

oceanic crust. We therefore consider the phase change instability which might exist between the crystalline peridotite of the upper mantle and partially molten peridotite (the partial melt fraction has the composition of basalt).

If the mantle at, say, a depth of 100 km were just above its temperature of beginning of melting (as has been suggested for the seismic low-velocity zone), the introduction of some pressure perturbation (conceivably arching of the lithosphere through tectonic activity, or some kind of surface rifting) might be sufficient to initiate a self-sustaining upward flow beneath the locus of the perturbation. If fragmentation of the lithosphere ensued, sea-floor spreading could take place with ridge formation, eruption of magmas, and generation of new crust. If the ridge remained stationary or nearly so, a relatively deep flow might ultimately become established. The boundary conditions for the motion of the new plates generated at the ridge might, however, be such that the ridge itself was constrained to migrate; in that the ridge is the locus of rifting, this amounts to the migration of a pressure perturbation. In turn, this should cause the phase change instability to propagate laterally at the level of the low-velocity zone; the instability should be able to sustain the horizontal surface flows as it migrated.

This suggestion is advanced somewhat tentatively, since the theory of the instability is only approximately valid for this phase change (it is at least divariant). It would, however, imply that slowly moving or static ridges, if they could be recognized, should be characterized by deep flows, and rapidly migrating ones by shallow flows. The perturbation of mantle isotherms associated with these two kinds of flow might ultimately be distinguishable by seismic methods. It is also possible that the two kinds of flow might be characterized by slightly different compositions of magmas extruded at the surface.

Finally, we test this melting instability quantitatively. We assume one phase to contain no liquid and the other phase to contain a partial melt fraction of 5 percent. We take the partial melt fraction to be basalt. The denser (no partial melt) phase lies beneath the lighter phase, and heat is absorbed in going from the denser to the lighter phase. For the 5 percent partial melt fraction, we take $Q = 5.4 \text{ cal g}^{-1}$, $c_p = 0.3 \text{ cal g}^{-1} \text{ } ^\circ\text{K}^{-1}$, $d = 100 \text{ km}$, $\gamma = 0.118$

$\text{kb } ^\circ\text{K}^{-1}$, $\rho = 3 \text{ g cm}^{-3}$, $g = 10^3 \text{ cm sec}^{-2}$, $T = 1500^\circ\text{K}$, and $\kappa = 10^{-2} \text{ cm}^2 \text{ sec}^{-1}$. We can determine the maximum value of the kinematic viscosity which will lead to an instability as a function of β . The result is given in Fig. 2. The maximum value of the kinematic viscosity for instability is of the order of $\nu = 10^{21} \text{ cm}^2 \text{ sec}^{-1}$. This value is not unreasonable, particularly since the presence of the partial melt is likely to reduce the viscosity significantly.

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Sea-Floor Spreading, Carbonate Dissolution Level, and the Nature of Horizon A

Abstract. Evidence from leg 2 of the Deep Sea Drilling Project suggests a constant spreading rate for the floor of the North Atlantic over the past 80 million years; a major lowering of the carbonate dissolution level during the early Pliocene; and an early to middle Eocene age for horizon A.

During leg 2 of the Deep Sea Drilling Project (DSDP), paleontological dates were obtained from sediments immediately above acoustical basement. These dates indicate that the North Atlantic Ocean has been opening at a constant rate of 1.1 cm per year during the past 80 million years, probably without any major episodes of quiescence or rapid spreading. In addition, a significant lowering in the calcium carbonate dissolution level is indicated during early Pliocene time. Pelagic sediments deposited near the compensation depth consist of calcareous ooze from early

Pliocene time on, but older sediments are barren of calcareous constituents. This may result from an increase in the amount of calcium carbonate in the ocean, caused indirectly by the onset of glaciation on the Antarctic continent. The prominent reflector, horizon A, appears to be present over parts of the eastern North Atlantic, as well as in large areas of the western North Atlantic. It formed during early and middle Eocene time, probably as a result of widespread silica deposition in the North Atlantic.

Leg 2 of the DSDP covered the 2-month period of October and November 1968; it involved crossing the North Atlantic from North America to Africa (1). Five sites (Nos. 8 to 12) were drilled by the *Glomar-Challenger* during this leg on a line from New York to Dakar (Fig. 1). Sites 8 through 11 are in the western half of the North Atlantic, roughly evenly spaced between the North American continent and the Mid-Atlantic Ridge; site 12 is in the eastern North Atlantic, northwest of the Cape Verde Islands.

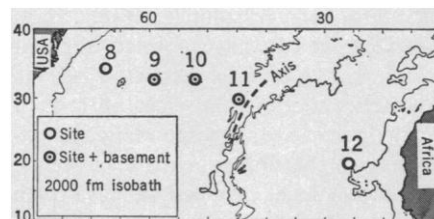


Fig. 1. Approximate location of sites drilled during leg 2 of the Deep Sea Drilling Project. [Courtesy of DSDP]

At sites 8 and 12, the westernmost and easternmost sites, the oldest sediment penetrated is of early Eocene age. This sediment is, however, not the oldest at these locations. At site 8 an additional 370 m of sediment occur below the bottom of the hole and above acoustical basement (1). If we assume that this sediment accumulated at the same rate as the 80 to 130 m of sediment immediately above it (about 0.29 cm per 1000 years), then the age of the sediment-basement contact should be approximately 185 million years, or close to the age of the Triassic-Jurassic boundary. The above extrapolation is of course only an approximation, as sediment accumulation rates at a particular location may vary through geologic time, especially at sites near the carbonate dissolution level. Also, stratigraphic gaps are common in the sedimentary record of the deep sea.

Site 8 is just southeast of the quiet magnetic boundary of the western North Atlantic. Heirtzler and Hayes (2), assuming a spreading rate of 1 cm per year for the North Atlantic, estimate an age of 200 to 250 million years for this boundary, the basement being older westward and younger eastward. If the age of the boundary is recalculated with a spreading rate of 1.1 cm per year, a figure probably more accurate (see below), the above age estimate is reduced by about 10 percent. Thus, an age of 185 million years for the basement at site 8 seems quite reasonable.

Applying a similar extrapolation to site 12, the age of the sediment-basement contact is estimated at about 153 million years, or approximately middle Jurassic. This estimate is, of course, subject to the same limitations as that for site 8.

In the western North Atlantic the ages of the sediment-basement contact at sites 8, 9, 10, and 11 support current thinking on the origin of the North Atlantic. The basement becomes, in fact, progressively younger as the Mid-Atlantic Ridge is approached: Triassic-Jurassic boundary at site 8 (extrapolated); Santonian to Campanian at site 9; early Campanian at site 10; and early middle Miocene at site 11. At two sites, Nos. 10 and 11, biostratigraphic control is excellent, and the sediments immediately above basement are dated as basal Campanian (approximately 80 million years) and early middle Miocene (*Orbulina* datum, or about 14 million years) (3). If we assume the center of the Mid-Atlantic Ridge to

have zero age, three points are available for calculating sea-floor spreading rates. The distance between sites 10 and 11 is about 730 km, and the age difference of the sediment-basement contact is 66 million years, yielding a spreading rate of 1.1 cm per year; from site 11 to the Mid-Atlantic Ridge the distance is about 155 km, and the age difference of the sediment-basement contact is 14 million years, again yielding a spreading rate of 1.1 cm per year. This exact agreement strongly suggests that the rate of sea-floor spreading did not vary greatly in the North Atlantic during the past 80 million years. This conclusion is in contrast to the view, advanced by Ewing and Ewing (4), that sea-floor spreading is episodic, with the present episode of spreading beginning about 10 million years ago and preceded by a long period of quiescence extending back to early Tertiary, and possibly to late Mesozoic time.

Sites 8, 9, and 12 are at depths very close to the modern carbonate dissolution level. At these localities, calcareous pelagic ooze is present on the surface of the ocean floor but soon gives way to barren clayey sediments below. The age of the transition from calcareous ooze to barren clay at site 8 is not known, the oldest calcareous sediments recovered at this site being Pleistocene in age. At sites 9 and 12 the transition occurs at similar stratigraphic levels—that is, near the base of the Pliocene. At site 9 the calcareous interval extends down to about 210 m and the transition appears to be gradual. At site 12 the calcareous sediments pass more abruptly into barren zeolitic clay between 30 and 40 m, a level marked by several prominent ash layers. As there is no evidence of turbidites at sites 9 and 12, it appears that a significant lowering of the carbonate dissolution level occurred in early Pliocene time in the North Atlantic.

It is difficult to explain a lowering of the carbonate dissolution level in the ocean at a time when the temperature was generally dropping, an event that could be expected, instead, to raise the carbonate dissolution level. Hays and Opdyke (5) have pointed out that widespread glaciation occurred on the Antarctic continent at least 4 million years ago, an age corresponding approximately to the time when the carbonate dissolution level dropped in the North Atlantic. Ice formation in Antarctica (and Greenland?) would result in about a 100-m drop in sea level, which in turn would cause large volumes of sedi-

ment to be transported to the ocean basins by rejuvenated streams. Glacial ice movement on Antarctica also would cause vigorous erosion and transport of sediments to the oceans. This sudden influx of large volumes of sediment, much of it carbonate, as well as meltwaters from glacial runoff charged with calcium carbonate, could explain the lowering of the carbonate dissolution level at a time when it might otherwise be expected to rise in response to a lowering of temperature.

An increase in the productivity of the North Atlantic, with an attendant increase in the production of calcareous skeletal material by planktonic organisms, might cause a lowering of the carbonate dissolution level as well, but the likelihood of this possibility is difficult to evaluate.

Horizon A is prominently developed over large parts of the western North Atlantic (6), and several sites on leg 2 reached or penetrated this reflector. Horizon A has an early to middle Eocene age at sites 8 and 9 and consists in both cases of chert beds associated with Radiolaria-rich siliceous ooze. At site 10 the reflector, present at about 170 m below ocean bottom, is a thin chert bed with which is associated a mixed radiolarian-calcareous nannofossil ooze. This ooze can be assigned to the *Discoaster lodoensis* zone of early Eocene age. Altogether, the reflector identified as horizon A in this part of the North Atlantic is Eocene in age or, more precisely, early to middle Eocene.

In the eastern North Atlantic, similar reflectors are also prominently developed. At site 12, northwest of the Cape Verde Islands where several weak reflectors can be recognized, chert was penetrated at the level of the first of these reflectors. Drilling was stopped when more chert was encountered and could not be penetrated. Radiolaria indicate an early to middle Eocene age, which makes a strong case for the presence of a coeval horizon A in the eastern North Atlantic.

On the basis of piston cores taken in the outcrop area east of the Bahama Plateau, Ewing *et al.* and Saito *et al.* (6, 7) proposed that horizon A represents an isochronous surface of late Cretaceous age formed by the extensive deposition of turbidites. According to these authors, the turbidites constructed a buried abyssal plain separating Cretaceous from Cenozoic sediments and marking an abrupt change in depositional conditions.

All the evidence obtained during leg 2 of DSDP points, however, to an early to middle Eocene age for the reflector identified here as horizon A. At site 8 turbidites are as much a part of the horizon A chert as the Radiolaria-rich siliceous ooze. But sites 9 and 10, which are farther from the continent, show little or no evidence for turbidite origin of the horizon A chert. The conclusion seems to be, therefore, that siliceous ooze sedimentation is really the factor determining the nature of horizon A and that turbidite association is only a function of proximity to the continent.

Ewing (8) has suggested that horizon A splits into 2 reflectors on the Bermuda Rise, the upper one being middle Eocene in age. Indeed, a middle Eocene reflector is developed at site 9, but the lower reflector seems to be represented only by some siliceous radiolarian sediments of Upper Cretaceous age. The lower reflector is absent at site 10.

Alternatively, it is possible that horizon A actually was never cored in the outcrop area, and the Upper Cretaceous sediments recovered there come from a stratigraphic level lower than horizon A. Lack of chert or siliceous ooze in any of the piston cores taken in the outcrop area may support this view or may be due entirely to coring difficulties.

The existence of an early to middle Eocene reflector (horizon A), synchronous in a broad sense and con-

sisting of one or more chert beds, seems well established. This reflector appears continuous over very large parts of the western North Atlantic, and possibly over parts of the East Atlantic as well. This reflector shows the least amount of chertification at site 10, a site that seems not to have been below the carbonate dissolution level at any time and that is farthest from continental influences. The most pronounced development of the reflector is at sites 8, 9, and 12, all of which were below the carbonate dissolution level at the time this reflecting layer was deposited and one of which (site 8) also received turbidite sediments.

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cent in Miocene shells. The object of the present work was to find out whether by gas-liquid chromatographic techniques, the degree of racemization during diagenesis of marine sediments can be detected within the first several thousand years.

Saanich Inlet is a stagnant fjord of Puget Sound on the coast of British Columbia where sediments are accumulating (3). A comprehensive geochemical study has been made of this inlet (4). Amino acids were one of a number of classes of compounds investigated. Four samples were selected for the determination of enantiomers of amino acids (Table 1). Core 4 (0 to 0.15) is from the top 15 cm of a 2.5-m surface core. Core 3B (7.9 to 8.2), core 3B (17.1 to 17.4), and core 3B (34.5 to 34.8) are from a deep core; the sample name includes the depth in meters below the sediment-water interface at which the sample was taken (4).

Sediment samples were first extracted with a mixture of benzene and methanol (7:3). The residue was dried and extracted twice with 1N HCl to remove free amino acids; these free amino acids were not examined for ratios of enantiomers. Subsequently, the bound amino acids were isolated by hydrolyzing the sediment in 6N HCl for 22 hours at 110°C. The acid hydrolyzate was separated from the sediment residue by filtration. The HCl was removed from the hydrolyzate at reduced pressure on a rotary evaporator. The residue was dissolved in citrate buffer and chromatographed on a Dowex 50 (H⁺) column which was eluted with H₂O followed by 0.2N NH₄OH. The ammonium hydroxide fractions, containing amino acids, were combined, concentrated, and chromatographed on a Dowex 1 (OH⁻) column which was eluted with H₂O followed by 0.1N HCl. The hydrochloric acid fractions, containing amino acids, were combined, concentrated, and evaporated to dryness in a vacuum desiccator at 40°C. Portions of each sample dissolved in citrate buffer were analyzed on an automated amino acid analyzer (Beckman 120C), and the concentration of each amino acid (Table 1) was calculated from the resulting chromatograms.

Although 20 amino acids were found in the samples only 13 are reported here; these are the amino acids whose presence was easily confirmed by gas-liquid chromatographic techniques. The enantiomeric distribution of seven of these amino acids was estimated. For

Racemization of Amino Acids in Sediments from Saanich Inlet, British Columbia

Abstract. *In sediments spanning the last 9000 years from Saanich Inlet L enantiomers of amino acids are most abundant, but the percentages of D enantiomers increase with age, apparently because of partial racemization. Of the amino acids measured, glutamic acid and alanine show the greatest degree of racemization; leucine, isoleucine, and valine show the least.*

Core samples of sediments ranging from 0 to 34.8 m deep from Saanich Inlet, British Columbia, contain a number of amino acids. With exceptions such as glycine and β -alanine, these amino acids have at least one asymmetric center and, therefore, can exist as enantiomers generally designated D and L. As in living systems, the L enantiomers are most abundant in these sediments, indicating biological precursors. With depth, and consequently with age, the abundance of D enantiomers relative to L enantiomers in-

creases slightly, apparently a reflection of partial racemization during the last 9000 years.

The presence and distribution of amino acids in cores of modern marine sediments have been documented (1). The degree of racemization of amino acids has been followed by enzymatic techniques in fossil *Mercenaria* shells from three ages—Recent, Upper Pleistocene, and Miocene (2). The percentages of D enantiomers increased from 0 in Recent shells to 20 to 40 percent in Upper Pleistocene shells to 50 per-