1% of this by volume and the Greenland ice cap about 10%; the rest is contained in the Antarctic ice sheet.

Changing the Volume of Ocean Basins

Three processes control the volume of the ocean basin itself: (1) changing the mean elevation of the sea floor by volcanic buildup or thermal doming; (2) modifying continental area by compression or extension of continental lithosphere; or (3) changing the lengths and/or spreading rates of mid-ocean ridges.

Large volcanic eruptions on the ocean floor (*e.g.*, the Nauru Basin basalt flow) or thermal uplift of large areas of oceanic lithosphere (*e.g.*, the Darwin Rise) raise global sea level and spill water onto the continents. Both effects decay with time, and sea level will subside as a result of isostatic adjustment and cooling subsidence, respectively. Harrison (1990) calculated that thermal uplift and subsequent subsidence of the equatorial Pacific alone accounted for nearly 60 m of sea-level fall since the peak thermal effect at 70 Ma.

The collision of the Indian, Arabian and African plates with those of Asia and Europe has thickened the lithosphere across this long suture zone, thereby reducing the total area of the Earth's landmass and increasing the area of the ocean basins. Harrison (1990) estimated that this has contributed to roughly 20 m of sea-level fall over the last 45 my.

Ridge crest changes are by far the largest contributor to changes in ocean basin volume (Hays and Pitman, 1973). The crucial factor is the area of sea floor formed per unit time, and this is controlled by changing spreading rates, ridge crest lengths, or a combination of the two. By reconstructing sea floor over the last 150 my, Kominz (1984) estimated that these processes contributed to a peak sea-level high of about 180 m above present at 80 Ma. She provided error analysis for this estimate that derives from uncertainties in the magnetic reversal time scale,

inaccurate estimates of ridge lengths, necessarily crude estimates of subducted ridges for which only remanent triple junctions remain, and additional features on the sea floor that have been entirely subducted. These uncertainties result in a peak global sea level that ranges from 317 m above present to roughly the present-day value.

In summary, spreading ridge processes dominate volumetric changes in magnitude (contributing to many tens to a few hundred meters of eustatic change) on time scales of several tens of millions of years. While ocean floor elevation changes and lithospheric collision act on roughly equivalent time scales, reasonable estimates of their effects are an order of magnitude smaller in scale.

RESEARCH OBJECTIVES

Documenting and Understanding Long-Term Eustatic Changes

Changes in global sea level on million-year time scales or longer can be determined by three techniques:

- (1) measuring foraminiferal $\delta^{18}\text{O}$, which reflects ice-volume and sea water temperature fluctuations (Shackleton and Opdyke, 1973; Miller et al., 1987; Prentice and Matthews, 1988);
- (2) identifying subaerial exposure surfaces within carbonate platforms and atolls (Fairbanks and Matthews, 1978; Halley and Ludwig, 1987; Major and Matthews, 1983); or
- (3) gauging sea-level changes relative to the continents (Sloss, 1963; Hallam, 1984; Vail, 1987).

The oxygen isotopic record is largely restricted to the past 100 my. While it is unsurpassed as an indicator of changes in global ice volume and can provide reliable timing and magnitude information regarding these changes, the oxygen isotopic record contains no information about other (*e.g.*, tectonically induced) mechanisms for sea-level change. Carbonate platform and atoll records represent "dip-sticks" in the ocean, and potentially can provide the most accurate magnitudes of eustatic change, regardless of the mechanism. Their value depends on the ability to identify subaerial exposure surfaces and freshwater diagenesis in samples from the submerged foundations of atolls and guyots. Though valuable, eustatic histories of this type are not especially common in the geologic record. By contrast, sea-level fluctuations observed on continents and their submerged margins are far more common, and interpretable records span a long time interval, certainly back to the beginning of the Paleozoic, and with limited resolution into the Precambrian (Christie-Blick *et al.*, 1988; Bond *et al.*, 1988). While the timing of eustatic changes may be determined reliably with this technique, separating the effects of local sediment supply and tectonism from those of eustatic changes complicates the measurement of eustatic change magnitudes.

Deep-Sea Oxygen Isotopic Records

Marine $\delta^{18}\text{O}$ records are potentially one of the best resources for reconstructing changes in the cryosphere, and hence, the Earth's glacio-eustatic history. Ice volume/sea-level changes are recorded in the marine $\delta^{18}\text{O}$ record because ice accumulation in polar regions stores water which is depleted in ^{18}O ($\delta^{18}\text{O}_{\text{ice}} = -35$ to -40‰), thus leaving the oceans enriched in ^{18}O . Fairbanks and Matthews (1978) showed that the sea water $\delta^{18}\text{O}$ value (δ_w) changes by 0.11‰ for every 10 meters of sea-level change. During the most recent deglaciation, sea level rose by 120 m, resulting in a global δ_w decrease of 1.3‰ (Fairbanks, 1989). This relationship can be used to estimate the ice volume component of Cenozoic sea-level changes from monospecific foraminiferal (planktonic and benthic) $\delta^{18}\text{O}$ records.

Two factors other than the δ_w changes associated with ice volume changes affect the $\delta^{18}\text{O}$ value recorded by foraminifera: (1) temperature of calcification; and (2) the local $\delta^{18}\text{O}$ value of the water. Foraminiferal $\delta^{18}\text{O}$ values change with temperature because of thermodynamic fractionation of oxygen isotopes during the precipitation of calcite. In general, foraminiferal $\delta^{18}\text{O}$ values decrease by $\sim 0.22\text{‰}$ for every 1°C increase in temperature.

The $\delta^{18}\text{O}$ value incorporated by foraminifera is affected also by the local δ_w value. Evaporation and precipitation patterns produce differences in the distribution of δ_w and salinity in the ocean. As water is transported from low to high latitudes, oxygen isotopes are fractionated, producing a range of surface δ_w values greater than 2.0‰ . Open ocean δ_w increases by 0.6‰ for every 1‰ increase in salinity (Craig and Gordon, 1965; Fairbanks, 1982). The range of deep water δ_w values varies by only $\sim 0.4\text{‰}$ (Ostlund *et al.*, 1987). As a result, variability in the $\delta^{18}\text{O}$ of deep water is considered to be less important in sea level and paleo-temperature estimates from benthic foraminifera. However, regions near potential low latitude deep water sources should be avoided as representatives of the global δ_w value during intervals of suspected warm, saline deep water (WSDW) production (*e.g.*, the early to middle Eocene and early Miocene (Berggren and Hollister, 1977; Kennett and Stott, 1990; Woodruff and Savin, 1989). The δ_w value of WSDW may differ from the δ_w values of high latitude deep water masses, thereby creating a local δ_w gradient in deep waters. However, this potential problem can be easily avoided by monitoring the deep water δ_w changes with deep Pacific sites.

Deep water $\delta^{18}\text{O}$ composites for the Cenozoic were first developed by Savin *et al.* (1975) and Savin (1977). These studies considered only long-term changes because of low sample resolution. Recently, these data have been reassessed, new measurements have been added, and the resulting composite curves have been smoothed and stacked to remove the higher frequency changes (Miller *et al.*, 1987; Prentice and Matthews, 1988). The general character in each of these syntheses shows that Cenozoic deep water $\delta^{18}\text{O}$ values decreased from the early Paleocene (0‰ from 65 to 60 Ma) to an early Eocene minimum (-1‰ at 52 Ma) and then increased to a maximum during the Last Glacial Maximum (LGM; -4.5‰ at 18 Ka) (Figure 1).

The total change in Cenozoic deep water $\delta^{18}\text{O}$ values of 5.5‰ reflects both temperature and ice volume changes. The LGM represents maximum Cenozoic ice volume and accounts for only 2.0‰ (36%) of the 5.5‰ increase. Deep water temperature changes must have accounted for the remaining 2.5‰ , which is equivalent to a temperature cooling of 10 to 11°C .

The $\delta^{18}\text{O}$ increase from the early Eocene to the present was not gradual; rather, it was punctuated by several large and rapid, step-like increases. Two of the largest changes ($>1\text{‰}$) occurred at the Eocene/Oligocene boundary (36 Ma) and in the middle Miocene (15 Ma). These events have been interpreted to represent, respectively, the development of the psychrosphere (warm surface ocean, cold deep ocean) near the Eocene/Oligocene boundary (Kennett and Shackleton, 1976) and the initiation of continental glaciation on Antarctica in the middle Miocene (Shackleton and Kennett, 1975; Savin *et al.*, 1975, 1981; Woodruff *et al.*, 1981), *i.e.*, the full development of the "ice house" world.

Matthews and Poore (1980) challenged these interpretations and suggested that large ice sheets existed prior to the middle Miocene. Recent drilling in the Southern Ocean (ODP Legs 113, 114, 119, and 120) and on the Antarctic continental margins (Barrett *et al.*, 1987) confirmed the existence of intermittent continental ice sheets on Antarctica between the early Oligocene and early Miocene (Miller *et al.*, 1991). Nonetheless, the contribution of deep water cooling vs. Antarctic glaciation to $\delta^{18}\text{O}$ increases between Eocene and middle Miocene time remains uncertain. Furthermore, there is still disagreement as to whether the middle Miocene increase represented the development of a permanent ice sheet in eastern Antarctica (*c.f.*, Matthews and Poore, 1980; Kennett and Barker, 1990).

The limited resolution of existing $\delta^{18}\text{O}$ histories provides very little insight into the ice-volume component of Cenozoic sea-level change. Seismic stratigraphic interpretations on

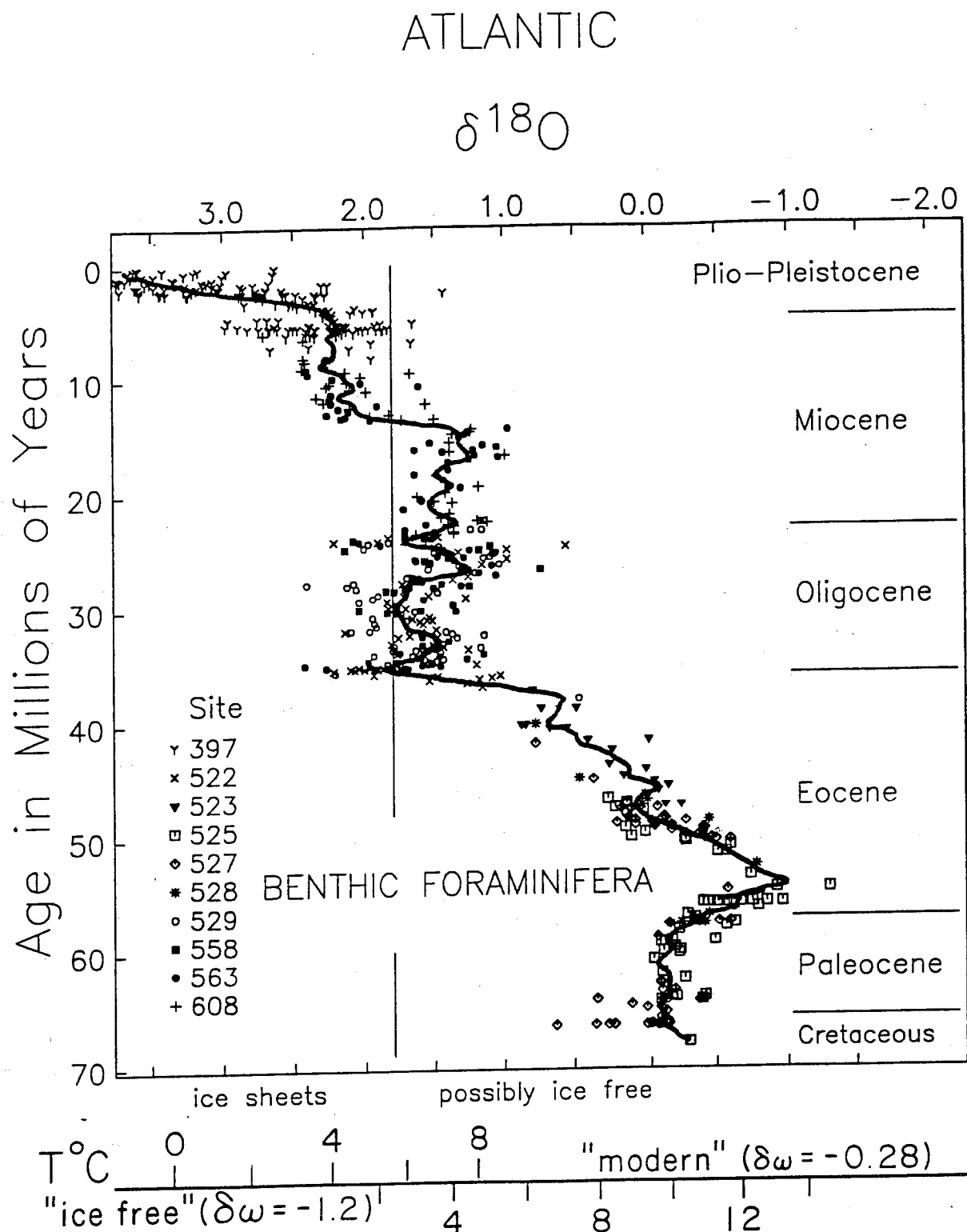


Figure 1: Composite of benthic foraminiferal (*Cibicidoides* spp.) $\delta^{18}\text{O}$ record for Atlantic DSDP sites (from Miller et al., 1987). The smoothed curve was obtained by linearly interpolating between data at 0.1 my intervals and smoothing with a 21-point Gaussian convolution filter which removed frequencies less than 1 my. The 1.8‰ line represents temperatures colder than 2°C under ice-free conditions which were interpreted to be inconsistent with ice-free polar regions. Therefore, intervals which recorded $\delta^{18}\text{O}$ values greater than 1.8‰ must represent times having some continental ice.

passive continental margins (*e.g.*, Haq *et al.*, 1987) indicate that there were at least 21 major sea-level cycles from the late Eocene to the late Pliocene. However, over the same interval of time, only four or five $\delta^{18}\text{O}$ increases can be recognized in the Savin (1977) compilation, and less than 10 $\delta^{18}\text{O}$ increases are distinguished in the Miller *et al.* (1987) synthesis.

Only recently have deep water $\delta^{18}\text{O}$ records been generated with sufficient resolution to identify the higher frequency (≤ 1.5 my) components comparable to the continental margin records. These studies identified several $\delta^{18}\text{O}$ increases (0.5 to 0.8‰) and interpreted them to reflect ice volume changes (Miller *et al.*, 1991; Wright and Miller, 1991; Wright *et al.*, in press). Each of these benthic foraminiferal $\delta^{18}\text{O}$ increases correlates well with a sea-level fall interpreted from sequence stratigraphy on passive continental margins (Haq *et al.*, 1987). Nonetheless, it is argued correctly by Matthews and Poore (1980) and Prentice and Matthews (1988) that the benthic foraminiferal $\delta^{18}\text{O}$ record is very susceptible to temperature changes.

On an annual basis, non-upwelling equatorial regions are the most thermally stable surface waters of the world's oceans (Ravelo *et al.*, 1990). These regions provide the best opportunity to obtain planktonic foraminiferal $\delta^{18}\text{O}$ time series which are less subject to temperature effects than other surface-water regions (Matthews and Poore, 1980; Prentice and Matthews, 1988). If sea surface temperatures (SST) in these regions have remained constant over time, then changes in planktonic foraminiferal $\delta^{18}\text{O}$ records from these waters should reflect ice volume changes. Under this premise, low latitude planktonic foraminiferal $\delta^{18}\text{O}$ records indicate that the world was ice-free prior to 35 to 40 Ma and that ice volume/sea level fluctuated between present day and 18 Ka levels since 35 Ma (Matthews and Poore, 1980; Prentice and Matthews, 1988).

Recent modeling studies have provided support for relatively constant western equatorial SST's (Ramanathan and Collins, 1991). These results suggest that large changes in the greenhouse gases produce feedbacks within the system which regulate the tropical SST's. Therefore, the thermal uniformity of tropical regions suggests that these can provide especially valuable $\delta^{18}\text{O}$ records of ice volume/sea-level changes. The only drawback with this approach is the lack of western equatorial records for many intervals in the Cenozoic.

The ability to extract an ice volume/sea-level record from benthic or planktonic foraminiferal $\delta^{18}\text{O}$ records alone is limited. However, one can apply a Pleistocene strategy which uses concomitant changes in benthic and low-latitude (non-upwelling) planktonic $\delta^{18}\text{O}$ records as indicators of ice-volume/sea-level change (Shackleton and Opdyke, 1973). Covariance between benthic and low-latitude planktonic $\delta^{18}\text{O}$ signals provides the strongest evidence for a global change in δ_w values resulting from ice-volume variations. The use of benthic-planktonic covariance does not require that either deep water or surface water temperatures remain constant. This approach attempts to isolate the ice volume signal common to both planktonic and benthic foraminiferal $\delta^{18}\text{O}$ records and eliminate the temperature signals. However, areas subject to large seasonal temperature variations (*e.g.*, upwelling regions) should be avoided.

These Pleistocene strategies have been applied successfully to the pre-Pleistocene $\delta^{18}\text{O}$ record (Miller and Fairbanks, 1985; Miller *et al.*, 1987, 1991; Wright *et al.*, in press). Low latitude planktonic $\delta^{18}\text{O}$ records from the western equatorial Atlantic, Pacific, and Indian Oceans have confirmed that many of the short-term $\delta^{18}\text{O}$ variations found in the benthic records match those in the surface water records. Still, it is noted that many more records are needed to document the ice volume/sea-level component of the marine $\delta^{18}\text{O}$ records.

The last step in this process is to correlate the $\delta^{18}\text{O}$ record of sea-level change with sedimentary deposits on and around Antarctica. While the relationship of a till or diamictite to ice-sheet size is uncertain, and the correlation of these glacial deposits to the deep sea record is tenuous at best, this linkage must be established. Once a correlation between the continental and the deep sea record is confirmed, the latter can provide a chronology of ice sheet size/sea-level change. Furthermore, planktonic-benthic covariance can provide limits on the size of the ice sheet.

Carbonate Atolls and Guyots

Atolls are the reef-fringed tops of volcanoes that have subsided due to cooling of the underlying lithosphere; carbonate guyots are simply drowned atolls that for one reason or another could not keep up with subsidence. The carbonate secreting organisms that constitute these structures usually grow within a few meters of sea level, and, therefore, are extremely sensitive to sea-level variations. Their exposure/flooding history can provide reliable and direct measurement of eustatic change. The "tectonic" (*i.e.*, thermal) subsidence history of an atoll is relatively simple and predictable, even though it is a combination of tectonics and eustasy. Studying the sediment accumulation, paleobathymetry, and subsidence history of an atoll minimizes the major uncertainties inherent in interpreting the stratigraphic records of passive continental margins. Furthermore, atolls and guyots are widely distributed across the Pacific Ocean where there are few continental margin records suitable for sea-level studies.

Eustatic lowstands are recorded in atolls as subaerial exposure surfaces and/or zones of freshwater diagenesis. These surfaces can create seismic reflectors and are recognized in core samples by their unique petrologic and stable isotopic signatures (Schlanger, 1963; Major and Matthews, 1983; Ludwig and Halley, 1988). Recent advances in shallow-water biostratigraphy (largely based on ostracods and benthic foraminifers), magnetostratigraphy, and strontium-isotope stratigraphy enable reasonable dating of atoll carbonates, especially in the Neogene (where precision is approximately ± 0.5 my). In cases where the topographic relief of the atoll exceeds the height of a sea-level fall, the magnitude of eustatic change can be measured as well. The accuracy of these measurements is greatest where subsidence rates are rapid relative to the rate of the eustatic fall, and where exposure surfaces are clearly separated by diagenetically unaltered sediments. When either of these criteria are not met, multiple events can overprint and obscure the sea-level record.

Sequence Stratigraphy on Continental Margins

The record of sea-level change on continental margins can be determined by: (1) mapping transgressions/regressions of the shoreline or of changes in water depth inferred from facies successions (Bond, 1978; Hancock and Kauffman, 1979; Harrison, 1989); or (2) analyzing regional unconformities with physical and seismic stratigraphy (Vail *et al.*, 1977; Vail, 1987; van Wagoner *et al.*, 1987) or with chronostratigraphy (Aubry, 1985; Miller *et al.*, 1985). The chief advantage to the sequence stratigraphic approach is that the formation of stratal discontinuities requires the lowering of depositional base level, and is therefore less sensitive to variations in sediment supply than is the position of the shoreline (Christie-Blick *et al.*, 1990). This approach provides a great deal of information about the timing of relative sea-level changes, but less certain information about their magnitudes.

Determining the influence of sea-level fluctuations on passive margin stratigraphy poses a paradox. Shallow-water (neritic) sections are most sensitive to sea-level changes, but are often difficult to correlate to a standard chronostratigraphy because of rare zonal marker plankton and discontinuous deposition. Conversely, deep-sea (bathyal to abyssal) sections often are better constrained and more complete, but the link between deep-sea deposition and sea-level fluctuations is complex and incompletely understood (Tucholke, 1981; Farre *et al.*, 1983; Farre, 1985; Mountain, 1987). A solution to this problem is to develop chronostratigraphic standards using deep sea sections (*e.g.*, Berggren *et al.*, 1985), monitor sea-level changes in shallow, passive continental margin sequences (Aubry, 1985; Olsson and Wise, 1987), and use facies-independent means (*e.g.*, seismic profiles, Sr-isotopes, or magnetostratigraphy) for correlating the two regions (Miller and Kent, 1987). Therefore, determining relative sea-level and eventually recognizing eustatic changes requires: (1) integrating studies from nearshore to deep-sea environments; (2) optimizing chronostratigraphic control in these diverse depositional environments by combining bio-, isotopic, and seismic stratigraphy; and (3) evaluating the records of several margins of different tectonic histories to establish clear distinctions between discontinuities caused by local tectonism and those of truly eustatic origin.

The primary product of chrono- and seismic stratigraphic studies along passive continental margins is improved estimates of the timing of eustatic changes, independent of assumptions about global cyclicity in sea-level change (*e.g.*, Haq *et al.*, 1987). Any global nature to regional unconformities/hiatuses can be established by comparing them with records from several passive margins and epicontinental seas. Ultimately, estimates can be made of the ages of eustatic falls (or more exactly, the ages of maximum rates of eustatic fall; Pitman, 1978; Vail, 1987; Christie-Blick *et al.*, 1990). This improved timing will allow comparison with the $\delta^{18}\text{O}$ record to determine the relationship between erosional events and inferred glacio-eustatic lowerings known to have occurred at least since the Oligocene. The ability to establish improved chronologies for glacial periods should allow researchers to apply these methods to assumed pre-glacial periods.

Besides the need to improve knowledge of the ages of eustatic changes, there is great need to constrain the magnitudes of these changes as well. The mechanisms for eustatic change are well understood, but with our currently limited ability to extract magnitude estimates from the geologic record, we are unable to improve our understanding of how these mechanisms have acted in the past. Determining ages of past eustatic changes will be a first step towards narrowing the field of responsible mechanisms; determining magnitudes will be the second.

Unfortunately, measuring magnitudes along passive continental margins is a more difficult task than measuring timing, and one that cannot be satisfactorily undertaken until we are more confident about which changes in the record are truly eustatic. The best approach to this problem on passive margins is to conduct subsidence analysis along transects (Figure 2) that are

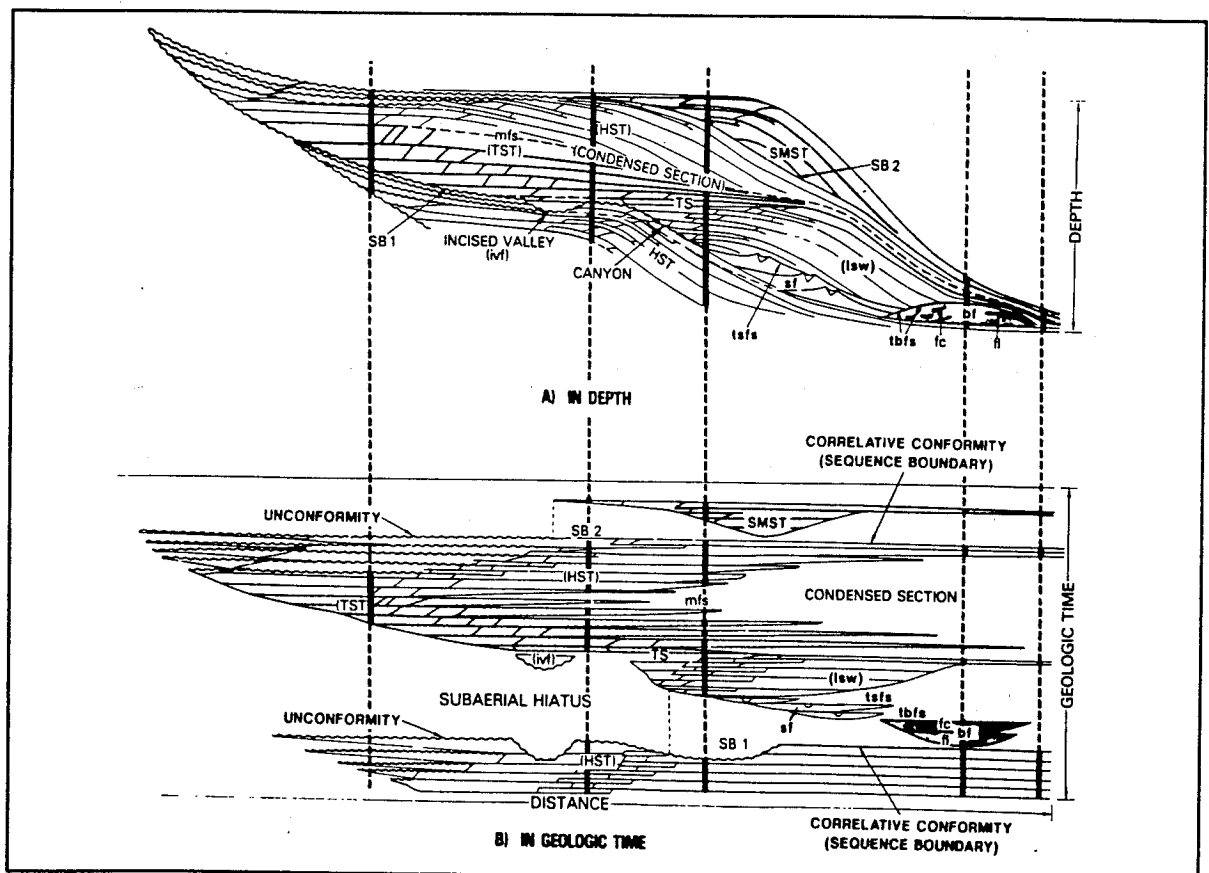


Figure 2: Idealized depositional sequence in depth (upper) and geologic time (lower) (from Vail, 1987). To evaluate properly the age and facies associated with sequence-bounding unconformities, conformable correlatives, and stratal relationships within sequences, it is necessary to core and log a transect of at least five holes as suggested in this diagram. Only after this has been completed on several margins of different rifting ages will it be possible to assess the global nature of sea-level fluctuations.

keyed to sequence and seismic stratigraphy as well as to appropriately positioned boreholes and available outcrops. Subsidence analysis provides a means for quantifying the relations between tectonic subsidence, eustasy, sediment accumulation, compaction and loading, and changes in paleobathymetry. Transects are required to: (1) date sequence boundaries precisely at their correlative conformities; (2) account for the influence of lithospheric flexure in controlling patterns of sedimentation; and (3) minimize errors in paleobathymetry. Many amplitudes of sea-level change are likely to be of the same order as water depth on a typical continental shelf, and will therefore be difficult to measure with precision.

The best estimates of paleobathymetry are obtained from sediments deposited at or near sea level. Early attempts to measure amplitudes of sea-level change directly from the geometry of seismic sequences (Vail *et al.*, 1977) did not account for the effects of subsidence or for the compaction and loading of the sediments, and this led to estimates of eustatic change that were markedly too large (Greenlee *et al.*, 1988; Christie-Blick *et al.*, 1990). Greenlee *et al.* (1988) proposed a modified technique in which the amplitude of sea-level fall is estimated from the downward shift in coastal onlap below the depositional coastal break for type 1 unconformities (*i.e.*, when sea level fell faster than the background rate of tectonic subsidence at the previous depositional coastal break). The depositional coastal break is identified in a seismic sequence as the point of steepest seaward dip (the "rollover point") of a clinoform, and is thought to occur in about ten meters of water (Vail, 1987; van Wagoner, 1987). The method of Greenlee *et al.* (1988) assumes that the lowest position of coastal onlap in the overlying sequence can also be identified correctly. An appropriately sited borehole is required to demonstrate this and to distinguish coastal (*i.e.*, non-marine) onlap from marine onlap. In general, the method of Greenlee *et al.* (1988) places a lower limit on the amplitude of eustatic fall because sea level is already falling at the time of the downward shift in onlap, and renewed coastal onlap above a sequence boundary begins before sea level starts to rise. Nonetheless, this method represents a considerable improvement in the quest for a rigorous quantitative approach to the investigation of eustasy.

No one technique for measuring long-term sea-level change is completely adequate. Integration of the deep-sea oxygen isotopic, atoll, and passive margin approaches is absolutely crucial. Furthermore, these complementary efforts will be most effective if they concentrate on three contrasting intervals: (1) when glacio-eustatic changes were clearly operating during the Oligocene to middle Miocene; (2) when geologic evidence suggests warm global temperatures prevented significant glacial-ice buildup in the Middle to Late Cretaceous; and (3) during the time for which debate continues over the existence of ice sheets in the middle Paleogene.

Future Sea-Level Change

We described in the preceding section three of the many lines of research that are needed to fully understand the history of sea level. In this section, we outline several aspects of sea-level research that are directly relevant to forecasting Global Change and rates of future sea-level variations: (1) conducting high resolution studies of the last deglaciation; (2) improving understanding of the role of CO₂ in the global heat budget; and (3) determining the contributions of specific large ice sheets to the last interglacial.

Modern Sea-Level Change — Separating Isostasy From Eustasy.

Tide gauge records of relative sea-level (RSL) change measured at coastal and island stations have been compiled and used to compute "global" sea-level change over the past century. Although there is a wide range of estimates due to numerous sources of local variability, nearly all estimates indicate rising sea level over the past 100 years. Most estimates range from 1.0 to 3.0 mm/year and the connection to possible global warming over the past century (Hansen *et al.*, 1981) has been suggested, but not demonstrated. Many authors have discussed in detail the causes of local variability, which include: (1) instrumentation problems; (2) sediment loading of the lithosphere at tide gauges near deltas of large rivers; (3) physical changes, natural or man-

made, affecting such things as river runoff, current changes, temperature, or salinity changes; and (4) regional climate effects such as changes in the onshore component of wind stress or changes in local atmospheric pressure.

To circumvent many of these problems, Barnett (1983 and 1990), following the lead of Fairbridge and Krebs (1962), selected a subset of 147 stations from the global tide gauge network which passed a list of quality control criteria (Figure 3). Barnett (1978, 1983) used empirical orthogonal function (EOF) analysis to statistically extract a sea-level signal which was coherent among the "key" tide gauge stations. Barnett (1990) analyzed the sea level results for two different time periods, 1903 to 1969 and 1930 to 1975, and found that the longer time period is fit by a linear trend of 1.51 ± 0.15 mm/yr. The results for the more recent interval are more variable, but also indicate a rising trend of 1.79 ± 0.22 mm/yr.

Debate continues over what fraction of this RSL is due to steric effects of thermal expansion of the oceans versus eustatic effects stemming from addition of water to the world's oceans. The magnitude of the steric effect is generally believed to be less than 5 cm over the past century (Gornitz *et al.*, 1982; Roemmich and Wunsch, 1984), implying that the remainder is due to melting of alpine (Meier, 1984) and continental glaciers.

Recently, Peltier and Tushingham (1991) concluded that sea level is rising at a faster rate than previously estimated based on substantial isostatic corrections to tide gauge records. They used a high resolution model of the glacial isostatic adjustment to remove this signal from more than 500 tide gauge records from around the world. The isostatically corrected results showed much less variance and revealed a globally coherent sea-level rise of 2.4 ± 0.9 mm/yr since 1900 (Figure 4), although their isostatic filter is extremely sensitive to small changes in the numerical analysis procedures.

The early post-glacial sea-level record is fundamental to the isostatic adjustment model, and more specifically, to the decay histories of the various ice sheets. The most recent RSL predictions made by Peltier and Tushingham (1991) used the ICE-3G reconstruction of Tushingham and Peltier (1990) (Figure 5). The ICE-3G reconstruction incorporates considerable glacial geomorphologic data as well as ^{14}C -dated sea-level records that are mostly from continental margins. Accuracy of the sea-level curve can be improved by extending the ^{14}C calibration beyond 9,500 years BP. As a result of secular variations in the atmospheric ^{14}C production, the ^{14}C time scale requires a correction of approximately 800 years at 9,500 years BP. In addition, there is evidence from lake sediments and fossil trees that there are intervals as long as 400 calendar years

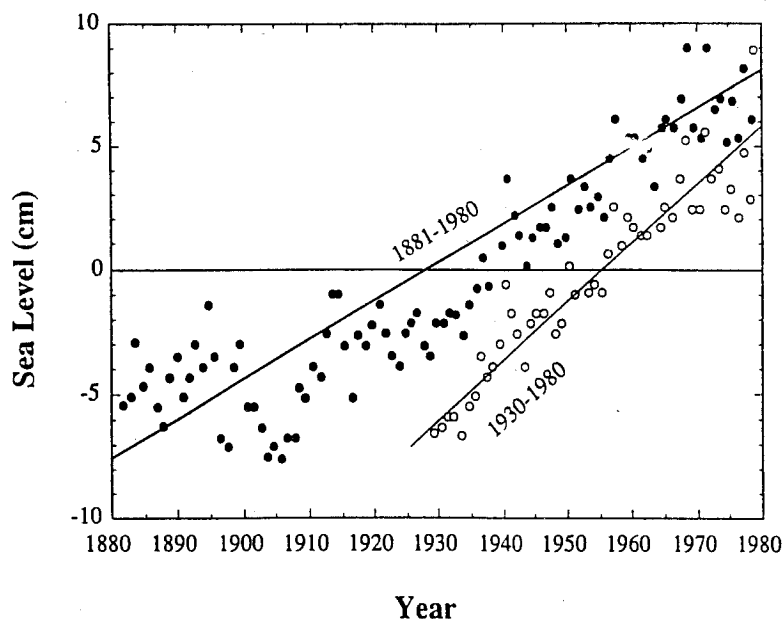


Figure 3: Estimates of "overall average" sea-level rise based on EOF analysis of regional tide gauge records for two different time periods. (from Barnett, 1984).

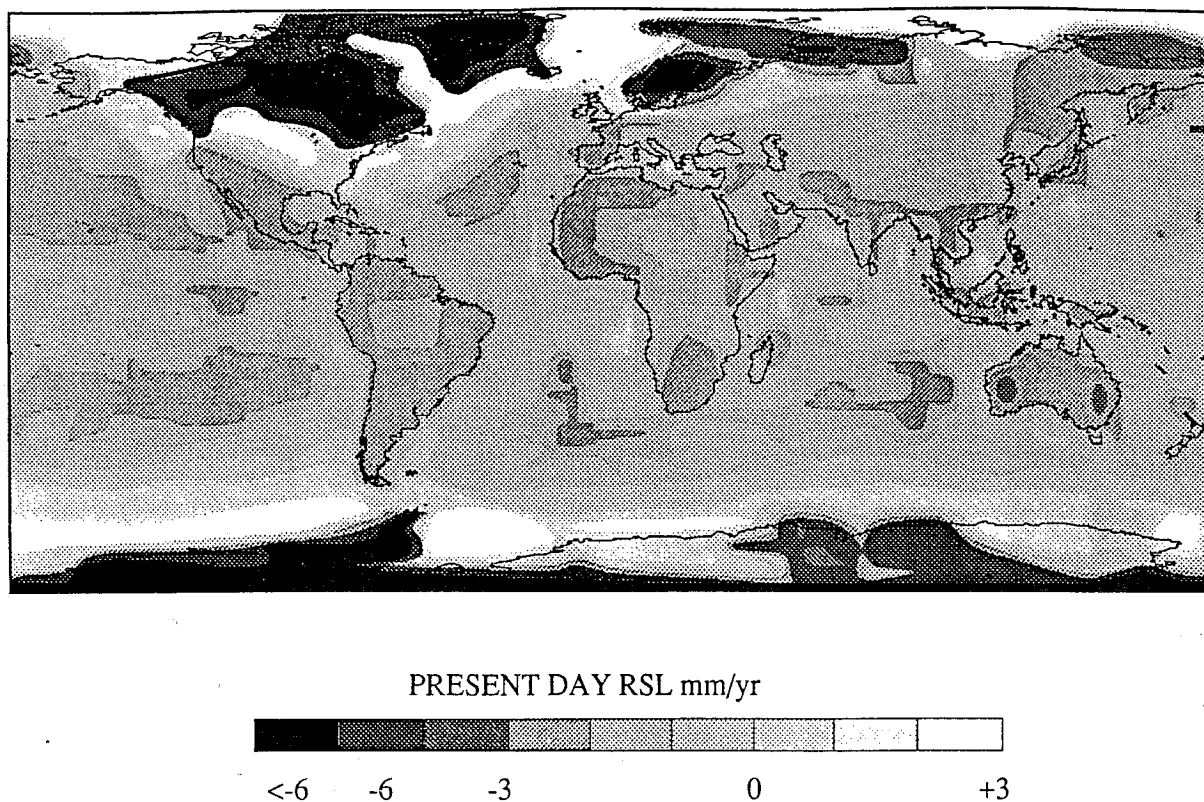


Figure 4: Modeling estimate of the modern rate of sea-level change solely due to glacial isostatic adjustment. (adapted from Peltier and Tushingham, 1991).

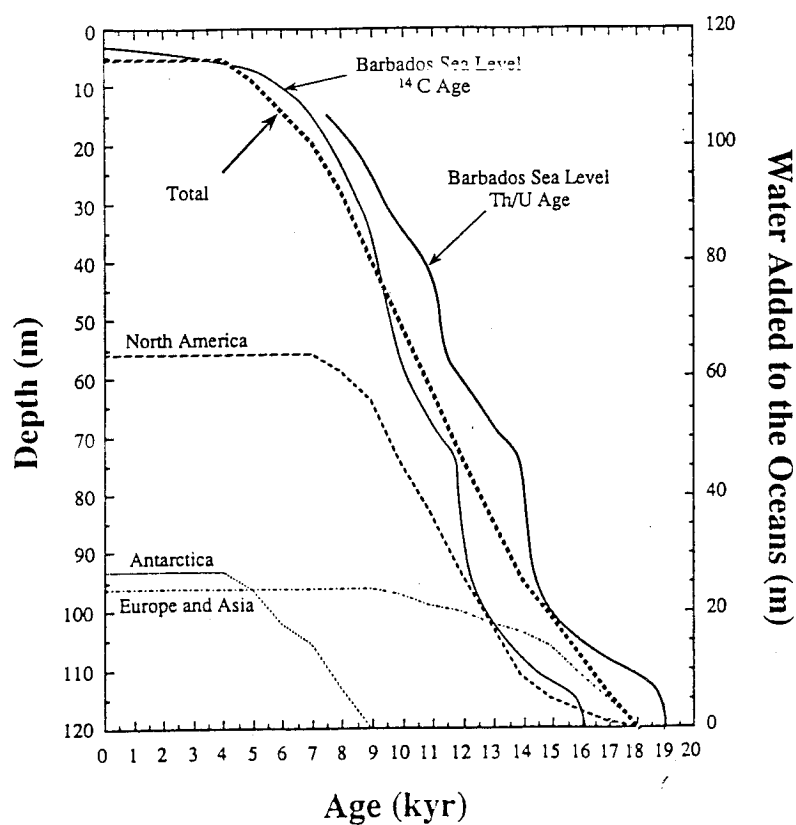


Figure 5: Estimated melting history and water volume of continental ice sheets in North America, Antarctica and Europe/Asia used in the modeling estimate of modern sea-level change due to glacial isostatic adjustment (estimates from ICE-3G model of Tushingham and Peltier, 1990). The ICE-3G total sea-level change is compared to the Barbados ^{14}C and $^{230}\text{Th}/^{234}\text{U}$ sea-level curves, which are uncorrected for any isostatic distortion.

when the ^{14}C values remained constant. ^{14}C -dendrochronology calibration becomes increasingly difficult to extend as the calibration is pushed beyond 10,000 years BP, and other techniques such as varved sediments or coral ^{14}C - $^{230}\text{Th}/^{234}\text{U}$ and ^{14}C - Pa/U calibrations must be relied on (Bard *et al.*, 1990). For example, the more accurate $^{230}\text{Th}/^{234}\text{U}$ -dated sea-level curve for the post-glacial indicates that the ^{14}C -dated sea-level curve was compressed by as much as 3,000 years (Bard *et al.*, 1990; Fairbanks, 1990) (Figure 5). The net result of this apparent age adjustment on the ICE-3G reconstruction is that the Laurentide Ice Sheet has a longer melting time, thereby indicating less ice decay in the Antarctic. Improvements in the resolution of the post-glacial sea-level record and, in particular, documentation of the melting histories of individual ice sheets, can contribute to an improved isostatic model useful for filtering modern tide gauge records.

Recent progress has been made in correlating terrestrial and marine records of deglaciation. The technique relies on identifying meltwater pulses dated with AMS ^{14}C in planktonic foraminifera near the great discharge areas of the Laurentide and Fennoscandian ice sheets, such as the Gulf of Mexico, Labrador Sea, Barents Sea, and North Sea (Jones and Keigwin, 1988; Broecker *et al.*, 1990; Lehman *et al.*, 1991) (Figure 6). Combined with geomorphologic and gravity anomaly data, measurements of the arrival time and duration of massive meltwater

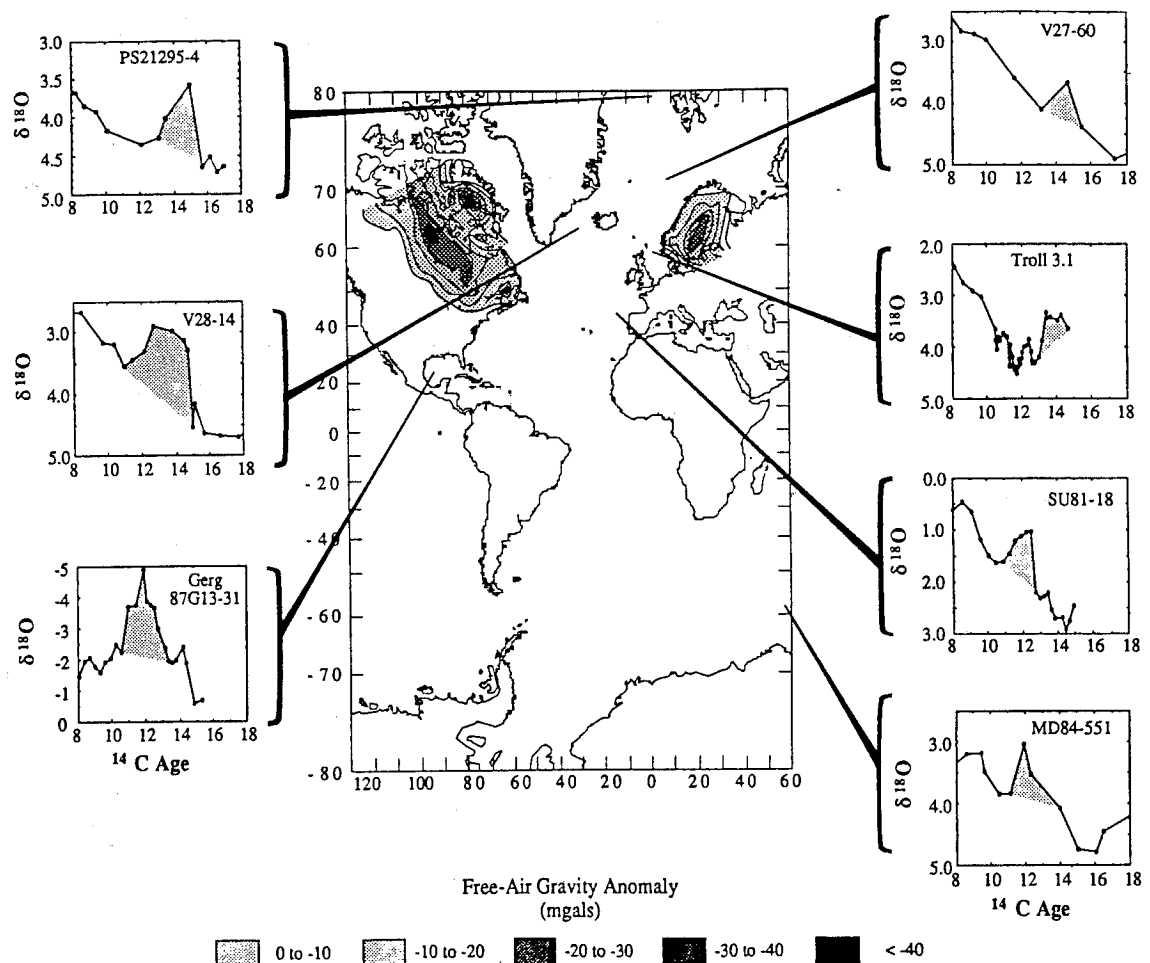


Figure 6: Oxygen isotopic records of planktonic foraminifera showing pulsed and sometimes sequential melting of different continental ice sheets plotted on a map of observed free-air gravity anomalies over Canada and Fennoscandia. The AMS ^{14}C dated discharge records are from Broecker *et al.* (1990)-Gulf of Mexico (Laurentide ice sheet); Jones and Keigwin (1988)-Fram Strait (Kara ice sheet); Bard *et al.* (1987)-Northeast Atlantic (Fennoscandian ice sheet); Labracherie *et al.* (1989)-Antarctica.

plumes in the Atlantic and Southern Oceans are crucial for dating the decay histories of individual ice sheets (Figure 6). Maps of these apparent meltwater plumes indicate that deglaciation swept from north to south with the earliest meltwater plume found in the Fram Strait at 15,000 ^{14}C yrs (Jones and Keigwin, 1988). However, modeling studies by Manabe and Stouffer (1988) indicate that changes in the rate of thermohaline circulation may produce large salinity anomalies in the circum-North Atlantic, which could produce $\delta^{18}\text{O}$ oscillations that are indistinguishable from meltwater pulses (Fairbanks *et al.*, 1991).

The H_2^{18}O tracer of large meltwater pulses ($15,000 \text{ km}^3/\text{yr}$) such as MWP-1A at 12,000 ^{14}C years BP can globally propagate by ocean mixing as well as by evaporation from both continental sources and the ocean surface mixed layer (Fairbanks, 1989; 1990). This can lead to confusion regarding the source and timing of the H_2^{18}O anomalies measured in deep sea cores because they may not be of local origin (Linsley and Thunell, 1990). Modeling efforts are necessary to approximate the atmospheric and oceanic transport of the $\delta^{18}\text{O}$ tracer associated with these global meltwater pulses (Duplessy *et al.*, 1991).

Post-Glacial Sea-Level Record — A Check on the CO_2 Amplifier Factor

A high priority research initiative of the U.S. global change program is the evaluation of the CO_2 fast-feedback factors, such as clouds, water vapor, albedo, and sea ice, collectively known as the amplifier factor (f) (Hansen *et al.*, 1984). A doubling of CO_2 without any feedback factors would increase the surface and tropospheric temperature by 1.2°C due to radiative forcing. The results of general circulation model (GCM) simulations indicate that a doubling of CO_2 leads to a global temperature increase ranging from 2 to 5°C , or an amplifier factor of 2 to 4. According to these simulations, the distribution of temperature change is not uniform with a doubling of CO_2 , but is enhanced in polar regions, especially in the northern hemisphere. Other simulations that use simple one-dimensional climate models show no net feedback (Ramanathan, 1988). The dramatic difference between these modeling results is primarily the result of cloud feedback parameterization.

Given this large discrepancy in the amplifier factor, Lorius *et al.* (1990) turned to the paleoclimate record in ice cores for an independent determination of the amplifier factor (f). They used D/H ratios to reconstruct the post-glacial temperature and studied the encapsulated atmospheric CO_2 record from the Vostok ice core in central Antarctica to conclude that (f) is ~ 3 . These ice core results are widely cited as confirmation of a positive and rather large CO_2 amplifier factor. Furthermore, Lorius *et al.* (1990) conclude that greenhouse gases contributed to $\sim 2^\circ\text{C}$ global warming between glacial and interglacial periods. It also is widely acknowledged that there are many untested assumptions concerning the accuracy of the temperature estimates based on D/H as well as the exact timing of the CO_2 changes. For example, in cold, low accumulation rate regions such as Vostok, the CO_2 signal is "locked in" 3,000 to 6,000 years after the snow has been deposited (Barnola *et al.*, 1991). Nevertheless, this is an excellent example of how paleoclimate reconstructions can and should be used to place independent constraints on modeling forecasts.

The post-glacial sea-level record can also be used as an independent check on the magnitude of the CO_2 amplifier factor. Snow accumulates more rapidly at coastal stations on Antarctica and on the Greenland ice cap than it does at Vostok, thus preserving a much higher-resolution CO_2 record. For example, the deglacial interval between 13,000 and 9,000 years BP at Vostok has only one CO_2 measurement, whereas Byrd (Antarctica) and Dye 3 (Greenland) have more than 10 CO_2 measurements in this interval and the gas closure time is on the order of only 100 years later than the surrounding ice. These records show that CO_2 concentration increased by $\sim 42\%$ during the deglacial, from 190 ppm to 270 ppm. The exact timing of the CO_2 change is poorly constrained for several reasons. The original Dye 3 record indicates the CO_2 shift occurred in 650 yrs (Stauffer *et al.*, 1985). However, new measurements in this interval were unable to reproduce the original values, and high variability in the new measurements casts doubt on the entire Dye 3 CO_2 data set (Staffelbach *et al.*, 1991). Possible absorption of CO_2 into seasonal

melt layers during annual warm intervals or during storage, apparently undetected during the core analysis and sampling, is cited as an explanation for the variable CO₂ values in Dye 3. The Byrd CO₂ record indicates that the CO₂ increased slowly from 190 ppm to 270 ppm between 20,000 and 10,000 years BP, although the age control for this ice core is poor (Paterson and Hammer, 1987; Staffelbach *et al.*, 1991).

The post-glacial sea-level record from Barbados is sufficiently detailed that accurate rates of sea-level change can be computed (Fairbanks, 1989; 1990). Supplemented by sea-level data from New Guinea (Chappell and Polach, 1991) and by isotopic records of meltwater discharge (Figure 7), these Barbados data show that the melting of the northern hemisphere ice sheets was not due to a slow change in accumulation versus ablation rates. Instead, this deglaciation was marked by two short episodes of dramatic disintegration. Accurate dating of these events with ²³⁰Th/²³⁴U and ¹⁴C methods permits a detailed comparison of ice response to the leading hypotheses of glacial-interglacial climate change, including atmospheric CO₂, thermohaline circulation (Broecker and Denton, 1990), and local insolation changes.

The insolation curve, the ice core CO₂ results, and the Barbados sea-level curve are all based on chronologies scaled in calendar years. The Atlantic deep and intermediate water $\delta^{13}\text{C}$ records dated by AMS ¹⁴C (Figure 7) can be converted to calendar years using the ¹⁴C calibration of Bard *et al.* (1990). Charles and Fairbanks (1991) show that the massive deglaciation between 14,200 and 13,800 ²³⁰Th/²³⁴U years BP, known as melt-water-pulse 1A, was exactly coincident with the start up of the North Atlantic Deep Water flux to the Southern Ocean (Figure 8). According to Charles and Fairbanks (1991), the initiation of the thermohaline circulation was apparently 1,500 years later than the estimates of NADW initiation based in the North Atlantic cores (Jansen and Veum, 1990; Boyle and Keigwin, 1987). This discrepancy may be a result of a strong mixing zone in the Atlantic Ocean basins between Antarctic Bottom Water and North Atlantic Deep Water. Cores within the Atlantic may record shifts in the location of this mixing zone that are not associated with net changes in the NADW flux.

Vulnerability of Greenland vs. the West Antarctic Ice Sheets

The scientific community is divided over the relative importance of the Greenland Ice Cap versus the West Antarctic Ice Sheet in global warming scenarios. The Sea Rise Group (1990; 1991) has focused on the instability of the West Antarctic Ice Sheet because it is the only remaining marine-based ice sheet which survived the deglaciation. Other groups have focused their research on monitoring the mass balance of the Greenland Ice Cap using satellite mapping techniques. Much of the justification for these scenarios comes from studies that suggest that the West Antarctic Ice Sheet surged during the last interglacial, causing a rapid 6 meter rise of sea level approximately 125,000 years BP (Mercer, 1968; Hollin, 1990). Based on ice core evidence, others have concluded that the Greenland Ice Cap melted during the last interglacial, causing the +6 meters relative sea level (Koerner, 1989) compared to modern sea level.

The development of the ²³⁰Th/²³⁴U dating technique via mass spectrometry (Edwards *et al.*, 1987a, 1987b; Edwards, 1988) has opened up new opportunities in sea-level research of the last interglacial. Previous alpha-counting dating techniques could resolve the last interglacial age to a precision of $\pm 15,000$ 2 σ (Harmon *et al.*, 1979). The new mass spectrometer dating technique allows precision of $\pm 1\%$ for 125,000 year old corals. For the first time, rates of sea-level change can be measured during the last interglacial cycle, allowing tests of the West Antarctic surge hypothesis, for example.

The Canadian ice caps were clearly in a state of disintegration near the time of the summer insolation maximum during the last deglaciation, but they have since recovered (Koerner and Fisher, 1990). If evidence for the destruction of the Greenland Ice Cap during the last interglacial is supported by new ice core and marine core data, then intensive studies of the rates of sea-level rise during the last 6 meters of the transgression would provide one measure of Greenland ablation rates. Obviously, no single climatic cycle in the past was exactly like any other climate

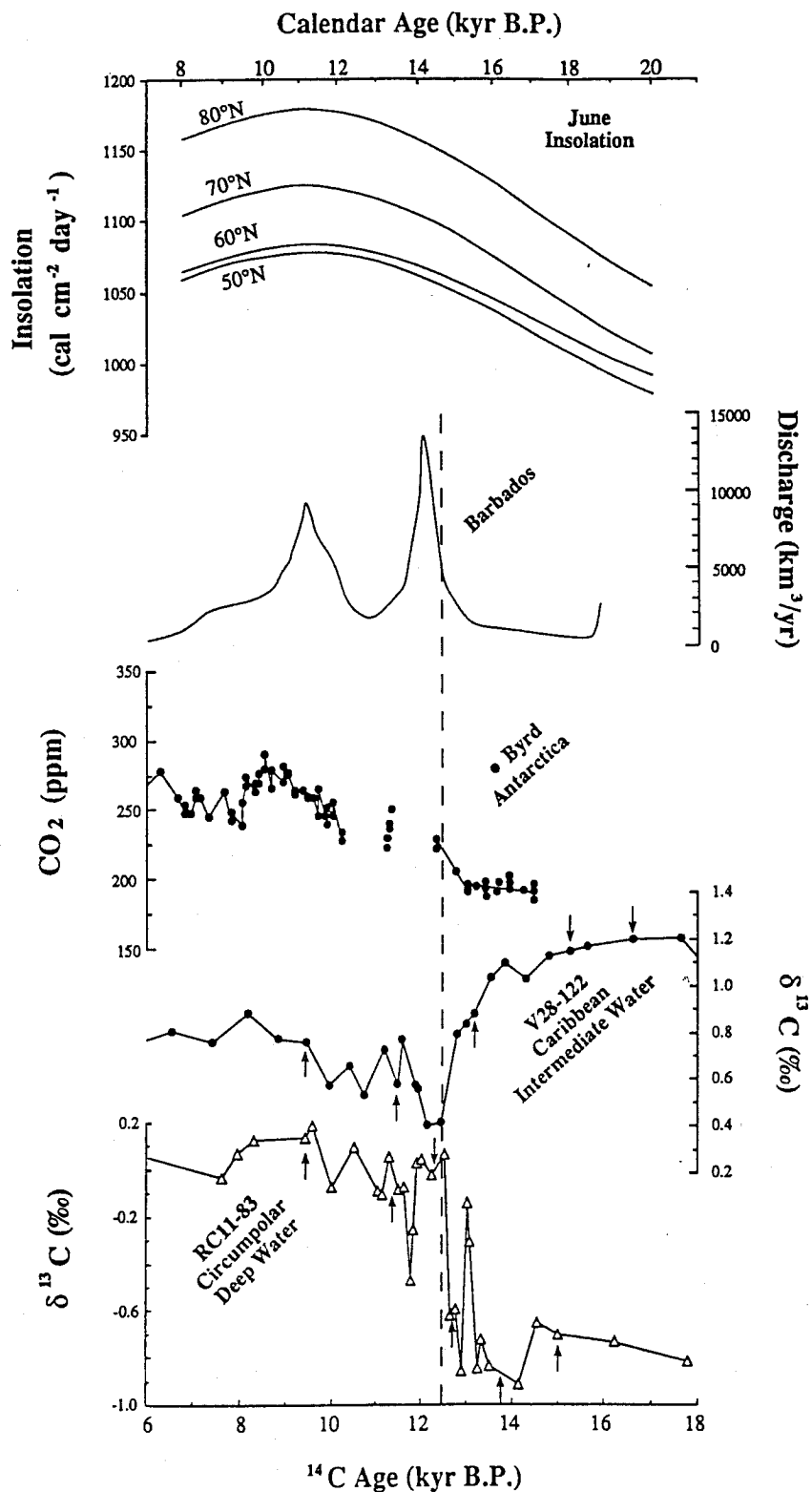


Figure 7: Record of atmospheric CO_2 from Byrd (Antarctica) polar ice core (Neftel et al., 1988) compared to global meltwater discharge curves (Fairbanks, 1991) reported in calendar or $^{230}\text{Th}/^{234}\text{U}$ years. Comparison of the Southern Ocean record of NADW flux (Core RC11-83, benthic foraminiferal $\delta^{13}\text{C}$ record) with a mid-depth $\delta^{13}\text{C}$ time series of Atlantic intermediate water chemistry (Core V28-122 benthic foraminiferal $\delta^{13}\text{C}$ record) (Charles and Fairbanks, 1991) adjusted to calendar years using the ^{14}C calibration of Bard et al. (1990) showing the coincidence of NADW production and deglaciation.

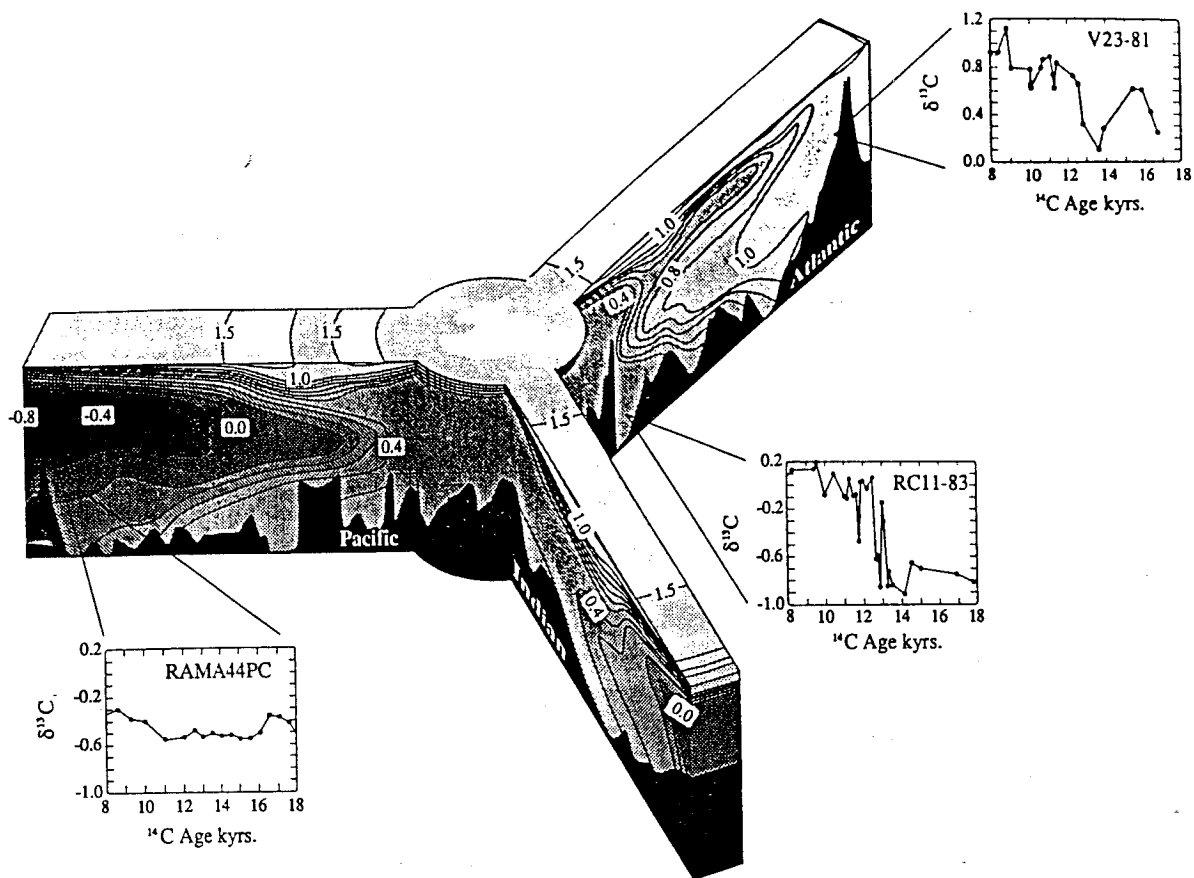


Figure 8: GEOSFCS $\delta^{13}\text{C}$ data from the different ocean basins (Kroopnick, 1985) with post glacial times series of $\delta^{13}\text{C}$ dated by AMS ^{14}C for the North Atlantic (Jansen and Veum, 1990), Southern Ocean (Charles and Fairbanks, 1991), and Pacific Ocean (Keigwin, 1987).

cycle, nor were the natural forces driving the melting 125,000 years BP the same as the anthropogenic influences expected over the next century. Nevertheless, the last interglacial is currently our best sea-level analogue. We can look also to the marine and terrestrial records to see if changes in the strength of thermohaline circulation, for example, were involved in the possible destruction of the Greenland Ice Cap during the last interglacial.

CONCLUSIONS

The geologic record contains widespread evidence of sea-level changes whose magnitudes and rates appear to match those of known Pleistocene glacio-eustatic cycles. Many of these changes occurred during periods of Earth history for which there is no evidence of extensive ice sheets. The possible existence of eustatic change mechanisms unrelated to ice volume changes suggests a complicated linkage among Earth system processes that may limit the accuracy of future sea-level change forecasts. However, before alternative mechanisms of sea-level change can be addressed meaningfully, there is a need for better estimates of the magnitudes and rates of sea-level changes during the pre-glacial and glacial past. It is especially important to determine reasonable upper limits to the rate at which sea level can rise. The constraints provided by better magnitude estimates could preclude several of the postulated and competing causes of sea-level change and help focus on the most likely possibilities.

Isostatic adjustment due to the melting of the continental ice sheets is a major component of the regional sea-level records measured during the past century. Filtering this "isostatic noise" from the global network of sea-level monitoring sites is a complex problem that is necessary in

order to assess global trends in ocean volume. This can be done statistically or numerically, and the numerical techniques require an accurate record of the melting histories and volumes for each ice sheet. AMS ^{14}C dating of meltwater plumes, measured by $\delta^{18}\text{O}$ of planktonic foraminifera, combined with a global meltwater discharge curve, is allowing researchers to piece together the sequential melting history of the great ice sheets. Rapid progress is anticipated in this area, partly due to the growing number of AMS ^{14}C dating facilities.

A doubling of atmospheric CO_2 leads to a global temperature increase of 1.2°C due to radiative forcing alone. GCM's indicate that fast feedback factors, such as water vapor content and clouds, will increase the thermal effects of CO_2 doubling by a factor of 3 to 4. However, simple one-dimensional energy balance models estimate the double CO_2 amplifier factor to be as low as 1 or even negative (Ramanathan, 1988). This amplifier factor is undoubtedly the most important parameter in the climate change forecasts, and the post glacial sea-level record may offer a qualitative test capable of resolving one extreme from the other.

Identifying whether Greenland or the West Antarctic ice sheet was responsible for higher sea level during the last interglacial is relevant to assessing the debate over the immediate vulnerability of these two ice masses under various scenarios of global warming. New dating methods, including the exciting accomplishments in mass spectrometry dating of carbonates using the uranium series, allows rates of sea-level change to be measured with high precision for times prior to the last interglacial. Accomplishments in land-sea correlations by measuring meltwater plumes with the oxygen isotope tracer can be extended back to the last interglacial cycle. New coring efforts will be needed because the existing piston core collections generally do not have material from the last interglacial cycle in high accumulation regions (*e.g.*, $> 25 \text{ cm/kyr}$). Changes in sea level have produced world-wide changes in shallow-water sedimentary and carbonate deposits that are recorded throughout the geologic record. There were numerous mechanisms responsible for these sea-level changes, including ocean temperature changes, variable storage of water on continents, and changes in the volume of the ocean basins. Combined with other lines of evidence, the magnitudes, rates, and frequencies of sea-level change have been estimated from analysis of seismic profiles and diagenetic patterns on atolls with various degrees of confidence.

REFERENCES

- Aubry, M.-P., 1985. Northwestern European Paleogene magnetostratigraphy, biostratigraphy, and paleoceanography. *Geology*, 13:198-202.
- Bard, E., Hamelin, B., Fairbanks, R.G., and Zindler, A., 1990. Calibration of the ^{14}C timescale over the past 30,000 years using mass spectrometric U-Th ages from Barbados corals. *Nature*, 345:405-410.
- Barnett, T.P., 1978. Estimating variability of surface air temperature in the Northern Hemisphere. *Mon. Weather Rev.*, 106:1353-1367.
- Barnett, T.P., 1983. Possible changes in global sea level and their causes. *Climate Change*, 5:15-38.
- Barnett, T.P., 1984. The estimation of "global" sea level change: A problem of uniqueness. *J. Geophys. Res.*, 89:7980-7988.
- Barnett, T.P., 1990. Recent changes in sea level: a summary. *Studies in Geophysics*. National Academy Press, Washington, D.C., 37-51.

- Barrett, P.J., Elston, D.P., Harwood, D.M., McKelvey, B.C., and Webb, P.N., 1987. Mid-Cenozoic record of glaciation and sea level change on the margin of the Victoria Land Basin, Antarctica. *Geology*, 15:634-637.
- Berger, A., 1978. Long-term variations of caloric insolation resulting from the Earth's orbital elements. *Quater. Res.*, 35:215-240.
- Berggren, W.A., and Hollister, C.D., 1977. Plate tectonics and paleocirculation - Commotion in the ocean. *Tectonophysics*, 38:11-48.
- Berggren, W.A., Kent, D.V., Flynn, J.J., and van Couvering, J.A., 1985. Cenozoic geochronology, *Geol. Soc. Am. Bull.*, 96:1407-1418.
- Bond, G.C., 1978. Speculations on real sea-level changes and vertical motions on continents at selected times in the Cretaceous and Tertiary Periods. *Geology*, 6:247-250.
- Bond, G.C., Komazin, M.A., and Grotzinger, J.P., 1988. Cambro-Ordovician eustasy: evidence from geophysical modeling of subsidence in Cordilleran and Appalachian passive margins. in Kleinspehn, K.L. and Paola, C. (eds.). *New Perspective in Basin Analysis*. Springer-Verlag, NY, 129-160.
- Charles, C.D. and Fairbanks, R.G., submitted 1991. North Atlantic deep water and its climate effects over the last deglaciation: evidence from Southern Ocean isotope records. *Nature*.
- Christie-Blick, N., Grotzinger, J.P., and von der Borch, C.C., 1988. Sequence stratigraphy in Proterozoic successions. *Geology*, 16:100-104.
- Christie-Blick, N., Mountain, G.S., and Miller, K.G., 1990. Stratigraphic and seismic stratigraphic record of sea-level change. in *Sea-level Change*. National Research Council Studies in Geophysics, 116-140.
- CLIMAP Project Members, *Geol. Soc. Am. Map and Chart series*, MC 36 (Geol. Soc. Am., Boulder, 1981).
- Craig, H. and Gordon, L.I., 1965. Deuterium and oxygen 18 variations in the ocean and the marine atmosphere. *Stable Isotopes in Oceanographic Studies and Paleotemperatures*, Consiglio Nazionale Delle Ricerch, Pisa, Italy, 277-373.
- Denton, G.H. and Hughes, T.J., 1981. "The Last Great Ice Sheets." Wiley-Interscience, NY.
- Duplessy, J.-C., Bard, E., Arnold, M., Shackleton, N.J., Duprat, J., and Labeyrie, L., 1991. How fast did the ocean-atmosphere system run during the last deglaciation? *Earth and Planet. Sci. Lett.*, 103:27-40.
- Fairbanks, R.G., 1989. A 17,000-year glacio-eustatic sea level record: influence of glacial melting rates on the Younger Dryas event and deep-ocean circulation. *Nature*, 342:637-642.
- Fairbanks, R.G., 1990. The age and origin of the "Younger Dryas climate event" in Greenland ice cores. *Paleoceanography*, 5:937-948.
- Fairbanks, R.G. and Matthews, R.K., 1978. The marine oxygen isotope record in Pleistocene coral, Barbados, West Indies. *Quat. Res.*, 10:181-196.
- Fairbanks, R.G., Charles, C.D., and Wright, J.W., 1991, in press. Origin of global meltwater pulses, in Kra, R. and Long, A., (eds.). *Four Decades of Radiocarbon Studies: An Interdisciplinary Perspective*. Elsevier.
- Fairbridge, R. and Krebs, Jr., O., 1962. Sea level and the southern oscillation. *Geophys. J. Roy. Astron. Soc.*, 6:532-545.
- Farre, J., 1985. The importance of mass wasting processes on the continental slope. Ph.D. thesis, 227pp. Columbia University, New York.

- Farre, J.A., McGregor, B.A., Ryan, W.B.F., and Robb, J.M., 1983. Breaching the shelfbreak: Passage from youthful to mature phase in submarine canyon evolution. in Stanley, D.J. and Moore, G.T. (eds.). *The Shelfbreak: A Critical Interface on Continental Margins*. SEPM Spec. Publ. No. 33:25-39.
- Gornitz, V., Lebedeff, L., and Hansen, J., 1982. Global sea level trend in the past century. *Science*, 213:1611-1614.
- Greenlee, S.M., Schroeder, F.W., and Vail, P.R., 1988. Seismic stratigraphic and geohistory analysis of Tertiary strata from the continental shelf off New Jersey—Calculation of eustatic fluctuations from stratigraphic data. in Sheridan, R. and Grow, J. (eds.). *Atlantic Continental Margin: U.S. Geol. Soc. Am., DNAG ser.*, 437-444.
- Gutenberg, B., 1941. Changes in sea level, postglacial uplift, and mobility of the Earth's interior. *Geol. Soc. Am. Bull.*, 52:721-772.
- Hallam, A., 1984. Pre-Quaternary sea-level changes. *Ann. Rev. Earth Planet. Sci.*, 12:205-243.
- Halley, R.B. and Ludwig, K.R., 1987. Disconformities and Sr-isotope stratigraphy reveal a Neogene sea level history from Enewetak Atoll, Marshall Islands, Central Pacific. *Geol. Soc. Am. Abstr. with Prog.*, 691.
- Hancock, J.M. and Kauffman, E.G., 1979. The great transgressions of the Late Cretaceous. *Jour. Geol. Soc. London*, 136:175-186.
- Hansen, J. and Lebedeff, S., 1988. Global surface air temperatures: update through 1987. *Geophys. Res. Lett.*, 15:323-326.
- Hansen, J. et. al., 1984. *Geophys. Monogr.*, 29:130-163.
- Hansen, J., Johnson, D., Lacis, A., Lebedeff, S., Lee, P., Rind, D., and Russell, G., 1981. Climate impact of increasing atmospheric carbon dioxide. *Science*, 213:957-966.
- Haq, B.U., Hardenbol, J., and Vail, P.R., 1987. Chronology of fluctuating sea levels since the Triassic (250 million years ago to present). *Science*, 235:1156-1167.
- Harmon, R.S., Ku, T.L., Matthews, R.K., and Smart, P.L., 1979. Limits of U-series analysis: Phase I: Results of the uranium-series intercomparison project. *Geology*, 7:405-409.
- Harrison, C.G.A., 1990. Long-term eustasy and epeirogeny in continents. in *Sea-Level Change*. National Academy Press, 141-158.
- Hay, W.W. and Leslie, M.A. 1990. Could possible changes in global ground water reservoir cause eustatic sea-level fluctuations? in *Sea-Level Change*. National Academy Press, 161-170.
- Hays, J.D. and Pitman III, W.C., 1973. Lithospheric plate motion, sea level changes and climatic and ecological consequences. *Nature*, 246:18-22.
- Jansen, E. and Veum, T., 1990. Two-step deglaciation: timing and impact on North Atlantic Deep Water circulation. *Nature*, 343:612-616.
- Jones, G. and Keigwin, L.D., 1988. Evidence from Fram strait (78°N) for early deglaciation. *Nature*, 336:56-59.
- Keigwin, L.D., 1987. North Pacific deep water formation during the latest glaciation. *Nature*, 330:362-364.
- Kennett, J.P. and Shackleton, N.J., 1976. Oxygen isotopic evidence for the development of the psychrosphere 38 Myr. ago. *Nature*, 260:513-515.
- Kennett, J.P. and Stott, L.D., 1990. Proteus and Proto-Oceanus: Paleogene Oceans as revealed from Antarctic stable isotopic results; ODP Leg, 113. *Proc. Ocean Drill. Prog., Sci. Rpt.*, 113:865-880.

- Koerner, R.M., 1989. Ice core evidence for extensive melting of the Greenland ice sheet in the last interglacial. *Science*, 244:964-968.
- Koerner, R.M. and Fisher, D.A., 1990. A record of Holocene summer climate from a Canadian high-Arctic ice core. *Nature*, 343:630-631.
- Kominz, M., 1984. Oceanic ridge volumes and sea-level change- an error analysis. in Schlee, J.S. (ed.). *Inter-regional Unconformities and Hydrocarbon Accumulation*. Am. Assoc. Pet. Geol. Mem. 36:109-127.
- Kroopnick, P., 1985. The distribution of carbon-13 in the world oceans. *Deep Sea Res.*, 32:57-84.
- Labracherie, M., Labeyrie, L.D., Duprat, J., Pichon, J., Bard, E., Arnold, M., and Duplessy, J.-C., 1989. *Paleoceanography*, 4:629-638.
- Linsley, B.L. and Thunell, R.C., 1990. The record of deglaciation in the Sulu Sea: evidence for the Younger Dryas Event in the tropical western Pacific. *Paleoceanography*, 5(6):1025-1039.
- Lorius, C., Jouzel, J., Raynaud, D., Hansen, J., and Le Treut, H., 1990. The ice-core record: climate sensitivity and future greenhouse warming. *Nature*, 347:139-145.
- Ludwig, K.R. and Halley, R.B., 1988. Strontium-isotope stratigraphy of Enewetak Atoll. *Geology*, 16:173-177.
- Major, R.P. and Matthews, R.K., 1983. Isotopic composition of bank margin carbonates on Midway Atoll: Amplitude constraint on post-early Miocene eustasy. *Geology*, 11:335-338.
- Matthews, R.K. and Poore, R.Z., 1980. Tertiary $\delta^{18}\text{O}$ record and glacio-eustatic sea-level fluctuations. *Geology*, 8:501-504.
- Meier, M.F., 1984. Contribution of small glaciers to global sea level. *Science*, 226:1418-1421.
- Mercer, J.H., 1968. Antarctic ice and Sangamon sea level. General Assembly of Bern (September-October 1967) Snow and Ice. IAHS Publ. 79:217-225.
- Miller, K.G., Mountain, G.S., and Tucholke, B.E., 1985. Oligocene glacio-eustasy and erosion on the margins of the North Atlantic. *Geology*, 13:10-13.
- Miller, K.G. and Kent, D.V., 1987. Testing Cenozoic eustatic changes: The critical role of stratigraphic resolution. in Ross, C. and Haman, D. (eds.). *Timing and depositional history of eustatic sequences: Constraints on seismic stratigraphy*. Cushman Foundation Forum. Res., Special Publication No. 24:51-56.
- Miller, K.G., Fairbanks, R.G., and Mountain, G.S., 1987. Tertiary oxygen isotope synthesis, sea-level history, and continental margin erosion. *Paleoceanography*, 2:1-19.
- Miller, K.G., Wright, J.D., and Fairbanks, R.G., 1991. Unlocking the Ice House: Oligocene-Miocene oxygen isotopes, eustasy, and margin erosion. *J. Geophys. Res.*, 96:6829-6848.
- Mountain, G.S., 1987. Cenozoic margin construction and destruction offshore New Jersey. in Ross, C. and Haman, D. (eds.). *Timing and depositional history of eustatic sequences: Constraints on seismic stratigraphy*. Cushman Foundation Forum. Res., Special Publication No. 24:57-83.
- Neftel, A., Oeschger, H., Staffelbach, T., and Stauffer, B., 1988. The CO_2 record in the Byrd ice core 50,000-5,000 years BP. *Nature*, 331:609-611.
- Oeschger, H., Beer, J., Siegenthaler, U., Stauffer, B., Dansgaard, W., and Langway, C.C., 1984. in Hansen, J.E. and Takahashi, T. (eds.). *Climate Processes and Climate Sensitivity*. AGU Geophysical Monograph Series, 29:299-306.

- Oeschger, H., Neftel, A., Staffelbach, T., and Stauffer, B., 1988. The dilemma of the rapid variations in CO₂ in Greenland ice cores. *Annals of Glaciology*, 10:215-216.
- Olsson, R.K. and Wise, S.W., 1987. Upper Paleocene to middle Eocene depositional sequences and hiatuses in the New Jersey Atlantic Margin. in Ross, C. and Haman, D. (eds.). *Timing and depositional history of eustatic sequences: Constraints on seismic stratigraphy*. Cushman Foundation Forum. Res., Special Publication No. 24:99-112.
- Ostlund, H.S., Craig, H., Broecker, W.S., and Spencer, D., 1987. *GEOSECS Atlantic, Pacific, and Indian Expedition, 7, Shorebased data and graphics*. National Science Foundation, Wash. D.C.
- Peltier, W.R. and Tushingham, A.M., 1991. Influence of glacial isostatic adjustment on tide gauge measurements of secular sea level change. *J. Geophys. Res.*, 96:6779-6796.
- Pitman, W.C., 1978. Relationship between eustasy and stratigraphic sequences of passive margins. *Geol. Soc. Am. Bull.*, 89:1389-1403.
- Prentice, M.L. and Matthews, R.K., 1988. Cenozoic ice-volume history: Development of a composite oxygen isotope record. *Geology*. 16:963.
- Ramanathan, V., 1983. *Science* 24:293-299.
- Ramanathan, V. and Collins, W., 1991. Thermodynamic regulation of ocean warming by cirrus clouds deduced from observations of the 1987. *Nature*, 351:27-32.
- Ravelo, A.C., Fairbanks, R.G., and Philander, S.G.H., 1990. Reconstructing tropical Atlantic hydrography using planktonic foraminifera and an ocean model. *Paleoceanography*, 5:409-431.
- Robin, G. de Q., 1987. Changing the sea level, projecting the rise in sea level caused by warming the atmosphere in Bolin, B., Doos, B.R., Jager, J., and Warrick, R.A. (eds.). *The Greenhouse Effect, Climate Change and Ecosystems*. John Wiley & Sons, 29:323-359.
- Roemmich, D. and Wunsch, C., 1984. Apparent changes in the climate state of the deep North Atlantic Ocean. *Nature*, 307:447-450.
- Savin, S.M., 1977. The history of the earth's surface temperature during the last 100 million years. *Annu. Rev. Earth Planet. Sci.*, 5:319-355.
- Savin, S.M., Douglas, R.G., and Stehli, F.G., 1975. Tertiary marine paleotemperatures. *Geol. Soc. Am. Bull.*, 86:1499-1510.
- Shackleton, N.J. and Kennett, J.P., 1975. Paleotemperature history of the Cenozoic and initiation of Antarctic glaciation: Oxygen and carbon isotopic analyses in DSDP Sites 277, 279, and 281. *Initial Rep. Deep Sea Drill. Proj.*, 29:743-755.
- Shackleton, N.J. and Opdyke, N.D., 1973. Oxygen isotope and paleomagnetic stratigraphy of Equatorial Pacific core V28-238: Oxygen isotope temperatures and ice volumes on a 10⁵ year and 10⁶ year scale. *Quat. Res.*, 3:39-55.
- Sloss, L.L., 1963. Sequences in the Cratonic Interior of North America. *Geol. Soc. Am. Bull.*, 74:93-114.
- Schlanger, S.O., 1963. Subsurface geology of Eniwetok Atoll. U.S. Geol. Survey Prof. Paper 260-BB:991-1066.
- Staffelbach, T., Stauffer, B., Sigg, A., and Oeschger, H., 1991. CO₂ measurements from polar ice cores: more data from different sites. *Tellus*, 43B:91-96.
- Tucholke, B.E., 1981. Geological significance of seismic reflectors in the deep western North Atlantic basin. in Warme, J.E., Douglas, R.G., and Winterer, E.L. (eds.). *The Deep Sea Drilling Project: A Decade of Progress*. Soc. Econ. Paleontol. Mineral, Spec. Publ. No. 32:23-37.

- Tushingham, A.M. and Peltier, W.R., 1990. ICE-3G: A new global model of late Pleistocene deglaciation based upon geophysical predictions of post-glacial relative sea level change. *J. Geophys. Res.*
- Vail, P.R., Mitchum, R.M., Todd, R.G., Widmier, J.M., Thompson III, S., Sangree, J.B., Bubb, J.N., and Hatlelid, W.G., 1977. Seismic stratigraphy and global changes of sea level, Seismic Stratigraphy-Applications to Hydrocarbon Exploration. in Payton, C.E. (ed.). *Seismic Stratigraphy — Applications to Hydrocarbon Exploration. Mem. Am. Assoc. Petrol. Geol.* 26:49-205.
- Van Wagoner, J.C., Mitchum, R.M., Posamentier, H.W., and Vail, P.R., 1987. Key definitions of sequence stratigraphy. in Bally, A.W. (ed.). *Atlas of Seismic Stratigraphy. Am. Assoc. Petrol. Geol. Studies in Geology*, 27.
- Wright, J.D. and K.G. Miller, 1991, in press. Miocene stable isotope stratigraphy, Site 747, Kerguelen Plateau. *Proc. of the ODP*, vol. 120, Part B.
- Wright, J.D., Miller, K.G., and Fairbanks, R.G. in press. Early and middle Miocene stable isotopes: Implications for deepwater circulation, climate and tectonics. *Paleoceanography*.
- Woodruff, F. and Savin, S.M., 1989. Miocene deepwater oceanography. *Paleoceanography*, 4:87-140.
- Woods, J.D., 1984. The upper ocean and air-sea interaction in global climate. in Houghton, J.T. (ed.). *The Global Climate*. Cambridge University Press.