CALCIUM CARBONATES

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Introduction

The ocean receives a continual input of calcium from riverine and groundwater sources and from the hydrothermal alteration of oceanic crust at mid-ocean ridge spreading centers. Balancing this input is the biological precipitation of calcium carbonate ($CaCO_3$) by shell- and skeleton-building organisms in both shallow marine and open-ocean environments. In the deep sea, the primary contributors to the carbonate budget of open-ocean sediments are the skeletal remains of calcareous plankton that have settled down from the surface after death. Seafloor sediments consisting of more than 30% by weight calcium carbonate are traditionally referred to as calcareous or carbonate ooze; such oozes accumulate at the rate of 1–4 cm per 1000 years and cover roughly half of the ocean bottom. Carbonate oozes are the most widespread biogenous sediments in the ocean.

While the biological production of calcium carbonate in oversaturated surface waters determines the input of carbonate to the deep sea, it is the dissolution of carbonate in undersaturated deep waters that has the dominant control on calcium carbonate accumulation in the open ocean. Since carbonate production rates in the surface ocean today greatly exceed the rate of supply of calcium, this 'compensation' through dissolution must occur in order to keep the system in steady-state. Increased dissolution at depth is largely a function of the effect of increasing hydrostatic pressure on the solubility of carbonate. However, superimposed on this bathymetric effect are regional preservation patterns related to differences in carbonate input and the carbonate chemistry of deep water masses. Carbonate oozes in the deep sea serve as a major reservoir of calcium and carbon dioxide on the Earth's surface. Their spatial and temporal accumulation patterns in the marine stratigraphic record are thus a primary source of data about the carbonate chemistry and circulation of past oceans, as well as of the global geochemical cycle of CO_2.

Carbonate Producers

The most important carbonate producers in the open ocean are planktonic coccolithophorids and foraminifera, unicellular phytoplankton and zooplankton respectively, which inhabit the upper few hundred meters of the water column (**Figure 1**). Coccolithophorids are the dominant carbonate-precipitating organisms on Earth. During part of their life cycle, they produce a skeletal structure (the coccosphere) consisting of loosely interlocking plates, often button-like in appearance, known as coccoliths. Deep-sea carbonates generally contain only the individual coccoliths, as the intact coccospheres are rarely preserved. Foraminifera produce a calcareous shell, or 'test', a few hundred microns in size that sinks after death or reproduction to the seafloor. Both coccolithophorids and the planktonic foraminifera construct their skeletal elements out of the mineral calcite, the more stable polymorph of $CaCO_3$. Calcareous sediments dominated by one or the other component are termed coccolith oozes or foraminiferal oozes, although in reality most carbonate-rich sediments are a mixture of both.

Coccolithophorids made their first appearance in the geological record in the earliest Jurassic, while planktonic foraminifers evolved somewhat later in the middle Jurassic. The appearance of these two dominant pelagic carbonate producers, and their rapid diversification in the Cretaceous, would have had major effects upon the carbonate geochemistry of the oceans. Before this, most carbonate was deposited in shallow seas, accounting for the high proportion of limestones among older rocks on the continents. Since the Mesozoic, deep-ocean basins have become enormous sinks for carbonate deposition.

Smaller contributions to the deep-sea carbonate budget come from a variety of other sources. Pteropods, free-swimming pelagic gastropods, construct a relatively large (several millimeters) but delicate shell out of the metastable form of $CaCO_3$ known as aragonite. However, while pteropods can be unusually abundant in certain environments, the increased solubility of aragonite leads to very restricted preservation of the shells and pteropod oozes are relatively rare in the ocean. In the vicinity of shallow, tropical carbonate platforms such as the Bahamas or Seychelles Bank, shedding of aragonitic bank-top sediments derived from algal and coral production can lead to aragonite-rich 'periplatform oozes' in deep waters around the perimeters of the platform. In general, contributions from bottom-dwelling organisms (e.g. benthonic foraminifera,
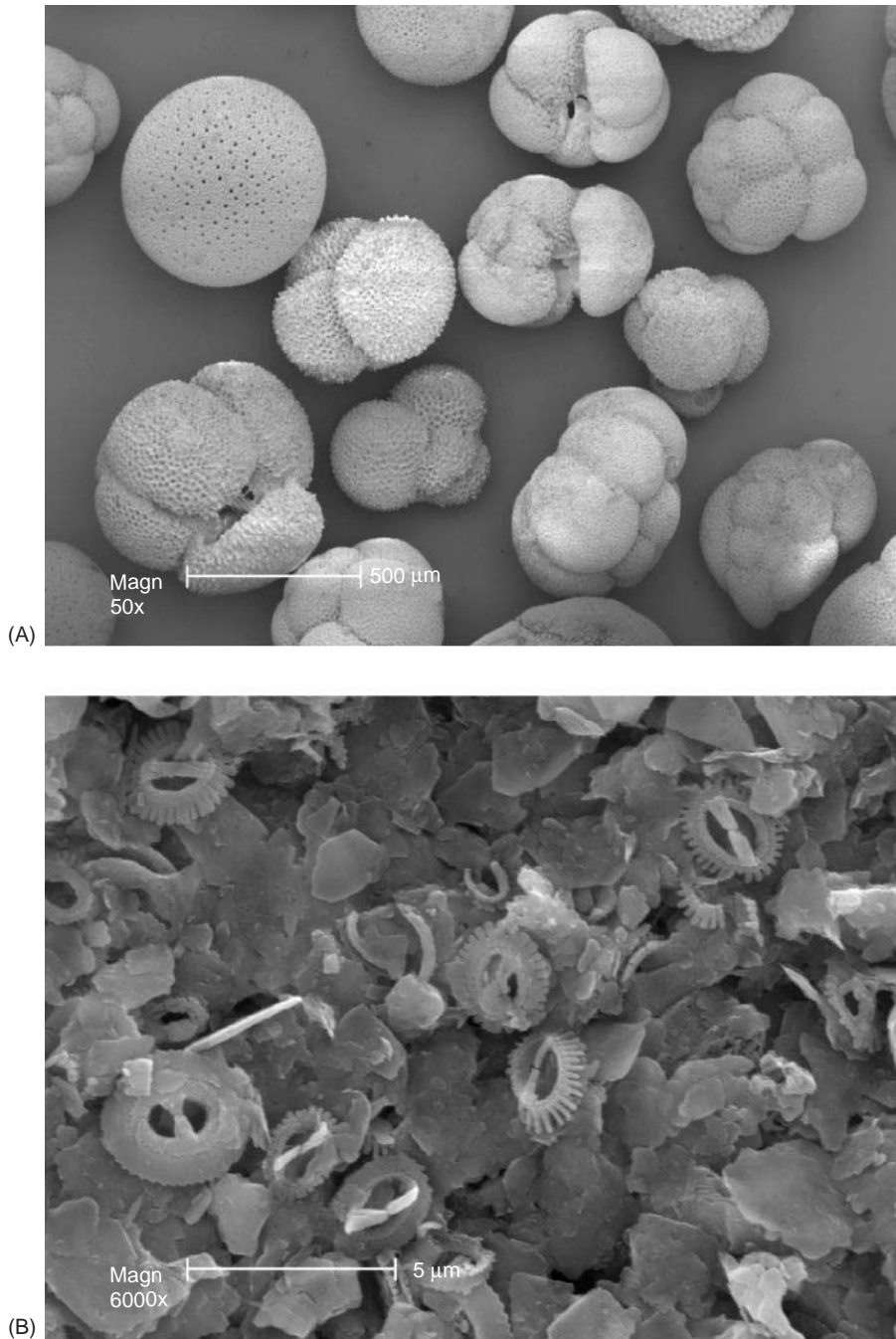
Figure 1 Carbonate oozes in the deep sea are dominated by the skeletal remains of (A) planktonic foraminifera ($\times 50$ magnification) and (B) coccolithophorids ($\times 6000$ magnification). Specimens shown here were isolated from a Caribbean sediment core.

ostracods, micromollusks) are negligible in deep-sea sediments.

Carbonate Distribution and Dissolution

The distribution of carbonate sediments in the ocean basins is far from uniform (**Figure 2**). If it were possible to drain away all of the ocean's water, carbonate oozes would be found draped like snow over the topographic highs of the seafloor and to be largely absent in the deep basins. The lack of carbonate-rich sediments in the deepest parts of the world's oceans has been recognized since the earliest investigations. Although surface productivity and dilution by noncarbonate sediment sources can locally
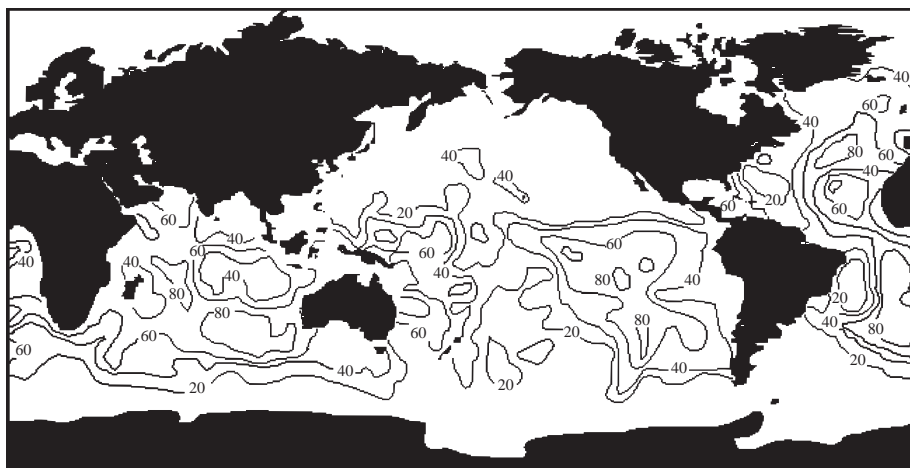
Figure 2 Global distribution of calcium carbonate (weight-% CaCO$_3$) in surface sediments of the ocean. Data compilation from Archer (1996); reproduced with permission from the American Geophysical Union.

influence the concentration of carbonate in deep-sea sediments, the clear-cut relationship between calcium carbonate content and water depth indicates that carbonate dissolution plays the major role in governing carbonate distribution patterns. To a first approximation, the dissolution of carbonate on the seafloor is a function of the corrosiveness or saturation state of the overlying bottom waters.

The amount of calcium carbonate that will dissolve in sea water if thermodynamic equilibrium is reached is governed by the following reaction:

$$CaCO_3(s) \leftrightarrow Ca^{2+}(aq) + CO_3^{2-}(aq)$$
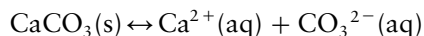
At equilibrium, the rate of carbonate dissolution is equal to the rate of its precipitation and the sea water is said to be saturated with respect to the carbonate phase. In the deep sea, the degree of calcium carbonate saturation (D) can be expressed as:

$$D = \frac{[Ca^{2+}]_{seawater} \times [CO_3^{2-}]_{seawater}}{[Ca^{2+}]_{saturation} \times [CO_3^{2-}]_{saturation}}$$

where $[Ca^{2+}]_{seawater}$ and $[CO_3^{2-}]_{seawater}$ are the *in situ* concentrations in the water mass of interest and $[Ca^{2+}]_{saturation}$ and $[CO_3^{2-}]_{saturation}$ are the concentrations of these ions at equilibrium, or saturation, at the same conditions of pressure and temperature. Since shell formation and dissolution cause the concentration of Ca^{2+} to vary by less than 1% in the ocean, the degree of calcium carbonate saturation (D) can be simplified and expressed in terms of the concentration of the carbonate ions only:

$$D = \frac{[CO_3^{2-}]_{seawater}}{[CO_3^{2-}]_{saturation}}$$

D is thus a measure of the degree to which a seawater sample is saturated with respect to calcite or aragonite, and so provides a measure of the strength of the driving force for dissolution. Values of D > 1 indicate oversaturation while values of D < 1 indicate undersaturation and a tendency for calcium carbonate to dissolve. Since the saturation carbonate ion concentration increases with increasing pressure and decreasing temperature, calcium carbonate is more soluble in the deep sea than at the surface. At the depth in the water column where D = 1, the transition from oversaturated to undersaturated conditions is reached. This depth is known as the saturation horizon (**Figure 3**). Aragonite is always more soluble than calcite, and its respective saturation horizon is shallower, because the saturation carbonate ion concentration for aragonite is always higher for the same conditions of pressure and temperature.

Observations from studies of surface sediments have allowed definition of regionally varying levels in the ocean at which pronounced changes in the presence or preservation of calcium carbonate result from the depth-dependent increase of dissolution on the seafloor. The first such level to be identified was simply the depth boundary in the ocean separating carbonate-rich sediments above from carbonate-free sediments below. This level is termed the calcite (or carbonate) compensation depth (CCD) and represents the depth at which the rate of carbonate dissolution on the seafloor exactly balances the rate of carbonate supply from the overlying surface waters. Because the supply and dissolution rates of carbonate differ from place to place in the ocean, the depth of the CCD is variable. In the Pacific, the CCD is typically found at depths between about 3500 and
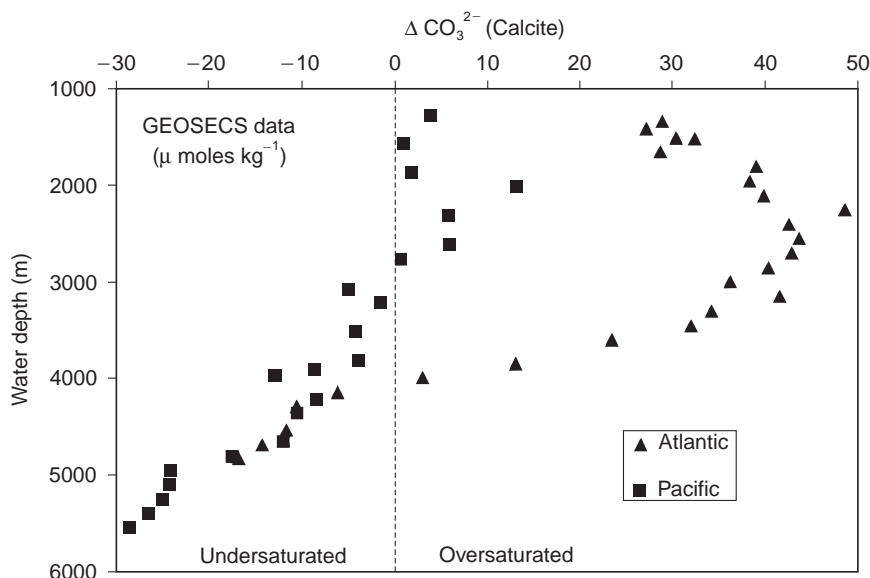
Figure 3 Bathymetric profiles of calcium carbonate (calcite) saturation for hydrographic stations in the Atlantic and Pacific Oceans (data from Takahashi *et al.* 1980). Carbonate saturation here is expressed as ΔCO_3^{2-}, defined as the difference between the *in situ* carbonate ion concentration and the saturation carbonate ion concentration at each depth ($\Delta CO_3^{2-} = [CO_3^{2-}]_{seawater} - [CO_3^{2-}]_{saturation}$). The saturation horizon corresponds to the transition from waters oversaturated to waters undersaturated with respect to calcite ($\Delta CO_3^{2-} = 0$). This level is deeper in the Atlantic than in the Pacific because Pacific waters are CO_2-enriched and $[CO_3^{2-}]$-depleted as a result of thermohaline circulation patterns and their longer isolation from the surface. The Atlantic data are from GEOSECS Station 59 (30°12′S, 39°18′W); Pacific data come from GEOSECS Station 235 (16°45′N, 161°23′W).

4500 m. In the North Atlantic and parts of the South Atlantic, it is found closer to a depth of about 5000 m. Close to continental margins the CCD tends to shoal, although much of this apparent rise can be attributed to carbonate dilution by terrigenous input from the continents. Rarely does carbonate ooze accumulate on seafloor that is deeper than about 5 km.

In practice, the CCD is identified by the depth transition from carbonate ooze to red clay or siliceous ooze that effectively defines the upper limit of the zone of no net $CaCO_3$ accumulation on the seafloor. Given the practical difficulty (e.g. analytical precision, redeposition) of determining the depth level at which the carbonate content of sediment goes to zero, some investigators choose instead to recognize a carbonate critical depth (CCrD), defined as the depth level at which carbonate contents drop to $< 10\%$ of the bulk sediment composition. The CCrD lies systematically and only slightly shallower than the CCD. A similar boundary to the CCD can be recognized marking the lower depth limit of aragonite-bearing sediment in the ocean, the aragonite compensation depth or ACD. Because of the greater solubility of aragonite as compared with calcite, the ACD is always much shallower than the CCD.

Above the CCD, the level at which significant dissolution of carbonate first becomes apparent is called the lysocline. As originally defined, the term lysocline was used to describe the depth level where a pronounced decrease in the preservation of foraminiferal assemblages is observed. It thus marks a facies boundary separating well-preserved from poorly preserved assemblages on the seafloor. This level is now more specifically referred to as the foraminiferal lysocline to differentiate it from the coccolith lysocline and pteropod lysocline, which may differ in depth because of varying resistance to dissolution or differences in solubility (in the case of the aragonitic pteropods). In addition, it is customary to recognize a sedimentary or carbonate lysocline as the depth at which a noticeable decrease in the carbonate content of the sediment begins to occur.

In theory, the lysocline records the sedimentary expression of the saturation horizon, that is the depth-dependent transition from waters oversaturated to waters undersaturated with respect to carbonate solubility (**Figure 4**). The lysocline thus marks the top of a depth zone, bounded at the bottom by the CCD, over which the bulk of carbonate dissolution in the ocean is expected to occur in response to saturation state-driven chemistry. The
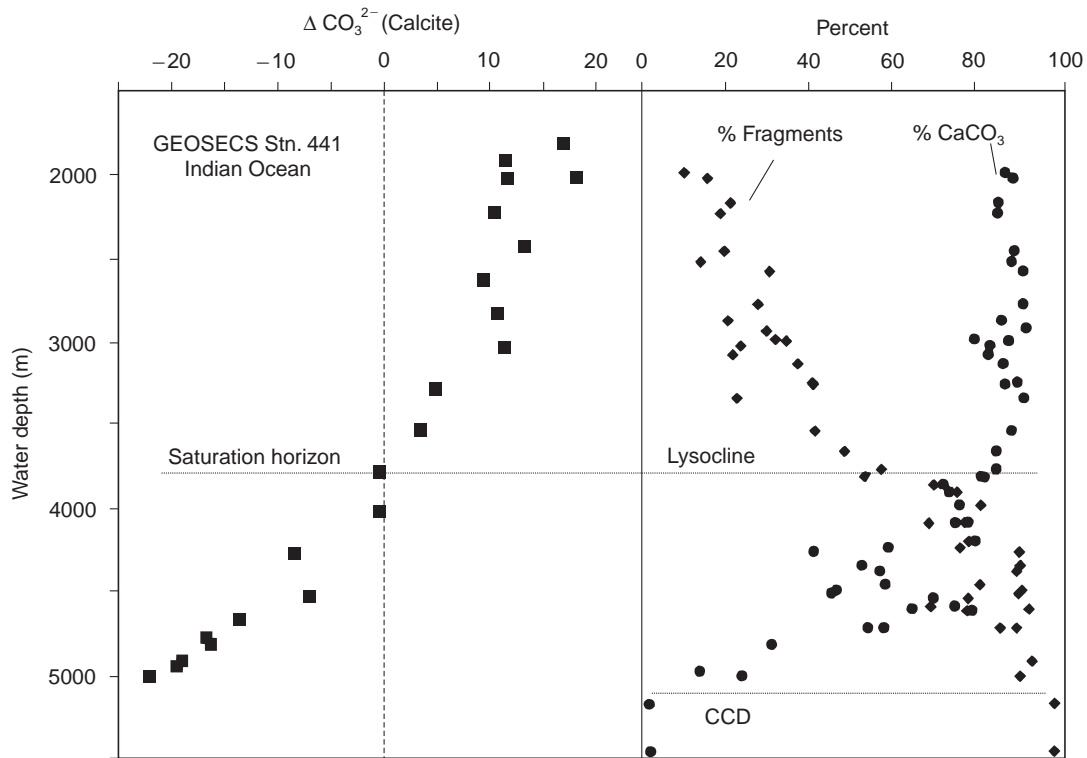
Figure 4 Comparison of carbonate saturation profile for the eastern equatorial Indian Ocean with measurements of foraminiferal fragmentation and carbonate content (weight-%) from depth-distributed modern sediment samples in this region. The saturation horizon with respect to calcite ($\Delta CO_3^{2-} = 0$) occurs locally in the water column at a depth of 3800 m. This level corresponds with both the foraminiferal lysocline and carbonate lysocline as recognized in the sediments. The carbonate compensation depth (CCD) in this region is found at a depth of approximately 5000 m. Increased foraminiferal fragmentation and decreases in sedimentary carbonate content are the result of dissolution and carbonate loss below the lysocline. Carbonate saturation data are from GEOSECS Station 441 (5°2′S, 91°47′E; Takahashi *et al.* 1980); modern sediment data are from Peterson and Prell (1985).

thickness of this sublysocline zone, as indicated by the vertical separation between the lysocline and CCD, is variable and is governed by the rate of carbonate supply, the actual dissolution gradient, and potentially by noncarbonate dilution in certain regions of the ocean.

While the term lysocline was originally used to define a preservational boundary, it has also been used in a fundamentally different sense to denote the depth at which dissolution rates of carbonate on the seafloor greatly accelerate. Whether these levels may or may not coincide, and the nature of their relationship to the saturation horizon or 'chemical lysocline', has been the subject of much discussion and debate. One of the reasons for uncertainty in this regard is the fact that both the carbonate content (%) of a sediment sample and the preservation of the calcareous microfossil assemblages therein can be surprisingly poor indicators of the extent to which dissolution has occurred. For example, the loss of carbonate (L) from sediment, expressed as a weight percentage of the total sediment, is given by:

$$L = 100(1 - R_o/R)$$

where R_o and R are the initial and final values of the noncarbonate (or residual) material. Thus, for a sample initially containing 95% carbonate and a R_o value of 5%, 50% of the carbonate in the sample must be dissolved in order to double the noncarbonate fraction and reduce the carbonate content to 90%. Since the carbonate fraction of the pelagic rain in the open ocean often approaches 95%, this inherent insensitivity means that significant loss of carbonate can occur before detectable changes in the carbonate content are observed. As a consequence, the carbonate lysocline, traditionally defined as the level where the carbonate content of sediments begins to sharply decrease with water depth, may lie deeper than the depth at which significant loss of carbonate to dissolution actually begins to occur.

Dissolution leads to an increase in surface area during the etching of carbonate skeletal material.

Etching produces roughness and widens pores, leading to weakening and ultimately to breakage. Because of their larger size, planktonic foraminifera have usually been the subject of dissolution studies that focus on the preservation state of the microfossils themselves. Planktonic foraminifera have a wide range of morphologic characteristics that enhance their abilities to remain suspended in the upper water column while alive. These same characteristics largely dictate their resistance to dissolution after death. Taxa living in warm, tropical surface waters, where density is generally low, tend to be open-structured with thin shells and porous walls. Taxa that live deeper in cooler, denser subsurface waters, or in colder surface waters at high latitudes, tend to be more heavily calcified with thicker shells and small or closed up pores. On the seafloor, the thin-shelled, more fragile species tend to dissolve more readily than the robust taxa. In effect, this means that individual species each have their own 'lysocline', which can be offset shallower or deeper from the foraminiferal lysocline determined from the total assemblage. There are additional consequences of this selective preservation of taxa that must be considered in paleo-oceanographic or paleoclimatic studies. For example, the selective preservation of more heavily calcified taxa tends to impart a generally 'cooler' appearance to the overall microfossil population and can bias attempts to derive paleotemperature information from seafloor assemblages, as well as other population properties such as diversity.

For carbonate particles produced in the upper ocean, settling rates play an important role in their distribution and preservation. Smaller planktonic foraminifers settle at about $150-250\,m\,d^{-1}$, while larger ($>250\,\mu m$) foraminifers may settle as much as $2000\,m\,d^{-1}$. These rates are rapid enough that little dissolution is thought to occur in the water column. Solitary coccoliths, on the other hand, sink at rates of 0.3 to $\sim 10\,m\,d^{-1}$, slow enough that dissolution within the water column should theoretically prevent their ever reaching the ocean bottom. However, sediment trap studies have shown that transport by fecal pellets is the dominant process by which small phytoplankton skeletons are transferred to the seafloor. Protection offered by the organic fecal pellet covering may also protect the coccoliths after deposition and account for the fact that the coccolith lysocline is generally observed to lie somewhat deeper than the foraminiferal lysocline.

While the seafloor depths of the lysocline and CCD can be readily identified from sedimentary criteria, this information is of limited use without realistic knowledge of the rates at which calcium carbonate is lost from the sediments to dissolution. In practice, it is much easier to determine carbonate accumulation in the deep sea than it is to estimate carbonate loss. Yet the latter information is clearly needed in order to close sediment budgets and to reconstruct changes in the carbonate system.

Carbonate-rich sediments deposited above the saturation horizon should experience little in the way of saturation-driven dissolution because they lie in contact with waters oversaturated with respect to calcite. Nevertheless, evidence for significant supralysoclinal dissolution has been found in a number of studies. Much of this dissolution at shallower water depths is thought to be driven by chemical reactions associated with the degradation of organic carbon in the sediments. Organic carbon arriving at the seafloor is generally respired as CO_2 or remineralized to other organic compounds by benthic organisms. The metabolic CO_2 generated by organisms that live within the sediment can contribute to the dissolution of calcite even above the lysocline by increasing the chemical corrosivity of the pore waters. Studies of organic matter diagenesis in deep-sea sediments suggest that rates of supralysoclinal dissolution vary greatly with location, ranging from minimal loss to $>40\%$ calcite loss by weight. Temporal and spatial changes in the rain rate of organic carbon relative to carbonate can affect this process.

Whether above or below the lysocline, carbonate dissolution is mostly confined to the bioturbated surface sediment layer (typically $\leq 10\,cm$ in the deep sea). As carbonate is depleted from this bioturbated layer, older 'relict' carbonate is entrained from the sediments below. This results in 'chemical erosion' and can produce substantial hiatuses or gaps in the record. Dissolution, and hence erosion, eventually stops when nonreactive materials fill up the mixed layer and isolate the underlying sediment from the overlying water. Many clay layers interbedded within carbonate-rich sequences are likely produced by this mechanism; the resulting lithologic contrasts often show up as subsurface seismic horizons which can be traced for long distances and tell a story of changing dissolution gradients and carbonate chemistry in the past.

Basin-to-Basin Fractionation in the Modern Ocean

Superimposed on the general depth-dependent decrease of carbonate accumulation observed everywhere in the deep sea are preservation patterns that differ between the major ocean basins. Today, carbonate-rich sediments tend to accumulate in the Atlantic Ocean, while more carbonate-poor

sediments are generally found at comparable water depths in the Indian and Pacific Oceans. This modern pattern is largely the product of the ocean's thermohaline circulation and has been termed 'basin-to-basin fractionation'. In the Atlantic, deep and bottom waters tend to be produced at high latitudes because cold temperatures and high sea surface salinities lead to the formation of dense water masses that sink and spread at depth. These young, relatively well oxygenated and $[CO_3^{2-}]$-enriched waters tend to depress the depth of the saturation horizon and allow carbonate to accumulate over much of the Atlantic basin, as manifested by a deep lysocline and CCD. In contrast, neither the Indian nor Pacific Oceans today experience surface conditions that allow deep or bottom waters to form; water masses at depth in these basins largely originate in the Atlantic sector as part of what is sometimes described as the ocean's conveyor belt circulation, with a general upwelling of waters from depth balancing the formation and sinking of deep waters in the Atlantic source areas. Since deep and bottom waters in the Indian and Pacific Oceans are further removed from their modern source areas in the Atlantic, they tend to be CO_2-enriched and $[CO_3^{2-}]$-depleted because of their greater age and the cumulative effects of organic matter remineralization along their flow path. In particular, the *in situ* decrease in $[CO_3^{2-}]$ concentration leads to an increase in undersaturation of the water masses and a progressive shoaling of the saturation horizon (**Figure 3**). Thus, Indian and Pacific deep waters are generally more corrosive to the biogenic carbonate phases than Atlantic waters at comparable depth, the lysocline and CCD are shallower, and a smaller area of the seafloor experiences conditions suitable for carbonate preservation and accumulation. This pronounced modern pattern of basin-to-basin fractionation is illustrated by the fact that roughly 65% of the present Atlantic seafloor is covered by carbonate ooze, while only 54% of the Indian Ocean floor and 36% of the Pacific Ocean floor share that distinction. Naturally, if thermohaline circulation patterns have changed in the past, then carbonate preservation and accumulation patterns will change accordingly. The mapping and reconstruction of such trends has emerged as a powerful paleoceanographic tool.

Temporal Changes in Carbonate Accumulation and Preservation

The patterns of carbonate accumulation and preservation in the deep sea contain important information about the chemistry and fertility of ancient oceans.

Numerous studies have now shown that variations in the carbonate system have occurred on a variety of timescales, both within and between ocean basins. On a local or even regional scale, such variations can often be used as a correlation tool. This has come to be known as 'preservation stratigraphy'.

A number of criteria have commonly been used as indicators of the intensity of carbonate dissolution in deep-sea sediments. Variations in the measured carbonate content of sediments are commonly used to correlate between cores in a region, but are difficult to interpret strictly in terms of dissolution and changing deep-water chemistry. This is because the weight percent carbonate content of a sample can also be affected by changing carbonate input (i.e. surface production) and by dilution from non-carbonate sources. More useful are indices based on some direct measure of preservation state, such as the percentage of foraminiferal fragments in a sample relative to whole shells (**Figure 5**). However, while clearly recording dissolution, preservation-based indices can also be affected by other factors, including ecologic changes that may introduce variable proportions of solution-susceptible species into a region over time.

Because carbonate dissolution is a depth-dependent process, it is best studied where existing seafloor topography allows for sampling of sediments over a broad depth range. Given this sampling strategy, one way to circumvent the problems of using measured carbonate content and other relative dissolution indices (e.g. fragmentation) is to calculate carbonate accumulation histories for the individual sampling locations and examine depth-dependent differences in accumulation rates and patterns. To do so requires an accurate knowledge of sedimentation rates (e.g. cm per thousand years) and measurements of sediment bulk density (in g cm^{-3}), in addition to the data on carbonate content. The product of these three measures yields a mass accumulation rate for the carbonate component expressed in g per cm^2 per thousand years. Differences in accumulation between depth-distributed sites can provide insights into dissolution gradients and carbonate loss.

As the relative importance of calcium supply from weathering and carbonate production vary through time, the depth of the CCD must adjust to control dissolution and to keep calcium levels in balance. Studies of CCD behavior during the Cenozoic (**Figure 6**) have generally shown that CCD fluctuations were similar in the various ocean basins and were likely to have been driven by a global mechanism, such as a change in sea level and/or hypsometry of the ocean basins or a change in supply of calcium
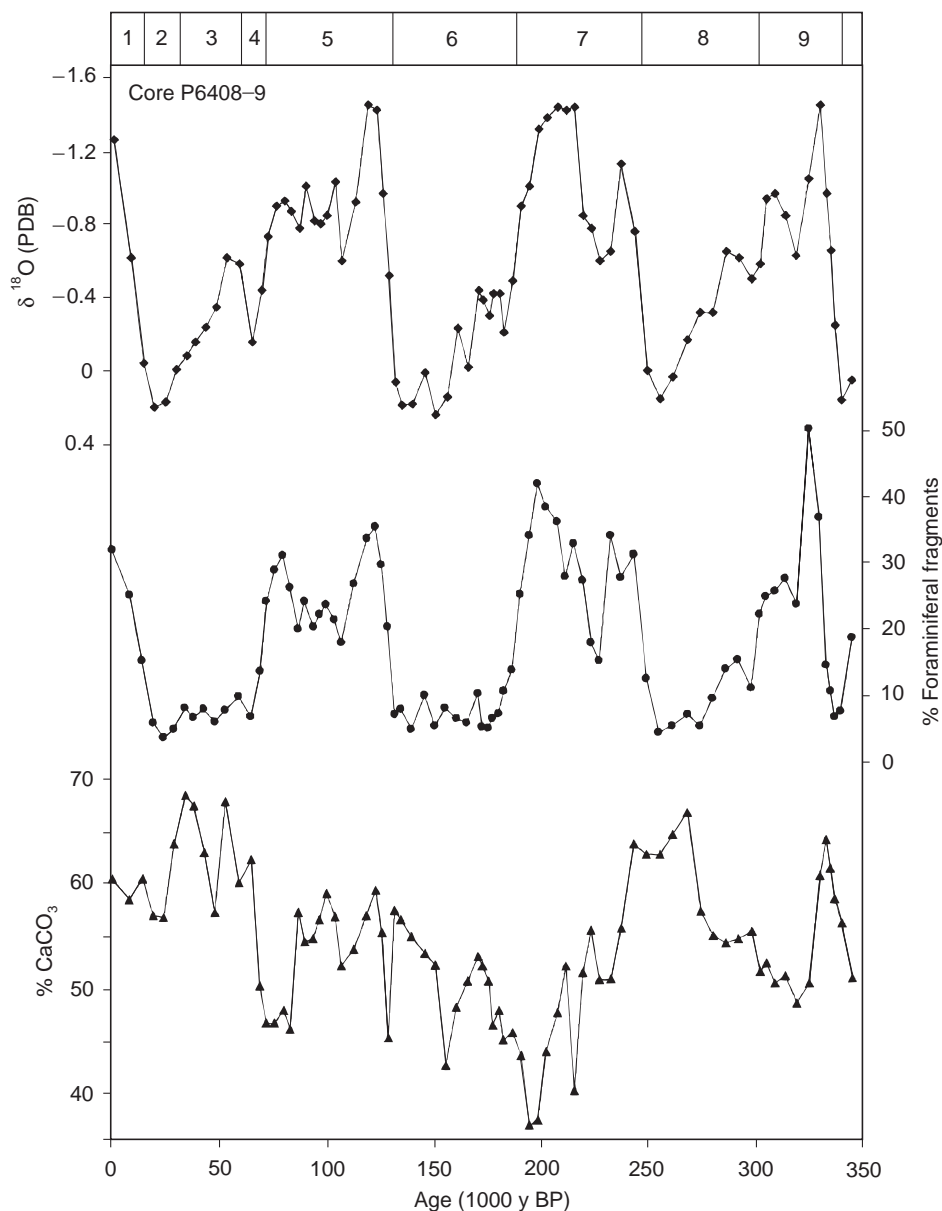
Figure 5 Measurements of foraminiferal fragmentation and calcium carbonate content (weight-%) spanning the last 350 000 years in Caribbean sediment core P6408-9. Stratigraphy and age control come from the oxygen isotope ($\delta^{18}O$) record shown at the top of the figure; odd-numbered stages are warm, interglacial intervals and even-numbered stages indicate cold, glacial climates with greatly expanded Northern Hemisphere ice cover. Variations in the ratio of foraminiferal fragments to whole shells can be directly related to the intensity of carbonate dissolution on the seafloor. Greatly increased preservation (i.e. decreased numbers of fragments) during cold, glacial stages indicates reductions in the chemical corrosivity of deep Caribbean waters in response to climate and ocean circulation changes. Note that variations in carbonate content at this location are not as clearly linked to climate-induced changes in deep-water chemistry as the fragmentation record. This is because the carbonate content of the sediments can also be affected by carbonate productivity at the surface and by dilution on the seafloor by noncarbonate sediment types. (Unpublished data from L. Peterson.)

to the oceans. There are, however, clear ocean-to-ocean differences in this general pattern that are likely to have been the result of changes in regional productivity and the interbasinal exchange of deep and surface waters. By examining such differences, estimates of past circulation and of the relative differences in carbonate productivity in different regions can be determined from regional offsets in the depth of the CCD.

Seafloor Diagenesis

With time and burial, carbonate oozes undergo a progressive sequence of diagenesis and are trans-
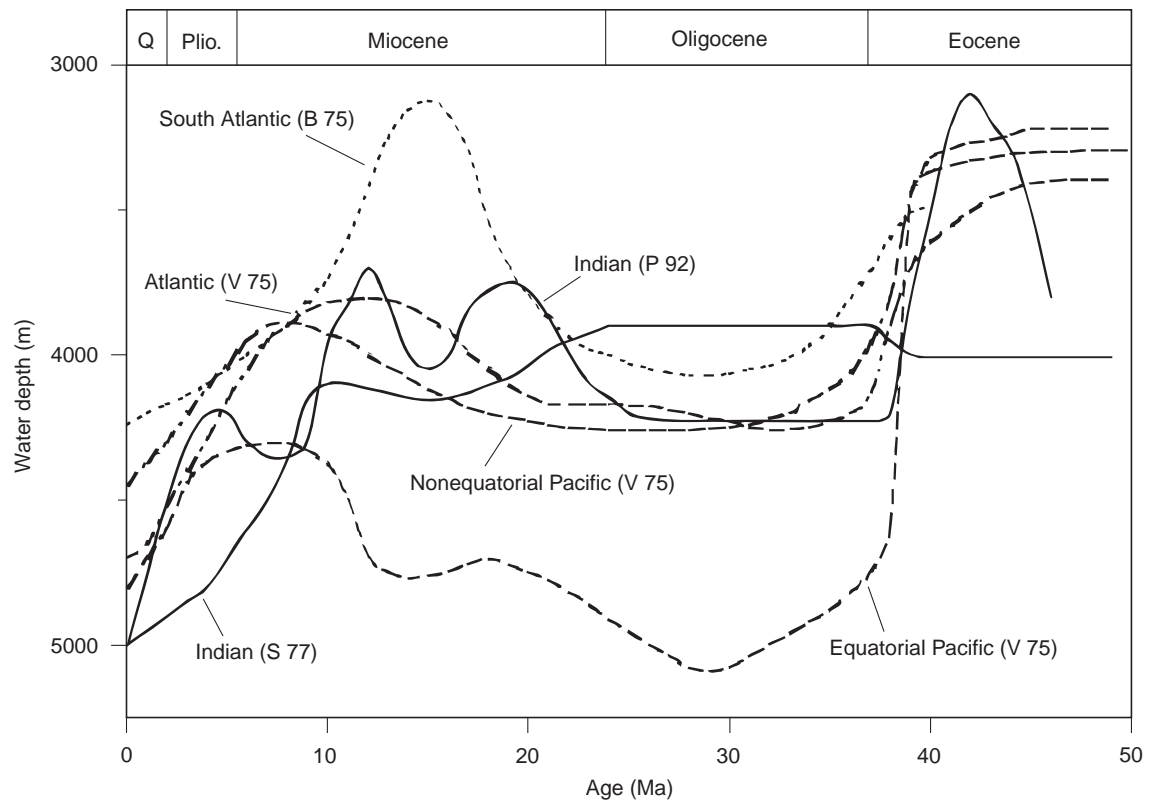
Figure 6 Compilation of reconstructed variations in the depth of the CCD from selected studies covering the last 50 million years for different oceanic regions. The overall similarity of the CCD behavior between regions suggests a common forcing mechanism, such as global sea level or a long-term change in the supply of calcium to the ocean. Variations between the oceans are probably the result of differences in regional surface productivity and deep circulation patterns. Cited CCD studies include: V75, van Andel (1975); B75, Berger and Roth (1975); S77, Sclater *et al.* (1977); P92, Peterson *et al.* (1992).

formed first to chalk and then to limestone through a combination of gravitational compaction, dissolution, reprecipitation, and recrystallization. Porosity is reduced from about 70% in typical unconsolidated carbonate oozes to roughly 10% in cemented limestones, while overall volume decreases by about one-third. Drilling results have shown that the transformation from ooze to chalk typically occurs within a few hundred meters of burial, while limestones are produced by further cementation under about 1 km of burial. Although the transformation of ooze to chalk to limestone is the expected diagenetic sequence, smaller scale reversals in lithification are often observed. Such reversals in pattern have led to the concept of diagenetic potential, which simply states that different sediments will take different lengths of time to reach equal stages of lithification depending upon the original character of the deposited sediment. Such factors as the original proportions of coccoliths to foraminifera (affecting grain size), the amount of dissolution experienced before burial, sedimentation rates, and numerous other subtle factors can influence the dia-

genetic potential of a carbonate sediment. To the extent that these factors reflect original oceanographic conditions, the sub-bottom acoustic reflectors that result from changing lithification state and diagenetic potential preserve a history of paleo-oceanographic events that can often be traced across large regions within ocean basins.

See also

Acoustics in Marine Sediments. Carbon Dioxide (CO$_2$) Cycle. Cenozoic Climate – Oxygen Isotope Evidence. Cenozoic Oceans – Carbon Cycle Models. Coral Reefs. Ocean Carbon System, Modelling of. Plankton and Climate. Pore Water Chemistry. Protozoa, Planktonic Foraminifera. Sea Level Change. Sedimentary Record, Reconstruction of Productivity from the. Thermohaline Circulation.

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CALIFORNIA AND ALASKA CURRENTS

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Introduction

The clockwise North Pacific Subtropical Gyre and the counterclockwise Subarctic or Alaska Gyre, the two principal current gyres of the North Pacific, have dimensions similar to those of the basins, i.e., several thousand kilometers. The West Wind Drift and Subarctic Current flow approximately zonally across the Pacific basin with origins in the Kuroshio and Oyashio Currents, respectively, in the western Pacific basin (**Figure 1**). As this broad eastward flow nears the west coast of North America, it bifurcates, splitting into the northward flowing Alaska Current, the eastern limb of the Alaska Gyre, and the southward flowing California Current, the eastern limb of the North Pacific Subtropical Gyre. This bifurcation takes place several hundred kilometers offshore and depends both on the ocean currents and the wind fields over the North Pacific. The water type is primarily Subarctic, relatively cool and fresh surface water. The volume transport within each of the currents is about 10–15 Sv. Both the California Current and the Alaska Current are coupled to current systems and processes along the adjacent continental margins, where the majority of the seasonal variability occurs. On both seasonal and ENSO (El Niño) timescales, the strengths of the two current systems vary out of phase. For each boundary current system (the Alaska system and the California system) both the large-scale gyres and the coastal current systems, as well as the interactions between them, are described in the text below.

The Alaska Current System

The Alaska Current, the eastern limb of the Subarctic Gyre, flows northward along the west coast of North America beginning at about 48–50°N (**Figure 2**). A companion coastal current flows parallel to the Alaska Current but closer to the coast. This