GEOLOGICA CARPATHICA, 46, 5, BRATISLAVA, OCTOBER 1995
257 - 266

CRETACEOUS PALEOCEANOGRAPHY

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(Manuscript received January 17, 1995; accepted in revised form June 14, 1995)

Abstract: The modern ocean is comprised of four units: an equatorial belt shared by the two hemispheres, tropical-subtropical anticyclonic gyres, mid-latitude belts of water with steep meridional temperature gradients, and polar oceans characterized by cyclonic gyres. These units are separated by lines of convergence, or fronts: subequatorial, subtropical, and polar. Convergence and divergence of the ocean waters are forced beneath zonal (latitude-parallel) winds. The Early Cretaceous ocean closely resembled the modern ocean. The developing Atlantic was analogous to the modern Mediterranean and served as an Intermediate Water source for the Pacific. Because sea-ice formed seasonally in the Early Cretaceous polar seas, deep water formation probably took place largely in the polar region. In the Late Cretaceous, the high latitudes were warm and deep waters moved from the equatorial region toward the poles, enhancing the ocean's capacity to transport heat poleward. The contrast between surface gyre waters and intermediate waters was less, making them easier to upwell, but because their residence time in the oxygen minimum was less and they contained less nutrients. By analogy to the modern ocean, "Tethyan" refers to the oceans between the subtropical convergences and "Boreal" refers to the ocean poleward of the subtropical convergences.

Introduction

In preparation for running a coupled climate-ocean general circulation model for the Cretaceous it has become necessary to think carefully about how ocean circulation operated then. Simplified models of the Cretaceous paleogeography and ocean have obscured significant similarities and differences between the structure and behavior of the ocean then and now. This report is intended to evaluate the relative importance of different aspects of ocean circulation and to determine the most important factors that must be included in a realistic simulation of the circulation of the Cretaceous ocean.

In the world today, the waters that are upwelled along tropical and subtropical coasts, in tropical thermal domes and along the equator are not ocean deep water, but are central waters, formed in the subtropical gyres, and intermediate waters, formed along the polar convergences or in marginal seas. Beneath the tropical-subtropical gyres, the intermediate water contains the oxygen minimum and hence is the major site of regeneration of nutrients. If this is to be taken into account, the simplest model that can describe the ocean today must contain a surface water layer, an intermediate water layer that outcrops along the poleward margin of the surface water, and a deep water layer that outcrops poleward of the intermediate water. A model with analogous resolution is required to describe the Cretaceous ocean.

Descriptive geographic and oceanographic terms are shown in Fig. 1, and used as follows: equatorial refers to the region within 15° of the equator; tropical refers to the region between the Tropic of Cancer (~23.5° N) and the Tropic of Capricorn (~23.5° S); subtropical refers to the region between the Tropics and the poleward boundaries of the great anticyclonic oceanic gyres at ~40° N and S; temperate to the region between ~40° N and S and the Arctic and Antarctic Circles (~66.5° N and S); polar refers to the region poleward of the Arctic and Antarctic Circles. Cyclonic circulation is counterclockwise in the northern hemisphere, clockwise in the southern hemisphere; anticyclonic circulation is opposite to cyclonic circulation. In both atmosphere and ocean, upward flow occurs in the center of cyclonic circulation, downward flow in the center of anticyclonic circulation. A positive fresh-water balance refers to an excess of the sum of precipitation and runoff over evaporation and induces estuarine circulation, as shown in Fig. 2A. A negative fresh-water balance refers to an excess of evaporation over precipitation and runoff, and results in anti-estuarine or lagoonal circulation (Fig. 2B).

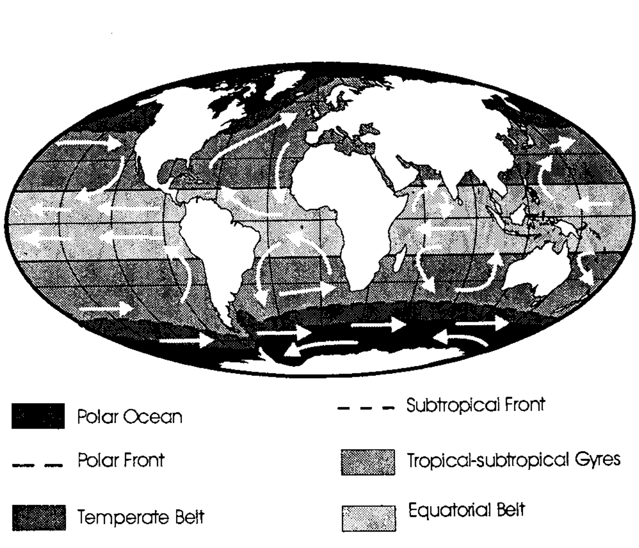

Fig. 1. Major regions, surface currents and fronts in the modern oceans.

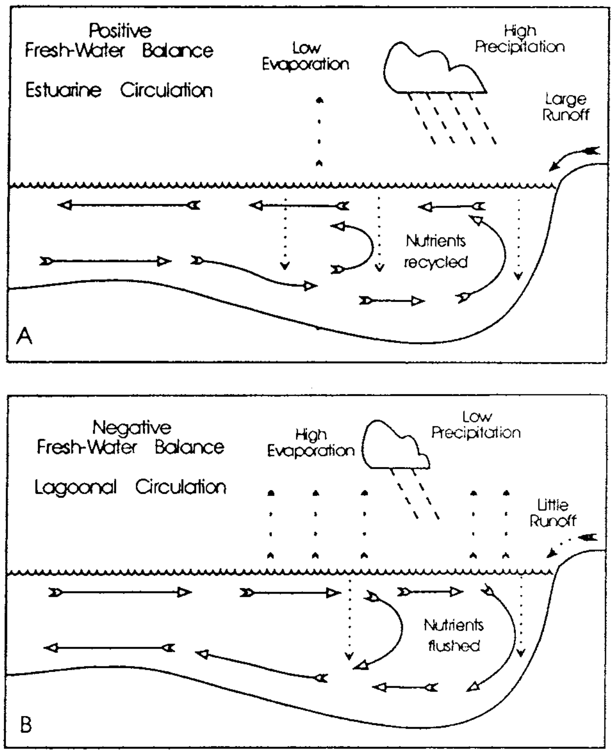

Fig. 2. Characteristics of positive (A) and negative (B) fresh-water balance seas.

Structure and circulation of the modern ocean

The view of the ocean presented here is based on classical descriptive oceanography, as presented in Tchernia (1980), Wells (1986), Brown et al. (1989), Pickard & Emery (1990), and Peixoto & Oort (1992) but emphasizes the role of ocean frontal systems, as suggested by Rooth (1982).

The modern ocean can be thought of as comprised of four units, three of which are mirrored in each hemisphere: an equatorial belt characterized by divergence and shared by the two hemispheres, stratified tropical-subtropical anticyclonic gyres, temperate mid-latitude belts of water characterized by convergence and steep meridional temperature gradients, and deeply convecting polar oceans characterized by cyclonic gyres. Convergence and divergence of the ocean waters are forced beneath zonal (latitude-parallel) winds. These features are shown schematically in Fig. 3.

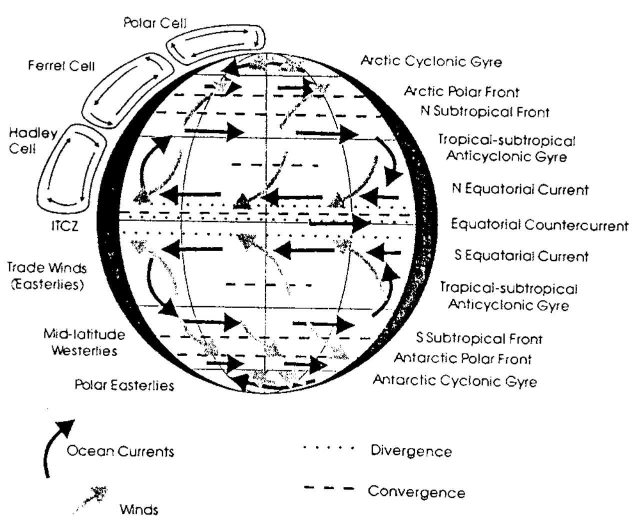

Fig. 3. Winds, ocean currents and oceanic fronts developed over a hypothetical symmetric large ocean basin.

The equatorial belt is a region of mostly westward surface flow beneath the influence of the easterly trade winds. It contains 28 % of the ocean surface area. Eastward subsurface flow returns some of the water, but much of the westward flow becomes incorporated into the western boundary currents of the tropical subtropical gyres. The eastward flow of the countercurrents may rise and break the surface in the eastern parts of the ocean basins. Increasing zonal velocities of the trade winds away from the atmospheric Intertropical Convergence Zone (ITCZ) induce divergence and upwelling in the waters of the equatorial belt. At present the ITCZ tends to be mostly in the northern hemisphere, and this results in the equatorial belt and its fronts and divergences being shifted asymmetrically into the northern hemisphere. As shown in Fig. 4, *Equatorial Surface Waters* are homogeneous; temperatures vary only from 26 °C to 28 °C. Salinities increase from less than 34.5 in the eastern sides of the basins to 35.5 on the western sides; density of the equatorial waters is low, $\sigma_t \leq 23$.

The most stable features of the modern ocean are the stratified tropical-subtropical gyres with warm saline surface waters. These lie in each ocean and in each hemisphere between ~15° and ~45° latitude. They occupy 46 % of the ocean surface. The *Tropical-Subtropical Surface Waters* are mixed by the wind, and locally form a homogeneous isothermal, isohaline layer. As shown in Fig. 4, temperatures of the surface waters of these gyres range from 28 °C along the equatorial margins to 15 °C along the polar margins. Salinities in the gyre centers range from ~35.5 in the North Pacific to >37.5 in the North Atlantic. Densities range from $\sigma_t = 23$ to 26.

The low-latitude easterly trade winds and mid-latitude westerly winds drive the gyres. The Coriolis force induces Ekman transport 90° to the right of the wind in the northern hemisphere and 90° to the left of the wind in the southern hemisphere, causing convergence in the center of the gyre. This convergence occurs beneath the dry descending air of the subtropical highs. Evaporation into this dry air raises the salinity of the waters in the center of each of the gyres, and reinforces the tendency to sink already forced by convergence. The tendency to sink is further enhanced by evaporative and sensible cooling of the surface waters, especially during winter. The sinking waters spread beneath the wave-mixed surface waters of the gyre as *Central Waters*. As shown in Fig. 4, central waters are saline, but cooler than the overlying surface waters. They range down to 15 °C and form the upper part of the main ocean thermocline. Residence time of central waters is in the order of a few decades.

The maximum of the zonal component of the westerly winds occurs at about 45° N and S, but because of the instability of the westerlies the maximum zonal velocity varies over about 10-15° of latitude. In this temperate region, which occupies about 11 % of the ocean surface, the ocean waters reach their greatest meridional velocity, flowing equatorward and tending to override waters at the adjacent lower latitudes. This forces the waters to sink along a series of convergences. The area of *Temperate Surface Waters* is bounded by what are termed the subtropical and polar fronts. As shown in Fig. 5, the subtropical fronts form an effective barrier to poleward ocean heat transport

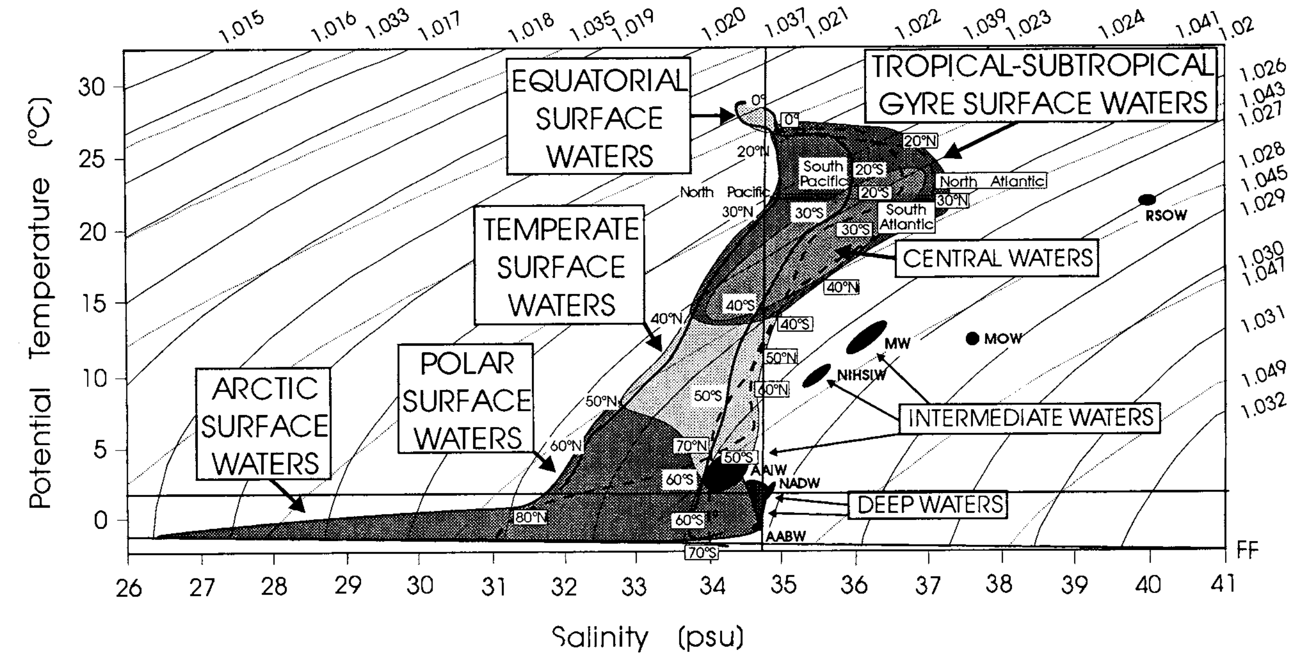

Fig. 4. The relation between temperature, salinity and density (Millero et al. 1980; Millero & Poisson 1981), and major modern oceanic water masses (Levitus 1982; Gorshkov 1974, 1977, 1980). Potential temperature is the temperature the water would have if moved adiabatically to the surface. **FP** - freezing point. Salinity is in "practical salinity units" determined by measurement of conductivity. Thin horizontal and vertical straight lines crossing at T = 3.51 ° C and S = 34.72 are the average temperature and salinity of the world ocean. Densities at the surface are shown by thin solid lines. Densities at 4000 decibars (~4000 meters depth) are indicated by thin dotted lines. Characteristics of Pacific surface water masses at different latitudes are indicated by a heavy solid line. Characteristics of Atlantic surface water masses at different latitudes (indicated by being enclosed in rectangles) are shown by a heavy dashed line. **RSOW** - Red Sea Outflow Water. **MOW** - Mediterranean Sea Outflow Water. **MW** - Mediterranean Water in the North Atlantic. **NIHSIW** - North Indian High Salinity Intermediate Water. **AAIW** - Antarctic Intermediate Water. **NADW** - North Atlantic Deep Water. **AABW** - Antarctic Bottom Water.

in the Pacific Ocean. However, as also shown in Fig. 5, heat is transported to the north throughout the Atlantic. This is a consequence of deep water formation in the North Atlantic that causes surface water to be resupplied across the subtropical fronts. The polar fronts lie along the poleward boundary of the region of sinking and form the equatorial boundary of the polar oceans. As shown in Fig. 4, the temperate waters between the subtropical and polar fronts typically have salinities ranging from about 35.0 to <34.0, mostly below the oceanic average. The temperature descends from 15 °C at the subtropical front to 4 °C along the polar fronts. Densities increase poleward, from $\sigma_t = 26$ to 27.

Figure 6 summarizes schematically the internal circulation of the ocean in an idealized N-S section. Oceanic fronts are not vertical, as they appear to be on sections with great vertical exaggeration, but slope equatorwards and grade into the quasi-horizontal stratification beneath the tropical-subtropical gyres. On the ocean surface the subtropical and polar fronts correspond to the lower half of the main ocean thermocline. The thermocline is the major pycnocline forming the floor of the tropical-subtropical gyres. The top of the thermocline is typically at a depth of less than 50 m on the eastern sides of the ocean basins and 100 m in the central parts of the basins. However, it lies at depths up to 500 m or more beneath the rapidly-flowing western boundary currents. Beneath the thermocline underlying the tropical-subtropical gyres, temperature of the ocean waters declines gradually with depth.

The tropical-subtropical thermocline water overlies *Intermediate Waters*. As indicated by the Antarctic Intermediate Water (AAIW) shown in Fig. 4, these are typically a cooler and less

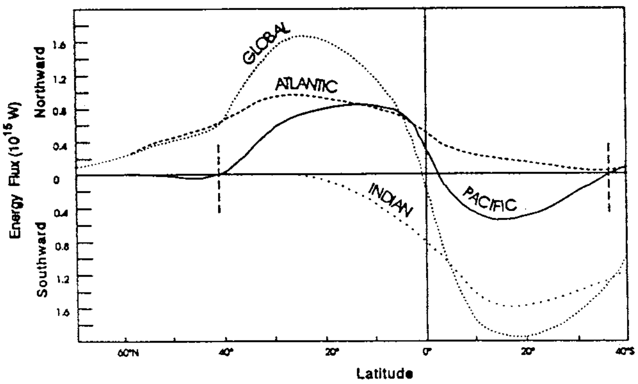

Fig. 5. Meridional heat flux in the modern ocean. Vertical dashed lines indicate the position of the Subtropical Fronts in the Pacific Ocean.

saline layer, that sinks along the polar fronts. There is, however, an alternative source of intermediate water for the ocean, marginal seas with a negative fresh-water balance. The major marginal basins supplying intermediate water to the ocean today are the Mediterranean and Red Seas. Outflow from these is warm and saline, as shown in Fig. 4. The outflowing waters mix with waters in the ocean interior to produce the intermediate water masses indicated in Fig. 4, the Mediterranean Water (MW) of the North Atlantic and North Indian High-Salinity Intermediate Water (NIHSIW) of the Arabian Sea. Residence time of intermediate waters are a few hundred years.

The polar oceans lie poleward of the polar fronts. Excluding the anomalously fresh Arctic Ocean, they occupy about 13 %

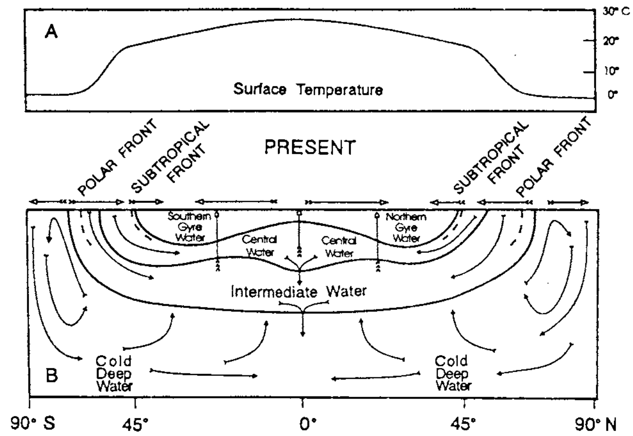

Fig. 6. Schematic profiles, showing **A** - surface temperatures, and **B** - flows in a modern large pole-to-pole ocean basin. Surface flows are indicated by arrows with open head and double tails. Interior flows are indicated by solid arrows with single tails. Low-latitude upwelling is indicated by arrows with open head and triple tails. Fronts are indicated by dashed lines.

of the ocean surface. Poleward of the polar fronts, the decreasing zonal wind velocity induces divergence in the water and causes cyclonic circulation, which also acts to pump water upward. As a result the polar oceans are only weakly stratified, and in the Circumantarctic region the ocean effectively convects from surface to bottom. Waters poleward of the polar fronts are cold and in the Antarctic have a salinity approximating the oceanic average. Below 5 °C, further cooling has little effect on the density of ocean water. To become dense enough to sink to the ocean floor, cold polar waters must increase in salinity. The most energy-efficient mechanism for increasing salinity is production of sea-ice. Newly formed sea-ice typically has a salinity of only 7, about 20 % that of the seawater from which it forms. During the freezing process most of the salt is expelled into the surrounding sea water, increasing its salinity and density. A major characteristic of the modern ocean is that it is filled with cold (<2 °C) water with a salinity ~34.7 that sank in the polar regions (for a review, see Hay 1993).

The present Arctic Ocean (2 % of the ocean surface) is an exceptional polar ocean; runoff dominates its fresh-water balance, resulting in abnormally low salinities in the surface water. However, it is also the site of formation of cold saline deep water that originates from formation of sea-ice from the saline Norwegian Current over the Barents and Kara shelves. The Arctic halocline is the most extreme pycnocline outside the equatorial belt separating cold low-salinity surface ($T = -1.8$ °C, $S = 30$, $\sigma_t = 24$) and high-salinity deeper ($T = -0.7$ °C, $S = 34.9$, $\sigma_t = 28$) water masses. The temperature profile is reversed, the surface waters being colder than the deep water (Pickard & Emery 1990).

The oxygen minimum, nutrients, and productivity

The oxygen minimum is produced by oxidation of organic matter, which consumes the O_2 the water had acquired when at the surface. Because there is no source of O_2 in the ocean interior, the loss of O_2 is cumulative and depends in part on the "age" (length of time since at the surface) of the water. The O_2 minimum forms beneath the surface mixed layer and its most intense local development is where the supply of settling organic particulate material is greatest, i.e. beneath areas of high productivity, as shown in Fig. 7.

In marginal basins, fresh-water balance plays a major role in concentration of nutrients in the intermediate and deeper waters. Basins with a positive fresh-water balance are estuarine (Fig. 2A). Surface waters flow out to the open ocean and there is subsurface inflow into the basin. This flow regime acts to concentrate nutrients in the basin. Both the ocean waters flowing in at depth and the river waters are relatively nutrient-rich, and contribute to the productivity in the basin. Nutrients are regenerated within the basin by settling and oxidation of organic particulate material. In the Black Sea, this results in anoxia of the entire deep water mass. Residence times of deep waters in positive fresh-water balance seas is typically long, e.g. 3,000 years for the Black Sea (Pickard & Emery 1990). Because of the Pacific's estuarine character, O_2 is minimal and nutrient concentrations reach their maximum in the North Pacific. North Pacific deep waters have a residence time of about 2,000 years.

In basins where the fresh-water balance is negative, the flow is anti-estuarine or lagoonal. There is surface flow of relatively nutrient-depleted surface ocean water into the basin and outflow at depth. Runoff is low, so few nutrients are introduced by rivers. Nutrients released into the deeper waters from oxidation of settling organic matter are flushed out of the basin with the deep outflow. The Mediterranean is a negative fresh-water balance basin marginal to the Atlantic, and the Atlantic is in turn a negative fresh-water basin marginal to the Pacific. For this reason, O_2 levels are high and nutrient concentrations low in the deep waters of the Mediterranean and North Atlantic. Residence times for deep water in positive fresh-water balance seas are short, e.g. 70 years for the Mediterranean, 200 years for the Red Sea, and 250 years for the North Atlantic.

With its positive fresh-water balance, it might be expected that the Arctic Ocean should also have an O_2-depleted layer. However, the production of saline deep water on the Barents and Kara shelves effectively ventilates the basin so that only a weak O_2 minimum develops. Residence time of deep waters in the Arctic Basin is only 20–25 years.

Productivity occurs wherever nutrients are introduced into the photic zone. Nutrient-rich subsurface waters can be upwelled by divergence of the surface waters forced by the wind. Upwelling forced by the winds must produce flows sufficient to overcome the density stratification of the water. The greatest density contrast in the ocean is between the surface waters of the equatorial belt and tropical-subtropical gyres, and the intermediate waters beneath them. The upwelled subsurface water must mix with surface water so that it remains on the surface and the nutrients can be completely utilized.

As shown in Fig. 7, open ocean upwelling occurs mostly in the equatorial region and in the polar oceans. In the equatorial region the nutrients can be fully utilized, but in the polar ocean only about 30 % of the nutrients introduced into the surface layer are incorporated into organic tissue. The remainder is returned to the deeper waters by convection. It is uncertain why the utilization in the polar ocean is so low, whether it is due to the rapidity of the convection, the relatively low light levels and absence of light during the winter months, or to an inadequate supply of iron. According to Martin & Fitzwater (1988), iron may be the ultimate limiting nutrient since it is required for production of chlorophyll and for fixation of nitrogen.

Productivity is also high where upwelling occurs on coasts having an orientation such that the wind causes offshore Ekman

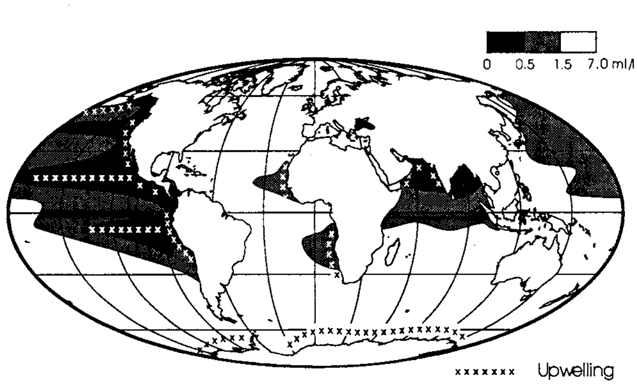

Fig. 7. Upwelling and O_2 concentrations on the absolute O_2 minimum in the modern ocean, after Levitus (1982). O_2 concentrations are shown in ml/l.

transport (for a review of upwelling processes, see Hay in press), as shown in Fig. 7. Because of the density contrast between surface and intermediate waters in the tropical-subtropical region, it is difficult to upwell water deeper than the Central Water masses, and these have relatively low levels of dissolved nutrients. The top of the Intermediate Water can be influenced by the wind only between about 25° and 15° latitude. Where the top of the intermediate water lies too deep to be affected by the wind, cyclonic circulation of the subsurface waters can pump nutrient-rich water upward and introduce it into the central water. However, there remains a strong contrast in productivity between areas where nutrient-rich intermediate water can be upwelled and those where only central waters are upwelled.

In the Atlantic and Pacific Oceans, the O_2 minimum is most strongly developed in the intermediate waters underlying the tropical portion of the gyres on the eastern sides of the ocean basins (see Fig. 7). Hence, the intermediate waters of the eastern tropical gyre margins contain most of the potential nutrient supply. This is predictable, because the highest productivity occurs along the eastern tropical margins of the gyres, and it is the high productivity that produces the organic matter that is decomposed in the intermediate water. It is the shallowness of the pycnocline on the eastern sides of the gyres that makes upwelling of nutrient-rich water particularly effective.

As shown in Fig. 7, the Indian Ocean is anomalous. The monsoons force upwelling in the Arabian Sea, and the Ganges-Brahamaputra river system brings nutrients to the Bay of Bengal, both processes producing O_2 minima in the northern Indian Ocean.

Comparison of the Pacific, Atlantic and Indian Oceans

Pacific and Atlantic water masses are indicated in Fig. 4; Indian Ocean water masses have intermediate characteristics. The Pacific operates almost as an ideal ocean should. Gyre surface and central waters are underlain by symmetrical northern and southern low-salinity intérmediate water masses formed along the polar fronts in both hemispheres. The intermediate waters in both hemispheres are nutrient-rich and support high productivity. Coastal productivity is highest in the southern hemisphere because of the relative stability of southern hemisphere winds and orientation of the western margin of South America. In contrast to the surface, central and intermediate waters, the deep water of the Pacific Ocean is asymmetrical; all Pacific Ocean deep water is produced in the Antarctic region. The deep water in the northern Pacific is older than that of the South Pacific, and is consequently more O_2-depleted and nutrient enriched.

The Atlantic and Indian Oceans are quite different from the Pacific. The South Atlantic and southern Indian Ocean gyres are similar to those of the Pacific, but the intermediate waters of the North Atlantic and northern Indian Ocean are dominated by warm saline outflows from the Mediterranean and Red Seas, replacing the cool, low salinity Intermediate water that characterizes a well-behaved ocean. In the North Atlantic the situation is made more complex by cold saline overflows of the Greenland-Scotland Ridge and deep waters formed in the Labrador Sea. This complex of water masses forms a northern source of deep water (North Atlantic Deep Water, NADW) that spreads through much of the Atlantic. The anti-estuarine circulation introduces O_2 and flushes nutrients from the North Atlantic. In the Indian Ocean the marginal sea outflows are too small to result in deep water formation with the consequence that there is no northern source for O_2 and nutrients accumulate in the aging deep water.

Structure of the Early Cretaceous Ocean

During the Early Cretaceous large ice sheets were absent, but the Earth displayed climatic contrasts (Fig. 8). Boreal (northern), Tethyan (low-latitude) and Austral (southern) marine faunal realms have been recognized. Boreal and Austral faunas have been interpreted as cool-temperate and Tethyan fauna as tropical-subtropical (Stevens 1971). Dropstones and rafted boulders, presumably transported by sea ice or icebergs have been reported from Alaska, Central Australia, Siberia, and Spitsbergen (Kemper 1987; Frakes & Francis 1988). Frakes & Francis (1990) suggested that the ice was seasonal and that there were no continental ice sheets. They reported well developed growth rings in high latitude vegetation. Glendonites have been cited by Kemper (1987) and Sheard (1991) as further evidence of cold polar temperatures. Robin (1987) speculated that the higher elevations in East Antarctica may have supported continental ice sheets of limited size.

The ocean of the Early Cretaceous may have been similar to the modern Pacific ocean, with a large marginal sea, the Central Atlantic-western Tethys. The general pattern of ocean temperatures, salinities, and densities must have resembled that of the modern oceans, shown in Fig. 4, with the difference that the Central Atlantic and western Tethys were more saline than the modern North Atlantic. In comparison with the modern ocean, "Tethyan" refers to the region between subtropical fronts, and "Boreal" and "Austral" the regions north and south of the subtropical fronts, respectively.

Kraus et al. (1978) suggested that if the outflow of the Mediterranean had a volume three times larger than it does at present, it would fill the world ocean with warm saline water. The present Mediterranean has a length of 3,500 km but is narrower, and more strictly confined to the arid zone than was the Early Cretaceous Atlantic. Production of saline waters would have occured where evaporation greatly exceeded precipititation; the water was made more dense by increasing its salt content (see Fig. 4). In the case of competing sources, the dense water with the greatest buoyancy flux becomes the water and the source with the lesser buoyancy flux becomes intermediate water. The buoyancy flux is product of two factors: 1 - the density difference between the denser water and the surrounding seawater, 2 - the volume flux at which the

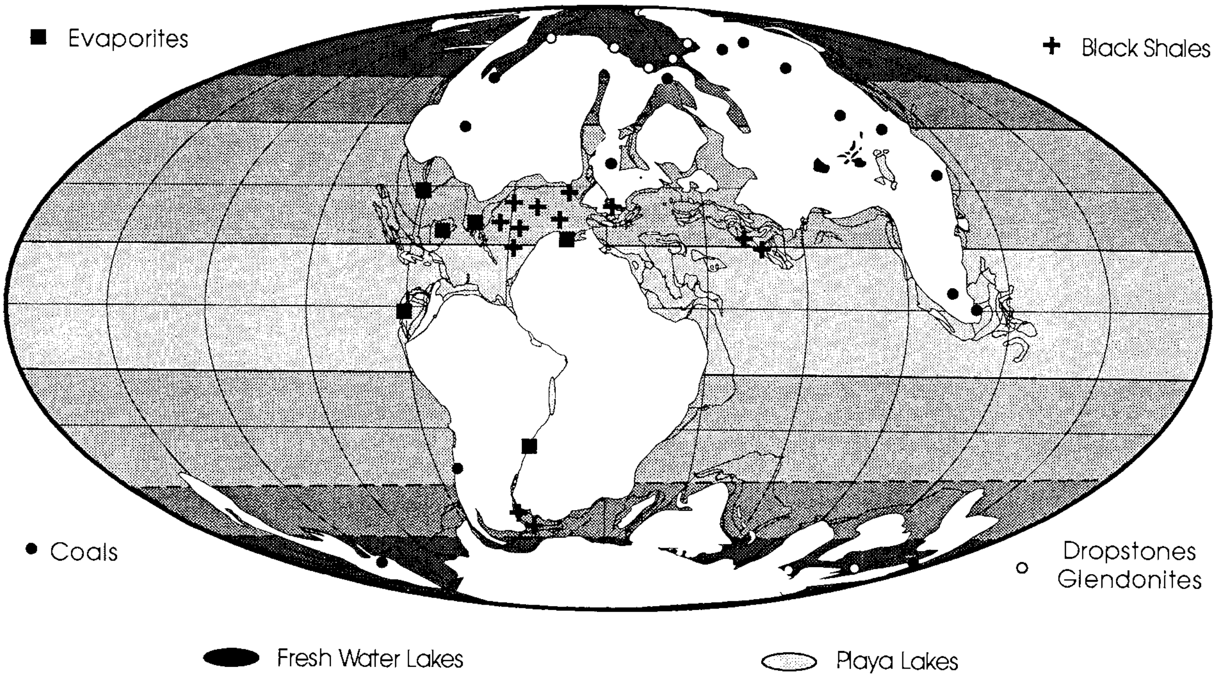

Fig. 8. Occurrence of some climate sensitive sediments and speculative distribution of major oceanic regions in the Early Cretaceous. New plate tectonic model by W.W. Hay, C.N. Wold, K.M. Wilson, S. Voigt & R.M. DeConto; paleogeography and distribution of climate sensitive sediments after Ronov et al. (1989). Water masses and fronts as in Fig. 1.

denser water is produced (Peterson 1979). Warm saline deep water production would be greatest before salts other than carbonate are precipitated. Once salt precipitation begins there is no further increase in density, but the volume flux is reduced and hence the buoyancy flux is diminished (Brass et al. 1982).

Woo et al. (1992) have shown that mid-Cretaceous waters of the Gulf of Mexico were much more saline than today, and might be a source of warm saline deep water. Ocean circulation simulations by Barron & Peterson (1990) used a mid-Cretaceous (100 Ma) paleogeographic configuration and indicated that the western Tethys was a hypersaline area. Weissert et al. (1985) noted that the western Tethys was analogous to the modern Mediterranean in deep water production. In the eastern Mediterranean, the water is made more saline by evaporation during the summer. Cold winter winds from the northern land areas cool the water in the shallow marginal seas. The cold and saline waters mix, forming waters dense enough to sink to the bottom. The continual replenishment of the deep water with oxygenated surface waters prevents anoxic conditions from developing at present. The Early Cretaceous Atlantic was also a paleogeographic analog of the modern Mediterranean. Extending E-W from the Gulf of Mexico to the western Mediterranean, it was 6,000 km long, centered on the arid zone at 25 N latitude. During the Early Cretaceous the Central Atlantic-western Tethys does not appear to have been large enough to have produced bottom water for the world ocean, although it would have been a major source of intermediate water to the Pacific. Evaporation creates dense water by making it more salty, but a similar result is obtained with much less expenditure of energy in making sea ice. As sea ice (salinity ≈ 7) forms, salt is expelled, increasing the salinity and density of the surrounding water. During times in the Early Cretaceous when sea ice was produced in the polar regions, it is likely that the sources of deep water would have been there.

Blocked from the Pacific by the entry of the Caribbean Plate through Central America, from the eastern Tethys by Apulia and other shallow blocks, from the South Atlantic by the Guinea-Brazil Straits, the Early Cretaceous Central Atlantic became the site of deposition of C_{org}. The black shales in the Central Atlantic and western Tethys has been attributed to stagnant pools of warm saline bottom water (Brass et al. 1982), but might also be due to O_2 depletion in response to the periodic development of positive fresh-water balance and estuarine circulation, as has been suggested for the Medierranean sapropels (Rohling & Hilgen 1991). Weissert (1990) suggested that the widespread occurrence of siliciclastic along the margins of the western Tethys in the Early Cretaceous indicate increased runoff from land. Föllmi et al. (1994) have proposed that increased runoff at that time brought nutrients to fertilize the ocean. Barron et al. (1989) report that climate simulations suggest the hydrologic cycle was about 25 % greater during the Cretaceous than it is today. Furthermore, warm Tethyan waters would cause a regional increase in the hydrologic cycle and increased runoff from the adjacent lands.

A special paleogeographic condition that may have played a role in the Cretaceous oceans was the presence of meridional seaways that connected regions with surface waters having similar densities but different temperatures and salinities. As shown in Fig. 4, the concave-to-the-right nature of the isopycnals on a T-S-D diagram indicates that if two water masses of equal density but differing salinities and temperatures mix, the resulting mixture is always more dense and will sink. If the water-mass characteristics differ only slightly, this process is gentle; e.g. it occurs on the isopycnals within the main ocean thermocline, causing the "thermocline mixing" indicated in Fig. 6. However, if the water masses have very different characteristics, their mixing results in catastrophic rapid downwelling, termed "cabelling." This occurs in the Japan Sea today, with the result hat the deep waters have very high oxygen levels. However, Hay et al. (1993) postulated that in the Western Interior Seaway of North America, the cabelling surface waters had a high biomass, with the result that the forced downward flux of organic matter overwhelmed the O_2 supply, producing anoxia at the source of intermediate or deep waters.

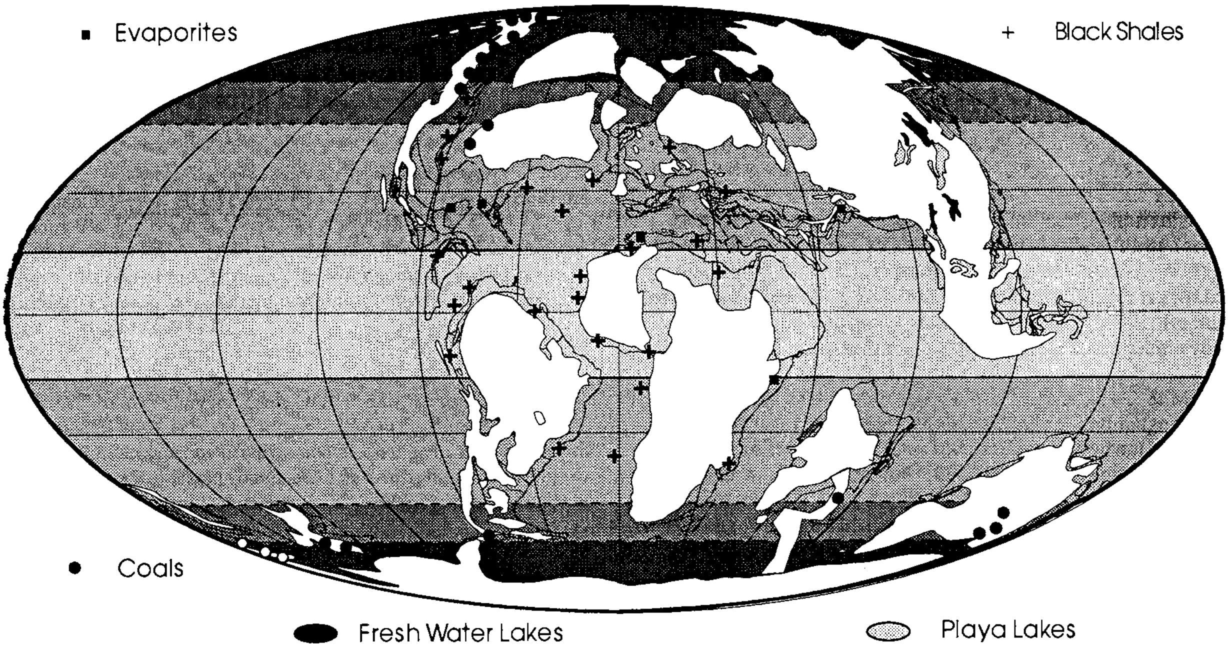

Fig. 9. Occurrence of some climate sensitive sediments and speculative distribution of major oceanic regions in the Late Cretaceous. New plate tectonic model by W.W. Hay, C.N. Wold, K.M. Wilson, S. Voigt & R.M. DeConto; paleogeography and distribution of climate sensitive sediments after Ronov et al. (1989). Water masses and fronts as in Fig. 1.

Structure of the Late Cretaceous Ocean

In contrast, the Late Cretaceous was a time of widespread warmth (Fig. 9). In the middle Tethys, oxygen isotopes indicate warm sea surface temperatures of 31 °C in the Cenomanian and Turonian, declining to 29 °C in the Santonian, with a minimum of 21 °C in the Late Campanian (Flexer et al. 1986). Douglas & Woodruff (1981) cited low latitude Pacific waters as being in the range 29-24 °C, with deep waters consistently about 10 °C cooler. Barrera et al. (1987) reported Antarctic (South Atlantic) shelf waters to be between 4 and 10.5 °C during the Late Campanian and Maastrichtian. There are no reports of ice-rafted detritus from the polar regions. Vegetation allows the land areas to be divided into two high latitude provinces, Northern Laurasian (*Aquillapollenites*) and Southern Gondwanan, and two lower latitude provinces, Southern Laurasian (*Normapolles*) and Northern Gondwanan (Crane 1987). In North America, the vegetation to 45° N has been termed megathermal, interpreted as reflecting temperatures of 30 to 20 °C (Wolfe & Upchurch 1987; Upchurch & Wolfe 1993). Vegetation between this region and 65° N paleolatitude has been termed mesothermal, and to represent temperatures of 20 to 13 °C. There is evidence of seasonality. At higher paleolatitudes temperatures were cooler, estimated to be 2-8 °C during the Campanian-Maastrichtian. Coals formed at both high and low latitudes, and evaporites can be used to define an arid zone. There was extensive volcanism associated with subduction, and this may have maintained high levels of CO_2 in the atmosphere. Overall, the Late Cretaceous climate was described by Frakes et al. (1992) as "very warm."

Most reconstructions of the Late Cretaceous emphasize the asymmetry of the Tethys seaway because "Tethyan faunas" are known mostly from the northern hemisphere. Dias-Brito (1992) has shown that calcisphere distributions support a hemispherically symmetrical Tethys (Megatéthis Mesocretácico) with a distribution of water masses closely resembling that suggested in Fig. 9. During the Late Cretaceous, the polar waters were cooler than the equatorial waters, but the thermal contrast was less than that of today. The ocean was significantly different from that of today, shown in Fig. 4, in that the lowest temperatures were well above 5 °C, the temperature below which cooling of the water has little effect on density. However the general shape of the T-S pattern of major surface water masses must have had the same shape (like a "7") as modern equatorial, tropical-subtropical and temperate water masses, and ocean deep waters must have had a high salinity. The tropical-subtropical gyres in the Pacific had high salinities, though not as high as in the Atlantic. Central waters would have formed as they do today. The subtropical and polar fronts, driven by the winds, must have existed. The subtropical front would form beneath the equatorward boundary of the zone of maximum westerly wind stress, and the polar front would be along the poleward boundary of this zone. Between the subtropical and polar fronts both the temperature and salinity of the water must have decreased, representing outcrop of the lower part of the tropical-subtropical pycnocline. Thus, as is the case today, the main cause of the pycnocline must have been temperature. Being along the poleward margin of the zone of maximum westerly wind stress, the polar front would have formed near where the precipitation exceeds evaporation and the ocean is relatively fresh. The intermediate waters may have been warmer than those of today.

If during the Late Cretaceous the high latitudes were warm and the polar oceans free of sea-ice, they may not have been effective as a source of ocean deep water. The major mechanism responsible for deep water formation today, salinization through the formation of sea-ices did not exist. For waters warmer than 5 °C density increases with both decreasing temperature and increasing salinity. Salinity increases in low-latitude marginal seas could well have dominated deep water production.

The Late Cretaceous Atlantic was significantly larger than that of the Early Cretaceous. It extended across 8,000 km at

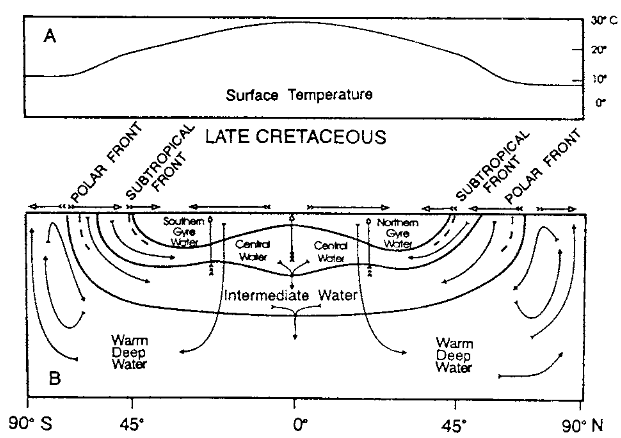

Fig. 10. Schematic profiles, showing **A** - surface temperatures, and **B** - flows in a Late Cretaceous large pole-to-pole ocean basin. Surface flows are indicated by arrows with open head and double tails. Interior flows are indicated by solid arrows with single tails. Low-latitude upwelling is indicated by arrows with open head and triple tails. Fronts are indicated by dashed lines.

latitude 25° N, and was about 3,000 km wide in the arid zone of the southern hemisphere. It was narrowest in the equatorial region, and the flow if equatorial rivers was directed away from it. The only part of the Atlantic that may have received large amounts of fresh water was the southern South Atlantic, with a zonal extent of 5,500 km at 60° S. The widest, deepest connection was between the Caribbean Plate and South America, located almost directly on the equator. At this time the Atlantic is a prime candidate to have been the global supplier of ocean deep water, acting as the source of warm saline water that could be upwelled in the polar regions of the Pacific as well as in the South Atlantic to supply heat to the polar regions. The Pacific ocean was probably fed by warm saline waters emerging from the Cretaceous Atlantic having characteristics similar to the modern Mediterranean outflow.

Low-latitude deep water sources would have an important effect on ocean heat transport, reversing the role of ocean deep water, as shown in Fig. 10. Deep waters would have flowed from the tropical region toward the poles, not from the poles toward the equator as is the case today. The result would be to transport heat poleward of the polar front, where it could become incorporated into the high-latitude convection and brought to the surface.

By analogy with the present we can expect that poleward heat transport in the "tropical-subtropical gyres" of the "Pacific" would have been $\pm 17 \times 10^{15}$ W in each hemisphere. Additional heat transport would have been provided by the poleward flowing deep waters. Upon reaching the polar regions these waters would be incorporated into the top-to-bottom convecting waters poleward of the polar fronts. During transport, the deep warm water was isolated and could not lose its heat. Every 10^6 m^3 of water carried poleward could release 4.2×10^{12} J for every degree it is cooled. For the Late Cretaceous the cooling was about 20 °C, so that a poleward deep water transport rate of 1 Sverdrup would release 84×10^{12} W. The area of the oceans poleward of the polar fronts was $\pm 50 \times 10^{12}$ m^2 in each hemisphere. If the radiation deficit was about 60 Wm^{-2}, deep water transport must have delivered about 3×10^{15} W to keep the poles free of sea ice, requiring poleward flow and upwelling of ±36 Sverdrups. The 60 Wm^{-2} is a present day value; obviously the outward radiation would be greater with a warmer ocean, but the albedo would also be lower, so more incoming radiation would be retained. Upwelling 36 Sverdrups in each hemisphere is reasonable; the present circumantarctic current has a vertical circulation of 70 Sverdrups.

During the Late Cretaceous, the contrast between surface gyre waters and intermediate waters was almost surely less than it is today, making them easier to upwell. However, because they were more easily upwelled, their residence time in the oxygen minimum was less and they contained less nutrients. This may be why Cretaceous ocean productivity appears to have been low (Bralower & Thierstein 1984; de Boer 1986).

A special peculiarity of the Late Cretaceous is the deposition of typically open ocean pelagic deposits, chalk, to the shoreline of shelf and epeiric seas. Today shelf seas are separated from the open ocean by fronts that form over the shelf break. The cause of these fronts is not well understood, but their significance is readily apparent from the informal terms "blue water" and "brown water" used by oceanographers to differentiate the open ocean and inshore waters. As shown in Fig. 11, sea level rose reaching its maximum during early in the Late Cretaceous, and remained high until in the Cenozoic. Although calibration of the sea level curve in absolute terms is questionable, it is evident that the time of highest sea level corresponds to the time of deposition of chalk in seas flooding the continental blocks. As shown schematically in Fig. 12, it appears that in the Late Cretaceous the water depths over the shelf break reached a level that no longer permitted the development of a shelf break front. As a result, for some 30 million years the ocean and epeiric seas behaved as a single unit.

Summary and conclusions

The structure and circulation of the Cretaceous ocean was probably quite similar to that of the modern ocean, with stratified equatorial and tropical-subtropical regions, rapid temperature decline across a mid-latitude frontal region, and deeply convecting polar oceans. The major difference occurred in the Late Cretaceous when deep water flow reversed from equatorward to poleward, vastly supplementing the oceans ability to transport heat to the polar regions.

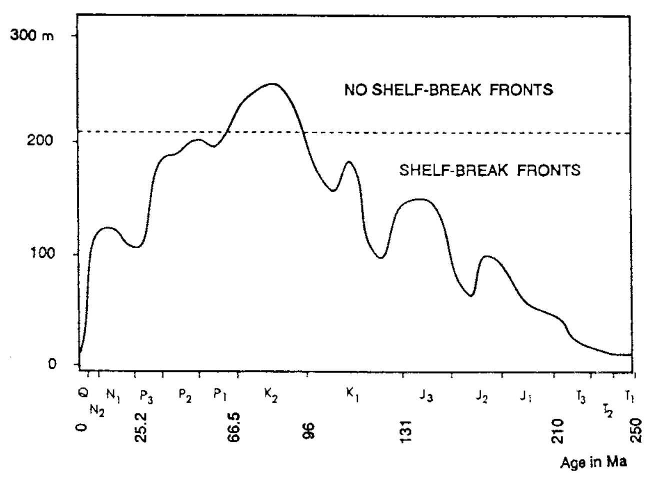

Fig. 11. Global sea level during the Mesozoic and Cenozoic, after Haq et al. (1988). Dashed line indicates he hypothesized level above which shelf break fronts cease to exist.

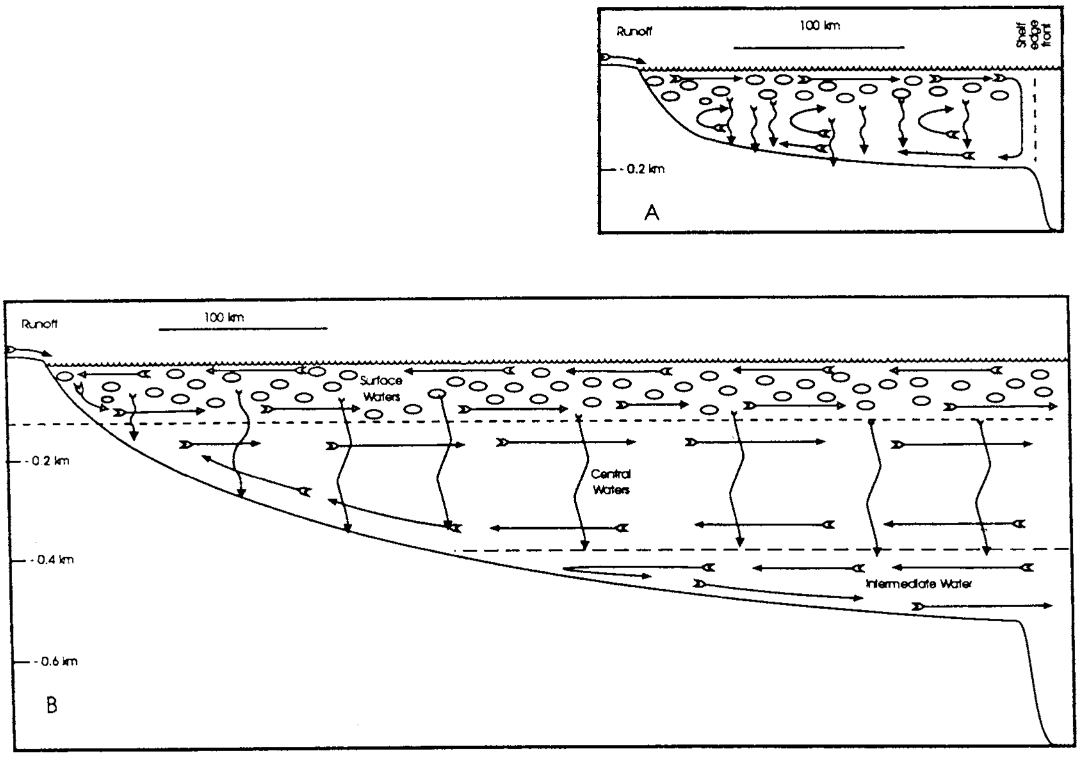

Fig. 12. Schematic views of circulation in **A** – a modern continental shelf sea, and **B** – a Late Cretaceous epicontinental sea.

Episodes of organic carbon deposition may reflect changes in the nutrient supply from land, production of saline bottom waters that formed stagnant pools, or periodic changes from a strongly negative toward a positive fresh-water balance in the Atlantic and Tethys seas.

A realistic ocean model for the Cretaceous must be able to simulate the formation of intermediate as well as deep water. It should also be able to simulate the flows into and out of major marginal seas with positive or negative fresh water balances.

Acknowledgements: This work was carried out with support from the Deutsche Forschungsgemeinschaft, from the Donors of The Petroleum Research Fund administered by the American Chemical Society, and through grants EAR 9320136 and EAR 9405737 from the Earth Sciences Section of the U.S. National Science Foundation.

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