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The role of salinity in circulation of the Cretaceous Ocean

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With 3 figures and 1 table in the text

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Abstract: The density of seawater is a complex function of temperature, salinity, and pressure. Because of the non-linearity of the equation of state of seawater, the densities of seawaters having the same temperature and the same salinity differences (with respect to the mean salinity of the ocean) will vary with the mean salinity of the ocean. Although this strange property of seawater is evident in a plot of the equation of state, it has never been considered in trying to reconstruct ancient ocean circulation. These differences in the density field may have caused the ocean to respond differently to atmospheric forcing in the past. The different response may hold the key to understanding "ocean anoxic events" and episodes of large-scale burial of organic carbon and production of petroleum source rocks.

Introduction

HOLSER et al. (1980) made a first attempt to track ocean salinity back through time, taking salt extractions into account. They came to the unexpected conclusion that the Cambrian ocean probably had a salinity of about 48 ‰ and that the ocean has been getting less saline throughout the Phanerozoic. A detailed history of salinity was ot presented, because they had no idea of how to reconstruct river delivery of salt to the ocean in the past and could only guess that it was somehow related to the erosion of previously buried evaporites. It is easy to estimate the effect of young evaporite extractions on lowering the salinity of the ocean, but because of the supply of salt from erosion of ancient evaporite deposits, the problem of estimating ocean salinity becomes difficult for more ancient times.

In studies of pre-Quaternary paleoceanography, most workers have simply disregarded the mean ocean salinity problem by assuming the ocean

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has always had a mean salinity of 34.7 ‰, the same as today (RAILSBACK et al. 1989). Others have made a correction for the present mass of fresh water in glacial ice on Antarctica and Greenland (SHACKLETON & KENNETT 1975, DUPLESSY et al. 1991, 1993), assuming a global mean salinity of 34.03 ‰, or more complex schemes taking into account the gradual buildup of Antarctic ice (ZACHOS et al. 1994). The extractions of salt from the ocean to form evaporite deposits have been disregarded, although it has been recognized that they must have a significant effect on ocean salinity (SOUTHAM & HAY 1981).

The effect of mean ocean salinity differences on circulation

ROOTH (1981) called attention to the fact that the mean salinity of the ocean has profound implications for the behavior of the ocean. To understand the implications of differences in mean ocean salinity for the themohaline circulation, one need only consider what would happen if the salinity of the ocean were significantly lower than it is today. Fig. 1 is a graphical representationof the equation of state for seawater (MILLERO & POISSON 1981). Below a salinity of 27.4 ‰, the maximum density of seawater lies above the freezing point, and it behaves like fresh water in that the coldest water will float. This means that the polar regions, which usually have lower than average salinities, become excluded as major sites of deep water formation even if they are very cold. In our preliminary Campanian ocean simulation, all of the Arctic is excluded as a potential site of deep water formation for this reason. It is easy to envision that all of the cooler areas of the planet might be covered by lower salinity water and excluded from deep water formation.

Another peculiarity arises when salinities are significantly higher than at present. Note that between salinities of 30-35 the density curve becomes almost vertical at temperatures of 0 °C and below. The cause for this is easy to see in the left part of Fig. 1. The density lines become vertical at the point of maximum density. In seawater, the maximum density lies below the freezing point. Becasue at S = 30-35, the maximum density is just below the freezing point, the temperature change as seawater cools from 0 °C to the freezing point (about -2 °C) has almost no effect on its density. Today, density changes in seawater below 0 °C are dominantly a function of salinity, and consequently high-latitude bottom water formation takes place where the water is saltiest. However, Fig. 1 shows that when the salinity reaches 40 ‰, the density curve never approaches the vertical, but slopes down to the freezing point. The densest water in the ocean will be the coldest water. This has profound implications for formation of high-latitude bottom waters in the past.

Today the thermohaline circulation is driven by minute density differences (HAY 1993) and this has undoubtedly been true in the past. Changing mean ocean salinity implies significant changes in the thermohaline circulation.

A new simulation of the Late Cretaceous (Campanian) climate (DeCONTO 1996, HAY et al. in press) has been used to drive a prototype ocean circulation model. The preliminary results, showing temperature and salinity of the ocean surface, are shown in Fig. 2. The heat transport generated by the ocean model is close, but not exactly equal to that

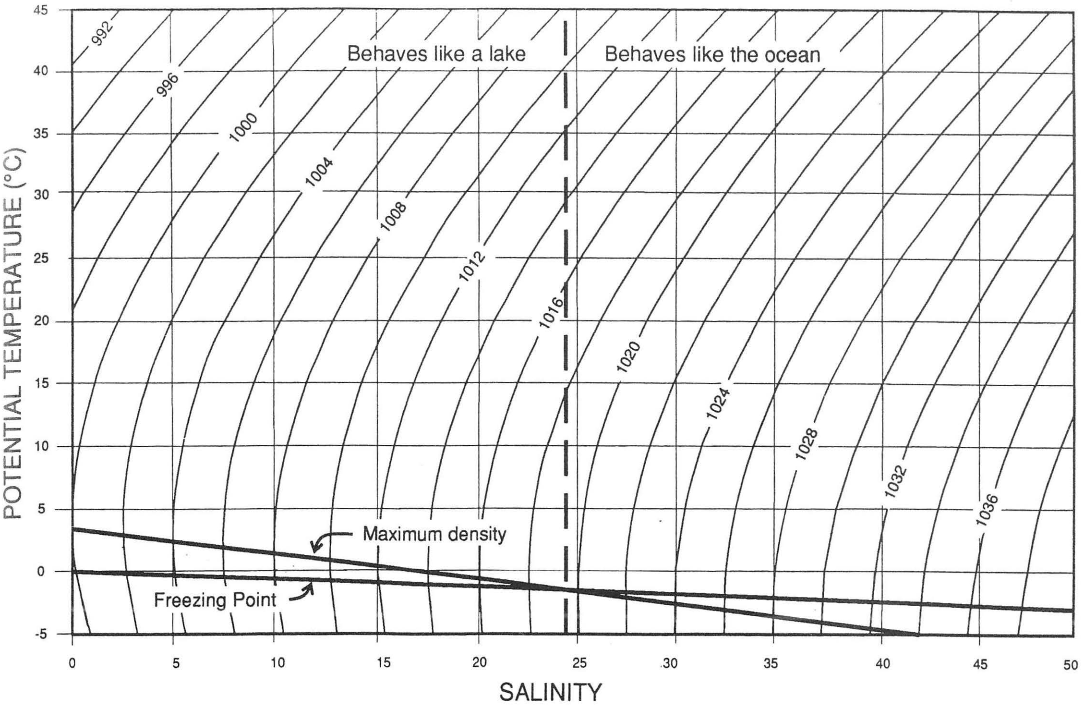

Fig. 1. Graphical representation of the equation of state of seawater 1980, relating temperature, salinity, and density at the sea surface (MILLERO & POISSON 1981). The curved lines are densities (kg/m^3). To the left of the vertical dashed line, the maximum density of seawater is above the freezing point. A temperature change at constant salinity is a vertical line. The different effect of temperature changes with different salinities is evident if one compares the slope of the density curves relative to a vertical line at different salinities. The differences are quite obvious if one compares the slopes of the density curves at salinities of 0 and 50. The differences are more subtle with lesser salinity changes, but they are inherent in the T-S-ρ relationship.

specified for the climate simulation. The salinities shown in Fig. 2A were generated assuming the mean ocean salinity to be 34.8 ‰. The most striking features of this preliminary ocean simulation are the salinity contrasts from place to place. At the present the greatest salinity contrast in the open ocean is 2 ‰, between the North Atlantic, where the high salinity waters of the gyre center are >37.5 ‰, and the North Pacific, where the gyre center waters are about 35.5 ‰. In the preliminary Cretaceous simulation, the contrast between the South Pacific and North Pacific gyre centers is 4.5 ‰, from 36.5 to 32 ‰, and the South Atlantic has salinities that range up above 40, while much of the Arctic has salinities well below 30 ‰. These great salinity contrasts in the Cretaceous simulation are the result of the greater evaporation and precipitation rates inherent in the higher global temperatures. They imply a much greater role for salinity in modifying the surface ocean circulation than is the

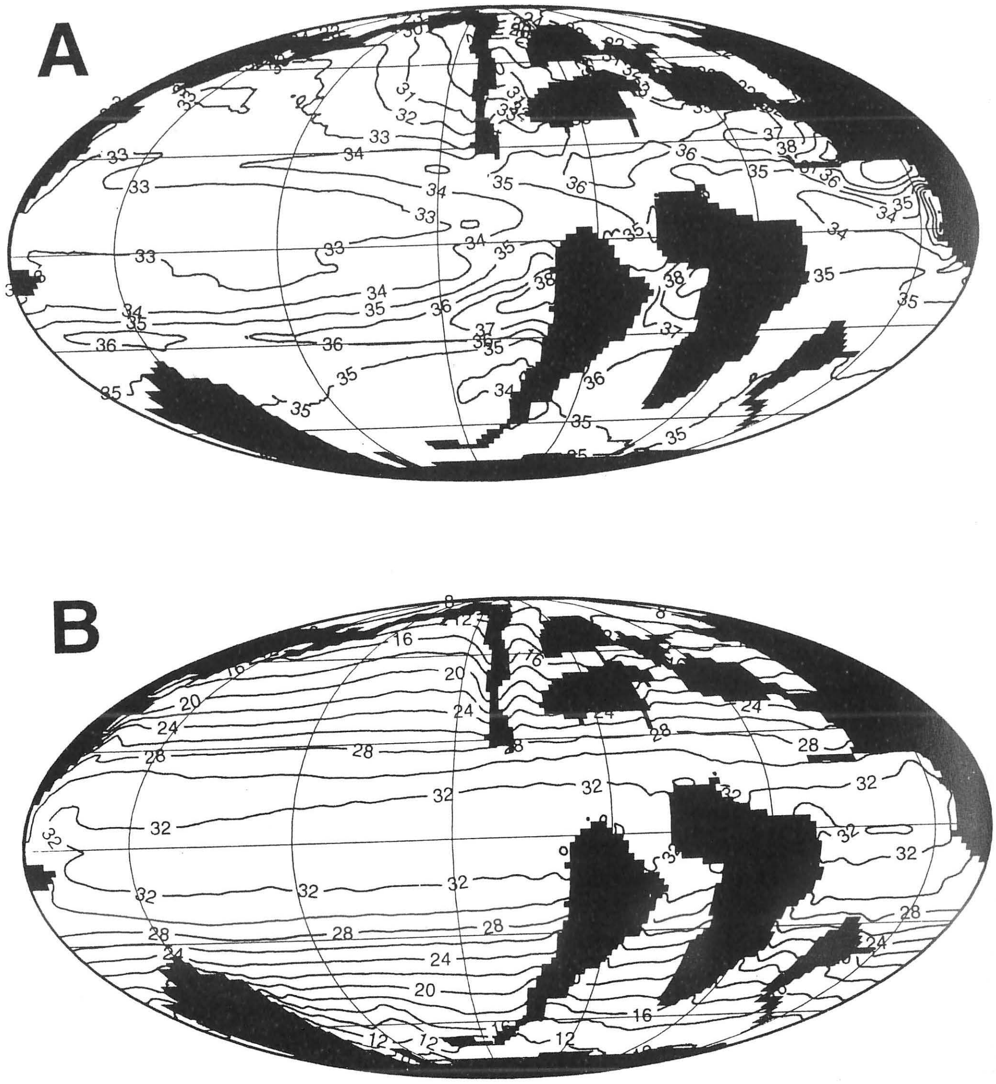

Fig. 2. Results of the GENESIS (version 2) Earth System Model's prototype ocean general circulation model simulation of the Campanian (80 Ma; Late Cretaceous). Mean ocean salinity assumed to be 34.8 ‰. A: Surface water salinities, contour interval 1 ‰. B: Surface water temperatures, contour interval 2 °C. Land areas are shown in black.

case at present. They also suggest that interior water masses might have formed differently in the past.

The mean salinity of the ocean in the past

Table 1 gives estimates of the volumes and masses of water, masses of salt, and paleosalinities of the ocean at a number of critical times in the history of the ocean. The masses of salt in the ocean in the past

Table 1. Preliminary estimates of ocean salinity during the Cenozoic and Cretaceous.

VOLUMES	(m^3)
Volume of the ocean at present	1.350E+18
Volume of H_2O in glacial ice at present (Antarctica, Greenland, Mountains)	0.024E+18
Probable volume of H_2O in glacial ice at Last Glacial Maximum	0.074E+18
Maximum possible volume of ice at Last Glacial Maximum	0.093E+18
Volume of seawater at Last Glacial Maximum	1.276E+18
Minimum possible volume of seawater at Last Glacial Maximum	1.266E+18
Volume of seawater in Early Pliocene (with present Antarctic ice)	1.353E+18
Volume of seawater in Early Pliocene (with 1/4 present Antarctic ice)	1.358E+18
Volume of Late Miocene Ocean (with present Antarctic ice)	1.353E+18
Volume of Early Miocene Ocean (with 1/2 present Antarctic ice)	1.363E+18
Volume of the ocean on a warm ice-free Earth (Eocene - Cretaceous)	1.375E+18
SEAWATER MASSES	kg
Mass of seawater in the ocean at present	1.397E+21
Mass of water in glacial ice at present	0.024E+21
Mass of additional glacial ice at Last Glacial Maximum	0.050E+21
Maximum possible amount of ice at Last Glacial Maximum	0.093E+21
Mass of seawater in the ocean at Last Glacial Maximum	1.347E+21
Minimum possible mass of seawater at Last Glacial Maximum	1.329E+21
Mass of seawater in Early Pliocene (with present Antarctic ice)	1.397E+21
Mass of seawater in Early Pliocene (with 1/4 present Antarctic ice)	1.403E+21
Mass of seawater in Late Miocene (with present Antarctic ice)	1.403E+21
Mass of seawater in the Early Miocene Ocean (with 1/2 present Antarctic ice)	1.414E+21
Mass of seawater in the ocean on an ice-free Earth (Eocene - Cretaceous)	1.422E+21
MASSES OF SALT	kg
Mass of salt in the ocean at present	48.514E+18
Mass of Salt in Ocean at Last Glacial Maximum	48.514E+18
Minimum possible mass of salt in the ocean at Last Glacial Maximum	48.473E+18
Mass of salt in Ocean in Early Pliocene (with Antarctic ice)	48.514E+18
Mass of salt in Ocean in Early Pliocene (without Antarctic ice)	48.514E+18
Mass of salt in Ocean in Late Miocene (before Red Sea and Mediterranean extractions)	51.616E+18
Mass of Salt in Ocean in Early Miocene	52.581E+18
Mass of salt in the ocean on an ice-free Earth (Eocene - Late Cretaceous)	50.800E+18
Mass of salt in the ocean on an ice-free Earth (Early Cretaceous)	60.131E+18
MASSES OF WATER	kg
Mass of H_2O in the ocean at present	1.349E+21
Mass of H_2O in the ocean at the Last Glacial Maximum	1.299E+21
Minimum possible mass of H_2O in the ocean at Last Glacial Maximum	1.289E+21
Mass of H_2O in the ocean in Early Pliocene (with present Antarctic ice)	1.349E+21
Mass of H_2O in the ocean in Early Pliocene (with 1/4 present Antarctic ice)	1.354E+21
Mass of H_2O in the ocean in Late Miocene (with 1/2 present Antarctic ice)	1.343E+21
Mass of H_2O in the Early Miocene Ocean (with 1/4 present Antarctic ice)	1.353E+21
Mass of H_2O in the ocean on an ice-free Earth (Eocene - Cretaceous)	1.367E+21
ESTIMATED SALINITIES (after Hay and Wold, 1997)	S (1)
Salinity of the ocean at present	34.723
Salinity of Glacial Ice	0.007
Salinity of additional ice at Last Glacial Maximum	0.007
Salinity of old sea ice in Arctic Ocean	4.000
Salinity of the ocean at Last Glacial Maximum	36.001
Maximum possible salinity at Last Glacial Maximum	37.607
Salinity in Early Pliocene (with present Antarctic ice)	34.659
Salinity in Early Pliocene (with 1/4 present Antarctic ice)	34.527
Salinity in Late Miocene (with present Antarctic ice)	38.995
Salinity of Early Miocene Ocean (with 1/2 present Antarctic ice)	38.698
Salinity of the ocean on an ice-free Earth: Eocene, 50 Ma	36.438
Salinity of the ocean on an ice-free Earth: Late Cretaceous, 70 Ma	35.830
Salinity of the ocean on an ice-free Earth: Late Cretaceous, 90 Ma	34.946
Salinity of the ocean on an ice-free Earth: Early Cretaceous, 110 Ma	38.557
Salinity of the ocean on an ice-free Earth: Early Cretaceous, 130 Ma	41.897

have been reconstructed by using principles of sedimentary cycling as discussed in detail by HAY & WOLD (in press). They concluded that during most of the Cenozoic mean ocean salinities have been higher than they are today. Each of the major salt extractions into the offshore caused a rapid decrease of oceanic salinity by a few per mille. In the Early Cretaceous mean ocean salinities ranged between 38 and 42 ‰, and the Jurassic and Triassic they were between 43 and 53 ‰.

The effects of significant variations in the mean salinity of the ocean are shown in Fig. 3, a temperature (T), salinity (S), and density (ρ, curved lines) plot for ocean surface waters. The dotted curve (A) shows 5° zonal averages of T, S, and ρ from the Arctic through the North Atlantic to the equatorial Atlantic. The dashed curve (D) shows 5° zonal averages of T, S, and ρ for the South Pacific. Solid curve C shows zonal averages of T, S, and ρ for the Cretaceous "South Pacific". Solid curve E is identical to C but is displaced to the right, to reflect an average ocean salinity of 38.0 ‰, our estimate of the highest salinity during the Late Cretaceous. Solid curve B is identical to C but is displaced to the left, to reflect an average ocean salinity of 33.6 ‰, our estimate of the lowest salinity during the Late Cretaceous. Although the solid curves are identical, the density contrasts along them are different. This is because the slopes of the density curves change with the assumed average salinity, a result of the non-linearity of the equation of state of seawater. The total density contrast in curve E is about 7.7 kg/m^3, whereas in B it is about 7.0 kg/m^3. The total density contrast on the modern South Pacific curve (D) is only 5.4 kg/m^3. In the interior of the modern ocean, slight density differences separate major water masses. Today the density difference between the Intermediate Water, which contains the oxygen minimum (HAY 1995a) and the Deep Water underlying it is about 0.3 kg/m^3, and the difference between warmer, more saline North Atlantic Deep Water and colder, less saline Antarctic Bottom Water is less than 0.1 kg/m^3.

For the modern ocean, density increases steadily from the warm equatorial region to the poles. As a result, the density surfaces in a pole-to-pole meridional are depressed at the equator, where the lowest density water is found, and rise to the surface at higher latitudes. In the Cretaceous the maximum densities are in the tropical and polar regions. In a pole-to-pole meridional section there are zones of lower density water both in the mid-latitudes and along the equator. Any region where surface waters have high density is a potential site of intermediate or deep water formation.

Were "ocean anoxic events" related to high ocean salinities?

During the Late Jurassic and Early Cretaceous C_{org}-rich sediments were deposited in many areas of the world. These are now the source rocks of most of the major producing oil fields. The paleoceanography of C_{org}-rich sediments has been the topic of a recent paper by HAY (1995b), but the origin of the Late Jurassic-Early Cretaceous deposits does not have a satisfactory explanation. The C_{org}-rich sediments of the Late Jurassic and Early Cretaceous accumulated when the mean ocean salinity was high (39-42 ‰), before the South Atlantic salt extraction. In contrast, C_{org}-rich deposits are rare in the Late Cretaceous, when the

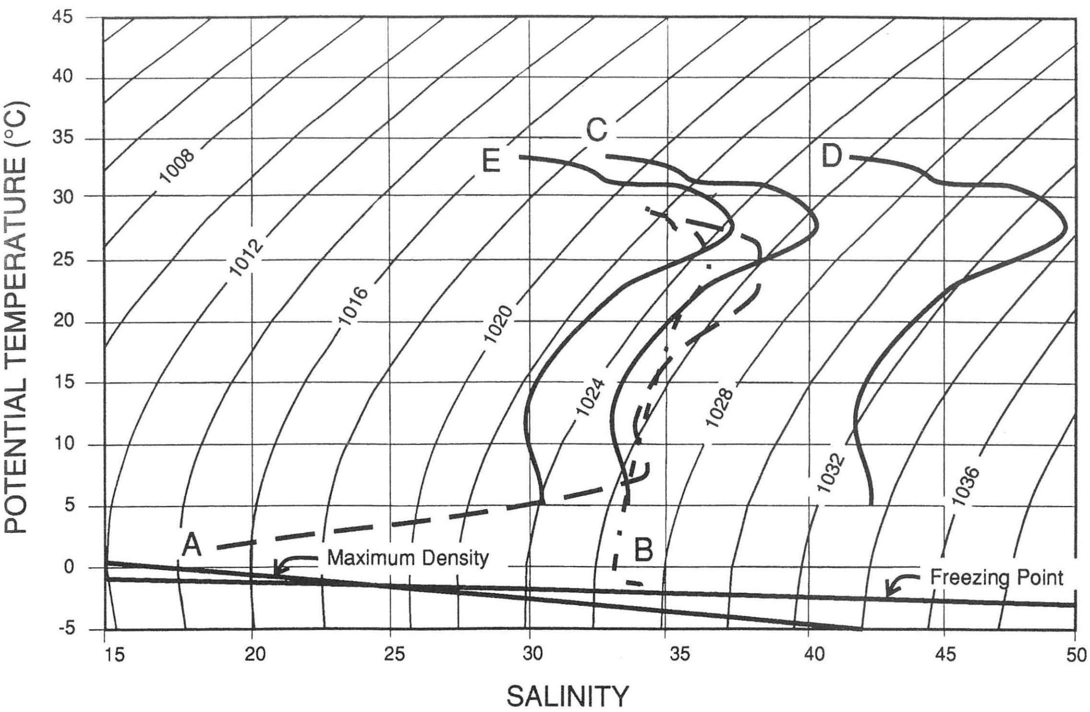

Fig. 3. The differences between modern surface ocean temperatures, salinities, and densities, and those simulated for the Late Cretaceous. The curved lines are densities (kg/m^3). Long-dashed curve A is the average for the surface of the present North Atlantic and Arctic oceans, and dashed-dot curve B is the average for the surface of the South Pacific (after LEVITUS 1982). Except for the Arctic Ocean, the densest waters are in the polar regions, with density declining steadily to the equatorial region. Solid curve C represents temperatures and salinities for the South Pacific from the Campanian simulations assuming the average ocean salinity to be 34.8 ‰. In the Campanian simulations the densest waters are the high salinity warm waters of the low latitudes and densities decrease toward both the equator and poles. Solid curve D is for the Campanian South Pacific assuming a mean ocean salinity of 43.6 ‰. Solid curve E assumes a mean ocean salinity of 32.9 ‰. The temperature range in the Campanian simulations is 5 to 34 °C; the present range is -1 to 28 °C. Campanian surface ocean salinity contrasts are significantly greater than present. Fig. 2A shows that outside the South Pacific, there are large areas of the simulated Campanian ocean where the maximum density of seawater would be above the freezing point.

salinity of ocean waters was much lower. This leads us to suspect that the behavior of the thermohaline circulation in the Late Jurassic and Early Cretaceous was fundamentally different from today. When ocean salinities are about 40 ‰ or higher, the most energy-efficient way to form deep water is simply by cooling the water at high latitudes. Sea ice formation results in salinization of the surrounding waters and is the most important means of deep water formation today, but it is the most efficient means of deep water formation only at salinities of 35 ‰ or lower. Evaporation also increases the salinity of surface waters and can lead to deep water formation, as in the modern Mediterranean, but it is the least energy-efficient mechanism.

If high mean ocean salinities prevailed in the Early Cretaceous,deep water formation may have been driven exclusively by cooling of the water in the polar regions. The waters at sites of deep water formation in the polar regions must have had a high salinity to acquire the necessary density to sink into the deep ocean, but because of precipitation and low evaporation rates polar waters have lower than average salinities. If deep waters formed by cooling in the polar regions, it implies that high salinity waters from the tropical-subtropical gyres must have been drawn across the subtropical and polar fronts to replace waters descending into the ocean interior at the sites of deep water formation.

The climatic transition from the Early Cretaceous with episodic occurrence of polar ice to the equable climates of the Late Cretaceous may be a result of the Early Cretaceous salt removal in the South Atlantic. The lower salinity may have assisted the transition from deep water formation in the polar regions to deep water formation in the mid-latitudes. This transition could be responsible for the reversal of ocean circulation postulated by CHAMBERLIN (1906) and BRASS et al. (1982).

Conclusions

There have been significant changes in the mean salinity of the ocean during the Mesozoic and Cenozoic. Reconstruction of paleosalinities indicates that during most of the Cenozoic mean ocean salinities have been higher than at present. Late Cretaceous paleosalinities were similar to those of the Late Cenozoic, but Early Cretaceous mean ocean salinities were much higher, ranging between 38 and 42 ‰. The major salt extractions of the S outh Atlantic (Early Cretaceous) and Red Sea and Mediterranean (Late Miocene) caused rapid decreases of oceanic salinity by a few per mille. Because of the non-linearity of the equation of state for seawater, the ocean may behave differently when the mean oceanic salinities are significantly different. At salinities of 40 ‰ or higher, the most energy-efficient way to form deep water is by cooling saline water at high latitudes. Sea ice formation expells salt into the surrounding waters and is the most important means of deep water formation today. However, it is an efficient mechanism for deep water formation only at salinities of 35 ‰ or lower. Evaporation also increases the salinity of surface waters and can lead to deep water formation, as in the modern Mediterranean, but it is the least energy-efficient mechanism.

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