Dysoxic/Anoxic Episodes in the Aptian-Albian (Early Cretaceous)

TIMOTHY J. BRALOWER

Department of Geology, University of North Carolina, Chapel Hill, NC

WILLIAM V. SLITER

Branch of Paleontology and Stratigraphy, USGS, Menlo Park, CA

MICHAEL A. ARTHUR

Department of Geosciences, Pennsylvania State University, University Park, PA

R. MARK LECKIE

Department of Geology, University of Massachusetts, Amherst, MA

DAVID ALLARD

Graduate School of Oceanography, University of Rhode Island, Narragansett, RI

SEYMOUR O. SCHLANGER

Department of Geological Sciences, Northwestern University, Evanston, IL

New biostratigraphic data suggest, for the first time, that the long Aptian-Albian Oceanic Anoxic Event was marked by at least three distinct, relatively brief episodes of widespread dysoxia/anoxia which interrupted generally oxygenated conditions. The high-resolution, integrated foraminiferal and nannofossil biostratigraphy derived here allows recognition of an apparently ocean-wide dysoxic/anoxic episode in the early Aptian (*Globigerinelloides blowi* foraminiferal Zone, *Chiastozygus litterarius* nannofossil Zone, *Conusphaera rothii* nannofossil Subzone, shortly after magnetic Chron CM0). Equally widespread, but distributionally patchier dysoxic/anoxic episodes occurred in the early Albian (*Hedbergella planispira* foraminiferal Zone, *Prediscosphaera columnata* nannofossil Zone, subzone NC8B) and early late Albian (*Biticinella breggiensis* foraminiferal Zone, *Axopodorhabdus albianus* nannofossil Zone, subzone NC9B). These episodes can be best recognized in highly carbonaceous sediments deposited in epicontinental basins, now exposed on land, including the lower Aptian "Livello Selli" of the Italian Apennines, the lower Albian "Niveau Paquier" of the Fosse Vocontienne, France, and the lower upper Albian Toolebuc Formation of the Eromanga Basin, Queensland, Australia. Our data indicate that these horizons correlate at the nannofossil subzonal level to carbonaceous intervals in DSDP/ODP sites which were deposited in pelagic and hemipelagic oceanic settings.

Although none of these episodes is associated with major biotic extinctions, they are characterized by changes, of variable magnitude, in the community structure of planktonic foraminifera, which commonly consist of a low-diversity assemblage of opportunistic taxa or are entirely absent. Nannofossil taxa in these horizons do not change as radically, but in some sites show marked nearshore affinities or, in other sequences, are possibly replaced by other phytoplankton. Dissolution events cannot be entirely ruled out as a cause of some of the changes in calcareous biota. The proposed dysoxic/anoxic episodes appear to correlate with sea level transgressions or highstands. Rising sea level and climatic consequences of volcanism conditioned the oceans to be prone to dysoxia/anoxia. The spatial and sedimentological variability of organic carbon-rich intervals, and the fact that at least two of them have a patchy geographic distribution, indicate that regional or local climatic, tectonic or oceanographic factors induced or triggered dysoxia/anoxia.

INTRODUCTION

Sy Schlanger's interest in the enigma of Cretaceous black shales stemmed from his life-long fascination with the geology of the Pacific Ocean. Some of the first truly pelagic organic-rich sediments were recovered in the Pacific during Deep Sea Drilling Project Legs 17 and 33, of which Schlanger was co-chief scientist. His classic paper coauthored with Hugh Jenkyns in 1976, based partly on the results of these legs, recognized that these organic-rich sediments were not deposited randomly through time, but tend to be concentrated in particular time intervals with suitable oceanographic and/or climatic conditions. Schlanger and Jenkyns [1976] termed these Oceanic Anoxic Events (OAEs), and this theory and term is now cemented in the literature. The original OAEs proposed were

The Mesozoic Pacific: Geology, Tectonics, and Volcanism
Geophysical Monograph 77

localized within the Aptian-Albian (Early Cretaceous) and Cenomanian/Turonian boundary (Late Cretaceous).

Subsequent work by Schlanger contributed to various aspects of black shale research. His work on Cretaceous volcanism in the Pacific, which culminated in a paper with Jenkyns and Isabella Premoli Silva published in 1981, documented the major episode of mid-plate volcanism in the Pacific from the Aptian through the Santonian (125-80 Ma). This paper correlated this volcanic episode to "the great Late Cretaceous transgression" [e.g., Hancock and Kauffman, 1979], which involved the highest sea levels of the Phanerozoic Era [Vail et al., 1977; Haq et al., 1987]. High sea level caused by intraplate volcanism and thermal rejuvenation of the sea floor are cited as critical factors by most authors who have discussed the origin of Cretaceous black shales [e.g., Jenkyns, 1980; Waples, 1983; Weissert, 1989]. The hypsographic and atmospheric effects of this volcanic activity both indirectly led to black shale deposition. High sea level, higher-than-present atmospheric carbon-dioxide concentrations [e.g., Berner et al., 1983] and vastly different paleogeography [e.g., Barron and Washington, 1982] led to a Cretaceous climate that is considered to have been warm and equable, with continental ice sheets generally absent. As the result of relatively low thermal gradients and an active hydrologic cycle, ocean circulation may have been influenced more by haline-dominated convection than at present [e.g., Brass et al., 1982; Barron and Peterson, 1990]. High mean temperatures and reduced oxygen solubility, as well as generally slow deep circulation, created an ocean highly susceptible to dysoxic/anoxic conditions [e.g., Thiede and van Andel, 1977; Roth, 1978; Thierstein and Berger, 1978; Arthur and Natland, 1979; Demaison and Moore, 1980; Pedersen and Calvert, 1990]. The profundity of Schlanger's work on Pacific volcanism is now just being realized with the discovery that individual volcanic episodes such as the formation of the Ongtong Java Plateau, coincide with OAEs [Larson, 1991a; Tarduno et al., 1991].

One of Schlanger's favorite projects concerned the detailed investigation of the origin and effects of the Cenomanian/Turonian boundary OAE on the marine environment. This OAE, which was clearly the most dramatic and widespread of these episodes, is marked by turnover of macrofossil groups [e.g., Raup and Sepkoski, 1982; Elder, 1989] and corresponds to a widespread excursion of carbon isotopes [e.g., Scholle and Arthur, 1980], apparently reflecting the rapid removal of large quantities of organic material from the oceanic system. In collaboration with Michael Arthur, Jenkyns and Peter Scholle, Schlanger carried out a detailed investigation of the sedimentology, distribution and carbon isotopic stratigraphy of organic-rich horizons in a number of Cenomanian/Turonian boundary sequences. The culmination of this project was the publication in 1987 of an oceanographic model that elegantly accounted for the global distribution of black shale horizons in this time interval. Through the flooding of shallow epicontinental seaways and the production of saline bottom waters, Schlanger and colleagues proposed widespread displacement and upwelling of nutrient-rich deep waters. This upwelling led to the expansion of oxygen minimum zones in shelf settings which was conducive to black shale deposition.

Schlanger was involved in the inception, funding and early stages of this present study which concerns the origin, spatial and temporal distribution of organic-rich sediments during OAE 1 in the Aptian-Albian. The term OAE is not strictly appropriate for this long (20 m.y.) interval. Although sediments enriched in TOC (Total Organic Carbon) are common, their spatial and temporal distribution has not been well documented. A salient question remains as to whether there are discrete intervals in which carbonaceous sediments are concentrated within the overall Aptian-Albian time window, or whether sedimentation of TOC-rich units occurred at different times in different places. Sliter [1989a] correlated organic-rich horizons from Pacific DSDP Sites 167, 305 and 463 to the early Aptian *Globigerinelloides blowi* foraminiferal Zone and proposed that they were the same age as the "Livello Selli" of Italy and the "Niveau Goguel" of southern France based on the work of Coccioni et al. [1987; 1989], Erba [1988] and Bréhéret [1988]. The regional nature of an early Albian organic-rich episode also was demonstrated in European land sections by Bréhéret [1985]. By correlating calcareous nannofossil and planktonic foraminiferal biochronologies, and by surveying a host of additional land and DSDP/ODP sections, we have taken these previous studies one step further by defining exact temporal and spatial limits for these intervals of oxygen deficiency. Following the terminology introduced by Arthur et al. [1990], we term these episodes "oceanic anoxic subevents".

This paper is dedicated to the memory of Sy Schlanger. Although he did not see this project come to fruition, his guidance and energy during its early stages greatly helped us address the problems that we faced. While Sy might not have been entirely convinced by our conclusions, we certainly attribute much of the scientific groundwork that led to this project to his previous studies of Cretaceous black shales.

TECHNIQUES

We have conducted a thorough literature review of 39 Aptian-Albian sections recovered by DSDP/ODP (Table 1). Certain sections were too poorly recovered (e.g., DSDP Site 418) or too poorly fossiliferous (e.g., DSDP Site 317) to be studied further. Other sections (e.g. DSDP Site 547) were almost devoid of organic material but were still an integral part of our analysis of the spatial distribution of anoxia/dysoxia. This study is based upon combined lithostratigraphic and biostratigraphic investigations of 6 land and 15 DSDP/ODP sections (Fig. 1). Certain sequences have been studied largely for biostratigraphic purposes. These sections are expanded, mostly TOC-poor, nearly continuously recovered and contain well-preserved microfossil assemblages. Other sequences, including all of the land sections, have been studied largely because they contain horizon(s) enriched in TOC. The latter sections generally contain less well-preserved microfossil assemblages characterized by moderate to extreme amounts of overgrowth and etching which systematically remove solution-susceptible taxa and reduce diversity [e.g., Thierstein, 1980]. There is some evidence that the TOC-rich units of the land sections are stratigraphically condensed relative to correlative deposits from certain deep-sea sites [Bralower, et al., submitted]. Several of the DSDP/ODP sections

TABLE 1. DSDP/ODP Sections Considered for this Investigation

Site	Location	Stage	Range TOC	Reason Unselected
105	W. N. Atlantic	Apt.-Alb.	0-4%	Poor recovery, preservation
144	Demerara Rise	U.Apt.	Unknown	Poor recovery
167	**Magellan Rise**	**U.Barr.-U.Apt.**	**Unknown**	
258	**Naturaliste Plat.**	**M.Alb.-U.Alb.**	**0-3%**	
288	Ontong Java Plat.	U.Apt., U.Alb.	~0	Poor recovery
289	Ontong Java Plat.	U.Apt.	~0	Poor recovery
305	Shatsky Rise	L.Apt.-U.Alb.	0-9.3%	Poor recovery, preservation
306	Shatsky Rise	Aptian, U.Alb.	~0	Poor recovery, preservation
317	Manihiki Plateau	L.Apt.-U.Alb.	0-29%	Poor preservation
327	Falkland Plateau	L.Apt-M.Alb	0-3.5%	Same section as Site 511
330	Falkland Plateau	L.Apt-M.Alb	0-4%	Same section as Site 511
356	Rio Grande Rise	U.Alb.	0-2%	Poor preservation
361	Cape Basin	L.Apt.-M.Alb.	0-14%	Poor preservation
363	Walvis Ridge	U.Apt.-U.Alb.	<1%	Poor preservation, low TOC
364	**Angola Basin**	**L. Apt**	**0-40%**	
367	Cape Verde Basin	L.Apt-U.Alb.	0-35%	Poor preservation
370	**Morocco Basin**	**L.Apt.-U.Alb.**	**0-7%**	
386	**Bermuda Rise**	**L.Alb-U.Alb.**	**0-15%**	
387	Bermuda Rise	L.Apt.-U.Alb.	0-4%	Poor preservation
391	Blake-Bahama Basin	L.Apt.-M.Alb.	0-3%	Poor recovery
398	**Vigo Seamount**	**U.Barr.-U.Apt.**	**0-3%**	
400	**Bay of Biscay**	**U.Apt.-U.Alb.**	**0-3%**	
402	**Bay of Biscay**	**L.Apt.-U.Apt.**	**0-12%**	
417	**Bermuda Rise**	**L.Apt.-U.Apt.**	**0-21%**	
418	Bermuda Rise	L.Apt.-U.Alb.	0-10%	Poor recovery
463	**Mid. Pac. Mtns.**	**L.Apt.**	**0-8%**	
464	Hess Rise	U.Apt.-U.Alb.	0-4%	Poor recovery
465	Hess Rise	U.Alb.	0-9%	Poor recovery, preservation
466	Hess Rise	U.Alb.	0-5%	Poor recovery
511	**Falkland Plateau**	**U.Barr.-U.Alb.**	**0-10%**	
530	Walvis Bay	U.Alb.	0-9%	Poor preservation
534	Blake-Bahama Bas.	L.Apt.-M.Alb.	0-3%	Poor preservation
535	Gulf of Mexico	L.Apt-M.Alb.	0-6%	Poor preservation, low TOC
540	Gulf of Mexico	L.Alb-U.Alb.	0-4%	Poor preservation, low TOC
545	**Mazagan Esc.**	**U.Apt.-U.Alb.**	**0-5%**	
547	**Mazagan Esc.**	**U.Alb-L.Cen.**	**0-1.5%**	
641	**Galicia Bank**	**U.Barr.-M.Alb.**	**0-26%**	
693	Weddell Sea	U.Apt.-L.Alb.	0-3%	No zonal markers
763	**Exmouth Plat.**	**U.Barr.-Cen.**	**<1%**	

Selected sites shown in bold. Stratigraphic intervals shown for these are those studied here.

investigated, including Sites 258, 364, 402 and 417, are poorly recovered (20-40%) but still reveal interesting trends.

Calcareous nannofossil biostratigraphy was carried out by Bralower on all of the sequences investigated. Thickly-concentrated smear slides were prepared so that even the rarest of occurrences could be noted. Preservation of nannofossils ranges from poor to good in the investigated sections. Special attention was paid to samples in critical stratigraphic intervals, especially towards the ends of species ranges, where slides were examined for several hours. Foraminiferal micropaleontology was carried out by Leckie on the Atlantic DSDP/ODP Sites and by Sliter on the Italian land sections and Pacific DSDP Sites. Leckie concentrated mostly on marlstones and claystones which could be washed using standard techniques, and Sliter worked predominantly on limestones using thin sections. The two workers collaborated closely and exchanged samples in critical intervals. Detailed foraminiferal taxonomy and stratigraphic distributions will be presented elsewhere [Leckie and Sliter, in prep.] A detailed analysis of the early Aptian interval has been offered in Bralower et al. [submitted]. That publication includes comprehensive nannofossil range charts of all of the sections investigated. In this publication, we display abbreviated range charts of the Albian sections investigated which include only the stratigraphic distributions of important biomarker taxa. Complete sample-by-sample range charts of all of the taxa identified in the other sections are illustrated in Sliter and Bralower [in press]. A list of all

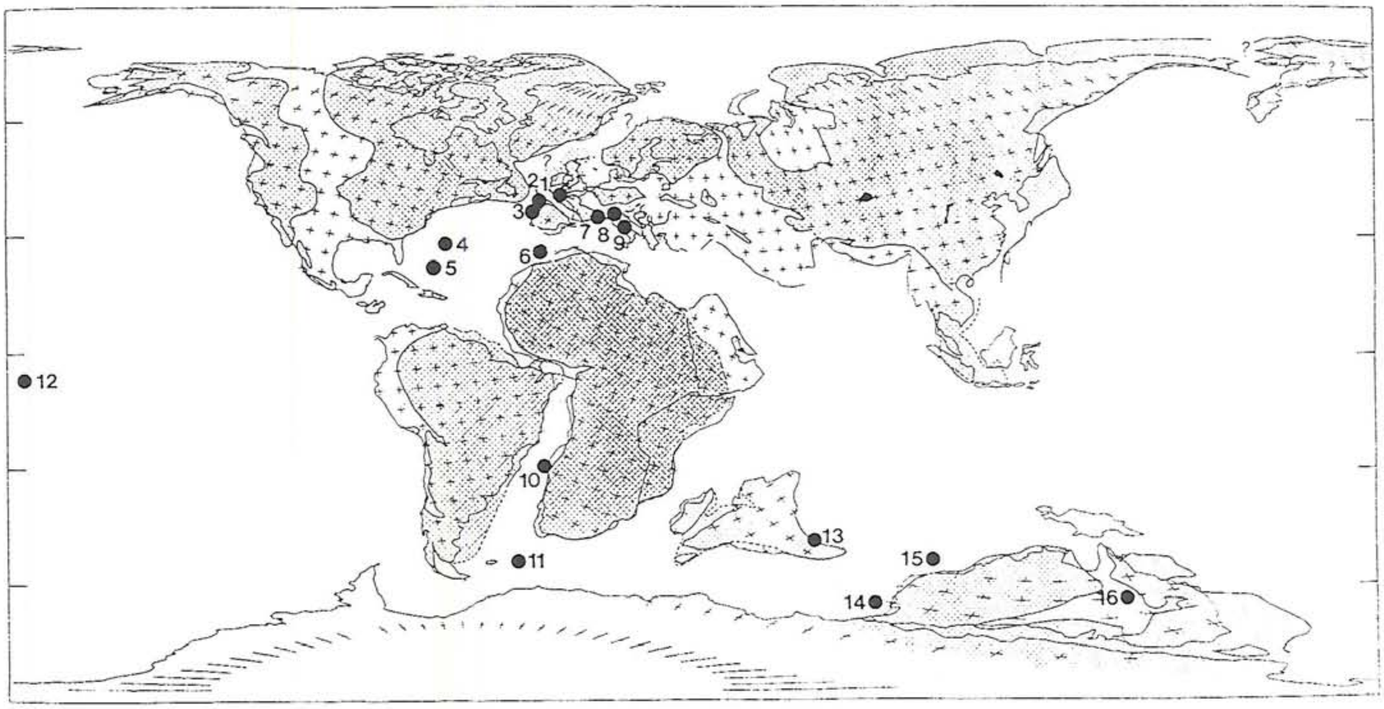


Fig. 1. Paleogeographic location of DSDP/ODP sites and land sections studied in this investigation. 1. Bay of Biscay (DSDP 400, 402), 2. ODP 641, 3. DSDP 398, 4. DSDP 386, 5. DSDP 417, 6. Morocco Basin, DSDP Sites 370, 545, 547, 7. Fosse Vocontienne (Col de Palluel, Vergons), 8. Cismon, 9. Apennines (Gorgo A Cerbara), 10. DSDP 364, 11. DSDP 511, 12. DSDP 463, 13. Cauvery Basin, 14. DSDP 258, 15. ODP 763, 16. Eromanga Basin. Paleogeographic reconstruction of Barron et al. [1981].

nannofossil and planktonic foraminifer taxa discussed here is provided in Table 2.

TOC and carbonate contents of samples from DSDP/ODP sites were taken from the literature and analyzed as part of this investigation (Table 3). Analyses were carried out either at GSO/URI or at UNC-CH using Coulometrics coulometers. A large number of samples of different lithologies at individual sites have been analyzed in order to avoid the problem of representative sampling and single TOC values for long intervals. The absolute concentration of organic carbon in any sample is a result of the carbon flux to the sea floor and decomposition rates during burial, as well as the supply of dilutants such as carbonate, silica and clastic material. Higher TOC contents or peaks in TOC profiles in DSDP/ODP sections have been interpreted to indicate enhanced organic matter preservation and accumulation regardless of the cause. Because supply rates of carbon and dilutants vary considerably from site to site, we have not erected criteria as to what constitutes "TOC-rich" or "TOC-poor" and instead we use relative terms such as "higher" and "lower" to describe trends in TOC in individual locations. We would prefer to interpret accumulation rates of TOC, a factor which takes into account differential compaction and negates the effect of dilutants on TOC contents. However, given the current uncertain status of radiometric control in the middle Cretaceous, it is not possible to calculate precisely the component accumulation rates.

INTEGRATED APTIAN-ALBIAN MICROFOSSIL BIOCHRONOLOGY

Original zonations for Aptian-Albian planktonic foraminifera and calcareous nannofossils were developed up to 20 years ago [e.g., Bolli, 1966; Moullade, 1966; Thierstein, 1971; Longoria, 1974] in sections from the Tethyan province. These zonations have been modified somewhat since in part by the addition of subzones [e.g., Manivit et al., 1977; Sigal, 1977; Roth, 1978]; however, the resolution remains low, averaging 4-6 m.y. per zone and 2-3 m.y. per subzone. Recently, there have been advances both in techniques and taxonomy. For both microfossil groups, techniques have evolved allowing the study of indurated sedimentary rocks such as limestones [e.g., Monechi and Thierstein, 1985; Sliter, 1989b]. Taxonomic advances which include the clarification of existing taxa, and the discovery of new taxa also lead to increased potential resolution as there are more species available for biostratigraphy. Continued operation of the Deep Sea Drilling Project and Ocean Drilling Program has led to recovery of new Aptian-Albian sections in the last twenty years. These include sites from high latitudes and from all of the different ocean basins. Many of these sections are expanded relative to those in which original zonations were developed. In this paper, we propose an upper Barremian-lower Cenomanian calcareous nannofossil biostratigraphic scheme with higher resolution than those presently available. In addition, we propose a preliminary integration of

TABLE 2. Microfossil Taxa Discussed

Calcareous Nannofossils
Parhabdolithus achlyostaurion Hill, 1976
Corollithion achylosum [Stover, 1966] Thierstein, 1971
Corollithion acutum Thierstein in Roth and Thierstein, 1972
Lithraphidites acutum Verbeek and Manivit in Manivit et al. [1977]
Cretarhabdus angustiforatus [Black, 1971] Bukry, 1973
Parhabdolithus angustus [Stradner, 1963] Stradner, Adimaker & Maresch, 1968
Prediscosphaera columnata [Stover, 1966] Perch-Nielsen, 1984
Grantarhabdus coronadventis [Reinhardt, 1966] Grün in Grün and Allemann, 1975
Axopodorhabdus decorus [Deflandre, 1954] Wind and Wise in Wise and Wind, 1977
Broinsonia enormis [Shumenko, 1968] Manivit, 1971
Eiffellithus cf. *E. eximius* [e.g., Hill and Bralower, 1987]
Sollasites falklandensis Filewicz et al., in Wise and Wind, 1977
Eprolithus floralis [Stradner, 1962] Stover, 1966
Percivalia hauxtonensis Black, 1973
Micrantholithus hoschulzii [Reinhardt, 1966] Thierstein, 1971
Parhabdolithus infinitus [Worsley, 1971] Thierstein, 1972
Rucinolithus irregularis Thierstein in Roth and Thierstein, 1972
Corollithion kennedyi Crux, 1981
Chiastozygus litterarius [Gorka, 1957] Manivit, 1971
Octocyclus magnus Black, 1972
Flabellites oblongus [Thierstein, 1973] Crux, 1982
Conusphaera rothii [Thierstein, 1971] Jakubowski, 1986
Corollithion signum Stradner, 1963
Prediscosphaera spinosa [Bramlette and Martini, 1964] Gartner, 1968
Nannoconus steinmannii [Kamptner, 1931] subsp. *steinmannii*
Crucicribrum striatum subsp. *constansii* Wise and Parker, in Wise, 1983
Eiffellithus trabeculatus Gorka, 1957
Eiffellithus turriseiffelii [Deflandre in Deflandre and Fert, 1954] Reinhardt, 1965
Planktonic Foraminifera
Biticinella breggiensis [Gandolfi, 1942]
Globigerinelloides algerianus Cushman and ten Dam, 1948
Globigerinelloides blowi [Bolli, 1959]
Globigerinelloides duboisi [Chevalier, 1961]
Globigerinelloides ferreolensis [Moullade, 1961]
Hedbergella gorbachikae Longoria, 1974
Hedbergella planispira [Tappan, 1940]
Hedbergella similis Longoria, 1974
Hedbergella trochoidea [Gandolfi, 1942]
Leupoldina cabri [Sigal, 1952]
Planomalina buxtorfi Gandolfi, 1942
Planomalina cheniourensis Sigal, 1952
Rotalipora appenninica Renz, 1936
Rotalipora brotzeni Sigal, 1948
Rotalipora subticinensis [Gandolfi, 1957]
Rotalipora ticinensis [Gandolfi, 1942]
Schackoina cenomana [Schacko, 1897]
Ticinella bejaouaensis Sigal, 1966
Ticinella praeticinensis Sigal, 1966
Ticinella primula Luterbacher, 1963
Schackoina spp.

biostratigraphies based on nannofossils and planktonic foraminifera.

The nannofossil zonation scheme developed (Fig. 2) is based largely on studies of continuous DSDP/ODP sites with good microfossil preservation. The sections and intervals applied include DSDP Sites 398 (upper Barremian to upper Aptian), 511 (upper Barremian to upper Albian), 545 (upper Aptian to upper Albian), 547 (upper Albian to lower Cenomanian), and ODP Sites 641 (upper Barremian to lower Albian) and 763 (upper Barremian to lower Cenomanian). Detailed range charts for these sections are illustrated in Bralower et al. [submitted; late Barremian-early Aptian interval of all sections], Bralower and Siesser [1992; ODP Site 763] and Sliter and Bralower [in press; upper Aptian to Cenomanian portions of Sites 511, 545, 547 and 641].

In order to be fully compatible with previously established zonations, we propose a series of informal subzonal divisions for the widely-used Roth [1978] nannofossil zonation which was modified after the zonation of Thierstein [1971, 1973]. The two subzones of the *Chiastozygus litterarius* Zone (NC6) have been formally defined by Bralower et al. [submitted]. Subzonal units are recognizable in all sections except those with the most poorly preserved microfossils. We propose a series of additional events, termed biohorizons, the relative order of which is not as well established partly because they can only be precisely determined in expanded sequences with good microfossil preservation (e.g., Sites 398, 545, 547, 641, 763). A full discussion of these biohorizons and a detailed correlation of zonal and subzonal units applicable in low latitude sites with those in high latitude settings is given in Bralower [1992]. We stress that the order proposed for these biohorizons, and their correlation with zonal and subzonal units, is tentative and will continue to be revised as new sequences are found and as taxonomic concepts are clarified.

The definitions of the zonal and new subzonal units, which are also classified using the "NC" prefix of Roth [1978] are as follows:

Chiastozygus litterarius Zone (NC6)
Interval from the first occurrence (FO) of *Rucinolithus irregularis* to the FO of *Eprolithus floralis* [Thierstein, 1971; emended by Wiegand, 1984; Erba and Quadrio, 1987; Erba, 1988].

Subzone NC6A: *Conusphaera rothii* Subzone
Interval from the FO of *Rucinolithus irregularis* to the last occurrence (LO) of *Conusphaera rothii*.
Subzone NC6B: *Grantarhabdus coronadventis* Subzone
Interval from the LO of *C. rothii* to the FO of *E. floralis*.

Comments: The last occurrence of *C. rothii* can be determined precisely in deep-sea sections, but this event is more difficult to establish in the land sections due to poorer preservation and the rarity of this taxon.

Parhabdolithus angustus Zone (NC7)
Interval from the FO of *E. floralis* to the FO of *Prediscosphaera columnata* [Thierstein, 1971, 1973].

TABLE 3. Sources of Lithological, Biostratigraphical and Magnetostratigraphical Data in Sections Investigated

Section	Lithostratigraphy	Biostratigraphy*	Magnetostratigraphy	Carbon/Carbonate
DSDP 167	Winterer et al., 1973	Roth, 1973 (n); Douglas, 1973 (f); Tarduno et al., 1989 (n,f)	Tarduno et al., 1989	----
DSDP 258	Davies et al., 1974	Thierstein, 1974 (n)	----	Davies et al., 1974; this study
DSDP 364	Bolli et al., 1978	Proto Decima et al., 1978 (n); Caron, 1978	----	Foresman, 1978; this study
DSDP 370	Lancelot et al., 1977	Cepek, 1977 (n); Pflaumann and Krasheninnikov, 1977 (f)	----	Erdman and Schorno, 1977 this study
DSDP 386	Tucholke et al., 1979	Okada and Thierstein, 1979 (n)	----	Cameron, 1979; Bralower, 1984
DSDP 398	Arthur, 1979; Sibuet et al., 1979	Blechschmidt, 1979 (n); Sigal, 1979 (f)	----	Sibuet et al., 1979; Erdman and Schorno, 1979; Johnson et al., 1979
DSDP 400	Montadert, Roberts et al., 1979	Dupeuble, 1979 (f); Müller, 1979 (n)	----	Deroo et al., 1979; this study
DSDP 402	Montadert, Roberts et al., 1979	Müller, 1979 (n)	----	Deroo et al., 1979; this study
DSDP 417	Donnelly et al., 1980	Gartner, 1980 (n); Siesser, 1980 (n)	----	Deroo et al.,1980; this study
DSDP 463	Dean et al., 1981	Cepek, 1981 (n); Roth, 1981 (n); Tarduno et al., 1989 (n,f)	Tarduno et al., 1989	Dean, 1981; this study
DSDP 511	Ludwig et al., 1983	Krashenninikov and Basov, 1983 (f); Wise, 1983 (n)	Salloway, 1983	Deroo et al., 1983; Bode, 1983; Parker et al., 1983; this study
DSDP 545	Hinz et al., 1984	Leckie, 1984 (f) Wiegand, 1984 (n)	----	Hinz et al., 1984; Deroo et al., 1984
ODP 641	Boillot et al., 1988	Applegate and Bergen, 1988 (n)	Ogg, 1988	Boillot et al., 1988; Dunham et al., 1988; this study
ODP 763	Haq et al., 1990	Bralower and Siesser, 1992 (n)	----	this study
Palluel	----	----	----	this study
Vergons	----	----	----	this study
Cismon	----	Channell et al., 1979 (n,f); Bralower, 1987 (n)	Channell et al., 1979	this study
Piobbico	----	----	----	this study
Cerbara	Coccioni et al., 1989	Lowrie et al., 1980 (n,f); Bralower, 1987 (n)	Lowrie and Alvarez, 1984	this study
Augathella	----	Shafik, 1985 (n)	----	----
Cauvery	----	Kale and Phansalkar, 1992 (n)	----	----

* n - nannofossils; f - foraminifera. The nannofossil biostratigraphy of all sections was investigated as part of this study. Foraminifera were investigated in Sites 370, 400, 511, 545 and 641 and in the Palluel, Vergons, Cismon and Cerbara Sections.

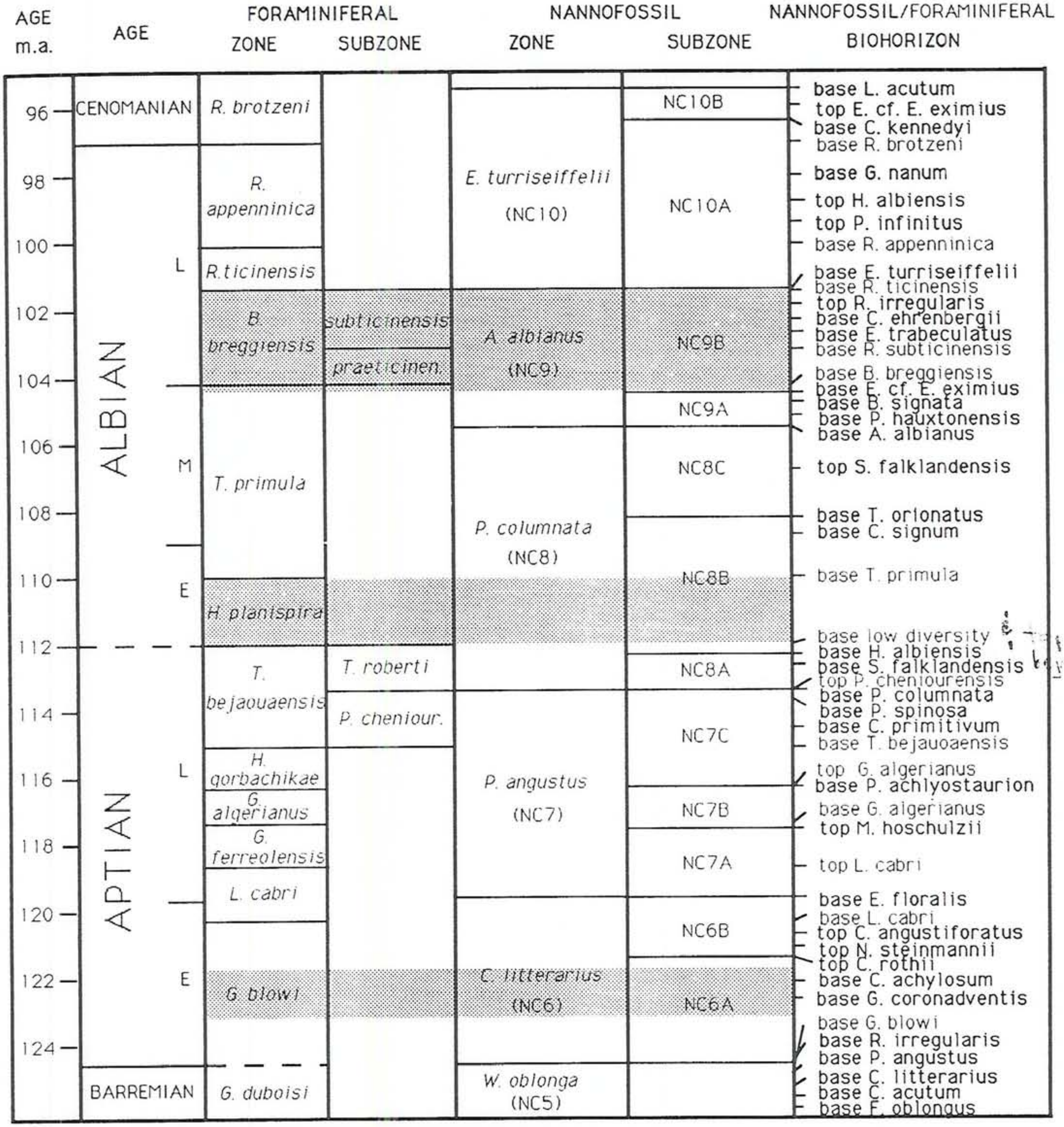


Fig. 2. Aptian-Albian planktonic foraminiferal-calcareous nannofossil biostratigraphy and the timing of the three proposed dysoxic/anoxic episodes (shown in grey). Planktonic foraminifera biostratigraphy is modified after Sigal [1977] and Sliter [1989b], calcareous nannofossil biostratigraphy and correlation with planktonic foraminifera biostratigraphy is described in the text. Nannofossil biohorizons in bold face. The durations of dysoxic/anoxic episodes illustrated are based on zonal definitions and are longer than those calculated in text which are based on thicknesses of TOC-rich horizons (see text for details).

Subzone NC7A
Interval from the FO of *E. floralis* to the LO of *Micrantholithus hoschulzii.*

Subzone NC7B
Interval from the LO of *M. hoschulzii* to the FO of *Parhabdolithus achlyostaurion.*

Subzone NC7C
Interval from the last occurrence of *P. achlyostaurion* to the first occurrence of *Prediscosphaera columnata.*

Comments: The original definition of Thierstein [1971, 1973] suggested that the first occurrence of *E. floralis* could substitute for the first occurrence of *P. angustus* in marking the base of the *P. angustus* Zone. Since we consistently find *P. angustus* to evolve close to or before the first occurrence of *R. irregularis*, this no longer serves as a suitable marker for the base of this zone. We therefore use the first occurrence of *E. floralis* as suggested previously by Erba [1988] and Applegate and Bergen [1988]. The long *P. angustus* Zone is not an easy interval to subdivide as there is little turnover of nannofossil taxa. Both of the events used to define subzones in this interval have problems. The last occurrence of *M. hoschulzii* (we include *M. obtusus* in this taxon) may have been affected by reworking at Sites 398 and 641 where rare specimens are observed up into the Albian. The first occurrence of *P. achlyostaurion* has been utilized by Erba [1988] as a zonal marker in the middle Albian. This species is extremely rare in the land sections, but we believe that its occurrence in the upper Aptian of Sites 364 and 545 is a distinctive event which should be recognizable elsewhere.

Prediscosphaera columnata Zone (NC8)
Interval from the FO of *P. columnata* to the FO of *Axopodorhabdus albianus* [Roth, 1978].

Subzone NC8A
Interval from the FO of *P. columnata* to the FO of *Hayesites albiensis.*

Subzone NC8B
Interval from the FO of *H. albiensis* to the FO of *Tranolithus orionatus.*

Subzone NC8C
Interval from the FO of *T. orionatus* to the FO of *A. albianus.*
Comments: The first occurrences of *H. albianus* and *T. orionatus* are extremely useful events which can be determined precisely in land and deep-sea sections.

Axopodorhabdus albianus Zone (NC9)
Interval from the FO of *A. albianus* to the FO of *Eiffellithus turriseiffelii* [Roth, 1978].

Subzone NC9A
Interval from the FO of *A. albianus* to the FO of *Eiffellithus* cf. *E. eximius.*

Subzone NC9B
Interval from the FO of *E.* cf. *E. eximius* to the FO of *E. turriseiffelii.*

Comments: The first occurrence of *E.* cf. *E. eximius*, a taxon which has been discussed in some length by Hill and Bralower [1987], can be determined precisely in sections with well-preserved nannofossil assemblages. This taxon is not usually observed in land sections.

Eiffellithus turriseiffelii Zone (NC10)
Interval from the FO of *E. turriseiffelii* to the FO of *Lithraphidites acutum* [Thierstein, 1971; emended by Manivit et al., 1977].

Subzone NC10A
Interval from the FO of *E. turriseiffelii* to the FO of *Corollithion kennedyi.*

Subzone NC10B
Interval from the FO of *C. kennedyi* to the FO of *L. acutum.*

Comments: The first occurrence of *C. kennedyi* has been used to define the Albian/Cenomanian boundary [see discussion in Bralower and Siesser [1992]), however, recent investigations by E. Erba [Pers. Comm., 1992] suggest that this event lies well within the lower Cenomanian. *C. kennedyi* has a cosmopolitan distribution, but is more abundant in high-latitude sites.

The resolution provided by this biostratigraphy varies through time from 1 to 4 m.y. based on nannofossil subzones, 0.5 to 3 m.y. when nannofossil subzones and foraminiferal zones are utilized together, and 0.25 to 2 m.y. based on subsidiary nannofossil biohorizons (Fig. 2). Correlation of zonal and subzonal units with stage and substage divisions is illustrated in Figure 2. These correlations are well determined for nannofossils in the Aptian and for both groups in the Albian, although the exact placement of the Barremian/Aptian and Aptian/Albian boundaries is still in question. Integration of calcareous nannoplankton and planktonic foraminiferal zonations is not without problems and will be the subject of further detailed study. The correlations illustrated are based largely on Sites 545 and 547 for the late Aptian to late Albian interval as well as results of detailed biostratigraphy of the Piobbico Core [Erba, 1988; Tornaghi et al., 1989]. Planktonic foraminiferal zonation schemes in the Barremian-early Aptian interval is in a state of flux (see Leckie and Bralower [1991], Coccioni et al. [in press], and Bralower et al. [submitted] for full discussion). Hence the correlation with nannofossil biostratigraphy and chronostratigraphy for this interval is tentative.

STRATIGRAPHIC AND GEOGRAPHIC DISTRIBUTION OF ORGANIC CARBON-RICH INTERVALS

The sites have been divided into two categories in terms of their TOC contents. In the first, major TOC-rich horizons are interbedded with sediments almost devoid of organic matter. At other sites, however, variable TOC enrichments occur throughout the Aptian-Albian interval. Correlation of the former category can

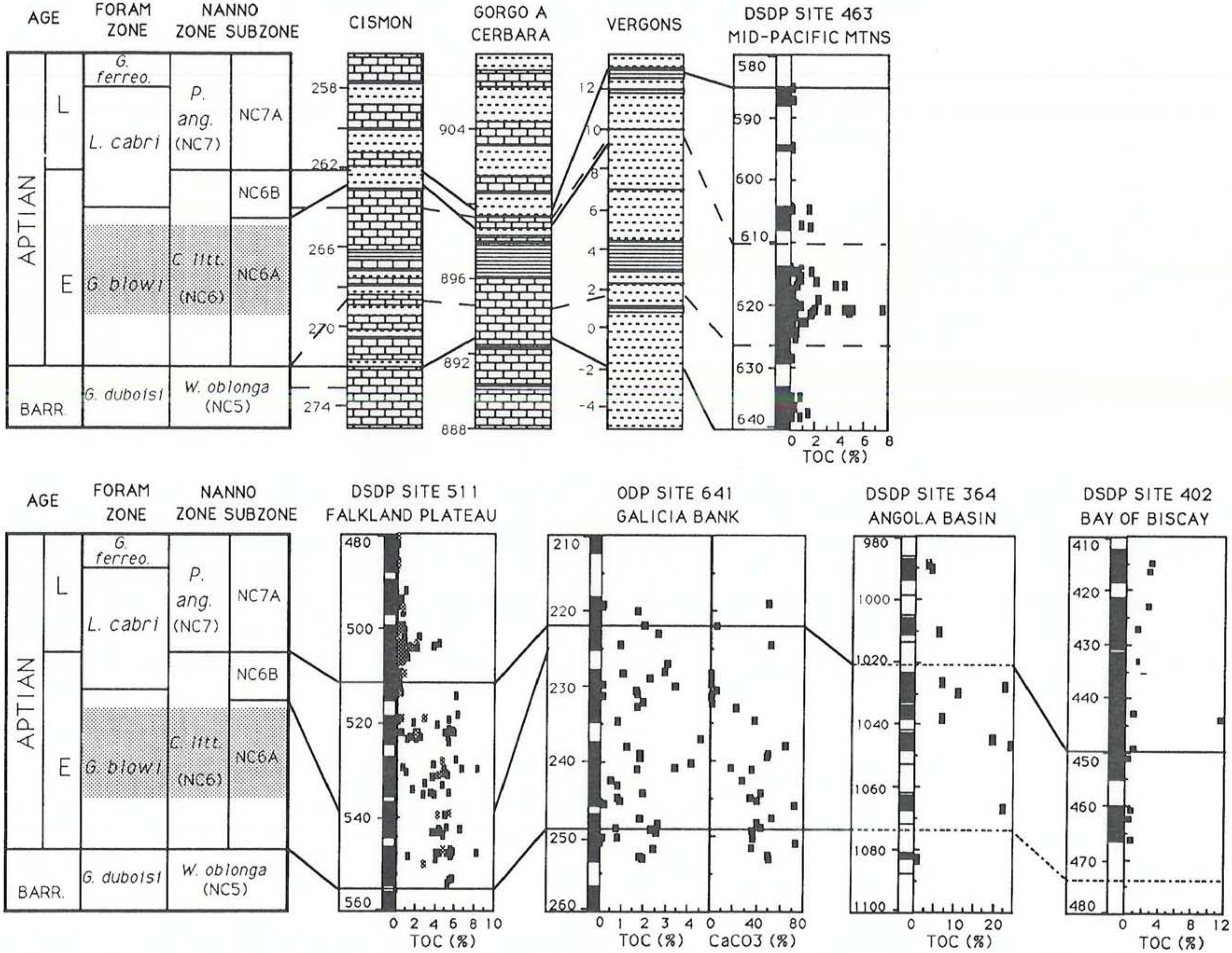


Fig. 3. The early Aptian dysoxic/anoxic episode. Planktonic foraminiferal zonation is from Sliter [1989b], nannofossil zonation is after Bralower et al. [submitted]. Solid thin lines: nannofossil events, dashed lines: foraminiferal events, solid thick lines: unconformities, dotted lines: uncertain event positions. Columns for DSDP/ODP sites indicate recovered (black) and unrecovered (white) intervals. See Bralower et al. [submitted] for detailed lithologic and biostratigraphic range data.

usually be established; this might be quite complicated for the latter type. We presume that enhanced preservation of organic matter in the horizons that we have studied resulted from some combination of oxygen deficiency, high fluxes of organic matter from surface waters, and reduced dilution of organic matter by other sedimentary components. However, there is currently no foolproof technique that allows detection of water column dysoxia/anoxia from the properties of ancient sedimentary rocks. Dysoxic (ie, <0.2 ml/l O_2) or anoxic conditions at or above the sediment/water interface are commonly indicated by the presence of sedimentary laminations [e.g., Rhoads and Morse, 1971; Savrda et al., 1984]. However, because seasonality of sedimentation might be suppressed or dysoxic/anoxic conditions need only extend slightly above the sediment/water interface, presence or absence of lamination is not an absolute indicator of the vertical extent or intensity of dysoxic/anoxic conditions. In several sites (e.g., DSDP Site 370 and ODP Site 641), laminated sediments occur intermittently through the Aptian-Albian interval. In such locations, TOC preservational events are inferred from a combination of discrete peaks in TOC and/or changes in planktonic microfossil assemblages (e.g., assemblage composition, diversity, barren intervals). A more complete discussion of the significance of these biotic indicators is given below.

Our biostratigraphic results indicate that laminated units most enriched in TOC, and sediments characterized by changes in the planktonic assemblages cluster in three narrow biostratigraphic intervals (Fig. 2): (1) the early Aptian *Globigerinelloides blowi* foraminiferal Zone, *Chiastozygus litterarius* nannofossil Zone, *Conusphaera rothii* Subzone (NC6A; Fig. 3); (2) the early Albian *Hedbergella planispira* foraminiferal Zone, *Prediscosphaera columnata* nannofossil Zone, subzone NC8B (Fig. 4), and (3) the early late Albian *Biticinella breggiensis* foraminiferal Zone, *Axopodorhabdus albianus* nannofossil Zone, subzone NC9B (Fig. 5). Using the terminology of Arthur et al. [1990], we term these subevents OAE 1a, 1b and 1c, respectively. It should be noted that Arthur et al. [1990] assigned a late late Albian age for OAE 1c. The proposed subevents correspond to lithologically-distinctive and TOC-rich intervals in pelagic/hemipelagic sections exposed on land including the lower Aptian "Livello Selli" of the Umbrian Apennines and Southern Alps of Italy, the lower Albian "Niveau Paquier" of the Fosse Vocontienne of France and the lower upper Albian Toolebuc Formation of the Eromanga basin, Queensland, Australia (Figs. 3-5). These horizons correlate well with carbonaceous units in the deep sea (DSDP/ODP sites), strongly increasing the significance of the three proposed subevents.

The lower Aptian "Livello Selli" at Cismon (Southern Alps) and several sections in the Italian Apennines correlates with the "Niveau Goguel" in the Fosse Vocontienne. These horizons lie within the *G. blowi* foraminiferal Zone which is defined based on the presence of the nominate species, but absence of *Leupoldina cabri*, the *C. litterarius* nannofossil Zone which is here defined as the interval between the first occurrences of *Rucinolithus irregularis* and *Eprolithus floralis*, and the *Conusphaera rothii* nannofossil Subzone (NC6A) defined as the interval between the first occurrence of *R. irregularis* and the last occurrence of *C. rothii* [Bralower et al., submitted; Figs. 2,3]. These horizons also correlate at the

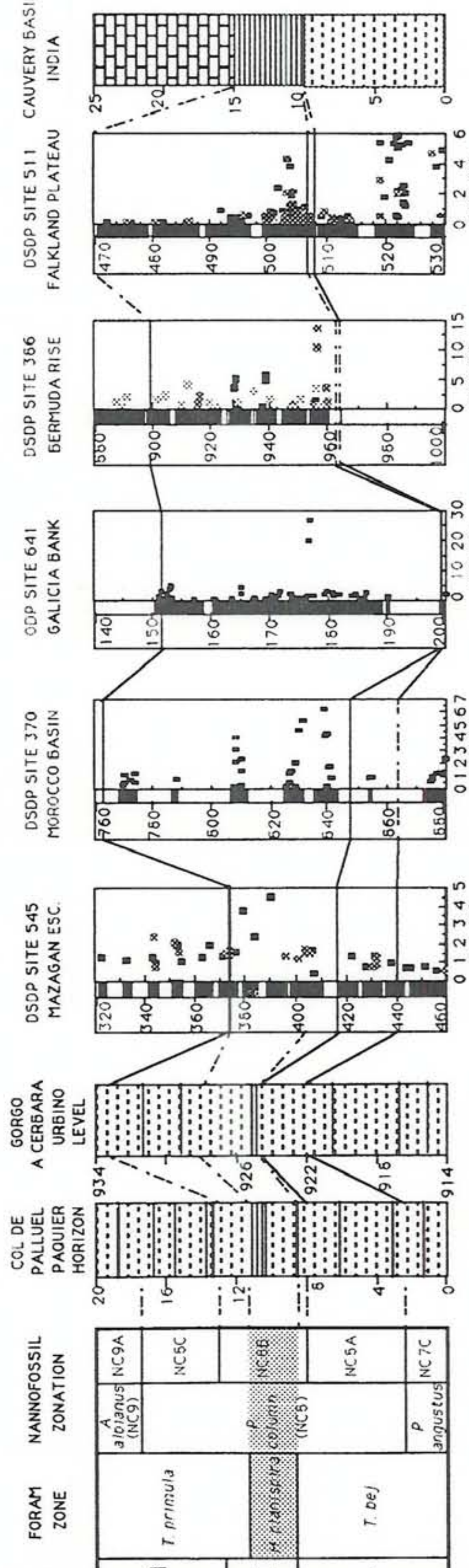
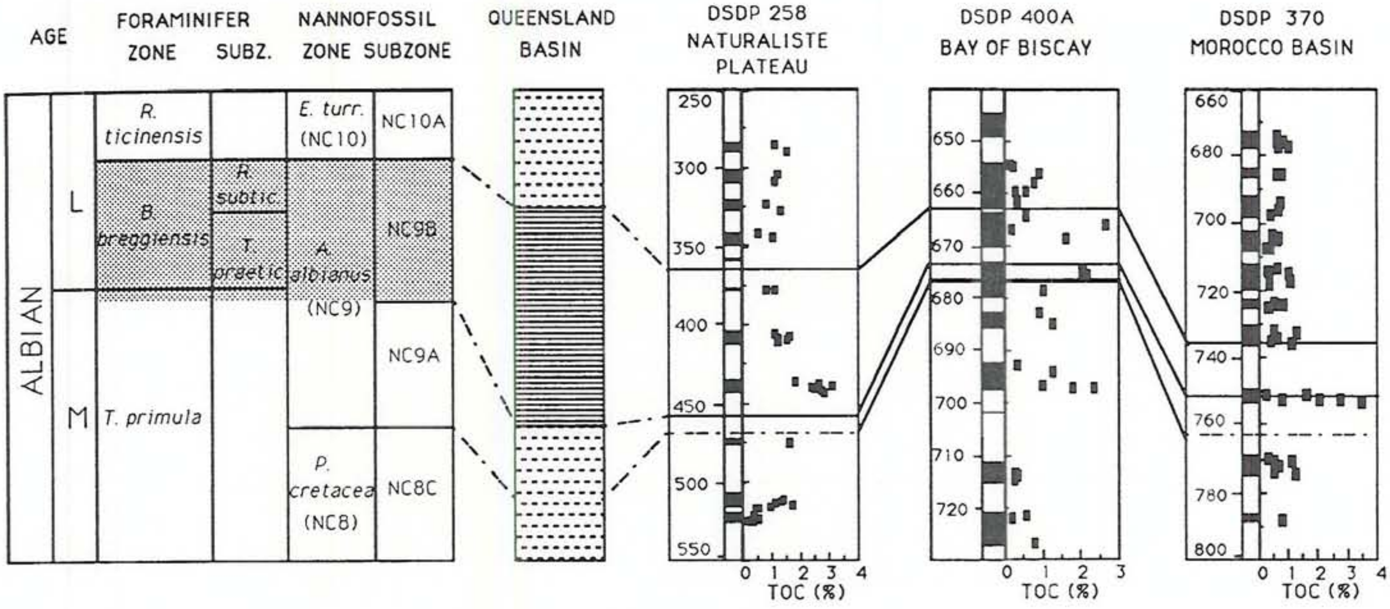


Fig. 4. The early Albian dysoxic/anoxic episode. Planktonic foraminiferal zonation is from Sliter [1989b], nannofossil zonation is described in this paper. Lines/columns as in Fig. 3. See Figs. 6-13 for more detailed biostratigraphic/lithostratigraphic data for each section.



KEY TO FIG.S 5 - 16

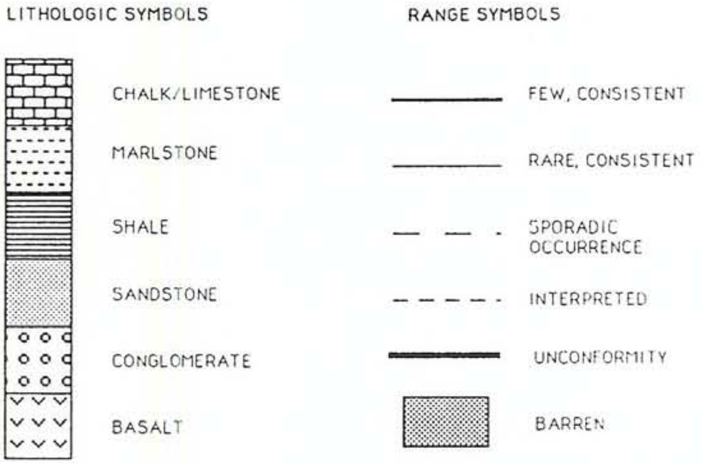
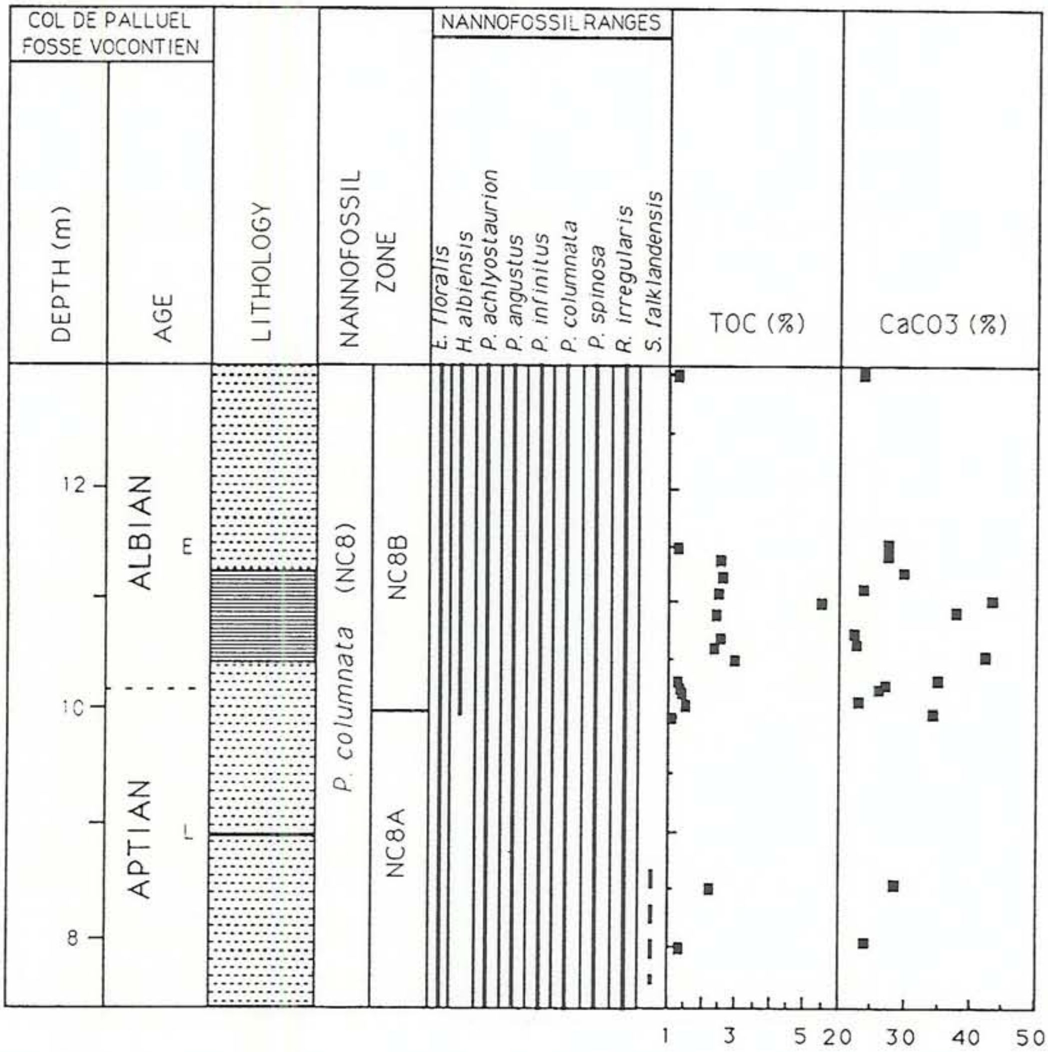


Fig. 5. The late middle Albian dysoxic/anoxic episode. Planktonic foraminiferal zonation is from Sliter [1989b], nannofossil zonation is described in this paper. Lines/columns as in Fig. 3. See Figs. 14-17 for more detailed biostratigraphic/lithostratigraphic data for each section.

nannofossil subzonal level with carbonaceous limestones (up to 8% TOC) in Cores 70 and 71 of DSDP Site 463 in the Mid-Pacific Mountains (Fig. 3), and at the zonal level with black shales in Cores 42 to 45 of DSDP Site 364 in the Angola Basin, South Atlantic (up to 24% TOC). TOC contents are variable throughout the *C. litterarius* Zone at ODP Site 398 drilled on Vigo Seamount, but the correlative subevent (interval barren of microfossils in Cores 398D-118 to -120) occurs at the top of an interval of higher TOC and lies just above the *C. rothii* Subzone (see discussion in Bralower et al. [submitted]). The age of TOC-rich sediments in Cores 60-62 of DSDP Site 511 on the Falkland Plateau is not easily established as the low diversity nannoflora lacks zonal marker species such as *R. irregularis* and planktonic foraminifera are absent. However, we place these sediments within and slightly above the *C. rothii* nannofossil Subzone based on the last occurrence of *C. rothii* in Section 61-2 and the absence of *E. floralis* throughout this interval (Fig. 3). TOC contents are also highly variable throughout the *C. litterarius* Zone at ODP Site 641 drilled on Galicia Bank with no identifiable peak associated with the interval barren of microfossils (Sections 641C-9R-1 and -2, 228-230 meters below sea floor; Fig. 3) that is thought to represent OAE 1a. This interval occurs within the *C. rothii* nannofossil Subzone and some 10 m above magnetic polarity zone M0 [Ogg, 1988] in a similar position to organic carbon-rich horizons at DSDP Site 463 and Italian land sections.

The "Niveau Paquier" of the Col de Palluel Section, Fosse Vocontienne, France, lies within the early Albian *H. planispira* foraminiferal Zone based on the presence of a low-diversity



KEY TO FIG.S 5 - 16

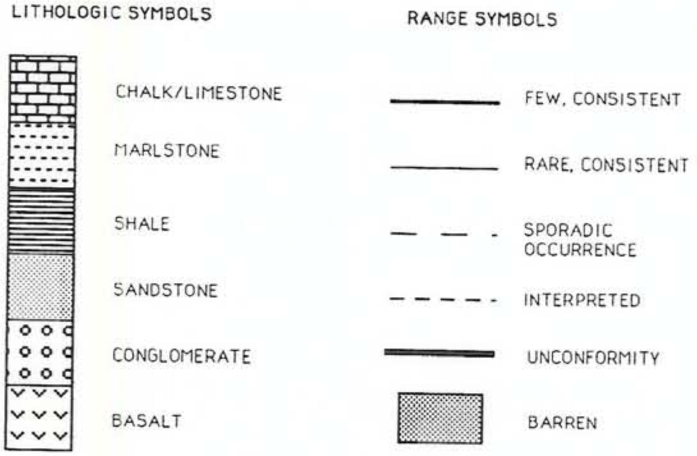
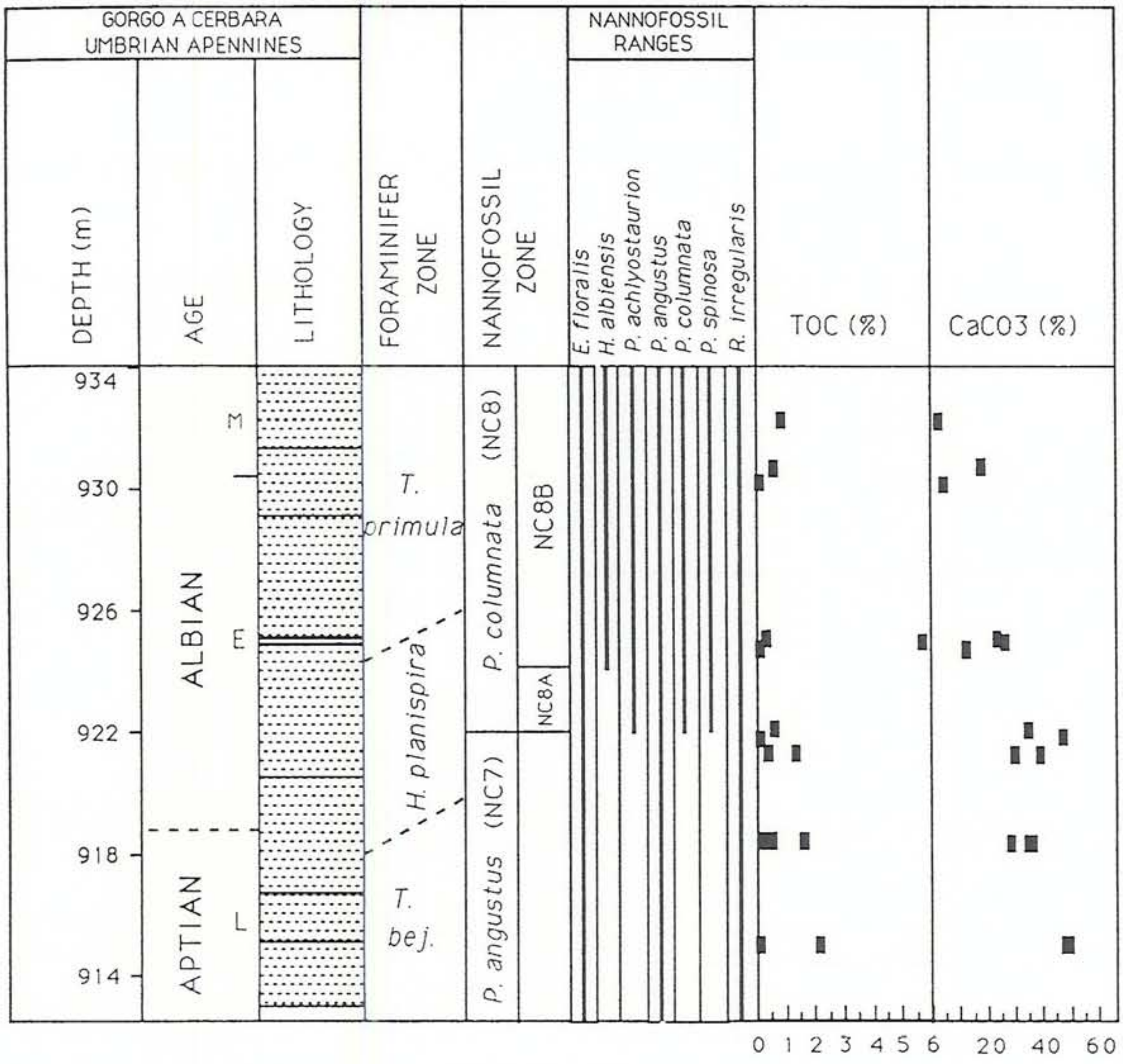


Fig. 6. Upper Aptian-lower Albian stratigraphy of the Col de Palluel Section, Fosse Vocontienne, France. Depths are from an arbitrary sampling scheme. Lithologic column shows position of the organic-rich "Niveau Paquier". Nannofossil biostratigraphy is part of this investigation. See key for explanation of lithologic and range symbols. Position of stage boundary is from nannofossil biostratigraphy. Organic carbon and carbonate contents are from this investigation.



KEY TO FIG.S 5 - 16

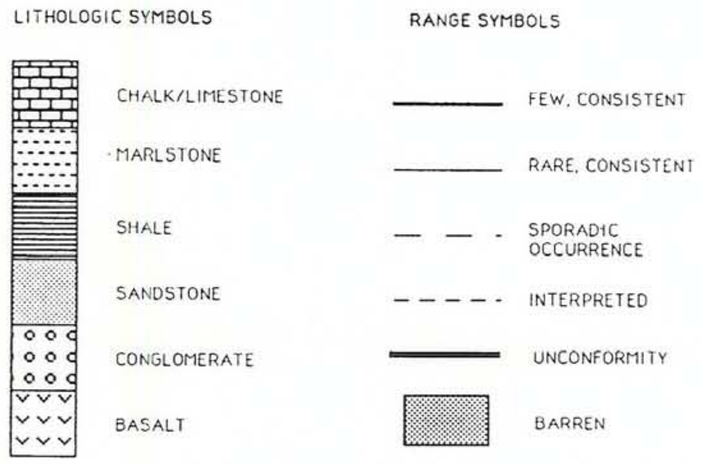
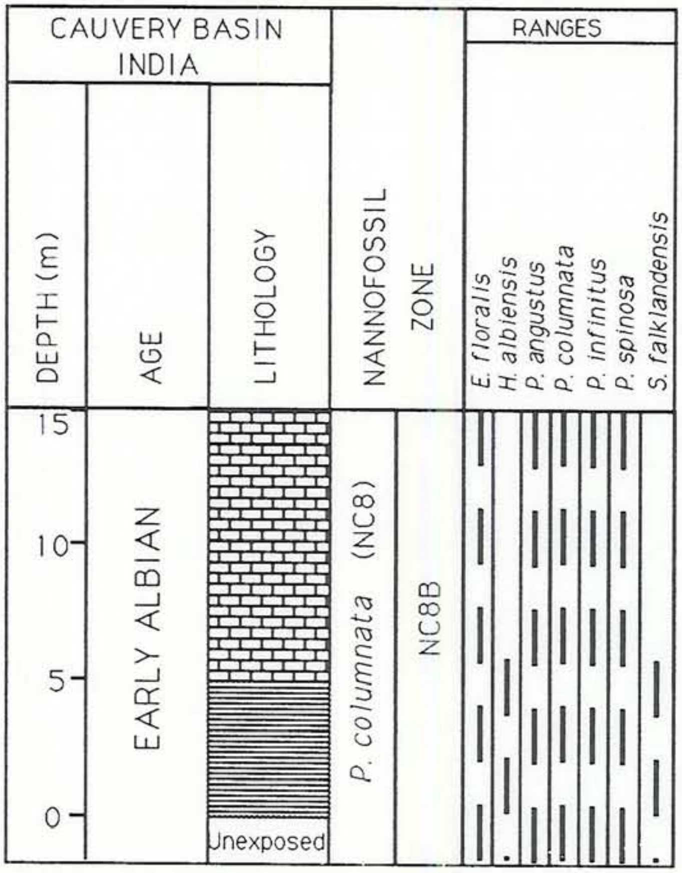


Fig. 7. Upper Aptian-middle Albian stratigraphy of the Gorgo a Cerbara Section, Umbrian Apennines, Italy. Depths are from sampling scheme which is an extension of that of Lowrie and Alvarez [1984]. Lithologic column shows position of the organic-rich "Livello Urbino" at 924.5m. Foraminifera and nannofossil biostratigraphy are part of this investigation. See key for explanation of lithologic and range symbols. Positions of stage and substage boundaries are from nannofossil biostratigraphy. Organic carbon and carbonate contents are from this investigation.



KEY TO FIG.S 5 - 16

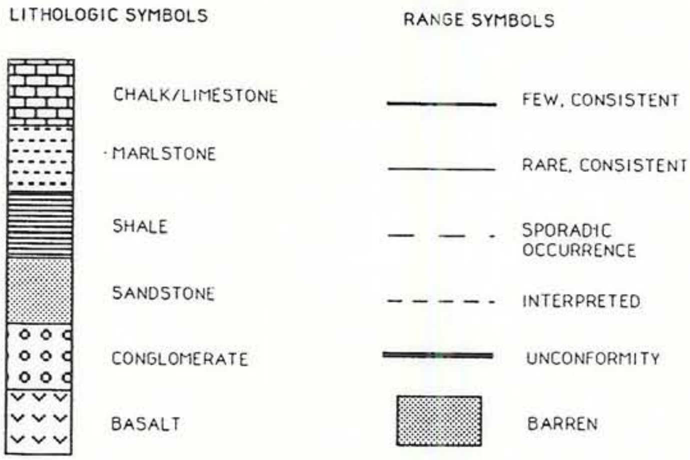


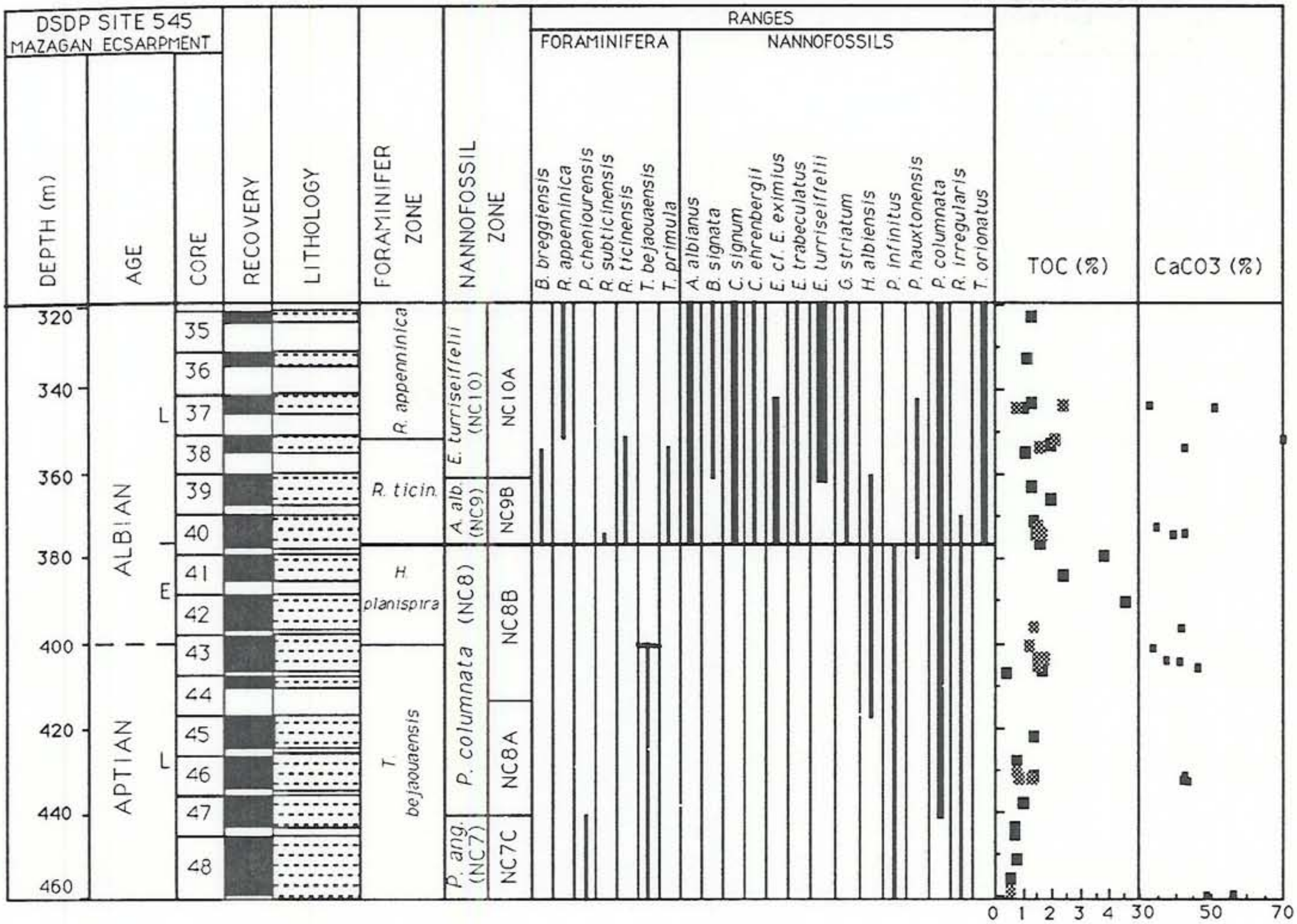
Fig. 8. Lower Albian stratigraphy of the Dalmiapuram Grey Shale, Cauvery Basin, India. Depths are an arbitrary scheme used in this investigation. Illustrated is generalized lithology. Nannofossil biostratigraphy was obtained as part of this investigation. See key for explanation of lithologic and range symbols.

assemblage of hedbergellids, and the absence of *Ticinella primula*, and nannofossil subzone NC8B based on the occurrence of *P. columnata* and *Hayesites albiensis*, and the absence of *Tranolithus orionatus* (Figs. 2, 4, 6). Based on identical nannofossil biostratigraphic criteria, the "Niveau Paquier" correlates with the "Livello Urbino" in the Gorgo a Cerbara (Fig. 7) and other sections from the Italian Apennines. Our planktonic foraminifera biostratigraphy does not disagree with this correlation, although because of the lack of limestones and the corresponding meagre number of samples analyzed in this interval, we cannot differentiate between the *H. planispira* and the overlying *T. primula* Zones (Fig. 7). An unresolved problem exists in that Tornaghi et al. [1989] have correlated the "Livello Urbino" not with the *H. planispira* foraminiferal Zone, but with the underlying *T. bejaouaensis* Zone. The nannofossil biostratigraphy carried out here would correlate the TOC-rich Dalmiapuram shales of the Cauvery Basin, India [Govinden, 1982] with the "Niveau Paquier" based on the presence of *H. albiensis*, but absences of *T. orionatus* and *A. albianus* in our samples (Fig. 8). However, Kale and Phansalkar [1992] have reported the latter two species from this unit and would thus correlate it with the early late Albian episode (based on the absences of *E.* cf. *E. eximius* and *E. turriseiffelii*).

Based on nannofossil biostratigraphy, these units exposed on land correlate with intervals which are not entirely TOC-rich, but which do possess individual TOC-rich horizons, at DSDP Site 545 off the Mazagan Escarpment (Fig. 9; up to 5% TOC), ODP Site 641 on Galicia Bank (Fig. 10; up to 27% TOC), DSDP Site 511 on the Falkland Plateau (Fig. 11; up to 4% TOC) and DSDP Site 386 on Bermuda Rise (Fig. 12; up to 14% TOC). This correlation is very clear at Sites 545 and 641 where the ranges of *P. columnata*, *H. albiensis* and *T. orionatus* have been established precisely, but slightly more difficult at Site 386, where long intervals are barren of calcareous microfossils. However, our data combined with those of Okada and Thierstein [1979] allow limits to be placed for the boundaries of subzone NC8B (Fig. 12). At Site 511, there is a general paucity of traditional marker species, however, a peak in TOC contents, and a microfossil-poor interval in Core 57 lies in nannofossil subzone NC8B based on the combined occurrence of *Prediscosphaera spinosa* (but not *P. columnata* [see Bralower, 1992]) and *H. albiensis* in a sample from Section 57-6 (Fig. 11). Poor sediment recovery precludes an in-depth study of DSDP Site 370; however, higher TOC contents (up to 7%) are observed in subzone NC8B than in NC8A (Fig. 13).

Foraminiferal biostratigraphy provides age constraints in some of the deep-sea sites. The peak in TOC contents at Site 545 lies within the *H. planispira* Zone (Fig. 9). This zone cannot be differentiated from the *T. bejaouaensis* Zone at Site 511 and the boundary between these combined units and the underlying *H. gorbachikae* Zone cannot be defined exactly based on the paucity of low latitude marker species. The peak in TOC in Core 511-57 lies within this transitional interval (Fig. 11). Although absence of markers inhibits zonal assignment, the interval proposed to correlate to OAE1b at Site 370 based on nannofossil biostratigraphy also appears to be early Albian in age based on the foraminifera *G. blowi* and "*G.*" *gyroidinaeformis* in Section 30-4 (Fig. 13). Cores 27-29 and 31 are largely barren of planktonic foraminifera.

The lower upper Albian carbonaceous Toolebuc Formation, Eromanga Basin, Queensland, Australia recovered in the Augathella Borehole lies in nannofossil subzone NC9B based on the combined occurrence of *Axopodorhabdus albianus* and *Eiffellithus* cf. *E.*



KEY TO FIG.S 5 - 16

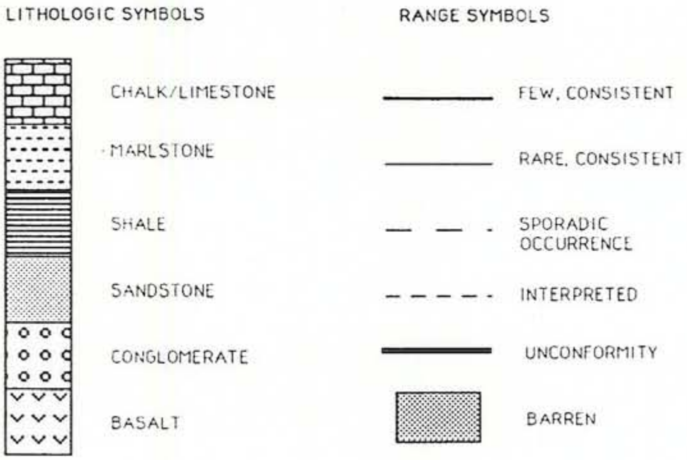
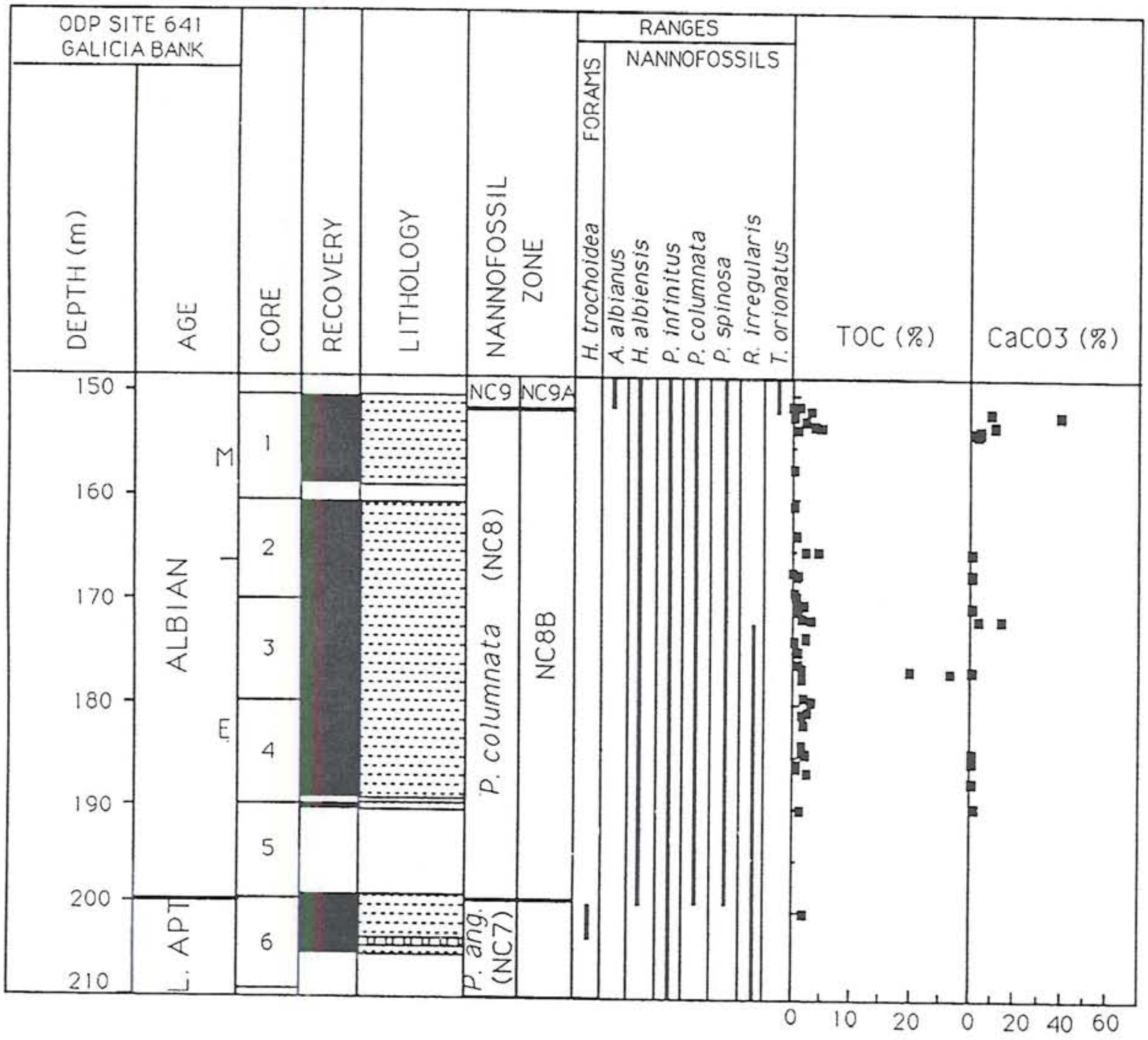


Fig. 9. Upper Aptian-upper Albian stratigraphy of DSDP Site 545, Mazagan Escarpment. Depths are in meters below sea floor. Illustrated are core recovery and generalized lithology. Foraminifera biostratigraphy is modified after Leckie [1984]. Recovery column shows unrecovered intervals in white. Thick horizontal line indicates position of unconformity. Nannofossil biostratigraphy was obtained as part of this investigation. See key for explanation of lithologic and range symbols. Positions of stage and substage boundaries are from nannofossil biostratigraphy. Organic carbon and carbonate contents are from Hinz et al. [1984; grey symbols] and Deroo et al. [1984; black symbols].



KEY TO FIG.S 5 - 16

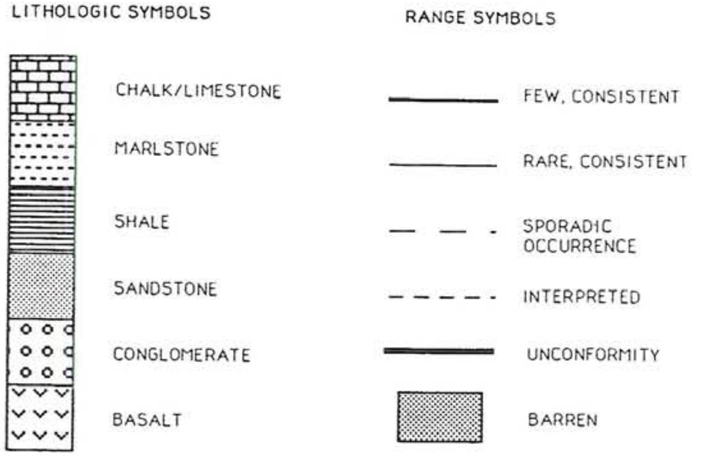
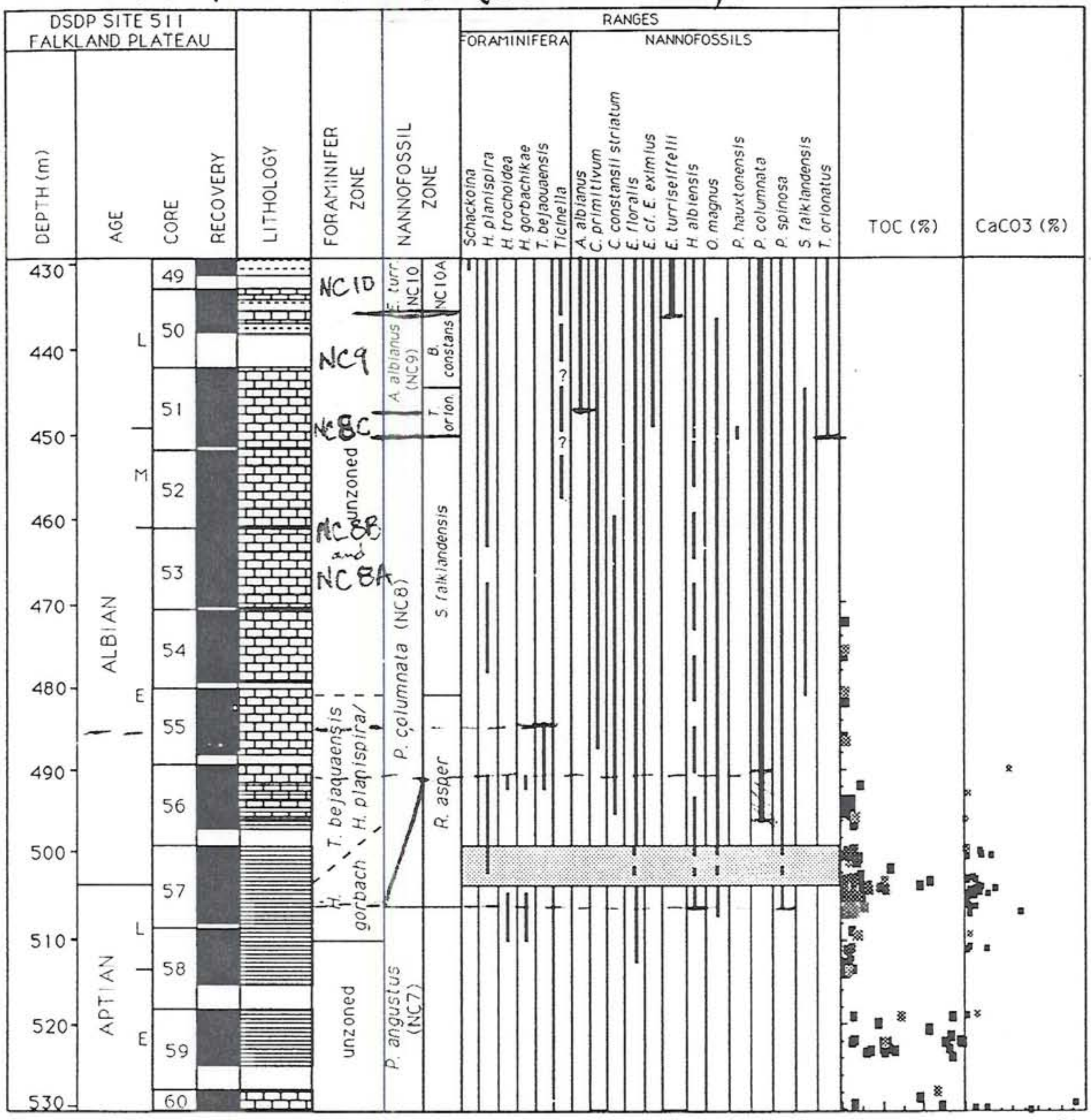


Fig. 10. Upper Aptian-upper Albian stratigraphy of ODP Site 641, Galicia Bank. Depths are in meters below sea floor. Illustrated are core recovery and generalized lithology. Foraminifera and nannofossil biostratigraphy were obtained as part of this investigation. See key for explanation of lithologic and range symbols. Positions of stage and substage boundaries are from nannofossil biostratigraphy. Organic carbon and carbonate contents are from Boillot et al. [1988] and this study.

base of NC8 depends if you believe P. columnata or P. spinosa
Bralower had more confidence in spinosa (because of TOC
spike in Core 57?)
also note v. low FO of albiensis (base of NC8B)

bejaouaensis
55-6



KEY TO FIG.S 5 - 16

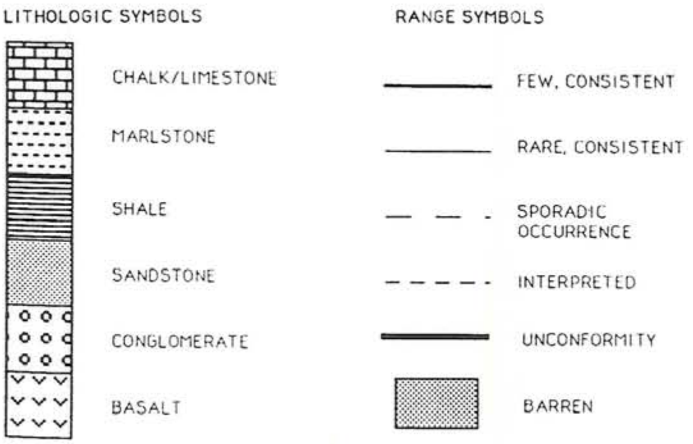
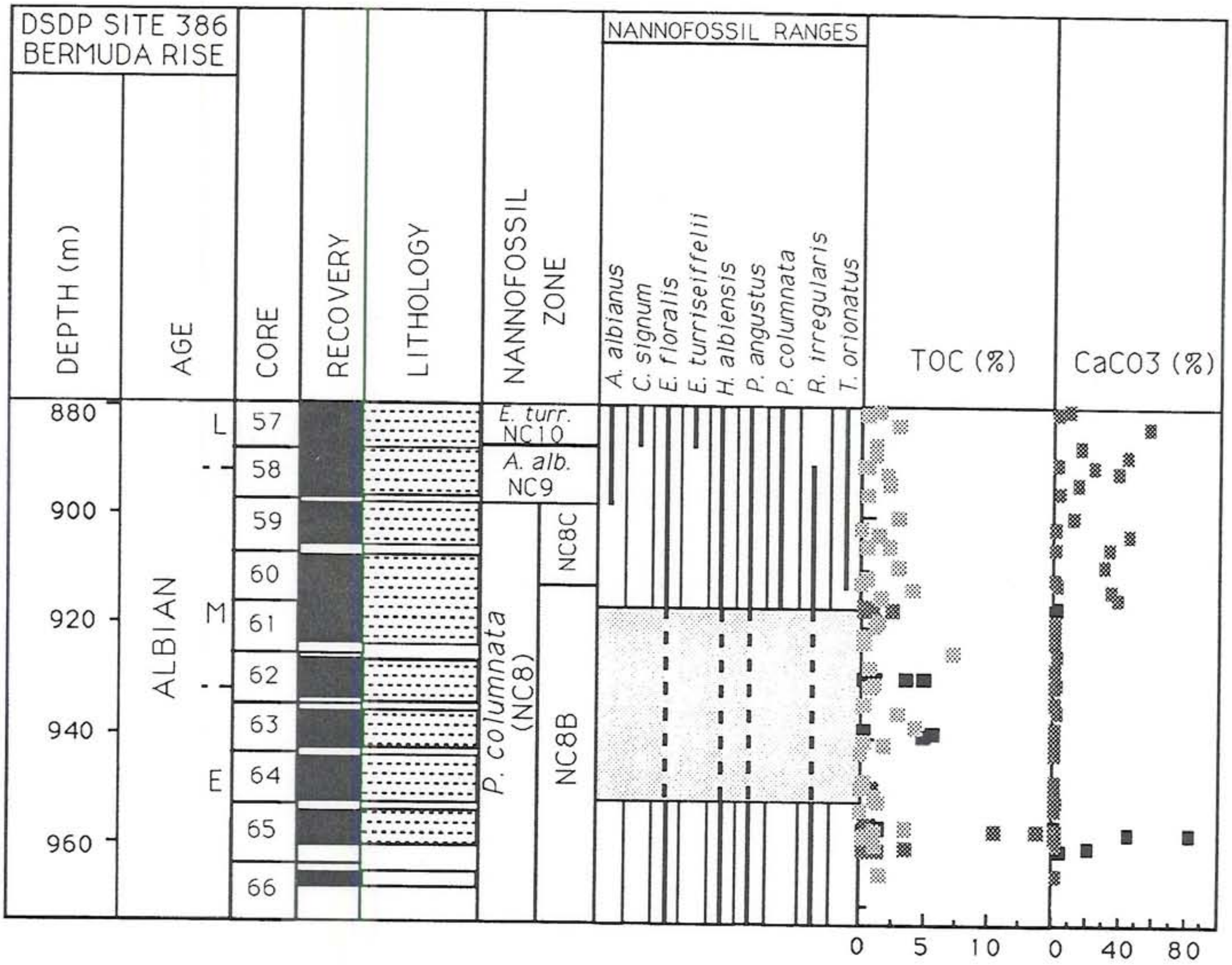


Fig. 11. Aptian-Albian stratigraphy of DSDP Site 511, Falkland Plateau. Depths are in meters below sea floor. Illustrated are core recovery and generalized lithology. Foraminifera biostratigraphy is modified after Krasheninnikov and Basov [1983]. Nannofossil biostratigraphy was combined from Wise [1983] and this investigation. Subzones shown are from the high latitude scheme of Wise [1983]. See key for explanation of lithologic and range symbols. Dotted range area indicates depths of unfossiliferous section. Positions of stage and substage boundaries are from nannofossil biostratigraphy. Organic carbon and carbonate contents are from Deroo et al. [1983], Bode [1983], Parker et al. [1983] (all grey symbols)] and this investigation (black symbols).



KEY TO FIG.S 5 - 16

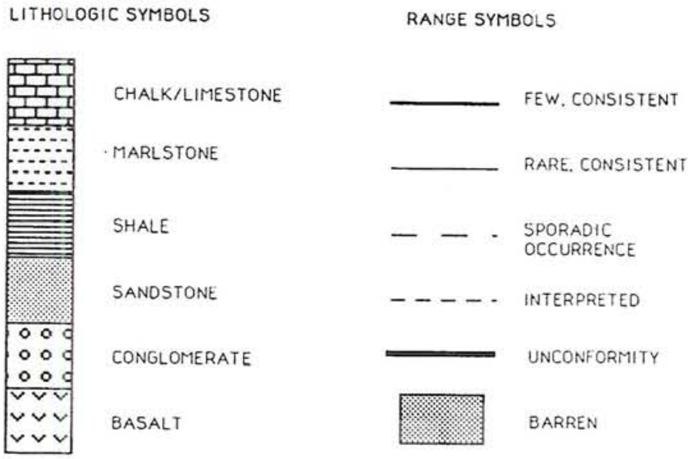
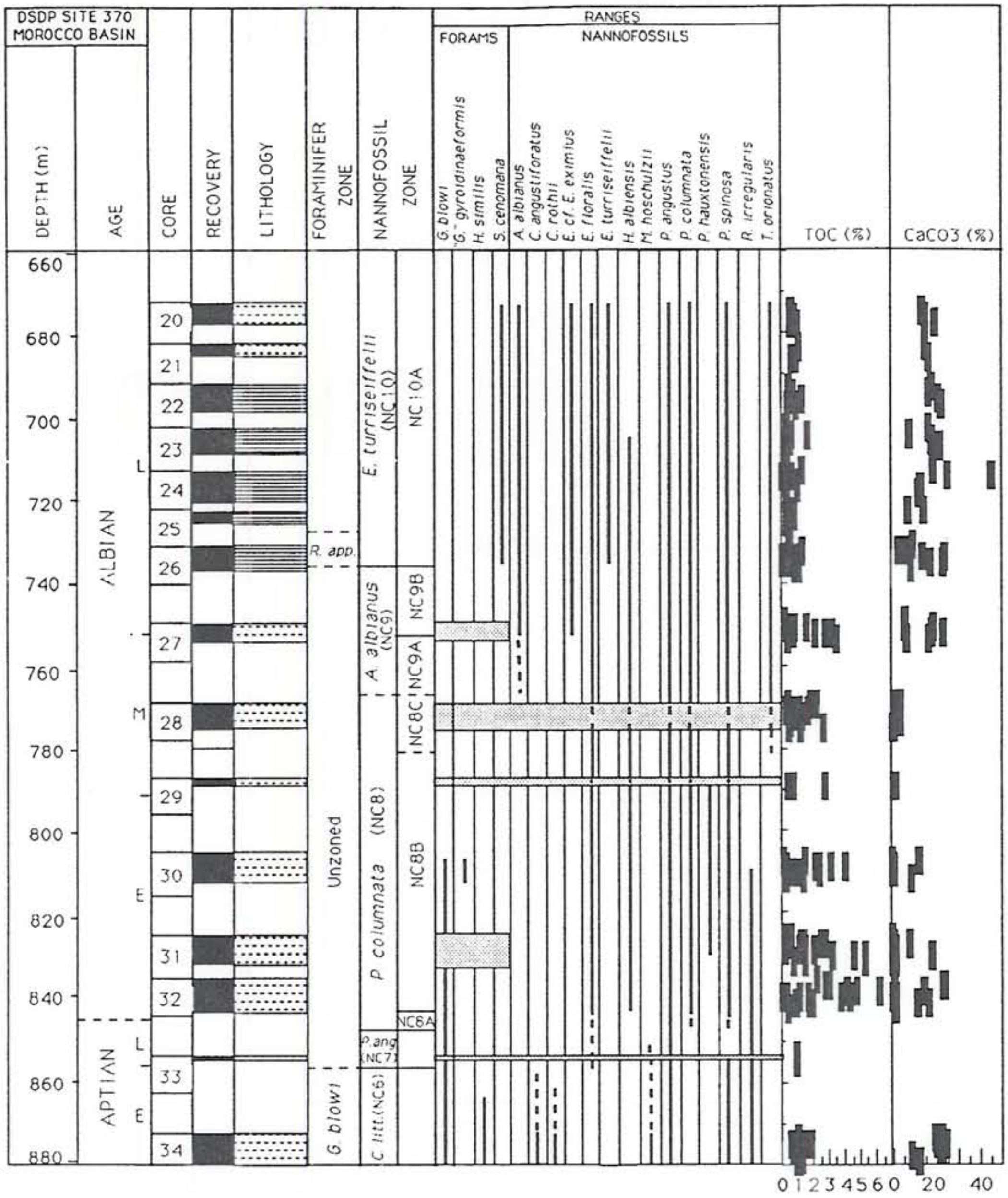


Fig. 12. Albian stratigraphy of DSDP Site 386, Bermuda Rise. Depths are in meters below sea floor. Illustrated are core recovery and generalized lithology. Nannofossil biostratigraphy was combined from Okada and Thierstein [1979] and the results of this investigation. See key for explanation of lithologic and range symbols. Dotted range area indicates depths of unfossiliferous section. Positions of substage boundaries are from nannofossil biostratigraphy. Organic carbon and carbonate contents are from Cameron [1979; light grey symbols] and Bralower [1984; dark grey and black symbols].



KEY TO FIG.S 5 - 16

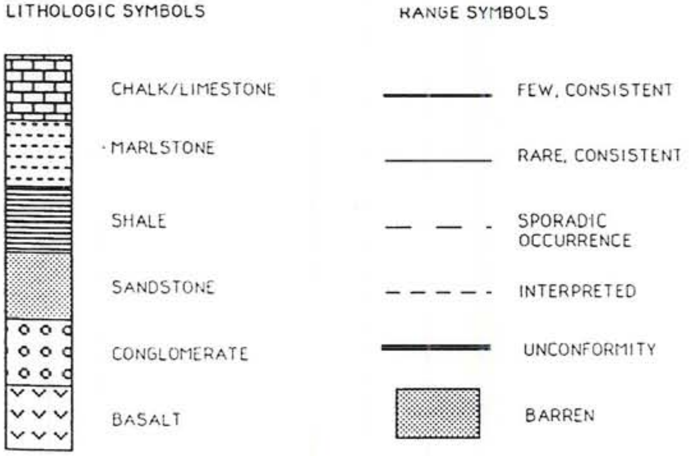
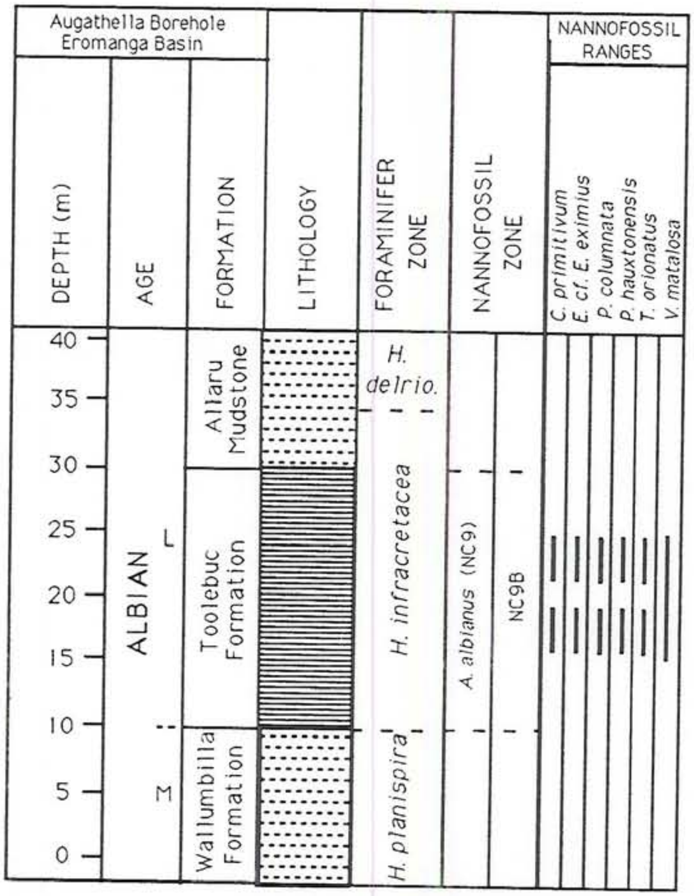


Fig. 13. Lower Aptian-upper Albian stratigraphy of DSDP Site 370, Morocco Basin. Depths are in meters below sea floor. Illustrated are core recovery and generalized lithology. Foraminifera biostratigraphy is modified after Pflaumann and Krashenninikov [1977]. Nannofossil biostratigraphy was obtained as part of this investigation. See key for explanation of lithologic and range symbols. Positions of stage and substage boundaries are from nannofossil biostratigraphy. Organic carbon and carbonate contents are from Erdman and Schomo [1977] and this study.



KEY TO FIG.S 5 - 16

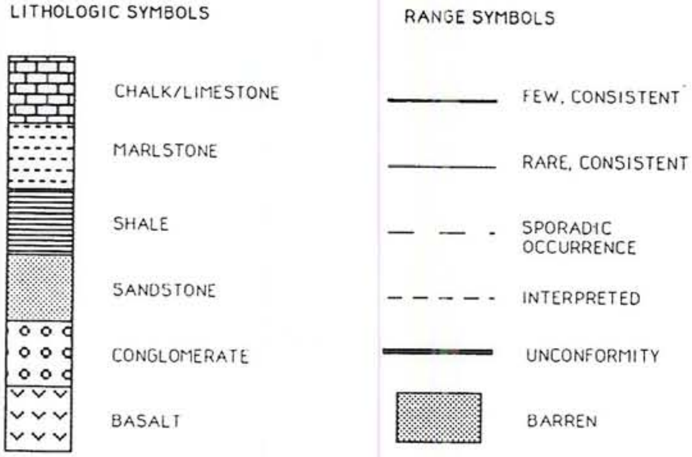


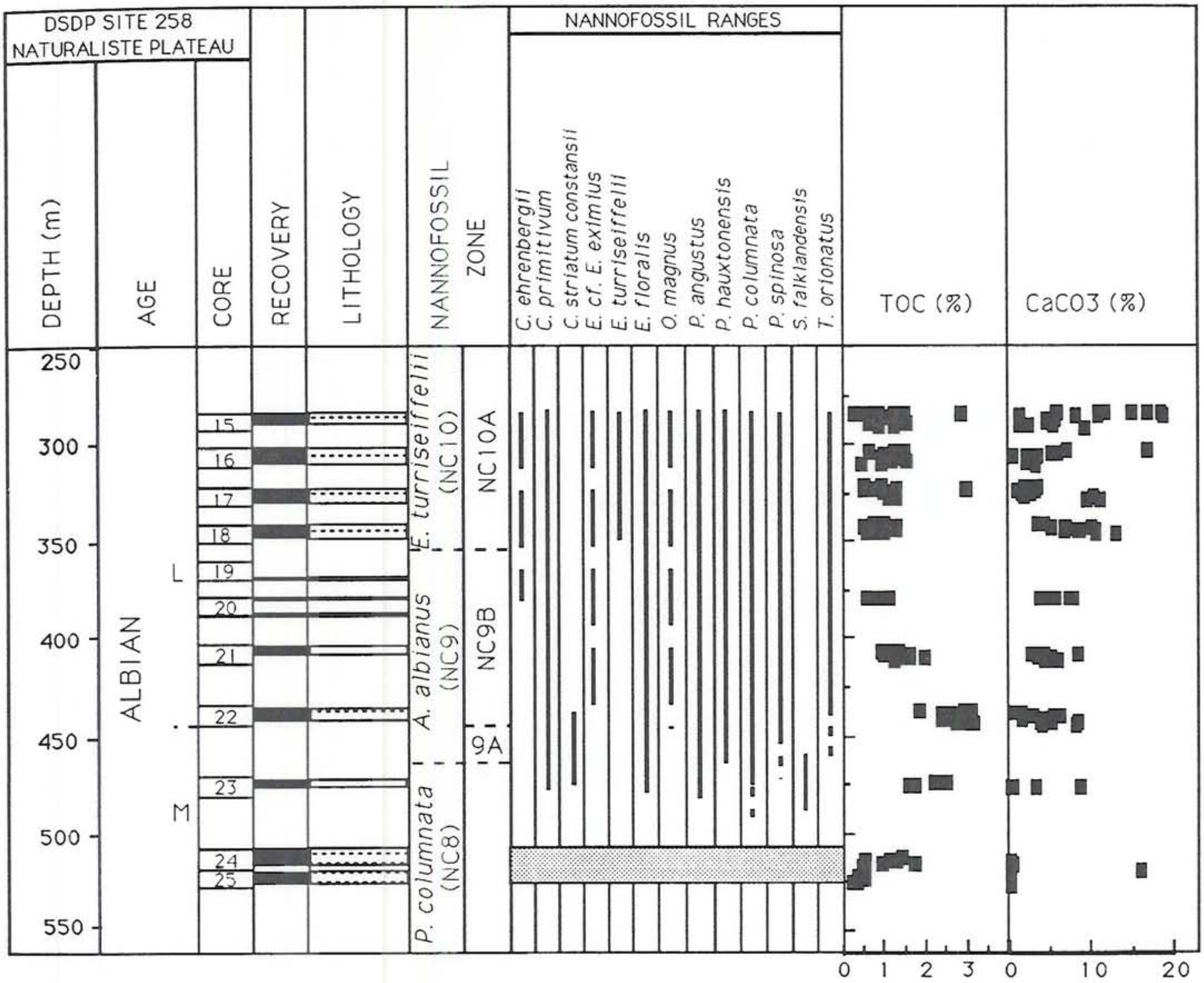
Fig. 14. Middle and upper Albian stratigraphy of Augathella Borehole, Eromanga Basin, Australia. Depths are an arbitrary scheme used in this investigation. Generalized lithology is illustrated. Foraminifera biostratigraphy is modified after Haig [1979]. Nannofossil biostratigraphy was combined from Shafik [1985] and this investigation. See key for explanation of lithologic and range symbols. Positions of substage boundaries are from nannofossil biostratigraphy.

eximius, but absence of *E. turriseiffelii* (Figs. 2, 14; see also Shafik, 1985). It is not possible to define exact zonal and subzonal boundaries in this section as the five fossiliferous samples observed all lie within the same subzone. Correlation of the Toolebuc with peaks in TOC contents in deep-sea sites is less certain than for the other subevents owing to poor recovery in most sections studied (Fig. 5). However, it is clear that this interval corresponds with peaks in TOC contents at DSDP Site 370 from the Morocco Basin (Fig. 13), Site 258 from the Naturaliste Plateau, Indian Ocean (Fig. 15) and Site 400 from the Bay of Biscay (Fig. 16). The TOC peak at Site 258 lies within Core 22, an interval lithologically identical to adjacent cores (Fig. 15). More significant than the absolute magnitude of the TOC peaks at Sites 370 and 400 (Figs. 13 and 16) is the similar decrease in TOC at the base of the overlying *E. turriseiffelii* Zone.

Foraminiferal biostratigraphy provides little information on the timing of proposed OAE1c as the sites investigated are either barren of foraminifera in the late Albian interval (Site 258) or contain a low diversity fauna which lacks diagnostic marker species (Toolebuc Formation; Haig [1979]). The late Albian peak of TOC at Site 400 lies within an interval in the transition between the *T. primula/H. planispira* and the basal *B. breggiensis* Zones. The relevant sediments at Site 370 are largely barren of planktonic foraminifera, however the assemblage in Section 26-4 appears to be latest Albian in age based on the occurrence of *Schackoina cenomana* and *Guembelitria* spp. indicating a possible unconformity between Cores 26 and 27. This assignment in not inconsistent with the correlation of OAE1c at this site.

Results of other studies confirm the identification of these three dysoxic/anoxic episodes. The lower Aptian TOC-rich Fischschiefer of northern Germany is a likely equivalent of OAE 1a [Gaida et al., 1981]. TOC contents are high throughout the Aptian interval of DSDP Site 361 in the Cape Basin, South Atlantic, but reach a peak of 25% in the *Chiastozygus litterarius* nannofossil Zone [Proto Decima et al. [1978]; Herbin et al. [1986, 1987]. Poor nannofossil preservation does not allow subzonal subdivision of this section. Koutsoukos et al. [1991] identified a peak in TOC contents in the lower Albian of several wells from the Sergipe Basin, offshore Brazil. Although no precise biostratigraphic control was cited, these sediments are possible equivalents of OAE 1b. The lower Albian bituminous Pariatambo Formation of Peru (*Douvilleiceras mammilatum* and *Hoplites dentatus* ammonite Zones; Benavides-Caceres [1956]; Jaillard [1987]) also is a likely equivalent. TOC-rich black shales of the Skull Creek Formation of the U.S. Western Interior (*Inoceramus bellvuensis* and *I. comanchaeanis* Zones of the early late Albian; Kauffman et al., in press) are likely equivalents of OAE 1c.

The occurrence of TOC-rich horizons in the three proposed dysoxic/anoxic intervals is by no means ubiquitous at all of the sites investigated. Figures 17a-c illustrate the geographic distribution of all sites investigated for the three intervals of interest. These figures differentiate, for each proposed subevent, sites at which TOC-rich sediments exist, or which show diagnostic changes in microfossil assemblages, and those which possess no evidence for dysoxia/anoxia. For these same sites, Figure 17d shows occurrences of TOC-rich sediments which lie in other calcareous nannofossil subzones. Certain localities are characterized by high TOC at times other than the proposed episodes. For example, the Aptian-Albian section in the Umbrian Apennines contains numerous other black shale horizons beside the "Livelli Selli" and "Urbino",



KEY TO FIG.S 5 - 16

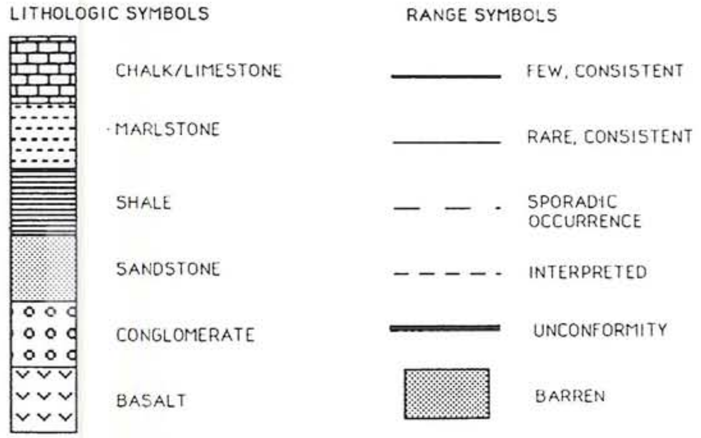
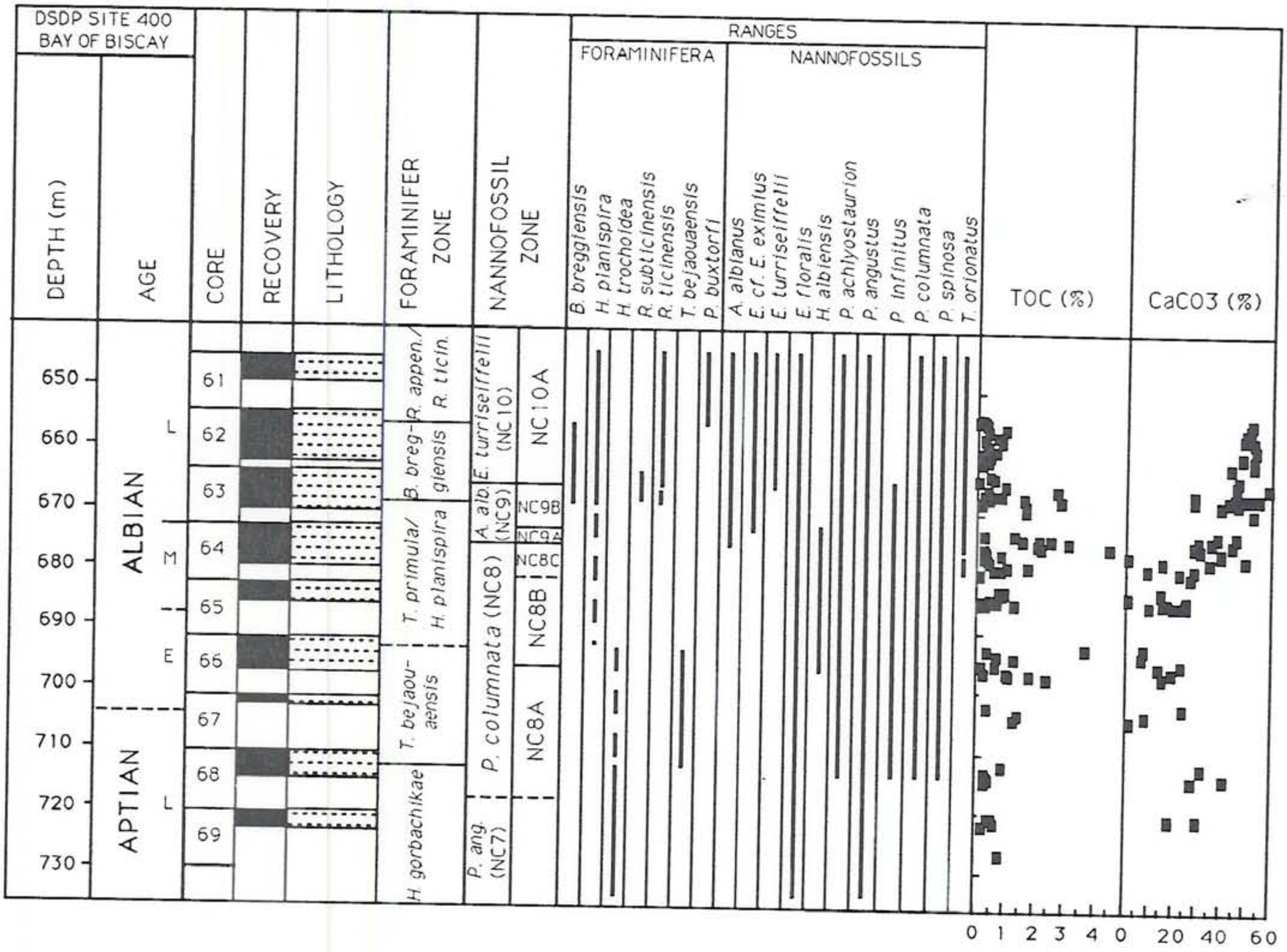


Fig. 15. Middle and upper Albian stratigraphy of DSDP Site 258, Naturaliste Plateau. Depths are in meters below sea floor. Illustrated are core recovery and generalized lithology. Nannofossil biostratigraphy was combined from Thierstein [1974] and this investigation. See key for explanation of lithologic and range symbols. Positions of substage boundaries are from nannofossil biostratigraphy. Organic carbon and carbonate contents are from Davies et al. [1974] and this investigation..



KEY TO FIG.S 5 - 16

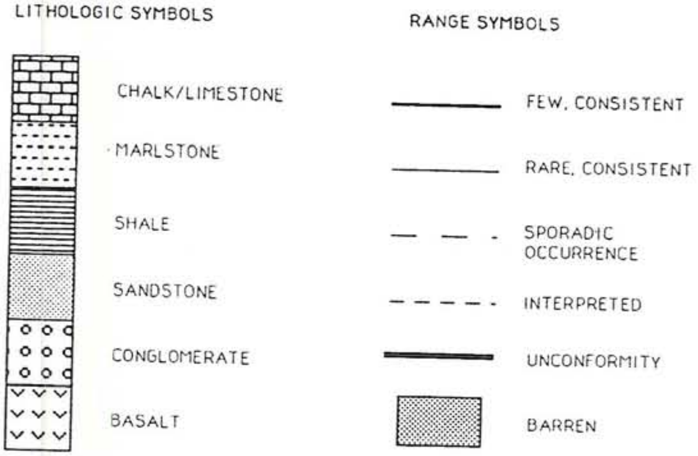


Fig. 16. Middle and upper Albian stratigraphy of DSDP Site 400, Bay of Biscay. Depths are in meters below sea floor. Illustrated are core recovery and generalized lithology. Foraminifera biostratigraphy is modified after Depeuble [1979]. Nannofossil biostratigraphy was obtained as part of this investigation. See key for explanation of lithologic and range symbols. Positions of substage boundaries are from nannofossil biostratigraphy. Organic carbon and carbonate contents are from Deroo et al. [1979] and this investigation.

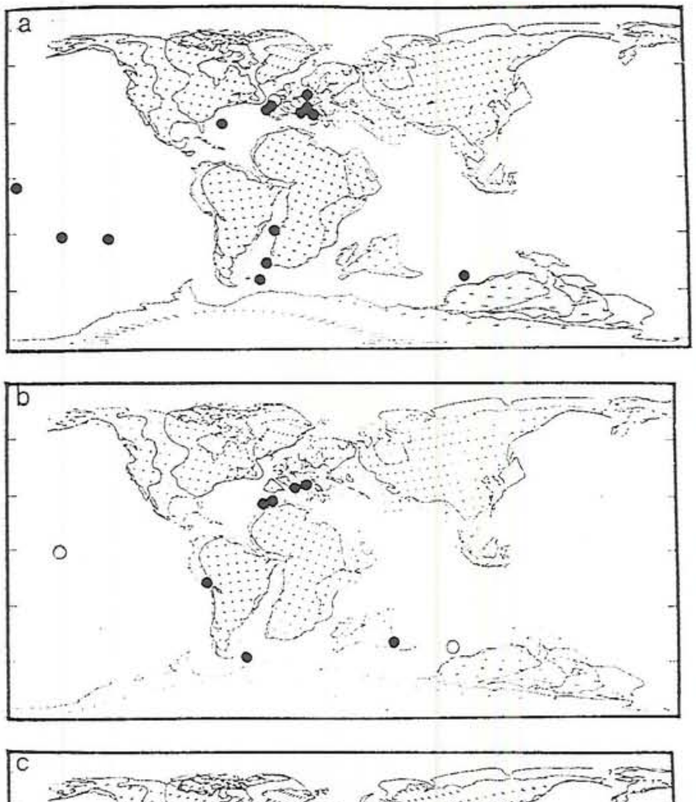


Fig. 17a-c Location of sites which show evidence of anoxia/dysoxia in the three proposed dysoxic/anoxic episodes. a. Early Aptian OAE1a, b. early Albian OAE1b and c. early late Albian OAE1c. These sites are shown by filled circles. Open circles indicate locations of sites which show no evidence of dysoxia/anoxia in these episodes. d. Location of sites which show evidence of anoxia/dysoxia in other intervals of the Aptian-Albian: triangles indicate late Aptian (*P. angustus* nannofossil Zone and subzone NC8A), squares indicate late early Albian-middle Albian (nannofossil subzones NC8C and NC9A) and circles indicate late late Albian (*E. turriseiffelii* nannofossil Zone). Paleogeographic reconstruction of Barron et al. [1981].

but these other horizons usually have lower organic carbon contents and can be differentiated by other geochemical and microfossil parameters [e.g., Pratt and King, 1986; Premoli Silva et al., 1989]. DSDP Site 465 on the Hess Rise in the paleo-equatorial Pacific includes a thick section of laminated, TOC-rich limestone in the upper Albian *E. turriseiffelii* nannofossil Zone. DSDP Site 402 in the Bay of Biscay recovered sediments with low TOC in the lower Aptian *C. litterarius* nannofossil Zone, and sediments with higher TOC contents in the overlying *P. angustus* Zone (Fig. 3). Certain sites (e.g., ODP Site 763, Exmouth Plateau) contain no evidence of high TOC or laminated sediments in the intervals of the proposed Albian subevents. South Atlantic DSDP sequences (e.g. Sites 361 and 364) contain TOC-rich Albian sediments, but with insufficient data and biostratigraphic control [Arthur and Natland, 1979; Graciansky et al., 1982; Müller et al., 1983] to be interpreted at this point.

At this time there is an insufficient data set to carry out a statistical analysis of the significance of correlation of the three subevents defined. In other words, it is not possible to quantify how closely all available TOC-rich samples correlate to the three subevents or, alternatively, how closely they approach a random distribution. In a purely qualitative sense there is a strong stratigraphic clustering of TOC-rich samples to the early Aptian subevent. However, as mentioned, in other sequences, this subevent is recognized by criteria other than TOC. Partly because of the smaller number of sites, there appears to be a weaker temporal clustering of TOC-rich horizons to the two Albian subevents. However, the validity of these episodes is strengthened considerably by the stratigraphic correlation of carbonaceous horizons deposited in deep-sea sites with prominent TOC-rich units which were laid down in epicontinental settings. Obviously, a final quantitative analysis of the objective basis of all of the subevents will have to await the recovery of more continuous Aptian-Albian sections and the refinement of zonations and chronologies.

Because of the various uncertainties described, the biostratigraphic intervals defined (Fig. 2) are considered as time "envelopes" for these subevents, the duration of which, at any given locality, may have been considerably shorter. Moreover, it is not possible at present, given current biostratigraphic resolution, to distinguish whether the subevents were synchronous or somewhat diachronous in different locations. This question is extremely difficult to answer in this interval in which magnetostratigraphy is inapplicable. The duration of any Mesozoic event is difficult to estimate because of the paucity of reliable radiometric ages, and uncertainties in the correlation of available radiometric ages with biostratigraphy. However, we can make some preliminary calculations. The durations of nannofossil subzones NC8B (4.2 m.y.) and NC9B (2.7 m.y.) were estimated assuming constant sedimentation rates at Site 763 utilizing the timescale of Harland et al. [1990]. We assume a duration of 0.56 m.y. for magnetic chron M0 [Harland et al., 1990] to estimate sedimentation rates at Site 641. Based on sedimentation rates and thicknesses of the correlative units at Site 641 (OAE 1a), Col de Palluel (OAE 1b) and Site 258 (OAE 1c), the durations of the subevents are as follows; OAE 1a and 1b: approximately 0.6 m.y., and OAE 1c: approximately 0.8 m.y.

A significant geographic trend emerges from the admittedly small data set analyzed. The early Aptian subevent appears to have affected a much larger number of sites than the later two subevents. These include locations in a variety of different paleogeographic settings such as open ocean plateaus, continental margins and epicontinental seaways (Fig. 17a). The other two subevents affected an equally widespread set of locations and oceanographic settings (Figs. 17b,c). However, it is clear that dysoxic/anoxic watermasses had a much patchier extent during these subevents as there are a large number of locations which show no evidence of dysoxia/anoxia. Site 763 provides a particularly striking comparison in this regards. The nearly complete upper Barremian-middle Turonian section recovered at this site includes TOC-rich sediments in the early Aptian (OAE 1a; Fig. 17a) and at the Cenomanian/Turonian boundary (OAE 2; e.g., Bralower and Siesser, 1992), however, the apparently complete Albian section contains no evidence of significant TOC enrichment resulting from OAEs 1b and 1c (Figs. 17b,c).

CHANGES OF PLANKTON ASSEMBLAGES

The effect of oxygen depletion on establishment of benthic communities had led to the association of such environments with intervals in sedimentary sections in which micro and macrobenthos are sparse or entirely absent [e.g., Eicher and Worstell, 1970; Savrda et al., 1984; Sageman et al., 1991]. In several sequences investigated, the intervals of the proposed dysoxic/anoxic episodes correspond to the absence of calcareous nannofossils and/or planktonic foraminifera. Most of these intervals are extremely enriched in calcified and pyritized radiolarians [e.g., Sigal, 1979; Thurow, 1988]. A complete lack of carbonate is observed in some of the most TOC-rich units such as the "Livello Selli".

In all of these cases, early diagenetic dissolution of relatively low amounts of carbonate may have occurred accompanying the oxidation of large quantities of organic matter [e.g., Emerson and Bender, 1981]. At other sites, such as DSDP Sites 398 and 511, and ODP Site 641, there are intervals in which all calcareous plankton have disappeared, and others in which planktonic foraminifera are absent but nannofossils are abundant, and in some cases, well preserved. Nannofossils are, in general, less susceptible to dissolution during diagenesis [Schlanger and Douglas, 1974] and this may explain their diversity and abundance in samples without planktonic foraminifera. However, the absence of obvious dissolution of the nannofossils in several samples from these sites within critical intervals indicates that the lack of microfossil carbonate in other samples within critical intervals may be, at least partially, of primary origin. Additionally, there is no apparent correlation between TOC and microfossil preservation in the sections studied, i.e. in some cases higher TOC values correspond to better microfossil preservation. However, as it is not possible to reconstruct the history of oxidation of organic matter in an individual horizon, a diagenetic interpretation cannot be totally ruled out.

It is possible that there was a rise in the CCD coincident with the onset of the overall OAE1 episode [e.g., Thierstein, 1979] and that the paleodepth of many of the sites studied was below the CCD for a short period of time. We view this as an unlikely explanation of the entire phenomenon due to the large range of paleodepths of the investigated sites, some of which are at least as shallow as several hundred meters (e.g., Vergons Section). Theoretically, nannoplankton and planktonic foraminifera should respond differently to dysoxic/anoxic conditions and associated changes in water column properties because of significant differences in life modes, habitats and ecological tolerances. Local disappearance of planktonic foraminifera may have been induced by changes in oxygenation or salinity. These organisms have fairly narrow salinity tolerances which do not differ by more than 5-6 ppt from open ocean values [e.g., Bé, 1977; Lipps, 1979; Hemleben et al., 1989]. Nannoplankton in general tend to be more tolerant to changes in water column conditions. Rare modern coccolithophores can endure salinities up to 250 ppt and below 5 ppt. However, the salinity tolerance of common species, such as *Emiliania huxleyi*, lies between approximately 16 and 40 ppt (see summary in Tappan, 1980). High latitude DSDP Site 511 on the Falkland Plateau is a location in which lowered salinity might have led to a low diversity, moderately well-preserved nannoflora and to the exclusion of planktonic foraminifera (Cores 511-60 to -62). The development of an oxygen-minimum zone also may have had a profound effect on planktonic foraminiferal assemblages, leading to a low diversity fauna of very shallow-dwelling opportunistic species [e.g., Caron and Homewood, 1983; Bréhéret et al., 1986; Arthur et al., 1987; Leckie, 1987].

A feasible explanation for the absence of nannoplankton in many of these TOC-rich intervals is related to productivity. These sediments are often enriched in dinoflagellates [Habib, 1979], and it is therefore possible that nannofossils were uncompetitive during a brief interval of high nutrient supply to surface waters. The production of modern coccolithophores, such as *E. huxleyi*, decreases in highly turbulent/highly fertile water masses [Margalef, 1978]. This is not a result of intolerance of coccolithophores to any ecological variable, but rather that diatoms and dinoflagellates have higher potential growth rates [Raymont, 1980], utilize nutrients more efficiently, and therefore outcompete coccolithophores under these conditions. Coccolithophores also become less competitive than dinoflagellates in highly stratified/low turbulent/low fertility water masses because they do not have resting cysts or functional flagella which allow them to leave the uppermost part of the water column in search of nutrients [Margalef, 1978]. Modern planktonic foraminifera seem to be more competitive than nannoplankton in eutrophic regimes [e.g., Hemleben et al., 1989], producing higher total abundances but lower species diversities than in oligotrophic settings. As with oxygenation, unstable, eutrophic environments led to the proliferation of opportunistic foraminifer taxa which can effectively compete with radiolarians and other zooplankters for food. There is no record of diatom productivity in these sections that might support an increase in nutrient supply, either because they were not present or abundant in the Early Cretaceous or because of dissolution and silica diagenesis. Our working hypothesis, therefore, is that these units which are barren of calcareous microfossils have a primary origin and were produced during brief eutrophic intervals.

One of the most pervasive questions concerning mid-Cretaceous organic carbon-rich sediments is whether they formed as a result of high oceanic productivity or effective preservation of organic carbon that was produced [e.g., Schlanger and Jenkyns, 1976; Waples, 1983]. Much of the debate is centered on the suitability of modern environments, such as the anoxic Black Sea, as modern analogs for ancient black-shale forming environments [e.g., Demaison and Moore, 1980; Pedersen and Calvert, 1990]. The largest source of error in these studies lies not with the calculation of organic carbon accumulation rates [e.g., Calvert et al., 1991], but rather with estimation of primary production rates which are known to be variable on short time-scales and difficult to measure accurately. Our paleoecological interpretation of these calcareous planktonic-free intervals follows some earlier ideas expressed by Arthur et al. [1987] and Leckie [1989] who attributed biotas of organic-rich horizons to higher productivity. We point out that such high productivity is intimately coupled with effective preservation of organic material on the sea floor in an oxygen-minimum zone setting in modern environments in which upwelling is perennial. The suggestion that individual TOC-rich units (which represent fractional zonal durations of less than 1 m.y.) were produced as a result of higher fertility conditions does not negate the conclusions that Bralower and Thierstein [1984], Dean et al. [1986] and Arthur et al. [1988] reached independently using different techniques, that average long-term mid-Cretaceous productivity (estimated at zonal durations of several millions of years) was a fraction of that at the present day. We also stress that this conclusion does not imply that other black shale horizons could not have been deposited under dramatically different conditions [e.g., Herbert and Fischer, 1986; Erba et al., 1989], and in particular, lower fertility.

PLANKTON EVOLUTION AND DYSOXIA/ANOXIA

Marine plankton experienced considerable diversification and turnover during the mid-Cretaceous [e.g., Tappan and Loeblich, 1973; Roth, 1987, 1989; Leckie, 1989]. Barremian-Aptian time was characterized by the first major radiation of planktonic foraminifera in terms of numbers of taxa and morphologic diversity [Leckie, 1989]. This same interval witnessed the demise of the dominant low-latitude assemblage of calcareous nannofossils, including *Nannoconus*, and the radiation of a more cosmopolitan coccolithophorid assemblage [Bergen, 1992]. Roth [1987] suggested that black shale deposition during the mid Cretaceous was associated with rising sea level, an increase in oceanic stratification, a decrease in oceanic fertility, and an increase in calcareous nannofossil diversification rates. Needless to say, there are numerous errors involved in the calculation of diversification and turnover rates and problems with the analysis of measures of species diversity but we show general trends in Figure 18. The problems with synthesizing such data are discussed in detail in the publications from which the data are taken. The relationship between calcareous plankton evolution and the development of widespread dysoxia/anoxia is complex as evidenced by the different diversity patterns between the three Oceanic Anoxic Subevents (Fig. 18).

Early Aptian OAE1a, probably the most widespread of the three Oceanic Anoxic Subevents, developed during a time of radiation in both the nannofossils and planktonic foraminifera (Fig. 18). According to Roth [1987], rates of nannofossil diversification peaked during early Aptian time. The planktonic foraminifera were in the midst of their first major radiation. The first presumed deeper dwelling morphotypes (e.g., *Ticinella* and *Planomalina*) did not appear until late Aptian time, and therefore, if OAE1a was associated with the development or expansion of an oxygen minimum zone, it would probably not have had a deleterious effect on a planktonic foraminiferal population composed exclusively of near-surface dwellers. Roth [1987] suggested that this time of overall nannofossil diversification represents the filling of niches to attain equilibrium diversity. This interpretation may be particularly relevant for the wide open niche-space being exploited by the planktonic foraminifera. In a study of the evolutionary history of Neogene planktonic foraminifera, Wei and Kennett [1986] suggested that equilibrium diversity is determined by the availability of resources or ecospace in a given physical environment. The late Barremian-early Aptian, somewhat similar to the early Miocene, was a time of overall diversification, where rates of speciation exceeded rates of extinction, despite the tectonic, climatic or oceanographic changes responsible for widespread dysoxia/anoxia during OAE1a.

In certain hemipelagic sites, blooms of nannoconids occur slightly above OAE 1a (e.g. at Site 641 and Gorgo a Cerbara), but species richness does not appear to change significantly. Nannoconid abundance increases in continental margin and epicontinental settings [e.g., Roth and Krumbach, 1986], but it is not possible to attribute this increase to an individual paleoecologic variable. It has been postulated [Busson and Noël, 1991] that this group are resting cysts produced by calcareous dinoflagellates in response to adverse conditions, in an analogous fashion to modern *Thoracosphaera*. While there are abundance peaks in certain samples, the late Barremian-early Aptian interval is accompanied by a significant overall decrease in the abundance of nannoconids (e.g., Coccioni et al., in press). The extinction of the dominant taxon, *N. steinmannii* occurred shortly after OAE1a.

Early Albian OAE1b, on the other hand, developed during a time of reduced simple diversities in the calcareous plankton and dinoflagellates, and reduced rates of nannofossil turnover and diversification (Fig. 18). Nearly all presumed deeper dwelling species of planktonic foraminifera disappeared, including the keeled morphotype [e.g., Leckie, 1989; Premoli Silva et al., 1989]. *H. planispira* dominates faunas in its nominate zone. Leckie [1989] suggested that this loss of planktonic foraminiferal diversity and morphologic variety was in response to the decrease of density gradients in the upper water column and proposed that increased rates of formation of warm, saline intermediate and/or deep water masses characterized latest Aptian and early Albian time. An alternative hypothesis, although physically improbable, is that an expanded oxygen-minimum zone occasionally reached to within tens of meters of the sea surface and thereby impinged on the niche of the deeper dwelling species of planktonic foraminifera. Increased oceanic fertility at this time may also have favored opportunists amongst the plankton and thus tended to lower overall

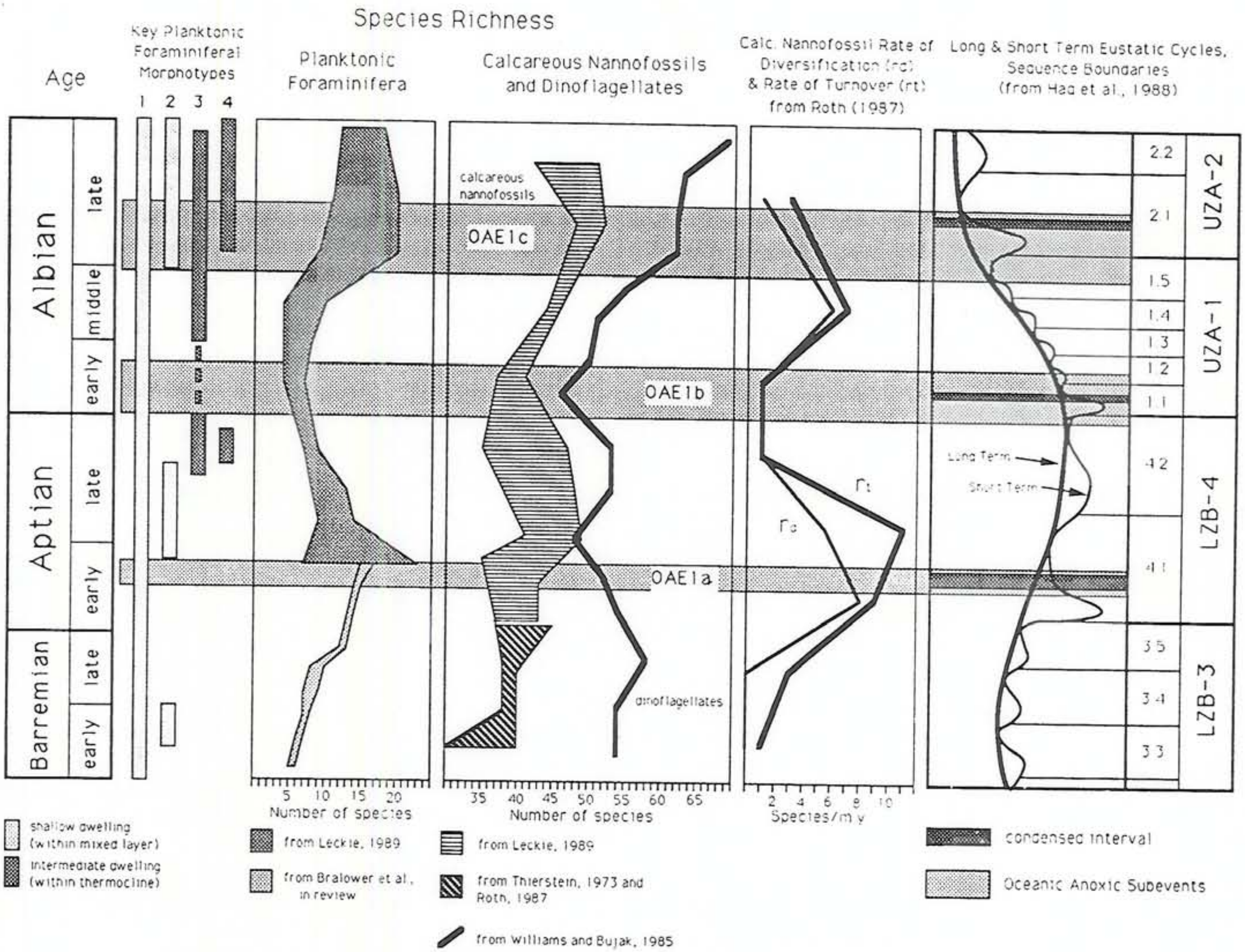


Fig. 18. Microfossil evolution during the late Barremian to Albian and relation to eustatic change. Illustrated from left are: A. Planktonic foraminiferal morphotypes: 1. Trochospiral and planispiral taxa (e.g., *Hedbergella, Globigerinelloides*); 2. Taxa with radially elongate chambers (e.g., *Schackoina, Leupoldina, Clavihedbergella*); 3. Taxa with complex apertural systems (e.g., *Ticinella*); 4. Keeled taxa (e.g., *Planomalina, Rotalipora, Praeglobotruncana*). B. Species Richness (number of species observed in individual samples; data sources indicated at base of figure) of planktonic foraminifera, calcareous nannofossils and dinoflagellates. C. Calcareous nannofossil rate of diversification and rate of turnover in number of species per million years [from Roth, 1987]. D. 2nd and 3rd order eustatic cycles [from Haq et al., 1988]. Also illustrated are the stratigraphic distribution of OASEs and condensed intervals. See text for full discussion.

diversity [Roth, 1987; Leckie, 1989]. These hypotheses are not mutually exclusive. Where observed, ammonite communities may indicate the effects of dysoxia/anoxia or related factors; the "Niveau Paquier" is dominated by Boreal and juvenile specimens of Tethyan taxa [Bréhéret et al., 1986] suggesting cooler, more fertile waters or a stressed environment. Calcareous nannofossil community structure does not appear to change nearly as dramatically. Abundance peaks of *Nannoconus* occur in coincidence with OAE 1b in some sites (e.g. in the "Niveau Paquier"; Bréhéret [1983]), also suggesting stressed or fertile surface waters.

Calcareous nannofossils experienced a marked increase in diversification rates during the mid-Albian, which began, perhaps coincidentally, in the same general time interval as OAE1b. Planktonic foraminifera, on the other hand, had an explosion in diversity and turnover rates during the late Albian, roughly coincident with the development of OAE1c (Fig. 18). The diversification of planktonic foraminifera during the late Albian signalled the development of strong density gradients close to or within the photic zone. New high-latitude water masses may have developed as sea level rose, epicontinental seas grew, and warm, saline water mass sources in the low latitudes were shut down, thereby permitting the establishment of enhanced upper water column density gradients [Leckie, 1989].

The evolutionary history of the calcareous plankton was very different between the time of OAE1a (early Aptian), OAE1b (early Albian), and OAE1c (late Albian) perhaps suggesting that the oceanographic processes responsible for the development of dysoxia/anoxia differed between these times, or alternatively, that factors controlling the preservation and accumulation of organic matter in the deep sea were decoupled from the uppermost water column where plankton evolution occurred. In addition, the evolutionary trends considered here are much longer-term phenomena than the relatively brief dysoxic/anoxic subevents. Even though nannoplankton and planktonic foraminiferal communities may have been affected by dysoxic/anoxic episodes in the Aptian-Albian, it is clear from the high diversity in samples overlying these intervals that the flora and fauna quickly reestablished themselves. The early Albian episode may have had a profound effect on the evolution of the calcareous plankton as this interval appears to correlate with major faunal and floral turnover and significantly reduced simple diversities. Both the early Aptian and early late Albian episodes come in the middle of an interval of diversification of nannofossil and planktonic foraminiferal assemblages. However, this diversification started earlier than the onset of dysoxic/anoxic conditions and there is no exact correlation with the few extinctions which occur in these intervals [Leckie, 1989; Bralower et al., submitted; Fig. 18]. The Aptian-Albian OAE has previously been evoked as having a major effect on the evolution of calcareous plankton [Roth, 1987, 1989; Leckie, 1989]; however this was clearly an interval of rapid paleoceanographic change and it is difficult to associate this evolution with an individual parameter.

The relative stasis of nannoplankton assemblages during the Aptian-Albian subevents compared to those of planktonic foraminifera and ammonites is similar to the situation at the Cenomanian/Turonian boundary OAE. In this latter interval, foraminifera and ammonites underwent dramatic changes in community structure, including decreased diversity and extinction of deeper-dwelling taxa [e.g., Hart, 1980; Caron and Homewood, 1983; Leckie, 1989; Elder, 1989], whereas nannoplankton community structure and diversity did not change significantly [Watkins, 1986; Bralower, 1988].

CAUSES OF "OCEANIC ANOXIC SUBEVENTS"

The correlation of many Aptian-Albian TOC-rich horizons to these three "oceanic anoxic subevents" indicates operation of a forcing mechanism over a widespread set of sites and oceanographic locations. Our results indicate that two of these events (OAE 1b and OAE 1c) had patchier, although apparently as widespread distributions as the other subevent (OAE 1a). Also, when compared in detail, it is clear that each episode left a variable stratigraphic record in each site. The stratigraphic and geographic variability displayed by these subevents leads us to distinguish between ocean-wide forcing mechanisms which caused water masses to be prone to oxygen-deficiency during the three time "envelopes" and factors which operate on a regional or even local scale, which actually induced or triggered dysoxia/anoxia in different water masses. In the following we briefly explore some of these global and regional forcing mechanisms [see Allard et al., in prep. for a more detailed analysis].

The concatenation of a number of global forcing mechanisms caused the oceans to be susceptible to oxygen deficiency during the three "ocean anoxic subevents". Evidence exists in lower Aptian sections in the Pacific Ocean basin that there may have been a relationship between volcanism and dysoxia/anoxia; for example, Vogt [1989] proposed that increased heat flow associated with mid-plate volcanic activity would lead to warming of bottom waters and would induce rapid turnover of nutrients, stimulating productivity followed by oxygen depletion in deep-water masses. Evidence for this suggestion includes lower Aptian carbonaceous limestones at DSDP Site 463 which are highly tuffaceous [e.g., Dean et al., 1981]. Recently, Tarduno et al. [1991] proposed that tuff-bearing sediments overlying basement on the Ontong Java Plateau correlate with the *G. blowi* foraminiferal Zone and that the production of this massive oceanic plateau occurred in a 1-3 m.y. episode in the latest Barremian to earliest Aptian. The interval of OAE 1a may, therefore, correlate with widespread midplate volcanic activity and the possible (but, as yet, undated) initiation of rapid seafloor spreading at the beginning of the Cretaceous long normal polarity zone [e.g., Larson, 1991a] (Fig. 19). Active volcanism may have led to "black shale" deposition as a result of increased pCO_2 (increased CO_2 outgassing rates), global warming and consequent decreases in oxygen solubility in deep-water masses [Arthur et al., 1991]. However, a more indirect mechanism may also be inferred.

Because of the absence of direct correlation, it is difficult to interpret fully the relationship between the proposed subevents and sea level. As a first approximation, however, the three OASE's appear to correspond in time with second order seismic sequence boundaries (super-cycles) as recognized by Haq et al. [1988] who defined three super-cycle boundaries in the Barremian through Albian interval (Fig. 19) [Arthur et al., 1990; Leckie et al., 1992]. All three represent Type 1 boundaries based on the recognition of low-stand fan deposits. The Haq et al. [1988] sequence stratigraphy and eustatic sea level curve represent a model. The stratigraphic expression of eustatic sea level varies from basin to basin. While the supercycles may represent second-order changes of global sea level, their boundaries may not everywhere be expressed as a Type 1 sequence boundary. The sequence stratigraphic interpretation of these three supercycles is that each is bounded by a relatively rapid rate of sea level fall, exceeding the rate of subsidence, followed by a relatively rapid rate of sea level rise. The subsequent sea level rise may be associated with the development of a condensed interval in offshore marine sections with the potential to accumulate greater concentrations of organic matter [Loutit et al., 1988]. The development of condensed intervals in such areas may be related to reduced sedimentary supply during relatively rapid transgressive sea level episodes, to a rising CCD as the locus of carbonate deposition shifts landward, or to increased geostrophic circulation and erosion or non-deposition near continental margins associated with increased deepwater renewal rates. However, it is not clear that OC concentration would increase simply as a result of reduced dilution (note that TOC concentration and accumulation rates in modern environments increase with increasing sedimentation rate [e.g., Müller and Suess, 1979; Sarnthein et al. 1988]). It appears more likely to us that

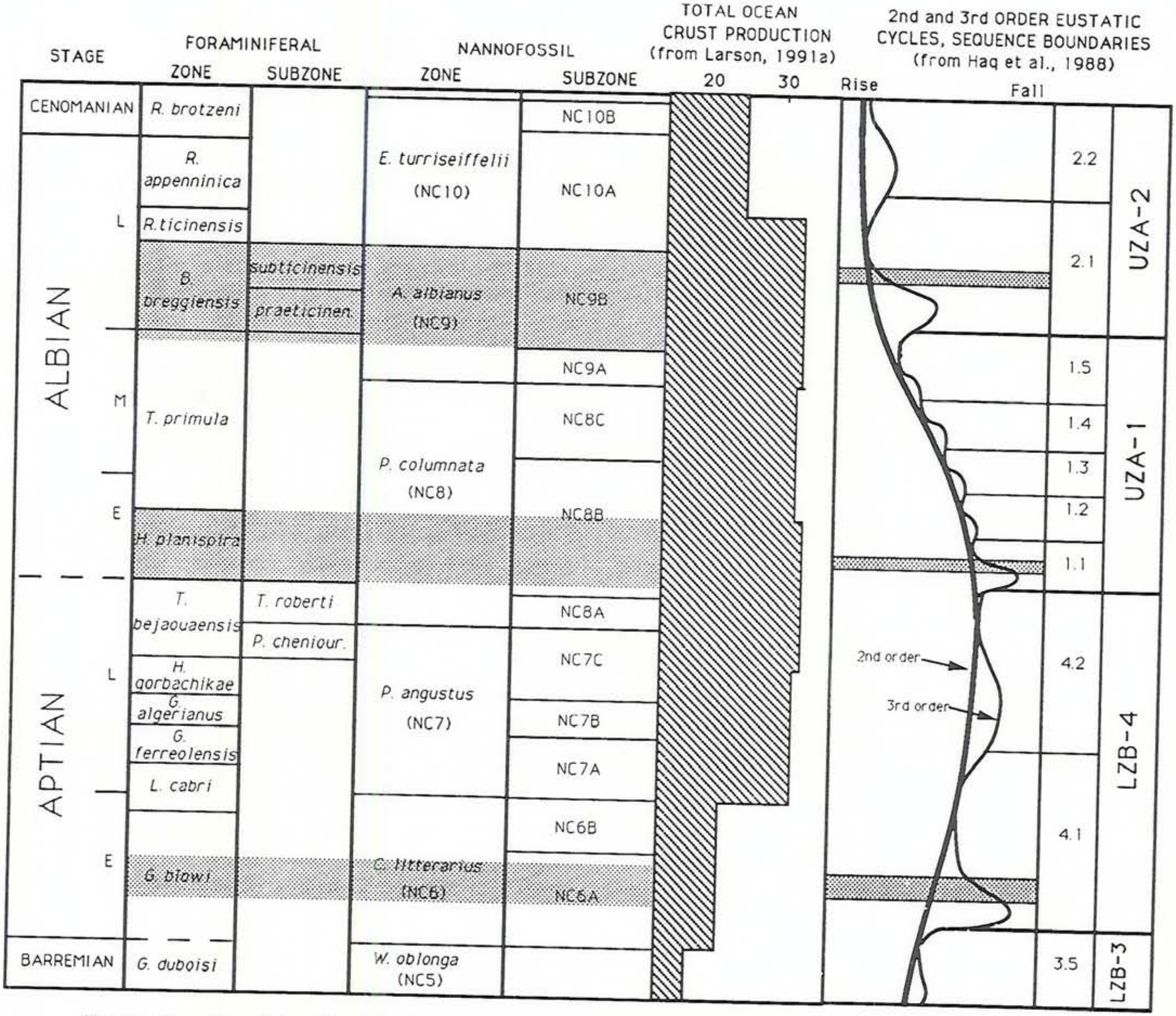


Fig. 19. Correlation of dysoxic/anoxic episodes with 2nd and 3rd order eustatic cycles and sequence boundaries of Haq et al. [1988] and total ocean crust production curve of Larson [1991a]. Terminology as in Haq et al. [1988]. Ages of eustatic cycles have been modified slightly based on apparent disparities in nannofossil zonations between Haq et al. [1988] and this study. Light shading refers to stratigraphic record of dysoxia/anoxia; dark shading refers to position of condensed section as predicted by sequence stratigraphy. Dark lines refer to Type 1 Sequence Boundaries. Crustal production rates are in 10^6 km^3/m.y. See text for full discussion.

some other mechanism is operating to enhance TOC flux or preservation at the same time.

Although this mechanism is not entirely clear, in some cases the intervals of the hypothesized oceanic dysoxic/anoxic subevents correspond to unconformities and condensed section. For example, Bralower et al. [submitted] showed that the interval encompassing or directly overlying early Aptian OAE 1a is condensed in numerous sequences. The *H. planispira* Zone (OAE 1b) is absent from a fairly continuous pelagic section in the Calera Limestone of California [Sliter, 1991]. The upper part of the *Rotalipora subticinensis* Subzone of the *B. breggiensis* Zone (OAE 1c) is also missing in the Calera Limestone. A pronounced unconformity correlates with OAE 1c at DSDP Site 545 off the Mazagan Escarpment (Fig. 9). Therefore the rock record, although extremely indirect, does support a correlation of Aptian/Albian organic carbon enrichments with transgressive phases in sea level.

A more direct relationship has been established for the interval of the Cenomanian/Turonian OAE which can be directly tied to a late transgressive phase of sea level in outcrops of the Western Interior Cretaceous Seaway [e.g., Kauffman, 1977] or in northwest

Europe [e.g., Hancock and Kauffman, 1979]. Based on this correlation, Arthur et al. [1987] proposed that flooding of low-latitude evaporative marginal seas led to increased rates of "warm, saline" deep-water formation, which consequently caused upwelling of nutrients over large areas of the oceans, particularly at ocean margins. Because increased seafloor spreading rates and mid-plate volcanism are likely driving mechanisms for sea level changes [e.g., Larson, 1991a, b], they may have indirectly led to the correlation of intervals of elevated sea level and Aptian-Albian dysoxic/anoxic episodes but were not implicated in the Cenomanian/Turonian event [Schlanger, 1986].

While the previously discussed factors might have caused the ocean to be more susceptible to oxygen deficiency, none of them could have directly induced a water mass to become fully oxygen depleted. A different set of factors, which operate on more regional or local scales, directly affect the supply or demand of oxygen in a water mass. These factors might include upwelling events and higher OC fluxes from surface waters which can stress the supply of oxygen in an individual water mass, but which are unlikely to have affected large areas of the ocean simultaneously, or tectonic factors which lead to restriction of a basin. Regional climatic fluctuations such as increased precipitation and ensuing runoff might create stable stratification on the longer term [e.g., Pratt, 1984] and lead to less replenishment of oxygen by mixing down of surface waters during storms or other events such as winter cooling and convection. Other potential factors depleting oxygen include: 1) advection of oxygen minima into epicontinental seaways or onto shelves upon sea level rise [e.g., Frush and Eicher, 1975; Heckel, 1977; Fischer and Arthur, 1977], and 2) rising temperature and consequent decrease of oxygen solubility of deep waters.

In conclusion, it appears that the oceans were more prone to dysoxia/anoxia during the early Aptian and Cenomanian/Turonian events than during the two Albian subevents. It is likely that similar global forcing mechanisms, such as high sea level and an overall warm, equable climate, operated in all of these time intervals. However the patchier extent of dysoxic/anoxic water masses in the Albian subevents indicates either that the oceans were less susceptible to dysoxia/anoxia at these times or that local/regional triggering factors had a stronger influence.

CONCLUSIONS

The Aptian-Albian Oceanic Anoxic Event is characterized by at least three individual dysoxic/anoxic episodes: OAE 1a, early Aptian *G. blowi* foraminiferal Zone, *C. litterarius* nannofossil Zone, *C. rothii* nannofossil Subzone; OAE 1b, early Albian *H. planispira* foraminiferal Zone, *P. columnata* nannofossil Zone (subzone NC8B); and OAE 1c, early late Albian (*B. breggiensis* foraminiferal Zone, *A. albianus* nannofossil Zone (subzone NC9B). These subevents are observed in sections deposited in pelagic environments and along continental margins in all ocean basins, and in several epicontinental seaways. All three episodes appear to have had a nearly ocean-wide extent, but the distribution of dysoxic/anoxic watermasses in the Albian intervals appears to have been much patchier than during the early Aptian episode.

Although none of the subevents is associated with major extinctions, they appear to be marked by changes in the community structure of planktonic foraminifera, but the pattern differs for each subevent. The absence of planktonic foraminifera in some intervals may be related to dissolution and/or adverse ecological factors such as low salinity and/or oxygen deficiency. Nannofossil taxa in these horizons may show marked epicontinental affinities or may have been excluded competitively by other phytoplankton.

The correlation of the subevents with sea level is only indirect at this time. However, rising sea level and climatic consequences of volcanism are likely factors which conditioned the oceans to be prone to dysoxia/anoxia through a variety of feedbacks. Regional or even local climatic or tectonic factors or high biologic productivity actually induced or triggered dysoxia/anoxia at mid- to deep-water depths.

Acknowledgments. We gratefully acknowledge discussions with W. Balch, W. Dean, E. Erba and J. Mutterlose. We thank D. Bottjer, D. Eicher, W. Elder, E. Erba, T. Herbert, H. Jenkyns, S. Kidwell, K. McDougall, J. Mutterlose, C. Paull and B. Sageman for critical reviews of a previous manuscript. We are grateful to C. Botelho, J.T. Ellis, C. Mellor, M. Parrow and L. Sierra for laboratory assistance. We thank S. Shafik and V. Phansalkar for generously donating material from Australia and India, respectively, and the Ocean Drilling Program for numerous samples. Research supported by NSF (EAR-8721350).

RML would like to thank Richard Cashman for labratory assistance. RML's contributions to this project have been partially funded by the Petroleum Research Fund, administered by the American Chemical Society.

REFERENCES

Applegate, J.L., and J.A. Bergen, Cretaceous calcareous nannofossil biostratigraphy of sediments recovered from the Galicia Margin, ODP Leg 103, *Proc. Ocean Drilling Program, Sci. Results*, 103, 293-319, 1988.

Arthur, M.A., North Atlantic Cretaceous black shales: the record at Site 398 and a brief comparison with other occurrences, *Initial Rep. Deep Sea Drilling Project*, 47B, 719-738, 1979.

Arthur, M.A., and J.H. Natland, Carbonaceous sediments in the North and South Atlantic: the role of salinity in stable stratification of early Cretaceous basins, in M. Talwani, W. Hay, and W.B.F. Ryan (eds.), *Deep Drilling in the Atlantic Ocean: Continental Margins and Paleoenvironment*, Maurice Ewing Series, 3, Am. Geophys. Union, Washington, D.C., pp. 375-401, 1979.

Arthur, M.A., S.O. Schlanger, and H.C.Jenkyns, The Cenomanian-Turonian Oceanic Anoxic Event, II, Paleoceanographic controls on organic matter production and preservation, in *Marine Petroleum Source Rocks, Spec. Publ.*, 24, J. Brooks and A. Fleet (eds.), Geological Society of London, pp. 399-418, 1987.

Arthur, M.A., W.E. Dean, and G.E. Claypool, Anomalous ^{13}C enrichment in modern marine organic carbon, *Nature*, 315, 216-218, 1988.

Arthur, M.A., H.-J. Brumsack, H.C. Jenkyns, and S.O. Schlanger, Stratigraphy, geochemistry, and paleoceanography of organic carbon-rich Cretaceous sequences, in: R.N. Ginsburg, and B. Beaudoin (eds.), *Cretaceous Resources, Events and Rhythms*, Kluwer Academic Publishers, pp. 75-119, 1990.

Arthur, M.A., L.R. Kump, W.E. Dean, and R.L. Larson, Superplume?, Supergreenhouse? *EOS*, 72, 301, 1991.

Barron, E.J., and W.M. Washington, Cretaceous climate: a comparison of atmospheric simulations with the geologic record, *Palaeogeogr., Palaeoclimatol., Palaeoecol.*, 40, 103-133, 1982.

Barron, E.J., and W.H. Peterson, W.H., Mid-Cretaceous ocean circulation: results from model sensitivity studies, *Paleoceanography*, 5, 319-338, 1990.

Barron, E.J., C.G. Harrison, J.L. Sloan, II, and W.W. Hay, Paleogeography, 180 million years ago to the present: *Ecolog. Geol. Helv.*, 74, 443-470, 1981.

Bé, A.W.H., An ecological, zoogeographic and taxonomic review of Recent planktonic foraminifera: in A.T.S. Ramsay (ed.), *Oceanic Micropaleontology*, Academic Press, New York, 1, 1-100, 1977.

Benavides-Caceres, V.E., Cretaceous System in Northern Peru, *Bull. Am. Mus. Nat. Hist.*, 108, 357-493, 1956.

Bergen, J.A., Assemblage turnover in Mesozoic calcareous nannofossils: periodicity and distinction from the terminal Cretaceous event, *5th North American Paleontological Congress, Abstracts with Programs*, 26, 1992.

Berner, R.A., A.C. Lasaga, and R.M. Garrels, The carbonate-silicate geochemical cycle and its effect on atmospheric carbon dioxide over the past 100 million years: *Amer. Jour. Sci.*, 283, 641-683, 1983.

Blechschmidt, G., Biostratigraphy of calcareous nannofossils: leg 47B, Deep Sea Drilling Project: *Initial Rep. Deep Sea Drilling Project*, 47B, 327-360, 1979.

Bode, G.W., Appendix II. Carbon and carbonate analyses, *Initial Rep. Deep Sea Drilling Project*, 73, 1185-1187, 1983.

Boillot, G., E.L. Winterer, A.W. Meyer, et al., *Proc. Ocean Drilling Program, Sci. Results*, 103, 663pp, 1988.

Bolli, H.M., Zonation of Cretaceous to Pliocene marine sediments based on planktonic foraminifera, *Bol. Inform. Asoc. Venezolana de Geol., Min y Petrol.*, 9, 3-32, 1966.

Bolli, H.M., W.B.F. Ryan, et al., *Initial Rep. Deep Sea Drilling Project*, 40, 1079 pp, 1978.

Bralower, T.J., Organic carbon and transition metal accumulation rates in Holocene and mid-Cretaceous sediments: data and techniques, *Scripps Institution of Oceanography Data Series*, 84-24, 55pp, 1984.

Bralower, T.J., Valanginian to Aptian calcareous nannofossil stratigraphy and correlation with the upper M-sequence magnetic anomalies, *Mar. Micropaleontol.*, 11, 293-310, 1987.

Bralower, T.J., Calcareous nannofossil biostratigraphy and assemblages of the Cenomanian-Turonian boundary interval: implications for the origin and timing of oceanic anoxia, *Paleoceanography*, 3, 275-316, 1988.

Bralower, T.J., Aptian-Albian calcareous nannofossil biostratigraphy of ODP Site 763, and the correlation between high and low latitude zonations, in: Duncan, R.A., and D.K. Rea, (eds.), *Am. Geophys. Union Monograph*, 70, 1992.

Bralower, T.J., and H.R. Thierstein, Low productivity and slow deep-water circulation in mid-Cretaceous oceans: *Geology*, 12, 614-618, 1984.

Bralower, T.J., and W.G. Siesser, Cretaceous calcareous nannofossil biostratigraphy of ODP Leg 122 Sites 761, 762 and 763, Exmouth and Wombat Plateaus, N.W. Australia, *Proc. Ocean Drilling Program, Sci. Results*, 122, 529-556, 1992.

Bralower, T.J., M.A. Arthur, R.M. Leckie, W.V. Sliter, D.J. Allard, and S.O. Schlanger, Oceanic anoxia/dysoxia in the early Aptian: lithostratigraphy, biostratigraphy and paleoecological consequences, *Paleoceanography*, submitted.

Brass, G.W., J.R. Southam, and W.H. Peterson, Warm saline bottom water in the ancient ocean, *Nature*, 296, 620-623, 1982.

Bréhéret, J.-G., Sur des niveaux de black shales dans l'Albien inférieur et moyen du domaine vocontien (sud-est de la France), étude de nannofacies et signification des paléoenvironments, *Bull. Mus. natn. Hist. nat., Paris*, 5, 113-159, 1983.

Bréhéret, J.-G., Indices d'un événement anoxique étendu à la Téthys alpine, à l'Albien inférieur (événement Paquier), *C. R. Acad. Sc. Paris*, 300, 355-358, 1985.

Bréhéret, J.-G., Épisodes de sédimentation riche en matière organique dans les marnes bleues d'âge aptien et albien de la partie pélagique du bassin vocontien, *Bull Soc. géol. France*, 8, 349-356, 1988.

Bréhéret, J.-G., M. Caron, and M. Delamette, Niveaux riches en matière organique dans Albien vocontien, *Docum. Bur. Rech. géol. min.*, 110, 141-191, 1986.

Busson, G., and D. Noël, Les nannoconidés, indicateurs environmentaux des océans et mers épicontinentales du Jurassique terminal et du Crétacé inférieur, *Oceanologica Acta*, 14, 333-356, 1991.

Calvert, S.E., R.E. Karlin, L.J. Toolin, D.J. Donahue, J.R. Southon, and J.S. Vogel, Low organic carbon accumulation rates in Black Sea sediments, *Nature*, 350, 692-695, 1991.

Cameron, D.H., Grain size and carbon/carbonate analyses, Leg 43, *Initial Rep. Deep Sea Drilling Project*, 43, 1043-1048, 1979.

Caron, M., Cretaceous planktonic foraminifers from DSDP Leg 40, Southeastern Atlantic Ocean, *Initial Rep. Deep Sea Drilling Project*, 40, 651-678, 1978.

Caron, M. and Homewood, P., Evolution of early planktonic foraminifers, *Mar. Micropaleontol.*, 7, 453-462, 1983.

Cepek, P., Mesozoic calcareous nannoplankton of the eastern North Atlantic, Leg 41, *Initial Rep. Deep Sea Drilling Project*, 41, 667-688, 1977.

Cepek, P., Mesozoic calcareous nannoplankton stratigraphy of the Central North Pacific (Mid-Pacific Mountains and Hess Rise), Deep Sea Drilling Project, Leg 62, *Initial Rep. Deep Sea Drilling Project*, 62, 397-418, 1981.

Channell, J.E.T., W. Lowrie, and F. Medizza, Middle and Early Cretaceous magnetic stratigraphy from the Cismon section, Northern Italy, *Earth Planet. Sci. Lett.*, 42, 153-166, 1979.

Coccioni, R., O. Nesci, M. Tramontana, F.-C. Wezel, and E. Moretti, Descrizione di un livello-guida "radiolaritico-bituminoso-ittiolitico" alla base delle Marne a Fucoidi nell Appennino Umbro-Marchigiano, *Boll. Soc. Geol. It.*, 106, 183-192, 1987.

Coccioni, R., R. Franchi, O. Nesci, F.-C. Wezel, F. Battistini, and P. Pallecchi, Stratigraphy and mineralogy of the Selli Level (Early Aptian) at the base of the Marne a Fucoidi in the Umbro-Marchean Apennines (Italy), in J. Wiedmann (ed.) Cretaceous of the Western Tethys, *Proc. 3rd Int. Cretaceous symposium, Tübingen*, 1989.

Coccioni, R., E. Erba, and I. Premoli Silva, Barremian-Aptian calcareous plankton biostratigraphy from the Gorgo a Cerbara section (Marche, central Italy) and implications for plankton evolution, *Cret. Res.*, in press.

Davies, T.A., B.P. Luyendyk, et al., *Initial Rep. Deep Sea Drilling Project*, 26, 1129pp, 1974.

Dean, W.E., Calcium carbonate and organic carbon in samples from Deep Sea Drilling Project Sites 463, 464, 465 and 466, *Initial Rep. Deep Sea Drilling Proj.*, 62, 869-876, 1981.

Dean, W.E., G.E. Claypool, and J. Thiede, Origin of organic carbon-rich mid-Cretaceous limestones, mid-Pacific mountains and southern Hess Rise, *Initial Rep. Deep Sea Drilling Project*, 61, 877-890, 1981.

Dean, W.E., M.A. Arthur, and G.E. Claypool, Depletion of ^{13}C in Cretaceous marine organic matter: Source, diagenetic or environmental signal?, *Mar. Geol.*, 70, 119-158, 1986.

Demaison, G.J., and G.T. Moore, Anoxic environments and oil source bed genesis, *Organic Geochem.*, 2, 9-31, 1980.

Deroo, G., J.P. Herbin, J. Roucaché, and B. Tissot, Organic geochemistry of Cretaceous shales from DSDP Site 398, Leg 47B, Eastern North Atlantic, *Initial Rep. Deep Sea Drill. Project*, 47, pt. B, 513-522, 1979.

Deroo, G., J.P. Herbin, J. Roucaché, and B. Tissot, Organic geochemistry of Cretaceous sediments from DSDP Holes 417D (Leg 51), 418A (Leg 52) and 418B (Leg 53), Western North Atlantic, *Initial Rep. Deep Sea Drilling Project*, vs. 51-53, 737-746, 1980.

Deroo, G., J.P. Herbin, and J. Roucaché, Organic geochemistry of Upper Jurassic-Cretaceous sediments from Site 511, Leg 71, Western South Atlantic, *Initial Rep. Deep Sea Drilling Project*, 71, 1001-1013, 1983.

Deroo, G., J.P. Herbin, and J. Roucaché, Organic geochemistry of Cenozoic and Mesozoic sediments from Deep Sea Drilling Project Sites 544 to 547, Leg 79, Eastern North Atlantic, *Initial Rep. Deep Sea Drilling Project*, 79, 721-742, 1984.

Donnelly, T., J. Francheteau, et al., *Initial Rep. Deep Sea Drilling Project*, 51, 52 and 53, 1613pp, 1980.

Douglas, R.G., Planktonic foraminiferal biostratigraphy in the Central North Pacific Ocean, *Initial Rep. Deep Sea Drilling Project*, 17, 607-694, 1973.

Dunham, K.W., P.A. Meyers, and E.S. Ho, Organic geochemistry of Cretaceous black shales and adjacent strata from the Galicia Margin, North

Atlantic Ocean, *Proc. Ocean Drilling Program, Sci. Results*, 103, 557-565, 1988.

Dupeuble, P.A., Mesozoic foraminifers and microfacies from Holes 400A, 401 and 402A of the DSDP Leg 48, *Initial Rep. Deep Sea Drill. Project*, 48, 451-474, 1979.

Eicher D.L., and P. Worstell, Cenomanian and Turonian foraminifera from the Great Plains, United States, *Micropaleontology*, 16, 269-324, 1970.

Elder, W.P., Molluscan extinction patterns across the Cenomanian-Turonian stage boundary in the Western Interior of the United States, *Paleobiology*, 15, 299-320, 1989.

Emerson, S., and M. Bender, Carbon fluxes at the sediment-water interface of the deep-sea: calcium carbonate preservation, *J. Mar. Res.*, 39, 139-162, 1981.

Erba, E., Aptian-Albian calcareous nannofossil biostratigraphy of the Scisti a Fucoidi cored at Piobbico (Central Italy), *Riv. It. Paleont. Strat.*, 94, 249-284, 1988.

Erba, E., and B. Quadrio, Biostratigrafia a Nannofossili calcarei, Calpionellidi e Foraminiferi planctonici della Maiolica (Titoniano superiore-Aptiano) nelle Prealpi Bresciane (Italia settentrionale), *Riv. It. Paleont. Strat.*, 93, 3-108, 1987.

Erba, E., I. Premoli Silva, and M.E. Tornaghi, The Aptian-Albian scisti a fucoidi, Italy, record changes in fertility and carbonate productivity, *Terra Abstracts*, 1, 42, 1989.

Erdman, J.G., and K.S. Schorno, Geochemistry of carbon, Deep Sea Drilling Project, Leg 41, *Initial Rep. Deep Sea Drilling Project*, 41, 849-854, 1977.

Erdman, J.G., and K.S. Schorno, Geochemistry of carbon: Deep Sea Drilling Project, Legs 47A and 47B, *Initial Rep. Deep Sea Drilling Project*, 47, pt. B, 553-560, 1979.

Fischer, A.G., and M.A. Arthur, Secular variations in the pelagic realm, *Soc. Econ. Paleont. Min., Spec. Publ.*, 25, 19-50, 1977.

Foresman, J.B., Organic geochemistry DSDP Leg 40, continental rise of southwest Africa, *Initial Rep. Deep Sea Drilling Project*, 40, 557-568, 1978.

Frush, M.P., and D.L. Eicher, Cenomanian and Turonian foraminifera and paleoenvironments in the Big Bend region of Texas and Mexico, in W.G.E. Caldwell (ed.), *The Cretaceous system in the Western Interior of North America, Geological Assoc. of Canada, Spec. Pap.*, 13, 277-301, 1975.

Gaida, K.-H., R. Gedenik, E. Kemper, W. Michaelis, R. Scheuch, H.-H. Schmitz, and W. Zimmerle, litholosche, mineralogische und organisch-geochemische Untersuchungen an Tonsteinen und Tonmergelsteinen der Unterkreide Nordwest-deutschlands (unter besonder Berücksichtigung der Schwarzschiefer), *Geol. Jahrb.*, A, 58, 15-47, 1981.

Gartner, S. Jr., Calcareous nannofossils, Deep Sea Drilling Project Holes 418A and 418B, *Initial Rep. Deep Sea Drilling Project*, 51, 52, 53, 815-821, 1980.

Govinden, A., Imprint of global "Cretaceous Anoxic Events" in east coast basins of India and their implications, *Bull. ONGC*, 19, 257-270, 1982.

Graciansky, P.C. de, E. Brosse, G. Deroo, J.P. Herbin, L. Montadert, C. Müller, J. Sigal, and A. Schaaf, Les formations d'Age Crétacé de l'Atlantique Nord et leur matière organique paléogéographie et milieux de dépôt, *Revue de l'Inst. France Petrole*, 37, 276-336, 1982.

Habib, D., Sedimentary origin of North Atlantic Cretaceous palynofacies, in M. Talwani, W. Hay and W.B.F. Ryan (eds.), *Deep Drilling Results in the Atlantic Ocean; Continental Margins and Paleoenvironment*, Maurice Ewing Series, v, 3, Am. Geophys. Union, Washington, D.C., 420-437, 1979.

Haig, D.W., Cretaceous foraminiferal biostratigraphy of Queensland, *Alcheringa*, 3, 171-187, 1979.

Hancock, J.M., and E.G. Kauffman, The great transgressions of the Late Cretaceous. *J. Geol. Soc. Lond.*, 136, 175-186, 1979.

Haq, B.U., J. Hardenbol, and P.R. Vail, Chronology of fluctuating sea levels since the Triassic (250 million years to Present), *Science*, 235, 1156-1167, 1987.

Haq, B.U., J. Hardenbol, and P.R. Vail, Mesozoic and Cenozoic chronostratigraphy and cycles of sea-level change, *Soc. Econ. Paleont. Min., Spec. Publ.*, 42, 71-108, 1988.

Haq, B.U., U. von Rad, S. O'Connell, et al., *Proc. Ocean Drilling Program, Sci. Results*, 122, 826pp, 1990.

Harland, W.B., R.L. Armstrong, A.V. Cox, L.E. Craig, A.G. Smith, and D.G. Smith, *A geologic time scale*, Cambridge University Press, 263pp, 1990.

Hart, M.B., A water depth model for the evolution of the planktonic Foraminiferida, *Nature*, 286, 252-254, 1980.

Heckel, P.H., Origin of phosphatic black shale facies in Pennsylvanian cyclothems of mid-continent North America, *Am. Assoc. Petrol. Geol. Bull.*, 61, 1045-1068, 1977.

Hemleben, C., M. Spindler, and O.R. Anderson, *Modern Planktonic Foraminifera*, 363pp, Springer Verlag, New York, 1989.

Herbert, T.D., and A.G. Fischer, Milankovitch climate origin of mid-Cretaceous black shale rhythms, central Italy, *Nature*, 321, 739-743, 1986.

Herbin, J.P., F. Magniez-Jannin, and C. Müller, Mesozoic organic-rich sediments in the South Atlantic: distribution in time and space, in Biogeochemistry of Black Shales, E.T. Degens, P.A. Meyers, and S.C. Brassell,(eds.), *Mitteilingen aus dem Geologisch-Paläontologischen Institut der Universität Hamburg*, 71-98, 1986.

Herbin, J.P., C. Müller, P.C. de Graciansky, T. Jaquin, F. Magniez-Jannin, and P. Unternehr, Cretaceous anoxic events in the South Atlantic, *Revista Brasieleira de Geociencias*, 17, 92-99, 1987.

Hill, M.E., and T.J. Bralower, Early evolution, stratigraphy and taxonomy of *Eiffellithus eximius* and related species, *Abh. Geol. Bundesanst.*, 39, 98-97, 1987.

Hinz, K., E.L. Winterer, et al., *Initial Rep. Deep Sea Drilling Project*, 79, 934pp, 1984.

Jaillard, E., Sedimentary evolution of an active margin during middle and Upper Cretaceous times: the north Peruvian margin from late Aptian up to Senonian, *Geol. Rund.*, 76, 677-697, 1987.

Jenkyns, H.C., Cretaceous anoxic events: From continents to oceans, *J. Geol. Soc. London*, 137, 171-188, 1980.

Johnson, D.L., R.D. McIver, and M.A. Rogers, Insoluble organic matter bitumens in Leg 47 samples, *Initial Rep. Deep Sea Drilling Project*, 47, pt. B, 543-546, 1979.

Kale, A.S., and V.G. Phansalkar, Nannofossil biostratigraphy of the Utatur Group, Trichinopoly District, South India, *Mem. di Scienze Geol., Univ. di Padova*, 63, 89-107, 1992.

Kauffman, E.G., Geological and biological overview, Western Interior Cretaceous basin, in Cretaceous facies, faunas and paleoenvironments across the Western Interior Basin: edited by E.G. Kauffman, *Mountain Geologist*, 14, 75-99, 1977.

Kauffman, E.G., B.B. Sageman, J.I. Kirkland, W.P. Elder, T.J. Harries, and T. Villamil, Molluscan biostratigraphy of the Western Interior Cretaceous Basin, North America, in W.G.E. Caldwell and E.G. Kauffman (eds.), *Cretaceous Evolution of the Western Interior Basin of North America, Geological Assoc. of Canada, Spec. Pap.*, in press.

Koutsoukos, E.A.M., M.R. Mello, N.C. de Azambuja Filho, M.B. Hart, and J.R. Maxwell, The Upper Aptian-Albian succession of the Sergipe Basin, Brazil: an integrated paleoenvironmental assessment, *Am. Assoc. Petrol. Geol. Bull.*, 75, 479-498, 1991.

Krasheninnikov, V.A., and I.A. Basov, Stratigraphy of Cretaceous sediments of the Falkland Plateau based on planktonic foraminifers, Deep Sea Drilling Project, Leg 71, *Initial Rep. Deep Sea Drilling Project*, 71, 789-820, 1983.

Lancelot, Y., E. Seibold, et al., *Initial Rep. Deep Sea Drilling Project*, 41, 1259pp, 1977.

Larson, R.L., Latest pulse of the Earth: evidence for a mid-Crataceous super plume, *Geology*, 19, 547-550, 1991a.

Larson, R.L., Geological consequences of superplumes, *Geology*, 19, 963-966, 1991b.

Leckie, R.M., Mid-Cretaceous planktonic foraminiferal biostratigraphy off

Central Morocco, Deep Sea Drilling Project Leg 79, Sites 545 and 547. *Initial Rep. Deep Sea Drilling Project*, 79, 579-620, 1984.

Leckie, R.M., Paleoecology of mid-Cretaceous planktonic foraminifera: a comparison of open ocean and epicontinental sea assemblages, *Micropaleontology*, 33, 164-176, 1987.

Leckie, R.M., An oceanographic model for the early evolutionary history of planktonic foraminifera, *Paleogeogr. Paleoclimatol. Paleoecol.*, 73, 107-138, 1989.

Leckie, R.M., and T.J. Bralower, New ages for planktonic foraminiferal datums in the Barremian-Lower Aptian, *Geol. Soc. America, Abs. with Prog.*, 23, 167, 1991.

Leckie, R.M., R. Scott, R., T.J. Bralower, and W.V. Sliter, Relationships between sequence boundaries and the evolutionary history of planktonic foraminifera, calcareous nannofossils and reef communities in the mid Cretaceous (Barremian-Cenomanian), *5th North American Paleontology Abstracts with Programs*, 179, 1992.

Lipps, J.H., The ecology and paleoecology of planktic foraminifera, in *Foraminiferal Ecology and Paleoecology*, SEPM Short Course, 6, 62-104, 1979.

Longoria, J.F., Stratigraphic, morphologic and taxonomic studies of Aptian planktonic foraminifera, *Riv. Esp. Micropal. Num. Extra*, 107pp, 1974.

Loutit, T.S., J. Hardenbol, P.R. Vail, and G.R. Baum, Condensed sections: the key to age determination and correlation of continental margin sequences, *Soc. Econ. Paleont. Min., Spec. Publ.*, 42, 183-213, 1988.

Lowrie, W., and W. Alvarez, Lower Cretaceous magnetic stratigraphy in Umbrian pelagic limestone sections, *Earth Planet. Sci. Lett.*, 71, 315-328, 1984.

Lowrie, W., W. Alvarez, I. Premoli Silva, and S. Monechi, Lower Cretaceous magnetic stratigraphy in Umbrian pelagic carbonate rocks, *Geophys. J.R. Astron. Soc.*, 60, 263-281, 1980.

Ludwig, W.J., V. Krasheninnikov, et al., Site 511 Summary, *Init. Rep. Deep Sea Drilling Project*, 71, pt. 1, 477pp, 1983.

Manivit, H., K. Perch-Nielsen, B. Prins, and J.W. Verbeek, Mid Cretaceous calcareous nannofossil biostratigraphy, *Proc. K. Ned. Akad. Wet.* Ser. B, 80, 169-181, 1977.

Margalef, R., Life-forms of phytoplankton as survival alternatives in an unstable environment, *Oceanologica Acta*, 1, 493-509, 1978.

Monechi, S., and H.R. Thierstein, Late Cretaceous-Eocene nannofossil and magnetostratigraphic correlations near Gubbio, Italy, *Mar. Micropaleontol.*, 9, 419-440, 1985.

Montadert, L., D.G. Roberts, et al., *Initial Rep. Deep Sea Drilling Project*, 48, 1183pp, 1979.

Moullade, M., Etude stratigraphique and micropaléontologique du Crétacé inférieur de la "Fosse vocontienne". *Doc. Lab. Fac. Sci. Lyon*, 15, 1-369, 1966.

Müller, C., Calcareous nannofossils from the North Atlantic (Leg 48), *Initial Rep. Deep Sea Drilling Project*, 48, 589-639, 1979.

Müller, C., A. Schaaf, and J. Sigal, Biochronostratigraphy des formations d'âge Crétacé dans les forages du DSDP dans l'ocean Atlantique Nord, *Rev. Inst. Fr. Pet.*, 38, 683-708, 1983.

Müller, P.J. and Suess, E., Productivity, sedimentation rate and sedimentary organic carbon in the oceans. 1. Organic carbon preservation, *Deep Sea Res.*, 26A, 1347-1362, 1979.

Ogg, J.G., Early Cretaceous and Tithonian magnetostratigraphy of the Galicia Margin (Ocean Drilling Program Leg 103), *Proc. Ocean Drill. Prog. Sci. Res.*, 103, 659-682, 1988.

Okada, H., and H.R. Thierstein, Calcareous nannoplankton-Leg 43, Deep Sea Drilling Project, *Initial Rep. Deep Sea Drilling Project*, 43, 507-574, 1979.

Parker, M.E., M.A. Arthur, S.W. Wise, Jr., and C.R. Wenkham, Carbonate and organic carbon cycles in Aptian-Albian black shales at Deep Sea Drilling Project Site 511, Falkland Plateau, *Initial Rep. Deep Sea Drilling Project,* 71, 481-550, 1983.

Pedersen, T.F., and S.E. Calvert, Anoxia vs. productivity: what controls the formation of organic-carbon-rich sediments and sedimentary rocks?, *Am. Assoc. Petrol. Geol. Bull.*, 74, 454-466, 1990.

Pflaumann, U., and V.A. Krasheninnikov, Early Cretaceous planktonic foraminifers from eastern North Atlantic, DSDP Leg 41, *Initial Rep. Deep Sea Drilling Project*, 41, 539-564, 1977.

Pratt, L.M., Influence of paleoenvironmental factors on preservation of organic matter in middle Cretaceous Greenhorn Formation, Pueblo, Colorado, *Am. Assoc. Petrol. Geol. Bull.*, 68, 1146-1159, 1984.

Pratt, L.M., and J.D. King, Variable marine productivity and high eolian input recorded by rhythmic black shales in mid-Cretaceous pelagic deposits from Central Italy, *Paleoceanography*, 1, 507-522, 1986.

Premoli Silva, I., E. Erba, and M.E. Tornaghi, Paleoenvironmental signals and changes in surface fertility in mid-Cretaceous Corg-rich pelagic facies of the fucoid marls (central Italy), *Geobios, mém. spéc.*, 11, 225-236, 1989.

Proto-Decima, F., F. Medizza, and L. Todesco, Southeastern Atlantic Leg 40 calcareous nannofossils, *Initial Rep. Deep Sea Drilling Project*, 40, 571-634, 1978.

Raup, D.M., and J.J Sepkoski, Mass extinction in the marine fossil record, *Science*, 215, 1510-1503, 1982.

Raymont, J.E.G., *Plankton and Productivity in the Oceans, 1, Phytoplankton*, Pergamon, Oxford, 489 pp, 1980.

Rhoads, D.C., and J.W. Morse, Evolutionary and ecologic significance of oxygen-deficient basins, *Lethaia*, 4, 413-428, 1971.

Roth, P.H., Calcareous nannofossils, Leg 17, Deep Sea Drilling Project, *Initial Rep. Deep Sea Drilling Project*, 17, 695-795, 1973.

Roth, P.H., Cretaceous nannoplankton biostratigraphy and oceanography of the northwestern Atlantic Ocean, *Initial Rep. Deep Sea Drilling Project*, 44, 731-759, 1978.

Roth, P.H., Mid-Cretaceous calcareous nannoplankton from the central Pacific: implications for paleoceanography, *Initial Rep. Deep Sea Drilling Project*, 62, 471-489, 1981.

Roth, P.H., Mesozoic calcareous nannofossil evolution: relation to paleoceanographic events, *Paleoceanography*, 2, 601-612, 1987.

Roth, P.H., Ocean Circulation and calcareous nannoplankton evolution during the Jurassic and Cretaceous, *Paleogeog. Paleoclimatol., Paleoecol.*, 74, 111-126, 1989.

Roth, P.H., and K.P. Krumbach, Middle Cretaceous calcareous nannofossil biogeography and preservation in the Atlantic and Indian Oceans: Implications for paleoceanography, *Mar. Micropaleontol.*, 10, 235-266, 1986.

Sageman, B.B., P.B. Wignall, and E.G. Kauffman, Biofacies models for organic-rich facies: tool for paleoenvironmental analysis, in *Cycles and Events in Stratigraphy*, edited by G. Einsele, W. Ricken, and A. Seilacher, Springer Verlag, pp. 542-564, 1991.

Salloway, J.C., Paleomagnetism of sediments from Deep Sea Drilling Project Leg 71, *Initial Rep. Deep Sea Drilling Project*, 71, 1073-1091, 1983.

Sarnthein, M., K. Winn, J.-C. Duplessy, and M.R. Fontugne, Global variations of surface ocean productivity in low and mid latitudes: influence on CO_2 reservoirs of the deep ocean and atmosphere during the last 21,000 years, *Paleoceanography*, 3, 361-399, 1988.

Savrda, C.E., D.J. Bottjer, and D.S. Gorsline, Development of a comprehensive oxygen-deficient marine biofacies model: evidence from Santa Monica, San Pedro, and Santa Barbara Basins, California Continental Borderland, *Am. Assoc. Petrol. Geol. Bull.*, 68, 1179-1192, 1984.

Schlanger, S.O., High frequency sea-level fluctuations in Cretaceous time: an emerging geophysical problem, in K.J. Hsu (ed.) *Mesozoic and Cenozoic Oceans, AGU Geodynamics Series*, vol. 15, pp. 61-74, 1986

Schlanger, S.O., and R.G. Douglas, The pelagic ooze-chalk-limestone transition and its implications for marine stratigraphy, *Spec. Publ. Int. Assoc. Sediment.*, 1, 117-148, 1974.

Schlanger, S.O., and H.C. Jenkyns, Cretaceous oceanic anoxic events: causes and consequences, *Geol. en Mijnbouw*, 55, 179-184, 1976.

Schlanger, S.O., H.C. Jenkyns, and I. Premoli Silva, Volcanism and vertical tectonics in the Pacific basin related to global Cretaceous transgression, *Earth Planet. Sci. Lett.*, 52, 435-449, 1981.

Schlanger, S.O., M.A. Arthur, H.C. Jenkyns, and P.A. Scholle, The Cenomanian-Turonian oceanic anoxic event, I. Stratigraphy and distribution of organic carbon-rich beds and the marine $\partial^{13}C$ excursion, in *Marine Petroleum Source Rocks, Spec. Publ.*, vol. 24, edited by J. Brooks and A. Fleet, Geological Society of London, pp. 347-375, 1987.

Scholle, P.A., and M.A. Arthur, Carbon isotope fluctuations in Cretaceous pelagic limestones: Potential stratigraphic and petroleum exploration tool, *Am. Assoc. Petrol. Geol. Bull.*, 64, 67-87, 1980.

Shafik, S., Calcareous nannofossils from the Toolebuc, Eromanga Basin, Australia, *BMR Jour. of Australian Geology and Geophysics*, 9, 171-181, 1985.

Sibuet, J.-C., W.B.F. Ryan, et al., *Initial Rep. Deep Sea Drilling Project*, 47B, 787pp, 1979.

Siesser, W., Calcareous nannofossils: Legs 51 and 52 of Deep Sea Drilling Project, *Initial Rep. Deep Sea Drilling Project*, 51, pt. B, 823-845, 1980.

Sigal, J., Essai de zonation du Crétacé méditerranéen à l'aide des foraminifères planctoniques, *Geol. Medit.*, 4, 99-108, 1977.

Sigal, J., Chronostratigraphy and ecostratigraphy of Cretaceous formations recovered on DSDP Leg 47B, Site 398, *Initial Rep. Deep Sea Drilling Project*, 47, pt. B, 287-327, 1979.

Sliter, W.V., Aptian anoxia in the Pacific Basin, *Geology*, 17, 909-91, 1989a.

Sliter, W.V., Biostratigraphic zonation for Cretaceous planktonic foraminifers examined in thin section, *Jour. Foram. Res.*, 19, 1-19, 1989b.

Sliter, W.V., Cretaceous hiatuses, black shales and Pacific paleoceanography, *Geol. Soc. Am. Abs. with Prog.*, 23, A341, 1991.

Sliter, W.V., and T.J. Bralower, USGS Open File report, in press.

Tappan, H., *The Paleobiology of Plant Protists*, Freeman,1028pp, 1980.

Tappan, H., and A.R. Loeblich, Jr., Evolution of the Oceanic Plankton, *Earth Sci. Rev.*, 9, 207-240, 1973.

Tarduno, J.A., W.V. Sliter, T.J. Bralower, M. McWilliams, I. Premoli Silva, and J.G. Ogg, M-sequence reversals recorded in DSDP Sediment Cores from the Western Mid Pacific Mountains and Magellan Rise, *Geol. Soc. Am. Bull.*, 101, 1306-1316, 1989.

Tarduno, J.A., W.V. Sliter, L. Kroenke, R.M. Leckie, H. Mayer, J.J. Mahoney, R. Musgrave, M. Storey, and E.L. Winterer, Rapid formation of Ontong Java Plateau by Aptian mantle volcanism, *Science*, 254, 399-403, 1991.

Thiede, J., and T.H. van Andel, The paleoenvironment of anaerobic sediments in the late Mesozoic South Atlantic, *Earth Planet. Sci. Lett.*, 33, 301-309, 1977.

Thierstein, H.R., Tentative Lower Cretaceous calcareous nannoplankton zonation, *Eclog. Geol. Helv.*, 64, 459-488, 1971.

Thierstein, H.R., Lower Cretaceous calcareous nannoplankton biostratigraphy, *Abhandlung. Geolog. Bund.*, 29, 3-53, 1973.

Thierstein, H.R., Calcareous nannoplankton-Leg 26, Deep Sea Drilling Project, *Initial Rep. Deep Sea Drilling Project*, 26, 619-688, 1974.

Thierstein, H.R., Paleoceanographic implications of organic carbon and carbonate distribution in Mesozoic deep-sea sediments, in M. Talwani, W. Hay, and W.B.F. Ryan (eds.), *Deep Drilling in the Atlantic Ocean; Continental Margins and Paleoenvironment*, Maurice Ewing Series 3, Am. Geophys. Union, Washington, D.C., pp. 249-274, 1979.

Thierstein, H.R., Selective dissolution of Late Cretaceous and Earliest Tertiary calcareous nannofossils, experimental evidence, *Cretaceous Res.*, 2,165-176, 1980.

Thierstein, H.R., and W.H. Berger, Injection events in ocean history, *Nature*, 276, 461-466, 1978.

Thurow, J., Diagenetic history of Cretaceous radiolarians, North Atlantic Ocean (ODP Leg 103 and DSDP Holes 398D and 603B), *Proc. Ocean Drill. Prog. Sci. Res.*, 103, 531-555, 1988.

Tornaghi, M.E., I. Premoli Silva, and M. Ripepe, Lithostratigraphy and planktonic foraminiferal biostratigraphy of the Aptian-Albian "Scisti a Fucoid" in the Piobbico core, Marche, Italy: background for cyclostratigraphy, *Riv. It. Paleont. Strat.*, 95, 223-264, 1989.

Tucholke, B.E., P.R. Vogt, et al., *Initial Rep. Deep Sea Drilling Project*, 43, 1115pp, 1979.

Vail, P.R., R.M. Mitchum, Jr., R.G. Todd, J.M. Widmier, S. Thompson, III, J.B. Sangree, J.N. Bubb, and W.G. Hatfield, Seismic stratigraphy and global changes in sea level, in, C.E. Payton, (ed.), *Seismic Stratigraphy-Applications to Hydrocarbon Exploration, Am. Assoc. Petrol. Geol. Mem.*, 26, 49-212, 1977.

Vogt, P.R., Volcanogenic upwelling of anoxic, nutrient-rich water: a possible factor in carbonate-bank/reef demise and benthic faunal extinctions?, *Geol. Soc. Am. Bull.*, 101, 1225-1245, 1989.

Waples, D.W., A reappraisal of anoxia and richness of organic material, with emphasis on the Cretaceous North Atlantic, *Initial Rep. Deep Sea Drilling Project*, 80, 999-1016, 1983.

Watkins, D.K., Calcareous nannofossil paleoceanography of the Cretaceous Greenhorn Sea, *Geol. Soc. Am. Bull.*, 97, 1239-1249, 1986.

Wei, K.-Y., and J.P. Kennett, Taxonomic evolution of Neogene planktonic foraminifera and paleoceanographic relations, *Paleoceanography*, 1, 67-84, 1986.

Weissert, H., C-isotope stratigraphy, a monitor of paleoenvironmental change: a case study from the early Cretaceous, *Surveys in Geophysics*, 10,1-61, 1989.

Wiegand, G.E., Cretaceous nannofossils from the Northwest African margin, Deep Sea Drilling Project Leg 79, *Initial Rep. Deep Sea Drilling Project*, 79, 563-578, 1984.

Williams, G.L., and J.P. Bujak, Mesozoic and Cenozoic dinoflagellates, in, *Plankton Stratigraphy*, H.M. Bolli, J.B. Saunders, and K. Perch-Nielsen, (eds.), Cambridge University Press, 847-964, 1985.

Winterer, E.L., J.I. Ewing, et al., *Initial Rep. Deep Sea Drilling Project*, 17, 930pp, 1973.

Wise, S.W., Jr., Mesozoic and Cenozoic calcareous nannofossils recovered by Deep Sea Drilling Project Leg 71 in the Falkland Plateau Region, Southwest Atlantic Ocean, *Initial Rep. Deep Sea Drilling Project*, 71, 481-550, 1983.

Timothy J. Bralower, Department of Geology, University of North Carolina, Chapel Hill, NC 27599-3315.

William V. Sliter, Branch of Paleontology and Stratigraphy, USGS, Menlo Park, CA 94025.

Michael A. Arthur, Department of Geosciences, Pennsylvania State University, University Park, PA 16802.

R. Mark Leckie, Department of Geology, University of Massachusetts, Amherst, MA 01003.

David Allard, Graduate School of Oceanography, University of Rhode Island, Narragansett, RI 02882.

Seymour O. Schlanger, Department of Geological Sciences, Northwestern University, Evanston, IL 60201.