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Evolving ideas about the Cretaceous climate and ocean circulation

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ABSTRACT

The Cretaceous is a special episode in the history of the Earth named for a unique rock type, chalk. Chalk is similar to modern deep-sea calcareous ooze and its deposition in epicontinental seas occurred as these areas became an integral part of the ocean. The shelf-break fronts that today separate inshore from open-ocean waters cannot have existed during the Late Cretaceous probably because the higher sea level brought the base of the wind-mixed Ekman layer above the sea floor on the continental margins.

A second peculiarity of the Cretaceous is its warm equable climate. Tropical and polar temperatures were warmer than today. Meridional and ocean-continent temperature gradients were lower. The warmer climate was a reflection of higher atmospheric levels of greenhouse gasses, CO_2 and possibly CH_4, reinforced by higher water vapor content in response to the warmer temperatures. Most of the additional energy involved in the meridional heat transport system was transported as latent heat of vaporization of H_20 by the atmosphere. Poleward heat transport may have been as much as 1 Petawatt (20%) greater than it is today. C_3 plants provided for more efficient energy transport into the interior of the continents.

Circulation of the Cretaceous ocean may have been very different from that of today. It is impossible for large areas of the modern ocean to become anoxic, but episodes of local anoxia occurred during the earlier Cretaceous and became regional to global during the middle of the Cretaceous. The present ocean structure depends on constant wind systems, which in turn depend on stability of the atmospheric pressure systems forced by polar ice. During most of the Cretaceous the polar regions were ice free. Without polar ice there were seasonal reversals of the high-latitude atmospheric pressure systems, resulting in disruption of the mid- and high latitude wind systems. Without constant mid-latitude westerly winds, there would be no subtropical and polar fronts in the ocean, no well-developed ocean pycnocline, and no tropical subtropical gyres dominating ocean circulation. Instead the ocean circulation would be accomplished through mesoscale eddies which could carry warmth to the polar regions.

Greater knowledge and understanding of the Cretaceous is critical for learning how the climate system operates when one or both polar regions are ice free.

1. Introduction

The Cretaceous has long been recognized as a special episode in the history of the Earth and several of the most important ideas in geology derive from the study of Cretaceous rocks. Among the major stratigraphic units into which Earth history is divided, only the Cretaceous and the Carboniferous are named for unique sedimentary deposits. For the Cretaceous, the special type of rock is chalk (fr. *craie* for the rock, *Crétacé* for the time interval). Chalk was immortalized in Huxley's famous public lecture to the working men of Norwich, England and published in Macmillans Magazine (Huxley, 1868). When deep sea deposits were recovered from the North Atlantic in 1853 it became apparent that such deposits were not restricted to the late Mesozoic, but that somehow the oceanic plankton producing the modern *Globigerina* ooze had thrived in the seas covering the continents during the Late Cretaceous. The question as to how open-ocean plankton penetrated into shallow seas and even to the shore awaits a definitive answer.

A second peculiarity of the Cretaceous is evidence for warm conditions extending into the polar regions, at first conjectured from the presence of limestone (chalk) deposits in Denmark and Sweden (Lyell, 1837). During the latter part of the 19th century there was debate over whether the Earth had a meridional temperature gradient and climate zones prior to the Tertiary. After Lord Kelvin's calculations of the loss of heat by the Earth (Lord Kelvin, 1863, 1864) it was thought that the internal heat flux from the cooling Earth exceeded the solar energy flux until the Cenozoic. Hence, most scientists believed that the climate of the entire planet had been initially hot, with gradual cooling during the Precambrian, Paleozoic and Mesozoic. The Cretaceous was the last epoch of geologic time during which the Earth was uniformly warmed by the heat flux from the interior of the Earth. With the beginning of the Cenozoic the radiation from the sun came to exceed the flux of

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energy from the interior of the Earth and permitted a meridional temperature gradient to develop. Today we know that the energy flux from the sun, measured in hundreds of watts per square meter, has dominated the heat flux from the interior of the Earth, measured in hundredths of a watt per square meter, since the early Archaean. However in the late 19th century cooling of the Earth was assumed to be not only responsible for pre-Tertiary climates, but to drive many geologic processes.

Neumayr (1883) countered the idea of homogenous climates prior to the Cenozoic by demonstrating that the biogeographic distribution of ammonites and other marine molluscs showed that meridional climate zones had existed at least since the beginning of the Jurassic. It was in this context that he applied the terms boreal, temperate, and equatorial to the latitudinally distinct fossil assemblages. He also recognized a number of distinct biogeographic regions and believed that the boundaries between the biogeographic provinces were determined by ocean currents. His ideas were challenged by Ortmann (1896) who argued that the distributions of fossils only reflected paleobiogeographic regions, not climatic differences, and that their boundaries had nothing to do with ocean circulation. However, Haug (1910), in his global account of stratigraphy, accepted Neumayr's ideas and recognized the importance of rudists as indicators of equatorial conditions. Haug used the terms "boréale, équatoriale, australe, and tempérée nord et sud" in describing climatic zones of the Mesozoic. Much work has been accomplished in using proxies to document Cretaceous temperatures and using numerical models to understand the climate of the Cretaceous, but uncertainties remain.

Shortly after the arguments for climate zones on a warm Earth had been developed, two revolutionary ideas were introduced into geology. Chamberlin (1899) proposed that changes in the amount of a greenhouse gas, CO_2 might be responsible for the alternation of ice ages and interglacials and account for globally warm climates. Subsequently, he suggested that a reversal of the thermohaline circulation of the ocean might also be responsible (Chamberlin, 1906).

Another major idea arose as the concept of climate-related biogeographic provinces in the Jurassic and Cretaceous was developed further by Uhlig (1911). Although overthrusts in the Alps had been proposed earlier by Bertrand and others, Uhlig recognized that the apparent sharpness of the boundaries between the Alpine-Carpathian faunas and those of northern Europe were a result of large-scale horizontal movement of parts of the Earth's surface. His ideas of mobility of the Earth's surface anticipated the more general work of Alfred Wegener and modern plate tectonics. The most important tools in making plate tectonic reconstructions of the continents and oceans are seafloor magnetic lineations. Unfortunately, these are unambiguous only for the Atlantic. They are incomplete guides for reconstruction of the Indian and Pacific Oceans and their surrounding areas, and offer little if any information on the past locations of terranes. Because much of the Cretaceous seafloor has been subducted (57% of the latest Cretaceous, 80% of the mid-Cretaceous, and 95% of the earliest Cretaceous!), there are a variety of reconstructions differing particularly in treatment of the Caribbean, Tethys, Indian and Pacific Oceans (for further information see the detailed discussion in Hay et al., 1999).

Finally, rare cores of Cretaceous deep sea sediment were recovered in expeditions after WWII. Some of these contained black shales. On the first leg of the Deep Sea Drilling Project more extensive recovery of Cretaceous black shales was made in the deep western Atlantic off the Bahamas. This initiated an ongoing discussion about whether and how the ocean could become anoxic.

Because the Cretaceous was so different from the modern world, because its sediments are widespread and often visible in outcrops, and because it is the oldest period for which plate tectonic reconstructions can still be constrained by seafloor magnetic lineations, it will always be the best example of a radically different state of the Earth. Although studies of the Cretaceous climate are not often cited in works dealing with future climate change, background knowledge of the Cretaceous has shaped many of the arguments. The Cretaceous is a laboratory for testing ideas about causes and effects of global climate change.

2. The chalk problem

Early Cretaceous strata resemble those of most other geologic periods, but starting in the Cenomanian and lasting in many parts of the world until the end of the Maastrichtian, conditions became unique in Earth history. Chalk is a porous, earthy, limestone made up of the calcitic remains of calcareous nannofossils and planktonic foraminifera. As a deposit it is analogous to modern deep-sea *Globigerina*/nannoplankton ooze. However, it was deposited over broad areas of the continental platforms. Understanding how this could happen may offer important insights into the behavior of the ocean at the time.

In part, the deposition of chalk on the continental blocks may have been simply due to the evolution of the calcareous plankton. Planktonic foraminifera have their origins in the Jurassic (BouDagher-Fadel et al., 1997) and the calcareous nannoplankton in the Triassic (Bown et al., 2004) as red plastids replaced green in the ocean (Falkowski et al., 2004). As indicated in Fig. 1, it is thought that both of these groups were originally restricted to epeiric seas, but they extended their habitat, colonizing the open ocean, in the Early Cretaceous (Roth, 1986) when they became deep-sea ooze and rock-forming, leaving a record of calcitic pelagic micrite in the Tethys (Maiolica), as well as in the Atlantic and Pacific. Except for the brief terminal Cretaceous episode, they have continued to form deep-sea oozes ever since. A wide variety of calcareous nannoplankton continued to occupy epicontinental seas in the early Tertiary but they never again became rock-forming in shallow waters. The planktonic foraminifera and other calcareous nannoplankton have remained important elements in the open-ocean plankton, and have become ever more effective as contributors of carbonate to deep sea sediments. During the Late Cretaceous and especially the Cenozoic, they transferred the bulk of carbonate deposition from shallow seas, where it had been throughout the earlier Phanerozoic, to the deep sea (Southam and Hay, 1981; Hay, 1999). By creating a deep sea carbonate sink, the calcareous plankton revolutionized regulation of ocean chemistry, allowing a control on the saturation state of calcium carbonate in the ocean that had not existed previously (Ridgwell, 2005).

It is most likely that, if the calcareous plankton originated in epeiric seas, those seas had a positive fresh water balance, so that surface waters would flow out onto the ocean. The calcareous plankton would initially colonize the surface mixed layer of the ocean. With evolutionary progress they could then colonize the denser waters of the pycnocline (density gradient - probably a thermocline or thermal gradient at the time although it might possibly have been a halocline or salinity gradient) and sub-pycnocline waters, as shown in Fig. 1. This hypothesis could be tested using information on the depth habitats of foraminifera of different ages.

Douglas and Savin (1978), Boersma and Shackleton (1981), D'Hondt and Arthur (1996), and Abramovich et al. (2003) have used stable isotopes to establish depth relations for a number of planktonic foraminiferal species. Abramovich et al. (2003) recognized deep, subthermocline, thermocline and mixed layer species. In general the smaller forms with lighter tests inhabit the upper layers, and species with heavier tests deeper layers. This corresponds to a zonation according to buoyancy, as proposed by Bé (1968), Bé et al. (1973), and Marszalek (1975, 1982). A similar depth zonation of calcareous nannofossil-bearing organisms has been

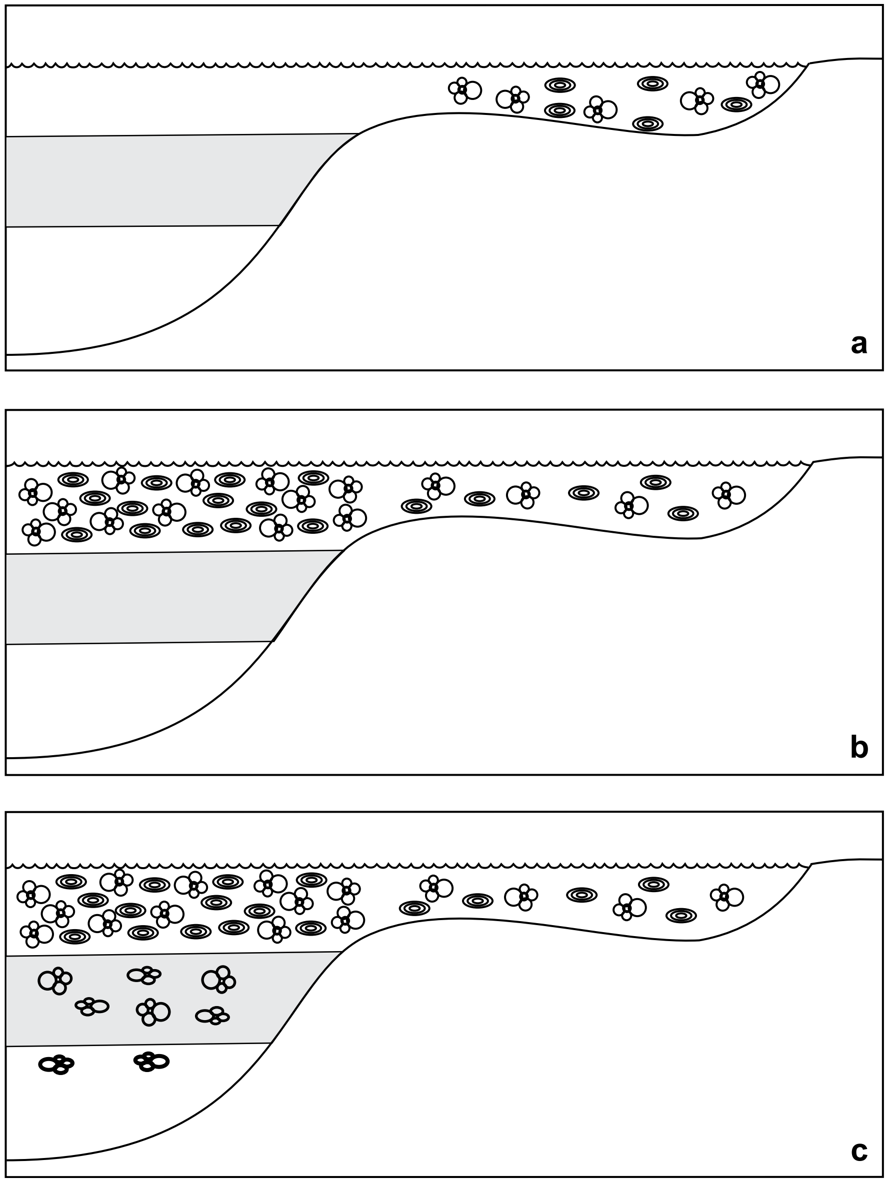

Fig. 1. Schematic history of calcareous plankton: foraminifers (joined circles) and calcareous nannofossils (nested ovals). a. Origination in epeiric seas in the Jurassic; b. spread into surface waters of the open ocean (Early Cretaceous); c. colonization of the pycnocline and subpycnocline waters (later Early Cretaceous).

proposed by Young (1994). This interpretation, which would be valid whether the pycnocline is a thermocline, a halocline, or both, has been applied to the Cretaceous Western Interior Seaway of North America by Fisher et al. (2003). D'Hondt and Zachos (1998) found evidence that some Late Cretaceous (Maastrichtian) serial planktonic foraminifera had symbiotic algae and clearly lived in the photic zone, while some trochospiral species clearly did not have photosymbionts.

Although the cause of the spread of calcareous plankton into the open ocean remains uncertain, it occurred after a significant decline in ocean salinity associated with deposition of salt in the South Atlantic during the Aptian (Southam and Hay, 1981; Hay et al., 2001, 2006). It also corresponds to the time when the Mg/Ca ratio in seawater fell below 2 (Stanley and Hardie, 1999; Berner, 2004) and when sulfur isotope ratios ($\delta^{34}S$) for seawater sulfate changed from late Early Cretaceous (Albian) values of about 15.5‰ to Late Cretaceous values of about18‰ (Paytan et al., 2004). Stanley et al. (2005) found that some modern species of coccolithophore can multiply more rapidly and secrete more coccoliths in waters with the lower Mg/Ca ratio thought to have prevailed in the Cretaceous whereas they are inhibited by modern seawater composition. They argue that the change in composition of seawater was critical to the massive deposition of chalk.

Unfortunately, whatever permitted the spread of calcareous plankton into the open ocean does not explain the subsequent rock-forming occurrence of the open-ocean plankton to produce the chalk deposited in epicontinental seas. This happened as the Late Cretaceous epicontinental seas became an integral part of the ocean.

2.1. Ocean frontal systems in and bordering epicontinental seas

Csanady (1982) noted that circulation in the coastal ocean differs fundamentally from that of the open ocean in that the large ocean basins are under the long-term influence of both the trade winds and the westerlies, whereas coastal regions are much more strongly influenced by passing weather systems. Furthermore, today the waters over most continental shelves are separated from those of the open ocean and internally subdivided by a spectrum of frontal systems, discussed below. Most modern epicontinental seas have a restricted communication with the ocean through narrow passages and are either more or less saline than the ocean, depending on their fresh-water balance.

Csanady (1997) has presented a general review of the state of knowledge of coastal ocean processes; the discussions below are brief descriptive accounts of aspects of the topic. For full mathematical treatment the reader is referred to the papers cited. The best general account of modern shallow water frontal systems is that of Tomczak (1998) which is not formally published but available at his website. Frontal systems are narrow bands in the ocean where the properties of the seawater change rapidly over short distances. They separate regions where the water-mass properties change only gradually over large distances. Six types of frontal systems occur on continental margins: tidal fronts, river plume fronts, estuarine fronts, caballing fronts, upwelling fronts and shelf-break fronts.

Tidal fronts form where the tides in shallow waters cause turbulence so that a mixed (inshore) water mass is juxtaposed against a more stratified (offshore) water mass. They depend on local bathymetric conditions and are generally only a few km to tens of km in length; they are ephemeral features unimportant in affecting the long-term separation of ocean and inshore waters.

River plume fronts develop where river outflows are large enough to overwhelm the mixing processes. They are very important today in differentiating the Arctic shelf waters from those of the central Arctic Ocean. However, the plumes of many large rivers (e.g., Amazon, Orinoco) extend across the continental shelf and are frequently sheared off as they become entrained in the boundary currents of the oceanic gyres. Where they are stable, river plume fronts force recycling of nutrients by being highly productive, exporting particulate organic matter to the underlying seawater where the nutrients are regenerated, and then pulling up the underlying seawater to become diluted. As a result sediments associated with river plumes are often very rich in organic matter.

Estuarine fronts are the analogs of river plumes, but develop in estuaries, coastal features where the riverine sediment load has been insufficient to fill in the coastal river valleys deepened during glacial low stands of sea level. Estuaries are a geomorphic peculiarity of the late Neogene. However, they have an upscaled analog in marginal seas with a strong positive fresh-water balance, such as the Baltic Sea and Hudson Bay. Positive fresh-water balance marginal seas have certainly existed throughout Earth history, and the great transcontinental meridional seaways of the northern hemisphere behaved as giant estuaries. From about 95 Ma until the Pliocene the Arctic Ocean Basin was largely isolated from the ocean proper and was brackish to fresh. The meridional seaways across North America, Europe, and Asia connected this fresher water in the north with the high salinity waters of the Tethys in the south.

Caballing fronts develop where water masses with the same density but different temperatures and salinities meet. The resulting mixed water is always denser than either of the parent waters, and sinks forming a front. Caballing fronts are important in the hyperventilation of the Japan Sea and in deep water formation in both the Arctic and Antarctic. The sudden transition from carbonate to non-carbonate sediments in the Cenomanian-Turonian of the North American Western Interior seaway in northern Wyoming and South Dakota has been interpreted as a caballing front (Hay et al., 1993; Fisher et al., 1994) which pulled in and killed the surface water plankton enriching the sediments in organic matter.

Upwelling fronts form where the wind blows parallel to the coast. The surface Ekman layer transport is directed 90° to the right of the wind direction north of the equator and to the left south of the equator. In response surface waters move offshore, lowering the sea surface near the coast and bringing deeper waters to the surface. Upwelling fronts may develop over the shelf where bathymetric irregularities can stabilize their position. On the eastern sides of the ocean basins the thermocline is shallow enough to be brought to the surface, forming an upwelling frontal system along the shelf break.

Shelf-break fronts are characteristic of most passive continental margins in the low and mid-latitudes. They are quasi-permanent features fixed to the shelf/slope break, and in spite of their widespread occurrence, they remain one of the most enigmatic features of the ocean and their origin is still not well understood. The shelf/slope break itself is closely associated with the transition from normal to thinned continental crust on passive margins; the slope/deep sea floor transition marks the boundary between thinned continental crust and ocean crust. That the fundamental geologic boundary marking the edge of the continental block should have a counterpart as a frontal system in the overlying waters seems at first inexplicable. The water masses and their biota on either side of shelf-break fronts are so different that oceanographers studying them are often described as green-water (shelf) or blue water (ocean) specialists. Numerical models usually assume that the edge of the ocean is a vertical boundary along the 200 m isobath in order to avoid the complexity of dealing with shelf-break fronts. Only stray specimens of planktonic foraminifera are found inshore of the shelf-break fronts, but calcareous nannoplankton can be abundant in the shelf water although genera and species are largely different from their open-ocean counterparts. The fronts themselves tend to be biologically productive, attracting fish and are therefore of great interest to the fishing industry. The shelf waters often contain suspended sediment, whereas open-ocean surface waters are virtually devoid of detrital sediment other than dust. Shelf-break fronts also serve to trap organic matter on the shelf and to force nutrient recycling on the shelf. Whereas the water sinks along most fronts, it rises along shelf-break fronts. Shelf-break fronts separate the oceanic and inshore components of the Ekman layer. During the episode of chalk deposition, the shelf break fronts must have disappeared, as shown in Fig. 2.

2.2. Shelf-break fronts and the Ekman layer

The Ekman layer is the well-mixed surficial layer of the ocean driven by wind and influenced by friction (described as eddy viscosity) and the Coriolis parameter (Stewart, 2005). Transport in the surface Ekman layer is directed 90° to the right (left) of the wind direction north (south) of the equator. In the open ocean, the effective depth of the Ekman layer (D_E) is taken to be where the flow is opposite to that at the surface. Because the Coriolis parameter is involved, the effective depth of the Ekman layer goes to infinity at the Equator, and thins toward the poles. Following Stewart (2005), and assuming a long term average wind speed of 10 m/s (about 20 knots), the depth of the Ekman layer in the open ocean would be 107 m at a latitude of 30° and 80 m at 65°. These values approximate those obtained from observational data.

Shelf-break fronts occur where the entire water column on the shelf is the inshore component of the Ekman layer. Here the motion of the waters is also influenced by tides and bottom friction. The tides cause short-term stirring of the waters. The bottom friction slows currents on the shelf, turning them to the left (right) in the northern (southern) hemisphere. It also deepens the effective depth of the Ekman layer. Equatorward longshore winds in the mid-latitudes produce offshore surface water movement in both hemispheres. This lowers the sea surface near the coast, causing both upwelling and mixing. It also results in a pressure gradient normal to the shore, driving a geostrophic current along the coast in the same direction as the wind. One region in which these effects have been examined in detail is the North Atlantic Bight.

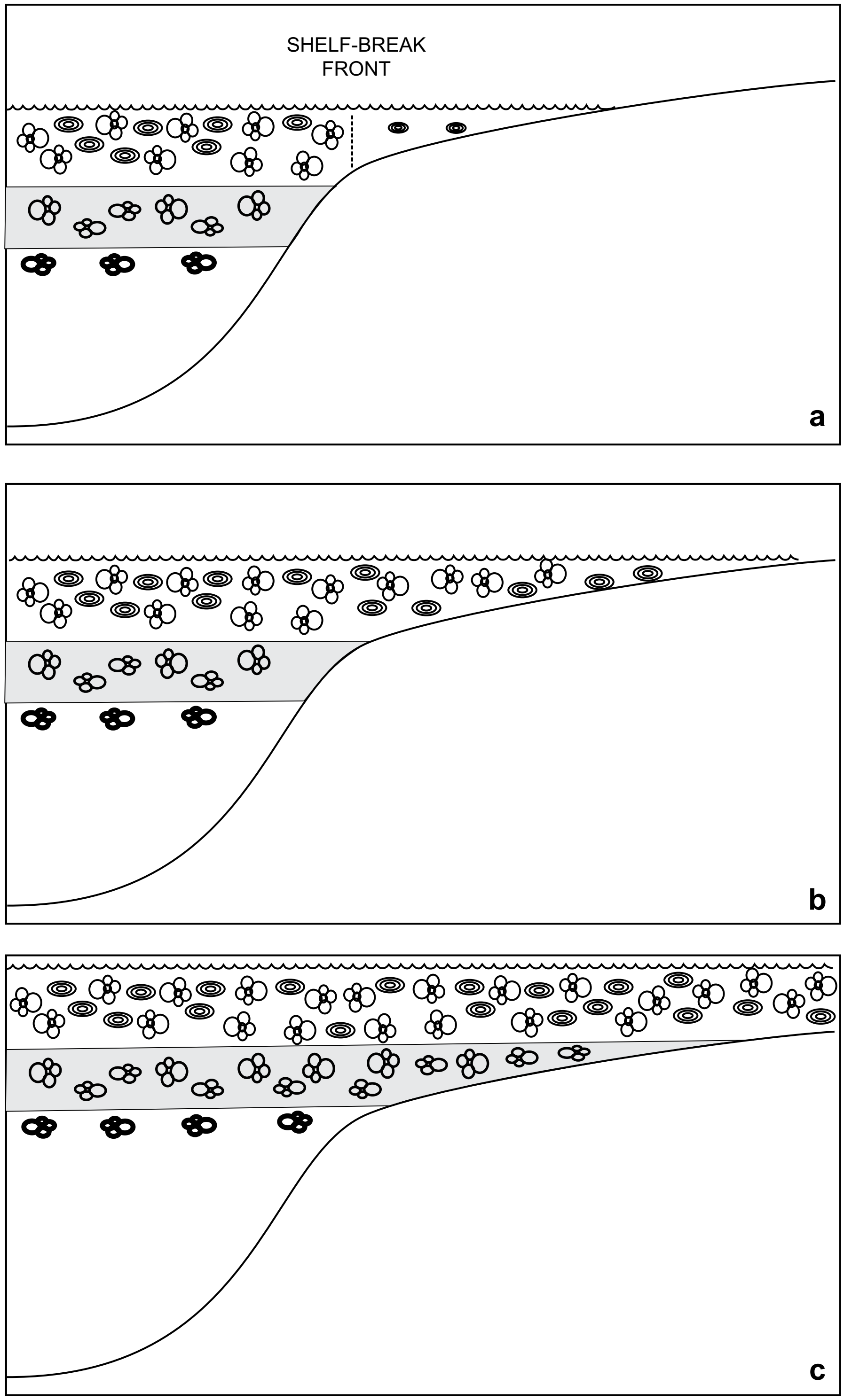

Fig. 2. Schematic history of calcareous nannoplankton (continued). a. Inhibition of most oceanic calcareous plankton from shelf seas by shelf break fronts (late Early Cretaceous and post mid-Maastrichtian); b. spread of surface water dwelling calcareous plankton onto shelves as sea level rises (Cenomanian); c. spread of surface, pycnocline, and subpycnocline calcareous plankton into epeiric seas at sea level maximum (Turonian-Campanian).

The shelf-break front of the North Atlantic Bight, off the east coast of North America, has been the subject of considerable study (Flagg and Beardsley, 1978; Ou, 1984; Beardsley et al., 1985, 1992; Chapman, 1986; Gawarkiewicz, 1991; Gawarkiewicz and Chapman, 1992; Gawarkiewicz et al., 1996, 2001). A cold southward current originating off Greenland flows over the shelf and is bordered offshore by the warm Gulf Stream. The Gulf Stream is a thick western boundary current, depressing the main thermocline well below the depths to which the wind can affect the water. Both the coastal current and the Gulf Stream are parts of the Ekman layer. Near the bottom the southward-flowing coastal water is slowed by friction, and turned to the left (seaward) by the Coriolis effect. The lighter inshore waters are forced beneath denser outer shelf waters, resulting in convective overturning and mixing. As a result well-mixed lower salinity, cooler shelf waters are juxtaposed against the more saline, warmer waters of the Gulf Stream. The frictional effect of the bottom on the shelf waters ends at the shelf break, which, in this area, is at a depth of about 100 meters, and the shelf waters rise toward the surface. Across the resulting frontal system there is a density contrast balanced by the momentum of the shelf waters. Gawarkiewicz and Chapman (1992) described and modeled the effect of the bottom friction. The offshore Ekman flux it induces is concentrated in the bottom boundary layer. To understand why the front should be tied to the shelf break, they changed the slope of the shelf in the model. Increasing the gradient of the shelf increased the influence of the shelf break, increasing the upward velocity of the water at the shelf break and intensifying the front. Decreasing the shelf gradient had the opposite effect. Modeling experiments by Condie (1993) confirm the stabilizing influence of steep topography below shelf-break fronts.

2.3. The depth of the shelf/slope break and epeiric seas now and in the Cretaceous

The presence of abundant oceanic plankton in Cretaceous epicontinental seas in such numbers as to form chalk implies that the shelf-break frontal systems that divide shelf and epicontinental sea waters from the open ocean today did not exist then. It could be argued that the breakdown of shelf-break fronts occurred simply because of the increased water exchange between the ocean and the epeiric seas, all of which would have had either a positive or negative fresh-water balance relative to the ocean proper. However, in many areas the onset of chalk deposition occurs well after the flooding of the epeiric sea began.

Older literature generally cites the depth of the shelf/slope break as 200 m, but this is simply the inflection point on hypsographic curves where the data were binned into 200 m intervals. Detailed bathymetric maps are needed to determine the depth of the shelf/slope break and these are not available except for a few areas. One of the best maps for examining the shelf/slope break is that for the East Coast and Gulf Coast of the United Sates (Belding and Holland, 1970). It shows that off Maine the shelf/slope break lies at about 130 m depth; from the Mid-Atlantic Bight southward it gradually shallows and in southern Florida it is only 30 m deep.

Why should the shelf break lie at such different depths? Hay and Southam (1977) argued that the shapes of the continental shelves and slopes today are a result of the peculiar nature of the Quaternary world. During most of the past 2 million years sea level has been much lower than it is today. They noted that where the shelf is covered by detrital sediment that can be eroded, the modern shelf break ranges between −100 and −150 m. Where the shelf consists of carbonates, which dissolves but does not erode, the shelf break is typically about 30 m. Along the eastern North American margin, the shallowing of the shelf/slope break is most pronounced off the Carolinas and Georgia where the transition from detrital to carbonate sedimentation occurs. Assuming that, except for brief periods of maximum glaciation, the Quaternary sea-level lowstands were about −100 m, and ignoring the complexities directly associated with ice loading, the shoreline would have been along the shelf break off Maine. Areas further south would have been 30, 50, and 70 meters above sea level respectively. The heads of the submarine canyons that occur between Maine and Virginia would have been cut by rivers emptying into the ocean. Hay and Southam (1977) suggested the depth of the shelf break was probably about 30 meters during times of relatively stable sea level before glacial oscillations began. This may have been the earlier Mesozoic "base-level" before the Cretaceous sea level rise. As sea level rose, not enough detrital material was delivered by rivers to aggrade the shelves to their former slope and they were "drowned."

How high was the Cretaceous sea level rise? Hay (1995a) noted that the episode of chalk deposition began in the mid-Cenomanian and ended in the latest Maastrichtian, corresponding to the highest sea-levels of the Phanerozoic. How deep was the water over the shelves and in the epeiric seas? Skelton et al. (2003) suggested that the minimum water depth required for the calcareous plankton to flourish in epicontinental seas must have been at least 200 meters, but this is based on direct analogy with modern conditions. There have been a number of attempts to calibrate long term sea level change and thereby estimate the depths in epicontinental seas using a variety of arguments or observations. The early attempts were based on estimates of the areas of the continents flooded, determined from paleogeographic maps (Sloss, 1963; Wise, 1974; Ronov, 1994). Conversion of area flooded to sea level requires the use of a hypsographic curve. Most conversions have used the modern hypsographic curve, but continental hypsography has changed with time (Bond, 1979; Harrison et al., 1981, 1983; Hallam, 1984, 1992; Wyatt, 1986, 1987; Harrison, 1990; Algeo and Wilkinson, 1991). Although based on different assumptions, most of the estimates are in general agreement that the episode of chalk deposition occurred when sea level was more than 130 m above that of today. Maximum sea-level rise deduced from continental flooding is about 200 m. The "Exxon" sea level estimates (Vail et al., 1977; Haq et al., 1987) have been based on coastal onlap in specific areas, but they also depend on assumptions concerning continental hypsometry. The later versions of the "Exxon curve" suggest that chalk deposition began and ended when sea level was about 180 m higher than today, and that maximum sea level was about +240 m. Haq and Al-Qahtani (2005) show the sea level as rising to about +300 m during the Aptian-Albian interval, and then declining beginning in the Maastrichtian. These estimates, derived from coastal onlap, are probably high because isostatic adjustment for water loading and erosion of sediment from the shelves during sea level lowstands were not taken into account. Others have tried to calibrate late Mesozoic sea-level by determining the position of shorelines on stable cratonic platforms. McDonough and Cross (1991) based their estimate of +265 to +286 m for the sea-level maximum on the eastern shoreline of the North American Western Interior Seaway. Sahagian and Holland (1991), Sahagian and Jones (1993), and Sahagian et al. (1996) used the Russian Platform for their calculations; their data suggest +130 m as the depth at the time chalk deposition began and ended, and a maximum sea-level of about 190 m. Finally, there are estimates based on displacement of seawater as a result of faster sea-floor spreading and emplacement of oceanic plateaus during the Cretaceous. Kominz (1984) estimated the rise in sea-level at 80 Ma due to sea-floor spreading to be 180 m. For the time of initiation of chalk deposition, this translates to +130 m. Harrison (1990) using sea floor spreading, Pacific volcanism, sediment accumulation, and several other factors arrived at an estimate of 227 m for the maximum, and about 200 m for the time of cessation of chalk deposition. In sharp contrast to earlier estimates, Miller et al. (2005), on the basis of backstripping sediments on the New Jersey margin, concluded that late Cretaceous sea levels were only 50–70 m higher than today, much lower than estimates based on other lines of reasoning. If true, this implies that interaction between shelf and open-ocean waters are extremely sensitive to sea-level changes.

2.4. Oceanographic significance of the breakdown of shelf-break fronts in the Late Cretaceous

Conditions changed from an earlier Cretaceous ocean, in which episodes of organic carbon deposition in the deep sea were both frequent and widespread, to a later Cretaceous ocean. According to Hu et al. (2005), oxidized deep sea deposits began to appear locally in the Albian. In the Tethys, continuous deposition of reddish carbonate-rich pelagic deposits (Scaglia rossa) began in the early Turonian However, the widespread, continuous occurrence of oxidized deep sea deposits coincides with the initiation of deposition of chalk on the continental blocks. It seems likely that shelf or epicontinental seas in close communication with the open ocean may have been sources for well-oxygenated dense waters which became the deep waters of the ocean basins. This requires that the waters over the shelves must have been stratified to the extent that surface ocean water could flow in, be modified, and denser water flow back into the ocean. How might this occur?

Assuming that the shelf-break fronts disappeared when sea level was about 130 m higher than today, that the shelf/slope break was originally about 30 m below sea level, and taking isostatic adjustment to the weight of the water into account, the shelf break would have been at a depth of −200 m. The implication is that the Ekman layer over the shelves was 200 m thick, which is reasonable as it is the depth that many ocean models take as the ocean boundary. Recall that the thickness of the Ekman layer over the shelves is greater than in the open ocean because of the increased turbulence due to bottom friction and tidal motions. As soon as the shelf-break fronts disappeared, the depth of the Ekman layer would be reduced to its oceanic value of about 100 m or perhaps even less, as discussed below. In effect, there would be a sudden change in the state of the ocean margin. Oceanic mixed-layer plankton could enter the shelf regions and penetrate into epeiric seas. This decrease in thickness of the Ekman layer would allow the upper part of the oceanic pycnocline to impinge on the shelf. As the water deepened, the entire pycnocline and finally sub-pycnocline waters could move onto the shelf and enter the epicontinental seas.

3. The problem of the warm, equable climate

As Neumayr (1883) and Uhlig (1911) had proposed, there was clearly a meridional climatic zonation, reflected in the modern use of the terms “Boreal” and “Tethyan” to describe the two major climate zones, even though the equator-pole temperature gradients were less than today. However, their “boreal realm” was in northern Europe, and on modern plate tectonic reconstructions it plots at a latitude of about 30° N (Voigt et al., 1999; Mutterlose et al., 2003). Hay (1995a) suggested that the Boreal and Tethyan realms are poleward and equatorward of the subtropical fronts that form the northern boundaries of the anticyclonic tropical-subtropical ocean gyres, but now it seems more likely that the boreal realm may be the region of high-salinity water associated with the descending limb of the Hadley cell (Voigt et al., 1999). The latitude corresponds with terrestrial geologic evidence for the Hadley cell boundary in North America (Ludvigson et al., 2004).

3.1. *Cretaceous temperatures and meridional temperature gradients*

Since the middle of the last century it has been recognized that the polar regions were warmer during the Mesozoic than they are today, producing a more “equable” climate (Sloan and Barron, 1990).

Today, the temperature difference between the equator and polar regions is about 50 °C in the Northern Hemisphere and 90 °C in the Southern Hemisphere, the difference being largely due to adiabatic cooling reflecting the elevation of Antarctica. Frakes (1999) summarized the data and estimates of Cretaceous sea surface and terrestrial temperatures. As shown in Fig. 3, there are four differing concepts concerning Cretaceous temperatures and meridional gradients: 1) tropical sea surface temperatures were the same as today, but polar temperatures were warmer - (5 to 8 °C) except when ice was present (0 to −5 °C); 2) the tropics were significantly cooler and mid-latitudes warmer than today; 3) tropical sea surface temperatures were 32 to 34 °C, with polar regions 10 to 18 °C; 4) tropical sea surface temperatures were about 42 °C and polar temperatures >18 °C.

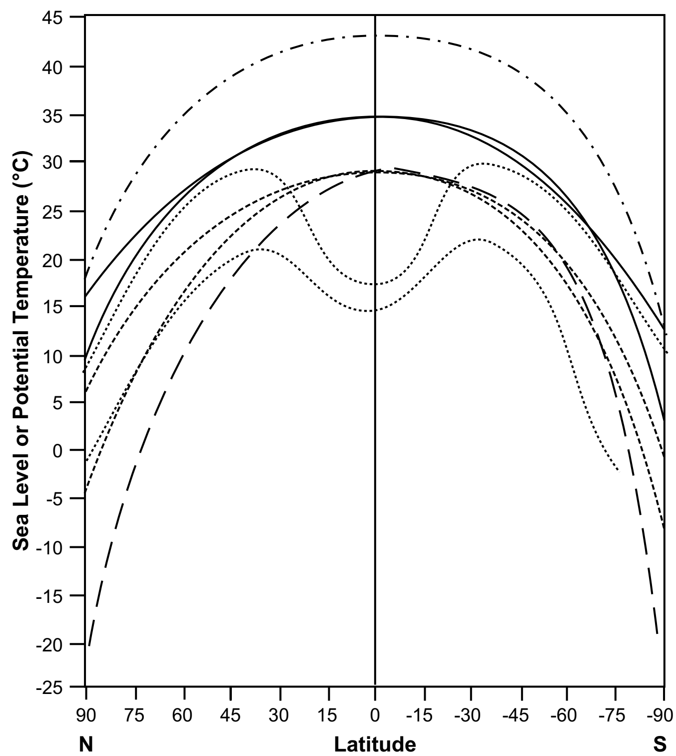

Fig. 3. Spectrum of ideas concerning meridional temperature gradients in the Cretaceous. The long dashed line represents the present day surface temperature at sea level, corrected for the elevation of Antarctica; note that the modern thermal equator is south of the Equator. There are four major groups of opinions concerning Cretaceous temperatures: (1) tropical sea surface temperatures were the same as today, but polar temperatures were warmer - (5 to 8 °C) except when ice was present (0 to −5 °C) - indicated by the two short-dashed lines; (2) the tropics were significantly cooler and mid-latitudes warmer than today - indicated by the upper saddle-shaped dotted line; the lower saddle-shaped dotted line is the present day potential temperature of waters at 100 m depth in the Atlantic; (3) tropical sea surface temperatures were 32 to 34 °C, with polar regions 10 to 18 °C - indicated by the two solid lines; (4) tropical sea surface temperatures were about 42 °C and polar temperatures >18 °C - indicated by the dot-dashed line.

Most geologic data have been interpreted as indicating that the Late Cretaceous meridional temperature gradient was 35 °C or less in both hemispheres, and that tropical temperatures were about 2 to 4 °C warmer than today (Barron, 1983a; Barrera et al., 1987; Barron et al., 1993; Pearson et al., 2001). Others (DeConto, 1996; Wilson and Opdyke, 1996; Hay et al., 1997; Norris and Wilson, 1998; DeConto et al., 1998, 1999; Clarke and Jenkyns, 1999; Pearson et al., 2001; Wilson et al., 2002) have argued for Cenomanian-Maastrichtian temperatures significantly warmer than today. Maastrichtian tropical temperatures of up to 32 °C have been interpreted from the $\delta^{18}O$ ratios of pristine planktonic foraminiferal tests by Pearson et al. (2001). On the basis of oxygen isotopes from fish tooth enamel from Tethyan carbonate platforms, Pucéat et al. (2003) concluded that during the Turonian the waters of the Tethys were 32 °C, or 4 °C warmer than the tropical oceans today. Recent interpretation of the evidence suggests that the temperature of the Atlantic equatorial region (Demarara Rise) may have been as warm as 42 °C in the late Turonian (Bice et al., 2006).

Polar temperatures have been cited as cold, based on Valanginian and late Aptian dropstones and glendonites, as discussed below, cool, based on Barremian to Albian southern hemisphere paleofloras (Rich et al., 1988) and permafrost-affected sediments bearing dinosaur fossils, warm (about 10 °C), based on Turonian-Coniacian paleofloras from the northern hemisphere (Parrish and Spicer, 1988; Herman and Spicer, 1996, 1997), both paleofloras and paleofaunas from the Late Cretaceous of the southern hemisphere

(Askin, 1992; Case et al., 1998) and on $\delta^{18}O$ ratios of planktonic foraminifera (Fassell and Bralower, 1999), or very warm (>14 °C) based on an Arctic fossil crocodilian (Santonian-Coniacian: Tarduno et al,. 1998; Huber, 1998). It is generally agreed that the warmest temperatures, both in the tropics and polar regions, occurred during the middle of the Cretaceous (Cenomanian-Coniacian), and that temperatures declined sharply during the Maastrichtian.

A special controversy has arisen over the interpretation of the $\delta^{18}O$ measured in the shells of planktonic foraminifera - termed the "cool tropics paradox" by D'Hondt and Arthur (1996). Maastrichtian sea surface temperatures were interpreted as being as low as 15 °C in the tropics, rising in the mid-latitudes and then declining toward the poles. Others (Sellwood et al., 1994; Huber et al., 1995; Price et al., 1997) have also argued that at times during the Cretaceous tropical sea surface temperatures were anomalously low (15–23 °C). Crowley and Zachos (2000) assessed pre-Pleistocene tropical sea surface temperatures derived from planktonic foraminifera and argued that the data show neither cooling nor significant tropical warming. Much of the argument over the "cool tropics paradox" rests in interpretation of the $\delta^{18}O$ in terms of what part of the signal is temperature and what part is the "salinity effect" (Poulsen et al., 1999; DeConto et al., 2000a). The "salinity effect" is actually the result of the hydrologic cycle differentiating the $\delta^{18}O$ of the upper layers of the ocean in response to evaporation and precipitation, both of which selectively leave the heavier isotope behind. Another aspect of this debate is that the interpreted cooler tropical sea surface temperatures are actually a diagenetic signal, perhaps reflecting cooler pore water temperatures in the sediment (Pearson et al., 2001).

No climate models have produced tropical temperatures cooler than those of the mid-latitudes. Horrell (1990), on the basis of energy balance considerations of the effects of increased cloud cover, increased zonal wind speed, and increased poleward energy transport, argued that it was unlikely that tropical sea surface temperatures could drop as low as 15 °C. The only way that surface waters in the tropics could become cooler than those of the mid-latitudes is if the upwelling of cool polar waters were so rapid and in such large volumes that it could overwhelm the solar input and wind mixing of the surface layer. In fact, something of this sort does happen today below the mixed layer. Fig. 3 shows the similarity of modern subsurface temperatures and the proposed cool tropics suggesting that the data were derived from Cretaceous foraminifera that lived below the mixed layer. The depression of modern tropical water temperatures below the mixed layer is the result of the general diffuse upwelling of water that sank at high latitudes. This is a complication in determining the absolute heat balance of the Earth, which still has great uncertainties. Because of this process the sea surface throughout the tropics is lower than it would be without this input of "negative heat" from below. The implication is that a similar process of ocean-wide upwelling of cool polar waters was operative in the Maastrichtian. This would be expected if, as some assert, there was polar ice in the latest Cretaceous.

3.2. *Was there ice in the Cretaceous greenhouse?*

In their reviews, neither Tarling (1978) nor Crowell (1982) cited any record of continental glaciation between Late Permian and late Paleogene times, even though the Antarctic continent was always near or at the South Pole. Fischer and Arthur (1977) placed the late Paleozoic (Carboniferous, Permian) and Triassic in their "Icehouse" climatic state and the Jurassic and Cretaceous in their "Greenhouse" state.

More recently it has been argued that there may have been extensive ice at some times during the Cretaceous, based on three lines of evidence: the sediments themselves, sea-level curves, and oxygen isotopes.

3.2.1. *Direct evidence from sediments*

Although there is no direct evidence for large continental ice sheets during the Mesozoic, it has been proposed that there was sea-ice formation and that small maritime glaciers may have existed. This evidence is chiefly in the form of dropstones ("outsized clasts") found in marine sediments (Frakes, 1979; Kemper, 1987; Frakes and Francis, 1988; Frakes et al., 1992; Ahlberg et al., 2002). However, there are arguments against the interpretation of dropstones as evidence for ice on land, particularly continental glaciers (Bennet and Doyle, 1996; Bennet et al., 1996). Pebbles and cobbles are often picked up by sea ice as it is driven onshore by the winds, and can subsequently be carried out to sea when the ice leaves the shore. Sea ice in the Arctic ocean contains sediment from a variety of sources (Nürnberg et al., 1994). In the modern Arctic "ice factory," the Laptev Sea, sediment-laden waters from the Yana, Lena and Khatanga rivers may flow out on top of the ice (Dethleff et al., 2000). In the Arctic today, coastal erosion is even more important than rivers in supplying sediment, and much of the eroded sediment is carried away by ice (Rachold et al., 2000).

The presence of glendonites has also been cited as evidence for Cretaceous ice (Kemper, 1987; Frakes and Francis, 1988, 1990; Sheard, 1991). Glendonites are pseudomorphs of stellate aggregates of crystals of ikaite ($CaCO_3$ $6H_2O$) (Shearman and Smith, 1985), a carbonate mineral that forms interstitially in rocks with organic-rich pore fluids under sub-zero conditions on the Antarctic shelf (Suess et al., 1982). However, ikaite is also known from the Zaire deep sea fan (Jansen et al., 1987), where the deep ocean temperatures are about 3 °C. Glendonites clearly indicate cold water but not necessarily perennial ice on the adjacent land areas.

The only unequivocal geologic evidence of ice is the recent discovery of a 2 m thick diamictite found in a Cretaceous sequence near the northern end of the Flinders Range of Australia (Alley and Frakes, 2003). Frakes et al. (1992) assigned the Middle Jurassic and Early Cretaceous to a "cool mode," because they believe there is evidence for significant sea-ice formation at high latitudes. Cooling events occurred at the Berriasian-Valanginian boundary, during the early late Valanginian and in the earliest Aptian. They assigned the Early Jurassic and Late Cretaceous to a "warm mode" because there is no evidence for ice-rafted detritus. The alternation between times with polar ice formation and times when there appears to have been no polar ice suggests significant changes in the state of the Earth's climate.

3.2.2. *Evidence from sequence stratigraphy and coastal off- and onlap*

The question of whether large amounts of ice may have been present on land during the Cretaceous has been raised ever since the Exxon sea-level curve was first proposed and a cause for the periodic Cretaceous eustatic variations sought. Cretaceous eustatic sea-level falls were originally thought by the Exxon group to be of the order of 100 m (Vail et al., 1977), and this was only slightly modified in subsequent updates (Hardenbol et al., 1981; Haq et al., 1987; Hardenbol et al., 1998; Haq and Al-Qahtani, 2005). This would require ice sheets larger than the present Antarctic (=59 m of sea level change). These and most other estimates of sea level change from coastal onlap studies did not take into account isostatic adjustment due to the changing water load and to the changing sediment load if erosion occurred during the sea level lowstand (see discussion in Hay and Southam, 1977). Assuming an eustatic sea level fall of 100 m, and consolidated sediment with a density of 2200 kg/m^3 eroded to the same cross-shelf gradient it had before the sea level fall, coastal onlap analysis would indicate a sea level fall of 300 m. However, the coastal plain sediments commonly involved coastal onlap studies would have been very young, less consolidated, and more susceptible to erosion at the time of the sea level fall. In this case the error involved in not taking isostatic adjustment and erosion into account could be as much as a factor of 6. Making this correction, it is evident that short term Cretaceous sea-level falls were probably more in the order of a few tens of meters at most.

Stoll and Schrag (1996) argued that two rapid sea level falls in the Berriasian-Valanginian were documented by changes in the Sr isotopic composition of seawater. They argue that as Sr-rich aragonite sediments were converted to Sr-poor calcite when carbonate banks were exposed to the atmosphere. They assumed the 50 m sea level fall of Haq et al. (1987) was correct and argued that the Sr evidence confirmed this magnitude of sea level change. They did not take isostatic adjustment into account. Miller et al. (1999), estimated a short-term early Maastrichtian sea-level fall to be about 40 m, correcting for water loading (but not for sediment erosion). Miller et al. (2003, 2004) used a complete backstripping routine and concluded that the sea-level falls were > 25 m, and the end Campanian- early Maastrichtian fall was probably 40 m. Gale et al. (2002) estimated that short-term sea level falls in the Cenomanian were 2 to 20 m.

Gale et al. (2002) noted that there are three possible causes of sea level change, 1) changes in the amount of groundwater stored in the pore space of sediments, 2) thermal expansion of seawater, and 3) ice on land.

As an alternative solution to the problem of explaining eustatic changes in ice free times, Hay and Leslie (1990) calculated the present volume of pore space in sediments above sea level that might be filled or emptied of groundwater to be about 25×10^6 km^3, slightly more than the amount of ice on land today. If this pore space were originally filled with air and then filled with water, it would produce a sea level lowering of 76 m, or 50 m after isostatic adjustment. However, the present conditions are an anomaly in earth history because of the massive offloading of uncompacted sediment from the continental blocks onto the ocean floor in response to the sea level changes since the initiation of glaciation of Antarctica at the beginning of the Oligocene, and especially after the initiation of northern hemisphere glaciation since the Miocene. They suggest that in the mid-Cretaceous the pore volume above sea level may have been as much as twice that of today. If so this implies that changes in groundwater storage in response to Milankovitch cycles or other climatic changes could be responsible for tens of meters of sea level change. However, from analysis of a Triassic climate model, Jacobs and Sahagian (1993) found that during that arid episode of earth history Milankovitch cycles could have caused changes in sea level of only about 10 m on a 100,000 year time scale.

Gale et al. (2002) cite 1 m per degree C as a value for the thermal expansion of seawater. The compressibility of sea water is strongly temperature and to a lesser extent salinity dependent. Calculated from the equation of state for seawater (Millero et al., 1980; Millero and Poisson, 1981), taking into account the variability of compressibility with pressure, assuming a salinity of 35‰ (Late Cretaceous, Hay et al., 2006), and assuming the temperature changes to be restricted to the waters below 100 m, the thermal expansion per degree C ranges from 0.27 m with a deep water temperature near 0 °C, to 0.65 m with a deep water temperature near 10 °C, to 0.96 m with a deep water temperature near 20 °C, and to 1.22 m with a deep water temperature near 30 °C. Deep water temperature changes of 4° to 8 °C, such as suggested by Stoll and Schrag (1996) would induce only a 4 to 8 m sea level change if the deep waters were relatively warm. However, going from a warm deep ocean to a cold deep ocean (25° to 0 °C) would cause a sea level fall of 20 m. This is the effect that might be expected in a switch from warm saline tropical to cold polar deep water sources. For the Early Cretaceous, when average ocean salinities were close to 40‰, the increased compressibility of the water would cause almost one additional meter of sea level fall as deep waters cooled from 25° to 0 °C.

3.2.3. Evidence from oxygen isotopes

Stoll and Schrag (1996) described two $< 10^6$ yr variations of δ^{18}O in the Early Cretaceous, one between 125 and 126 Ma of about +1‰ and the other between 128 and 129 Ma of about +1.4‰. They noted that these coincide both with sea-level falls on the Haq et al. (1987) sea level curve and with spikes in the Sr isotope curve. Abreu et al. (1998) noted a number of coincidences between sea level variations deduced from sequence stratigraphic studies and the oxygen isotope record, suggesting that the presence of ice on land is the most reasonable cause. Barrera et al. (1997) and Barrera and Savin (1999) presented evidence for general cooling of ocean surface waters toward the end of the Cretaceous indicated by a gradual δ^{18}O increase of about 0.5‰. Superimposed on this long-term trend they found two short episodes during the Maastrichtian (71–69.5 Ma and 68–67.5 Ma) when both the δ^{18}O ratios of surface-dwelling planktonic foraminifera and of benthic foraminifera increased by about 1‰. They interpreted these events as reflecting changes in the thermohaline circulation of the ocean that resulted in cooling of oceanic intermediate waters. Subsequently, Miller et al. (1999) suggested a glacioeustatic mechanism for this cooling episode because of its synchroneity with the rapid sea-level fall they estimated to be of the order of 25 to 40 m. Further short-term cooling episodes have been suggested for the late Turonian because of the coincidence of cooling indicated by δ^{18}O of brachiopod shells and the southward migration of boreal invertebrates into the Tethys region (Voigt and Wiese, 2000; Voigt, 2000). Stoll and Schrag (2000) reported positive isotopic excursions of about 1‰ in lower Cenomanian and upper Turonian-Coniacian carbonates in Italy.

The problem inherent in interpretation of δ^{18}O ratios in the carbonate fossils is that they are dependant on four factors: 1) temperature at which shell formation took place, 2) vital effects of the organism secreting the skeletal mineral, 3) the δ^{18}O composition of the seawater, and 4) carbonate concentration of the seawater. The temperature effect is well known from experimental data (Erez and Luz, 1983; Bemis et al., 1998). Vital effects are also known from culturing experiments, and investigators making paleotemperature estimates are careful to use shells from organisms where they are minimal or where the correction factor is known. Unfortunately, the δ^{18}O composition of the water itself depends on four factors 1) the secular trend in δ^{18}O (about −0.016‰/my: Veizer and Hoefs 1976), 2) the volume and δ^{18}O of fresh water in streams, lakes, and in the groundwater system (whereby usually only the latter is large enough to affect the value of δ^{18}O in the ocean), 3) the evaporation/precipitation balance at the sea surface (commonly referred to as the "salinity effect" as discussed above), and 4) the volume and δ^{18}O composition of ice on land. The first two factors have been generally neglected, at present the third affects only surface dwelling organisms, and hence attention has been concentrated on the ice volume effect.

The "salinity effect" is very small in the deep sea today because most ocean deep waters come from cold saline (34.7‰) sources that have similar characteristics. Only isolated marginal seas with unusual deep water sources, such as the Mediterranean and Red Seas, have different compositions of deep water δ^{18}O. This may not have been the case in the Cretaceous when there may have been competing warm and cold deep water sources with different salinities. Not enough is known about the paleobathymetry of the Cretaceous oceans to be sure that the δ^{18}O values of benthic foraminifera measured at one location are representative of the ocean as a whole.

Recent advances in understanding of Cretaceous paleogeography suggest that the amount of fresh water may not be negligible for the Cretaceous. Changes in the volume of groundwater probably have a minor effect on δ^{18}O because the isotopic composition of warm rain is close to that of the ocean and most of the land area was in lower latitudes. The volume of water in rivers is always negligible as is that of modern lakes. However, during the Cretaceous the Arctic Basin became isolated and may have been a very large lake (Kazmin and Napatov, 1998). Throughout most of the Early Cretaceous the Arctic was open to the Pacific as the South Anyui Ocean Gulf. As the connection with the Pacific closed during the Albian, the Arctic became isolated, communicating with the world ocean only through shallow seaways across western Siberia, between Norway and Greenland, and through the western interior of North America. Paleontologic evidence indicates that throughout much of the Late Cretaceous the Arctic Ocean had a low salinity (Fisher et al., 1994). The balance of salt in the Arctic Ocean would have been very delicate because inflow of saline waters could only occur through the depths of the long epicontinental seaways. Short-term changes in the fresh water balance, such as might occur on Milankovitch time scales could have affected $\delta^{18}O$ ratios in the world ocean. Today the volume of the deep Arctic Basin is about 12.3×10^6 km^3 (Jakobsson, 2002); the maps of Kazmin and Napatov (1998) indicate that at 100 Ma it was about 2/3 as large (about 8×10^6 km^3). If it were filled with high latitude rainwater ($-20‰$), its effect on the $\delta^{18}O$ composition of the seawater would have been about $+0.12‰$; if it were filled with water from snowmelt ($-30‰$) the effect wold have been about $+0.19‰$.

The ice volume correction for the Quaternary was correctly approximated by Shackleton (1967) based on the assumption that the temperature of ocean deep waters did not change appreciably between glacials and interglacials. This meant that the change in $\delta^{18}O$ of about 1.8‰ measured in deep-sea benthic foraminifera could be interpreted in terms of ice volume. His initial results were unexpected and controversial because in order to match the sea-level change between the last glacial and present they implied that the $\delta^{18}O$ of the ice sheets had to be $\leq -40‰$, not the -20 to $-30‰$ measured in polar snow today. However, his results were later confirmed by ice cores on Greenland and Antarctica. Further refinements have shown that the temperatures of ocean deep waters do change slightly between glacials and interglacials, but these are a relatively minor correction. Unfortunately, we cannot use Shackleton's method to estimate Cretaceous ice volumes because there is no unequivocal independent evidence for the temperatures of ocean deep waters during the Cretaceous.

The discovery that the carbonate concentration of seawater affects the $\delta^{18}O$ of the foraminiferal tests is a complication with implications for the Cretaceous that have not been explored. Spero et al. (1997) found that with a carbonate ion concentration range of 100 to 642 μmol/kg, the $\delta^{18}O$ in shells of *Orbulina universa* changed by 1.5‰. They note that range in carbonate concentrations during the late Quaternary is thought to be from about 100 to 400 μmol/kg. Almost nothing is known about carbonate ion concentration in the Cretaceous when atmospheric CO_2 was much higher and ocean chemistry different from that of today (Lowenstein et al., 2003; Paytan et al., 2004; Berner, 2004; Holland, 2005).

Barrera and Savin (1999) and Miller et al. (1999) used the Maastrichtian change in $\delta^{18}O$ of 0.5‰, and the assmption that the Cretaceous ice had a composition of $-40‰$ to argue that an ice sheet about 40% as large as that of modern Antarctic would be required to account for a 25 m sea-level fall. Alternatively, it might be assumed that ice formed during the Cretaceous would be more like that of polar precipitation today and would have a $\delta^{18}O$ value near 20‰. In this case the ice sheet required would be about the same size as the Antarctic today.

3.2.4. Where was the ice?

Assuming that there was ice on land during part of the Early Cretaceous and during the late Maastrichtian, the question is - where was it?

Oxygen isotopes in freshwater molluscs have been cited as evidence of alpine glaciation in the Late Cretaceous-early Paleocene Laramide mountains of the northern United States (Carpenter, 2003). A noted above, the first direct evidence of moving ice, a 2 m thick diamictite of Early Cretaceous age, has been found in southern Australia, but this may also represent mountain glaciation.

Mountain glaciation alone cannot account for significant sea level changes. All of the Quaternary mountain glaciers accounted for only about 10 m of sea level change. If there was extensive ice during the Cretaceous, it must have been as an ice sheet on a continental block. The Cretaceous geology of North America and Eurasia is relatively well known and does not have areas of uncertainty large enough to accommodate an ice sheet of the size required to account for 30–40 m of sea level change. The only likely candidate is Antarctica.

DeConto and Pollard (2003) have performed numerical experiments to investigate the Oligocene initiation of glaciation on Antarctica. They found that lowering of atmospheric CO_2 concentrations to 840 ppm coupled with favorable orbital configurations resulted in initiation of glaciation on higher elevations. Their topographic map of pre-Oligocene Antarctica was based on present bedrock topography isostatically adjusted after removal of the present ice cover. Cretaceous topography may have been significantly different. Rifting of Antarctica from the other parts of Gondwana was initiated at the end of the Hauterivian, about 130 Ma (Hay et al., 1999). Hay (1981) and Southam and Hay (1981) proposed that continental rifting involves a long (100 my) pre-rifting period of uplift of a broad region (800 km). The uplift, in the order of 1.5 to 2 km culminates in the 20 my prior to the actual rifting, and decays away over a period of 20 my as the rift widens. For the Antarctic, this would mean a 7500 km long 1000 km wide ridge, which would develop and decay between 150 and 110 Ma. The uplift is shown schematically in Fig. 4, a plate tectonic reconstruction for 130 Ma. Adding 1.75 km to the 1 to 2 km elevations typical of pre-glacial Antarctica, this implies a maximum area of 7.5×10^6 km^2 with elevations ranging from 1.5 to 4 km.

The extent of possible ice cover depends on the area of land above the 0 °C annual isotherm. Today the 0 °C annual isotherm is "timberline," the boundary between evergreen forests and tundra. Otto-Bliesner and Upchurch (1997) and DeConto et al. (1998, 1999) have shown that this boundary is critical for the accumulation of permanent snow and ice. Because of the meridional temperature gradient at sea level, this boundary is low at the poles and high at the equator. Its elevation depends on the lapse rate, the decline in temperature with elevation. The lapse rate depends on the degree of saturation of the air with water vapor, and can vary from place to place. Today the lapse rate is typically cited as 0.6 °C/100 m. We do not know the lapse rate for the possible episodes when ice on land may have been present during the Cretaceous, but it must lie between the dry adiabatic lapse rate (air without any water vapor), and the wet adiabatic lapse rate (air saturated with water vapor), and was probably closer to the wet adiabat than it is today. Fig. 5 shows the change in temperature of air with elevation for dry and wet adiabats, after Hay (1996).

Fig. 5 is not wholly accurate for the Cretaceous because large uplifts, notably Tibet and western North America, and the ice caps on Antarctica and Greenland have displaced air onto a lowered sea level, changing sea level pressure from what it was in the Cretaceous and Paleogene. Modern sea level pressure was originally intended to be 1 Bar (1000 mb or hPa) but it was established by averaging pressure in Europe and North America which are lower than the global average. Modern sea level pressure (global average) is 1013.25 hPa.

Based on Fig. 5, Fig. 6 shows the elevations at which the temperature would be 0 °C assuming lapse rates for either dry or wet (water vapor saturated) air, and sea surface temperatures for three models, with -20 °C, 0 °C, and 15 °C in the polar regions and 28 °C, 32 °C and 42 °C at the equator. It also shows schematically the

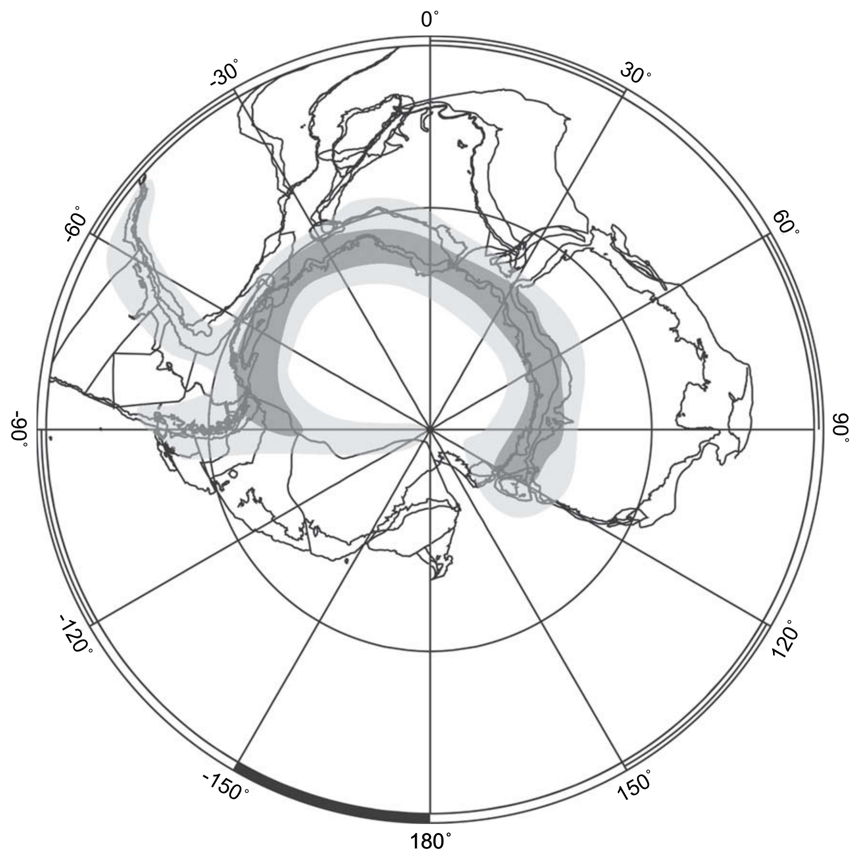

Fig. 4. Areas of higher elevation on Antarctica and the surrounding fragments of Gondwana shortly before and after separation (130 Ma). The grey areas could have been sites of upland ice sheets.

elevations of Cretaceous uplifts and mountains in North and South America, Eastern Asia, and Antarctica (after Ronov et al., 1989). It is evident that, although there may have been ice on mountains at higher latitudes, the only area in which large ice sheets could occur is Antarctica. The accumulation of ice results in further elevation of the surface, but after isostatic adjustment 1 km of ice raises the surface by only about 0.7 km. Accumulations of 2 km of ice over the entire uplift would result in a sea level fall of 48 m, corresponding to the estimate of Stoll and Schrag (1996). However, it is likely that the area of any ice sheet was probably less and the thickness greater; it is unlikely that the ice would have reached the shoreline.

The Cretaceous diamictite found near the northern end of the Flinders Range of Australia (Alley and Frakes, 2003) lies opposite the sector of Antarctica having the lowest elevation (DeConto and Pollard, 2003), so this is an area in which mountain glaciation rather than an ice sheet might be expected. Additional evidence, in the form of diamictite, may lie hidden beneath the modern Antarctic Ice Sheet, and beneath the sediments of the Falkland Plateau and other areas of the South Atlantic. However, as Hay (1981) noted, the uplift hypothesis postulates that subsequent to the subsidence after rifting, the regional slope reverses, from away from the rift to toward the rift. As a result, the material eroded from the more elevated areas and deposited on the flanks is then re-eroded and deposited in the rift valley. Thus it is to be expected that most or all of the diamictite deposited on the flanks of the uplift would have been destroyed and redeposited as normal marine detrital sediment.

Explaining Maastrichtian ice is more difficult. The areas uplifted by the Laramide Orogeny are too small to have been able to support an ice sheet of the size required (>25 m of sea level fall). Antarctica was now almost completely surrounded by relatively warm water which could have been an ample source for snow if atmospheric greenhouse gas levels were low enough to permit polar cooling, i.e., relatively brief glaciations in the interior of Antarctica of the type described by DeConto and Pollard (2003) as leading to the continents full scale glaciation. Siberia is problematic as the site of an ice sheet unless the Arctic Ocean had a perennial sea ice cover so that the polar sea-surface air temperatures were well below 0 °C.

3.3. The cause of the warmer climate

The cause of the global warmth of the Mesozoic has been an enigma that has puzzled both geologists and atmospheric scientists. Brooks (1928) had hypothesized that during periods of reduced meridional temperature gradient, the atmospheric circulation must have decreased in intensity. He also suggested that the major winds would shift 10 to 15° poleward. Flohn (1950) proposed that the position of the subtropical highs is a function of the meridional temperature gradient, moving poleward as the gradient decreases. Damon (1968) had noted that times of global warmth coincided with high sea level stands, suggesting that there might be a relation between global warmth and albedo and/or ocean heat transport. Kraus et al. (1979) concluded that atmospheric heat transport must be less during periods of reduced meridional temperature gradient.

When simulation of pre-Pleistocene paleoclimates using general circulation modeling began in the late 1970's, the intent was to select a climate as different from those of today and the Pleistocene glacials as possible, e.g., a time of global warmth and reduced meridional temperature gradient. The Cretaceous was selected as the initial target because its paleogeography and globally warm

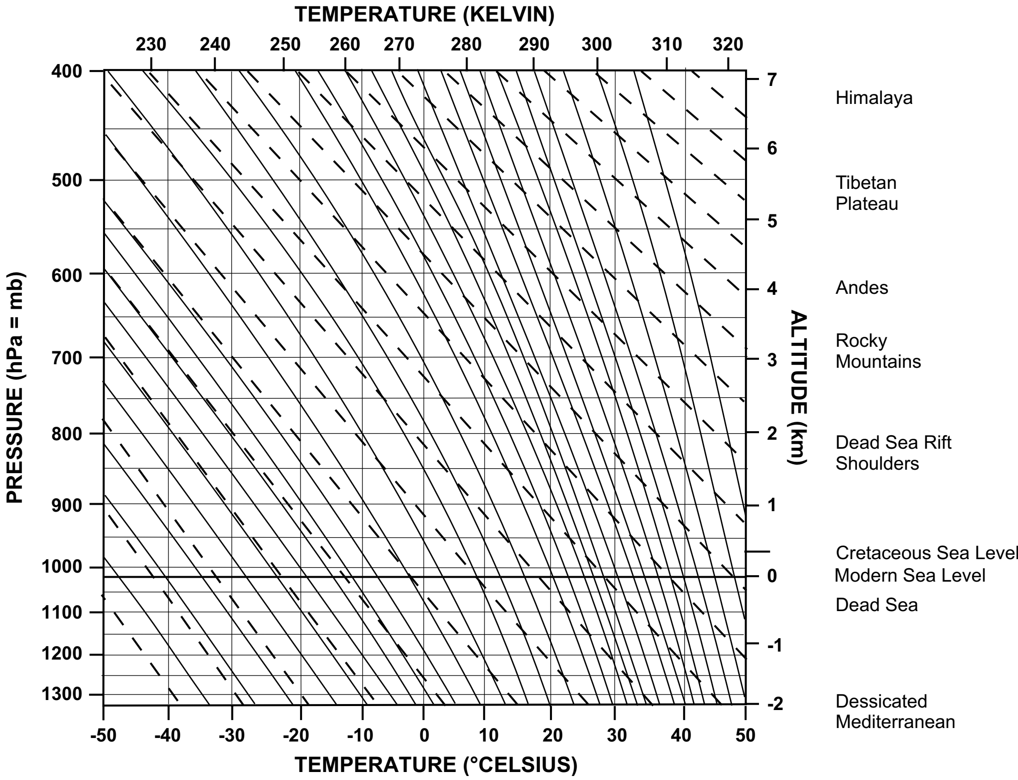

Fig. 5. Dry and wet (saturated air) adiabats for a range of temperatures and pressures that may have existed on the Earths surface during the Phanerozoic. The adiabats show the change in temperature with elevation. Changes in sea level pressure due to plateaus, mountains, below sea-level depressions and changes in the volume and/or mass of the atmosphere are not taken into account. The dry adiabats are essentially straight lines sloping from lower right to upper left, shown by dashes. The wet adiabats are steeper and curved, shown by solid lines.

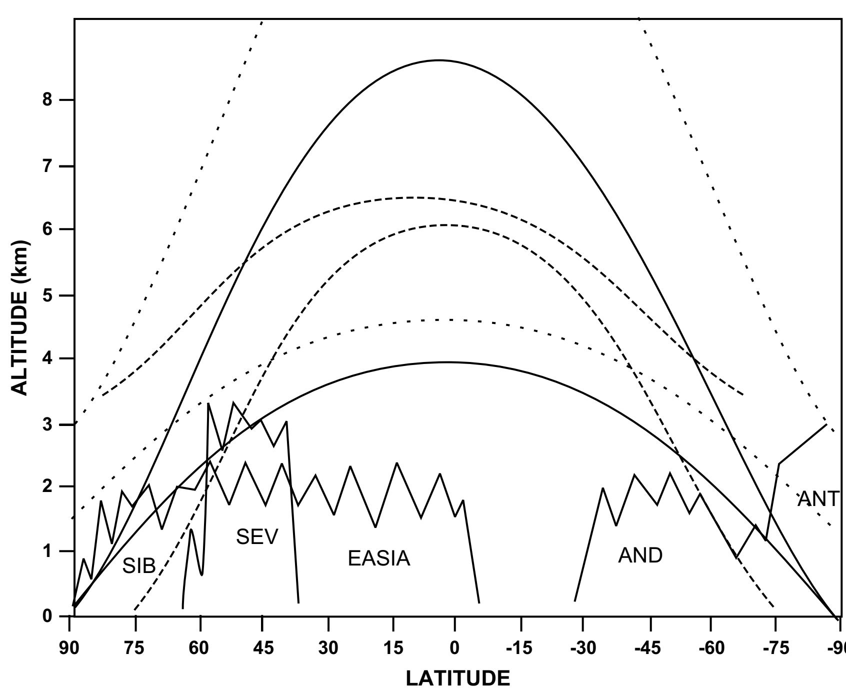

Fig. 6. Elevation of the 0 °C isotherm assuming polar temperatures of about −20 °C and equatorial temperature of 28 °C (modern day), shown by the dashed line, polar temperatures of 0 °C and equatorial temperature of 32 °C, shown by the solid lines (Early Cretaceous and Maastrichtian?), and polar temperatures of 15 °C and equatorial temperature of 42 °C shown by the dotted lines (extremely warm Late Cretaceous). The lower of each pair of lines is the dry adiabat, the upper is the wet adiabat. The real lapse rate, which determines the real elevation of the 0 °C isotherm lies near the wet adiabat. The locations and probable elevations of Cretaceous mountains and plateaus are indicated: SIB = Siberia; SEV = Sevier Highlands of western North America; EASIA = lower latitude Eastern Asia; AND = Cretaceous Andean mountains; ANT Antarctica, mostly plateau.

climatic conditions are well known (Barron, 1983a,b; Crowley, 1983).

The greatly reduced meridional temperature gradient of the Cretaceous has been regarded as a climate paradox that arises from three paradigms:

(1) both the atmospheric and ocean heat transport systems of the Earth are driven by the winds which are forced chiefly by the meridional temperature gradient;
(2) during the Cretaceous the meridional temperature gradient was much less than it is today; which implies that the atmospheric and ocean heat transport systems were much more effective then than they are today;
(3) the reduced meridional temperature gradient means that the forcing of the winds controlling Cretaceous atmospheric and ocean heat transport was less than it is today.

Many of the modeling experiments that explored basic aspects of this paradox were performed in the early 1980's (Barron et al., 1981b; Barron, 1984a,b, 1985; Barron and Washington, 1982a,b, 1984, 1985). It was soon realized that the "paradox" resulted from the assumption that the components of meridional heat transport by the atmosphere, sensible heat (including potential energy), latent heat, and kinetic energy (presently about 40%, 60%, and <1% respectively) did not change their proportions on a warmer Earth.

The contrasts between modern and Cretaceous climates have been attributed to the differences in paleogeography, greater greenhouse gas content in the atmosphere, differences in atmospheric and ocean heat transport, or to some combination of all of these factors.

3.3.1. *Effects of different paleogeography*

Initial investigations suggested that the displacement of the continents and the high sea level would account for the global warmth (Barron et al., 1980). A planetary albedo model (Thompson and Barron, 1981) suggested that the greater extent of low albedo water and reduced area of more reflective land during the mid-Cretaceous could be a significant factor in explaining Cretaceous warmth. Using an energy balance model, Barron et al. (1981b) determined the meridional energy transport required to produce a gradient with equatorial temperatures the same as today but with mean annual polar temperatures of 0 °C. For Cretaceous paleogeography the model suggested that ice-free polar conditions could exist if the present-day meridional energy transports could be maintained. An experiment with elevated tropical temperatures showed an increase in poleward latent heat flux, but this was offset by a decrease in the poleward sensible heat flux. Both models suggested a planetary warming of 2 °C.

The initial results of experiments related to simulation of the mid-Cretaceous climate (100 Ma) were reported by Barron and Washington (1982a,b). A complete account was presented by Barron and Washington (1984), and the simulations have been discussed in the light of subsequent studies by Barron and Moore (1994). Five sensitivity experiments were performed, using mean annual solar insolation. A baseline experiment compared the results of simulations for the present with the Cretaceous paleogeography of Barron et al. (1981a), with the Antarctic divided into East and West Antarctica and the blocks translated to conform with Tarling (1978), using the paleotopography of Scotese et al. (1981) as shown in the Cenomanian maps of Parrish and Curtis (1982, figs. 3, 4). For each of the five sensitivity experiments leading from the present to Cretaceous paleogeography one geographic variable was changed, and the results of each experiment were compared with the results of the preceding experiment. The first experiment used present-day geography and topography and explored the effect of changing the albedo of the surfaces of Antarctica and Greenland, from snow-covered to snow free. The effect was to increase the global average temperature 0.8 °C and to raise the temperature of the Antarctic by 10–15 °C. The second experiment was to remove the topography from the present-day continents to explore the effect of relief on the climate. The global average temperature increased 1.1 °C although the temperature of present lowland areas decreased 1–7 °C; the temperature of the Antarctic increased by 15 °C. The third experiment moved the flat continents to their Cretaceous positions, but keeping present-day shorelines. This produced the largest change, an increase of 3.1 °C in global average temperature. The greatest difference was in the northern polar region, where temperatures increased 21 °C. The fourth experiment changed the shorelines to reflect the higher sea-level of the mid-Cretaceous. Although it was expected that the much greater area of epicontinental seas would result in significant warming, the global average temperature actually decreased by 0.1 °C. The fifth experiment added Cretaceous topography; the global average temperature decreased 1.1 °C, exactly offsetting the effect of removing topography from the present land areas carried out in experiment 2. Comparing the simulation with mid-Cretaceous geography directly to the simulation for the present, Barron and Moore (1994) cited a global average temperature increase of 4.8 °C for the Cretaceous. Tropical temperatures increased 2 °C, but the North Pole became 15 °C and the South Pole 39 °C warmer than in the simulation for the present-day. The atmospheric circulation patterns were not shifted poleward in the Cretaceous simulations, as Brooks (1928) and Flohn (1950) had suggested should be the case, but the subtropical highs in the Northern Hemisphere moved slightly southward to be located over the Tethys. Barron and Washington (1982b) noted that the atmospheric circulation was not "sluggish" in spite of the lowered meridional temperature gradients. They suggested that this might be because either equatorial temperatures were slightly higher as a result of increased evaporation and latent heat transport, or the polar warming eliminated the polar temperature inversion. These experiments demonstrated that continental displacement, topography, and sea level alone or in combination cannot fully account for the warmth of the polar oceans or the globally higher temperature of the mid-Cretaceous. Experiments with higher sea level did not have the expected effect of lowering the Earth's albedo and increasing the amount of energy absorbed because the shallow seas became cloud-covered and reflective.

Experiments to determine the effect of the changes between modern and highly schematic Cretaceous paleogeography using a model with a seasonal cycle were performed by Crowley et al. (1986). They found increases of the Cretaceous summer temperatures over Greenland and Antarctica of 10–12 °C and 5–8 °C respectively. They noted that these changes were of a magnitude comparable to those predicted for increased atmospheric CO_2.

A numerical atmospheric-ocean-land climate system model, GENESIS, was developed by Starley Thompson and David Pollard of NCAR (U.S. National Center for Atmospheric Research) specifically to investigate climates of the Earth's past (Wilson et al., 1994; Pollard and Thompson, 1995, 1997; Thompson and Pollard, 1995a,b, 1997). Using GENESIS Barron et al. (1992, 1993, 1994, 1995) simulated the mid-Cretaceous climate using the same paleogeography that served as a boundary condition for the earlier AGCMs. The Cretaceous simulation indicated a global temperature 0.2 °C less than today, although the polar regions were warmed significantly. The South Pole was 47 °C warmer in summer and 27 °C warmer in winter.

Poulsen et al. (2001) investigated the effect of the Early Cretaceous opening of the gateway between the North and South Atlantic, using NCAR's Parallel Ocean Climate Model (PCOM). They found that with $4 \times CO_2$ the interior waters of the ocean were significantly warmed by opening the gateway, although surface temperatures remained about the same. Poulsen et al. (2003) used a simplified Cretaceous paleogeography and the Fast Ocean-Atmosphere Model (FOAM) to reinvestigate the Atlantic sector. In these simulations they found that opening the North Atlantic-South Atlantic gateway could indeed have contributed to the Cenomanian-Turonian thermal maximum. They suggested that some of the thermal effect of the proposed CO_2 increase attributed to a "hidden pulse of sea-floor spreading" proposed by Wilson et al. (2002) and Bice and Norris (2002) might actually reflect the opening of the gateway.

Paleogeographic changes alone are insufficient to explain the Cretaceous climate. Alternative solutions to the problem have been sought in increased greenhouse gas content of the atmosphere and in increased poleward ocean heat transport. Until recently it was assumed that there was only one significant greenhouse gas, CO_2.

3.3.2. Effects of increased atmospheric greenhouse gases and clouds

A simple calculation of the temperature of the Earth's surface required to reradiate energy in balance with the incoming solar insolation (238 W m^{-2}) using the Stefan-Boltzmann Law indicates that the surface temperature should be 255 K or −18 °C (Peixoto and Oort, 1992). However, the mean surface temperature of the Earth today is about 15 °C; the 33 °C difference between the −18 °C calculated and the +15 °C observed is due to the greenhouse effect of triatomic and more complex atmospheric gases (water vapor, CO_2, O_3, and CH_4), the water droplets forming clouds, and aerosols.

Shortly after Arrhenius (1896) discovered the role of CO_2 as a greenhouse gas, Chamberlin (1899) suggested that changes in atmospheric CO_2 might be responsible for the late Cenozoic ice ages. He later concluded that water vapor, being a more powerful greenhouse gas, was the responsible agent. The major greenhouse gas today, and probably throughout the Phanerozoic is undoubtedly water vapor, but its concentration is a function of temperature. Water vapor provides a positive feedback to increase the greenhouse effect introduced by other gases. A review of the development of ideas relating to perception of the significance of greenhouse gases has been presented by Fleming (1998).

After Chamberlin's rejection of the idea, the potential role of increased CO_2 was neglected until revived by Budyko and Ronov (1979) and Budyko et al. (1987). In the context of exploring possible future climate change, Manabe and Wetherald (1980) using an AGCM with a "swamp ocean" (in AGCM modeling, a "swamp ocean" has no heat capacity and performs no heat transport, but can serve as an infinite source of moisture) found that increasing atmospheric CO_2 by a factor of four resulted in an increase of tropical temperatures by 4 °C, but that the warming effect is concentrated in the polar regions where temperatures increased 15 °C. This launched the modern concern over the potential climatic effects of burning fossil fuels. They concluded that the smaller meridional temperature gradients were maintained by a large poleward latent heat transport. Using the mid-Cretaceous paleogeography, mean annual insolation experiment described by Barron and Washington (1984) as a base, Barron and Washington (1985) found that increasing atmospheric CO_2 by a factor of four increased the global average temperature (presently 15 °C) by 3.6 °C. The warming took place at all latitudes. They noted that because of the warmer temperatures induced by higher levels of atmospheric CO_2, the hydrologic cycle would have been strengthened.

Barron and Moore (1994) reported the result of experiments using the mid-Cretaceous GENESIS (V. 1) simulation with a seasonal cycle as a base, and two, four and six times present atmospheric CO_2. With two times present CO_2 equatorial ocean temperatures were close to 30 °C. Four times present CO_2 produced equatorial ocean temperatures slightly above 30 °C and a warming of the global average temperature of 5.5 °C compared to the GENESIS simulation with present-day CO_2. This approached the minimum temperature estimate for the mid-Cretaceous based on the paleontologic evidence available at the time (Barron, 1983a). Polar temperatures warmed more than those in the tropics although they remained cold enough for sea ice to form. Six times present CO_2 produced equatorial temperatures near 32 °C.

DeConto (1996) completed a more realistic simulation of Late Cretaceous (Campanian) climate using the GENESIS (V.2) Earth System model. The results were reported briefly in Hay et al. (1997) and more extensively in DeConto et al. (1998, 1999, 2000b). This new simulation produced a meridional temperature gradient that falls within the envelope of possible temperatures suggested by paleontologic data. Most importantly, the simulation indicated a global average temperature of 28 °C, 13 °C warmer that at present. Tested with a proxy formation model, it correctly defined the regions of occurrence of Late Cretaceous evaporites (Wold et al., 1995). It also correctly predicted the composition of high-latitude floras (DeConto et al., 1998).

All of these models had a suspicious feature: the atmospheric pressure systems retained polar highs throughout the year even though there was no polar ice. With a water covered North Pole surrounded by land, it would be expected that the land would be warmer than the water in summer and cooler in winter, resulting in polar atmospheric highs in summer and lows in winter. For an ice-free Antarctic, where the paleogeographic situation is exactly the opposite, with polar land surrounded by water, there should be a low in summer and a high in winter. The fact that these simulations failed to show the expected seasonal reversals caused Hay (2000b) to suspect that something was missing.

Flögel (2001) performed simulations of different orbital configurations using GENESIS (V.2) with Turonian paleogeography and 1882 ppm CO_2 (5 × yr 2000 CO_2, 7 × preindustrial CO_2) concentrations. The temperature distributions were similar to those of DeConto (1996) but there was an important difference. These simulations, using higher atmospheric CO_2 levels than previous experiments, showed the expected seasonal reversals of the atmospheric pressure systems at the poles. Why the seasonal reversals do not occur in simulations with lower CO_2 concentrations remains unclear, but I suspect that "tuning" of the model to simulate modern conditions introduces a bias which requires high greenhouse gas concentrations to overcome.

Recently, it has been suggested that part of the greenhouse effect might be due to methane, which, by weight, has a significantly higher radiative effect than CO_2 . Today natural methane emissions are chiefly from ruminant animals and marshes; both have their Cretaceous counterparts. Chin et al. (1991), in a study of coprolites, suggested that dinosaur flatulence may have contributed significant amounts of methane to the Cretaceous atmosphere. Ludvigson et al. (2002, 2003) used oxygen and carbon isotopes to suggest that there were significant methane fluxes from high latitude mid-Cretaceous wetlands. Bice et al. (2006) concluded that if CO_2 alone were to account for the very warm (about 42 °C) tropical temperatures it would have to be at concentrations of 3500 ppm. They suggested that the greenhouse may have been the result of a combination of CO_2 and methane. Jenkyns (2003) proposed that the sudden warming at the onset of the early Aptian OAE 1a (Selli Event) and the middle Albian OAE 1b (Paquier Event) may reflect releases of methane as a result of the dissociation of clathrates.

An important aspect of the Cretaceous climate which has been largely neglected is the effect of cloud types and cloud cover. Most of the climate models used for Cretaceous simulations have used the cloud parameterization of Slingo and Slingo (1991) or some variant thereof, but very different results were obtained using the Hadley Meteorological Center cloud parameterization (Valdes, 2002).

Sloan and Pollard (1998) have proposed that polar stratospheric clouds might be responsible for polar warmth. Water vapor clouds form in the stratosphere in regions with very low temperatures and are essentially restricted to the polar regions during polar night. By inducing radiative warming in the lower stratosphere, these clouds could cause up to 20 °C of surface warming during the winter. It is believed that the water vapor in the stratosphere originates from the decomposition of methane. If methane were more abundant in the Cretaceous atmosphere, polar stratospheric clouds could indeed have played a significant role in polar warming.

3.3.3. Does the Earth have a thermostat?

As soon as quantitative estimates of warmer Cretaceous temperatures were discussed in the late 1970's, they were questioned by some atmospheric scientists. It was argued that the Earth has a natural thermostat which would limit temperature increases. Newell et al. (1978) and Newell and Dopplick (1979) proposed that there is a natural limit to tropical temperatures, imposed by the large amounts of latent heat involved in evaporation. They argued that a very large increase in insolation would be required to provide the energy necessary for the increased evaporation from a warmer sea surface. Ramanathan and Collins (1991, 1992, 1993) proposed that the thermostat is more complex. They argued that surface evaporation cannot be the only limiting mechanism because evaporation adds moisture to the boundary layer, enhancing the greenhouse effect, and because the Hadley circulation advects water vapor into the warmest regions of the ocean, reducing evaporation there. They proposed that increased evaporation and atmospheric vapor content would induce formation of cirrus clouds high in the troposphere; these would reflect incoming solar radiation and limit surface temperature increases. Recently this effect has been documented, and has been termed "Global Dimming" (Stanhill and Cohen, 2001). The specific cause of the change in solar energy reaching the Earth's surface was not known but it was suspected that it was anthropogenic. Recently these effects have been documented more thoroughly (Charlson et al., 2005; Wielicki et al., 2005; Pinker et al., 2005; Wild et al., 2005). Strong indication that the dimming effect is indeed anthropogenic and is related to the contrails produced by jet aircraft in the stratosphere came from analysis of temperatures measured in the United States before and after the terrorist attack of September 11, 2001. All commercial planes over the US were grounded for three days, and contrails were eliminated regionally. The result was an increase in the daily temperature range, with an increase in the maximum and decrease in the minimum temperatures (Travis et al., 2002). Wild et al. (2005) argue that dimming was occurring until the late 1980's, but since then the CO_2 increase has begun to overwhelm the increased cloudiness due to anthropogenic activities. If there is a natural thermostat effect caused by increasing cloudiness in response to global warming, it has yet to be detected. The proposed increased-cloudiness thermostat is the opposite of the effect for polar warming via stratospheric clouds proposed by Sloan and Pollard (1998).

Walker et al. (1981) proposed that CO_2 is the natural thermostat that has regulated the temperature of the Earth since shortly after the planet accreted. The idea was that because the carbonation weathering reaction is temperature controlled, it would act as a negative feedback mechanism to keep long-term atmospheric CO_2 variations within a narrow range. Explanations for the "Snowball Earth" (Hoffman et al., 1998) demonstrated that under certain circumstances, the CO_2 thermostat could fail.

The idea of an effective thermostat mechanism that would limit sea surface temperatures to approximately their present values is in conflict with our current understanding of the geologic evidence from the Cretaceous.

3.3.4. Effect of vegetation on the climate and the cold continental interior paradox

Over the past decade climate simulations have begun to include the effects of vegetation. One of the most important features of realistic plant cover is the effect of evergreen forests versus tundra in enhancing polar warmth (DeConto, 1996; Otto-Bliesner and Upchurch, 1997; DeConto et al., 1998, 1999, 2000b; Upchurch et al., 1998, 1999). Under winter conditions, a snow-covered tundra produces a high albedo so that cold conditions can persist into the spring and summer months, whereas evergreens mask the high albedo of snow and result in higher uptake of solar energy after the sun reappears.

The "cold continental interior paradox" arises because Cretaceous climate simulations generally predict below freezing winter temperatures in the continental interiors, even with warm poles (Schneider et al., 1985). Different climate models have produced very different results (Jones, 2001), some indicating winter temperatures as cold as about −30 °C in the continental interiors (DeConto et al., 1999). Paleofloral assemblages in North America (Wolfe and Upchurch, 1987) and Siberia (Vakhrameev, 1991; Spicer et al., 2002) and crocodiles in Mongolia (Lefield, 1971) all indicate winter temperatures above freezing. DeConto (1996) provided a possible solution to this dilemma. In his simulation the Campanian vegetation was generated iteratively by the Equilibrium Vegetation Ecology (EVE) model (Bergengren et al., 2001). Only plant life forms with known Late Cretaceous analogues were included, and water-conserving C_4 plants (e.g., tropical and subtropical grasses), which did not emerge until the latest Cretaceous (Gale et al., 2001) were excluded. For the Campanian simulation, the C_4 grasses were replaced by herbaceous C_3 plants (forbs). Winter temperatures in the continental interiors rose significantly in response to the increased latent heat flux resulting from the higher transpiration rates of the C_3 plants.

3.3.5. Effects of changes in the meridional energy transport systems

Oceanic and atmospheric energy transports are intimately interrelated to each other and to the global radiation balance. The present day poleward transport reaches a maximum at about 43° N and S, and is estimated to be about 5 Petawatt (10^{15} W) (Trenberth and Caron, 2001). The total transport is divided between atmospheric and oceanic components. At present, about 1/3 of the global transport is oceanic and 2/3 atmospheric (Vonder Haar and Oort, 1973; Oort and Vonder Haar, 1976; Trenberth, 1979; Carissimo et al., 1985), but the relative proportions of atmospheric and oceanic transport vary with latitude and ocean. Today ocean heat transport dominates over atmospheric energy transport at low latitudes and at higher latitudes the situation is reversed. Stone (1978) proposed that the total meridional energy flux must remain constant, and that any change in heat transport by the ocean would be compensated by a change in the atmospheric energy transport. If this were true, the low meridional temperature gradients of the Cretaceous would be very difficult to explain.

Atmospheric energy transport. On a global scale most of the energy is transported by the atmosphere; hence, a first approximation to solving the problem of Mesozoic warmth should come from atmospheric general circulation model (AGCM) simulations. These typically show that Cretaceous atmospheric water vapor content was higher and the winds slower than at present. A greater proportion of the energy advected by the atmosphere was accomplished through latent heat transport.

Barron et al. (1989) and Pierrehumbert (2002) have recognized that temporal variation of the intensity of the hydrologic cycle had broad implications for paleoclimatology. DeConto's (1996) simulation of Late Cretaceous climate demonstrated that the mechanisms of atmospheric energy transport are quite sensitive to the overall planetary temperature. This is because of the nonlinear relation of water vapor pressure and temperature (Clausius-Clapeyron equation, see discussion in Peixoto and Oort, 1992).

Fig. 7 shows that there is a much greater potential for water vapor to enter the atmosphere at higher surface temperatures, and that the global temperature must have a strong effect on the total amount of water vapor in the atmosphere. The increased potential for latent heat transport is the major difference between modern and Cretaceous climates.

The temperature-saturation vapor pressure relation is important for three reasons: First, the phase conversion from ice or water to vapor and back involve conversions of large amounts of energy into latent heat. The latent heat of melting of ice at 0 °C is about 0.335×10^6 J kg^{-1}. The latent heat involved in the transformation of water to vapor depends on the temperature; it is 2.428 10^6 J kg at 20 °C and 2.258 10^6 J kg at its sea-level boiling point, 100 °C. If ice sublimates to vapor at 0 °C, the latent heat of sublimation is about 2.836×10^6 J kg^{-1}, the sum of the latent heat of fusion and vaporization at that temperature. The latent heat involved in the transformation of water or ice to vapor is more than three orders of magnitude greater than the specific heat of air, 1004 J kg^{-1} K^{-1} at constant pressure or 717 J kg^{-1} K^{-1} at constant volume. Latent heat is the most efficient means by which the atmosphere can transport energy. Today it is largely the release of latent heat at the Intertropical Convergence Zone that drives the circulation of the Hadley Cells.

Secondly, water vapor is the most important greenhouse gas, affecting in particular the radiative balance of the troposphere. It is responsible for net radiative cooling of the troposphere compensated by latent and sensible heating resulting from convection (Manabe and Strickler, 1964). It is also the most effective greenhouse gas in the troposphere, trapping much of the infrared radiation ($\lambda = 5$–$7\ \mu$, 15–100 μ) from the surface of the Earth and heating the surface and lower atmosphere though reradiation. This produces a powerful positive feedback, with the warmer surface releasing more water to the atmosphere, enhancing the greenhouse effect. Part of the radiative cooling from emission of long wave radiation by H_2O is compensated by latent heat release. The vapor condenses to form water droplets which are opaque to radiation passing through the remaining gap in the center of the Earth's infrared radiation ($\lambda = 7$–$15\ \mu$) spectrum. The effect of water droplet condensation is, by forming clouds, to completely close the window allowing radiation from the Earth's surface to escape directly into space.

Lastly, water vapor changes the molecular weight of the air, and with it the gas constant for air. The molecular weight of dry air (78% nitrogen, 21% oxygen, 1% argon) is 28.9×10^{-3} kg mol^{-1}, and the specific gas constant for dry air is 287.2 J kg^{-1} K^{-1}. The molecular weight of water is 18×10^{-3} kg mol^{-1}. Under constant temperature and pressure, equal volumes of air contain equal numbers of molecules (Avogadro's Law). Moist air, containing lighter vapor molecules has a lower molecular weight; vapor-saturated air at 40 °C contains about 37 g H_2O per kg air; it has a molecular weight of 28.5×10^{-3} kg mol^{-1}, and its specific gas constant is 291.0 J kg^{-1} K^{-1}. The change in density of the air is a decrease of 1.4%, which is the same as the change of density that results from heating dry air about 5 °C. At temperatures above 30 °C, the changes in water vapor content of the air begin to rival temperature changes in their effects on the density of the air, and differences in absolute humidity can play a significant role in regional pressure distributions.

The atmosphere is not saturated with water vapor; the relative humidity (proportion of vapor observed to the proportion of vapor at saturation) varies from about 70% over most ocean areas to about 40% over deserts (Peixoto and Oort, 1992, 1996). However,

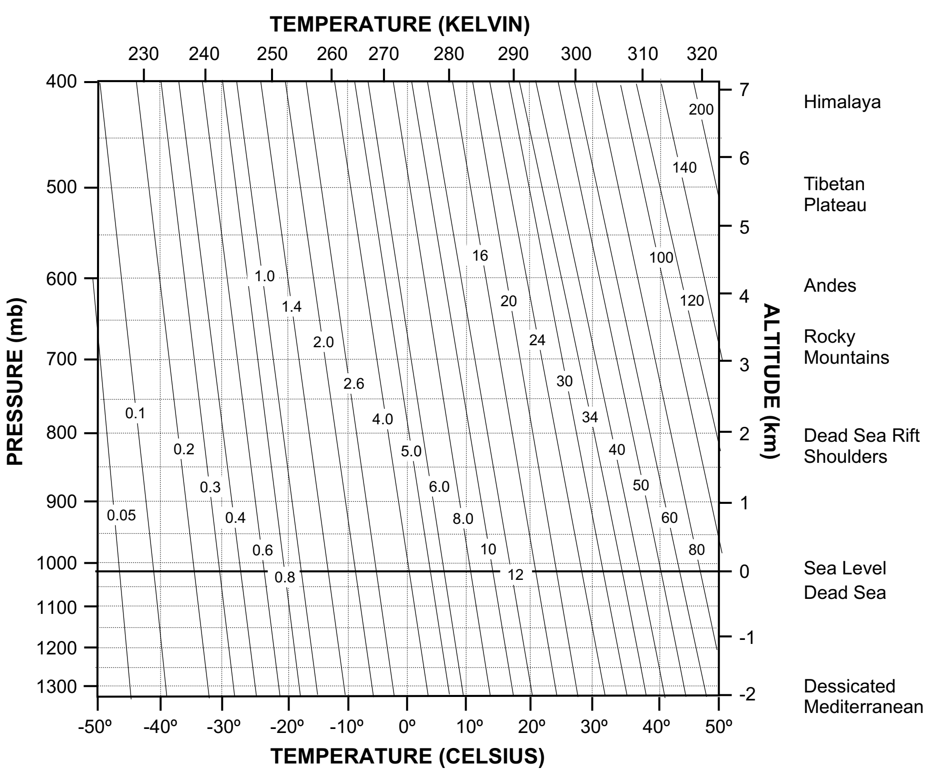

Fig. 7. Quantities of water vapor at saturation in air for a range of temperatures and pressures that may have existed on the Earth's surface during the Phanerozoic. Water vapor content is expressed in terms of g H_20/kg air. For the very warm tropical ocean temperatures postulated by Bice et al. (2006) saturated air at sea level would be 8% water vapor. The molecular weight of dry air is 28.97; the molecular weight of air at 42 °C saturated with water vapor is only 28.09 so that the presence of water vapor begins to rival temperature in controlling atmospheric pressure.

discounting the regional differences due to presence or absence of water surfaces and vegetation, the relative humidity of the lower atmosphere tends to remain constant even with large temperature fluctuations. This means that the absolute humidity (g H_2O per kg air) varies greatly with temperature. Today absolute humidities over the surfaces of the equatorial and polar oceans are 16 and 2 g kg^{-1} respectively. Despite these differences, the relative humidity at the 850 hPa (=millibar) level in the atmosphere in both regions is 70% . The absolute humidity over the Sahara is about 6 g kg^{-1}, and the relative humidity at 850 hPa is about 30%.

In a study examining the effect of changing poleward ocean heat transport during the mid-Cretaceous using an idealized coupled ocean-atmosphere model, Schmidt and Mysak (1996) found that the largest increase in meridional energy transport compared to models of the present day was actually in the atmospheric latent heat transport. Their scenarios with sinking tropical ocean waters resulted in an especially enhanced hydrologic cycle contributing up to 1 PW of additional poleward energy transport.

Ocean heat transport. Most discussions of ocean heat transport are concerned only with the surface ocean. Today the tropical-subtropical gyres carry energy in the form of sensible heat poleward, but the transport is blocked at 45° to 50° latitude by the Subtropical and Polar frontal systems. Only where there is large-scale high-latitude sinking of water into the ocean interior can the warm water cross these fronts and be advected to higher latitudes. However, the surface transport is only half of the ocean energy system. The other half is in the thermohaline circulation at depth, which today brings cold polar waters to low latitudes where it upwells slowly beneath the tropical-subtropical gyres, cooling the surface waters from below. Because of the paucity of relevant direct measurements, most estimates of the present-day magnitude of ocean heat transport have been calculated by indirect methods. The heat transport is different in each ocean basin, but reaches a maximum at latitudes between 10° and 30° in each.

Estimates of modern meridional ocean heat transport vary widely. To determine modern global values, Bryden and Hall (1980), Roemnich (1980), and Hall and Bryden (1982) used hydrographic observations to make direct estimates, and found the global ocean heat transport maximum to be 1 to 2 PW. Hastenrath (1982) calculated the meridional ocean heat transport from the heat fluxes at the ocean surface and obtained an estimate of the maximum of about 3.5 PW. Using a planetary energy balance method, Vonder Haar and Oort (1973), Oort and Vonder Haar (1976), Trenberth (1979), and Carissimo et al. (1985) inferred ocean heat transport from satellite measurements of the radiation at the top of the atmosphere and global data on atmospheric transports from weather balloons and rockets. These estimates are higher than those made using direct methods. Carissimo et al. (1985) proposed the highest ocean heat transports, 3.6 PW at 30° N, where they estimated the atmospheric transport to be only 2 PW. Ganachaud and Wunsch (2000) derived estimates from data produced by the World Ocean Circulation Experiment (WOCE), arriving at values of 1 PW for net southward transport in the southern hemisphere and 2 PW for northward transport in the northern hemisphere.

The effects of possible variations in ocean heat transport have been explored by Covey and Barron (1988) and Covey and Thompson (1989). Unreasonable values for current transports are required if the Cretaceous polar warmth were to be explained by increased ocean heat transport alone.

Chamberlin (1906) proposed a reversal of the internal circulation of the ocean as a means of explaining the difference between the globally warm climate of the Late Cretaceous and that of today. Today, heat is carried poleward by the surface currents of the ocean and cold water that sinks in the polar regions returns equatorward at depth to upwell in the tropics. Chamberlin reasoned that if the interior circulation were reversed, deep water sinking in the tropics would flow poleward and upwell there, adding to the heat supplied by the surface currents. Brass et al. (1982a,b) described this possible mode of ocean circulation as part of the "warm saline bottom water hypothesis."

Johnson et al. (1996) described the latitudinal fluctuations of the boundaries of tropical reefs in the Cretaceous of the Caribbean region. They concluded that these fluctuations were related to thermal variations resulting from changes in the ocean heat transport system. They believed that at times of high sea-level stands poleward movement of both surface and subsurface waters increased, reducing temperatures in the tropics and causing the reef belt to contract.

Most climate models specify the ocean heat transport as a boundary condition. However, Schmidt and Mysak (1996) conducted experiments to determine how the ocean energy transport might have differed in the Cretaceous. They concluded that ocean heat transport was greater than for the present day, but that, as noted above, latent heat transport by the atmosphere was even more important.

On the basis of an ocean GCM, Hotinski and Toggweiler (2003) concluded that the circumglobal Tethyan Passage might have induced wind-driven upwelling of relatively cold deep water at low latitudes. They found that without any change in radiative forcing, a low latitude circumglobal passage increased northern high-latitudetemperatures by 3 to 7 °C, while tropical temperatures cooled by up to 2 °C. With increased radiative forcing (outgoing longwave radiation reduced by 4%, about equal to 6 times modern CO_2) the Tethyan flow produced warming of northern high latitudes by 7 to 11 °C, while tropical temperatures remained within 3 °C of those of the present.

All of these estimates on ocean heat transport in the Cretaceous have assumed that the circulation of the ocean was analogous to that of today. Hay et al. (2005) proposed that the modern structure of the ocean did not arise until bipolar ice appeared on the Earth at the end of the Eocene. They postulate a very different structure and circulation patterns for the ocean in ice-free states of the Earth.

4. The anoxic ocean problem

Broecker (1969) published an informative abstract entitled "Why the deep sea remains aerobic," observing that today the supply of oxygen to the deep ocean overwhelms any possible supply of organic carbon from the surface waters. At the same time, the Deep Sea Drilling project recovered black shales from the deep Atlantic off the Bahamas. This initiated a debate which has become the "anoxic ocean conundrum." The problem has plagued many paleoceanographers: to achieve anoxia in the deeper waters replenishment of O_2 from the surface must be slower than the rate of decomposition of organic matter; this would seem to imply sluggish vertical circulation. However, the large export of organic matter from the photic zone required to use up all of the available oxygen implies high productivity in the surface waters. Because the supply of nutrients from land is very small compared that from upwelling of nutrient-rich intermediate and deeper waters, the high productivity of the surface waters implies that there must be vigorous circulation! The conundrum exists because many of us have made the assumption that even though organic carbon-rich sediments were deposited over large areas, the ocean structure and behavior were similar to what we observe today (Southam et al., 1987; Arthur and Sageman, 1994; Hay, 1995b; Kruijs and Barron, 1990; Bralower et al., 1994).

A detailed discussion of oceanic anoxic events is beyond the scope of the paper, but modern insightful accounts and syntheses have been published by Jenkyns (1999), Larson and Erba (1999), Jones and Jenkyns (2001), Erbacher et al. (2001), Leckie et al. (2002), Erba (2004), Tsikos et al. (2004), Nederbragt et al. (2004), Robinson et al. (2004), Pancost et al. (2004), Kuypers et al. (2004), Bornemann et al. (2005), Gallagher et al. (2005), and Heimhofer et al. (2006) among others.

4.1. The structure of the modern ocean

Today's ocean is characterized by low latitude stratification with anticyclonic gyres and high latitude convection with cyclonic gyres, as shown schematically in Fig. 8. Why does the modern ocean have this structure? The densest water on the ocean surface is the cold water of the polar regions. The density surfaces (isopycnals) that crop out on the ocean surface descend to depth equatorward. Relatively stable wind systems, with tropical Easterlies, mid-latitude Westerlies, and polar Easterlies, force the gyres in each ocean basin in each hemisphere. The day to day atmospheric circulation may seem to be chaotic, especially at higher latitudes, but the ocean currents integrate the wind signal over longer periods of time. The boundaries between the tropical-subtropical anticyclonic and polar cyclonic gyres are frontal systems that form beneath the most intense westerly winds where the curl of the wind stress changes most rapidly. The frontal systems act as barriers to the poleward transport of heat by the ocean. They are particularly effective in the modern Pacific Ocean (Rooth, 1982; Hay, 1995a, 2002).

The frontal region between the gyres is bounded equatorward by the subtropical front and poleward by the polar front. Because of the Coriolis effect, the water underlying the strongest of westerly winds is forced equatorward more intensely than that at lower latitudes. It is also more dense than the water at lower latitudes. As a result it sinks into the ocean interior along and adjacent to the Subtropical Front to form the thermocline/pycnocline which floors the tropical-subtropical anticyclonic gyres, as shown in Fig. 8. Water sinking along and adjacent to the Polar Front forms the cool low-salinity Intermediate Water that underlies the pycnocline. Alternatively, warm, high-salinity Intermediate Water can also enter the ocean as outflow from negative fresh water balance marginal seas. Today the Mediterranean and Red Seas produce large, distinctive water masses in the North Atlantic and Arabian Sea respectively.

In today's anticyclonic gyres between the Subtropical Fronts at 45° to 50° latitude and the equator, the ocean mixed-layer (Ekman layer) thickness generally ranges between 40–60 m on the eastern side and 120 to 160 m on the western side of the major ocean basins. At the center of the ocean gyres, at latitudes 20°, the mixed layer thickness may increase to about 250 m. Although the anticyclonic gyres of the tropics and subtropics are huge, they are superficial features involving only about 5% of the ocean's water. Their rotation forces relatively less dense warm water downward onto the more dense cold deep water that sinks in the polar regions. The water downwelling in the center of the anticyclonic gyres forms the Central Water masses of the lower part of the gyres. This reinforces development of a well-defined pycnocline floor to the anticyclonic gyres. Today the pycnocline is generally a thermocline, creating the modern temperature-stratified ocean structure of the mid- and lower latitudes. The modern mixed layer is underlain by the "ventilated thermocline," which forms the transition to the 100–200 m thick "main thermocline," overlying the cold waters of the ocean interior. The ventilated thermocline is derived from the downwelling of water in the tropical-subtropical anticyclonic gyres, and hence actively exchanges water with the Ekman layer. The main thermocline is more isolated.

Although the ocean thermocline is such an important feature, it is still not well understood (Boccaletti et al., 2004). Today it represents a remarkably stable balance between the wind-driven vertical mixing of the Ekman layer, the downwelling forced by the anticyclonic rotation of the oceanic gyres, the upwelling of deep water that sank in the polar regions, and the salt and heat diffusion processes usually described as "vertical diffusivity" or "vertical eddy viscosity," as shown in Fig. 9. Depending on the external forcing and the temperature and salinity differences between the surface and deeper waters, the maintenance of the pycnocline can be quite complex (Figueroa, 1996).

Beneath the anticyclonic gyres the internal pressure gradients reverse, so that the water below them has cyclonic rotation. This

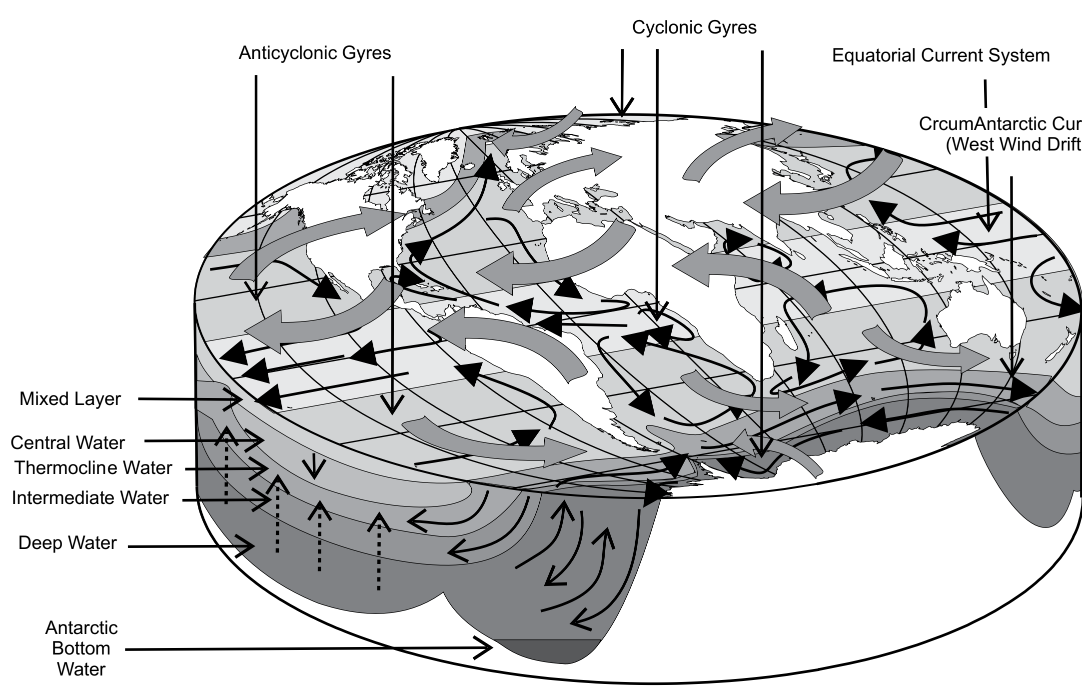

Fig. 8. Structure of the modern ocean (after Hay et al., 2005). Note that the general circulation is anticyclonic in the tropics and subtropics, and cyclonic in the polar regions. There is one notable exception, a region of cyclonic circulation in the eastern tropical South Atlantic (Stramma and England, 1999).

forces the denser deep ocean waters upward, further intensifying the pycnocline. The deep sea cyclonic gyres of the tropics and subtropics extend from the pycnocline to the ocean floor. Their circulation is strongly affected by the bathymetry so that they are not strict mirror images of the surficial anticyclonic gyres.

In the Arctic Ocean, the water temperatures are near freezing and temperature changes have only a small effect on density. Because of the strong positive fresh-water balance of the Arctic due to inflow from rivers, the surface layer is freshened and the well-developed pycnocline is a halocline.

Elsewhere in the polar regions the cyclonic rotation of the gyres forces waters upwards, preventing the formation of a stable pycnocline and permitting convection from the ocean surface to the ocean floor. This is the outcrop area of the ocean's deep water which is about 85% of the ocean's volume.

With this overall structure the ocean will always be well-supplied with oxygen, and an oxygen minimum can form only within and just below the pycnocline flooring the tropical-subtropical anticyclonic gyres. Because of the effect of the Earth's rotation on the dynamics of the anticyclonic gyres, the pycnocline is shallowest on the eastern margin of the ocean basin, rising to less than 50 m depth. Organic matter produced in the photic zone there sinks to decompose and form the nutrient-rich waters of the oxygen minimum. It can then be upwelled back to the surface by winds blowing Equatorward along the coasts. This positive feedback ensures that the highest productivity will occur along the eastern ocean margin overlying the most intense oxygen minimum and cannot spread over large areas of the ocean.

The modern, highly structured ocean and its circulation depend on the stability of the winds. Storms perturb the winds on short time scales, the ocean sees the long term integral. The present stability of the winds throughout the year is a result of the positive albedo feedback of the permanent ice cover of the polar regions, which insures year-round atmospheric highs at the poles. The change in sea surface temperature across the Subtropical and Polar frontal systems provides another positive feedback serving to stabilize the location of the Westerlies.

Until recently, it was assumed that the Cretaceous ocean had a structure and behavior closely resembling that of the modern ocean (Gordon, 1973; Lloyd, 1982; Haq, 1984; Hay, 1995a, 2002). Interpretations of the geologic evidence and model simulations based on this assumption have resulted in differing conclusions regarding major ocean features, such as flow through the Tethys (Seidov, 1986; Barron and Peterson, 1989, 1990; Föllmi and Delamette, 1991; Bush, 1997).

4.2. An alternative structure: the eddy ocean

The idea of an alternative structure for the ocean came as a result of studies of the distribution of calcareous plankton. They suggested that during warm, ice-free times the polar and subtropical oceanic fronts did not exist (Hay, 2000a). Evidence that the cause might be instability of the winds in response to a seasonal alternation of pressure systems at the poles came from Flögel's (2001) paleoclimate simulation for the Turonian. The idea that the transition from one state to another depends on whether the poles are ice covered or not arose from an attempt to explain the global nature of the Eocene-Oligocene climate change (Hay et al., 2005).

The ocean structure would be very different if there were no ice at the poles. It is intuitively obvious that there should be seasonal reversals of atmospheric pressure in the polar regions. A water-covered pole surrounded by land would be relatively warm in winter, generating an atmospheric low pressure system, and cool in summer, producing a high. A land-covered pole surrounded by water should be relatively cold in winter, producing a high, and warm in summer producing a low. There are seasonal reversals of the polar pressure systems in both hemispheres, but because of the paleogeographic peculiarity of one pole being water, the other land, they are synchronized. Both poles have lows in January and highs in July as shown in Fig. 10.

Climate simulations of the Turonian (Flögel, 2001) show the seasonal pressure reversals. The result is a much more complex atmospheric circulation than observed today. The tropical Easterlies remain constant and strong throughout the year, forcing an Equatorial Current system in the ocean. However, at higher latitudes the circulation varies with the seasons. In particular, the Westerly winds become unstable when there is a polar low and may even reverse direction.

Without the steady forcing of the Westerlies there would be no well developed low-latitude anticyclonic or high-latitude cyclonic ocean gyres. The subtropical to polar frontal systems would not exist. Without the frontal systems there would be no ocean-wide well-developed pycnocline and no barrier to poleward ocean heat transport. Instead, the effect of variable winds would be to generate a multitude of mesoscale eddies, some rotating anticyclonically and pumping water downward, and others rotating cyclonically and pumping water upward. Eddies of this sort are common in the ocean today, but they are mostly trapped within the overall circulation of the larger gyre systems. Today's eddies are not restricted to the surface ocean, but occur in the deep sea as well (Dengler et al., 2004).

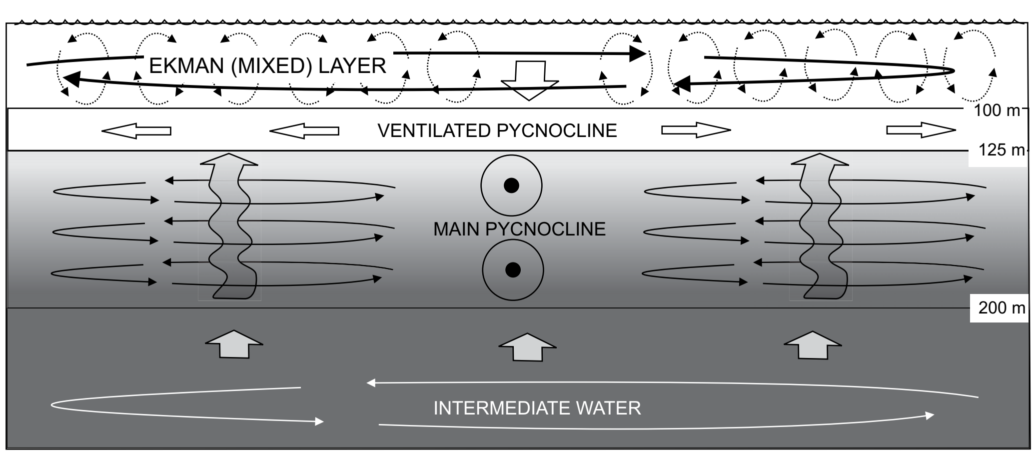

Fig. 9. Schematic cross section through the upper part of the ocean in the mid-latitudes of the northern hemisphere at present. The surface waters are the Ekman mixed layer with its general anticyclonic rotation forming the tropical subtropical gyre. Below are the ventilated pycnocline (about Central Water), the main pycnocline/thermocline, and the Intermediate Water mass. Flow of water in the Main Pycnocline is Equatorward, but there is additional circulation on the isopycnal surfaces. Overall the water is moving slowly upwards (1 to 2 m per year) in response to the production of deep water at high latitudes.

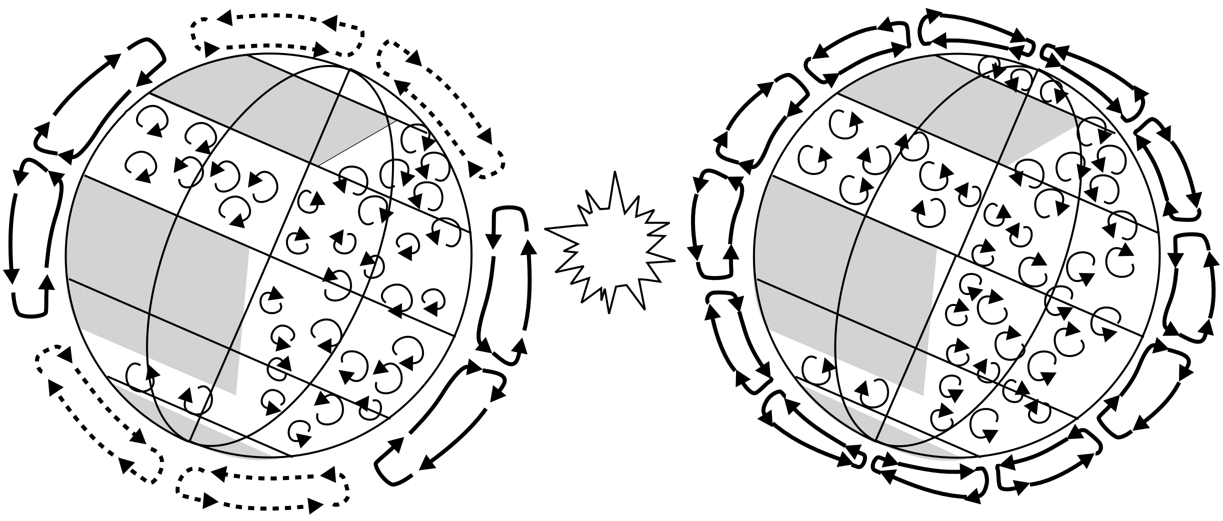

Fig. 10. Seasonal atmospheric and oceanic circulation on an Earth with schematic Cretaceous paleogeography and ice-free poles.

Because the oceanic Ekman layer thickness depends on the wind speed, it is possible to estimate it for the Cretaceous from climate model simulations. Cretaceous winds are generally thought to have been somewhat weaker than today, hence the oceanic Ekman layer thickness would have been less. Following Stewart (2005) average wind speeds of 8 m/sec (about 16 knots), 80% of today's average, would result in an Ekman layer depth of 120 m at 15 ^{o} and 72 m at 45°. The thickness of the pycnocline is determined by the wind-induced downward mixing, which can be estimated from climate modeling, the rate of upwelling of deeper waters, which is much more difficult to model or evaluate, and assumptions about the vertical eddy viscosity, which has values ranging from 10^{-4} to 10^{-1} m^2/s according to different sources (Wang et al., 1996). A clue to the nature and thickness of Cretaceous pycnoclines at times of ocean anoxia lies in the discovery that some organic-rich Cretaceous sediments contain biomarkers for green sulfur bacteria (Sinninghe Damsté and Köster, 1998; Simons and Kenig, 1998; Pancost et al., 2004). Green sulfur bacteria thrive in dysaerobic waters that lie withing the photic zone, and occur in great numbers at 100 m depth in the Black Sea today (Manske et al., 2005). A thin pycnocline implies a high rate of intermediate and/or deep water formation.

Fig. 11 shows what the circulation may have looked like during much of the Cretaceous. The only steady ocean currents would have been those of the equatorial systems, forced by the constant Easterly winds. At mid and high latitudes the ocean would have been filled with eddies taking trajectories that would have been much more chaotic than what is observed today. The larger, more deeply penetrating eddies would be steered by the bathymetry as are those in the cyclonic deep-convecting Circum-Antarctic Current today.

With globally warmer temperatures, the density differences in the ocean would be established mainly by the salinity fields resulting from evaporation and precipitation patterns. The temperature - salinity - density relationships are shown in Fig. 12. Today the density differences in surface ocean waters range from about 1027 kg/m^3 in the polar regions to about 1022 kg/m^3 near the equator. The decrease in density is generally evenly spread across latitude with the exception of a sharper gradient between the Subtropical and Polar fronts. The pattern of density surfaces in the Cretaceous would be very different. Using information from Cretaceous climate simulations (DeConto, 1996; Brady et al., 1998; Hay and DeConto, 1999; Flögel, 2001; Hay, 2002; Otto-Bliesner et al., 2002), we can speculate that as a result of the increased global warmth and enhanced hydrological cycle, equatorial densities in the Pacific would be lower (about 1018 kg/m^3) than beneath the Subtropical Highs (about 1024 kg/m^3). However, from the subtropics to the polar regions the declining temperature could offset the effect of declining density to maintain densities close to 1024 kg/m^3. This means that about half of the entire ocean surface would have almost the same density, and that this would approximate the density of the deep water. Throughout this region storm and wind-generated eddies could pump water up and down. The eddy-ocean would lack the highly-developed structure we see in today's stratified ocean, and the vertical circulation would be much more local and chaotic.

Some idea of the ability of eddies to pump water upward or downward can be gained from studies of those associated with the Gulf Stream. Worthington and Wright (1970) presented temperature and salinity sections across the North Atlantic in which eddies are clearly displayed. Converting the temperatures and salinities to density shows that at a distance of less than 100 nautical miles the North Atlantic eddies can juxtapose waters having a differential density of 0.4 kg/m^3 at a depth of 500 m. Since the total density contrast in waters in the Atlantic today is only of the order of 5 kg/m^3, this is a significant perturbation. However, in the Cretaceous almost the entire density gradient would have been between the equator and the subtropical high at 30° N and S. With much of the ocean at higher latitudes having densities around 1024 kg/m^3, surface and deep waters would have very little density contrast (0.5 kg/m^3). Eddies generating a difference of 0.4 kg/m^3 could allow the perturbation to extend from the surface to the ocean floor in many areas, producing benthic storms (Hollister and McCave, 1984). More energetic eddies could cause convection throughout the entire water column.

One aspect of Cretaceous ocean circulation that has not yet been explored is the effect of the higher evaporation and precipitation rates on the topography of the ocean surface. At present the evaporation over the surface of the North Atlantic beneath the descending limb of the Hadley Cell at 25° N is about 1.5 m/yr. The rotation of the gyre forces the water at that latitude upward, and the actual topographic relief is about +1 m. Without the evaporation the elevation in the gyre center would be 2.5 m. This would force the gyre to rotate more rapidly. At the higher temperatures that prevailed in the Cretaceous, the effect of evaporation would be

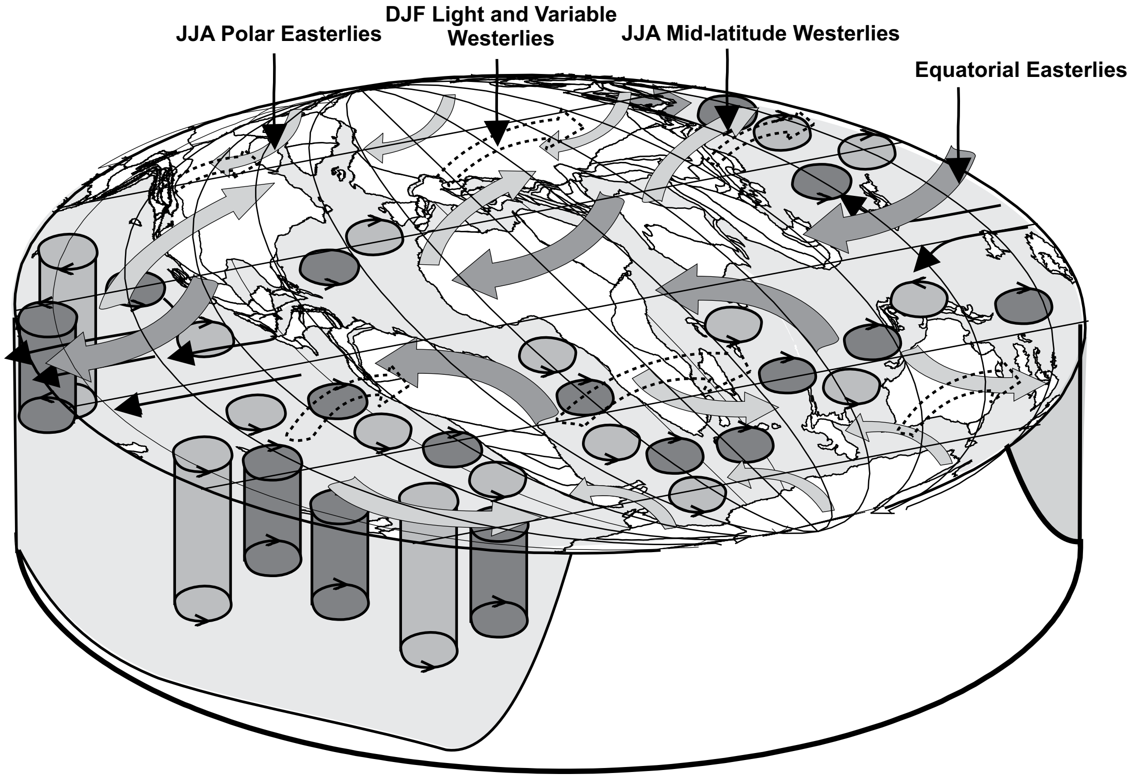

Fig. 11. Schematic view of the structure of an "Eddy Ocean" such as may have existed during much of the Cretaceous and Paleogene. Easterly equatorial winds, shown in dark grey are relatively constant throughout the year; Away from the equator, January winds are light and variable, shown as transparent dotted arrows. They are shown as weak westerlies, but may cease or even reverse. July winds are more like those of today's world, shown in light grey. Away from the Equatorial current systems the ocean is filled with mesoscale eddies and lacks the large gyres characteristic of the modern ocean.

greater; with tropical temperatures of 34 °C, the annual loss of water would be about 2.1 m. Even if there were steady westerly winds to force the gyre, they would be weaker, and the topography resulting from evaporation and wind-forcing would begin to approach 0. Rotation would cease. If tropical temperatures were 42 °C the evaporation would be more than 3 m/yr, creating a depression in the sea surface. Even if there were westerly winds the rotation of the gyre would be cyclonic. Ocean models usually

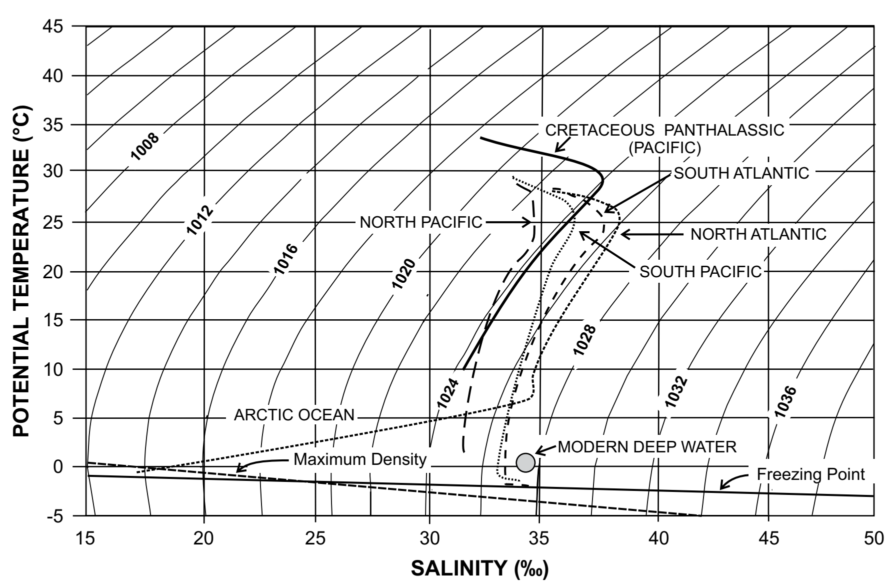

Fig. 12. Relationship between temperature, salinity, and density at the ocean surface (Millero and Poisson, 1981). Density is in kg/m^3. Dotted and dashed lines represent the surface waters of the modern North and South Atlantic and Pacific. Solid line represents the surface waters in the center of the Cretaceous Panthalassic (Pacific) Ocean. The warmest temperatures are in the Equatorial region, the coolest in the polar regions. The highest salinities are beneath the descending limb of the Hadley cells, about 20–30° latitude. The Cretaceous line (after Hay and DeConto, 1999) assumes Equatorial temperatures of about 34 °C. Note that in the Cretaceous there is a very large area poleward of the high salinity region where the density of the surface waters is almost constant. If Cretaceous tropical temperatures were 42 °C there would be a much enhanced hydrologic cycle and much greater meridional salinity differences.

include a "rigid lid," i.e., for computational purposes the topography of the surface is not included, so these effects are not apparent in current simulations

The Cretaceous was a time of significant changes in ocean salinity (Hay et al., 2001, 2006). The salt extractions that began in the Triassic continued through the Early Cretaceous as salt was deposited in the South Atlantic during the Aptian. Average ocean salinities of the Late Cretaceous were about the same as today (34.7‰), but before the Aptian the earlier Cretaceous salinities were probably about 3‰ higher (Hay et al., 2006). As can be seen in Fig. 12, with higher salinities the ocean begins to behave differently in response to temperature changes. In particular there is a greater change in density as the water approaches the freezing point making sea-ice formation more difficult. The higher salinities of the earlier Early Cretaceous, prior to the Aptian South Atlantic salt extraction, may have had a perceptible effect on ocean circulation.

How would oxygen reach the deep sea in an eddy-filled ocean? As can be seen in Fig. 13, the warmer surface waters would contain only a fraction of the oxygen of modern polar waters. More importantly, there would be no mechanism for constant replenishment of oxygen in the deep sea. Only occasionally, when strong anticyclonic eddies would force water to the bottom or cyclonic eddies would open a passage could oxygen be introduced to the deep interior. Most of the time the eddies would promote mixing in the upper half of the ocean, with ephemeral up- and downwelling sites, but overall the mixing rate would be much less than today. Because of the lack of sharp stratification, the oxygen minimum would be more diffuse and could extend to the ocean floor.

Where and how much biological productivity would exist in an eddy-filled ocean? As is the case today, nutrients would have been supplied to the surface ocean from two sources: upwelling of deep water and introduction by rivers and groundwater entering along the ocean margins. Because of the lack of stratification, nutrients would be more evenly distributed throughout the deep sea, with the result that upwelled waters would have lower concentrations than today. The dynamics of the atmospheric circulation would force equatorial upwelling as today, but upwelling in the tropics and subtropics would not be restricted to the eastern margins of the ocean. Instead, it would be most effective along E-W margins beneath the Easterlies at low latitudes, as described by Villamil et al. (1999). Today's large-scale high-latitude upwelling in response to the deep convection in cyclonic polar gyres would be greatly reduced. Instead, upwelling could occur throughout the open ocean at all latitudes away from the equator, whenever strong cyclonic eddies would force waters upward from the ocean interior. Areas of persistent high productivity could develop where the bathymetry might stabilize eddies. Stabilization occurs today in the western Arabian Sea off eastern Africa, where the Southwest Monsoon spins up the anticyclonic "Great Whirl" off Somalia and the Socotra eddy just northeast of it in much the same place year after year (Schott, 1983; Schott et al., 1990; Schott and McCreary, 2001; Wirth et al., 2002).

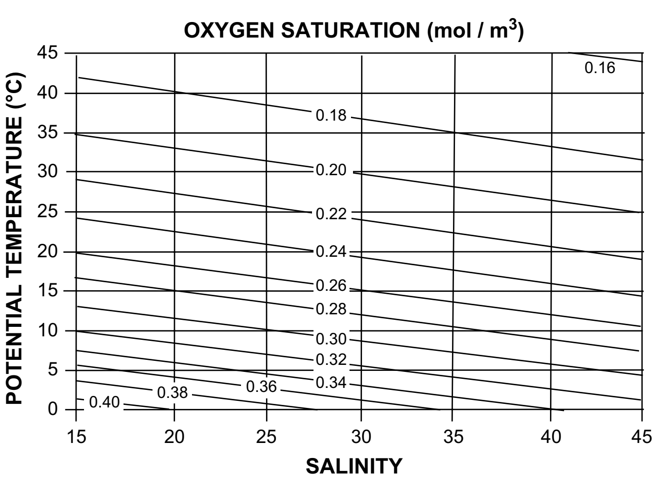

Fig. 13. Oxygen saturation as a function of temperature and salinity. Modern polar waters have temperatures near 0 °C and salinities about 34.7. Cretaceous waters could have temperatures as high as 15–20 °C in the polar regions and 42 °C in th tropics, and global salinities in the range of 33 to 38‰ (Hay et al., 2006). Because of its large volume, the salinity of the ocean's interior and deep water would be close to the average global salinity.

The overall balance between biological productivity and oxygen demand would be much closer than today, and local and basin-wide anoxia more likely. Truly global events, such as that at the Cenomanian-Turonian boundary would require that a strong persistent deep water source force turnover of most if not all of the ocean. This would be an event converting the hypothetical eddy-ocean into a stratified ocean more like that of today. In Italy, the Cenomanian-Turonian boundary "Bonarelli Event" corresponds to the change from a less oxic ocean in which the Scaglia bianca was deposited to the more highly oxic conditions of accumulation of the Scaglia rossa and may represent such an event.

4.3. Deep water formation today and in the Cretaceous

Where would the sites of deep water formation be located? Although usually cited as 20 to 40 Sv, the rates of deep water formation for the modern world are not well known, and may change on century to millennial time scales (Broecker et al., 1999). Stommel (1962) observed that the sites of formation of deep waters that drive the thermohaline circulation of the ocean are very small and largely represent geographic accidents. Dense water formation requires a restricted area in which seawater can be modified to increase its density either by increasing its salinity through evaporation or sea-ice formation, or by cooling. There must, in addition, be a path from this restricted area to the deep sea. Killworth (1983) noted that there are two regions of deep water formation, in shallower waters along continental margins and in the open ocean. Each region is responsible for about half of the total deep water production. Paradoxically, cyclonic circulation, which initially pumps water upward, seem to be a prerequisite for the formation of bottom water. The upwelling destroys the density stratification in the ocean interior creating an open passage to the deep ocean interior. Then the denser water generated by evaporation and cooling at the sea surface can fall down through the passage to the ocean floor. However, to become ocean deep water it is essential that there be an open path between the restricted site of dense water formation and the ocean basin proper. Hay (1993) reviewed modern deep water formation and speculated on the origin of the present system. Today, the densest deep water originates from sea-ice formation over the shallow Barents shelf, but is trapped in the Arctic Basin. About half of today's oceanic deep water originates over shelves in the Greenland-Iceland-Norwegian Sea and in the Weddell Sea. The open-ocean sites of deep water formation are the Labrador Sea, and in the region of Maude Rise off the Antarctic.

Virtually nothing is known for certain about deep water formation during the Cretaceous. Current ocean models simulating the Cretaceous do not include the details of shallow seas, and only the open-ocean component of deep water formation is simulated; deep convection introduces dense ocean surface water into the interior. Even though models include only the open-ocean component, they commonly overestimate total modern deep water production by a factor of two. Modern analogies for deep Cretaceous marginal seas are the Arctic and Mediterranean. The densest deep water in the ocean is in the Arctic Basins where cold saline waters reach densities more than 1028.1 kg/m^3. These deep waters can penetrate into the Norwegian-Greenland Sea where bottom waters can have a density of 1027.9 kg/m^3. They are prevented from entering the world ocean proper by the Greenland-Iceland-Faeroes Ridge. As discussed in Hay (1993) the waters that overflow the Ridge into the North Atlantic are not deep water, but lighter intermediate water masses with densities closer to 1027.5. Similarly, by virtue of their salinity, the deep waters of the Mediterranean are denser than those of the North Atlantic. The waters flowing out through the Strait of Gibraltar are initially denser than any water in the North Atlantic, but they become lighter through turbulent mixing and finally reach equilibrium with other North Atlantic waters at a depth of about 1,500 m. Paleontologic evidence suggests that the extensive "boreal" seas of Europe had salinities well above that of the Tethys (Voigt et al., 1999) but they have not yet been considered possible sources of ocean deep waters.

Simulations using different ocean models for the same Cretaceous time interval give different results. In a model with Campanian paleogeography, Brady et al. (1998) found deep convection in the sub-Antarctic, although the waters sinking there had been salinized by evaporation in the lower-latitude South Atlantic, in the eastern Tethys, and in the western North Pacific. Another model with the same paleogeography (Otto-Bliesner et al., 2002), showed deep convection only in the circum-Antarctic and North Pacific. Other models, with Albian and Turonian paleogeography (Poulsen et al., 2001), showed deep convection in the circum-Antarctic, northwestern Pacific, and along most of the northern Tethyan margin in the Albian simulations but only in a few small areas of the northern Tethyan margin in the warmer Turonian simulation. In reality, all of these areas were characterized by complex bathymetry, with subsea ridges and plateaus separating deeper basins. The configuration of the northwestern Pacific changed continuously as terranes from the western Pacific collided with the Asian margin (see paleogeographic/paleolithofacies maps of Kazmin and Napatov, 1998). The complex paleogeography of shallow Cretaceous seas and margins is also evident on paleogeographic maps for the Tethys and Peri-Tethys (Dercourt et al., 1990, 1993, 2000). It is unlikely that there were many unobstructed passages from these potential sites of dense water formation into the Panthalassa (Pacific) basin. The paleobathymetry of the Atlantic and Tethys would make it likely that each basin might have its own source for deep water, its own relatively oxygen-depleted waters, and its own nutrient supply and recycling system. Waters sinking along the Antarctic margin and in the northwest Pacific would be more oxygen-rich, but again their spread might be retarded by complex paleobathymetry. It is likely that there would be different deep water masses in the individual basins, and it is likely that there would be many sources for ocean deep waters rather than few, as is the case today. In the real Cretaceous world, the sites of dense water formation were most likely to have been on shelves or in epicontinental seas not bounded by shelf-break fronts; their areas were more than double those of today. Quantitative estimates of the rates of intermediate and deep water formation would make an important contribution to our understanding of Cretaceous pycnoclines and processes of ocean mixing.

5. Summary and conclusions

The Cretaceous is a special episode in the history of the Earth. Its uniqueness was early recognized by giving the name of a unique rock type, chalk, to this interval of geologic time. Chalk turned out to be similar to modern deep-sea calcareous ooze, but the question remains; why was it deposited on the continental blocks only during the Cretaceous. In part, this may have been simply due to the evolution of the calcareous plankton. Planktonic foraminifera have their origins in the Jurassic and the calcareous nannoplankton in the Triassic but they did not colonize the open ocean until the Cretaceous. The deposition of chalk in epicontinental seas occurred as these areas became an integral part of the ocean.

Today shelf-break fronts separate inshore from open-ocean waters. These fronts are characteristic of most passive continental margins in the low and mid-latitudes. They are quasi-permanent features fixed to the shelf/slope break, and in spite of their widespread occurrence, they remain one of the most enigmatic features of the ocean and are still not well understood. The shelf/slope break itself is closely associated with the transition from normal to thinned continental crust on passive margins. It is remarkable that the fundamental geologic boundary marking the edge of the continental block should have a counterpart as a frontal system in the overlying water.

During the Late Cretaceous calcareous oceanic plankton invaded the epicontinental seas, and it is argued here that this was in response to sea level rise. As the water depth over the shelves and passages into epeiric seas increased, they fell below the base of the wind-mixed Ekman layer, and shelf-break fronts could not be maintained.

A second peculiarity of the Cretaceous is its warm equable climate. Meridional temperature gradients were much lower than today, and the continental interiors warmer in winter. There is an emerging consensus that tropical temperatures were significantly warmer than today and that polar temperatures were well above freezing during most of the Late Cretaceous. The extreme view is that tropical temperatures may have been as high as 42 °C and polar temperatures may have approached 20 °C. It is thought that the warmer climate was the result of a greater atmospheric content of primary greenhouse gasses, CO_2 and possibly CH_4, with their effect strongly reinforced by the atmosphere's higher water vapor content in response to the warmer temperatures.

Most of the additional energy involved in the meridional heat transport system was transported as latent heat of vaporization of H_20 by the atmosphere, and poleward heat transport may have been as much as 1 Petawatt (20%) greater than it is today.

Vegetation consisting almost exclusively of C_3 plants provided for more efficient energy transport into the interior of the continents, resulting in milder winters and cooler summers. The evolution and spread of C_4 plants near the end of the Cretaceous may be responsible for the Cenozoic spread of desert conditions.

Finally, the Cretaceous has been very important in providing evidence that the circulation of the ocean may have been very different in the past. With its present structure and circulation it is impossible for large areas of the ocean to become anoxic. Yet there were episodes of widespread or global anoxia particularly during the mid-Cretaceous. The present ocean structure depends on constant wind systems, which in turn depend on stability of the atmospheric pressure systems in the polar regions. Today this stability is forced by polar ice.

If there was polar ice in the Cretaceous it was only for short intervals of time. During most of the Cretaceous the polar regions were ice free, The northern polar region was a gulf off the Panthalassa (Pacific) Ocean during the Early Cretaceous, but became isolated as a deep basin connected to the open ocean only through shallow epicontinental seaways during the Late Cretaceous. The southern polar region was occupied by Antarctica, which separated from the other Gondwanan fragments in the Early Cretaceous. Without polar ice there were seasonal reversals of the atmospheric pressure systems. However, because one pole was covered by water and the other by land, both would have lows in January and highs in July. This would result in disruption of the mid and high latitude wind systems.

Without today's constant westerly winds, there would be no subtropical and polar fronts in the ocean, no well-developed ocean pycnocline, and no tropical subtropical gyres dominating ocean circulation. Instead the ocean circulation would be accomplished through mesoscale eddies which could carry warmth to the polar regions.

Today the formation of deep water is divided between the open ocean and shallow shelves. Production of deep water was more likely to have occurred from salinization or chilling of water in the much more extensive epeiric seas of the Cretaceous. Without the global thermohaline circulation characteristic of the modern ocean, it would have been much more likely that individual basins could become dysaerobic or anoxic. Global episodes of anoxia might reflect general turnover of the ocean with widespread upwelling as stable sources of deep water developed.

The Cretaceous is a unique laboratory for understanding how the climate system operates when one or both polar regions are ice free. Further study of the Cretaceous is vital to developing a better understanding of how the Earth's climate system works.

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