Lecture 24
The Oceans

Reading: White, Ch 9.1 to 9.7.1 (or digital p370-400)

Last Time

1. The oceans: currents, stratification and chemistry

Today

2. the marine carbon cycle

The Marine Carbon Cycle

The marine biosphere and ocean currents control non-conservative chemical signatures in marine water masses.

Photosynthetic productivity in the surface waters requires a flux of nutrient-rich waters in to the photic zone. This is variable around the globe.

Respirative decomposition of organic matter throughout the marine water column is also variable around the globe.

If the oceans did not convect in a semi-steady-state manner, the cycle that controls the concentrations of C, O, N and P in various water masses and the relative biomass that is sustained there would not exist.

Today we investigate the movement of carbon in today's oceans.

next week we will look to the geologic record to see how past climatic conditions on Earth altered the movement of water masses the motions of carbon through the oceans, plus we consider our future.

Deep Ocean Circulation Effects:

The circulation patterns in the 2 oceans that we discussed last time results in what is sometimes referred to as basin-to-basin fractionation.

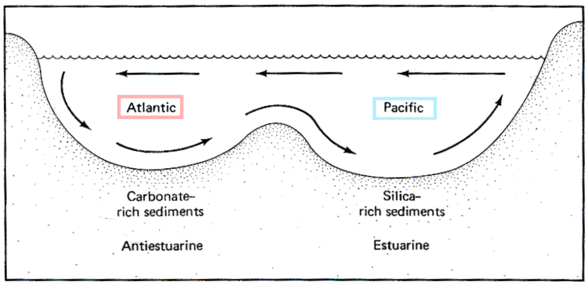

Figure 14–2 Concept of antiestuarine-estuarine circulation system between the Atlantic and Pacific oceans. The Atlantic forms young oxygenated bottom waters at high latitudes allowing carbonate to accumulate over most parts of the ocean basin. The Pacific receives older deep water at depth and gives up surface waters (like an estuary). Carbonate-poor sediments accumulate over much of the North Pacific since older deep waters are rich in CO_2 and thus are more acidic. Silica-rich sediments accumulate in areas of upwelling.

This circulation pattern affects nearly every aspect of the marine carbon cycle.

Pacific: deep water inflow is in the south. Return flow is by both upwelling and by diffuse upflow of deep water in the North Pacific. This *"in on the bottom out on the top"* circulation is like that in most estuaries, thus it is known as *estuarine circulation*.

Atlantic: has deep water formation from shallow waters: anti-estuarine circulation

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The diversity of Carbon-bearing phases in the Oceans

Carbon is present in different forms in the ocean:

- dissolved inorganic carbon (DIC): $\Sigma CO_2(aq)$
- particulate inorganic carbon (PIC): predominantly $CaCO_3$
- dissolved organic carbon (DOC): diverse mixture of simple and complex organic molecules
- particulate organic carbon (POC): dead and variably decomposed biomass
- living biomass carbon: organic carbon in living organisms, further subdivided into:
    - i. carbon in *"primary producers"* (photosynthetic organisms)
    - ii. carbon in consumer organisms (marine heterotrophs)

"Primary productivity" refers to the fixation of inorganic carbon into biomass by photosynthesis.

High primary productivity occurs in regions of the oceans that receive large nutrient inputs (e.g., upwelling zones). Low productivity occurs in areas receiving a small supply of nutrients (e.g., the centers of the oceans).

The diversity of Carbon-bearing phases in the Oceans

Chemical analysis of sea water shows the following approximate mass distributions of **organic carbon**:

Type	Concentration (µg/L)	Total Reservoir (g)
DOC	700	1×10^{18}
POC	20	3×10^{16}
Plankton	0.3	5×10^{14}
Bacteria	0.07	1×10^{14}
Other, i.e., fish	0.16	2.4×10^{14}

Note: *DOC:POC = 35:1*
By contrast, in the freshwater hydrosphere this ratio varies from 1.5:1 to 10:1

Organic and Inorganic Carbon

Redfield ratio

The **organic cycle of carbon** in the oceans involves RR carbon:
CO_2 atmosphere $\rightleftarrows$ ΣCO_2 (*aq*) $\rightleftarrows$ $(CH_2O)_n$ or $C_{106}H_{263}O_{110}N_{16}P$

The **inorganic cycle of carbon** in the oceans involves:
CO_2 atmosphere $\rightleftarrows$ ΣCO_2 (*aq*) $\rightleftarrows$ $CaCO_3$ shells

The two are inextricably linked. Both $CaCO_3$ and $(CH_2O)_n$ are associated with living and dead organisms.

Chemical analysis of authigenic sediment caught in traps left in the oceans to accumulate particulate matter over some period of time shows the following relationships:

POC : PIC	~4 :1
C : $O_{\text{particulate organic matter}}$	~1 : 1
C : $O_{\text{particulate inorganic matter}}$	~1 : 3
$O_{\text{bound in particulate organic matter}}$: $O_{\text{bound in particulate inorganic matter}}$	~1.3 : 1

Carbon-Oxygen Relationship

CO_2 and O_2 production/consumption are related by the Redfield equation, which are nutrient (N and P) limited.

The rate of oceanic primary productivity, or "how fast the mill grinds" (to coin a phrase used by Wallace Broecker, one of the persons who has helped elucidate the factors governing the marine carbon cycle) is nutrient-limited

Production of organic matter produces O_2

Digestion of organic matter consumes O_2

The molar consumption of O_2 occurs at ~1.3 x the molar destruction of Redfield Ratio POC. $\boxed{\Delta O_{2\,(+)}/\Delta CO_{2\,(-)} = 1.3}$

O_2 in the surface oceans is usually saturated to slightly supersaturated with respect to the atmosphere due to primary production.

DOC stays with the water it’s in, but POC settles out of the photic zone.

O_2 in the deep oceans is at a lower value due to respirative decomposition of falling organic matter (POC) from above.

Carbon-Oxygen Relationship

Occasionally aerobic respiration of settling particles is not fast enough to consume POC and O_2 before it reaches the sea bed. Sometimes this organic matter reaches the is buried.

For **organic carbon to be preserved in sediments** on the seafloor, the fluxes (Φ) of carbon and oxygen must follow these rules:

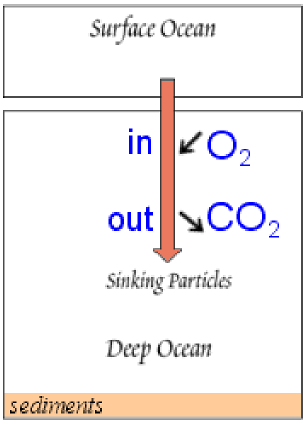
$\Phi_{C\ resp} < \Phi_{C\ photosyn} \quad = \Phi_{POC\ out\ of\ photic\ zone}$

$\Phi_{C\ photosyn} - \Phi_{C\ resp} \quad = \Phi_{POC\ to\ sediments}$

and

$\Phi_{O2\ photosyn} - \Phi_{O2\ resp} \quad = \Phi_{O2\ to\ atmosphere}$

$\boxed{\text{where each } \Phi_{O2} = 1.3\ \Phi_C \text{ (from the Redfield ratio)}}$



Variations in the amount of organic matter preserved in marine sediments in the geologic record constrain how the marine C and O cycles operated in the past on Earth.

Carbon-Oxygen Relationship

- Total Primary Productivity/yr = 197 x 10^{12} g C ~ 2 x10^8 tons C.
- Total Respiration/yr = 240 x 10^{12} gC = 2.4 x10^8 tons C
(of which ~ 25% is caught for human consumption)
- Presently oceanic respiration > photosynthesis by ~4 x 10^7 tons C/yr (15%), although these estimates are basically equal, given the large relative errors.

How much O_2 loss does this imply?

How might human impacts on nutrient cycling affect this?

C_{org} in sediments

Given these conditions, we don't expect significant organic matter preservation in marine sediments on a global scale today .

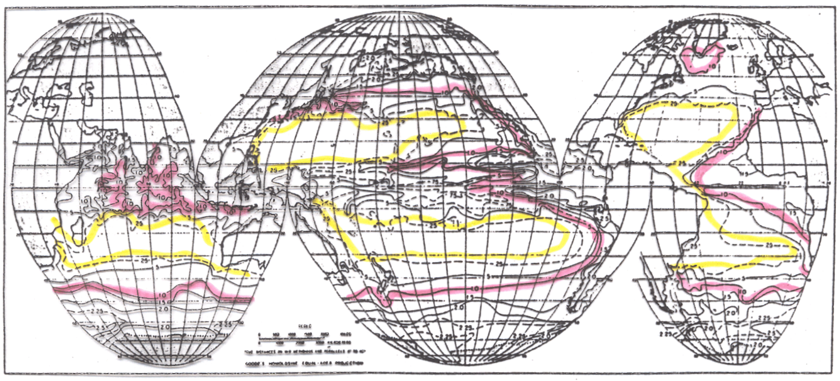
In fact, we only see appreciable C_{org} preservation in near surface sediments under relatively high productivity zones (which account for only ~0.5% of the overall marine C_{org} production, but which happen in just 0.1% of the ocean).

Clearly, there have been times in the geologic past (e.g., the carboniferous period) where this balance was different.

Geographic distribution of $\Phi_{C\ org}$ in the oceans

The circulation patters of the oceans cause predictable patterns in net photosynthesis and respiration rates in the oceans, as well as in water chemistry and in marine sediment composition.

This plot shows DIP contours in surface waters (red is high, yellow is low)



DIP

Fig. 8. Distribution of inorganic phosphate-phosphorus (µg-at/ℓ) at 100 m in the world ocean; Indian Ocean part adapted from WYRTKI (1971, Fig. 73), Pacific from REID (1962, Fig. 3a); Goode's Projection copyright by the University of Chicago Department of Geography.

Geographic distribution of $\Phi_{C\ org}$ in the oceans

(Plant) and Animal distribution broadly follows DIP. (In this case we consider just plankton, which are aquatic organisms that move with the currents).

DIP

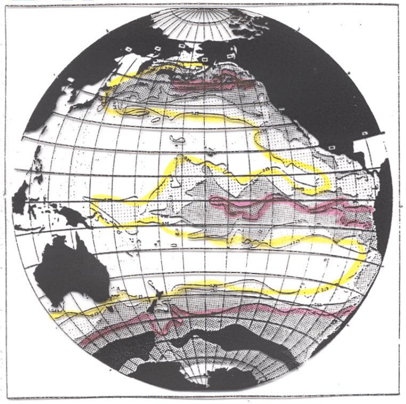


Fig. 7. Distribution of inorganic phosphate-phosphorus (μg-at/l) at the surface of the Pacific Ocean (REID, 1962, Fig. 3b).

Animals

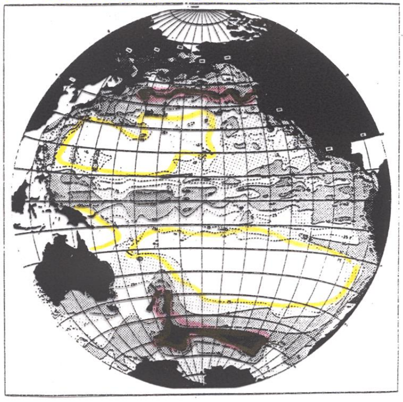


Fig. 11. Distribution of zooplankton volume (parts per 10^8 by volume) in approximately the upper 150 m of the Pacific Ocean (REID, 1962, Fig. 4b).

Geographic distribution of $\Phi_{C\ org}$ in the oceans

The global distribution of oceanic Primary Productivity

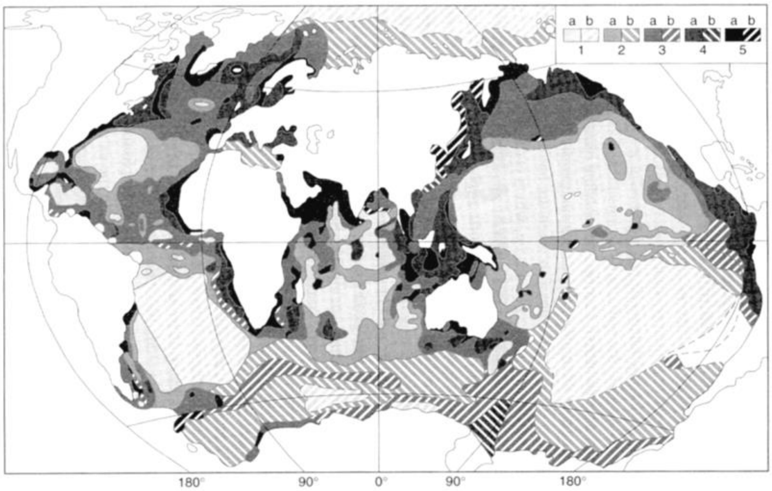


Figure 8.4. Rates of organic matter production (in mg C/m²/day) for the oceans: (1) less than 100; (2) 100–150; (3) 150–250; (4) 250–500; (5) more than 500. a = data from direct ^{14}C measurements; b = data from phytoplankton biomass, hydrogen, or oxygen saturation. (After O. J. Koblentz-Mishke, V. V. Volkovinsky, and J. G. Kabanova, "Plankton Primary Production of the World Ocean." In *Scientific Exploration of the South Pacific*, ed. W. S. Wooster, p. 185. Copyright © 1970 by the National Academy of Science, reprinted by permission of the publisher.)

Berner and Berner, "Global Environment"

Geographic distribution of $\Phi_{C\ org}$ in the oceans

Photosynthesis Breakdown:			
	open ocean	coastal zones	coastal and other upwelling zones
area km^2	326×10^6	32.6×10^6	0.33×10^6
area %	90	9.9	0.1
PP $(gC)/(m^2/yr)$	50	100	300
Total Prod$(x10^{12}gC/yr)$ calc. from area x PP x 10^4	163	32.6	1
% of Total	83	17	0.5
Respiration Breakdown:			
trophic levels*	5	3	1.5
*the number of food chain steps between producers and upper level consumers.			
Efficiency of C use	0.1	0.15	0.2
Neuston/zooplankton productivity (tons/yr)	16×10^5	12×10^7	12×10^7
% of Total	0.66	49.7	49.7

Inorganic Carbon in the oceans

Calcium Carbonate solubility.

- $CaCO_3$ (s) $\rightleftarrows$ Ca^{2+} (aq) + CO_3^{2-} (aq) is an important equilibrium in the oceans. Many organisms produce $CaCO_3$ shells and the production of carbonate minerals by this reaction lowers ΣCO_2 (*aq*) in sea water.

- K_{sp} of $CaCO_3$ varies strongly with T and P

Table 2-2: (*Broecker "Chemical Oceanography"*)

		Saturation Carbonate Ion Content, 10^{-6} moles/liter	
Temperature, °C	**Pressure, atm***	**Calcite**	**Aragonite**
24	**1**	**53**	**90**
2	**1**	**72**	**110**
2	**250**	**97**	**144**
2	**500**	**130**	**190**

*** a pressure of 100 atm is achieved at close to 1000 meters depth in the sea.**

Thus...

K_{sp} of $CaCO_3$ $\Uparrow$ as T $\Downarrow$ K_{sp} of $CaCO_3$ $\Uparrow$ as P $\Uparrow$	both characteristics are the opposite of those for many ionic solids

- Because with increasing depth into the in the Oceans..

T $\Downarrow$ and P $\Uparrow$	K_{sp} of $CaCO_3$ $\Uparrow$ with depth.

Inorganic Carbon in the oceans

Calcium Carbonate solubility.

So $CaCO_3$ is more soluble with depth in the oceans.

Biogenic Silica solubility.

In contrast, siliceous (SiO_2) shells and tests dissolve more easily in shallow waters, where SiO_2 is the least saturated.

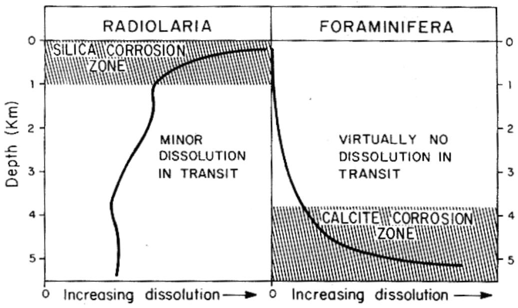


Figure 14-1 Comparison of dissolution profiles of radiolaria and planktonic foraminifera based on field experiments. Most dissolution of radiolaria (and diatoms) occurs in shallow waters. In contrast, most dissolution of carbonate microfossils occurs on the sea floor at depths greater than 3.5 km. (From W. H. Berger, 1976)

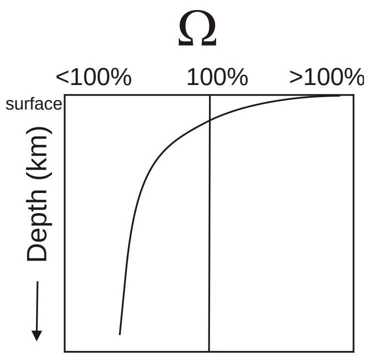
Inorganic Carbon in the oceans

Calcium Carbonate solubility.

- We define a parameter Ω as the % carbonate saturation to characterize water oceanic masses with respect to $CaCO_3$ solubility

$$\Omega = \frac{m_{Ca2+}\, m_{CO3\,2-}}{K_{sp}'} \times 100\%$$

(where $K_{sp}' = K_{sp}/a_{Ca2+}\, a_{CO3\,2-}$)



simplified oceanic profile

- Because oceanic m_{Ca2+} and γ_{Ca2+} vary by only a % or so

$$\Omega \sim \frac{m_{CO3\,2-}}{m_{CO3\,2-\text{ at CaCO3saturation}}} \times 100\%$$

molality

molality at saturation

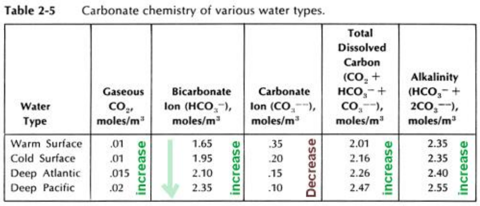


Table 2-5 Carbonate chemistry of various water types.

Water Type	Gaseous CO_2, moles/m^3	Bicarbonate Ion (HCO_3^-), moles/m^3	Carbonate Ion (CO_3^{--}), moles/m^3	Total Dissolved Carbon (CO_2 + HCO_3^- + CO_3^{--}), moles/m^3	Alkalinity (HCO_3^- + $2CO_3^{--}$), moles/m^3
Warm Surface	.01	1.65	.35	2.01	2.35
Cold Surface	.01	1.95	.20	2.16	2.35
Deep Atlantic	.015	2.10	.15	2.26	2.40
Deep Pacific	.02	2.35	.10	2.47	2.55
	increase	increase	Decrease	increase	increase

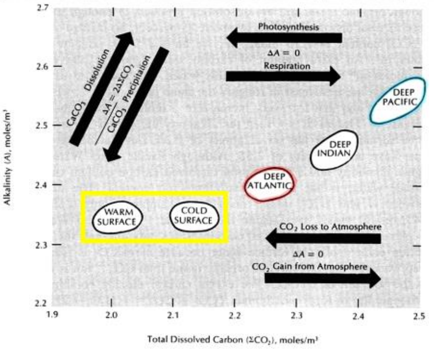


Figure 2-4 Relationship between the total dissolved inorganic carbon content ΣCO_2 and the alkalinity *A* of waters from various parts of the ocean (a so-called "Deffeyes diagram"). Arrows indicate the effects of various processes occurring within the sea. Of these, only the formation and the destruction of $CaCO_3$ cause the alkalinity to vary. The warm–cold surface ocean difference is due to the transfer of gaseous CO_2 to and from the atmosphere. The cold surface–deep water sequence is due to a combination of respiration and $CaCO_3$ dissolution. Broecker, "Chemical Oceanography"

Inorganic Carbon in the oceans

The oceanic currents and deep circulation patterns we discussed last time affect ΣCO_2 (*aq*) with depth and also with water mass age.

This in turn affects $CaCO_3$ distribution in surface sediments

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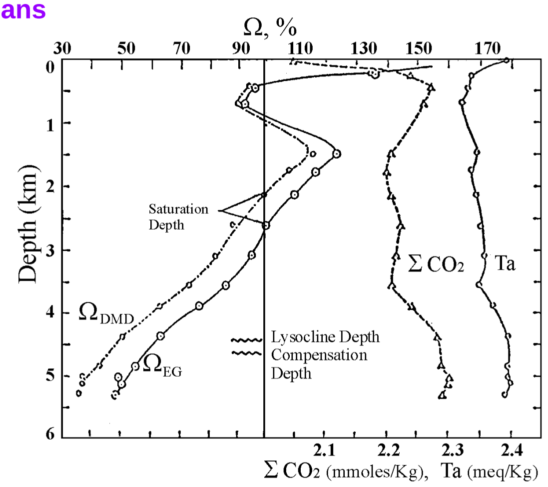
Inorganic Carbon in the oceans

Calcium Carbonate solubility.



👁 It is useful to look at Ω from the perspective of Alkalinity, Ca^{2+} and ΣCO_2 (*aq*) variations with depth in the oceans.

notice that Ω decreases even though ΣCO_2 and Total Alkalinity (TA) are roughly constant.

Also notice that ΣCO_2 and TA track very closely except in the very topmost 500 m or so of the oceans, where CO_2 is being consumed by photosynthesizers at a rate that is faster than the replenishment rate by equilibration with the atmosphere.

pH effects of the carbon cycle in the oceans

Photosynthetic production of $(CH_2O)_n$ consumes CO_2 (present mostly as bicarbonate in seawater) and H^+ and **raises the pH**, as summarized in the Redfield equation.

Production of $CaCO_3$ from bicarbonate ion liberates H^+ and therefore **lowers the pH.**

Because the ratio of marine POC:PIC ≈ 4:1,

marine phytoplankton use up more H^+ in producing organic matter

than they liberate in producing $CaCO_3$ tests and shells.

This makes the water more basic. For this reason, pH is generally higher in the surface waters of the ocean than it is at depth.

$CaCO_3$ deposition/preservation on the sea floor.

Geographic variations in surface sediment biogenic **$CaCO_3$ preservation** reflect variability in **productivity** and **deep water composition**.

Geologists examine the temporal variability of geographic variations by looking at cores drilled into the sediments of the world's ocean basins.

Important considerations:

- biogenic $CaCO_3$ comes in 2 different mineral forms, **calcite** (e.g., by coccolithophorae and foraminifera) and **aragonite** (e.g., by pteropods)
- K_{sp} for the two are different. Aragonite is meta-stable (thermodynamically) at all depths in the sea.
- $CaCO_3$ in either form is more soluble at greater depths.
- The depth of the saturation horizon (where **Ω** = 100%) ***varies from place to place, but there are no outward signs of its existence.*** Therefore, this horizon cannot be located and mapped without conducting a large number of measurements of composition, temperature, and pressure. This is impractical.

$CaCO_3$ deposition/preservation on the sea floor.

$CaCO_3$ dissolution kinetics are slow so $CaCO_3$ can persist on the sea floor even in waters where $\Omega < 100\%$.

This is analogous to the accumulation of snow produced in the troposphere (at T< 0°C) on the ground when T > 0°C, if it is falling fast enough.

If supply to the sediments is great from high $CaCO_3$ production in overlying waters, dissolution is "overwhelmed" by the supply and $CaCO_3$ will accumulate in sediments below the saturation horizon.

The compensation depth

- This depth below which $CaCO_3$ is generally not preserved in surface sediments tells us about carbon supply from above and corrosiveness of deep water for $CaCO_3$.
- The compensation depth is a kinetic horizon, rigorously defined as the location where dissolution rate/supply rate = 1.
- Beneath low $CaCO_3$ production regions the preservation horizon is close to or at the equilibrium saturation depth, but elsewhere it can reach 4500m.

$CaCO_3$ deposition/preservation on the sea floor.

Some details:
There are 2 compensation depths:
- calcite compensation depth (CCD): deeper
- aragonite compensation depth (ACD): shallower (aragonite solubility is higher as a function of decreasing temperature and increasing pressure).

$CaCO_3$ dissolution kinetics.
- dissolution rate is much faster at lower Ω.
- aragonitic tests dissolve faster than calcite tests.
- not all calcite tests are "built" the same, which influences dissolution rate:
  - a. Shape and size - determines how fast they settle
  - b. ornamentation - more ornamental and porous types (e.g., foram tests) dissolve more quickly than small, compact Coccolith tests.
  - c. extent of organic coatings affect dissolution as well

  Some of these features in turn reflect conditions in which the organism lived, such as temperature and water depth.

$CaCO_3$-rich marine sediments are known as calcareous oozes:
- Foramineferal ooze: sand sized, calcite. easy to dissolve.
- Nanofossil or Coccolith ooze: fine grained, calcite, harder to dissolve
- Pteropod ooze: large sand sized, aragonite, dissolves very easily

$CaCO_3$ deposition/preservation on the sea floor.

Various confounding factors mean that some shell types are preserved better than others on the sea floor, causing a range of horizons between the saturation depth (Ω = 100%) and the CCD.

The increasing rate of $CaCO_3$ dissolution and increased effect on more types of shells with depth leads to a loss of species diversity preservation in the sediments with increasing sea floor depth, even at constant species diversity in overlying waters (see diagram).

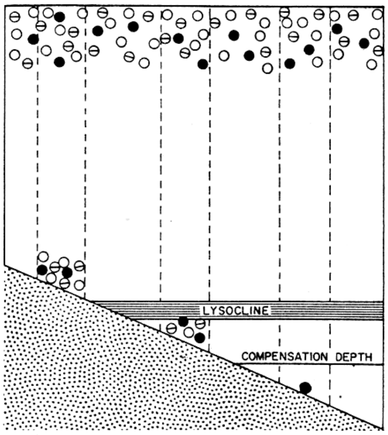


Figure 14–10 Schematic diagram shows increasingly selective dissolution of planktonic foraminiferal species with increasing water depth. Most species remain in sediments above lysocline, below which there is rapid decrease in diversity of species and morphotypes and concentration of robust forms. Below the CCD virtually no species remain. (With permission from A.W.H. Bé, *Oceanic Micropalaeontology*, vol. 1, p. 49, 1977. Copyright by Academic Press, Inc. (London) Ltd.)

For practical reasons, oceanographers and marine chemists often examine geographic depth variations of one or more of these easier-to-identify horizons and relate this back to the saturation horizon or CCD.

The Lysocline

An experiment using polished calcite spheres that sat for 4 months attached at various ocean depths to a wire hung from a buoy demonstrated that significant dissolution occurred at a critical depth level, **termed the *"lysocline***. The rate of dissolution increases with depth below this.

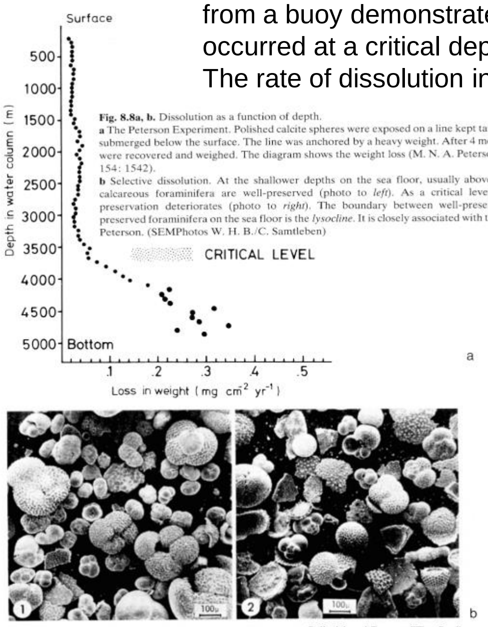


Fig. 8.8a, b. Dissolution as a function of depth.
a The Peterson Experiment. Polished calcite spheres were exposed on a line kept taut by a large buoy submerged below the surface. The line was anchored by a heavy weight. After 4 months the spheres were recovered and weighed. The diagram shows the weight loss (M. N. A. Peterson, 1966, Science 154: 1542).
b Selective dissolution. At the shallower depths on the sea floor, usually above 3,000 m or so, calcareous foraminifera are well-preserved (photo to *left*). As a critical level is approached, preservation deteriorates (photo to *right*). The boundary between well-preserved and poorly preserved foraminifera on the sea floor is the *lysocline*. It is closely associated with the critical level of Peterson. (SEMPhotos W. H. B./C. Samtleben)

Seibold and Berger, "The Sea"

Depths of various horizons in the area of the experiment:

equilibrium saturation horizon	~0.5 km
the lysocline	3.7 km
CCD	~4.7 km

The lower figure demonstrates that the most sensitive shells can dissolve above the lysocline, while other more robust forms will not.

$CaCO_3$ deposition/preservation on the sea floor.

These preservation variations lead to a generic depth distribution pattern (note: Pteropods = aragonite; globigerina = foram = calcite).

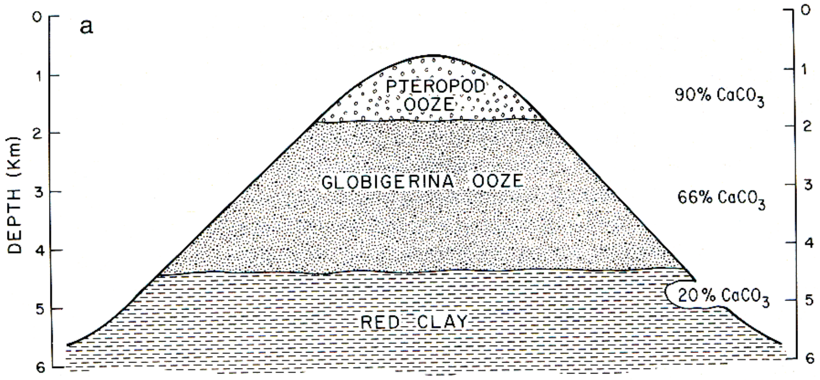


Fig. 8.6a, Depth distribution of calcareous deep sea sediments. Idealized bathymetric zonation of deep sea deposits, produced by increasing dissolution of carbonate with depth. (according to J. Murray, J. Hjort, 1912, The depths of the ocean. Macmillan, New York).

Seibold and Berger, "The Sea"

Geographic variations in $CaCO_3$ preservation

Both $CaCO_3$ precipitation and photosynthesis by phytoplankton **reduce seawater ΣCO_2** in surface water. These solids which sink to deeper water.

As deep waters flow from the Atlantic through the Indian to the Pacific ocean they acquire chemical signatures of continued POC & PIC input.

“Older” waters of the north Pacific have **higher ΣCO_2** and **lower pH** compared to the Atlantic.

These have competing $CaCO_3$ solubility effects:

- increased H^+ should help dissolve $CaCO_3$
- increased CO_3^{2-} should help preserve it

$$CaCO_3 + H^+ \rightleftarrows Ca^{2+} + HCO_3^-.$$

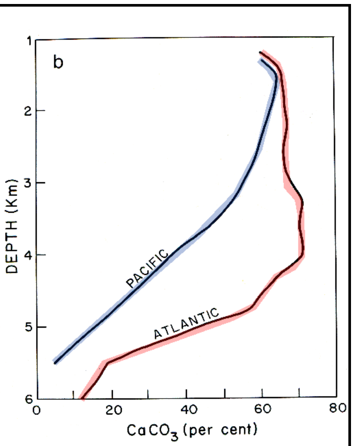


Fig. 8.6 b. Depth distribution of calcareous deep sea sediments.
Generalized depth profiles for carbonate content in deep sea sediments. (After R. R. Revelle, 1944, Carnegie Inst Wash Publ 556)

Seibold and Berger, "The Sea"

The net increase of H^+ from respiration of CO_{org} "wins out".

There is less $CaCO_3$ preservation on the deep Pacific seafloor than in the deep Atlantic, effectively "shallowing" the various $CaCO_3$ preservation horizons in the Pacific relative to the Atlantic.

Geographic variations in $CaCO_3$ preservation

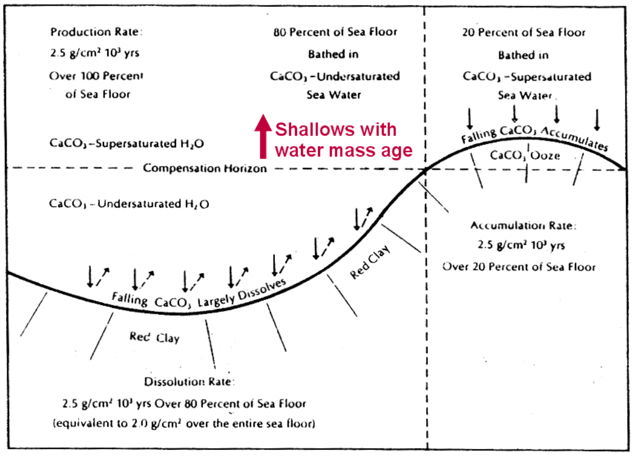


Figure 3-5 Cycle of $CaCO_3$ in the sea. Organisms generate $CaCO_3$ at a fairly uniform rate over the sea surface. About 20 percent of this $CaCO_3$ falls onto portions of the sea floor projecting above the saturation horizon and is preserved in the sediments. The remainder falls into undersaturated water and largely dissolves. The model predicts that for each square centimeter of sea floor 2.5 grams of $CaCO_3$ fall from the surface every 1000 years. Since 80 percent of this dissolves, the deep sea receives an average of 2.0 grams of $CaCO_3$ every 1000 years for each square centimeter of ocean surface. The model also predicts that where the sea floor projects above the saturation horizon, the accumulation rate should be 2.5 g/cm² 10³ yrs.

The CCD gets shallower as deep waters migrate from the Atlantic through the Indian to the Pacific ocean.

$CaCO_3$ deposition/preservation on the sea floor.

Local depth variations in the CCD

In general:

increased surface water primary productivity → increased $CaCO_3$ production → increased $CaCO_3$ supply to deep waters → a deeper CCD.

At upwelling zones:

Some of the productivity in these waters is caused by diatoms, which fix organic matter and produce SiO_2 tests.

The numbers of respiring organisms increase beneath upwelling zones, lowering the seawater pH via respiration of $(CH_2O)_n$.

This occurs without much increase in $CaCO_3$ flux, resulting in a shallower CCD than we'd otherwise predict.

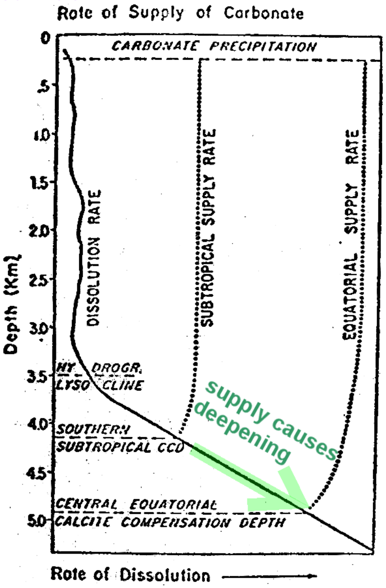


Fig. 8. Simple model accounting for equatorial CCD depression

Berger, Adelseck, Mayer

$CaCO_3$ deposition/preservation on the sea floor.

Depth variations of $CaCO_3$ preservation along north-south transects through the eastern and western Atlantic demonstrate these competing effects.

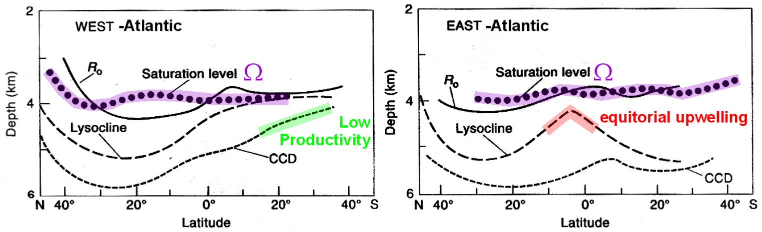


Figure 8.14. Plots of the depth of the carbonate compensation depth (CCD), lysocline, saturation level, and R_0 depth for surface sediments of the Atlantic Ocean as a function of latitude. The R_0 depth is that where evidence for dissolution is first encountered, and the saturation level is that depth in the water column below which calcite becomes undersaturated and can therefore dissolve. (After W. H. Berger, "Carbon Dioxide Excursions in the Deep Sea Record: Aspects of the Problem." In *The Fate of Fossil Fuel CO_2 in the Oceans*, ed. N. R. Andersen and A. Malahoff, p. 512. Copyright © 1977 by Plenum Press, reprinted by permission of the publisher.)

- The eastern Atlantic shows a CCD and lysocline bump near the equator due to a increased abundance of siliceous phytoplankton and zooplankton in overlying waters of the equatorial upwelling zone.
- The western Atlantic has a similar pattern in the northern hemisphere and at the equator.
- In the southern hemisphere, $CaCO_3$ production in overlying waters is very low, resulting in the CCD and other horizons closely approaching the saturation horizon.
- The effect of reduced preservation under high-productivity coastal upwelling zones is even more dramatic. The supply of organic matter from above is so high that pH can be lowered so much that $CaCO_3$ can dissolve even on shallow continental slopes at water depths of just a *few hundred meters*.

Plate Stratigraphy/oceanic sediment distribution

Away from continental margins,

the shallow spots on the sea floor

are almost always **volcanic constructs**, many of which are active (such as the globe encircling mid-ocean ridge system).

Oceanic lithosphere cools after it forms and thus deepens as it migrates from the volcanoes, eventually passing below the CCD. It might also pass beneath equatorial or other upwelling zones along the way.

The movement of the oceanic lithospheric plates and the circulation patterns of overlying waters lead to global distribution patterns of sediment types (see Table 14-1 on the next slide)

This leads to a typical sequence (A-D) of sediments overlying the basaltic part of the plate crust as well as to (also on the next slide).

General sequence

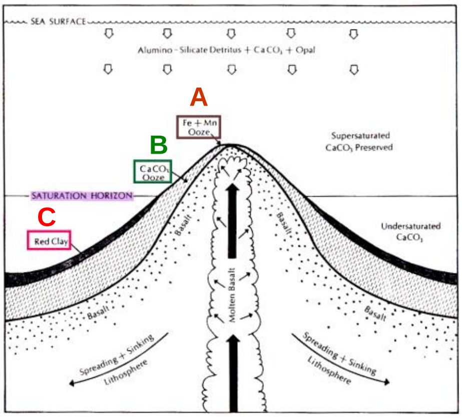


Figure 2-8 Sequence of sediment types accumulated by the great lithospheric plates as they move away from the crests of mid-ocean rises. The first sediment to be deposited is iron (Fe) + manganese (Mn) oxide, a product of volcanism. When a point a few kilometers away from the crest is reached, the sediment no longer receives volcanic products and is dominated by $CaCO_3$ falling from the surface. At a point several hundred kilometers from the crest, the plate subsides below the saturation horizon for calcite and $CaCO_3$ no longer accumulates. Beyond this point, continental detritus and perhaps opal dominate the sediment. A core bored through sediment capped with red clay would encounter buried $CaCO_3$ and then a thin layer of Fe + Mn-rich sediment before entering the underlying basalt (hard rock).

An example based on lithosphere produced along the equatorial East Pacific Rise)

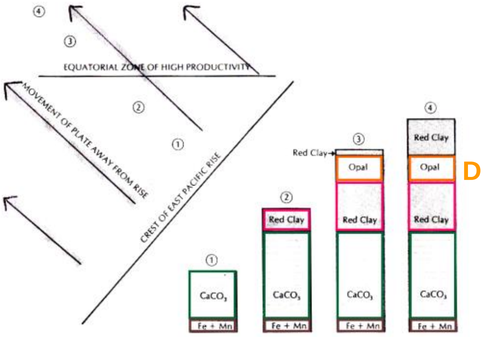


Figure 2-9 Sequence of sediments accumulating on a plate which crosses the high productivity equatorial belt in its movement away from a ridge crest. The numbers relate the map locations to the corresponding sedimentary sequences found in borings made at these points.

TABLE 14-1 PERCENTAGE OF PELAGIC SEDIMENT COVERAGE IN THE WORLD OCEANS[a]

	Atlantic	Pacific	Indian	Total
Foraminiferal ooze	65	36	54	47
Pteropod ooze	2	0.1	—	0.5
Diatom ooze	7	10	20	12
Radiolarian ooze	—	5	0.5	3
Red clay	26	49	25	38
Relative size of ocean (%)	23	53	24	100

[a] From Berger [1976]. Data from Sverdrup and others [1942]. Pacific pteropod ooze area from Bezrukov [1970]. Area of deep-sea floor = 268.1 × 10^6 km^2.

Plate Stratigraphy/oceanic sediment distribution

A. Metalliferous sediments form at ridge axes when low pe hydrothermal waters and high pe ocean waters mix because soluble Fe^{2+} and Mn^{2+} become oxidized, and precipitate as insoluble Fe^{3+} and Mn^{4+} oxides. The large localized flux of precipitating Fe and Mn oxides are mixed with the $CaCO_3$ sedimentation from above.

B. purer $CaCO_3$ rich "oozes" are deposited over the metal rich strata within a few km of the ridge axis (if sea floor is above the CCD).

C. The crust passes below the CCD after spreading some distance from the ridge axis (subsidence cooling is inversely proportional to the spreading rate). “Red clays” slowly accumulate (terrigenous wind-blown debris plus some clays made at deep sea volcanoes are deposited all over the oceans but they are swamped out by other sediment types where the latter occur).

D. Siliceous oozes will be found in the sediments if the sea floor passes beneath a high productivity zone (where siliceous plankton live) at any point along this path,. "The high SiO_2 zone" can be found in sediments with or without $CaCO_3$, depending on depth and location (above below the CCD).