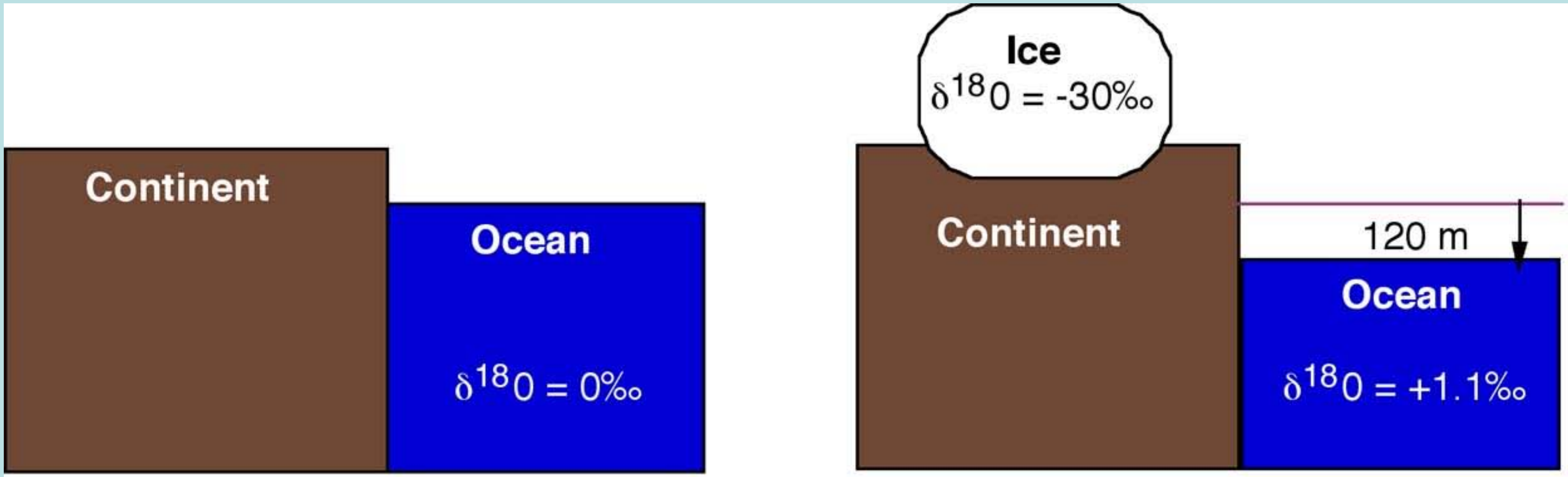
12.740 Paleoceanography
Spring 2008

Coastline evidence for sea level change

12.740 Lecture 3 Spring 2008

Effect of glaciation on the oxygen isotope composition of the ocean



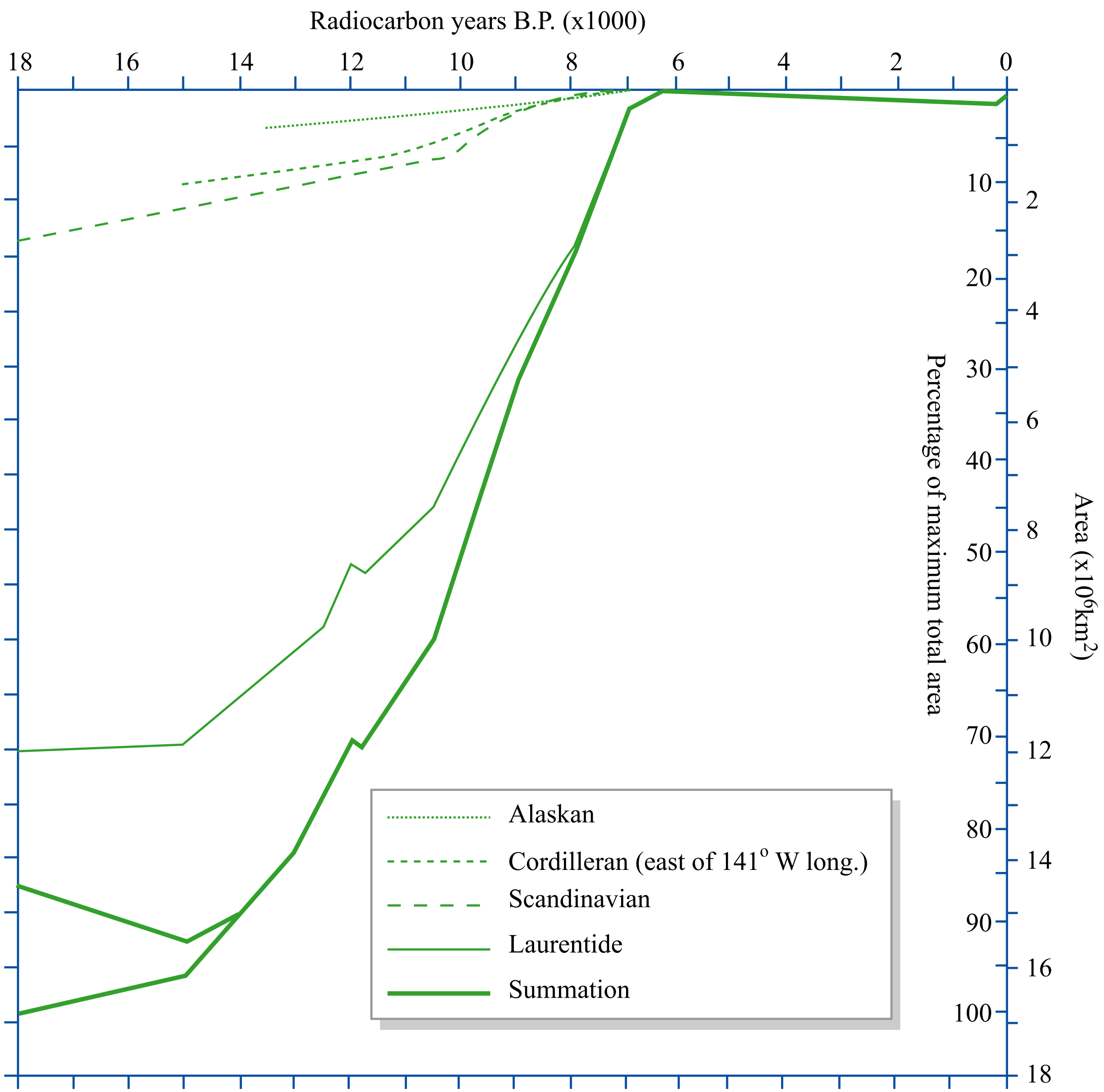
Modern Ocean

Last Glacial Maximum
(18,000 years Before Present)

Isotope Mass Balance Equation:

$$M_o\delta_o + M_i\delta_i = M_t\delta_t$$

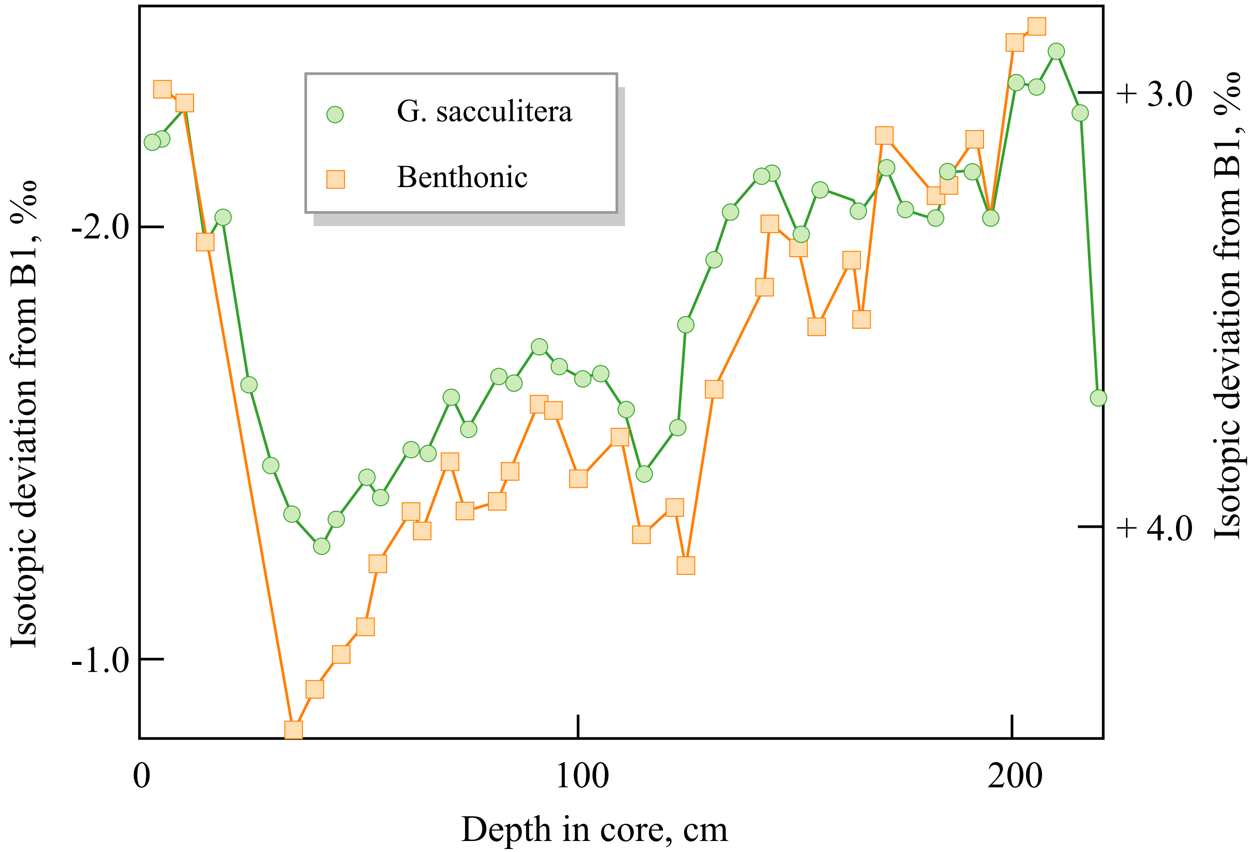
Area of ice sheets vs time (Bloom, 1971)



Areas of major ice sheets that melted in the last 18,000 years.

Figure by MIT OpenCourseWare.

Shackleton and Opdyke (1971) western tropical Pacific planktonic-benthic $\delta^{18}O$ comparison



Comparison of planktonic and benthonic oxygen isotopic record from eore V28-238. The two sequences are plotted to the same scale of isotopic change, but with scale zero-points differing by 5.3‰, the present-day planktonic-benthonic difference.

Figure by MIT OpenCourseWare.

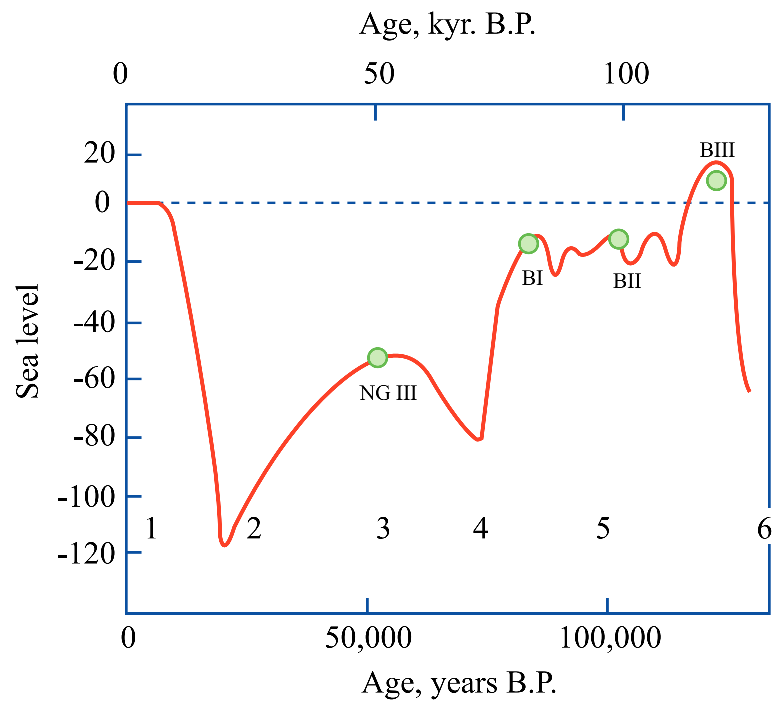
Estimates for sea level 30,000 years ago (as of 1980)

Reference	Sea Level Estimate
Curray (1965)	-15 m
Milliman and Emery (1968)	-10 m
Emery (1971)	-60 m
Morner (1971)	-15 m
Shackleton and Opdyke (1973)	-80 m
Chappel (1974)	-50 m
Weyer (1978)	0 m
Chappel and Veeh (1978)	-40 m
Geyh et. al. (1979)	-30 to -70 m
Blackwelder et. al. (1979)	-20 m

Estimates for sea level 30,000 years ago

Figure by MIT OpenCourseWare.

Shackleton and Opdyke (1971) sea level comparison



Glacio-custatic sea level curve for the past 130,000 yr derived from oxygen isotopic measurements in V28-238, compared with eastimated sea levels derived from work on Barbados (BI, BII, BIII) and New Guinea (NG III).

Figure by MIT OpenCourseWare.

Insert slides on continental shelf grab samples, C14 dating

Texas shelf sea level from calcareous fossils

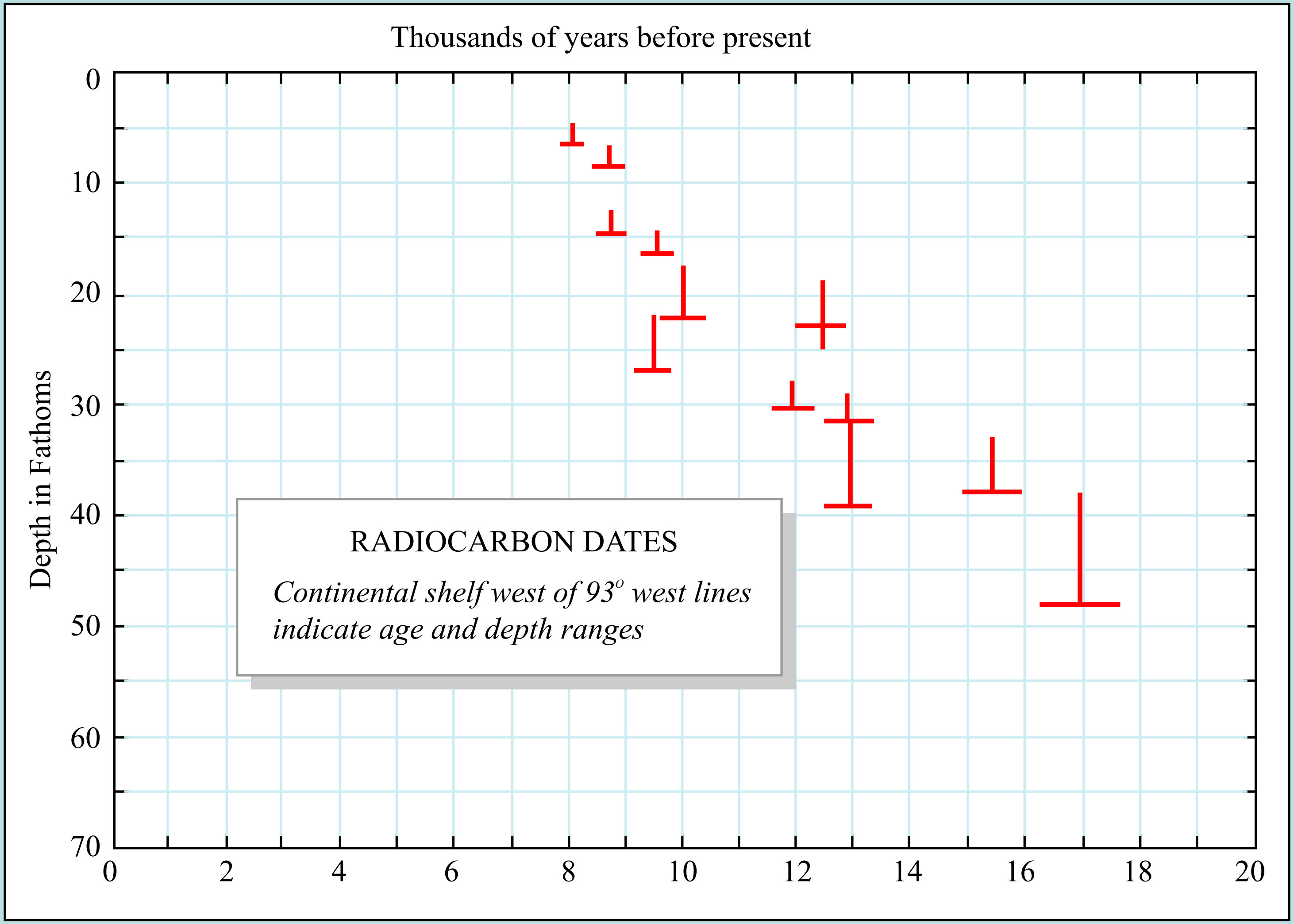
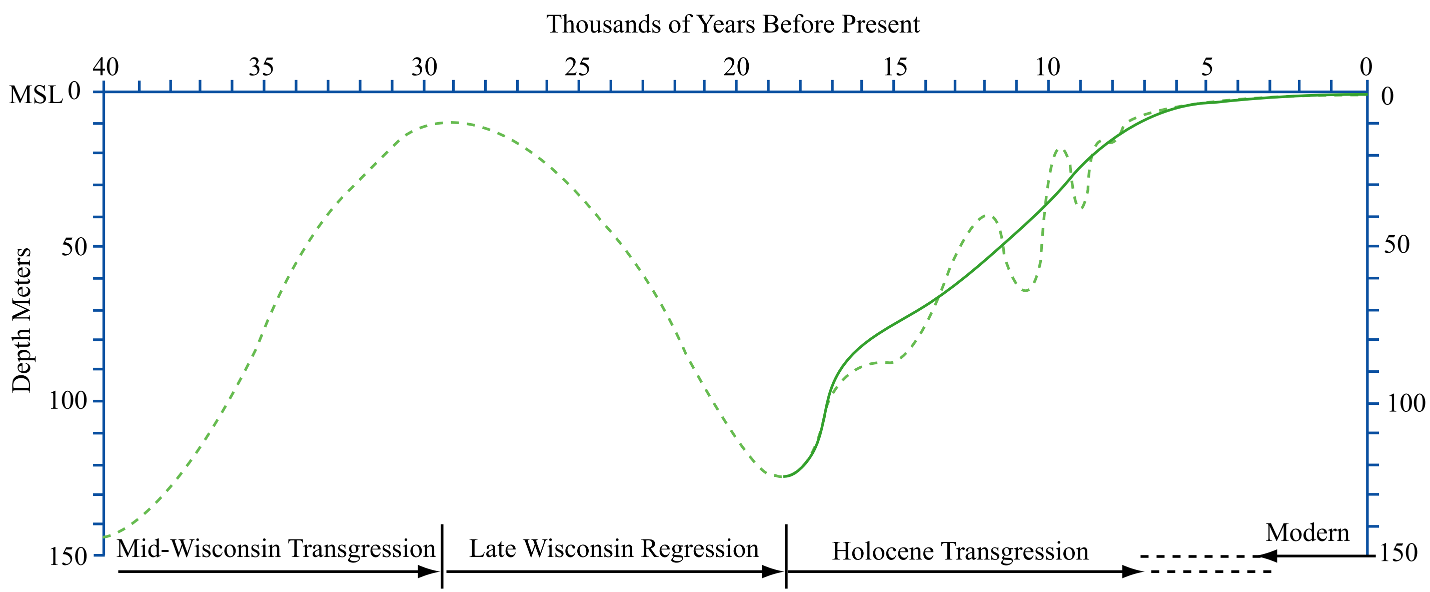


Figure by MIT OpenCourseWare.

Curray sea level curve



Late-Quaternary fluctuations of sea level, from compilation of published and unpublished radiocarbon dates and other geologic evidence. Dotted curve estimated from minimal data. Solid curves shows approximate mean of dates compiled. Dashed curve slightly modified from Curray (1960, 1961). Probable fluctuations since 5,000 B.P. not shown here.

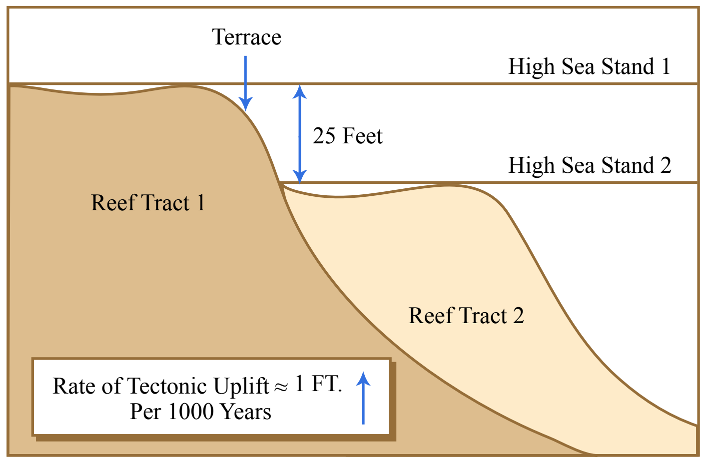
Figure by MIT OpenCourseWare.

TABLE III. DATES REJECTED FOR SEA LEVEL CURVE

Station	*Depth (fathoms)*	*Shell*	*Date*	*Reason for Rejection*
J-114 27°59.2′ 96°39.5′	11½	*Chama congregata*	<130	Thought previously to be shallow subtropical.
J-370 28°06.4′ 95°46.6′	21¾	*Aequipecten i. concentricus*	<140	Depth range greater than previously believed or possibly misidentified.
J-468c 26°46.0′ 96°42.0′	39	*Crassostrea rhizophorae*	3,200 ± 100	Possibly fortuitous aberrant population, or possibly transported.
J-468a 26°46.0′ 96°42.0′	39	*Crassostrea virginica?*	3,690 ± 170	Possibly *Crassostrea rhizophorae*.
J-482*a* 26°34.2′ 97°05.3′	14½	*Ostrea equestris* *Crassostrea virginica*	4,280 ± 175	*Ostrea* depth range greater than previously believed.
J-654 28°00.8′ 93°07.2′	47–65	*Ostrea equestris* *Trigonicardia media* *Turritella acropora* *Strombus alatus* *Oliva sayana*	6,880 ± 250	Living ranges too deep to be of value in determining a minimum age for deposit.
J-383 28°50.4′ 95°08.0′	8½	*Crassostrea virginica*	26,900 ± 1,800	Sample of beach rock and coquina from Freeport rock. Slightly recrystallized so should be older, probably interglacial or interstadial.
J-526 28°41.2′ 95°35.4′	8	*Rangia cuneata* *Crassostrea virginica*	32,500 ± 3,500	To the west and in line with Freeport rocks. Interglacial or interstadial.

Curray data table (…groan!)

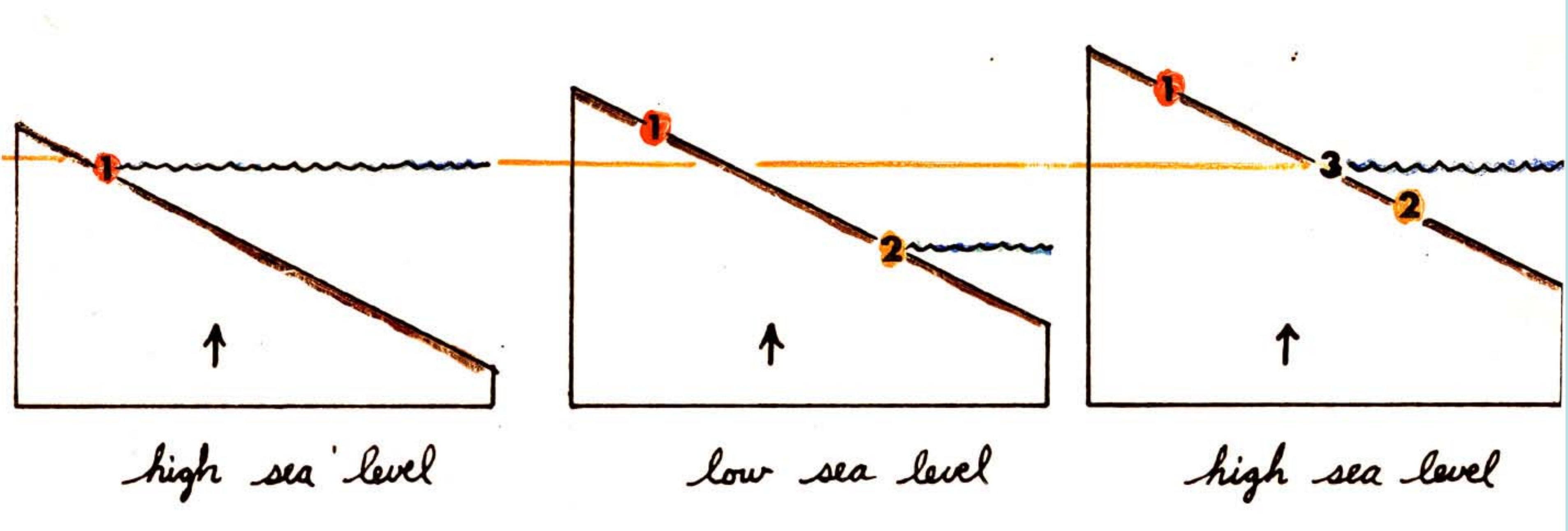
Model for Barbados coral terrace formation



Model for reef terrace formation. In this hypothetical example, reef tract no. 1 forms around Barbados during interglacial high sea stand no. 1. Emergence of reef tract no.1 takes place during an interval of lowered sea level which has a duration of 25,000 years. Interglacial high sea stand no.2 attains the same worldwide level as high sea stand no.1, but, relative to Barbados, high sea stand no.2 appears to be 25 feet lower, due to gradual tectonic uplift of the island at about 1 foot per 1,000 years. Therefore, the record of the eustatic oscillation is the addition of a 25-foot reef terrace to the coral cap.

Figure by MIT OpenCourseWare. Adapted from source: Mesolella et al. (1969).
D. Broecker et al. (1968) dated 3 Barbados terrace complexes.

Terrace formation on a coastline with changing sea level



$^{230}Th/U$ dating of corals

^{238}U ----------> ^{234}U --------> ^{230}Th ------->

$t_{1/2}$: 4.5×10^9 yrs 248,000 yrs 75,200 yrs

^{235}U -------> ^{231}Pa ----->

0.7×10^9 yrs 32,800 yrs

U (VI) is relatively soluble in seawater [carbonato complexes: e.g. $UO_2^{2+}CO_3^{2-}$; $UO2^{2+}(CO_3^{2-})_2$], occurs at about 13 nmol/kg (2.3 dpm/kg), and appears to be conservative in the ocean. ^{230}Th and ^{231}Pa are particle-reactive; i.e. it tends to attach to surfaces rapidly, and so it is removed from seawater on a time scale of ~30 years. Hence it occurs at fairly low concentrations in seawater (<0.1 dpm/100kg at the surface; ~1 dpm/100kg in deep waters).

Corals incorporate uranium (~2ppm) but very little ^{230}Th.

Assumptions for ^{230}Th dating of corals

1. Closed system : except for radio-decay and production, no uranium or thorium enters or leaves the object.

2. $^{230}Th_{initial}$ = O.

3. Initial $^{234}U/^{238}U$ activity ratio = 1.15 [this is the ratio observed in seawater; disequilibrium results from alpha-recoil damage in continental rocks and subsequently higher weathering rate of ^{234}U; we assume that this ratio has been stable in the past – but is this justified?

4. Then:

$$\frac{\partial^{234}U}{\partial t} = \lambda_{^{238}U}\,{}^{238}U - \lambda_{^{234}U}\,{}^{234}U$$

$$\frac{\partial^{230}Th}{\partial t} = \lambda_{^{234}U}\,{}^{234}U - \lambda_{^{230}Th}\,{}^{230}Th$$

where the radionuclides are expressed in <u>concentration units</u> (atoms per g); to convert to activity units (dpm/g), must convert each term:

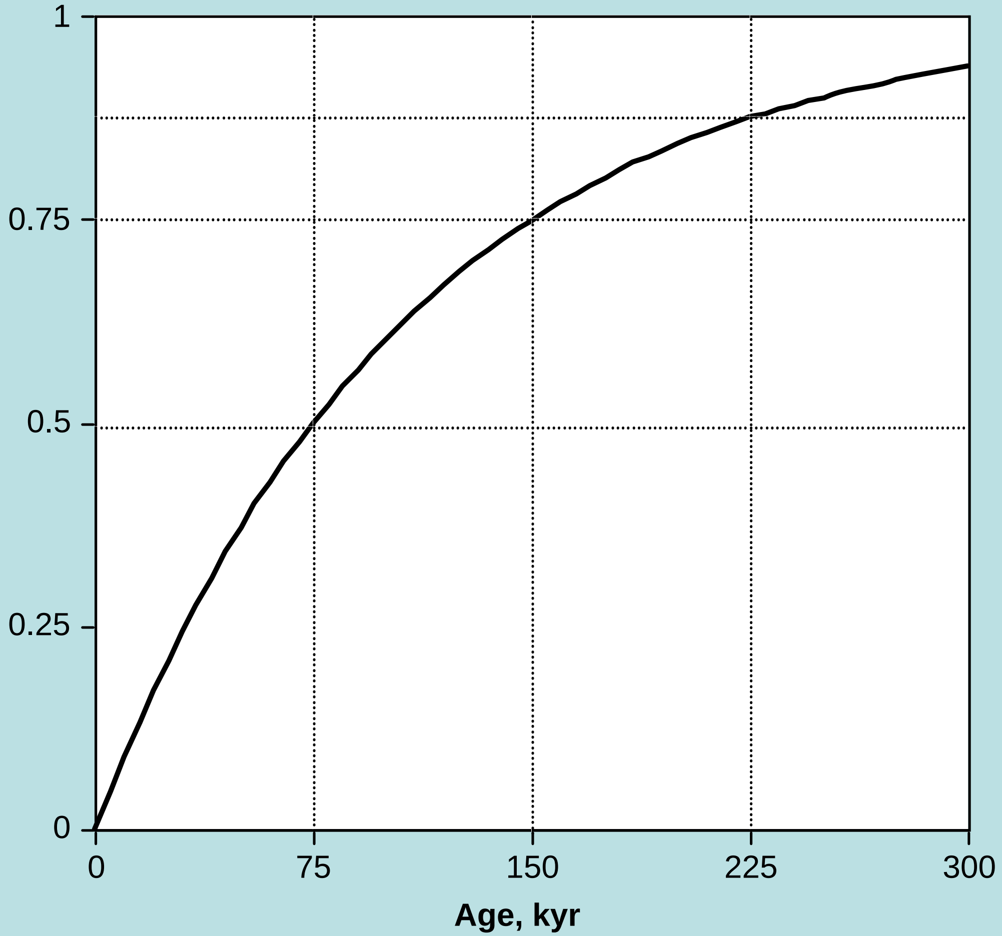
$$^{230}Th = \frac{A_{^{230}Th}}{\lambda_{^{230}Th}}$$

Hence:

$$\frac{\partial A_{^{234}U}}{\partial t} = \lambda_{^{234}U} A_{^{238}U} - \lambda_{^{234}U} A_{^{234}U}$$

$$\frac{\partial A_{^{230}Th}}{\partial t} = \lambda_{^{230}Th} A_{^{234}U} - \lambda_{^{230}Th} A_{^{230}Th}$$

Solutions: Assume ^{238}U activity is constant (because half-life is 4.5E+9 yr), solve ^{234}U equation first, then plug into ^{230}Th equation) (left as an exercise for the reader!). If we start out with a system where $A_{234U}/A_{238U} = 1$, the solution is simple:



this solution can be derived graphically by noting that every ^{230}Th half-life, the $^{230}Th/^{234}U$ ratio approaches 50% closer to 1.00

Because $^{234}U/^{238}U$ is initially above the radioactive equilibrium level, it is necessary to correct for the decay of the initial excess. The complete equations for calculation of ^{230}Th ages are:

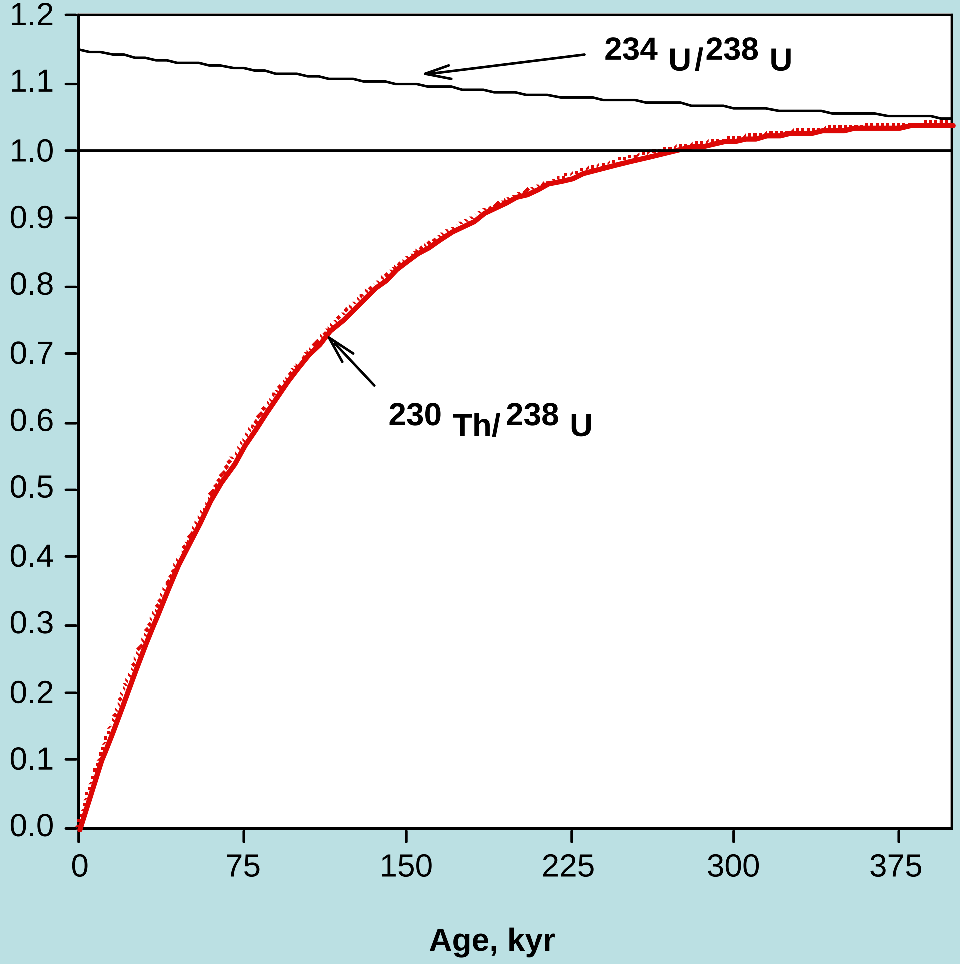
$$\frac{A_{^{230}Th}}{A_{^{238}U}} = 1 - e^{-\lambda_{230_{Th}} t} + \left(\frac{\delta^{234}U(t)}{1000}\right)\left(\frac{\lambda_{230_{Th}}}{\lambda_{230_{Th}} - \lambda_{234_U}}\right)\left(1 - e^{\left(\lambda_{234_U} - \lambda_{230_{Th}}\right)t}\right)$$

$$\delta^{234}U(t) = \delta^{234}U(t=0)e^{-\lambda_{234_U} t}$$

$$\delta^{234}U = \left(\frac{\left({}^{234}U/{}^{238}U\right)}{\left({}^{234}U/{}^{238}U\right)_{eq}} - 1\right) * 1000$$

t is the age of the sample (kyr before present)

Full solution of ^{230}Th/U dating method



Measurement: alpha counting

- Th, U are separated by ion chromatography, then put onto a metal surface as a thin film
- When the atoms decay, they emit alpha particles of defined energy
- The alpha particles are detected by a crystal scintillator, where the amount of light emitted (pulse height) is proportional to the energy of the alpha particle.
- Counting statistics: square root of n

Applications of the ^{230}Th dating of corals: alpha-counting era

- Veeh (1966) found 120 kyr bp high stand on Bermuda:
  +5 ± 1 m relative to present.

- **125** ± 4 kyr high stand found to be present on many oceanic islands.

- Barbados terraces dated by Broecker and Ku (1968)

1. Terrace III (40 m above present sea level) gave ages:

   128 ± 6 ka

   120 ± 6

   124 ± 6

   124 ± 6

   120 ± 6

   Average: 123, $\sigma = 3$

   If we assume that this terrace was 5 m above present sea level when it was formed, it implies that Barbados is being uplifted $^1/_3$ m kyr^{-1}

Applications of the ^{230}Th dating of corals: alpha-counting era

- Two other terraces were dated as well (assuming constant uplift) :

-13 m	**-10 m**
83 ± 4	103 ± 2
82 ± 2	103 ± 4
81 ± 4	
82 ± 4	

- And one more problematical date at 111 ± 9
- Broecker proposed that these terraces should be coeval with the oxygen isotope 5a-5c-5e record, and thus inferred by extrapolation in a sediment core that Termination II was at 127 ka
- A few years later, Shackleton and Opdyke (1971) placed the O18 record in a sediment core with the Brunhes-Matuyama magnetic reversal; interpolation between the core top and the reversal gave an age scale consistent with the Broecker age scale.
- More recently, Edwards Chen and Wasserburg (1986) developed mass spectrometric measurement of ^{230}Th, ^{234}U, and ^{238}U that was more sensitive and more precise than alpha counting. Redating the 5e terrace, they obtained ages of

  122.1 ± 1.1 , 122.7 ± 1.3, and 124.5 ± 1.

  Broecker was very happy (his average age was 123)!

- Older stands (correlated to MIS 7) were dated as well (through less precisely).

Mass spectrometric measurement of U, Th

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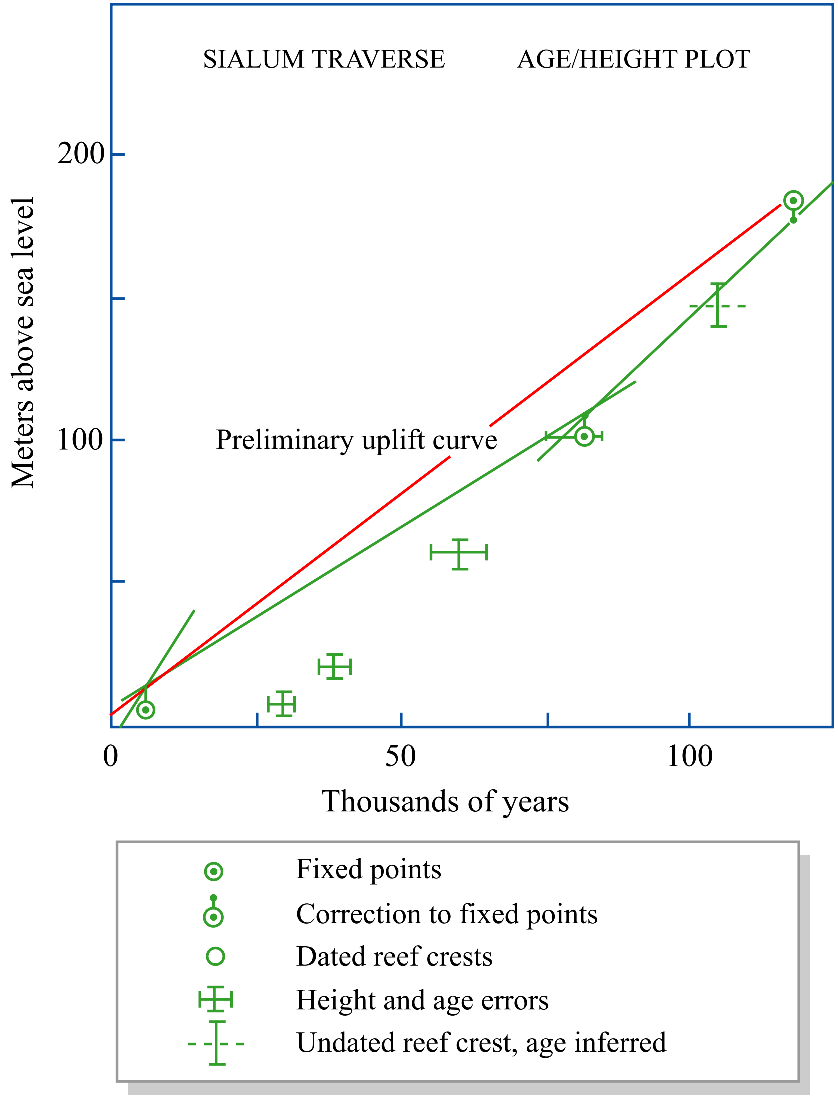
Edwards, Chen, & Wasserburg (1986/7) EPSL 81:175)

New Guinea elevations



First steps in disentangling tectonic movements from sea-level movements. Placement of fixed points and location of errors for a single section.

Figure by MIT OpenCourseWare.

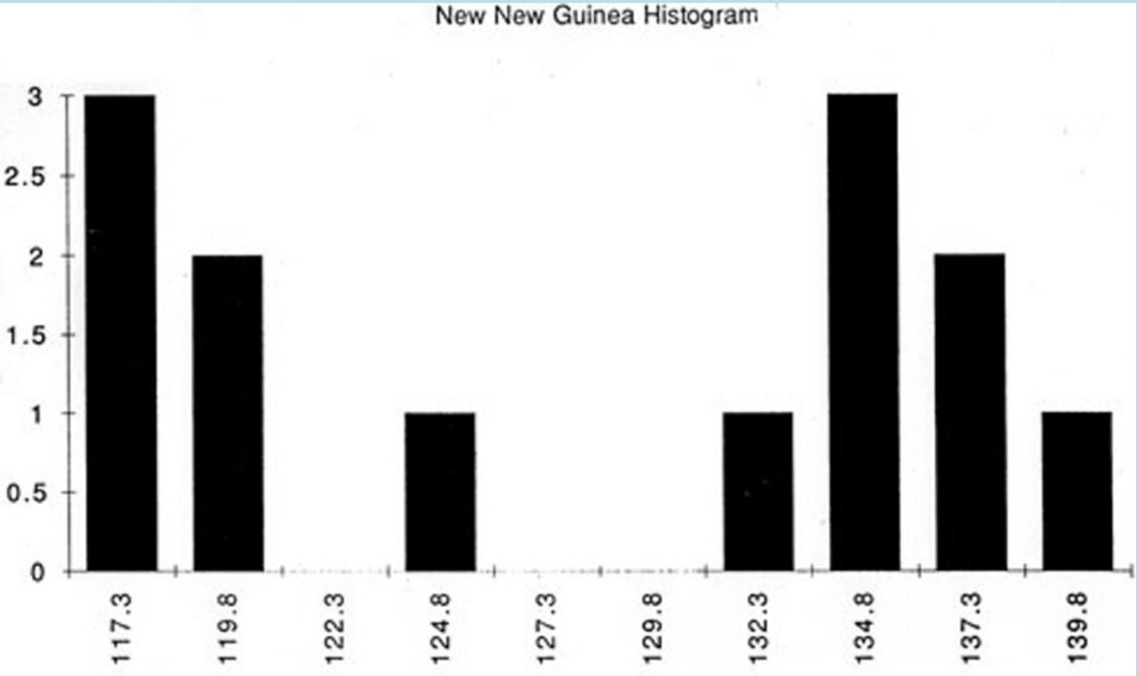
New Guinea ^{230}Th/U ages

Elevations (two-stage uplift model):

-13 m (assumed)	**-15 m**	**+5 m** (assumed)
84 ± 4	107 ± 9	142 ± 8
86 ± 4	107 ± 6	116 ± 7
		119 ± 7
		140 ± 10
		133 ± 10

More recent mass spec dates for New Guinea terraces:

Edwards, Chen, and Wasserburg (1986) have dated two New Hebrides coral samples (a "5e doublet" reef complex) that had been dated at 141+16 kyr by alpha counting. The new mass spec. ages were 129.9+1.1 and 125.5+1.3 kyr. While these are a bit older than the Barbados dates, they are not so deviant as the 141 kyr date made them seem to be. More recent measurements indicate that many other samples give ages near 134 kyr (Stein et al., 1993).



Larry Edwards (personal communication) argues that the New Guinea terraces are particularly susceptible to diagenetic alteration because of the high rainfall, and that he prefers to disbelieve the New Guinea dates and accept the Barbados dates.

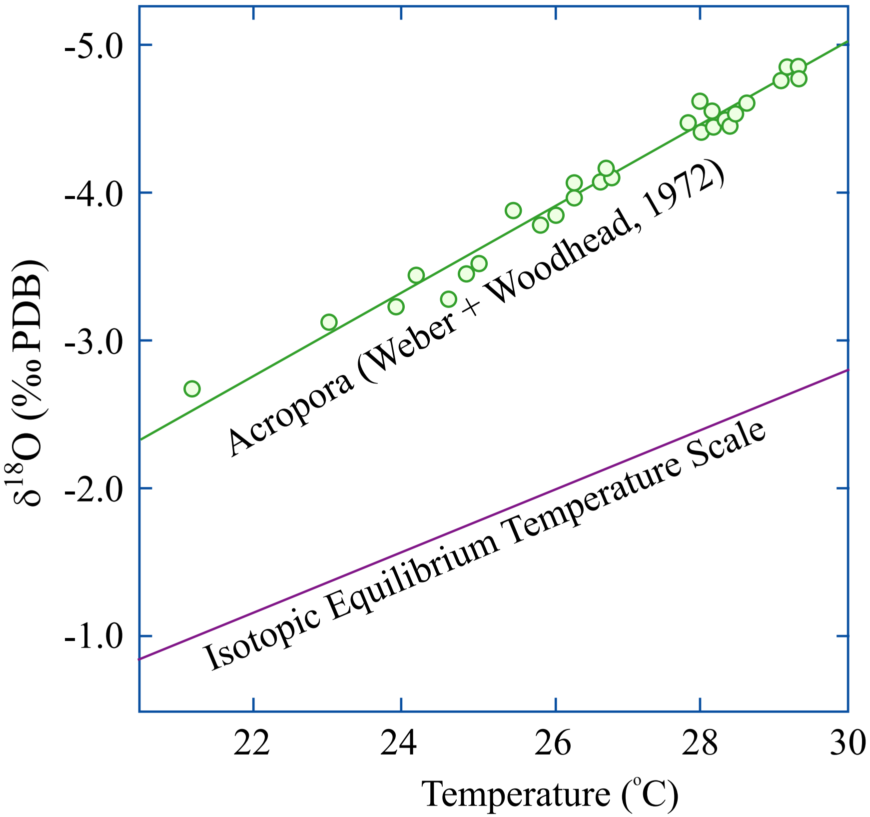
Comparison of MIS 5 high stands from different emerging islands

AGE AND SEA-LEVEL ESTIMATES FROM RISING-ISLAND CORAL TERRACES

BERMUDA (STABLE): +5(±1) m, 125±4 KYR BP

BARBADOS (BROECKER, MATTHEWS, KU)				
	82 ±4	106 ± 5	123 ± 6	KYR B.P.
NEW GUINEA (BLOOM, CHAPPELL, BROECKER)				
	85 ±4	107 ± 6	130 ± 8	"
HAITI (DODGE, FAIRBANKS, BENNINGER)				
	81 ± 3	108 ± 5	130 ± 6	"
SEA LEVEL:	−18 m	−15 m	(+5) m	

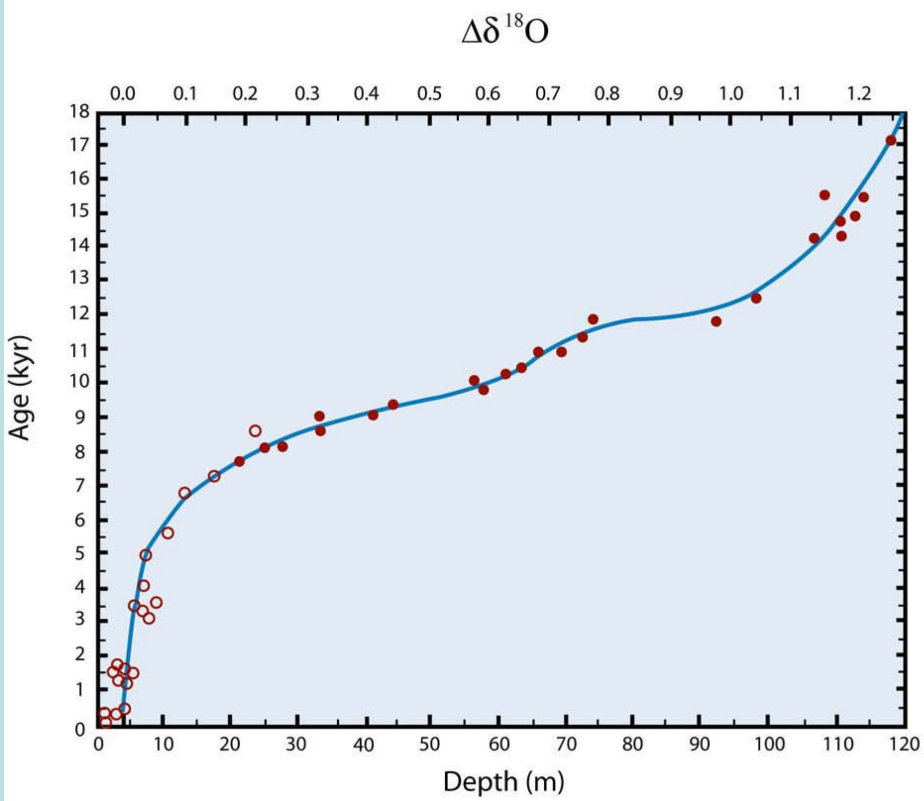
Coral δ^{18}O-T relationship



The mean $\delta^{18}O$ values for Acropora from 0-6m depths plotted against mean annual temperature of ocean water for 27 localities (835 specimens) from around the world. The lower curve is isotopic equilibrium temperature scale for the calcite water system as determined by McCrea (1950) and Epstein et. al. (1953) (adapted from Weber and Woodhead, 1972).

Figure by MIT OpenCourseWare.

Sea level over the past 20,000 years: submerged Barbados drill samples

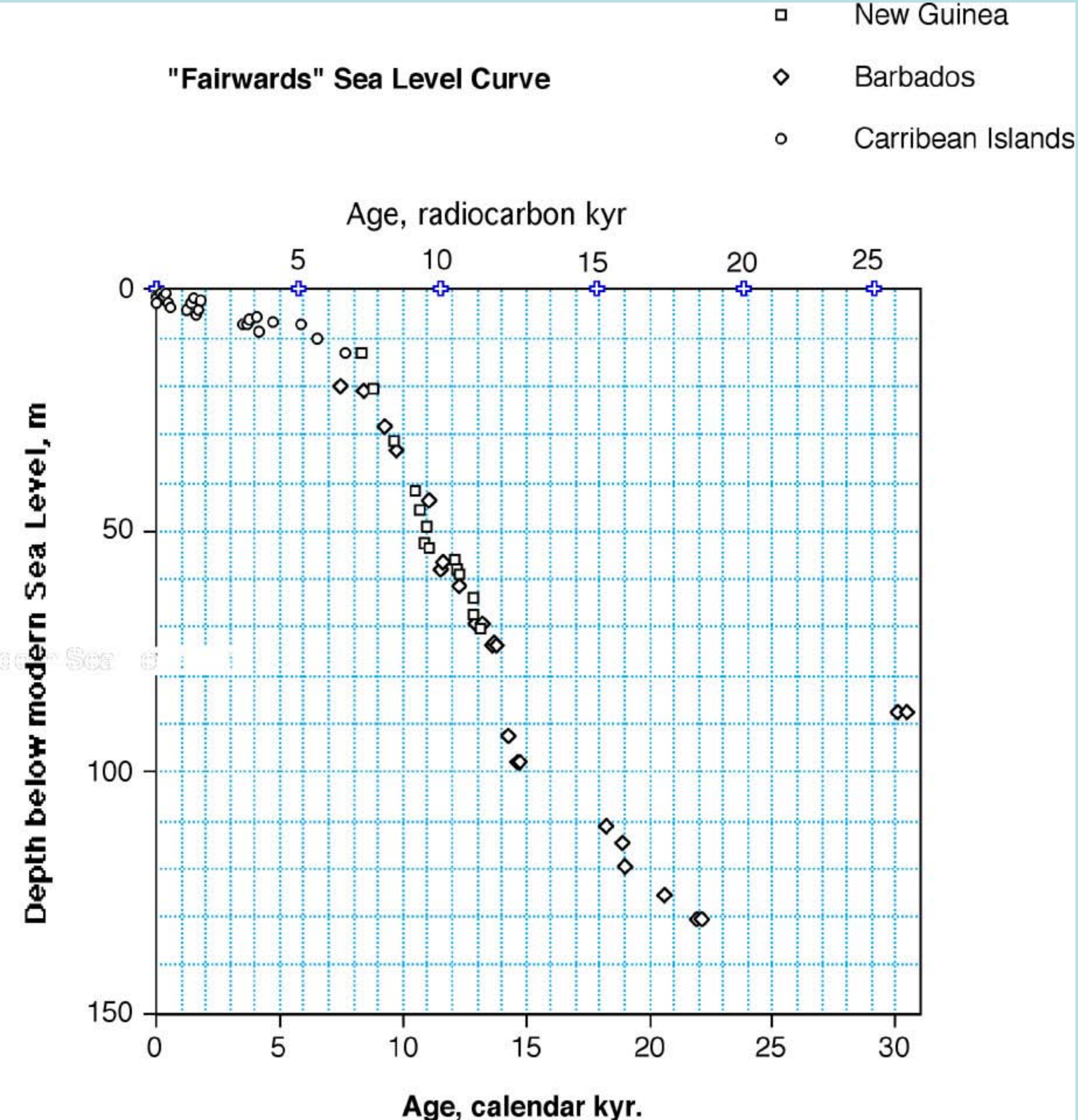


Barbados sea level curve based on radiocarbon-dated *A. palmata* (filled circles) compared with *A. palmata* age-depth data[1] (open circles) for four other Caribbean island locations. All radiocarbon ages in this figure are corrected for local seawater $\Delta^{14}C$ by subtracting 400 yr from the measured radiocarbon ages[8] but they are not corrected for secular changes in atmospheric ^{14}C levels[15]. The Barbados data are corrected for the estimated mean uplift of 34 cm kyr^{-1}. The top axis of the Barbados sea level curve (solid line) is scaled to the estimated $\delta^{18}O$ change of mean ocean water using the calibration in ref. 5.

Note steps between core breaks (MWPIa and Ib)

Figure by MIT OpenCourseWare. Adapted from source: Fairbanks et al. (1989).

Using ^{230}Th/U dating (Bard et al. 1990) and adding in data from submerged New Guinea corals (Edwards et al. 1993)



Note difference between ^{230}Th/U and ^{14}C time scales

Adding in Tahiti (Bard et al., 1990)

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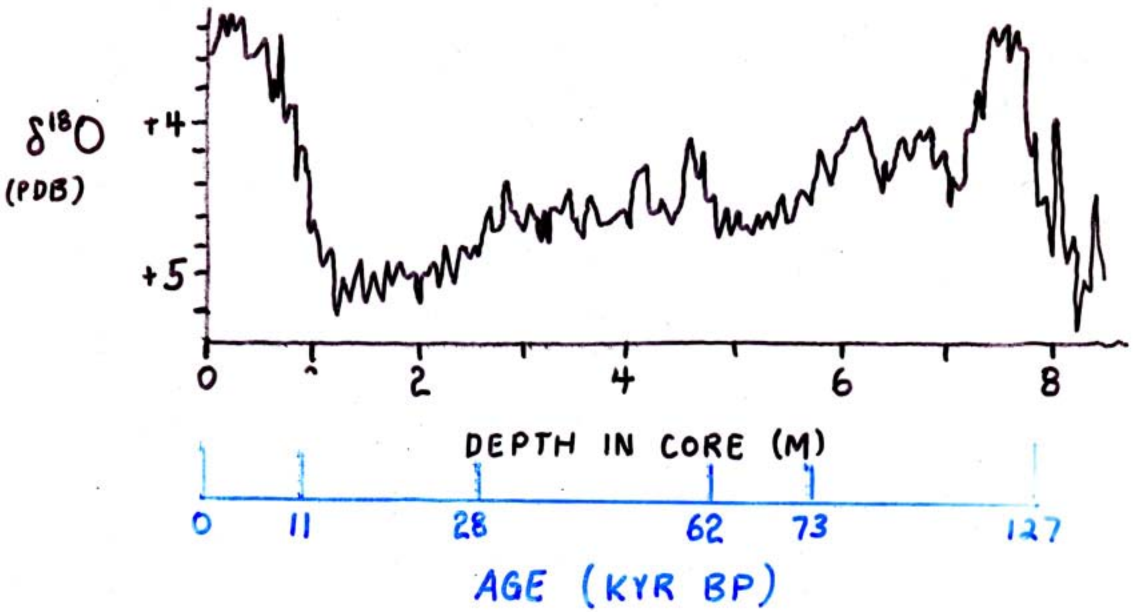
NATURE - VOL 382 - 18 JULY 1996

V19-30 benthic $\delta^{18}O$

SHACKLETON, IMBRIE, AND HALL (1983)



IF we assume glacial ice ≈ -35‰ and
temperature effect is negligible, then ΔS.L. ~ -180 m

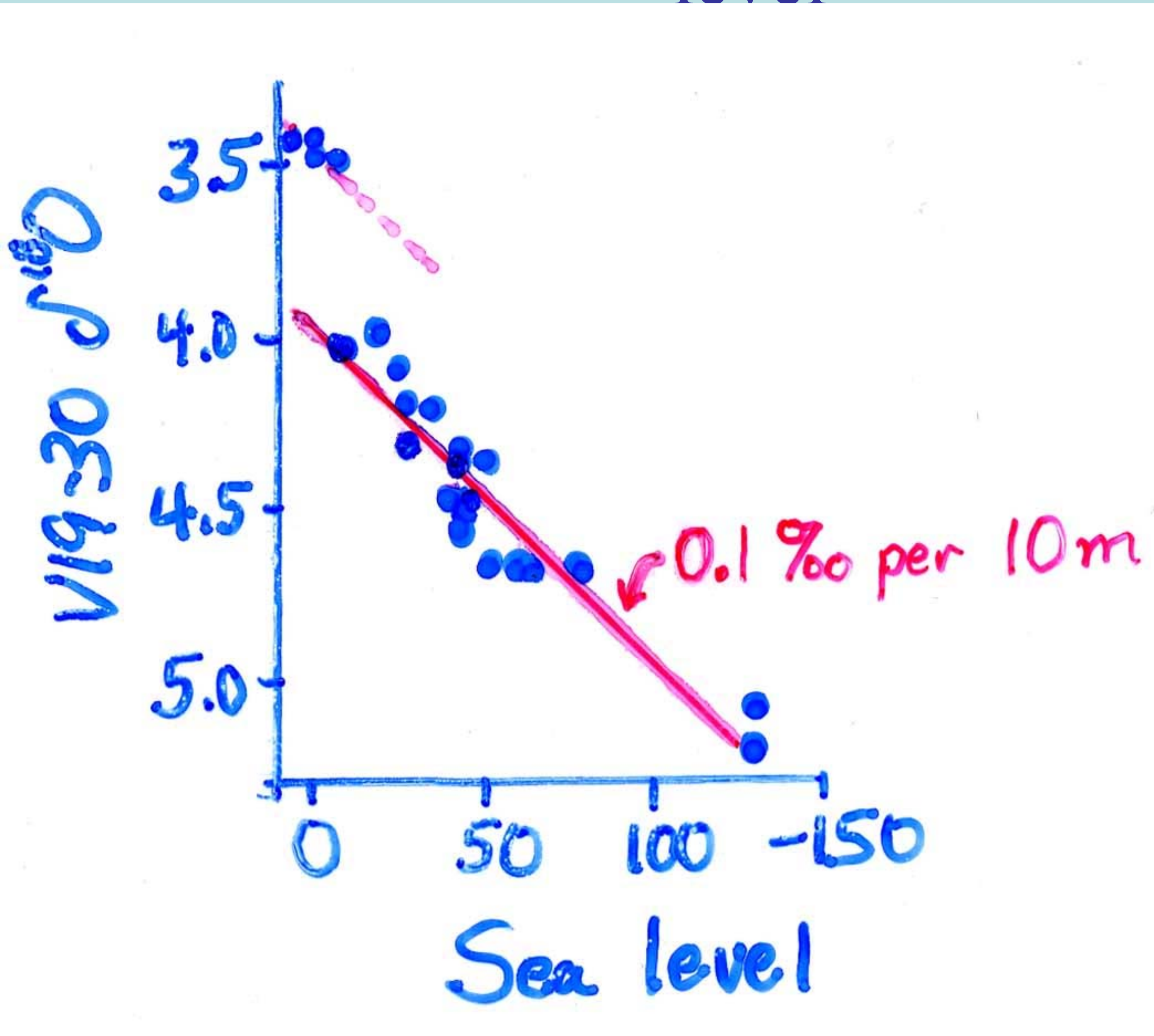
Reconciliation of sea level and $\delta^{18}O$: temperature <u>and</u> ice volume

• If we assume that the discrepancy between $\delta^{18}O$ and the various sealevel indicators is due to the effect of temperature on ice volume, then we can reconstruct paleo-deep-temperature histories from benthic $\delta^{18}O$, which implies that the deep ocean was colder by anywhere from 1.5°C (Pacific) to 3.0°C (Atlantic).

• Most of this cooling appears to have taken place going from oxygen isotope stages 5e to 5d, and the greatest warming going from stage 2 and 1. The rest of the time the temperature variations have been more subtle; i.e., the deep ocean must generally have been about 2°C colder than it is at present. This is known because the uplifted coral terraces suggest that sea level change between stages (5a,5c) and 5e was only 10-20m, but benthic $\delta^{18}O$ would have a sea-level change equivalent to more than 40m. So even though the sea level was not very much different during stages 5a,b and 5e, the deep sea temperature was about 2°colder.

• Recently Chappell et al. (1996) have shown that the original New Guinea sea level estimates were flawed by errors in $^{230}Th/U$ ages (between 30-70 ka) and elevation estimations (all previous dates were too young, so uplift was underestimated). Note that the "sea level from $\delta^{18}O$" curve is based on taking the "smoothed" difference between the planktonic RC17-177 record (Shackleton) and the benthic V19-30 record, and determining the difference; i.e., this $\delta^{18}O$ based sea level record assumes that there is no temperature (or local salinity) temporal signal in core RC17-177.

Benthic $\delta^{18}O$ vs New Guinea sea level



Chappell and Shackleton, (1986) Nature

Revised New Guinea sea level estimates compared to (modified) $\delta^{18}O$

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Previous

"Previous" version was based on ages that were uniformly too young - so uplift was underestimated.

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Revised

Fig. 1. Upper: Previous estimates of sea levels for the last 140 ka. v s sea levels deduced from coral terraces at Huon Peninsula ꟽ ⨯ I s estimates based on combined benthic and planktic deep sea core $\delta^{18}O$ data ꟽ ⨯Lower: v s new results between 30 and 75 ka this paper . and previous results 0–20 ka and 75–145 ka from HP; ' s undated sea levels based on lowstand deposits in the raised reef and fan-delta tracts at HP cf. ꟽ3,25 ⨯ I s isotopic sea levels as above.

Chappel et al. (1996) EPSL 141:227

New Guinea high stand sea levels compared to benthic $\delta^{18}O$

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New Guinea high stand sea levels compared to planktonic $\delta^{18}O$

The “poetic” Norwegian Sea $\delta^{18}O$ stack

If there was a site where we knew that temperature did not change, then $\delta^{18}O$ from this site could be used to assess ice volume $\delta^{18}O$ and relative temperature change at other sites. Unfortunately, it is difficult to know that temperature did not change somewhere. One constraint occurs in places where the potential temperature is near the freezing point of seawater. It can't get any colder (although it might have become warmer). It has been argued on various grounds (Labeyrie et al., 1987) that except for the most severe glaciations, NADW formation occurs in the Norwegian/Greenland Sea. In that case, the water must have been near the freezing point (as it is today), so that the temperature must have remained constant. A somewhat poetic "stack" of several N/G sea records was reconstructed (and incorporated with the oxygen isotope and sea level constraints for the last glacial maximum) to obtain a continuous estimate for the mean oxygen isotope composition of seawater. (the poetry involves how the cores were stacked, as the usual 5a-5b-5c-5d-5e sequence was not clearly evident in most of the records).

Direct Determination of the Timing of Sea Level Change During Termination II

Christina D. Gallup,[1*] H. Cheng,[2] F. W. Taylor,[3] R. L. Edwards[2]

An outcrop within the last interglacial terrace on Barbados contains corals that grew during the penultimate deglaciation, or Termination II. We used combined ^{230}Th and ^{231}Pa dating to determine that they grew 135.8 $\pm$ 0.8 thousand years ago, indicating that sea level was 18 $\pm$ 3 meters below present sea level at the time. This suggests that sea level had risen to within 20% of its peak last-interglacial value by 136 thousand years ago, in conflict with Milankovitch theory predictions. Orbital forcing may have played a role in the deglaciation, as may have isostatic adjustments due to large ice sheets. Other corals in the same outcrop grew during oxygen isotope (δ^{18}O) substage 6e, indicating that sea level was 38 $\pm$ 5 meters below present sea level, about 168.0 thousand years ago. When compared to the δ^{18}O signal in the benthic V19-30/V19-28 record at that time, the coral data extend to the previous glacial cycle the conclusion that deep-water temperatures were colder during glacial periods.

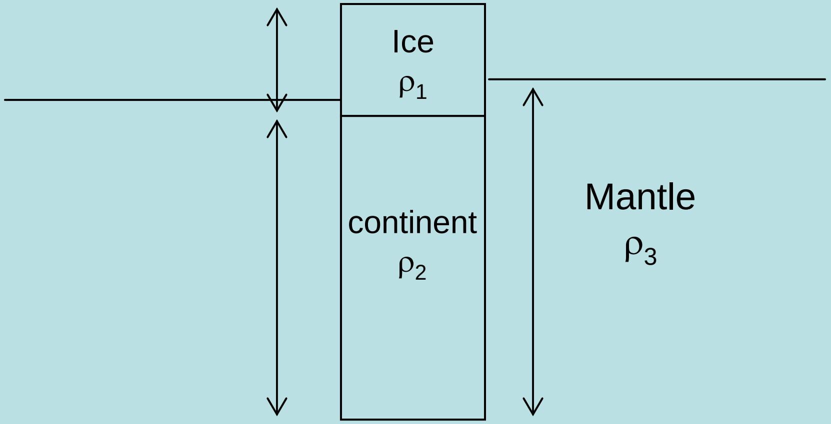
Sea level, isostasy, and the viscoelastic earth

Eustatic vs. Relative Sea level changes

Relative sea level is what you measure: the position where the sea meets the land (relative to some benchmark). Relative sea level includes a component of the vertical movements of the land.

Eustatic sea level is the level of the ocean corrected for the vertical movements of the land – and generally assumed to be the same everywhere on earth (except for some corrections – see later).

Isostasy: As major ice sheets build up on the earth's surface, the solid earth below sinks to maintain isostatic equilibrium: in effect, the (continental block + ice sheet) is floating in the mantle:



Isostatic equilibrium is: $\rho_1 \mathbf{z}_1 + \rho_2 \mathbf{z}_2 = \rho_3 \mathbf{z}_3$

One caution: plates can tilt or flex

Another factor: the earth's mantle acts like a very viscous fluid. Imagine putting an ice cube on a boat floating in molasses: it takes a while (many thousands of years for the earth) for the system to reach its equilibrium level.

After large ice sheets melt, the earth slowly rises, creating a series of terraces along the coastline:

Image removed due to copyright restrictions.

By dating deposits on the terraces, we can construct a relative sea level curve (that is dominated by local uplift).

Uplift since 6000 ka

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This evidence indicates that the (late stage) Laurentide ice sheet had two domes

source: R.F. Flint, Glacial and Quaternary Geology (QE696 .P6245)

Current uplift from GPS

Image removed due to copyright restrictions.

This evidence indicates that the (late stage) Laurentide ice sheet had two domes

source: Vanicek and Nagy (1980) Trans. Am Geophys. Union 61:146

The earth is a toothpaste tube…

A third complication: the earth is somewhat like a toothpaste tube: if you squeeze it in one place, it tends to pop up somewhere else!

These considerations lead to the use of visco-elastic models of the earth's interior. Basically, removing water from the ocean basins causes them to start moving up a little bit (because of the decreased overburden of water). Putting the ice sheets on Europe and North America of course causes the underlying continent to sink significantly (e.g. ice sheets were about 5 km maximum thickness; density of ice ~1 and mantle ~3, so continent must sink about $^5/_3$ km) and to flex near the margins. The displaced mantle then flows elsewhere, causing other places to move up. This all takes place on a finite time scale by laws of fluid dynamics, and depends heavily on the viscosity (structure) of the mantle (in fact, the only way we know the viscosity of the mantle is from studies of the rates of crustal rebound). All of this is reversed as the ice melts. Most of the isostatic adjustment occurs within a few thousand years, but there are parts of Hudson's bay that are still rebounding at rate of 1 m/century.

An interesting sidelight of this process is that estimates for how fast modern sea level is rising must take into account this crustal rebound and the associated mantle flows.

The ongoing rebound can be estimated from precise surveys; ancient rebound can be estimated by (^{14}C) dating shells of organisms that grow in the shallow seashore.

The sea-level change expected for a "static" ocean island depends on where the island is relative to the ice sheets and continents

Image removed due to copyright restrictions.

source:Clark et al. (1978) Quat. Res. 9:265-287

Other factors:

- Tilting and flexture
- Effect of gravity on sea level (Mitrovica)

Reading list

READING LIST:

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