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

An Earth System History Initiative of the Global Change Research Program

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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 $\delta^{18}O$ 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 CO_2 amplifier factor is central to international climate research. The thermodynamic effects of 2X CO_2 are well known; however, the amplifier factor resulting from fast climate feedbacks ranges from 1 to 4. The natural changes in CO_2 vary from 180 to 270 ppm and offer a natural experiment to evaluate this CO_2 amplifier factor. Radiocarbon dating of the ice core CO_2 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 $\delta^{18}O$, 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 $\delta^{18}O$ 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 $\delta^{18}O$ record because ice accumulation in polar regions stores water which is depleted in ^{18}O ($\delta^{18}O$ice = -35 to -40‰), thus leaving the oceans enriched in ^{18}O. Fairbanks and Matthews (1978) showed that the sea water $\delta^{18}O$ 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) $\delta^{18}O$ records.

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

The $\delta^{18}O$ 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 ~0.4‰ (Ostlund *et al.*, 1987). As a result, variability in the $\delta^{18}O$ 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 $\delta^{18}O$ 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 $\delta^{18}O$ 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 $\delta^{18}O$ 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 $\delta^{18}O$ 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‰) 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 $\delta^{18}O$ 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 $\delta^{18}O$ histories provides very little insight into the ice-volume component of Cenozoic sea-level change. Seismic stratigraphic interpretations on

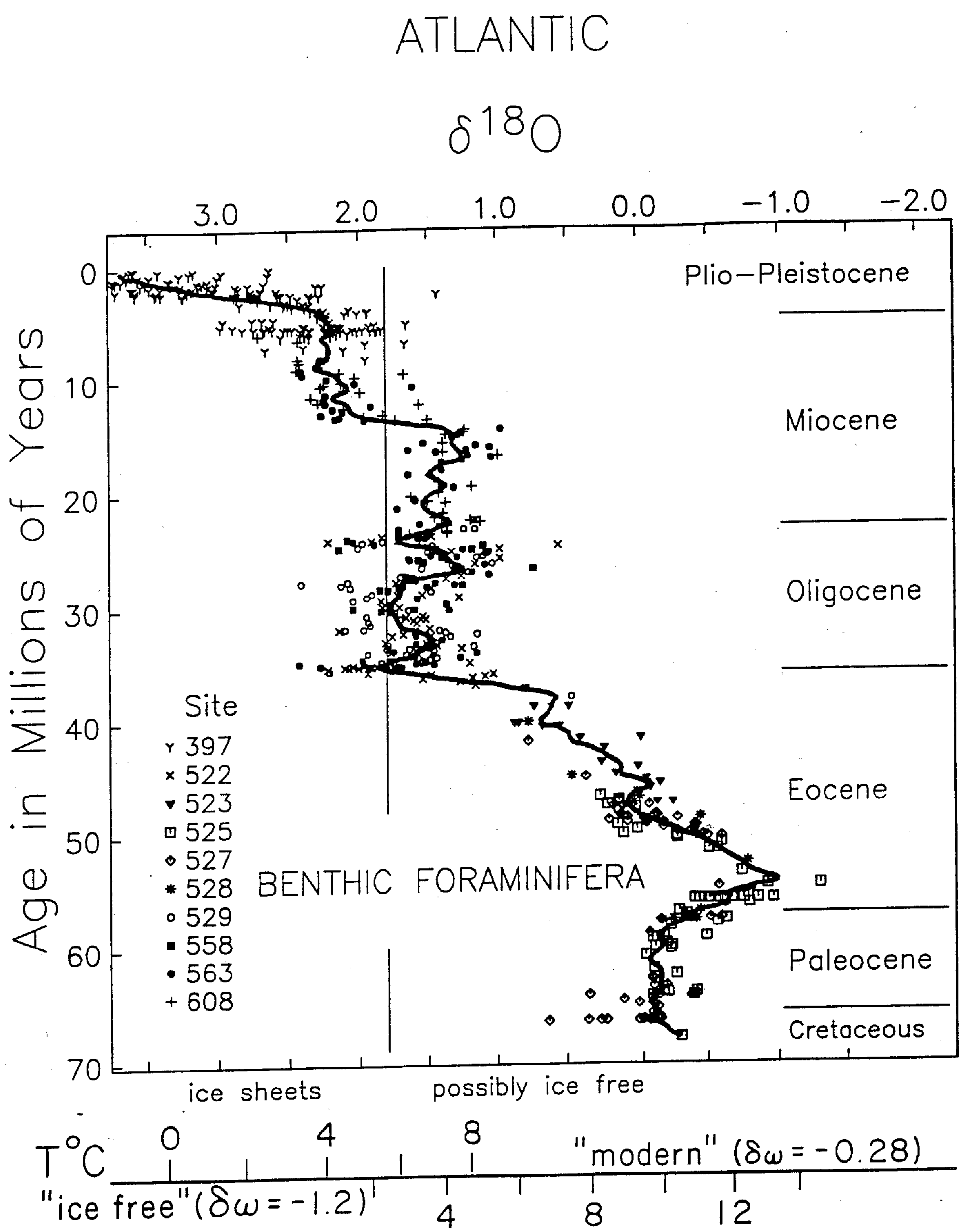

***Figure 1:** Composite of benthic foraminiferal* (Cibicidoides spp.) *$\delta^{18}O$ record for Atlantic DSDP sites (from Miller* et al., *1987). The smoothed curve was obtained by linearly interpolating between data at 0.1 my intervals and smoothing with a 21-point Gaussian convolution filter which 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 $\delta^{18}O$ 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 $\delta^{18}O$ increases can be recognized in the Savin (1977) compilation, and less than 10 $\delta^{18}O$ increases are distinguished in the Miller *et al.* (1987) synthesis.

Only recently have deep water $\delta^{18}O$ 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 $\delta^{18}O$ 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 $\delta^{18}O$ 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 $\delta^{18}O$ 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 $\delta^{18}O$ 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 $\delta^{18}O$ records from these waters should reflect ice volume changes. Under this premise, low latitude planktonic foraminiferal $\delta^{18}O$ 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 $\delta^{18}O$ 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 $\delta^{18}O$ records alone is limited. However, one can apply a Pleistocene strategy which uses concomitant changes in benthic and low-latitude (non-upwelling) planktonic $\delta^{18}O$ records as indicators of ice-volume/sea-level change (Shackleton and Opdyke, 1973). Covariance between benthic and low-latitude planktonic $\delta^{18}O$ 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 $\delta^{18}O$ 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 $\delta^{18}O$ record (Miller and Fairbanks, 1985; Miller *et al.*, 1987, 1991; Wright *et al.*, in press). Low latitude planktonic $\delta^{18}O$ records from the western equatorial Atlantic, Pacific, and Indian Oceans have confirmed that many of the short-term $\delta^{18}O$ 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 $\delta^{18}O$ records.

The last step in this process is to correlate the $\delta^{18}O$ 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 eustacy. 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 formininfers), 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 $\delta^{18}O$ 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

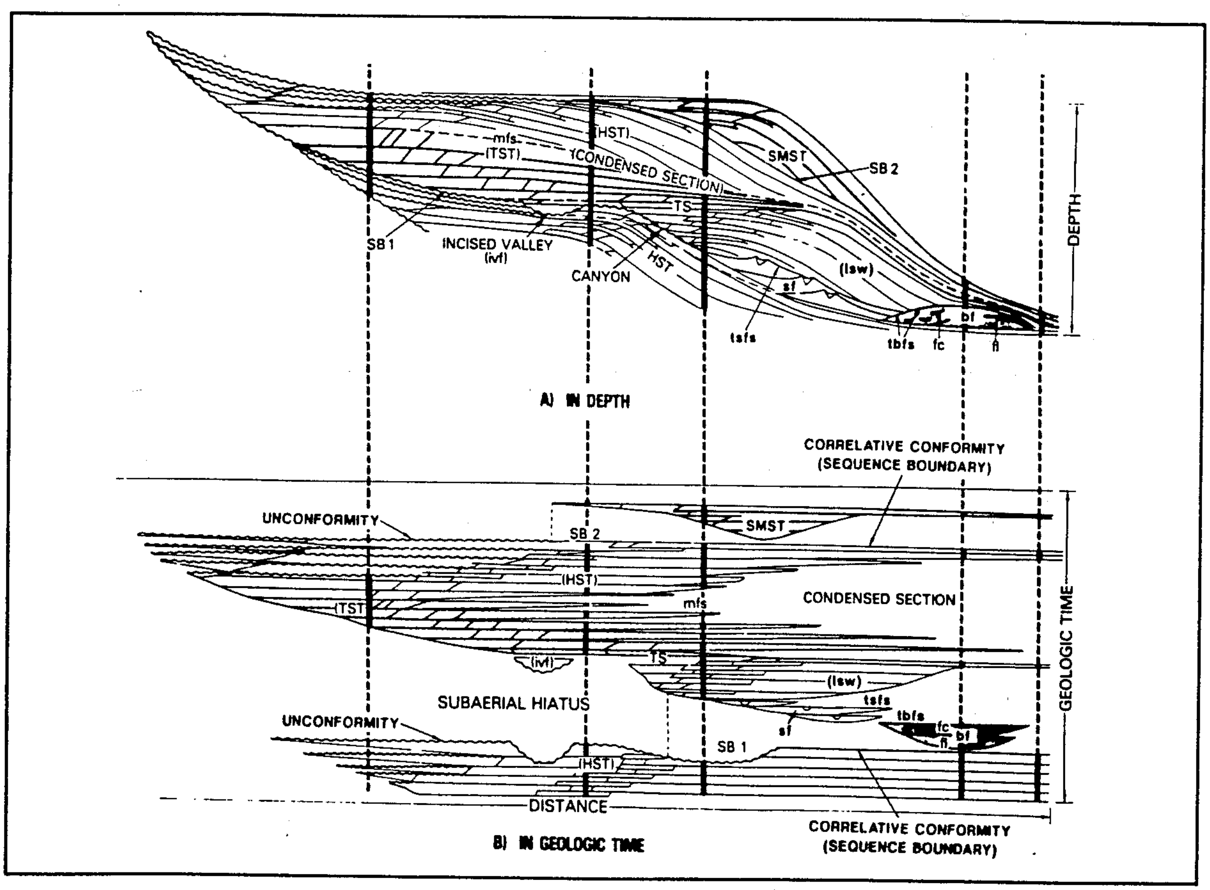

***Figure 2:** Idealized depositional sequence in depth (upper) and geologic time (lower))from Vail, 1987). To evaluate properly the age and facies associated with sequence-bounding unconformities, conformable correlatives, and stratal relationships within sequences, it is necessary to core and log a 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, eustacy, 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 eustacy.

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 CO_2 in the global heat budget; and (3) determining the contributions of specific large ice sheets to the last interglacial.

Modern Sea-Level Change — Separating Isostacy From Eustacy.

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 manmade, 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

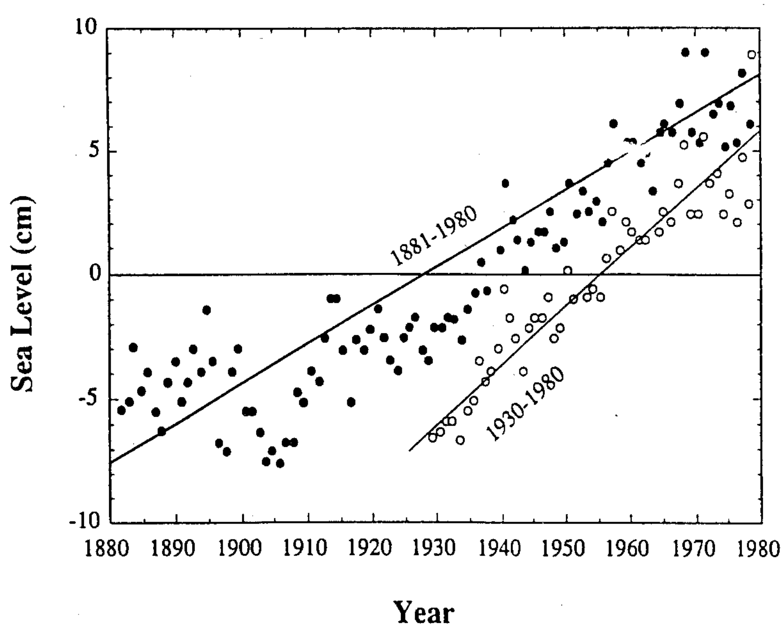

Figure 3: Estimates of "overall average" sea-level rise based on EOF analysis of regional tide gauge records for two different time periods. (from Barnett, 1984).