CARBONATE MINERAL SATURATION STATE AND CRATONIC LIMESTONE ACCUMULATION

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ABSTRACT. Sea-surface temperature/saturation data, sediment accumulation rates from tropical and temperate shallow marine platforms, and paleolatitudes of ancient cratonic carbonate sequences have been compiled to determine the influence of seawater chemistry on the distribution and rate of carbonate deposition. Tropical to subtropical accumulation rates are strongly dependent on the duration of time over which they are measured but, when considered at a time interval of $5 \text{ kyr}$, range to a maximum mean rate of about $2 \text{ m/ky}$ at a latitude of about $20^\circ$ and drop off precipitously between $25^\circ$ and $30^\circ \text{N}$.

Accumulation rates exhibit a functional relation with respect to latitude and sea-surface saturation state and can be described as: $R = k (\Omega - 1)^n$, where $R =$ accumulation rate, $n =$ order of the "process", $k =$ a rate constant, and $\Omega =$ calcite or aragonite saturation. Agreement between calcium carbonate accumulation rate data and rates calculated from heterogeneous-growth laboratory experiments of carbonate minerals suggests that spatial variation in physicochemical parameters of seawater composition is indistinguishable from biologic/ecologic factors in controlling areas and rates of shallow water carbonate accumulation in global oceans.

Rate of shallow water carbonate deposition is strongly dependent on the equator to pole temperature/saturation gradient. Comparison of the latitudinal extent of Cretaceous shallow water carbonates with latitudinal limits of carbonate accumulation in modern settings indicates that Cretaceous limestone deposition occurred to latitudes some $10^\circ$ more poleward than at present. This difference suggests that mean annual Cretaceous seawater surface temperature gradient between the equator and $38^\circ$ north latitude was about $5^\circ \text{C}$ lower (warmer) than that observed in modern oceans.

INTRODUCTION

Areas of platform carbonate generation and accumulation generally correspond to regions of shallow tropical water, far removed from sources of terrigenous input (Wilson, 1975). Given that modern marine carbonate sediment is largely confined to such settings (Hopley, 1982; Milliman, 1974; Scoffin, 1987), the widespread presence of limestone and dolostone in the rock record has led many to examine the supposition that carbonate accumulation has always occurred at tropical to subtropical latitudes. While a uniformitarian perspective of the relation between of low-latitude and carbonate accumulation has been broadly confirmed in the rock record (Blackett, 1961; Briden and Irving, 1964; Opdyke, 1959, 1962; Ziegler and others, 1984), secular variation in the poleward limits of limestone deposition has also been documented (Opdyke and Wilkinson, 1989).
Although finite spatial limits of shallow water carbonate production are generally recognized throughout the geologic record, processes that control the location of these boundaries are not well understood, and considerable question exists as to the relative importance of biologically-versus chemically-mediated factors in predicing suitable areas for carbonate production. This uncertainty partly reflects the fact that depositional styles of shallow water carbonate accumulation are far from uniform. In some areas such as the Bahama Banks, accumulation occurs over wide areas (Ginsburg and James, 1974), while in others, sediment generation is focused around reefal buildups that represent a relatively small percentage of shallow shelf area (Maxwell, 1968). Carbonate accumulation on the Great Barrier Reef, for example, accounts for only 15 to 20 percent of total shelf area along the northeast coast of Australia (Davies and Hopley, 1983), and globally, reefal systems only account for 2 percent (0.6 x 10^6 km^2 of the 27 x 10^6 km^2) of shelf area between 30°S and 30°N lat (Smith, 1978; Keir and Berger, 1984).

Shallow depths, tropical temperatures, and sea-level rise or platform subsidence are generally perceived as requisite conditions for significant carbonate accumulation. Therefore, it is not unreasonable to assume that areas of elevated ambient seawater saturation largely correspond to those areas where carbonate sediment is generated. Conversely, because many marine communities are extremely sensitive to changes in salinity, sunlight, and nutrient and particulate fluxes and concentrations (Bathurst, 1975; Dodge, Aller, and Thomson, 1974; Glynn and Stewart, 1973; Hollock and Schlager, 1986; Lees, 1975; Lees and Buller, 1972; Liang, 1985; Milliman, 1974; Morelock and Koenig, 1967; Roy and Smith, 1972; Scoffin, 1987; Schlager, 1981), one might also anticipate a relation between these environmental parameters and rate of carbonate deposition independent of temperature and saturation state.

Gaining a better perspective on the relative importance of biological versus physicochemical control on areas and rates of carbonate accumulation is important for a number of reasons. For example, shallow water deposits presently comprise about 40 percent of the global Neogene carbonate reservoir (Davies and Worsley, 1981; Hay, 1985; Khain, Ronov, and Balukhovskiy, 1979), but Paleozoic and Mesozoic cratonic deposits accounted for up to 90 percent of total limestone accumulation (Opdyke and Wilkinson, 1989). This difference largely reflects a long-term shift in sites of carbonate accumulation from shallow cratonic to deep (pelagic) oceanic reservoirs and is perhaps the most important global change in elemental flux patterns over this time interval. Given that shallow water carbonate precipitation is an integral part of the global carbon cycle (Keir and Berger, 1984), patterns of carbonate deposition have been used as important proxy indicators of Phanerzoic weathering rate and volcanic input of CO_2 to the atmosphere (Berner, Lasaga, and Garrels, 1983; Owen and Rea, 1985; Raymo, Ruddiman, and Froelich, 1988). In addition, the latitudinal distribution of carbonate deposits may
serve as an important record of paleoclimatological gradients in seawater temperature/saturation state.

In order to evaluate the significance of different environmental parameters in determining rates and areas of carbonate accumulation, surface seawater chemical and temperature data have been compiled for the modern ocean. Comparisons are then made to available data on modern and ancient rates and areas of carbonate accumulation in order to evaluate the relative importance of physicochemical processes in carbonate generation.

TEMPERATURE AND SATURATION STATE

The degree of aragonite saturation of surface sea water was calculated from the Geochemical Ocean Sections Study (GEOSECS) and Transient Tracers in the Ocean (TTO) data of Bainbridge (1981), Broecker, Spencer, and Craig (1982), Spencer and others (1983), Takahashi and others (1980), and Takahashi, Chipman, and Volk (1982). These sources include temperature, alkalinity, $P_{CO_2}$, and $CO_3^2-$ data necessary to calculate aragonite saturation ($\Omega_{arg}$), defined as $[CA^{2+}][CO_3^{2-}]/K_{arg}^*$. Surface water $Ca^{++}$ concentrations were assumed to be conservative with salinity (Culkin and Cox, 1966), and stoichiometric solubility constants for aragonite ($K_{arg}^*$) (Mucci, 1983) were evaluated at appropriate values of temperature and salinity.

On the basis of these measurements, it is apparent that surface ocean values of temperature and saturation state are positively correlated (fig. 1), and that each exhibits systematic variation with latitude (fig. 2). This is due to the fact that $P_{CO_2}$ equilibration is nearly achieved everywhere (Broecker and Peng, 1982). In equatorial regions ($0^\circ - 15^\circ$) saturation values are depressed relative to this correlation (fig. 2), but toward higher latitudes, temperature and carbonate saturation decrease more or less linearly. Lower saturation state in equatorial regions is due to higher atmospheric precipitation, lower rates of evaporation, and, in some areas, higher concentration of $CO_2$ related to upwelling of deeper marine water (Broecker and Takahashi, 1984).

ACCUMULATION RATE

To assess relations between ambient seawater saturation state and carbonate deposition, accumulation rates were tabulated from published investigations that employed $^{14}C$ and/or U/Th dating techniques. Studies that used values of sediment thickness and duration of accumulation determined from local sealevel curves relative to elevation of the underlying Pleistocene unconformity were utilized as well. This tabulation of 220 rate values includes a wide variety of facies including temperate and tropical sediment, as well as reefs, ooid shoals (Ginsburg and James, 1974; Martin and Ginsburg, 1965), and carbonate mud.

A striking aspect of these data is that the most important factor influencing accumulation rate is neither latitude nor method of rate
Fig. 1. Temperature and aragonite saturation values for Atlantic, Pacific, and Indian Ocean surface waters calculated from GEOSECS and TTO data in Bainbridge (1981), Broecker, Spencer, and Craig, (1982), Spencer and others (1983), Takahashi and others (1980), and Takahashi, Chipman, and Volk (1982). $R^2$ for this regression is 0.76.

determination nor sediment type. Rather, accumulation rate ($A$) is most strongly dependent on section duration ($D$), where:

$$\log A = -0.73 \log D + 1.40$$

In general, such a relation is typical of most sedimentary sequences (Sadler, 1981) and most commonly reflects decreasing variation in hialal time/depositional time ratio with increasing time interval over which accumulation rate is determined (fig. 3). Conceptually, rate of change in the elevation of any sediment surface can be considered as being distributed about a mean value intrinsic to that particular setting, with some average thickness/time change occurring most commonly and with “rarer” erosion/deposition events occurring less frequently (Hsu, 1983). Therefore, for any series of arbitrarily-chosen sequences such as those represented in this tabulation, thinner intervals might incorporate either greater or lesser amounts of depositional or hialal time, while thicker intervals should more closely approximate the mean erosion/deposition rate specific to that particular setting. In this context, only time intervals that represent some finite amount of accumulation are preserved (albeit incompletely) in any sequence. Operative limits of precision in determining accumulation/duration values for the thinnest of these also precludes
Fig. 2. Sea-surface temperatures and aragonite saturation from GEOSECS and TTO data as a function of latitude. Trends defined by best-fit second order polynomials; vertical lines are one standard deviation about mean values at 5° lat. (A) Note increase in temperature standard deviation with latitude and little difference between mean values at 0° and 10°. (B) Note latitudinal saturation maximum near 20°.
Fig. 3. Holocene carbonate accumulation rates (n = 220) in platform settings. (A) Accumulation rate relative to time duration of accumulation. (B) Log accumulation rate versus log duration for various carbonate facies (filled symbols) and 5 ky duration values (A5) calculated from duration (Dr) and rate (Ar) data as: log A5 = (log Ar) + (0.73 log Dr) + 1.40 as open symbols.

their inclusion in any tabulation (Anders, Krueger, and Sadler, 1987). Hence, time duration/accumulation rate data commonly exhibit an apparent logarithmic distribution.

Because it is evident from Holocene data and theoretical considerations that accumulation rate is strongly dependent on the duration of time over which accumulation is measured, it is therefore necessary to adjust rates determined over different durations to values representing some reference interval of time when attempting to arrive at a typical Holocene accumulation rate (fig. 3). If all Holocene thickness/age data are considered at a 5 ky duration, a typical length of time for the accumulation of many late Holocene sections, they yield a mean accumulation rate of about 1.1 m/ky. In addition, such duration-normalized rates exhibit comparable values regardless of the lithology of sediment under consideration; carbonate sand, mud, and reefal components exhibit a similar range of thickness and accumulation values.

When considered as a function of latitude (fig. 4), duration-normalized rates increase from a mean equatorial value of about 1 m/ky to a maximum of about 2 m/ky between 10° and 25° and then decrease
rather abruptly toward temperate regions. Throughout, minimum accumulation rates approach zero, and poleward variation is therefore only defined by higher rates at any particular latitude. On the basis of these data, rate of carbonate deposition in Holocene settings is greatest between latitudes of 15° and 20° and decreases both poleward and toward equatorial regions, though more abruptly toward temperate latitudes.

Such poleward decrease is similar to, but more dramatic than, that indicated by latitude/thickness curves of McKenzie, Bernoulli, and Schlanger (1980), Schlanger and Konishi (1975), and Scoffin (1987), who inferred a more uniform decline toward higher latitudes. When data on composition of carbonate facies from Schlanger and Konishi (1975) are plotted relative to mean accumulation rate, it is apparent that patterns of deposition are directly or indirectly related to the taxonomic affinities of the dominant calcifying groups in those areas (fig. 5). In addition, the pattern of change in accumulation rate with latitude is very similar to that defined by saturation values (fig. 2B), suggesting that rate of carbonate generation may be more closely linked to physicochemical factors such as ambient temperature than to biologic factors related to ecologic demands of various calcifying taxa that inhabit areas of carbonate sediment deposition. In order to examine further relations between physicochemical parameters and rates of carbonate generation, it is informative to com-

![Graph](image)

*Fig. 4.* Holocene carbonate accumulation rate data (calcite and aragonite) based on $^{14}$C and U/Th dates and on section thicknesses, here normalized to a 5 ky interval duration. Vertical lines are one standard deviation about mean accumulation values over 5° lat (horizontal lines). Note that highest rates occur between about 10° and 25°.
Fig. 5. Mean accumulation rates from figure 4 relative to data on dominant facies from data in Schlanger and Konishi (1975). Abrupt poleward decrease in deposition rate generally corresponds to an absence of reefal systems, a decline in the abundance of skeletal aragonite, and an absence of abiotic components as ooids and marine cement.

pare actual accumulation rate values with rate data from laboratory experiments that relate rate of abiotic precipitation to temperature and saturation state.

**PRECIPITATION RATE**

In laboratory systems, temperature, saturation state, and precipitation rate ($R$) are commonly related by the empirical expression:

$$ R = k (\Omega - 1)^n $$

where $k$ is the rate constant, $\Omega$ is saturation state defined as the equilibrium Ion Activity Product (IAP) divided by the solubility product $K_{sp}$ for the phase in question, and $n$ is the order of the reaction (Burton and Walter, 1987; Morse, 1983; Mucci and Morse, 1983). Because $\Omega = 1$ represents equilibrium, $\Omega - 1$ quantifies degree of solution super saturation. The expression:

$$ \log R = \log k + n \log (\Omega - 1) $$

is the linear form, where the order of the reaction ($n$) is the slope and $\log k$ (the rate constant) is the intercept; different expressions have been determined for calcite and aragonite (Burton and Walter, 1987).

Because here we are primarily concerned with differences in accumulation rate as a function of typical latitudinal values of temperature and
saturation state, the pivotal question is: to what degree does variation in accumulation rate calculated from laboratory experiments agree with latitudinal variation in measured accumulation rates? Given temperature and saturation state data of the global oceans (fig. 1), the above equation can be used to calculate anticipated changes in carbonate accumulation rate from low to high latitudes and to evaluate relations between physico-chemical parameters (T, Ω) and carbonate accumulation in natural settings.

Obviously, any appropriate value of k that relates carbonate accumulation rate data to ambient saturation state and temperature in natural systems will be dramatically different from that used in relating parameters from laboratory studies because Holocene rates have been tabulated as meters per kiloyear and not as micromoles per square meter of seed crystal surface per hour as is the convention in laboratory systems (Burton and Walter, 1987; Mucci and Morse, 1983). Also biotic carbonate precipitation is often locally enhanced by microscale supersaturation due to photosynthesis.

Moreover, biotic and abiotic carbonate components exhibit significant change in taxonomic makeup, composition, and mineralogy with increasing latitude. Lees (1975) divided the continuum of temperature-to-equatorial sediment types into foramol, chlorozoan, and chloralgal assemblages based largely on taxonomic affinities of contained shelly debris. Foramol sediment is typical of temperate shallow shelves and is dominated by benthic forams, molluscs, barnacles, bryozoans, and calcareous red algae, while more equatorial chlorozoan assemblages are dominated by the important reef building taxa and calcareous green algae, although they also contain some elements of the foramol assemblage. Chloralgal assemblages, on the other hand, accumulate under conditions of extreme temperature and/or salinity (>37 ppt and/or 32°C), contain few corals, and are generally dominated by green algae. Importantly, Lees (1975) has also shown that there are distinct temperature-salinity fields in which each of the three assemblages dominate, thus implicating a role of water temperature and composition in determining both sites and taxonomic compositions of cratonic carbonate.

Latitudinal change in the dominance of different calcifying groups is also generally coincident with a change in the mineralogy and composition of carbonate components. Low-latitude sediment is commonly dominated by magnesian calcite and aragonite, while in temperate regions calcite-rich sediments contain less magnesium, and fewer aragonitic components. More equatorial facies also incorporate inorganic precipitates such as ooids and marine cement that are usually absent at higher latitudes (Pilkey and Blackwater, 1968; Scoffin, 1987). Transition from mostly aragonite to primarily low-Mg calcite sediment roughly corresponds to the temperature minimum that reef-building coral communities can withstand; Grigg (1982) called this ecological limit the “Darwin Point,” near 18°C in the Northern Pacific.
If values of seawater saturation from GEOSECS and TTO data are put into the rate equation while employing interpolated values of \( n \) and \( k \) from data in Burton and Walter (1987) at sea-surface ambient temperatures (\( n_{\text{calcite}} = 0.6, 1.9, 2.3 \), and \( k_{\text{calcite}} = 14.0, 3.9, 3.7 \) at \( T = 5^\circ, 25^\circ, \) and \( 37^\circ \), respectively; \( n_{\text{aragonite}} = 0.4, 1.7, 2.3 \), and \( k_{\text{aragonite}} = 21.8, 40.6, 45.1 \) at \( T = 5^\circ, 25^\circ, \) and \( 37^\circ \), respectively) calculated aragonite precipitation rates are significantly higher than those for calcite at low latitudes, and rates for both polymorphs converge in a poleward direction (fig. 6A).

Moreover, good agreement between measured accumulation rates and those anticipated from laboratory experiments results when sediment thickness/duration accumulation rates are about 200 times less than mole/area/time precipitation rate values reported by Burton and Walter (1987) and when the aragonite-to-calcite transition is assumed to occur at about 30\(^\circ\) lat (fig. 6B). Perhaps not surprisingly, changes in precipitation rate with latitude anticipated from relations between \( \Omega \), \( T \), and \( n \) as determined in laboratory settings yield generally lower equatorial and temperate region values, with maximum accumulation at a latitude of about 20\(^\circ\). Given obvious differences in the nature of laboratory and natural systems, these results are remarkably similar and further implicate the importance of seawater composition in determining sites, rates, and compositions of carbonate sediment deposits.

There are several reasons for some of the discrepancy between empirically-predicted and measured accumulation rate data, particularly in temperate regions where measured rates are uniformly lower than those anticipated from seawater compositions. First, because the taxonomic composition of communities of calcifying organisms responsible for most carbonate production changes as a function of latitude and temperature (fig. 5), these differences may result in differing styles of deposition. Much low-latitude carbonate accumulation, for example, is associated with coral reefs that form a sediment-trapping framework. Because temperate regions lack framework building organisms, the poleward extent of such taxa may be of considerable significance in defining the global variation of shallow water calcium carbonate accumulation rates. Moreover, many taxonomic groups are most sensitive to extremes of low temperature near their latitudinal boundaries (below 16\(^\circ\)C near 30\(^\circ\) lat for corals; Emery, 1968; Wells, 1957). Where reefal systems are absent, carbonate sediment components are more susceptible to lateral transport by wave and other currents to deeper water (Smith, 1972), and a measured deposition rate at any high energy locality would therefore be lower than predicted. Also, the composition of normal marine waters will evolve with passage over shallow carbonate banks (Morse, 1984).

In spite of these complicating factors, variation in modern rates of carbonate accumulation with latitude exhibit good agreement both with poleward changes in temperature/saturation state and with variation in precipitation rate expected from relations determined in laboratory settings. In other words, equator to pole surface water temperature/
Fig. 6. (A) Sea-surface temperatures and theoretical aragonite/calcite precipitation rates calculated from: \( R = k (\Omega - 1)^n \), where \( R \) = precipitation rate, \( k \) = a rate constant, \( \Omega \) is saturation state, and \( n \) is order of the reaction (Burton and Walter, 1987; Morse, 1983; Mucci and Morse, 1983). (B) Measured and theoretical accumulation rates calculated from the rate equation when assuming aragonite dominance at latitudes less than 30° and sediment accumulation rates about 1/200th of precipitation rate values reported by Burton and Walter (1987).
saturation gradients are relatively well recorded by the occurrence of shallow water carbonate sediment. If accumulation rate is indeed related to ambient seawater composition, given the linear nature of temperature and saturation state, past latitudinal limits of carbonate accumulation may ultimately be employed to ascertain past gradients of sea-surface temperature/saturation state. In this regard, the lower precipitation rate/latitude gradient for calcite suggests that low-magnesium calcite components such as pelagic debris may not exhibit a significant change in accumulation rate as a function of temperature/saturation gradient (Fig. 6).

**PHANEROZOIC LIMESTONE**

To evaluate Phanerozoic patterns of platform carbonate accumulation, names, ages, and geographic locations of 2568 type stratigraphic columns were tabulated from data in Kerhoer (1966, 1970) for all named Phanerozoic “Chalk,” “Limestone,” and “Dolostone” sections within the United States. Present localities were then rotated to paleolatitudes of deposition using appropriate paleopoles from data in Van der Voo (1989).

When section paleolatitudes are plotted as a function of time it is apparent that limestone deposition occurred over much of the area now occupied by the United States during most of the past 550 my, as this region lay at relatively low latitudes during the majority of Phanerozoic time. Exceptions to this observation include the Quaternary and the Permo-Triassic when North America was generally emergent. Over the past 150 my, however, a significant portion of the contiguous United States extended northward of 30°, and, during part of this time interval, many Cretaceous limestone and dolostone sequences were deposited well beyond latitudinal limits of modern shallow water carbonate accumulation.

Moreover, while much of this Cretaceous cratonic carbonate is pelagic in origin (Comanche Series), a significant portion also includes facies (coquinas, oolites, rudist reefs) that indicate shallow water deposition at latitudes up to 38°, a latitude about 10° beyond the present northern limits of carbonate accumulation along the Florida coast and about 5° north of modern reefal sequences in Bermuda. If temperature/saturation conditions requisite for shallow water carbonate accumulation during the Cretaceous were similar to those defined by modern limits, then greater latitudinal extent of Cretaceous carbonate must either record higher values of seawater saturation at poleward latitudes and/or lower sea-surface temperature gradients. Given that higher atmospheric pCO$_2$ most likely existed during the Cretaceous (Berner, Lasaga, and Garrels, 1983), holding Cretaceous seawater temperature/saturation relations equal to those in modern oceans yields a conservative valuation of the Cretaceous equator to pole temperature gradient because, other factors being equal, raising pCO$_2$ will lower saturation state. While it is not clear whether higher pCO$_2$ was caused by increased volcanic outgas-
ing or by a change in weathering due to higher sea levels, it is likely that higher temperatures and latitudes of carbonate deposition were necessary in order to balance enhanced weathering fluxes of calcium and carbonate ions to global oceans having higher pCO₂ and a shallower CCD. Other potential changes in ocean chemistry, such as Mg/Ca ratio, carbonate alkalinity, and/or sulfate concentration notwithstanding, it is informative to contrast patterns of Cretaceous and Holocene calcium carbonate deposition in order to document probable changes in temperature/saturation state gradients.

Data indicate a 10° latitudinal reduction in areal extent of carbonate accumulation between Cretaceous (fig. 7A; 38° lat) and Holocene (figs. 4 and 7B; 28° lat) seas. Temperature/saturation relations in modern marine settings (figs. 1 and 2) yield an omega (aragonite) value of about 3.9 and a temperature of about 22°C at the northern limits of shallow water carbonate deposition (fig. 8). Projection of these values to 38° north lat requires considerable flattening of both saturation and temperature

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**Fig. 7.** Paleolatitudes of Phanerozoic carbonate type sections in the continental United States from data in Kerher (1966, 1970) and Van der Voo (1989). (A) Supplied area represents the latitudinal extent of the cratonic U.S. over the Phanerozoic. Horizontal dashed lines represent present latitudinal limits of significant carbonate accumulation at about 27° lat. (B) Detail of type section latitudes for the past 180 my showing the distribution of deeper-water pelagic/marly sections (squares), shallow-water sequences, shelly/reefal/ooolitic (triangles), and sections for which water depth cannot be confidently determined (diamonds). Note that during much of the Cretaceous, reefal/ooolitic carbonates accumulated at latitudes up to about 38°.
Fig. 8. Cretaceous saturation/temperature relations determined from modern sea-surface data and from paleolatitudes of Cretaceous shallow-water sequences. (A) Present latitude-saturation profile (solid line) and that inferred for the Cretaceous (dotted line) wherein modern saturation limits of significant carbonate accumulation (3.9) are extended to poleward limits of Cretaceous carbonate accumulation at about 38°. Higher latitudes of Cretaceous carbonate suggest a flattening of the seawater saturation gradient (stippled area) by about 0.35 at 38°. (B) Present (solid line) and Cretaceous (dotted line) latitude-temperature profiles. Paleolatitudes of Cretaceous sections indicate flattening of the seawater temperature gradient (stippled area) by about 5° at the northern latitude of Cretaceous carbonate accumulation. (C) Modern zonally averaged surface temperatures (solid line) in comparison with Cretaceous temperatures from Barron and Washington (1985). Note similarity between these estimates and those suggested by seawater temperature-saturation and carbonate section paleolatitude data.
gradients and, at 38° north lat, yield Cretaceous sea-surface saturation states and temperatures minimally 0.35 higher and 5° warmer than today (fig. 8).

Moreover, these results are attained assuming no secular change in low-latitude seawater composition or temperature. In fact, abundant data now suggest that, unlike carbonate in modern settings, little aragonite precipitated from equatorial Cretaceous seawater, and that this lack of primary aragonite as ooid cortices and marine cement may reflect higher atmospheric pCO₂ and lower seawater saturation state at Cretaceous low latitudes (Sandberg, 1985). If so, Cretaceous saturation/temperature gradients were probably even lower than those inferred from data on the distribution of cratonic carbonate sequences.

Two potential criticisms inherent in the data base used for these estimates merit some comment. The most apparent is that stemming from the use of open marine seawater compositions and temperatures in conjunction with platform carbonate accumulation rates and distributions to determine seawater saturation and temperature gradients for Cretaceous oceans. While temperatures and compositions in these two settings are probably dissimilar at any given latitude, justification for such an approach only requires some systematic relation between variation in rate of shelf accumulation and variation in surface seawater composition over the range of latitudes in question. In other words, we assume open ocean surface waters will evolve in a similar manner chemically over carbonate platform depositional areas given the same initial saturation/temperature values, whether the platform is Cretaceous or Holocene in age. What we hope to sample are specific differences in the equator to pole saturation/temperature gradients, by comparing the limits of deposition of similar facies.

Second, it should be noted that many of the highest latitude carbonate sections in the United States (fig. 6B) were deposited along the cordilleran Cretaceous seaway. Particularly along the most northern and more laterally-restricted extent of this water body (Kansas, Dakotas, Montana), it is unclear if normal marine conditions persisted during the accumulation of these primarily pelagic carbonates. Southward toward Texas, Oklahoma, New Mexico, and Arizona, however, many higher-latitude sequences contain rudist reefs and oolite that record the temporal persistence of normal marine conditions. Moreover, the lower Cretaceous temperature gradient determined from seawater composition and carbonate accumulation data (fig. 7B) is strikingly similar to that reported by Barron and Washington (1985) who determined that mid-Cretaceous northern hemisphere mean annual surface temperatures were 15° to 25°C warmer than at present, whereas equatorial temperatures were the same or only slightly warmer (5°) than at present (fig. 8C). Based on agreement between these two estimates of Cretaceous paleotemperature, latitudinal distributions of Phanerozoic carbonate sequences may well provide significant data for the reconstruction of paleoclimate during other intervals of geologic time.
In conclusion, tabulated shallow water calcium carbonate deposition rates average 1.1 m/ky for the late Holocene. Accumulation rates drop precipitously at latitudes between 25° and 30°. Given that temperature and carbonate mineral saturation state are closely correlated in the global ocean, average shallow water carbonate deposition can be parameterized using the rate equation \( R = k (\Omega - 1)^n \). By demonstrating that global average carbonate deposition can be described as an organized relation essentially as a function of temperature/saturation state, the high latitude limits of massive limestone buildups can be utilized as a tracer for paleotemperature gradients. This relation should prove useful to those geoscientists who wish to model carbon cycles at Phanerozoic time scales.

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