Haq et al. Eustatic Sea Level Curve: Implications for Sequestered Water Volumes[1]

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ABSTRACT

The eustatic sea level curve of Haq and co-workers differentiates a long and a short term signal. The mechanism that controls the long term curve is primarily variation in the volume of the ocean basins, while the short term curve reflects changes in ocean water volume. Only glacial eustasy operates at the frequency and with sufficient magnitude to control the short term curve. At each increment of time the volume of water sequestered from the ocean basins to account for the difference between the long and short term curves is computed, incorporating both hypsometric and water loading factors. An amount equivalent to that of present day Antarctica was sequestered at an approximately 5 Ma frequency since 145 Ma, but direct geological evidence of Late Jurassic to Early Tertiary continental ice of this equivalent magnitude is lacking. Because the existing ice coverage of Antarctica inhibits direct observation, we use the known fractionation affects of sequestration of water on the oxygen isotopic composition of sea water as a direct proxy for evidence of removal of large volumes of water from the oceans. Using fractionation factors estimated for different modes of water extraction of between +0.4‰ and +1.6‰ for a drop in sea level of 120 to 130 m, the changes in the isotopic composition of sea water oxygen are retrodicted using the volumes of sequestered water computed from the Haq et al. eustatic sea level curves. The retrodicted oxygen isotopic compositions for the interval between 60 and 30 Ma are compared with measured values of planktonic formanifera from several DSDP sites. This comparison reveals virtually no correspondence between the retrodicted and observed $\partial^{18}O$ records. This explicit test of the implicit requirements of the Haq et al. sea level curve indicates that the magnitudes and very existence of many of the short term eustatic oscillations are unsubstantiated. We then compute the coastal onlap history implicit in the long term curve, which shows many of the major shifts recorded by the coastal onlap curve of Haq et al. but lacks the highest frequency signals. We infer that the observed coastal onlap curve is dominated by the long term curve and that much more subtle changes in sedimentation and or subsistence are responsible for the sequences recorded by the coastal onlap curve. The short term may in part be an artifact of the assumed correlation of onlap with times of increased rate in sea level rise.

Introduction

A knowledge of the history of eustatic sea level change is important in understanding the marine and coastal plain stratigraphic record. McDonough and Cross (1991) review the various approaches employed to determine past sea levels. Workers from the petroleum industry, and particularly from Exxon, have developed a wide array of seismic stratigraphic techniques to interpret the stratigraphic sequences in the subsurface and have compared these with their up-dip surface exposures (Baum et al. 1988). This work has led to the examination of the factors that control coastal onlap and offlap, with particular emphasis on the role of eustatic sea level changes. The recent Society of Economic Paleontologists and Mineralogists Special Publication "Sea Level Changes: An Integrated Approach" (Wilgus et al. 1988) provides the most comprehensive summary of this approach to the interpretation of sequence stratigraphy and its implications for sea level history over the past ~255 Ma. As a part of these presentations Haq et al. (1988*a*) integrate biostratigraphic, chronostratigraphic, and sequence stratigraphic information in terms of a large format chart portraying the Haq et al. (1987) time scale, and correlations of magnetostratigraphy, biostratigraphy, sequence chronostratigraphy, as well as presenting a eustatic sea level curve.

Haq et al. (1988*a*) distinguish two eustatic

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curves, a "long term" curve and a "short term" curve, that are portrayed as heights of past sea level (in meters) relative to present sea level. According to Haq et al. (1988*a*), the scaling to meters represents "our best estimate of sea level rises and falls compared to the modern average mondial sea level" (p. 93). The long term curve is calibrated using the difference between thermo-tectonic subsidence and calculated geohistories of offshore well data (Hardenbol et al. 1981). The analysis takes into account thermal subsidence, isostatic effects of paleo-water depth changes, and eustatic differences. The high Late Cretaceous (Turonian) value (~255 m) was taken from Harrison (1986). The short-term sea level curve is a best estimate from the relative magnitudes of changes from seismic and sequence stratigraphic data (Haq et al. 1988*a*, p. 93) derived from the interpretation of coastal onlap and offlap.

Much criticism has been directed toward the Haq et al. (1987, 1988*a*) sea level curves primarily because neither the actual data upon which it was based nor the algorithm by which the seismically observed sequences have been transformed into changes in height have been published. As a consequence, these criticisms cannot be direct, but are instead based on presumed methodologies (e.g., Christie Blick et al. 1988, 1990), or based on a comparison with alternative interpretations of the eustatic sea level history (e.g., Matthews 1988). These criticisms are themselves criticized by Haq et al. (e.g., Haq et al., 1988*b*, Vail and Haq 1988) because they do not provide explicit tests of the curve.

We have therefore pursued the Haq et al. (1987, 1988*a*) interpretation from an alternative perspective of looking at the inherent implications of their interpretation for both history of water sequestration and oxygen isotope record of sea water. We develop explicit predictions directly from the Haq et al. curves and compare these with other monitors or models of sea level.

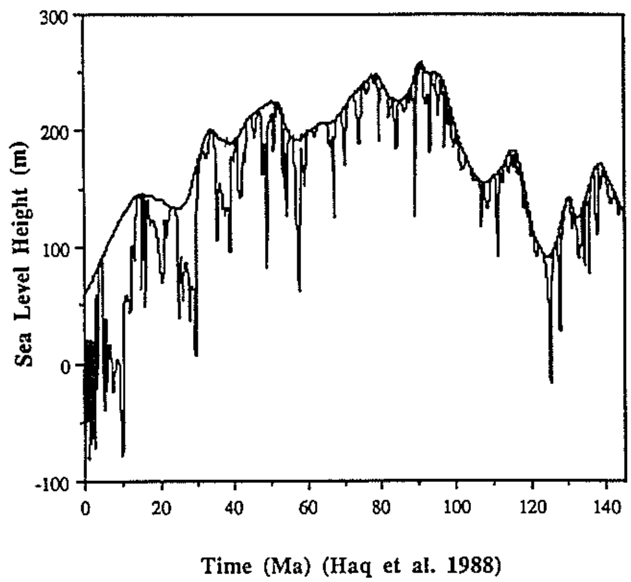


Figure 1. Plot of the long term and short term curves of Haq et al. (1988*a*) for the time interval of 145 Ma to present.

Causal Mechanisms of Sea Level Change

Figure 1 is a plot of the short term and long term curves for the period from 145 Ma to the Present; the characteristics are summarized in table 1. To remove the known development of the Antarctic ice cap, table 1 also summarizes the characteristics of the interval from 145 to 40 Ma separately. This comparison suggests that there is no demonstrable difference between times when glacial-eustasy is widely acknowledged and times before 40 Ma, when glacial eustasy remains uncertain. Although the average rates of change for the short term curve are quite high, the maximum rates inferred by Haq et al. (1988*a*) are an order of magnitude faster, which becomes important in discussion of possible mechanisms that might have controlled these proposed changes in sea level.

Haq et al. (1988*a*) do not discuss an explicit

Table 1. Summary of the Haq et al. (1988) Long and Short Term Eustatic Sea Level Curves

Time (Ma)	No. of Cycles	Frequency (Cycles/Ma)	ΔSL (m)	Av. Rate (m/m.y.)	Max. Rate (m/m.y.)
Short Term Curve					
0 to 145	89	.62	±41	±52	>800
40 to 145	62	.59	±36	±42	>600
Long Term Curve					
0 to 145	8	.06	±38	±5.1	>10

Table 2. Factors and Rates of Mechanisms Controlling Short and Long Term Eustatic Sea Level

	Magnitude (m)	Duration m.y.	Rate m/m.y.
Climatic Controls: Ocean Water Volume			
Glacial	120	20×10^{-3}	6.0×10^3
Dessication[a]	10	3×10^{-3}	3.3×10^3
Ocean temperature[b]	10	1×10^{-3}	10×10^3
Tectonic Controls: Ocean Basin Volume			
LT Ridge Vol. (max)[c]	300	75	4.0
LT Ridge Vol. (mean)[c]	200	75	2.7
LT Ridge Vol. (min)[c]	100	75	1.3
Max Interval Ridge Vol.[c]	45	5	9.0
Collisional decrease in continental area[d]	⇓70	50	1.4
Extensional increase in continental area[e]	⇑15	70	.2
Sediment infilling[d]	⇑35	50	.7
Hot spots[f]	⇑35	70	.5
Sum of Tectonic Rates:		LT (max)	4.0
		LT (mean)	2.7
		ST (max)	9.0

Note. LT = long term; ST = short term.
[a] Based on Mediterranean (Hsu et al. 1973).
[b] Based on Sahagian (1988).
[c] Based on Kominz (1984).
[d] Based on Rowley (unpub. data) derived from India-Asia collisional system.
[e] Based on Heller and Angevine (1985).
[f] From Pitman and Golvechenko (1991).

mechanism that has controlled short term sea level, except to mention that, "At least since the Oligocene, sea-level falls are in part due to increasing influence of glaciation" (Haq et al. 1988*a*, p. 93). They state further that "Today's long- and short-term sea levels diverge because the long term eustatic curve is estimated assuming an ice free world." It is not clear whether the common divergence of these curves further in the past is similarly assumed to reflect the difference between an ice free (long term) versus an ice present (short term) world. It is clear, however, that Haq et al. (1988*a*) conceive of the short term curve as depicting the actual history of sea level change while the long term curve maps the overall trend in sea level. These views are supported by more explicit discussions in Vail and Haq (1988, p. 599) suggesting that "shorter term eustatic changes . . . are most probably due in large part to changes in polar ice volume," although they also may be controlled in part by some other, as yet unknown, mechanism (p. 599).

The difference in frequency and particularly rate of change of the long versus short term curves leads to a natural distinction that reflects a difference in their respective causal mechanisms (Pitman 1978; Harrison 1986; Vail and Haq 1988). The long term curve reflects changes in volume of the ocean basins through time, whereas the short term curve reflects changes in ocean water volume. This view derives directly from estimates of maximum rates of change as a consequence of the various mechanisms that control ocean basin and ocean water volume. Table 2 lists total sea level changes and rates of change (dSL/dt) for some of the potential contributors to eustatic sea level change. It is obvious from table 2 that only desiccation and refilling of isolated ocean basins, such as apparently happened during Messinian times in the Mediterranean (Hsü et al. 1973), and glacial buildup and destruction are characterized by rates of change compatible with the short term curve. Because the history of desiccation and refilling of isolated ocean basins is quite limited (Mediterranean and Red Sea-Miocene, South Atlantic-Aptian, Gulf of Mexico-Callovian) and the amplitude of the effect is probably quite small (10–15 m), the importance of this mechanism as a potential driver of the Haq et al. short term sea level curve appears to be quite limited. We therefore conclude that the only plausible mechanism that might control the short term curve is glacial eustasy. This is supported by the general similarity of the pre- and post-Eocene parts of the Haq et al. curve.

Table 3. Distribution of Area Versus Elevation from −200 m to +300 m from Cogley (1985)

Elevation Intervals (m)	Area/100 m (sq. km)	Cumulative Area (sq. km)
−199– −100	74522	−186998
−99–0	112476	−112476
0–99	115827	115827
100–199	113250	229077
199–299	303940	533017

Data

The data employed in this analysis are a digitized version of the large chart from Haq et al. (1988*a*) in which the elevation of both long and short term cycles were sampled at even 100,000 yr intervals from 145 Ma to the Present. Note that the short term record is truncated at −70 m, but this only affects the Pliocene and Pleistocene record. The Haq et al. (1987, 1988*a*) time scale is used so as not to introduce potentially spurious effects due to variations in time scales. For reference, the Haq et al. time scale places the Jurassic-Cretaceous boundary at 131 Ma and the Cretaceous-Tertiary boundary at 66.5 Ma. The time range examined extends from the Late Jurassic (base of the Kimmeridgian) to the Present with the specific intent of sampling the Cretaceous, a time interval for which good paleogeographic control exists and for which we can assess with some confidence the implications of the Haq et al. curve for ice-related storage of water on the continents.

As part of the computation of the volume of sequestered water, we have used the compilation of continental hypsometry of Cogley (1985) to include the effects of continental flooding. Table 3 and figure 2 (inset) show the cumulative area versus elevation at 100 m intervals between −200 m and +300 m. Figure 2 (inset) shows a very nearly linear correlation between elevation (E) and area (A) between −200 m and +200 m elevation defined by the least squares fit at

$$A_{-200\text{ to }+200} = 99240 * E \tag{1}$$

Applying equation 1 to elevations >200 m results in an approximately 48% underestimate of the area at 255 m elevation. For this reason, the areas at elevations >200 m are computed by linearly interpolating using

$$A_{+200\text{ to }+300} = (E - 200)/100 * A_{300} + A_{cum} \tag{2}$$

where A_{300} is the area between 200 and 300 m elevation and A_{cum} is the cumulative area from 0 to 200 m elevation computed from equation 1.

For the present analysis we assume that the distribution of area per elevation interval, between +255 m and −130 m, has not changed significantly since the end of the Jurassic. We believe that the error incorporated as a result of this assumption minimizes the implied volumes of sequestered water because during much of the Cretaceous and Early Tertiary the area within 300 m of present sea level was most likely greater than today, as judged by the distribution of Cretaceous marine sediments now exposed in areas above 255 m present elevation. As will be shown below, the dominant term in the volume calculation is A_0, the present sea surface area (364,134,000 km^2) (Times Atlas

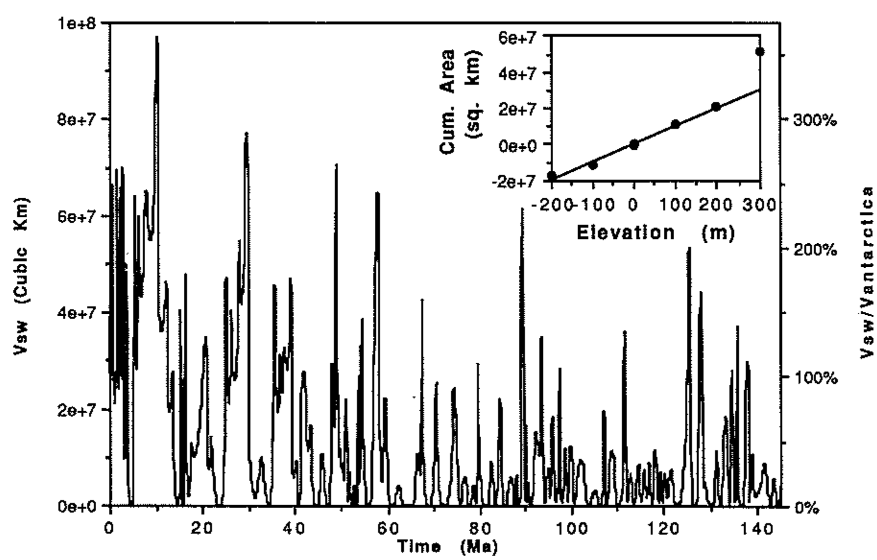


Figure 2. (Inset) Plot of elevation versus area for elevations from −200 m to +300 m from Cogley (1984). Plot of the volume of sequestered water over the interval 145 Ma to Present. Right-hand axis scaled to presently sequestered volume of water in grounded ice in Antarctica. For reference, the volume of sequested water at the last glacial maximum is equal to 7.67 × 10^7 km^3.

1988). For the interval since 145 Ma, A_0 represents >90% of the total area covered by sea water.

Relation between Sequestered Water Volume and Sea Level

The relation between the volume of sequestered water, V_{sw}, (and note $V_{sw} = 0.92*V_{ice}$ and change in the sea level height (dSL) is described by equation 3 for sea levels between -200 m $\leq dSL \leq$ 200 m,

$$V_{sw} = 1/2\,A\,dSL + A_0\,dSL \tag{3}$$

where A_0 is the present surface area covered by sea water and A the cumulative area. The 1/2 reflects the fact that cross-sectional area is approximated as that of a right triangle. Equating dSL with a change in E and substituting from equation 1 for A, equation 3 can be rewritten as:

$$V_{sw} = 49620*dSL^2 + A_0dSL \tag{4}$$

which can be solved directly for sea level change associated with the complete melting of grounded ice from Greenland, Antarctica, or all grounded ice (see table 3 for volumes), yielding values of ~7 m, ~74 m, and ~81 m, respectively, for the instantaneous effect of melting. Incorporating effects of water loading, the longer term rises would be ~5 m, ~56 m, and ~62 m, respectively. Antarctica and Greenland ice sheets account for ~97% of all grounded ice, with most of the rest on Baffin and Ellesmere islands, the Himalayas and Tibet, Alaska, and the Andes. The amount of water that needs to be sequestered to yield a known drop in sea level (for example, that needed for the last glacial maximum drawdown of between -120 m and -130 m) can be computed from equation (4). The additional volume of water locked up in ice at the last glacial maximum was between 4.3 and 4.6 × 10^7 km^3 more than today.

In order to determine V_{sw} using the Haq et al. sea level curves, in which the elevations of the long and short term curves are specified, the volume of water sequestered at each interval, V_{si}, is described in terms of the areas at elevations corresponding to the long (A_{li}) and short (A_{si}) term curves at each increment time (i) multiplied by $SL_{(l\text{-}s)i}$,

$$V_{si} = [A_0 + A_{si} + 1/2(A_{li} - A_{si})]*1.309*SL_{(l\text{-}s)i} \tag{5}$$

the difference in elevation of the long and short term curves at increment i. The factor 1.309 reflects the effects of Airy water loading ($\Delta SL = dSL + \rho_w/\rho_m*dSL = dSL + 1.03/3.33*dSL$).

In figure 2, the volumes of sequestered water over the past 145 Ma account for the differences in elevation of the long and short term cycles computed using equation 5. It shows that over this interval significant volumes of water are implied to have been sequestered from the ocean basins. The mean value of V_{si} is ~1.3 × 10^7km^3 over the entire 145 Ma record, but there is a very large variation about the mean. To put these volumes into some perspective, in figure 2 we also compare these volumes with the presently sequestered water in Antarctica. During 19 intervals between 145 and 40 Ma more water was sequestered than the modern day grounded ice of Antarctica. This corresponds to a mean frequency of approximately once per 5.5 Ma, or about once per stage. The mean volume is between 3 and 4 times the present volume of the Mediterranean (Times Atlas 1988) and approximately 16 to 17 times the volume of the Black Sea (Times Atlas 1988). Thus it is extremely difficult to imagine that this much water could be kept completely isolated from the main ocean basins in the liquid state, and we reinforce the previous suggestion that the most likely mechanism for sequestering this much water would be as ice.

Relationship between Volume of Sequestered Water and Areal Coverage of Ice

The volumes implied are significant but hard to appreciate from a geological perspective. In order to gain a better appreciation of what is inherent in the Haq et al. curves, the volumes can be converted into areal coverage, assuming that it is all locked up in a single ice sheet with a circular footprint. We tried methods that employed approximations based on both parabolic and elliptical cross-sectional ice sheets (Paterson 1969), but when compared to the present sizes of Antarctica and Greenland, neither method yielded satisfactory estimates of known area from their known volumes (table 4). A third technique, using an empirical correlation between areal ice coverage and volume, is based on the volumes and areal coverage of an Ice Free World, Greenland, Antarctica, and the Last Glacial Maximum. The least squares fit to the data that relates area of coverage to volume shown in Fig. 3, (inset) is

$$A_{ice} = 0.48*V_{ice}. \tag{6}$$

Figure 3 plots A_{ice} corresponding to these V_{si} derived from the Haq et al. curve based on equation 6.

Table 4. Comparison of Areal Coverage Computed from Known Volumes of Sequestered Water with Observed Area Coverage for Greenland and Antarctica

	Volume ($10^6 km^3$)		Area ($10^6 km^2$)						
	Ice	Water	Obs.[a]	Parabolic[b]		Elliptical[b]		Empirical[b]	
Greenland (1)	2.6	2.4	1.7	1	59	1	59	1.3	73
Antarctica (2)	29.6	27.2	12	7.2	60	11	92	14.3	119

Sources. (1) Lydolph (1985), (2) Drewery (1983).
[a] Observed.
[b] Computed % for each method employed (see text).

The striking feature of figure 3 is that the computed areas covered by ice frequently approximate or exceed that of present Antarctica. The average implied ice coverage is 40% of Antarctica. Figure 3 also shows that the maximum implied coverage is in excess of 200% of Antarctica prior to 40 Ma. Clearly at times when implied areal coverage exceeded that of Antarctica, other continental regions contributed to the sequestering of water. The Haq et al. eustatic curves requires significant ice, not just small isolated mountain ice sheets that today represent <3% of the total grounded ice volume. The Haq et al. curves imply an earth history significantly effected, if not dominated by, ice over the past 145 Ma.

It is also worth noting that the predicted paleoclimatic effects associated with increased albedo on such large ice sheets would be even greater, because we have only computed the grounded ice, and not that of the equally climatically significant floating ice. At present, the maximum extent of floating ice is ~2.0 times that of the grounded ice ($20 \times 10^6 km^2$ for southern hemisphere winter (Zwally et al. 1983) plus $8 \times 10^6 km^2$ (Parkinson et al. 1987) for northern hemisphere summer). By analogy we would expect that total ice coverage in the past might have attained similar proportions, thereby accentuating the predicted paleoclimatic effects.

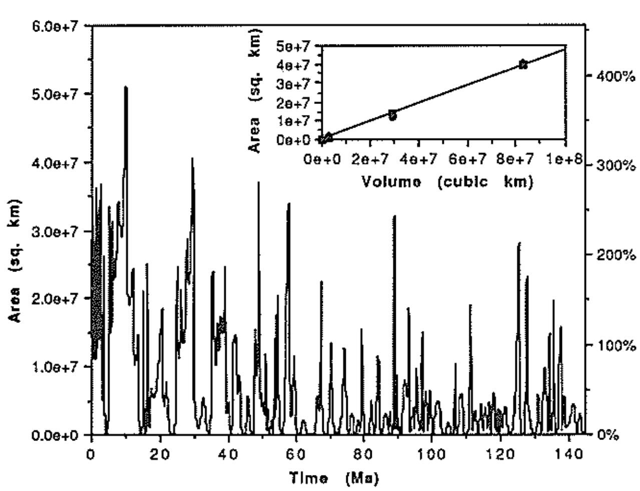


Figure 3. Plot of the area of ice coverage vs. time based on the area-volume correlation. Right-hand axis scaled to the area of grounded ice in present day Antarctica. Inset: Graph showing the correlation of the area of ice coverage versus volume of ice.

It should be clear that if the Haq et al. eustatic curves are corect, the geological record should clearly reflect a history of ice. In their comprehensive survey of the pre-Pleistocene glacial record, Hambrey and Harland (1981) report no examples of undisputed glacial deposits from the Late Jurassic to the Cretaceous. More recently. Frakes and Francis (1988) have discussed Cretaceous ice, reiterating earlier interpretations (Brown 1894; Woolnough and David 1926) of potential glacial dropstones in the Bulldog Shale of eastern Australia. However, the absence of glacial pavements or substantial sequences of other potentially glacial-marine sediments of Late Jurassic to Early Tertiary age suggest that even if these deposits were left by ice, the ice was not extensive and potentially not continental, as re-interpreted by Frakes and Francis (1990). The absence of unequivocal Jurassic to Early Tertiary evidence of continental ice is in obvious and marked contrast with the widely distributed and well described Late Precambrian, Late Ordovician, and Permo-Carboniferous glacial deposits and glacial pavements summarized in Hambrey and Harland (1981).

Geological and paleontological evidence of the history of ice on Antarctica is limited particularly for East Antarctica (Webb et al. 1984), where a record of Late Jurassic to Early Tertiary glaciations may exist under the ice. However, existing evidence from Antarctica (Truswell 1989; Drinnan and Crane 1989; Spicer 1989) and surrounding waters of the southern oceans indicate much less extreme conditions at high southern latitudes in the

Jurassic and Cretaceous (Truswell 1983, 1989), as can also be said for high northern latitudes during the same time (Spicer and Parrish 1986; Francis 1988).

At present we are left with rather conflicting situations. If the Haq et al. sea level curve is correct, significant volumes of water must have been sequestered. Ice provides the mostly likely sequestering mechanism considering the quantities involved and the rates at which this water must be removed from and then equally rapidly replaced into the oceans. At the same time, studies of the geology and paleontology of high latitude sites of the Jurassic, Cretaceous, and Early Tertiary suggest that widespread continental ice sheets did not exist. In addition, questions remain concerning the calibration of the long and short term curves (see Burton et al. 1987; Christie-Blick et al. 1990), the relationship of the coastal onlap curve to sea level variations (Pitman 1978; Pitman and Golvechenko 1983), and the geological evidence pertaining to past climates. The arguments outlined above hinge on negative, but incomplete, evidence and are therefore not very satisfactory. It would be preferable to have a direct proxy for the amount of water removed from the oceans through time.

Oxygen Isotopes as Indicators of Sequestered Water Volumes

The isotopic composition of oxygen in sea water is controlled primarily by temperature and the volumes of water sequestered from the oceans. Fractionation of oxygen isotopes occurs as a consequence of energetically favored evaporation of ^{16}O-enriched water leading to an enrichment of ^{18}O in the oceans. A direct and explicit test of the Haq et al. (1988) curve is possible using oxygen isotopes as a direct measure of the volumes of water extracted from the ocean basins.

Equation 8, from Shackleton (1974), provides the standard equation to extract temperature from measurements of the oxygen isotopic composition of fossil material ($\partial^{18}O_c$) based on an *assumed* model of the history of changing $\partial^{18}O_w$,

$$T(°C) = 16.9 - 4.38\,(\partial^{18}O_c - \partial^{18}O_w) + 0.1\,(\partial^{18}O_c - \partial^{18}O_w)^2 \qquad (7)$$

where $\partial^{18}O_c$ and $\partial^{18}O_w$ are respectively the isotopic composition of calcite and sea water relative to some standard. The fundamental problem is that there is one equation and two unknowns, T(°C) and $\partial^{18}O_w$. This has led various workers to develop different models based upon different assumptions concerning either the history of $\partial^{18}O_w$ to determine T(°C) (for example Shackleton 1986) or the constancy of tropical sea surface temperatures at ~28°C, allowing T(°C) to be fixed and thereby allowing $\partial^{18}O_w$ to be determined. In this case $\partial^{18}O_w$ can be interpreted to reflect the amount of sequestered water (Matthews 1984; Prentice and Matthews 1988).

Emiliani and Shackleton (1974) pointed out that bounds on the effect of extraction of water equivalent to a 120 m drawdown of the oceans (equivalent to sequestering 4.3×10^7 km³) can be placed between +0.4 and +1.6‰. They argue that a shift +0.4‰ would result "if the water removed had the oxygen isotopic composition of water evaporated in isotopic equilibrium with the subtropical sea surface" (Emiliani and Shackleton 1974, p. 511). A shift of +1.6‰ would result "if the water removed had an isotopic composition similar to that of snow in the interior of Antarctica" (p. 511). These bounds correspond to an isotopic shift per volume of water removed of 9.30×10^{-9} ‰/km³ and 3.72×10^{-8} ‰/km³. Fairbanks and Matthews (1978) independently estimated the variation in $\partial^{18}O$ from an analysis of Pleistocene reef traces on Barbados. They used the difference in $\partial^{18}O$ and elevation to determine a correlation of 0.011‰/m, used by Matthews (1984), Williams (1988), and Prentice and Matthews (1988) to estimate past sea levels. The Fairbanks and Matthews' value corresponds with a value of 3.02×10^{-8} ‰/km³. This independent calibration supports the bounds suggested by Emiliani and Shackleton (1974) and points toward the higher end of this spectrum rather than the lower for ice-related sequestration of water. The importance of the lower bound estimated by Emiliani and Shackleton (1974) is that if some other mechanism, for example storage in ground water or lakes not connected to the oceans, contributed to the sequestration of water, then a fractionation effect closer to this value would be expected.

Figure 4 shows the hindcasted shifts in the isotopic composition of seawater based on the values suggested by Emiliani and Shackleton (1974) and computed from the amounts of sequestered water from Figure 2. Note that these are retrodicted shifts in $\partial^{18}O$ of seawater, assuming that 0‰ represents the ice free world. Figure 5 focuses in more detail on the interval between 30 and 60 Ma showing quite large and rapid shifts retrodicted at 30 Ma, 35 Ma, ~39 Ma, 49 Ma, and between 57 and 58.5 Ma. For glacial sequestration of water the shifts should be >1‰ to almost 2.5‰.

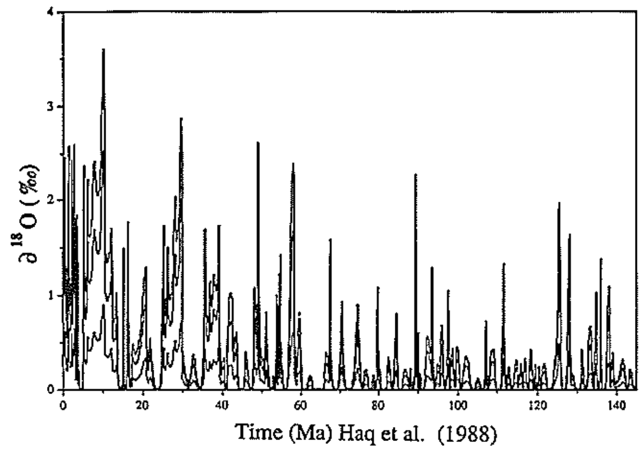


Figure 4. Plot of retrodicted oxygen isotopic composition of sea water based on the volumes shown in figure 2 and fractionation factors corresponding with a +0.4‰ (lowest curve), +1.1‰ (middle curve), and +1.6‰ shift associated with a 120 m drawdown at the Last Glacial Maxima (see text for discussion).

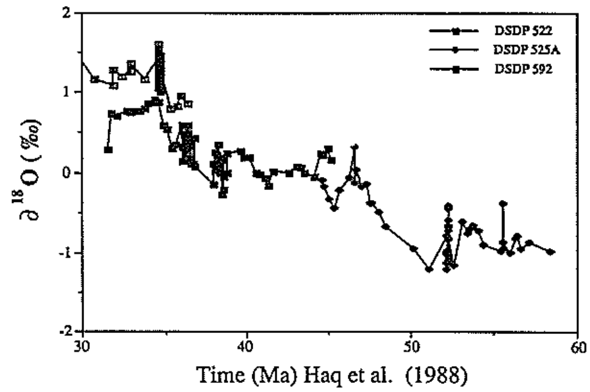


Figure 6. Plot of measured $\partial^{18}O$ of planktonic foraminifera from DSDP sites 522, 525A, and 592. Data for DSDP site 522 are based on Poore and Matthews (1984) for *Globigeroides euapertura* and *Globigerina galavasi*. DSDP site 525A data on *Morozovella* sp. are for Shackleton et al. (1984). DSDP site 592 data are from Murphy and Kennett (1986) for *Globigerina angiporoides*. Plots show all reported values for the given taxa at those sites.

Existing $\partial^{18}O$ Data

Prentice and Matthews (1988) have recently summarized Cenozoic $\partial^{18}O$ values of both tropical shallow-dwelling planktonic foraminifera and deep-water benthic foraminifera. They have constructed composite $\partial^{18}O$ records from a number of different DSDP cores and applied both vital and inferred latitudinal corrections in these Cenozoic sub-tropical composites. Because the corrections employed by Prentice and Matthews (1988) are rather large (ca. $\geq$1.7‰), we have analyzed the actual measured values used by Prentice and Matthews (1988) but uncorrected for the inferred latitudinal effects.

We employ data from sites DSDP 522 (Poore and Matthews 1984) for 30 to 36 Ma, 525A (Shackleton et al. 1984) from the Walvis Ridge area of the South Atlantic for 45 to 60 Ma, and 592 (Murphy and Kennett 1986) from the Lord Howe Rise in the Tasman Sea for 31 to 45 Ma (figure 6). We use the specific foraminifera employed by Prentice and Matthews (1988), and have converted the reported ages to the Haq et al. (1988) time scale using a linear interpolation between correlative stage boundaries as tie points. Figure 7*a–c* compares the measured values against the retrodicted history prescribed by Haq et al. (1988*a*) sea level curves (figure 4 and 5).

The primary purpose of figure 7 is to compare the presence or absence of correlations in the times of change and in relative magnitudes of the shifts. We tacitly assume that the signals recorded by the $\partial^{18}O$ of the planktonic foraminifera reflect primarily changes in ice volume, following the assumption that changes in latitude of the individual sites is sufficiently small such that the sites are unlikely to move significantly with respect to the latitudinal sea surface temperature gradient. As a first-order approximation, 1°C changes of sea water temperature result in a shift in $\partial^{18}O_w$ of $\sim\pm0.2$‰, which would similarly apply to variations in water depth occupied by the measured foraminifera. Thus some, and possibly much, of the variations portrayed in these data may reflect small variations in sea water temperature, which presumably account for the differences between measured values at the same heights within the cores of the same species of foraminifera, as for example seen in the plots as vertically arranged points.

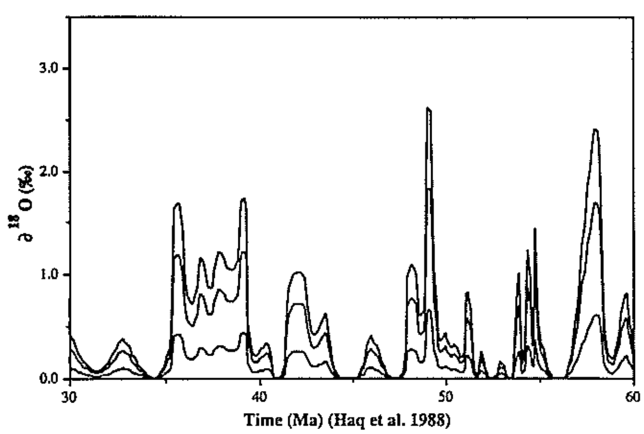


Figure 5. Enlargement of figure 4 focusing specifically on the interval between 30 and 60 Ma. Three curves are identical to those in figure 4.

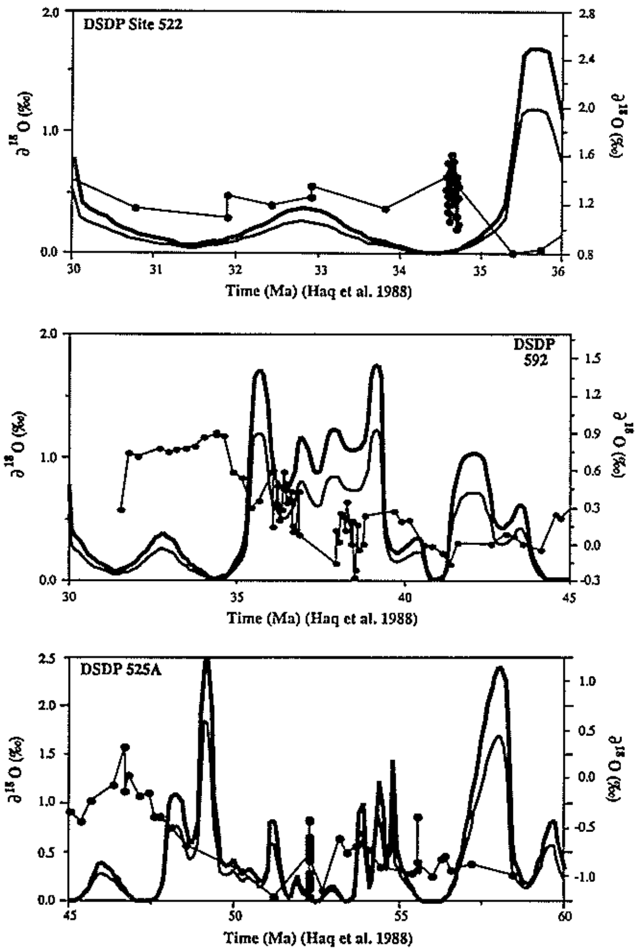


Figure 7. Comparison of the retrodicted $\partial^{18}O$ from figure 4 showing only the middle and upper two curves with the measured values from DSDP site 522, 592, and 525A, plotted related to the Haq et al. (1988*a*) time scale. The left hand axis scales the shifts in $\partial^{18}O$ based on retrodicted volumes and on fractionations of 2.56×10^{-8} ‰/km^3 and 3.72×10^{-8} ‰/km^3 of the two continuous curves, while the right hand axis scales the measured vaules. The scales have been shifted in order that the range of the right and left scales are identical, and that the minimum reported value is arbitrarily correlated with the zero ice volume. (*a*) Site 522; (*b*) Site 592; (*c*) Site 525A.

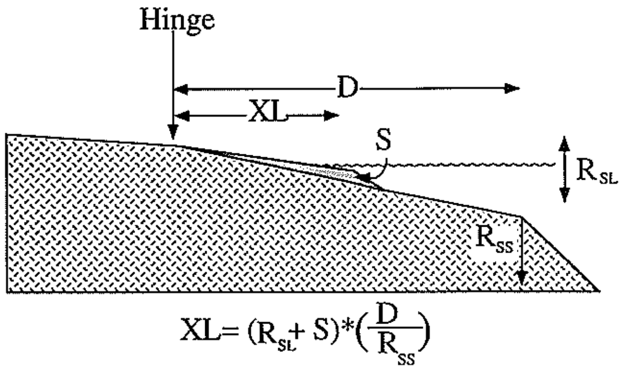
The important point illustrated by figures 7*a–c* is that although there are temporal variations in measured $\partial^{18}O$ values, the times and magnitudes of change do not correlate with those retrodicted by the Haq et al. sea level curves. In fact, there are few correlated changes and many anti-correlated events. For example, between 35 and 36 Ma, the Haq et al. curve hindcasts positive shifts of between +1.1‰ and +1.6‰, while the measured records of both DSDP sites 522 and 592 are characterized by minima. For the subsequent intervals, both DSDP sites 522 and 592 show increasingly enriched values while Haq et al. retrodicts more depleted values. Within the frequency band appropriate for the short term curve (i.e., 1 to 2 m.y.) none of the measured $\partial^{18}O$ records are characterized by >± 0.5‰ shifts, a factor of three to five less than the shifts expected from the Haq et al. (1988*a*) curves. The fact that neither the times nor the magnitudes agree with the retrodicted values suggests that, even if diagenetic alteration were to dampen the measured signal, it would not effect the times or directions of the shifts (F. Richter pers. comm. 1991).

Since the oxygen isotopic composition of sea water is accepted as a monitor of the development and sequestration of water (Prentice and Matthews 1988; Shackleton 1986) and since we have shown that the Haq et al. (1988*a*) sea level curves make explicit predictions concerning the volumes of sequestered water, the comparison shown in figure 7*a–c* can be viewed as an explicit test of the validity of the Haq et al. (1988*a*) curves. The lack of correlation of the eustatic sea level curve with the observed record of variations in the isotopic composition of sea water indicates that the inversion of the coastal onlap curve to a short term eustatic sea level curve is invalid, and that the magnitudes of the eustatically driven short term sea level changes must be significantly less than estimated by Haq et al. (1988*a*).

First-Order Curve and Coastal Onlap and Offlap

Negation of the Haq et al. (1988*a*) interpretation of the coastal onlap curves raises the important question as to the driving mechanism(s) of coastal onlap. Pitman (1978) demonstrated that the record of coastal onlap and offlap of a differentially subsiding ramp is not one of alternately rising and falling sea level, but instead, of changes in rates of rise or fall of sea level. However, Vail et al. (1977) and subsequent publications have relied heavily on inferring the short term sea level history from coastal onlap and offlap. And although the shapes of the "sea level" curves have changed significantly from 1977 to 1988, there remains a one to one correlation between times of coastal onlap and offlap and sea level rise and fall on the short term curve (see Haq et al. 1988*a* chart). These workers may have confounded coastal onlap and offlap with rapid eustatic sea level changes, as opposed to variations in rates of sea level change.

To examine this in more detail we accept as a starting point the long term curve of Haq et al. as a measure of changes in ocean basin volume. We



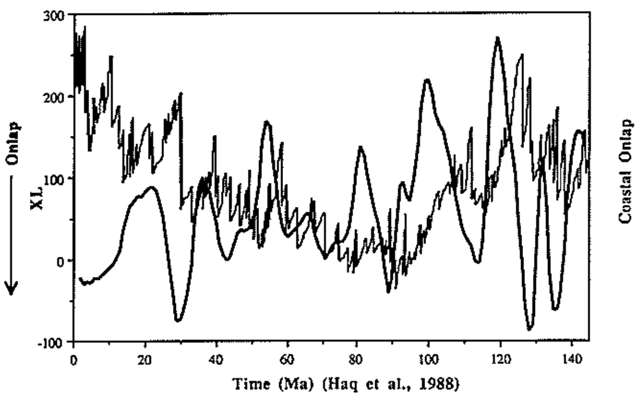


Figure 8. (*a*) Pitman (1978) cross-section showing the relation between strandline (X_L) and differential shelf subsidence. Where a shelf of width *D* measured from the hinge ($X = 0$) to the shelf edge ($C = D$), characterized by differential subsidence such that the rate of subsidence increases linearly from 0 and $x = 0$ to R_{SS} at $X = D$, but is partially offset by coastal plain sedimentation at rate *S*. The equilibrium position of strandline is defined by $X_L = (R_{SL} + S)^*(D/R_{SS})$, where R_{SL} is the rate of sea level change. (*b*) Plot of X_L vs. time in which *D*, *S*, and R_{SS} are held constant, and R_{SL} varies according to the Haq et al. long term sea level curve. Also shown is the coastal onlap curve of Haq et al. (1988*a*). Note that actual values of $X_L > 0$ are meaningless, they only indicate that sea level was rising faster than the rate of sedimentation *S*, and therefore the strandline has migrated landward of the hinge.

acknowledge that the rates of change implied by this curve locally exceeded theoretically possible rates, but that on average the rates of change could be sustained by the known mechanisms.

Pitman (1978) showed that the equilibrium strandline position on a hinged, differentially subsiding ramp is the point on the ramp that is subsiding at the same rate as sea level is falling (figure 8*a*). In its simplest form, the equilibrium position (X_L) of the strandline can be computed from a knowledge of the subsidence rate at the shelf edge (R_{SS}), the width of the shelf (*D*) measured from the hinge at $X = 0$ to shelf edge ($X = D$), the rate of coastal plain sedimentation (*S*), and the rate of sea level change (R_{SL}), such that

$$X_L = (R_{SL} + S)^*(D/R_{SS}). \qquad (8)$$

If $S = 0$, equation 8 simplifies to $X_L = R_{SL}^*(D/R_{SS})$, which only makes sense if sea level is falling. The effect of coastal plain sedimentation (*S*), when $S \neq 0$, is to shift the equilibrium strandline position out farther onto the shelf by an amount equal to $S^*(R_{SL}/R_{SS})$. During periods of sea level rise, coastal plain sedimentation maintains the position X_L seaward of the hinge until the rate of sea level rise exceeds the rate of sedimentation. Pitman (1978) and Pitman and Golvechenko (1983) have expanded equation 8 to incorporate the effects of changing subsidence rates through time as, for example, a passive margin matures, as well as effects associated with thermally driven re-attainment of equilibrium rates of sea level change as a consequence of changes in rates of seafloor spreading. We will ignore these effects and simply focus on the consequences of changes in rates of sea level change. We thus hold everything in equation 8 fixed, except R_{SL}. We adopt values of $D = 250$ km, $S = 5$ m/Ma ($= 0.5$ cm/1000 *a*), and $R_{SS} = 20$ m/Ma. We use the Haq et al. long term curve as input, by computing average values of R_{SL}. The computed X_Ls using equation 8 are shown in figure 8*b*. Because the rate of change on the Haq et al. curve is seldom constant, stasis of the strandline is only rarely attained. This plot clearly shows an inherent history of coastal onlap and offlap embedded in the long term curve. Figure 8*b* also compares the computed history of coastal onlap and offlap with the observed record of coastal onlap and offlap of Haq et al. (1988*a*). The observed curve portrays the Haq et al. (1988*a*) interpretation of the form of the coastal onlap curve characterized by progressive onlap followed by a very rapid seaward shift in the position of coastal onlap. This curve cannot be directly converted to the position of the strandline relative to the hinge, or any other metric that can be compared precisely to the computed curve. For our purposes the important comparison is the correspondence between times of landward or seaward shifts of the strandline and not to the specific shape or details of the magnitudes of the changes as they are not comparably scaled. In addition, it is not clear whether the long term secular variations portrayed on the observed onlap curve are real or simply reflect a general conformity to the long term sea level curve. When viewed in this way, the curves are quite similar with many well correlated events.

The important consequence of this exercise is that the long term curve has embedded in it an inherent history of coastal onlap and offlap not driven by higher frequency oscillations of the height of sea level. Thus, if the history of coastal onlap and offlap is used to infer such shorter term oscillations, which clearly do characterize post-Oligocene times, then events associated with changes in rates must somehow be filtered out first. Because of the large uncertainties in the long term sea level curve, we are not yet in a position to derive a rigorous short term curve from coastal onlap alone.

Judging from the general similarity in form of the inherent and observed histories of coastal onlap, the Haq et al. long term curve may provide a first approximation of the history of the eustatic sea level, whereas the short term curve, which has been interpreted as reflecting changing heights of eustatic sea level, might instead more closely portray the history of changing rates of sea level rise and/or fall. Pitman's (1978) original suggestion that coastal onlap and offlap are the best and most direct recorder of rates of sea level change may thus be born out of the Haq et al. data.

Conclusion

The Haq et al. eustatic sea level curve can be used to compute the volumes of sequestered water needed to generate the difference between the long and short term curves. Comparisons of the rates of change associated with each of these curves, and comparison with known and postulated mechanisms of changing sea level, shows that only grounded ice can drive the short term curve. The computed volumes, which take into account both hypsometric and water-loading effects, imply that about once every stage since the Kimmeridgian an Antarctica-size ice sheet evolved and dissipated within an approximately 1–2 m.y. interval, and that during most of the time significant ice was persent. The Haq et al. curve implies a world in which ice, in the form of large continental ice sheets has played a significant, if not dominant role. In turn the Haq et al. sea level curve implies that long term climatic conditions have not varied significantly, being characterized by alternating glacial and interglacial intervals.

Existing geological and paleontological evidence from this same interval, and particularly from the Late Jurassic to the Eocene, suggest that large continental ice sheets did not exist then. The geological and paleontological record of high paleolatitude sites of Late Jurassic to Eocene age instead contain a record of cool to cold temperate climatic conditions, in direct contradiction of that implied by the Haq et al. sea level curve. A comparison of the retrodicted $\partial^{18}O$ of sea water with measured values for the early Cenozoic demonstrates virtually no correlation between the hindcast and measured records. This explicit test of the Haq et al. interpretation, indicates that their inversion of the coastal onlap curve to a short term eustatic sea level record is not correct.

A partial solution to this dilemma is found in Pitman's (1978) earlier suggestions that coastal onlap and offlap are generally recorders of changes in rates of sea level rise or fall, not of up and down oscillations. A theoretical coastal onlap curve computed using Pitman's equations and the Haq et al. long term curve reveals a general correspondence of the overall trends and many specific onlap events inherent in the Haq et al. long term curve with their observed coastal onlap curve. This suggests that at least significant components of the short term curve do not reflect vertical oscillations of sea level, but instead record changes in rates of sea level change. It is possible that, if the inherent signal of the long term curve were removed, only quite small amplitude short term events would be needed to account for the finer details of the observed coastal onlap curve.

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REFERENCES CITED

Baum, G. R., and Vail, P. R., 1988 Sequence stratigraphic concepts applied to Paleogene outcrops, Gulf and Atlantic basins, *in* Wilgus, C. K., et al., eds., Sea-level changes: An integrated approach, SEPM Spec. Pub. 42, p. 309–327.

Brown, H. Y. L., 1894, From Strangways to Mount Paisley: Geol. Survey South Australia Ann. Rept. Gov. Geol., 10 p.

Burton, R.; Kendall, C. G.; and Lerche, I., 1987, Out of our depth: on the impossibility of fathoming eustasy from the stratigraphic record: Earth Sci. Rev., v. 24, p. 237–277.

Christie-Blick, N.; Mountain, G. S.; and Miller, K. G., 1988, Sea level history, discussion: Science, v. 241, p. 596.

———; ———; and ———, 1990, Seismic stratigraphic record of sea level change, *in* U.S. Nat. Res. Council, Geophysics Study Comm., Sea Level Change: Washington, DC, Nat. Acad. Press, p. 116–140.

Cogley, J. G., 1985, Hypsometry of the continents: Zeitsch. Geomorph. supp. 53, 48 p.

Drewry, D. J. (ed.), 1983, Antarctica: Glaciological and Geophysical Folio: Cambridge, Scott Polar Res. Inst., 9 p.

Drinnan, A. N., and Crane, P. R., 1989, Cretaceous paleobotany and its bearing on the biogeography of austral angiosperms, *in* Taylor, T. N., and Taylor, E. L., eds., Antarctic Paleobiology: Its Role in the Reconstruction of Gondwana: New York, Springer-Verlag, p. 192–219.

Emiliani, C., and Schackleton, N.J., 1974, The Bruhnes Epoch: isotopic paleotemperatures and geochronology: Science, v. 183, p. 511–514.

Fairbanks, R. G., and Matthews, R. K., 1978, The marine oxygen isotope records in Pleistocene coral, Barbados, West Indies: Quat. Res., v. 10, 181–196.

Frakes, L. A., and Francis, J. E., 1988, A guide to Phanerozoic cold polar climates from high-latitude ice-rafting in the Cretaceous: Nature, v. 333, p. 547–549.

Francis, J. E., 1988, A 50 million year old fossil forest from Strathcona Fiord, Ellesmere Island, Arctic Canada: evidence for a warm polar climate: Arctic, v. 41, p. 314–318.

Hambrey, M. J., and Harland, W. B. (eds.), 1981, Earth's pre-Pleistocene Glacial Record: Cambridge, Cambridge Univ. Press, 1004 p.

Haq, B. U.; Hardenbol, J.; and Vail, P. R., 1987, Chronology of fluctuating sea level since the Triassic: Science, v. 235, p. 1156–1167.

———; ———; and ———, 1988*a*, Mesozoic and Cenozoic chronostratigraphy and cycles of sea-level change, *in* Wilgus, C. K., et al., eds., Sea-level changes: An integrated approach: SEPM Spec. Pub. 42, p. 71–108.

———; ———; and ———, 1988*b*, Sea level history: response: Science, v. 241, p. 601–602.

Hardenbol, J.; Vail, P. R.; and Ferrar, J., 1981, Interpreting paleoenvironments, subsidence history, and sea-level changes of passive margins from seismic and biostratigraphy, *in* Proceedings of the 26th Int. Geol. Congress, Paris: Oceanologica Acta, Supp. to v. 3, 33–44.

Harrison, C. G. A., 1986, Long-term eustasy and epeirogeny in continents: Nat. Acad. Sci. Studies in Geophysics ??, p. 111–131.

Hsu, K. J.; Ryan, W. B. F.; and Cita, M. B., 1973, Late Miocene dessication of the Mediterranean: Nature, v. 242, p. 240–244.

Kominz, M.A., 1984, Oceanic ridge volumes and sea-level change—an error analysis, *in* Schlee, J. S., ed., Interregional unconformities and hydrocarbon accumulation: AAPG Mem. 36, p. 109–127.

Lydolph, P. E., 1985, The Climate of the Earth: New Jersey, Rowman and Allanheld, 386 p.

Matthews, R. K., 1988, Sea Level history, discussion: Science, v. 241, p. 597–599.

McDonough, K. J., and Cross, T. A., 1991, Late Cretaceous sea level from a paleoshoreline: Jour. Geophys. Res., v. 96, p. 6591–6607.

Murphy, M. G., and Kennett, J. P., 1986, Development of latitudinal thermal gradients during the Oligocene: oxygen isotope evidence from the southwest Pacific, *in* Kennett, J. P., and van der Borch, C. C., eds., Init. Rept. Deep Sea Drilling Project, v. 90, p. 1347–1360.

Parkinson, C. L.; Comiso, J. C.; et al., 1987, Arctic sea ice 1973–1976: satellite passive-microwave observations: NASA SP-489, 296 p.

Pitman, W. C., 1978, Relationship between eustasy and stratigraphic sequences of passive margins: Geol. Soc. America Bull., v. 89, p. 1389–1403.

———, and Golvechenko, X., 1983, The effect of sea level change on the shelf edge and slope of passive margins, *in* Stanley, D. J., and Moore, G. T., eds., The shelfbreak: critical interface on continental margins: SEPM Spec. Pub. 33, p. 41–58.

Poore, R. Z., and Matthews, R. K., 1984, Late Eocene-Oligocene oxygen and carbon isotope record from South Atlantic Ocean, Deep Sea Drilling Project, site 522, *in* Hsu, J. J.; LeBreque, J. L.; et al., eds., Init. Rept. Deep Sea Drilling Project, v. 73, p. 725–735.

Prentice, M. L., and Matthews, R. K., 1988, Cenozoic ice-volume history: development of a composite oxygen isotope record: Geology, v. 16, p. 963–966.

Shackleton, N. J., 1986, Oxygen isotope evidence for Cenozoic climate change, *in* Brenchley, P., ed., Fossils and Climate: Chichester, UK, Wiley, p. 27–34.

———; Hall, M. A.; and Boersma, A., 1984, Oxygen and carbon isotope date from Leg 74 foraminifers, *in* Moore, T. C.; Rabinowitz, P. D.; et al., eds., Init. Rept. Deep Sea Drilling Project: v. 74, p. 599–612.

Slos, L. L., 1963, Sequences in the cratonic interior of North America: Geol. Soc. America Bull., v. 74, p. 93–114.

Spicer, R. A., 1989, Reconstructing high-latitude Cretaceous vegetation and climate: Arctic and Antarctic compared, *in* Taylor, T. N., and Taylor, E. L., eds., Antarctic Paleobiology: Its Role in the Reconstruction of Gondwana: New York, Springer-Verlag, p. 27–36.

———, and Parrish, J. T., 1986, Paleobotanical evidence for cool north polar climates in the mid-Cretaceous (Albian-Cenomanian): Geology, v. 14, p. 703–706.

Times Atlas of the World, 1988, Times Atlas of the World (Seventh Comp. ed.): London, Times Brooks, 228 p.

Truswell, E. M., 1983, Recycled Cretaceous and Tertiary pollen and spores in Antarctic marine sediments: a catalogue: Palaeont. Abt. B., v. 186, p. 121–174.

———, 1989, Cretaceous and Tertiary vegetation of Antarctica: a palynological perspective, *in* Taylor, T. N., and Taylor, E. L., eds., Antarctic Paleobiology: Its role in the Reconstruction of Gondwana: New York: Springer-Verlag, p. 71–88.

Vail, P. R.; Mitchum, R. M., Jr.; and Thompson, S., III, 1977, Seismic stratigraphy and global changes in sea level, *in* Payton, C. E., ed., Seismic stratigraphy—applications to hydrocarbon exploration: AAPG Mem. 26, p. 83–97.

———, and Haq, B. U., 1988, Sea level history, response: Science, v. 241, p. 599.

Watts, A. B., and Steckler, M. S., 1979, Subsidence and eustasy at the continental margin of eastern North America, *in* Talwani, M., et al., eds. Deep drilling results in the Atlantic ocean: continental margins and environment: Am. Geophys. Union Ewing Ser. 3, p. 218–234.

Webb, P. N.; Harwood, D. M.; McKelvey, B. C.; Mercer, J. H.; and Stott, L.D., 1984. Cenozoic marine sedimentation and ice-volume variation on the east Antarctic craton: Geology, v. 12, p. 287–291.

Wilgus, C. K.; Hastings, B. S.; Kendall, C. G. S.; Posamentrer, H. W.; Ross, C. A.; and van Wagoner, J. C. (eds.), 1988, Sea-level changes: An integrated approach: SEPM Spec. Pub. 42, 407 p.

Williams, D. F., 1988, Evidence for and against sea level changes from stable isotopic record of the Cenozoic, *in* Wilgus, C. K., et al., eds., Sea-level changes: An integrated approach: SEPM Spec. Pub. 42, p. 31–36.

Wise, D. U., 1974, Continental margins, freeboard, and the volumes of continents and oceans through time, *in* Burk, C. A., and Drake, C. L., eds., The Geology of Continental Margins: New York, Springer-Verlag, p. 45–58.

Woolnough, W. G., and David, T. W. E., 1926, Cretaceous glaciation in central Australia: Quat. Jour. Geol. Society London, v. 82, p. 332–351.

Zwally, J. J.; Comiso, J. C.; et al., 1983, Antarctic sea ice was 1973–1976: satellite passive-microwave observations: NASA SP-459, 206 p.