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Chemostratigraphy and biostratigraphic correlation : exercises in systemic stratigraphy

Chemostratigraphy
Biostratigraphy
Systemic stratigraphy

Chimiostratigraphie
Biostratigraphie
« Stratigraphie systémique »

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ABSTRACT

Global changes in marine geochemistry, on scales between one thousand and one million years permit the detailed correlation of sedimentary sequences in different ocean basins. The condition is that the geochemical signals are at least approximately dated by biostratigraphy (and magnetostratigraphy, where applicable). Through mutual reenforcement of chemostratigraphy and biostratigraphy unusually high stratigraphic resolution can be obtained.

The integration of chemostratigraphy and biostratigraphy opens new avenues for analyzing the record-producing system within the framework of *systemic stratigraphy*. This type of stratigraphy focuses on global change in sea level, climate, and general geologic setting. It attempts to identify the underlying causes of global stratigraphic signals by considering 1) changes in input of matter and energy ; 2) changes in spatial distribution of sediments ; and 3) temporary changes in the partitioning of materials between active geochemical reservoirs. Examples from Pleistocene and Neogene deep-sea records illustrate the concepts and tools of this type of analysis.

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RÉSUMÉ

La chimiostratigraphie et les corrélations biostratigraphiques : essai sur la stratigraphie systémique.

Les variations observées dans la géochimie de l'océan mondial à l'échelle du millier ou million d'années permettent d'établir des corrélations précises entre les séquences sédimentaires des différents bassins océaniques. Pour cela, les signaux géochimiques doivent être datés, au moins approximativement, par la biostratigraphie et si possible par la magnétostratigraphie. L'intercalibration entre stratigraphie géochimique (chimiostratigraphie) et biostratigraphie permet ainsi d'obtenir une résolution stratigraphique extrêmement fine.

Cette intégration des deux méthodes ouvre de nouveaux horizons pour l'étude du système d'enregistrement que représentent les sédiments océaniques et définit une *stratigraphie systémique*. Ce type de stratigraphie est centré sur les variations globales du niveau de la mer, du climat et du cadre structural géologique. Son but est d'identifier les facteurs responsables des changements stratigraphiques globaux en tenant compte 1) des variations dans la nature et la quantité de matériaux et dans l'énergie reçues par l'océan ; 2) des changements dans la répartition des sédiments ; et 3) de l'évolution de la répartition des divers éléments entre réservoirs géochimiques actifs. L'utilisation des concepts et outils mis en œuvre pour ce type d'analyse est illustrée à partir d'exemples pris dans certaines séries sédimentaires océaniques du Pléistocène et du Néogène.

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INTRODUCTION

Correlation based on fossils (biostratigraphy) generally is superior to correlation based on lithologic properties (lithostratigraphy) when the goal is chronologic equivalence of rock sequences. This commonplace concept is deeply anchored within stratigraphic tradition ; it goes back at least to Charles Lyell and is expressed in the epoch names of the Cenozoic from "Eocene" to "Pliocene" and "Pleistocene", names which denote the similarity of fossils to living organisms. Also as a matter of tradition, stratigraphic units belonging to various intervals of geologic time have been associated with "type localities", commonly the places where rock sequences were described in an early phase of geologic mapping. The concept that such a local sequence defines a stratigraphic unit on a more than local scale is pervasive throughout most of the stratigraphic nomenclature, from "Cambrian" to "Flandrian".

Notable exceptions among stratigraphic names are the labels "Carboniferous" and "Cretaceous", which express the observation that global changes in geochemistry accompany the course of geologic time. This basic observation has become increasingly significant for shorter and shorter time scales in recent years, especially in deep-sea sediments. It opens up new possibilities for high-resolution stratigraphic correlation between widely separated realms of deposition.

The stratigraphic record may be seen as the product of a geochemical system consisting of geologic setting, climate, and processes of sediment production. This "exogenic system" responds to tectonic and astronomic input signals, producing both lithologic and paleontologic output signals some of which are correlated on a global scale (e.g. Tappan, 1968). We have recently addressed certain aspects of internal feedback mechanisms within the exogenic system, especially as they relate to step-like transitions from one climatic state to another (Berger *et al.*, 1981, see Fig. 1). Such step-like transitions have especially high potential for global correlation.

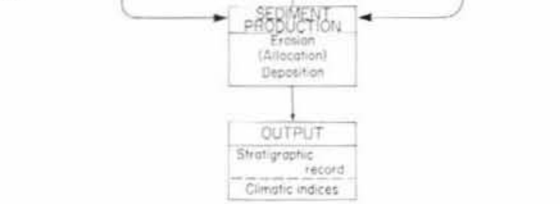


Figure 1

The exogenic system as a machine with internal feedback control with variable input from external sources and with feedback from climatic variations. Output is in terms of sediment patterns, including climatic indices. Instability is introduced by isolated temporary storage bins (arrow) whose contents become available all of a sudden, mainly through endogenic changes in geography (e.g. rapid regression). From Berger et al. *(1981).*

Here we summarize basic propositions about how the exogenic system generates chemostratigraphic signals in the marine realm. We then examine the interaction between chemostratigraphic and biostratigraphic correlation, and implications for high-resolution global stratigraphy. On the assumption that certain chemostratigraphic signals are global, the comparison with biostratigraphy represents a test of synchroneity of first and last appearance datum (FAD and LAD) information from various regions. Conversely, when well-established chronostratigraphic paleontologic markers are at hand, the nature of the chemical signal can be explored.

CHEMOSTRATIGRAPHY VERSUS BIOSTRATIGRAPHY : BASIC PROPOSITIONS

The reason that the subdivision of geologic history is based on biostratigraphy rather than lithostratigraphy is simply this : evolutionary change is more nearly synchronous on a global scale than is lithofacies change, and evolutionary changes are not repeated, while lithologic changes commonly are. However, just as any given paleontological record has both an evolutionary (global) component and an ecological (regional or local) one, thus any given lithologic record may be presumed to contain global signals in addition to regional ones. This is expected because lithology depends on sea level and climate, both of which can change on a global scale. The task is to extract the global signal from the lithologic record, a record which is commonly swamped with regional and local information (proximity of mountains, size of depositional basin, circulation within the basin). Chemical signals within the lithologic record of marine sequences, we think, are prime candidates for discovering global effects, because they depend to some extent on the chemical state of the world ocean, and not only on that of the basin of deposition.

There is one type of chemical record which has a minimum of regional interference, namely that of the deep sea. It is here, therefore, that global geochemical signals are most easily detected. The fundamental papers on the subject are relatively recent : Arrhenius (1952) on the carbonate cycles in the equatorial Pacific (Fig. 2) and Emiliani (1955) on the

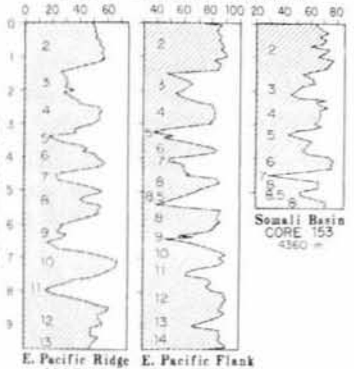


Figure 2

Carbonate dissolution cycles in the Indo-Pacific, Swedish Deep-Sea Expedition. Cores 45 (7°40'N, 106°21'W) and 59 (3°05'N, 133°06'W) : stratigraphic record and numbering system from Arrhenius (1952). Core 153 (2°18'S, 55°33'E) : record from Olausson (1960) ; numbers added by us. Note the striking similarities in these records, but also the difficulty in correlating older cycles without additional information.

oxygen isotope cycles in the Atlantic (Fig. 3). These papers, and the approach to stratigraphy which they introduced, now provide the framework for Pleistocene history. It is to these chemostratigraphic (i.e. lithostratigraphic) sequences that the traditional subdivisions (Günz, Mindel, Riss, Würm ; Nebraskan, Kansan, Illinoian, Wisconsin) must be correlated to have more than regional meaning.

The shift of frame of reference from regional to global signals occurred despite the mixed success of the chemostratigraphers in attempting to correlate their sequences with the traditional schemes, and despite the initial reluctance of the land-based stratigraphers to recognize the superiority of the chemostratigraphic signal over traditional schemes of subdivision.

The chemostratigraphic signals of Arrhenius and Emiliani resemble more or less irregular wave records, and show autocorrelation if shifted along the time axis for a certain amount. Short sequences within such records, therefore, cannot be uniquely identified, and chronologic placement depends on appropriate dating methods. Obviously, biostratigraphy plays a crucial role in such placement.

Chemostratigraphic signals such as carbonate cycles and isotope cycles appear closely linked to sea level changes, at least in the Neogene. This link provides the main bridge to the land record. On land the stratigraphy of sedimentary rocks has evolved as a record of transgressions and regressions, with control on equivalence of strata provided mainly by fossils. To the extent that the regression/transgression histories have a large global component (Grabau, 1940 ; Fairbridge, 1961 ; Hallam, 1977 ; Vail *et al.*, 1977) they produce changes in the chemistry of the ocean which can be detected in deep-sea carbonates, as fluctuations of isotopic composition, carbonate content, and preservation state of fossils (Berger, Winterer, 1974 ; van Andel, 1975 ; Berger, 1977 ; Fischer, Arthur, 1977 ; Hay, Southam, 1977 ; van Andel *et al.*, 1977 ; Berger, 1979 ; Thierstein, 1979 ; van Andel, 1979 ; Scholle, Arthur, 1980). In addition, sea level changes bear on sedimentation rate changes (see Davies *et al.*, 1977) and on the distribution of hiatuses in the pelagic record (see Moore, Heath, 1977).

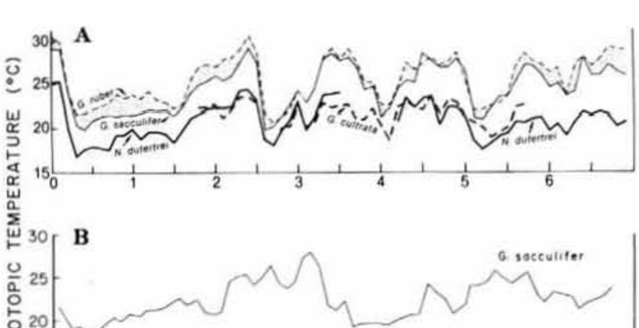


Figure 3

Oxygen isotope stratigraphies of a) Core A179-4, Caribbean Sea and b) Core 234, equatorial Atlantic. From Emiliani (1955). a) the stratification of upper waters is reflected in the differences in composition of planktonic foraminifera living at various depths. Note that the O-18 fluctuations of deep-living foraminifera are subdued ; b) note similarity of this record to Caribbean one in upper one half of the cores. Also note high rate of change during deglaciation, following maximum "cooling" ("isotopic temperature" is a measure of $\delta^{18}O$ content, and hence of ice mass and sea level variation as well).

Some chemostratigraphic signals may not be related to global sea level change, and consequently may not be strictly contemporaneous in all basins. Biostratigraphy can put limits on the chronology of chemostratigraphic events, especially in conjunction with magnetostratigraphy. Once a chemostratigraphic signal is shown to be global it becomes a superior time-marker to which biostratigraphy must defer. It is then possible to extract regional ecologic effects on the diachronism of biostratigraphic horizons (Fig. 4).

Our basic proposition, then, is that chemostratigraphy and biostratigraphy are equal partners in the reconstruction of global geologic history. They are fused in what may be called "systemic stratigraphy", a concept which in principle goes back to geologic antiquity.

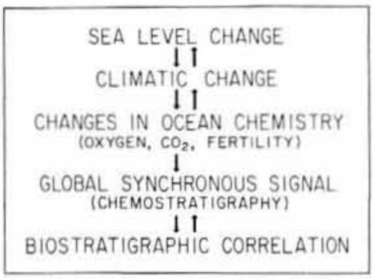


Figure 4

Dominant position of sea level in producing systemic changes which are useful in global correlation. Feedback from climate to sea level occurs during ice ages.

SYSTEMIC STRATIGRAPHY

We have made certain assertions regarding the relative efficacy of chemostratigraphy and biostratigraphy in structuring the geologic record ; in particular we have claimed that there are global stratigraphic signals which are produced by the "exogenic system". These signals — whose nature we shall examine presently — are of the utmost significance for chronologic correlation. They contain both chemical and biological information. They are best studied in terms of a systems approach, that is, in terms of signal input, modulation within the system, and signal output. We propose the term "systemic stratigraphy" to characterize this approach. The principal elements of this type of stratigraphy are summarized in Figure 5.

The various elements listed are basic to all of stratigraphy in one form or another. They are, however, at times buried below the masses of detailed inventory necessary to establish correlations among different stratigraphic sequences, including more or less complete "type sections".

Unlike "lithostratigraphy", "biostratigraphy", "magnetostratigraphy", and "chemostratigraphy", *systemic stratigraphy* does not prescribe the properties and contents of the record which should be catalogued. Instead, it provides a focus on global change in sea level, climate, and general geologic setting. Operationally, it consists of the extraction of global signals from a given stratigraphic sequence, and the interpretation of these signals in terms of regional modulation.

Two sets of tools are necessary for doing systemic stratigraphy. One is an understanding of the mechanisms by which systemic signals are produced. This allows one to concentrate on the type of information which may have systemic

SYSTEMIC STRATIGRAPHY

A. DEFINITION:

Determination and correlation of global climate-related trends, cycles, and events of the "exogenic system" (ocean/atmosphere/reactive lithosphere/biosphere) as recorded in lithologic and paleontologic sequences

B. BASIC RULES:

1) A change in any part of the system produces changes in the others, hence all systemic stratigraphic signals are correlated

2) Every regional signal has a systemic component, which may be amplified or obscured through regional factors

C. BASIC DRIVING FUNCTIONS:

Fluctuations of irradiation and sea level

D. PRIMARY EFFECTS THROUGH:

Changes in temperature and in evaporation-precipitation patterns, ice-water balance, fertility in the ocean and on land, carbonate saturation, CO_2 pressure in air

E. SECONDARY EFFECTS ON:

Biogeography, ecology, evolution, and (chemical) facies distributions

Figure 5

Summary of propositions regarding the main elements of systemic stratigraphy.

value. The other is a criterion of globality, which allows the separation of regional signals and the rejection of local ones (= "noise"). The first set of tools is provided by the general principles of physical, chemical, and biological oceanography. The second is a matter of performing correlations between signals of known and of unknown significance.

In the present context we are interested in "ocean geochemical cycles" and their effects on the geologic record. Hence, the signals of systemic stratigraphy of interest here are mediated through changes in ocean chemistry. We briefly review the mechanisms which lead to such changes. For general correlations between geochemical signals we refer to Fischer and Arthur (1977) and to van Andel *et al.* (1977). We shall return to specific correlations when examining the Pleistocene.

Ocean water contains major ions ("salt") and minor constituents involved in biological cycling ("nutrients"). It also contains gases, notably oxygen and carbon dioxide. Furthermore, a given parcel of water is characterized by a certain temperature. All of these properties have a certain chance of being recorded in sediments, especially in biogenous sediments which are chemical precipitates from sea water.

The composition of calcareous materials in a stratigraphic sequence, for example, reflects something about the saturation of sea water with aragonite and calcite, which is a function of concentration of calcium, of CO_2 content, and of pH, as well as of temperature. Assemblages of phosphate minerals and of biogenous silica in a record bear messages about the fertility of the ocean. Types and abundances of organic matter allow one to reach conclusions about the oxygenation of sea water.

There are three ways to change the stratigraphic signal by changing the chemistry of the ocean : 1) a change in input of matter and energy ; 2) a change in spatial distributions ; 3) a temporary change in the partitioning between ocean reservoirs. We shall address each of these factors in turn.

INPUT CHANGES AND SPATIAL FRACTIONATION

The input of salt, nutrients, carbon, etc. depends on the rate of erosion and the composition of the eroded material, as well as on hydrothermal processes on the seafloor and other volcanic activity.

One would expect, in general, that uplift and regression increase rates of erosion on land, and hence cause an increase in delivery of salt, nutrients, and carbon to the ocean. The reverse — transgression — has been thought to lead to nutrient starvation. This idea, first proposed by Bramlette (1965) to explain plankton extinctions at the end of the Cretaceous, is a classic example of systemic stratigraphy. A change in climate would produce a concomitant change in the input of materials also. If tropical-type weathering is increased, for example, the silica supply will increase, and *vice versa* (Leclaire, 1974 ; Hay, Southam, 1977).

The discovery of hydrothermal activity in crestal regions of the Mid-Ocean Ridge, and the realization that this activity plays a large role in the chemistry of sea water (Edmond *et al.*, 1979), opens up new source and sink functions for the stratigrapher (unfortunately poorly controlled ones). Would a change in spreading rate change the input of calcium, or the availability of phosphorus ? It appears that this is so : about one-third of the calcium budget of the ocean may be delivered from ridge crest activity, while substantial amounts of phosphorus may be extracted by precipitation of metalliferous sediments.

Input changes are frequently invoked when stratigraphers are faced with high abundance signals : for example, accumulations of chert may be ascribed to volcanic activity (Bramlette, 1946 ; Gibson, Towe, 1971). On the other hand, spatial distributions may be just as important. As Calvert (1966) showed, volcanism is not necessary to explain the silica accumulation in the Gulf of California : it is a matter of internal redistribution within the ocean. The Gulf acts as a silica trap because of intense seasonal upwelling.

The redistribution patterns are contingent upon bathymetry (basin and shelf morphologies) and circulation, which is in turn controlled by climate (temperature and precipitation-evaporation). Basically, distributions reflect geographic fertility patterns which translate into production-preservation patterns of biogenous materials (Berger, 1970).

Basins with deep-water inflow collect nutrient-rich, CO_2-rich waters and produce nutrient-rich sediments (C_{org}, Si, P). Basins with deep-water outflow collect nutrient-poor waters and produce carbonate-rich sediments (such as the Mediterranean for example). The sapropel layers in the eastern Mediterranean Pleistocene record, sandwiched between carbonates, must record reversals of water exchange patterns (Kullenberg, 1952 ; Olausson, 1961). To reverse the deep circulation of the Mediterranean, the upper waters must freshen relative to the deep ones. The required freshening has recently been documented by stable isotope analysis of foraminiferal shells taken from the sapropel layers (Williams *et al.*, 1978). The point is, according to the tenets of systemic stratigraphy, the reversal of the Mediterranean cannot have been without consequences for the rest of the world ocean, whose deep circulation depends to some degree on the salty input from the Straits of Gibraltar (Reid, 1979).

Besides basin-basin fractionation (and perhaps even more important), there is basin-shelf fractionation (Berger, Winterer, 1974 ; Hay, Southam, 1977) and also there is latitudinal fractionation on the shelves (Seibold, 1970). The various fractionation processes are ultimately driven by temperature and salinity gradients : when gradients are strong they work well and separate "evaporite-type" facies (including carbonate) from "nutrient-type" facies, but when they are weak, these two types of facies are intercalated or mixed.

The efficiency of the fractionation processes depends on the level of fertility of the ocean : sterile oceans do not precipitate and redissolve biogenous materials. The question of which factor or factors determine the overall productivity of the sea is crucial to systemic stratigraphy. Bramlette (1965) thought it was the nutrient supply to the sea, on a geologic time scale. In a discussion to Bramlette's paper, Newell (1966) stressed the obvious factor of internal cycling. Broecker (1969), Berner (1973) and Piper and Codispoti (1975) emphasized the sink functions : if nutrients are more readily removed from the system they will become less available (have a shorter residence time) in the ocean and hence fertility will drop. For example, if phosphorus is trapped in the metalliferous sediments at the ridge crest as proposed by Berner (1973), an increase in ridge crest activity will decrease the residence time of phosphorus in sea water, so that a given phosphorus atom will be recycled fewer times than before. Or again, a rise in sea level and a concomitant warming and decrease in oxygenation of the ocean would increase the areas on which organic-rich sediments are deposited. Such areas trap phosphorus, and they also are a site for denitrification where biologically available nitrogen is converted to inert N_2. Hence, certain nutrient concentrations would be expected to decrease on the whole. Shelf seas with a deep water inflow, of course, would not necessarily be impacted by a general decrease in ocean fertility : they act as nutrient traps and recycling can be intense here.

How can we recognize "high fertility" and "low fertility" oceans in the geologic record ? Arguments based on diversity of organisms, length of food chain, and other ecologic phenomena (Tappan, 1968 ; Lipps, Mitchell, 1976 ; Fischer, Arthur, 1977), are somewhat involved although they are indeed crucial to the understanding of the paleontologic record and hence to biostratigraphy. A straightforward way to distinguish "low fertility" from "high fertility" oceans is to consider the type of biogenous sediment produced in terms of *distance from thermodynamic equilibrium*.

An equilibrium ocean ("Sillén Ocean") and a biologically active disequilibrium ocean ("Harvey Ocean") produce different deposits in detail although not necessarily in bulk (Berger, Roth, 1975 ; Berger, 1977). The chemical deposits of the Sillén Ocean reflect the high degree of saturation of sea water with biosubstances. For example, carbonates appear as oolites and aragonite is widely preserved. If a dearth of available nitrogen (or of trace elements) is the cause of low fertility, phosphate and silicate concentrations can rise in shallow waters, so that widespread phosphatization and silicification of shelf carbonates become possible. Conversely, carbonates of the Harvey Ocean are of the resistant kind : heavy calcitic shells of foraminifera and molluscs. Phosphate and silicate concentrations are low because of biologic extraction, hence the areas of accumulation of phosphatic and siliceous deposits are restricted to regions of strong upwelling.

TRANSIENT RESERVOIRS

We have briefly (perhaps too briefly) indicated how circulation patterns, driven by climatic gradients, can lead to spatial fractionation of biogenous deposits, implying that changes in these climatic gradients will then produce changes in the stratigraphic sequences of major ocean basins. Such sequences are correlated from one site to another because climate and ocean circulation are global phenomena. We have argued that strong climatic gradients and a high oceanic fertility produce disequilibrium, that is, a "Harvey Ocean", in which fractionation and hence separation of different types of chemical facies are at a maximum.

Disequilibrium also can be produced through temporal storage and reintroduction of materials to which ocean circulation and ocean fertility are sensitive. We have proposed the term "transient reservoirs" to emphasize this concept (Berger *et al.*, 1981), which draws analogies between glacial ice caps, Arctic brackish waters, Mediterranean brines, Aptian salt deposits of the South Atlantic, forests and soil carbon, shelf carbonates, and surficial deep-sea carbonates. The transient reservoirs are part of the "geologic setting" within the record-producing system (Fig. 1).

The best-known transient reservoir is the glacial ice cap on the northern hemisphere. It temporarily stores about 3 % of the ocean's water. Storage and release produces the familiar sea level fluctuations which are the heartbeat of Pleistocene history (the Milankovitch mechanism provides the "pacemaker" : Hays *et al.*, 1976 ; Pisias, 1976). The oxygen isotope record is closely tied to this fluctuation and can be used to monitor it within the deep-sea record (Fig. 6).

The record shows (Fig. 3) that release of water through melting of the ice caps can be very fast, producing on the order of one to ten centimeters of sea level rise per year for considerable periods of time. Such a rate is a substantial fraction (1 to 10 %) of the average annual evapo-precipitation rate pattern, which is closely tied to deep-sea circulation. The point is, the rate of introduction of fresh water from the transient reservoir is large enough to change the density patterns of the ocean and affect deep-sea circulation on a large scale. It is a major source of instability (Olausson, 1965 ; Worthington, 1968 ; Berger *et al.*, 1977).

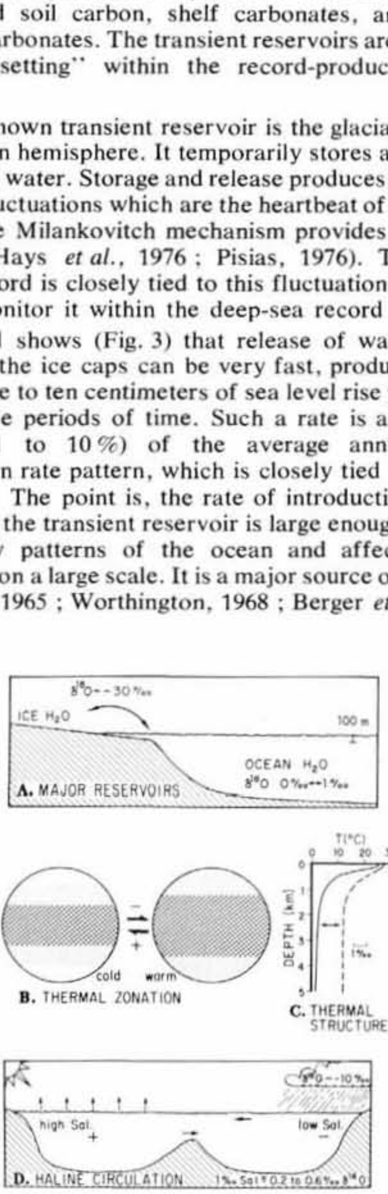


Figure 6

Main factors producing changes in the oxygen isotope stratigraphies of pelagic carbonates (disregarding diagenesis). A : Ice effect in the Quaternary ocean. B : Surface temperature changes. C : Temperature profile changes (cold and warm oceans). D : Evaporation-precipitation patterns.

The rapid merging of previously isolated ocean basins may have effects somewhat analogous to the melting of an ice cap. That large basins were isolated at some time in their evolution is apparent from the presence of salt deposits in their margins. The breakup of Pangaea as conceived by Wegener (1915) generated many opportunities for the development of such transient reservoirs. It is quite obvious that salt deposits, such as in the Mediterranean at the end of the Miocene (Hsü *et al.*, 1973) or in the South Atlantic in the Aptian (Kumar *et al.*, 1977) imply the production and potential availability to the world ocean of heavy brine. Injection of such brine would be a source of global chemostratigraphic change (Thierstein, Berger, 1978), presumably the bottom-covering brine would act as a trap for organic matter and nutrients. In turn this organic matter and the nutrients would constitute transient reservoirs which upon dissipation of a deep salty layer would become available to cause eutrophication and associated algal blooms.

The salt deposits in the spreading basins have other implications as well. They indicate a potential source of rapid sea-level variation — as much as 10 m for the Mediterranean and 50 m for the South Atlantic (Berger, Winterer, 1974). In the Jurassic, the early North Atlantic Basin may have provided a source of instantaneous transgression and regression, due to repeated emptying and filling. If so, the stratigraphic signals associated with such floodings and dry-ups as recorded in Jurassic shelf deposits should be correlatable between all continents.

Less obviously, and paradoxically, salt deposits may also indicate a potential for a transient reservoir of fresh water. This seems to be the case in the Mediterranean, judging from diatom-rich layers of late Miocene age (Hajós, 1973 ; Schrader, Gersonde, 1978). The salt merely shows that a basin was sufficiently isolated from the global ocean to produce brines *at times* : it need not be filled with brine all the time.

Finally, as far as salt deposits, they are themselves a transient reservoir and may influence global carbonate stratigraphy through lock-up and release of calcium in sulfates as well as through the sulfur cycle (cf. Holser, 1977).

We have referred to a potential transient reservoir of carbon in connection with temporary widespread anaerobism. Periods of increased anaerobic deposition of varying lengths have been proposed for certain times in the Cretaceous (Barremian to Turonian ; Ryan, Cita, 1977 ; Thiede, van Andel, 1977) and some have been thought to involve the entire ocean ("Anoxic oceanic events", Schlanger, Jenkyns, 1976 ; Jenkyns, 1980). If anoxic events are real, the associated lock-up and release of organic carbon must lead to changes in pCO_2 in the atmosphere, which in turn produces climatic fluctuations with feedback for the circulation and chemistry of the oceans. We believe that the climatic cycles discovered in non-glaciated periods, and yet apparently of Milankovitch-type dimensions (Dean *et al.*, 1978) may be tied to such carbon-dominated feedback mechanisms. If so, it should be possible to monitor such cycles in the carbon isotope record (Fig. 7).

Of course, marine organic carbon is not the only transient reservoir to be considered in this context. There is the biosphere (forests, swamps, bogs) and soil carbon, and there are shallow water and deep-sea carbonates which are subject to buildup and erosion with redissolution.

We emphasized stable isotopes as monitoring devices for systemic stratigraphy because we favor this particular tool over others in pursuing global chemostratigraphy and hence systemic stratigraphy. The reason is that the global signal, especially in oxygen isotopes, has a good chance of shining through the regional and local overprint in marine sequences : the ocean dominates within the oxygen reservoirs of the exogenic system. Likewise, but to a lesser extent, the ocean dominates the assemblage of active carbon reservoirs. Furthermore, the factors providing for excursions in oxygen and carbon isotopes in the carbonate record are a manageable number : fractionation between water masses (including ice masses during certain periods), temperature, effects of diagenesis for oxygen ; fractionation between carbonate and organic carbon (including both marine and terrestrial carbon), vital effects, and diagenesis for carbon. For detailed discussions of these various factors, see Duplessy (1978), Savin (1977) and Shackleton (1977 *b*) on oxygen' and Kroopnick *et al.* (1977), Shackleton (1977 *a*) and Scholle and Arthur (1980) on carbon.

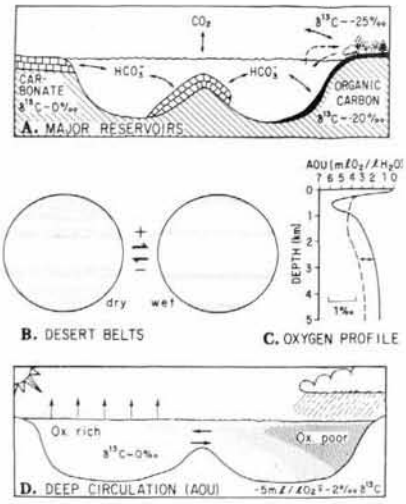


Figure 7

Main factors producing changes in the carbon isotope stratigraphies of pelagic carbonates (disregarding diagenesis). A : Reservoir exchange (organic carbon versus carbonate). B : Climatic change and effect on biosphere and soil carbon abundance. C : Change of oxygen profiles (and associated CO_2 derived from respiration). D : Evaporation-precipitation patterns governing apparent oxygen utilization (AOU) and hence abundance of respirative (^{12}C-rich) CO_2. (oxygen-rich corresponds to CO_2-poor and vice versa). *Note the analogies between the schemes in Figures 6 and 7.*

We shall next illustrate a number of the concepts discussed, using examples from the deep-sea record of the Quaternary and the late Tertiary. It should become clear, as these examples are developed, that the reasoning in establishing the globality of a chemostratigraphic signal is not entirely circular, although some biochronologic control is assumed which is itself in need of verification.

SYSTEMIC STRATIGRAPHY OF THE QUATERNARY

The standard chemostratigraphic signal of the Quaternary is the oxygen isotope record. It approaches the ideal signal : one which describes the fluctuations of a single causal factor and which is global and closely related to basic driving functions of climatic change.

The oxygen isotope record is largely a reflection of the buildup and decay of continental ice, hence both of climate and sea level (Shackleton, Opdyke, 1973). Globality is readily demonstrated by correlating Atlantic and Pacific signals (Emiliani, Shackleton, 1974 ; see Fig. 8).

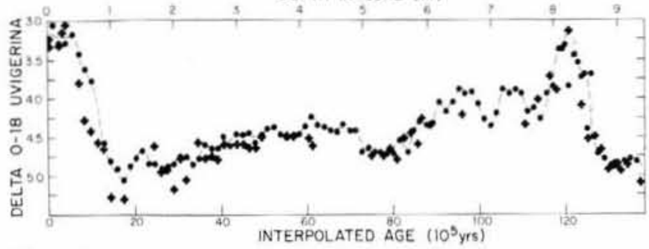


Figure 8

Comparison of oxygen isotope records from Pacific and Atlantic Oceans. Composition is for the benthic foraminifera genus Uvigerina, *off NW Africa ("Meteor") and in the east equatorial Pacific ("Vema"). Data from Shackleton (1977b), cores Meteor 12392 (2573 m) and V19-29 (3157 m) as plotted in Berger (1981). Note the striking agreement between the two records.*

One must assume that the excellent correlation shown in Figure 8 is based on strictly synchronous changes in the chemistry of the ocean (temperature changes are of minor importance well below the thermocline). Based on this assumption, the synchroneity of biostratigraphic markers can be tested. Such a test was performed by Thierstein *et al.* (1977), on late Quaternary coccolith datum levels. These authors demonstrated the synchroneity of the extinction datum of *Pseudoemiliania lacunosa* throughout the world ocean within the early part of oxygen isotope stage 12 (Fig. 9). Similarly they established the synchroneity on a global scale of the first appearance of *Emiliania huxleyi* in late oxygen isotope stage 8. By the same method of correlation of biostratigraphic events with oxygen isotope stages, Thierstein *et al.* (1977) were able to show that the replacement of *Gephyrocapsa caribbeanica* by *E. huxleyi* as the dominant coccolith species occurred earlier in the tropics than in high latitudes. The ecologic and evolutionary significance of this replacement is as yet unknown.

Thus, the systemic nature of a chemostratigraphic signal (e.g. oxygen isotope record) is established through ocean-ocean correlation. Having verified its global validity, through biostratigraphy, we can now explore, through correlation of new signals to the established one, the systemic nature of such new signals. One such signal is preservation of calcareous shells, a recordable measure of carbonate saturation of the ocean (Fig. 10).

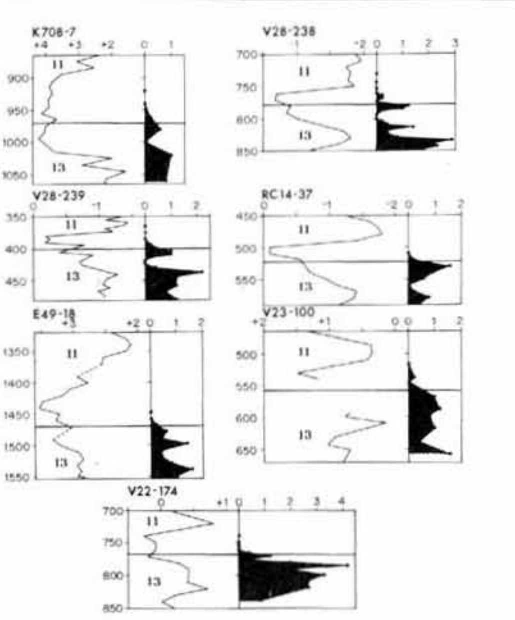


Figure 9

Analyses of seven piston cores from North Atlantic (K708-7 and V23-100), South Atlantic (V22-174), West Equatorial Pacific (V28-238, V28-239), as well as Equatorial Indian Ocean (RC14-37) and Southern Indian Ocean (E49-18). Left : oxygen isotope values for stages 11, 12, and 13. Right relative abundances of Pseudoemiliania lacunosa *given in percent. Both are plotted to depth (centimeters in core). Solid horizontal line = position of extinction level of* P. lacunosa. *From Thierstein* et al. *(1977).*

Clearly, the preservation of foraminifera, as reflected in the "solution index" calculated on the basis of species composition, is closely related to the oxygen isotope variation. Hence, the preservation signal is systemic, not just local.

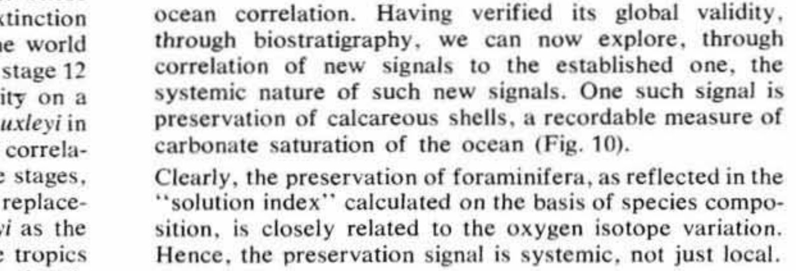
Carbonate curves are more difficult to interpret than preservation signals. Instead of only one factor (dissolution intensity) it is dependent on three (supply, dissolution, dilution by non-carbonate). Nevertheless, Quaternary car-

Figure 10

Comparison of oxygen isotope stratigraphies with preservation and carbonate records.

Left : *Oxygen isotope and preservation stratigraphies of Core V28-238, western equatorial Pacific. $\delta^{18}O$ values for* Globigerinoides sacculifer *and benthic foraminifera from Shackleton and Opdyke (1973). Solution index values (S.I.) for the > 149 µm-size fraction (high values = poor preservation), from Thompson (1976). I : last deglaciation, —11,000 yrs, II : penultimate deglaciation, —127,000 yrs ; III : prepenultimate deglaciation, —250,000 yrs ("terminations" and time scale of Broecker, van Donk, 1970). Diagram after Thompson (1976).*

Right : *Oxygen isotope and preservation stratigraphies of Core IC-5, equatorial Indian Ocean. Data from Oba (1969). Solution index calculated as* Globorotalia menardii/(G. menardii + Globigerinoides) *(see Volat* et al., *1980). Note the high degree of correlation between the records from the two oceans.*



bonate fluctuations can be correlated throughout the Indo-Pacific, being dominated here by dissolution effects (Fig. 2). The carbonate cycles of the Atlantic do not fit this scheme. Apparently these cycles are largely dilution cycles, with relatively little influence from dissolution. Also, there may be a seesaw effect between Pacific and Atlantic concerning the preservation of carbonate (Volat *et al.*, 1980). Therefore, although the Atlantic cycles are systemic in nature, they do not parallel those of the Pacific, rather they are largely anti-parallel (this relationship, of course, is a close *correlation* also).

A closer examination of the isotope curve and the solution curve in Figure 10 reveals the possibility of a phase shift between the two signals (see also Luz, Shackleton, 1975 ; Ninkovich, Shackleton, 1975 ; Shackleton, Opdyke, 1976 ; Moore *et al.*, 1977). It appears that the changes in preservation lag behind those of oxygen isotope composition. At least two processes have to be considered to attempt an explanation for this phenomenon provided it is real and not just an effect of differential mixing at the seafloor. First, the circulation of the ocean changes in response to the subtraction and addition of water, and to cooling and warming. The overall fertility and the CO_2 distribution in deep waters should be affected by such changes, which in turn would change carbonate saturation. Second, changes in carbonate mass balance which accompany sea level changes could produce dissolution cycles. After a rise of sea level the enhanced precipitation of shelf carbonates would successfully compete with the deep ocean for the available carbonate, while after a fall of sea level shelf carbonates would be exposed to erosion and transfer to the deep sea. It is significant that preservation is best, however, during a rapid rise of sea level, and worst during a fall (see below). Thus, mass balance alone cannot explain the phenomena seen.

Another signal whose systemic nature is obvious from correlation with the oxygen isotope stratigraphy is temperature of surface waters (Fig. 11). On the whole, the fluctuations of the surface temperature calculated from foraminifera in a Caribbean core, correlates very well with the isotope record from the same core. On the other hand, the curve of abundance of *Globorotalia menardii* only correlates partially with the isotope curve. Thus, there is a strong regional signal, presumably contingent upon regional extinctions and recolonization. In general, *G. menardii* becomes extinct in the Atlantic during glacials, and is reintroduced during interglacials, presumably from the Indian Ocean, past the Cape of Good Hope (Parker, 1973).

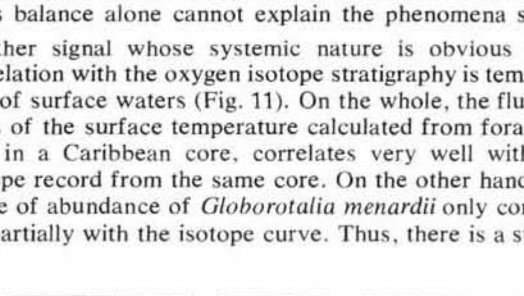


Figure 11

Comparison between Globorotalia menardii *abundance record, oxygen isotope stratigraphy, and transfer temperature record (calculated from foraminiferal assemblage compositions), as given by* Imbrie et al. *(1973). Emiliani stage numbers added to isotope record. Letters Z through T are Ericson zones defined on the basis of the presence or absence of* G. menardii. *Note the correlations, and that* G. menardii *remained absent during a warm period after regional extinction during isotope stage 14.*

The pattern is not simple, however. Although glacial conditions exclude *G. menardii*, the levels of extinction and reintroduction do not seem to have an obvious, well-defined relationship to the environment as reflected in oxygen isotopes and water temperature. Perhaps rates of change, or interactions between different factors are more important than position in the glacial cycle and temperature *per se*. The point is, the global signals can be used to explain but a portion of the regional one, the residual variation must be sought in more local conditions.

The identification and use of systemic signals is of great importance for the intercalibration of biostratigraphy, magnetostratigraphy and regional chemostratigraphy. A summary of Late Pleistocene biostratigraphic datum levels which have been calibrated in this fashion is provided in Figure 12. Note the preservation spikes coinciding with rapid sea level rise, and note the higher-than-expected coincidence of LAD's with these "termination" events. These are the periods of the most rapid and pronounced climatic, oceanographic and geochemical change within the system, during the Quaternary.

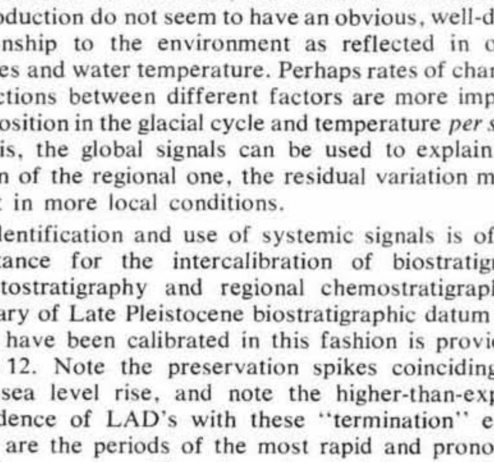


Figure 12

Correlation of biostratigraphic datum levels to the oxygen-isotope stratigraphy for the Brunhes normal polarity Epoch (0-700 ka). Composite $^{18}\delta O$-curve from Emiliani (1978). Data on the calibration of paleontologic events to the oxygen-isotope stratigraphy are from Shackleton and Opdyke (1973), Thompson and Saito (1974), Gartner and Emiliani (1976), Hays and Shackleton (1976), Thompson (1976), Burckle (1977), Thierstein et al. *(1977), Morley and Shackleton (1978) and Thompson and Sciarrillo (1978). FAD : first appearance datum ; LAD : last appearance datum ; N : calcareous nannofossil ; F : planktonic foraminifera ; R : radiolarian ; D : diatom. These datum levels followed by * are restricted to the tropical Indo-Pacific whereas the others apply from low to high latitude of the world ocean.*

Center : *calcareous fossil preservation record for the equatorial Pacific. Solid black areas : maximum preservation ; vertically ruled areas : poor preservation ; dotted and white areas : good and medium preservation (preservation data from Thompson, 1976). Note that the preservation spikes (PSI, PSII, ...) coincide with the deglaciation events (numbering scheme is that of Broecker, van Donk, 1970, and also of "Solution Cycles" of Thompson, 1976).*

SYSTEMIC STRATIGRAPHY OF THE NEOGENE

In principle, the Neogene has many characteristics of the Quaternary : for almost two-thirds of this period, there is an Antarctic ice cap, and during the last 3 million years there is evidence for northern glaciations (Savin, 1977). However, the changes of climate and the associated fluctuations of systemic signals are much less violent in the Miocene and Pliocene than in the Quaternary. In that these fluctuations are more gentle they resemble those of earlier periods of ocean history, and hence what we learn about these fluctuations may be more easily applicable to Paleogene and Cretaceous conditions than our Quaternary-derived understanding.

Without the large-scale rapid buildup and decay of continental ice which characterizes the Quaternary, oxygen isotope stratigraphy is much less pregnant with information and significance. The signal no longer signifies mainly fluctuations in ice mass and associated sea level change but is produced by a combination of a number of different factors (Fig. 6). Global synchroneity, therefore, is no longer guaranteed, and neither is the profound climatic-geochemical significance which derives from the parallelism with sea level.

The somewhat decreased importance of oxygen isotope stratigraphy is compensated by an increase in the significance of the carbon isotope signal. As mentioned, the balance between the water reservoirs ice and ocean has an analogue in the balance between the carbon reservoirs ocean/atmosphere/biosphere/reactive carbonate/reactive carbon (Fig. 7).

From the large number of carbon reservoirs it is immediately obvious that the carbon balance is more complex than the ice-ocean balance. Also, the fact that carbon is the element of life and is much less abundant than water, leads to complications through fractionations of isotopes associated with the life cycles of organisms.

An example of a detailed carbon isotope record from the Late Miocene is given in Figure 13. It is based on the analysis of benthic and planktonic foraminifera from DSDP Site 238 in the tropical Indian Ocean. Four carbon isotope signals are shown, three from benthic species and one from *Globigerinoides sacculifer*, a planktonic form. The signals are shifted to provide for maximum congruency in the lower one-third of the graph. Thus, we assume the lower part of the record represents the "normal" situation, with *G. sacculifer* being heavier than *Oridorsalis umbonatus* by about 2.5 per mil in its carbon composition as measured in $\delta^{13}C$ values. *G. sacculifer* is heavier, because it records the $\delta^{13}C$ of surface waters, and these waters have been preferentially stripped of ^{12}C by organic particle production and settling (Fig. 7 *c*). The difference in $\delta^{13}C$ of the various benthic foraminifera is unexplained, but is related to their life cycles (habitat, food and growth rates ; see Vincent *et al.*, 1981 *a* ; 1981 *b*).

We note that the several $\delta^{13}C$ signals (Fig. 13) are not clearly congruent in the lower Upper Miocene, after being shifted in the manner indicated. Also, these fluctuations show no obvious relationship to the less than coherent oxygen isotope stratigraphies of the same species.

Clearly, this portion of the stratigraphy is difficult to interpret : several unidentified factors, none strong and dominant, appear involved in producing the signals.

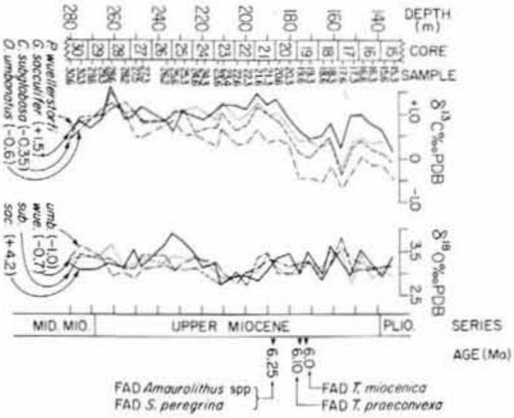


Figure 13

Upper Miocene carbon and oxygen-isotopic stratigraphies for the three benthic foraminifera Planulina wuellerstorfi, Cassidulina subglobosa *and* Oridorsalis umbonatus *and the planktonic species* Globigerinoides sacculifer *at DSDP Site 238, tropical Indian Ocean.* Left : *superimposed $\delta^{13}C$ curves ;* P. wuellerstorfi *(dotted line) plotted according to $\delta^{13}C$ values in upper scale,* C. subglobosa *(dashed line) shifted by — 0,35 per mil,* O. umbonatus *(dash/dotted line) shifted by — 0,6 per mil,* G. sacculifer *(solid line) shifted by + 1,5 per mil.*
Right : *superimposed $\delta^{18}O$ curves ;* C. subglobosa *(dashed line) plotted according to $\delta^{18}O$ values in upper scale,* P. wuellerstorfi *(dotted line) shifted by — 0,7 per mil,* O. umbonatus *(dash/dotted line) shifted by — 1,0 per mil,* G. sacculifer *(solid line) shifted by + 4,2 per mil. First Appearance Datum (FAD) for each of the species* Stichocorys peregrina *(radiolarian),* Amaurolithus *spp (calcareous nannofossils),* Thalassiosira praeconvexa *(diatom) and* T. miocenica *(diatom) shown to right. Ages follow Mankinen and Dalrymple (1979) time scale. Data from Vincent* et al. *(1980).*

The situation changes above 200 m depth in the core, where both carbon and oxygen stratigraphies become much more coherent than before. One might interpret this change as the emergence of a dominant signal-guiding factor. Shortly after, all carbon signals show a shift toward lighter values. Apparently a change in the system has now taken place which affects the chemistry and the climate of the ocean. Since both benthic and planktonic foraminifera show the effect, it involves both deep and shallow waters, and hence the entire ocean.

Before we reflect on the nature of the change, let us test the globality and synchroneity of the phenomenon. As before, we compare biostratigraphic events with the isotopic signal (Fig. 14).

The "E6 Carbon Shift" (it occurs in Magnetic Epoch 6 ; we follow the principle established by Hays *et al.*, 1969, and Dunn and Moore, 1981, for naming chemostratigraphic signals according to their position within magnetic epochs) can be detected in the isotope stratigraphies of a large number of DSDP sites. It is invariably preceded by the FAD of *Amaurolithus* spp (Haq *et al.*, 1980), supporting synchroneity both for the isotope record and the *Amaurolithus* FAD. That this reasoning is not wholly circular becomes evident when we contemplate the alternative : that the association of the FAD with the shift in each locality is a matter of coincidence. The FAD of *Amaurolithus* has furthermore been shown to be synchronous in paleomagnetically-dated piston cores within the middle normal polarity event of Epoch 6 (Haq *et al.*, 1980).

What event or events might have caused the carbon isotope excursion ? This question has been discussed by various authors (Keigwin, 1979 ; Bender and Keigwin, 1979 ; Vincent *et al.*, 1980 ; Keigwin and Shackleton, 1980) ; no

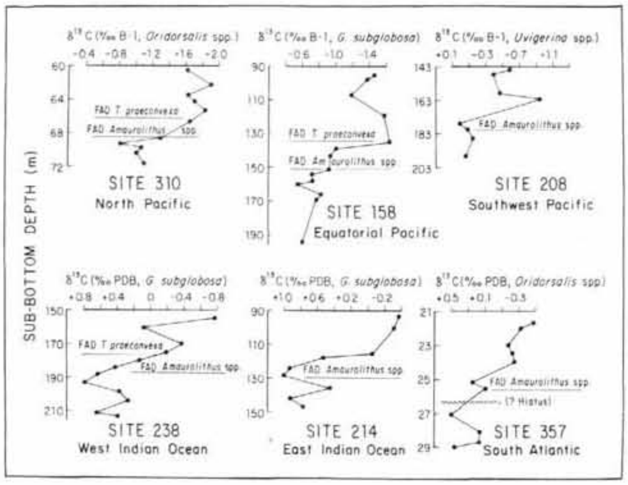


Figure 14

Upper Miocene carbon-isotopic stratigraphies for benthic foraminifera at DSDP sites and occurrence of two phytoplankton datum levels. Modified from Haq et al. *(1980).*

consensus has been reached. Our graph may contain some clues to the answer. First, we already noted that the effect is global, hence it involves large-scale geochemical balances and/or the overall circulation and fertility of the ocean. We must ask that the $\delta^{13}C$ values of the dissolved CO_2 in the ocean (mainly HCO_3^-) are changed, on the whole, by about —1 per mil at the crucial time (ca. 6 million years B.P.). The date of the event is significant : it is the time the Mediterranean enters its phase of partial isolation (in the earliest part of the younger reversed event of Magnetic Epoch 6 ; see Cita, 1979). This isolation subsequently resulted in desiccation cycles and buildup of thick salt deposits (Ryan, Hsü *et al.*, 1973 ; Ryan *et al.*, 1974 ; Hsü, Montadert *et al.*, 1978). These desiccation cycles are presumably due to a critical elevation of the entrance sill of the Mediteranean, brought about by a lowering of sea level (Van Couvering *et al.*, 1976 ; Adams *et al.*, 1977). The climatic variation paralleling the desiccation cycles (albedo changes in and around the Mediterranean !) may have been amplified by ice mass variations on Antarctica, and by variations in snow field expansion in Arctic regions, and this may explain the apparent increase in the coherence and energy of the oxygen isotope signal after the event (see Fig. 13).

The consequences of a drop in sea level are manifold. One that is thought to apply here is the large-scale introduction, through surficial erosion of shelf sediments and continental soils, of ^{12}C-rich organic matter. A decrease in forest cover also may be involved (cf. Shackleton, 1977 *a*). The shift seen could be produced by an addition of an organic carbon mass between twice and three-times the mass of carbon in the atmosphere.

The E6 Carbon Shift may be largely or partially produced by changes in fertility patterns of the ocean. We note in Figure 13 that the *Globigerinoides sacculifer* signal diverges from the *Oridorsalis umbonatus* signal at the beginning of the event, with *G. sacculifer* tending toward heavier values relative to *O. umbonatus*. A divergence is also seen in the benthic foraminifera. This divergence cannot be produced solely by a general change in $\delta^{13}C$ composition of ocean waters : it must involve deep circulation and fertility patterns which allow increased differentiation of surface and deep waters, and increased differential fractionation between species. Differences in carbon fractionation between co-occurring benthic foraminifera appear to increase with increasing fertility (Vincent *et al.*, 1981 *a*), hence, the E6 Carbon Shift in all probability marks a fertility increase.

An additional effect may be derived from a change in deep circulation. The semi-isolated Mediterranean must have functioned as a source of heavy (saline) water to the North Atlantic, favoring the production of deep water there. An onset or an increase in production of North Atlantic bottom water during this time could have led to relatively higher CO_2 content of Indo-Pacific deep waters. Such augmentation in respirative CO_2 leads to lower $\delta^{13}C$ values (Kroopnick *et al.*, 1977).

It is interesting that the benthic foraminifera *Uvigerina*, whose abundance increases markedly near the E6 Event, appears to prefer CO_2-rich bottom waters (Streeter, Shackleton, 1979) and regions of elevated fertility in the present ocean (Vincent *et al.*, 1981 *a*).

We lay no claim that this brief discussion of the E6 Carbon Shift exhausts the possible explanations. However, it does exemplify the subtle nature of carbon isotope stratigraphy. This stratigraphy contrasts markedly with the simplicity of oxygen isotope stratigraphy, which is interpreted essentially in terms of temperature and ice volume, neglecting other factors (Savin, 1977 ; Shackleton, 1977 *b* ; Duplessy, 1978).

Carbonate preservation remains as a source of strong chemostratigraphic signals in pre-Quaternary times. In fact, it shows extreme excursions at certain times (Fig. 15) which can be used for correlation over wide geographic areas under different oceanographic regimes (Kaneps, 1973 ; Vincent, in press). These excursions or "dissolution spikes" have global geochemical significance, therefore. The mechanisms producing carbonate fluctuations and excursions in terms of ocean geochemistry are still largely unknown. The Pleistocene record (Fig. 12) suggests that there are at least two such mechanisms in operation : one which is responsible for an overall correlation between oyxgen isotope and carbonate record, and another which is associated with the transition between warm and cold stages.

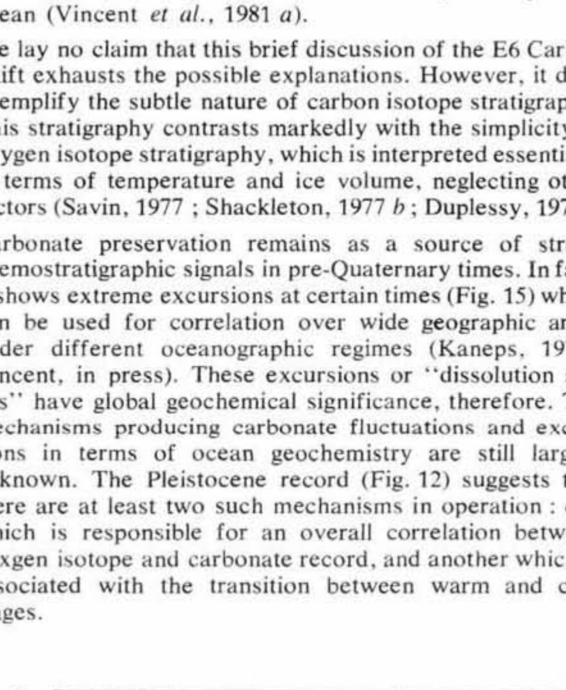


Figure 15

Carbonate stratigraphy and magnetostratigraphy of Core RC 12-66, equatorial Pacific, from Saito et al. *(1975). Terminology of pronounced carbonate minima (M 17, M 21, GU 3 and GI 17) according to Hays* et al. *(1969) and Dunn and Morre (1981).*

The first mechanism presumably is associated with equilibrium states, and mass balance arguments such as shelf-basin partitioning of carbonate would seem appropriate. In essence, during times of high sea level the shelves would rob the deep-sea of carbonate, and *vice versa*. The second mechanism, it appears, is associated with temporary disequilibrium conditions, and hence transient changes in ocean chemistry must be invoked. Two pronounced Pliocene carbonate minima, M 21 and GU 3 (Fig. 15), are marked by a distinct enrichment in ^{18}O of deep-sea benthic foraminifera in both Pacific equatorial and transitional waters (Vincent, in press). The co-occurrence of dissolution spikes and cooling/regression events is also seen in the Pleistocene record (Fig. 12). We speculate that this coincidence signifies release of organic carbon to the ocean system, from transient carbon reservoirs (shallow marine organic-rich sediments, marshes, swamps, soil, and forests). The dissolution spike GI 17 may be produced by a similar process, following the end-of-Miocene regression. If our suggestion holds, the dissolution spikes should be associated with (negative) $\delta^{13}C$ shifts in the carbon isotope record. There is some evidence that this is so.

The mechanism of organic-carbon injection from a transient reservoir is attractive because by titrating CO_2 with carbonate a pronounced dissolution spike can be produced quite readily. However, alternative (or additional) mechanisms must also be considered. A general increase in fertility of the ocean would raise the CCD (Broecker, 1974), that is, produce a dissolution spike if temporary. Such a temporary increase could be produced, for example, by increased erosion and supply of phosphate from land. Also, a change in deep circulation could produce an Indo-Pacific dissolution spike. Regression and cooling affects the North Atlantic region especially intensely, through increased evaporation and through albedo feedback on the surrounding land areas. Hence, regression and cooling favor North Atlantic bottom water supply, which in turn favors trapping of carbonate by the North Atlantic Basin (as at present), and hence increase in dissolution elsewhere. Such a seasaw mechanism may apply in the Pleistocene (Volat *et al.*, 1980) and can be tested through basin-basin correlation.

We make no claim that these suggestions are more than preliminary working hypotheses at this point. When we have learned to interpret the carbonate fluctuations properly, we will then be in a much better position to interpret the $\delta^{13}C$ record as well. Both records, after all, are signals stemming from the carbon cycle, and we would expect a close correspondence between the two.

CONCLUSION

The ocean, through sea-level changes and through its control of climate, is the principal regulator of the exogenic system which produces the marine sedimentary record, and to some extent the terrestrial record as well. The activity of this regulator is written into the deep-sea record whence it can be extracted as chemostratigraphic and biostratigraphic signals of global significance. Once these signals are available they can be used as systemic stratigraphic standards. Through detailed correlation with specific stratigraphic sequences, the standards will allow recognition of the global, regional, and local components of such sequences. Such analysis is the basic task of stratigraphy : it will allow the full use of biostratigraphic information in terms of ecology and evolution. Eventually it will render obsolete the concept of "type section" for those periods for which a continuous deep-sea record is available.

Acknowledgements

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