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2. CRETACEOUS PLANKTONIC FORAMINIFERS FROM SITES 865, 866, AND 869: A SYNTHESIS OF CRETACEOUS PELAGIC SEDIMENTATION IN THE CENTRAL PACIFIC OCEAN BASIN[1]

William V. Sliter[2]

ABSTRACT

Cretaceous planktonic foraminifers were recovered from Sites 865, 866, and 869 during Ocean Drilling Program (ODP) Leg 143 in the western central Pacific Ocean Basin. Rare planktonic foraminifers from the shallow-water platform limestone at Site 865 on Allison Guyot and Site 866 on Resolution Guyot range from possible Aptian to Albian in age in agreement with ages provided by benthic foraminifers. The oldest overlying Cretaceous pelagic sediment is early to middle Turonian in age and was recovered from a cavity in the platform limestone at Site 865. Manganese-encrusted pebbles at the top of the Cretaceous pelagic cap contain rare, early Maastrichtian planktonic foraminifers. Possible emergence and/or drowning of the edifice at Allison Guyot, and presumably also at Resolution Guyot, took place during or shortly after the late Albian and prior to deposition of the Turonian sediments. At Site 869 on the archipelagic apron in the Marshall Islands, planktonic foraminifers date a sequence of volcaniclastic deposits as late Cenomanian to Maastrichtian in age. The sequence contains major pulses of turbiditic sedimentation in the Cenomanian and Campanian and smaller pulses in the Turonian, Coniacian, and Maastrichtian.

The record of sedimentation from Leg 143 compared to that from selected Deep Sea Drilling Project (DSDP) sites in the central Pacific Ocean Basin shows a similar pattern of pelagic sedimentation and hiatuses. Further, the pattern is similar for sites from topographic highs and deep basins. Pelagic sediments are preserved in the Turonian, Coniacian, upper Santonian to lower Campanian, and upper Campanian to middle Maastrichtian sequences. Hiatuses occur in the lower-to-middle Cenomanian, Turonian, lower Santonian, lower-to-middle Campanian, and upper Maastrichtian sections.

The transition from Albian limestone to younger chalk at pelagic sites in the Mid-Pacific Mountains and southern Hess Rise takes place in the uppermost Albian analogous to that of the platform carbonate rocks where it also is associated with a hiatus. Correlation of this transition to similar events in the Tethys and Atlantic Ocean indicates that the demise of carbonate platform sedimentation in the western Pacific is related to a change in sea level in the late Albian *Rotalipora appenninica* Zone; either a brief, eustatic regressive-transgressive cycle or the initiation of platform drowning.

The regional hiatuses are ascribed to times of intensified ocean currents. In the mid-Cretaceous the intensified currents are linked to the production of warm, saline bottom-water during sea-level highstands. Increased upwelling and the rise of the calcite compensation depth during these times produced variations in dissolution intensity at intermediate water-depths and hiatuses in pelagic sedimentation. Pelagic buildups during the intervening quiescent intervals supplied the material that was later redeposited in the basins by mass wasting or subsequent mechanical erosion. The variations in dissolution intensity are likely linked to climatic oscillations. Thus, hiatuses represent warmer periods, whereas, pelagic accumulation correlates with cooler periods. The Late Cretaceous climatic oscillations apparently represent an extension of climate cycles begun in the mid-Cretaceous "greenhouse" state.

INTRODUCTION

Age-diagnostic Cretaceous planktonic foraminifers were recovered primarily from pelagic and hemipelagic sediments at Sites 865, 866, and 869 during Ocean Drilling Program (ODP) Leg 143 in the western central Pacific Ocean Basin (Fig. 1). The cruise was planned to investigate the age and origin of Cretaceous guyots in the western central Pacific Basin and, more specifically, the age of volcanic edifice building and history of carbonate platform development. In this context, the history of pelagic sedimentation in the Cretaceous sequence, as defined by the distribution of planktonic foraminifers, provides a critical record of platform drowning and the timing of fluctuations in sea level and erosional events.

Three of the four localities drilled during Leg 143 were targeted to recover Cretaceous sediments: two from the summits of drowned guyots (Site 865 from Allison Guyot; Sites 866 and 867/868 from Resolution Guyot) and one from the archipelagic apron adjacent to a paired atoll-guyot in the Marshall Islands (Site 869). The fourth locality (Site 870) was an engineering drilling test in Anewetak lagoon. Drilling at the two guyots penetrated the pelagic caps and underlying shallow-water carbonate platforms and reached the volcanic pedestal at Site 866 on Resolution Guyot. Cretaceous planktonic foraminifers were recovered primarily from the Cretaceous part of the pelagic cap at Sites 865 and the deep-water apron sediments at Site 869. Only very rare, poorly preserved, and often questionably identified specimens were found in the platform carbonate rocks at Sites 865 and 866.

Site 865 is located atop Allison Guyot (18°26.41′N, 179°33.34′W) in the Mid-Pacific Mountains at a water depth of about 1518 m (Fig. 1). Three holes were drilled, and Cretaceous planktonic foraminifers were recovered from Holes 865A and 865B. The Cretaceous pelagic material consists of both sparse, mineralized pebbles and cavity fills in the underlying shallow-water limestone. In addition, very rare planktonic foraminifers were identified from slightly more marine intervals in the predominantly Albian limestone at Hole 865A.

At Site 866 on Resolution Guyot in the Mid-Pacific Mountains (Fig. 1), very rare pieces of manganese-encrusted limestone of Cretaceous age also were recovered from Maastrichtian to Pliocene nannofossil-foraminiferal ooze, but these were not studied for foraminifers. Very rare planktonic foraminifers were identified in the underlying shallow-water platform carbonate deposits. Site 866, located at 21°19.95′N, 174°18.84′E at a water depth of 1346 m, and the adjacent Sites 867/868 were drilled as part of a transect across the perimeter mound on the north side of Resolution Guyot. No Cretaceous planktonic foraminifers were recovered from the latter sites.

Site 869 is located at 11°00.01′N, 164°44.97′E, at a water depth of 4827 m, 45 nmi southwest of the atoll-guyot pair, Pikinni Atoll and Wodejebato Guyot, in the northern Marshall Islands. Two holes were drilled, and Cretaceous planktonic foraminifers were recovered from

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[1] Winterer, E.L., Sager, W.W., Firth, J.V., and Sinton, J.M. (Eds.), 1995. *Proc. ODP, Sci. Results,* 143: College Station, TX (Ocean Drilling Program).

[2] U.S. Geological Survey, 345 Middlefield Road, MS 915, Menlo Park, CA 94025, U.S.A.

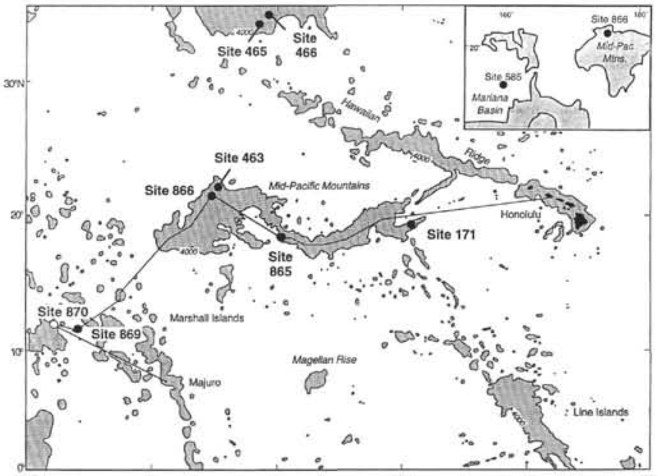

Figure 1. Location of Leg 143 drill sites and nearby DSDP sites that contain Cretaceous planktonic foraminifers (black circles). Shaded areas are shallower than 4 km. Line shows track of *JOIDES Resolution.* Inset: Location of DSDP Site 585 in the East Mariana Basin. Shaded areas are shallower than 5 km.

about 588 m of volcaniclastic materials interlayered with nannofossil and radiolarian claystone at Hole 869B.

Previous studies in the western Pacific have documented all or parts of the Cretaceous pelagic sequence that rests characteristically at shallower sites on truncated shallow-water limestone of Albian age (e.g., Hamilton, 1953, 1956; Heezen et al., 1973; Winterer, Ewing et al., 1973; Matthews et al., 1974; Lincoln et al., 1993) or at deeper sites on pelagic limestone at the limestone/chalk transition of Albian age (Thiede, Vallier, et al., 1981; Grötsch and Flügel, 1992; van Waasbergen, 1993). In addition, hiatuses of both regional and widespread extent have been identified in the more continuous Cretaceous sequences (Douglas, 1971, 1973; Thiede, Vallier, et al., 1981; Sliter, 1992). Thus, one of the important tasks of Leg 143 was to place the Cretaceous pelagic cores and rock fragments in a regional chronostratigraphic framework and compare the results to the previous record from the western central Pacific Ocean.

Here, my purpose is (1) to document the occurrence of Cretaceous planktonic foraminifers at Sites 865, 866, and 869, (2) to compare these occurrences with others at sites in the western Pacific, and (3) to derive a history of pelagic sedimentation. Stratigraphic distributions are shown in Tables 1 through 3; selected planktonic and benthic species are illustrated in Plates 1 through 3.

METHODS

Samples throughout the Cretaceous sequence of Sites 865, 866, 867/868, and 869 were examined for planktonic foraminifers using thin-section and whole specimen techniques, depending on the lithology. The stratigraphic distribution of planktonic foraminifers in Tables 1 through 3 is shown by their presence or absence. For examination of thin sections, the relative abundance was estimated by scanning the section at a magnification of ×150 using the following scale:

A = abundant (>20% of the total assemblage),
C = common (10%),
F = few (5%),
R = rare (1%), and
VR = very rare (<1% or 1–3 specimens).

Preservation was determined qualitatively, based on the degree of dissolution and/or overgrowth exhibited by the planktonic foraminifer examined in thin section using the following scale:

G = good (individual specimens exhibit little dissolution or recrystallization);
M = moderate (dissolution and recrystallization (overgrowth) of some individuals is apparent; species identifications are usually not impaired);
P = poor (specimens show strong dissolution and/or recrystallization; many specimens are represented by relics or "ghosts" that make identifications impossible); and
VP = very poor (most specimens are represented by ghosts; species identifications are rare to very rare; generic identifications may be rare).

For whole specimens, most samples were disaggregated in a hot solution of hydrogen peroxide and Calgon. Samples were washed over a 43-μm sieve and dried under a heat lamp. The relative abundance of planktonic foraminifers was based on a visual examination of the >43-μm residue using the following scale:

A = abundant (>50 specimens),
C = common (26–50),
F = few (11–25),
R = rare (3–10), and
VR = very rare (1–3).

Preservation of whole specimens was determined as follows:

G = good (little or no evidence of overgrowth, dissolution, or abrasion);
M = moderate (calcite overgrowth, dissolution, or abrasion are common but minor; species identifications usually not impaired);
P = poor (substantial overgrowth, dissolution, or fragmentation; species or generic identifications often impaired).

Sample notation follows the standard ODP format, given as drilling leg followed by site, core, core type (R = rotary, X = extended core barrel), section, and interval (centimeters below top of section). The abbreviation "CC" refers to a core-catcher sample taken from the bottom of the cored interval.

BIOSTRATIGRAPHY

The biozonation used is that of Caron (1985), modified by Sliter (1989), for Cretaceous planktonic foraminifers examined in thin section (Fig. 2). Use of the standard zones have been retained where possible and have been abbreviated by the numerical zone notation KS. The zonal scheme has been correlated to the magneto-geochronology of Harland et al. (1990).

Hole 865A

Two age groups of Cretaceous planktonic foraminifers were identified in thin section at Hole 865A (Table 1). The older group is composed mostly of poorly preserved, fragmented, and questionably identified species found in the more open-marine layers of the shallow-water platform limestone that were ascribed to the upper Albian (Sager, Winterer, Firth, et al., 1993). The majority of these forms below Core 143-865A-28R are now thought to be benthic foraminifers and of questionable value for dating the sequence. The rare planktonic specimens in and above Core 143-865A-28R are retained in the upper Albian. A specimen from Sample 143-865A-79R-1, 32–33 cm, is important as it suggests an older age near the Aptian/Albian boundary. The specimen illustrated in Plate 3, Figure 10, most closely resembles specimens from the *Hedbergella praetrocoidea–H. trocoidea* lineage or perhaps a juvenile *H. trocoidea.* Regardless of the taxonomic interpretation, the suggested age near the Aptian/Albian boundary agrees closely with the independent determination of Jenkyns et al. (this volume) based on strontium-isotope ratios.

Table 1. Occurrence of Cretaceous planktonic foraminifers, Hole 865A.

Stage	Core, Section, Interval (cm)	Sample Type	Abundance	Preservation	*Hedbergella* sp.	*H. trocoidea*	*Praeglobotruncana stephani*	*Heterohelix globulosa*	*Pseudotextularia elegans*	*Globotruncana linneiana*	*G. mariei*	*Heterohelix striata*	*Globotruncanella petaloidea*	*Hedbergella holmdelensis*	*Globigerinelloides bollii*	*Globotruncanita stuartiformis*	*Laeviheterohelix pulchra*	*Rugoglobigerina rugosa*	*Globotruncanita stuarti*	*Globotruncana bulloides*	*G. hilli*	*Contusotruncana plummerae*
	143-865A-																					
Maastrichtian	16R-CC	T	R	P				X	X	X				X		X			?			
	20R-CC	T	F	P				X	X	cf	X	X	X	X	X	cf	X	cf	cf	cf	X	cf
Albian	28R-CC	T	VR	P			?															
Aptian/Albian	79R-1, 32-33	T	VR	P		cf																
	88R-1, 36-38	T	VR	P	X																	

Notes: Sample: T= thin section. Abundance: VR= very rare, R= rare, F= few. Preservation: P= poor. cf= probable identification. ?= questionable identification.

Table 2. Occurrence of Late Cretaceous planktonic foraminifers, Hole 865B.

Stage	Core, Section, Interval (cm)	Sample Type	Abundance	Preservation	*Helvetoglobotruncana helvetica*	*Whiteinella baltica*	*W. aprica*	*Heterohelix reussi*	*Praeglobotruncana gibba*	*Hedbergella delrioensis*	*Dicarinella imbricata*	*D. hagni*	*Marginotruncana sigali*	*M. renzi*	*M. pseudolinneiana*	*Hedbergella* sp.	*Heterohelix* sp.	*Globigerinelloides* sp.
	143-865B-																	
Santonian/ Maastrichtian	16X-CC, 0-5	T	R	P												X	X	X
Turonian	17X-CC, 14-16	W	A	M	X	X	X	X	X	X	X	X	X	X	X			

Notes: Abbreviations as in Table 1, with the addition of the following: Sample: W=whole specimen. Abundance: A=abundant. Preservation: M= moderate.

The second group of planktonic foraminifers was limited to several isolated manganese-encrusted pebbles found at the bottom of Cores 143-865A-16R and -20R, the latter undoubtedly representing cavings from above. This assemblage includes *Globotruncanella petaloidea* and specimens tentatively referred to as *Globotruncanita stuarti* and *Ruguglobigerina rugosa*, among others, that identify a Maastrichtian age. These taxa and the absence of younger species suggest an age equivalent to the early Maastrichtian *Globotruncana aegyptiaca* Zone (KS 29). Figure 3 is a graphic representation of the planktonic foraminiferal age distribution at Hole 865A.

Hole 865B

Two age groups of Cretaceous planktonic foraminifers again were identified in Hole 865B; however, both are younger than the late Albian shallow-water carbonate deposits (Table 2). No planktonic foraminifers were found in the limited limestone samples from Hole 865B. The older assemblage consists of whole specimens washed from brown micrite cavity fill in the upper Albian limestone of Sample 143-865B-17X-CC, 14–16 cm (Fig. 4). This assemblage contains abundant, moderately well-preserved foraminifers that include the nominate species of the early to middle Turonian *Helvetoglobotruncana helvetica* Zone (KS 21). The age of this sample is interesting, as it represents the oldest pelagic material recovered from either Allison or Resolution guyots (Fig. 3). The second assemblage of planktonic foraminifers is limited to rare, poorly preserved specimens represented by ghosts in manganese-encrusted limestone fragments from the bottom of Core 143-865B-16X. Although the specimens resemble species that first appear in the upper Santonian and range into the Maastrichtian (as shown in Fig. 3), they most likely represent the age equivalent of the Maastrichtian assemblage found in Hole 865A.

Hole 866A

Cretaceous planktonic foraminifers at Site 866 are limited to rare, poorly preserved specimens from the shallow-water platform carbon-

Table 3. Occurrence of Late Cretaceous planktonic foraminifers, Hole 869B.

Stage	Core, Section, Interval (cm)	Sample Type	Preservation	Abundance	*Hedbergella* sp.	*Rotalipora greenhornensis*	*Dicarinella algeriana*	*Hedbergella planispira*	*Praeglobotruncana stephani*	*Whiteinella aprica*	*Hedbergella delrioensis*	*Heterohelix moremani*	*Globigerinelloides ultramicrus*	*Rotalipora deeckei*	*R. cushmani*	*Praeglobotruncana gibba*	*P. aumalensis*	*Dicarinella hagni*	*Whiteinella baltica*	*Heterohelix reussi*	*Dicarinella imbricata*	*Helvetoglobotruncana helvetica*	*Dicarinella canaliculata*	*Whiteinella brittonensis*	*Marginotruncana sigali*	*Whiteinella* sp.	*Marginotruncana schneegansi*	*M. pseudolinneiana*	*M. sinuosa*	*Dicarinella primitiva*	*Whiteinella praehelvetica*	*Globigerinelloides messinae*	*G. bollii*	*Heterohelix globulosa*	*Hedbergella holmdelensis*	*Heterohelix carinata*	*Globotruncana ventricosa*	*G. arca*	*Contusotruncana fornicata*	*Pseudotextularia elegans*	*Globotruncanita stuartiformis*	*Globotruncana mariei*	*Contusotruncana patelliformis*	*Globotruncana orientalis*	*Globotruncanita elevata*	*Globotruncana rosetta*	*G. bulloides*	*G. linneiana*	*Pseudoguembelina costulata*	*Heterohelix striata*	*Globotruncana subspinosa*	*Hedbergella monmouthensis*	*Rugoglobigerina* sp.	*Globotruncanita stuarti*	*Globotruncana aegyptiaca*	*Rugoglobigerina rugosa*	*R. hexacamerata*	*Laeviheterohelix dentata*
	143-869B-																																																									
Maastrichtian	10R-CC	W	P	C																												X	X		X			X	X	X	X	X	X			X		X	X			X		X	X	X	cf	X
	11R-1, 4-6	T	P	R																																																	cf					
	11R-3, 143-145	T	P	R																																												X										
	11R-CC	W	P	R																													X	X									cf				X			X								
	12R-CC	W	P	F																												X						X	X		X				X		X	X										
	13R-4, 43-44	W	P	F																													X	X				X	X		X	X					X		X	X	X	X	X					
	13R-4, 44.5-46	W	P	F																																		X	X		cf	X						X										
	13R-CC	W	P	F																													X					X			X							X	X	X								
	14R-CC	W	P	VR																																						cf																
Campanian	15R-CC	W	P	C																																		X	X		X	X		X														
	16R-CC	W	P	C																																		X	X		X	X	X			X	X											
	17R-CC	W	P	F																																			X			X	X			X												
	19R-CC	W	P	VR																																			cf		X				X													
	20R-CC	W	P	F																																		X	X		X	X		cf	X													
	22R-3, 26-27	T	P	R																															X																							
	23R-CC	W	P	A																																		X	X		X	X	X															
	25R-CC	W	P	R																														X			X	cf	X	X																		
San./Cam.	28R-CC	W	P	R																														X	X	X																						
Santonian	30R-1, 24-26	T	P	R				X												X												X	X	X	X																							
Coniacian	31R-1, 72-74	W	P	A		X				X					X	X			X	X	X		X	X	X		X	X	X	X	X																											
?	32R-1, 80-81	T	P	F																						X																																
Turonian	34R-1, 142-144	T	P	F			X			X	X	X				X			X	X	X	X	X	X	X																																	
	35R-1, 44-46	T	P	F			X			X	X					X		X	X	X	X	X	X																																			
	36R-3, 13-14	W	P	F		X			X					X	X	X	X	cf																																								
	37R-CC	W	P	R		X		X			X	X	X																																													
	44R-1, 127-128	W	P	F		X	X		X																																																	
	49R-1, 28-30	W	M	VR		X			X	X																																																
Cenomanian	50R-2, 80-82	T	P	R		cf	X	X																																																	
	52R-5, 12-15	T	P	R	X	cf																																																				
	57R-2, 122-125	T	P	R	X																																																					

Notes: Abbreviations as in Table 1, with the addition of the following: Abundance: C= common.

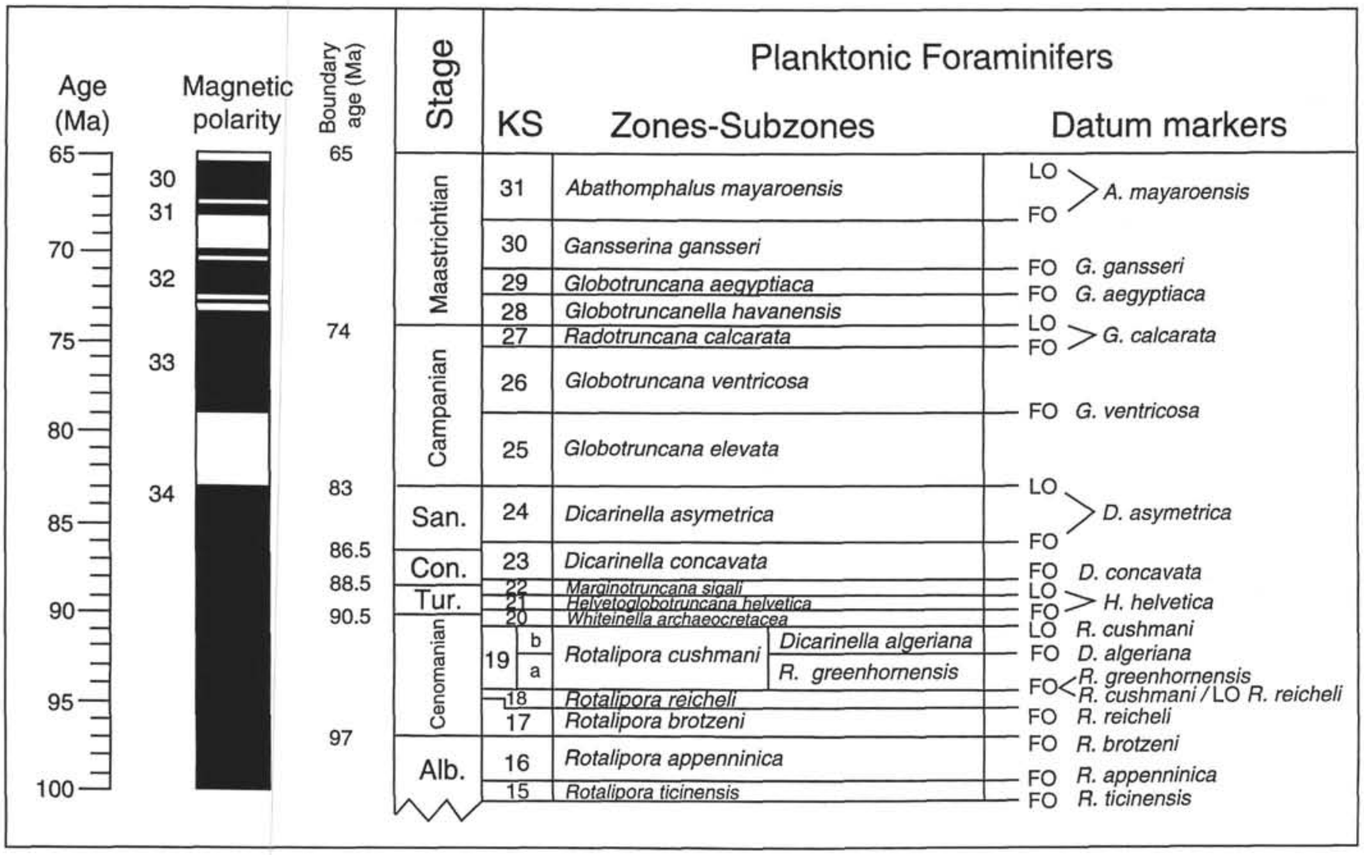

Figure 2. Zonation for Cretaceous planktonic foraminifers examined in thin section (after Sliter, 1989; modified from Caron, 1985). Magneto-geochronology after Harland et al. (1990). KS = Cretaceous zone notation; FO = first occurrence; and LO = last occurrence.

ate rocks identified in thin section. The specimens are mostly questionably identified and consist of small globular forms resembling hedbergellids. Perhaps the most convincing is the specimen from Sample 143-866A-36R-1, 2–4 cm, tentatively attributed to *Ticinella primula* and illustrated in Plate 3, Figure 12. The value of these questionable occurrences is less for age interpretation than for evidence of marine conditions.

Hole 869B

Cretaceous planktonic foraminifers were recovered from Cores 143-869B-10R to -57R and range from Cenomanian to Maastrichtian in age (Table 3). Preservation is generally poor owing to calcite dissolution, and the abundance and size of the specimens fluctuate considerably throughout the sequence. These fluctuations are attributed to the rapid emplacement of volcaniclastic turbidites that characterize the mode of sedimentation, especially at the base of the sequence. The supply of this material waned during the Turonian to early Campanian, compared to the background sedimentation of radiolarian claystone and siltstone, and then recommenced in the Campanian and Maastrichtian with the development of turbidites.

Planktonic foraminifers in Cores 143-869B-36R to -50R are representative of the *Dicarinella algeriana* Subzone (KS 19b) of the *Rotalipora cushmani* Zone of late Cenomanian age (Fig. 3). Present are the nominate species plus *Rotalipora greenhornensis, R. deeckei, Whiteinella aprica*, and *Praeglobotruncana gibba*, among others. Planktonic foraminifers become increasingly sparse downhole to Core 143-869B-57R and are then replaced by radiolarians and nannofossils that date the bottom of the hole as middle to late Cenomanian in age. The presence of a tentatively identified *Rotalipora greenhornensis* in Core-143-869B-52R indicates that the *Rotalipora cushmani* Zone extends at least to that depth.

Cores 143-869B-34R and -35R contain a Turonian assemblage assigned to the *Helvetoglobotruncana helvetica* Zone (KS 21) by the occurrence of the nominate species plus *Dicarinella canaliculata, Marginotruncana sigali*, and *Whiteinella baltica*, among others. A more abundant and diversified assemblage of early Coniacian age occurs in Core 143-869B-31R. Present are *Dicarinella primitiva, Marginotruncana schneegansi, M. pseudolinneiana*, and *Whiteinella brittonensis*, among others that identify the *Dicarinella primitiva* Zone of Caron (1985), here combined with the *Marginotruncana sigali* Zone (KS 23). Associated with these species are reworked middle to upper Cenomanian forms, such as *Rotalipora cushmani* and *R. greenhornensis*.

The first occurrence of small, size-sorted specimens of *Heterohelix globulosa, H. reussi, Globigerinelloides messinae*, and *G. bollii*, together with *Hedbergella planispira* in Core 143-869B-30R, are suggestive of a Santonian age. These are followed in Core 143-869B-28R by the occurrence of *Heterohelix carinata* that yields a late Santonian to early Campanian age. More normal size specimens were found in Core 143-869B-25R that include *Globotruncana ventricosa* and *Contusotruncana fornicata*, among others, which indicate a middle Campanian age (KS 26).

The Campanian assemblage increases in diversity uphole into Core 143-869B-12R with the addition of species such as *Globotruncanita stuartiformis, P. elegans, Globotruncana linneiana, G. bulloides*, and *Radotruncana subspinosa.* Planktonic foraminiferal assemblages from Core 143-869B-12R, instead, are rare, poorly preserved, and size sorted to smaller specimens. These specimens broadly indicate a late Campanian to Maastrichtian age, but were assigned to the late Cam-

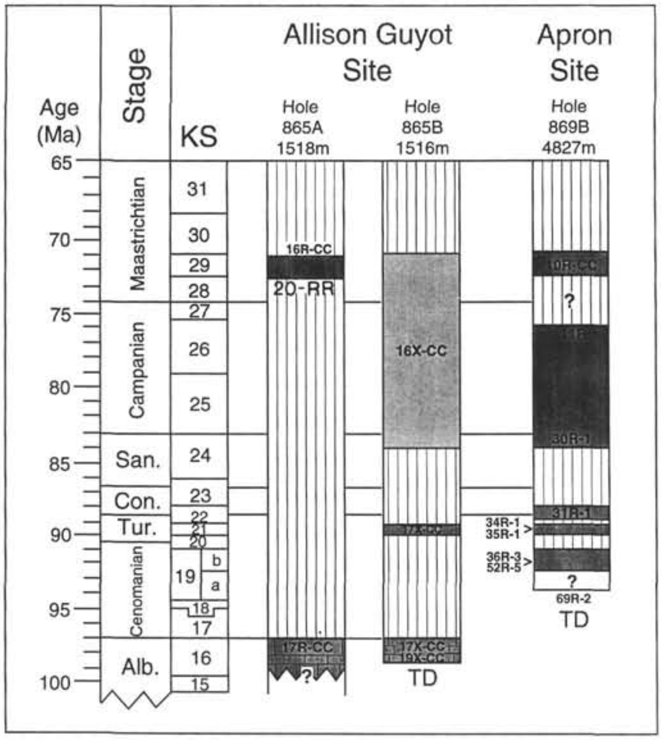

Figure 3. Age of Cenomanian to Maastrichtian sediments, dated by planktonic foraminifers from Leg 143. Albian shallow-water limestone shown by lithologic symbol, dated primarily by benthic foraminifers (Arnaud-Vanneau and Sliter, this volume). Zone assignment as in Figure 2. Dark stipple = dated to zone; light stipple = poorly dated interval. TD = total depth.

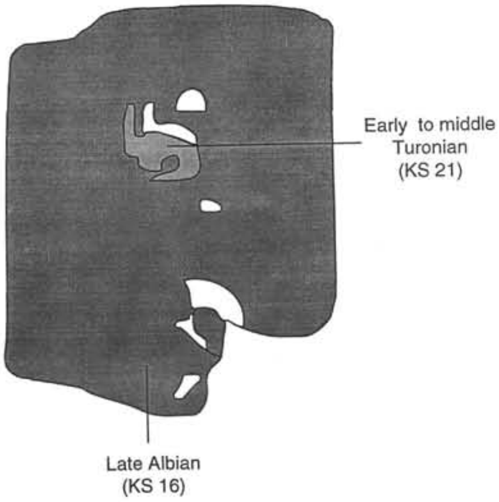

Figure 4. Sketch of core sample of late Albian shallow-water limestone from Section 143-865B-17X-CC showing several solution cavities (open areas) and location of early to middle Turonian cavity fill (light stipple) containing planktonic foraminifers.

panian on the basis of associated nannofossils (Sager, Winterer, Firth, et al., 1993). Maastrichtian planktonic foraminifers first occur in Core 143-869B-10R and include *Globotruncana aegyptiaca*, *Globotruncanita stuarti*, and possible *Rugoglobigerina hexacamerata*. This association in the absence of younger species indicates an early Maastrichtian age equivalent to the *Globotruncana aegyptiaca* Zone (KS 29).

HISTORY OF GUYOT/APRON PELAGIC SEDIMENTATION

The stratigraphic distribution of planktonic foraminifers from Leg 143 identifies two different histories of pelagic sedimentation from the guyot and apron sites (Fig. 3). At both Allison and Resolution guyots, sedimentation of the shallow-water carbonate materials ceased during the late Albian (Arnaud-Vanneau and Sliter, this volume). The precise timing of this occurrence at the two guyots is poorly constrained, however, dredges from the slope of Allison Guyot recovered planktonic foraminifers from the latest Albian *Rotalipora appenninica* Zone mixed with shallow-water debris (Grötsch and Flügel, 1992; Winterer et al., 1993; van Waasbergen, 1993). The irregular platform surface and the presence of apparent solution cavities at both guyots suggests karstification but the nature of the dissolution remains ambiguous (Sager, Winterer, Firth, et al., 1993). Whereas the depth of surface irregularities at both guyots and the presence of speleothems in cavities at Site 867 on Resolution Guyot suggests subaerial exposure during a subsequent decline in sea level or elevation of the regional seafloor (Winterer and Metzler, 1984), the absence of blocky meteoric water cements at both guyots and the increase in more normal/less restricted marine biofacies in the uppermost platform sediments at Site 866 imply platform drowning and possible submarine dissolution.

Following submergence, the presence of two generations of pelagic sediments provide a limited record of the post-drowning history. Planktonic foraminifers of early to middle Turonian age (about 89–90 Ma), in solution cavities at the platform surface, identify the oldest pelagic sediments. A second generation of pelagic sediments is indicated by lower Maastrichtian species in the mineralized limestone crust at the very top of the Cretaceous sequence. Both the limited record and the considerable hiatuses in the pelagic sequence suggest long periods of erosion and/or nondeposition, likely related to periods of accelerated current winnowing and past changes in productivity. Although the record is broadly similar to that of the Pacific Ocean Basin (Sliter, 1992), the extended duration of the hiatuses perhaps was augmented by the local subsidence history and by the northward plate motion at Allison and Resolution guyots that resulted in contact with current regimes in different water masses.

At Site 869 on the archipelagic apron, sedimentation of planktonic foraminifers shows a close correspondence to the rate and type of turbiditic sedimentation in the volcaniclastic sequence. Planktonic foraminifers were rare in the rapidly deposited coarse grain and debris flows of the Cenomanian, increased in abundance as sedimentation rates decreased during the Turonian to Santonian, and decreased in abundance as sedimentation rates increased during the Campanian (Table 3). The assemblages of planktonic foraminifers, usually confined to the more calcareous turbiditic layers in the volcaniclastic sequence, are associated with radiolarians, rare redeposited shallow-water material, and in-situ abyssal benthic foraminifers (Pl. 2, Fig. 16). In general, the influx of redeposited shallow-water material increases upsection from the upper Cenomanian, whereas the occurrence of radiolarians decreases.

Below Core 143-869B-57R, planktonic foraminifers are absent, and pelagic assemblages consist of radiolarians and nannofossils. The exclusion of planktonic foraminifers in the presence of well-preserved calcareous nannofossils likely indicates size sorting in the turbiditic environment, rather than differential preservation. The lack of shallow-water material at the bottom of the hole suggests that the shallow-water platform was not well established and was not exporting material in abundance or, alternatively though less likely, may reflect sediment dilution and/or a deeper-water volcanic source.

Sparse shallow-water material first occurs in Interval 143-869B-51R-4, 25–40 cm, represented by coalified woody fragments, followed in Sample 143-869B-49R-1, 28–30 cm, by a gastropod, orbitolinid foraminifer, and several small benthic foraminifers. Together, this material indicates the presence of one or more emergent volcanic islands. Redeposited material increases in the Turonian and includes ooids, coated grains, bivalve and echinoid fragments, sponge spicules, and smaller calcareous and agglutinated foraminifers. This assemblage continues into the Coniacian with the addition of gastropods, *Inoceramus* prisms, and ostracodes. Campanian samples contain a similar assemblage plus neritic and bathyal smaller calcareous benthic foraminifers (Pl. 1, Fig. 5) and bryozoans. In addition, orbitolinids were found in Samples 143-869B-10R-CC and -11R-1, 4–6 cm, at the Campanian/Maastrichtian boundary.

The pattern of pelagic sedimentation at the three sites, although meager at the guyot sites and strongly overprinted by local volcanism at the apron site, shows a general correlation in the Turonian to Santonian and in the Maastrichtian. The major difference between the sites is the local volcanic events at Site 869 that flooded the site with turbidites during Cenomanian and Campanian time.

CRETACEOUS PELAGIC SEDIMENTATION IN THE WESTERN CENTRAL PACIFIC OCEAN BASIN

Comparison of the stratigraphic record from DSDP sites in the western central Pacific Ocean shows a broad similarity in the pattern of sedimentation. The record is interrupted by several hiatuses of apparent regional extent (Sliter, 1992) that represent breaks in the stratigraphic sequence owing to erosion, dissolution, or nondeposition. Correlation of these hiatuses is hampered by (1) poor dating in the indurated intervals, which with planktonic foraminifers previously relied on rare, whole specimens washed from marly or chalky residues, and (2) poor recovery associated with the presence of chert. Typically in the North Pacific, both lithification and silicification become more pronounced in the Turonian and older record, thus making precise correlations more difficult.

Four sites drilled on topographic highs in the general area are of particular interest (Fig. 1); Site 463 from the Mid-Pacific Mountains, drilled on the sediment apron flanking Resolution Guyot about 24 nmi northeast of the northeastern rim at a water depth of 2532 m; Site 171, drilled to the east on Horizon Guyot at a water depth of 2290 m; and Sites 465 and 466, drilled to the north on southern Hess Rise at water depths of 2161 and 2665 m, respectively. A fifth site, Site 585 in the East Mariana Basin drilled at a water depth of 6109 m, is included and has been compared with the record from Site 869 as it contains pulses of redeposited shallow-water material.

The Cretaceous sequence from Site 463 consists of Barremian to Aptian interbedded pelagic and bioclastic limestone at the bottom of the hole, overlain by Aptian tuffaceous and carbonaceous limestone, followed by pelagic limestone and chalk of Albian to Maastrichtian age. Redeposited platform debris in the basal bioclastic limestone includes ooids, bivalve and echinoid fragments, rare glauconite, and basalt fragments presumably derived from Resolution Guyot.

Several hiatuses were identified in the Cenomanian, Santonian, Campanian, and Maastrichtian (Fig. 5), based on the distribution of planktonic foraminifers and calcareous nannofossils (Boersma, 1981; Čepek, 1981). The limestone/chalk transition between Cores 62-463-52 and -50 takes place in the upper Albian *Rotalipora appenninica Zone (KS 16), based on the presence of Planomalina buxtorfi* and *Ticinella primula,* among others, in chalk from the latter core (Boersma, 1981). The lower Cenomanian (KS 17), identified in Core 62-463-49 by the presence of *Rotalipora evoluta* (=*R. brotzeni*), apparently extends through the highly disturbed chalk in Core 62-463-48, based on both planktonic foraminifers and nannofossils (Boersma, 1981; Čepek, 1981). A hiatus was placed between the latter core and Core 62-463-44, which contains *R. greenhornensis of the middle to upper Cenomanian R. cushmani* Zone (KS 19).

The sequence apparently is continuous from the upper Cenomanian to the Coniacian in Core 62-463-27, based on the presence of zonal indicator species. A hiatus that involves the lower Santonian is indicated by the occurrence of an upper Santonian assemblage in Core 62-463-26 that includes *Dicarinella asymetrica, Globotruncanita elevata*, and *Globigerinelloides prairiehillensis*. Core 62-463-25 contains a mixed assemblage of upper Santonian and lower Campanian species and is separated by a major hiatus from the upper Campanian in Core 62-463-24, which is identified by the first occurrence of *Radotruncana calcarata* (KS 27). From Core 24 upsection, the sequence appears complete from the upper Campanian through the middle Maastrichtian (KS 30).

A similar record is apparent at Site 171 on Horizon Guyot (Fig. 1). The sedimentary sequence at the base of the hole consists of limestone, with shallow-water biogenic material presumably resting on vesicular basalt, represented by several pieces in Core 17-171-33. The limestone is mostly devoid of pelagic organisms; however, a few planktonic foraminifers recovered from sand-sized drilling debris of friable limestone at the top of the limestone sequence in Sample 17-171-29-CC by Douglas (1973) included *Hedbergella brittonensis* (=*Whiteinella brittonensis*) and indicate an age no older than late Cenomanian (KS 19b). It is also possible that these foraminifers represent downhole contamination. Based on this latter interpretation and the record from Site 463, the shallow-water limestone at the base of the hole, previously considered to be questionably Cenomanian in age (Winterer, Ewing, et al., 1973), may represent the widespread Albian limestone.

Overlying the limestone in Cores 17-171-28 and -27 is an undated interval of volcanic-rich limestone, hyaloclastite, conglomerate with clasts of shallow-water limestone and basalt, and apparently weathered vesicular basalt flows. This interval is overlain by volcanic siltstone, with planktonic foraminifers assigned to the *Helvetoglobotruncana helvetica* Zone (KS 21) of early-to-middle Turonian age. No nannofossils were reported below the Turonian interval that begins in Core 17-171-26 (Roth, 1973), lending support to a hiatus between the shallow-water limestone at the base of the hole and the volcanic siltstone facies. A hiatus between Cores 17-171-22 and -21 is indicated by both planktonic foraminifers (Douglas, 1973) and nannofossils (Roth, 1973) that corresponds to the interval from the Coniacian to early Santonian *Dicarinella concavata* Zone (KS 23) to the late Campanian *Radotruncana calcarata* Zone (KS 27). Above Core 17-171-21, the sequence is complete to the upper Maastrichtian *Abathomphalus mayaroensis* Zone (KS 31).

At Site 465 on southern Hess Rise, laminated, chert-rich pelagic limestone rests on trachyte that shows possible evidence of subaerial extrusion. The laminated limestone, having laminae composed of concentrations of radiolarians and planktonic foraminifers, is overlain by chert-rich nannofossil-foraminifer ooze. The pelagic environment represented by the limestone sequence quickly deepened upsection from upper-slope water depths at the base with redeposited mollusk debris and organic matter to middle bathyal water-depths at the limestone/chalk transition (Boersma, 1981).

The laminated limestone from Cores 62-465-40 to -26 contains elements of an upper Albian assemblage that has been attributed to the *Rotalipora appenninica* Zone (KS 16). Important species include *Rotalipora appenninica*, *R. gandolfii*, *Planomalina* sp. cf. *P. buxtorfi*, and *Ticinella primula* (Boersma, 1981). This sequence is separated by a hiatus from the overlying chert and chalk facies of Cores 62-465-25 to -23, which contain *Dicarinella asymetrica*, *D. concavata*, and *Globotruncanita elevata*, among others, that yields a late Santonian age (KS 24). Core 62-463-21 contains a mixture of upper Santonian and lower Campanian species. The occurrence of *Radotruncana calcarata* (KS 27) in Core 62-465-20 identifies a hiatus involving the lower to upper Campanian, whereas above this level, the sequence apparently is continuous into the upper Maastrichtian *Abathomphalus mayaroensis* Zone (KS 31).

Site 466 is located about 50 km northeast of Site 465 on southern Hess Rise. Poor recovery owing to the abundance of chert and poor preservation of the planktonic foraminifers complicates correlation of the Cretaceous sequence. Coring was terminated in chert-rich nannofossil limestone and chalk (Fig. 5). The limestone is overlain by cherty nannofossil ooze that extends to the top of the Cretaceous sequence. An abbreviated section of chert that is overlying black

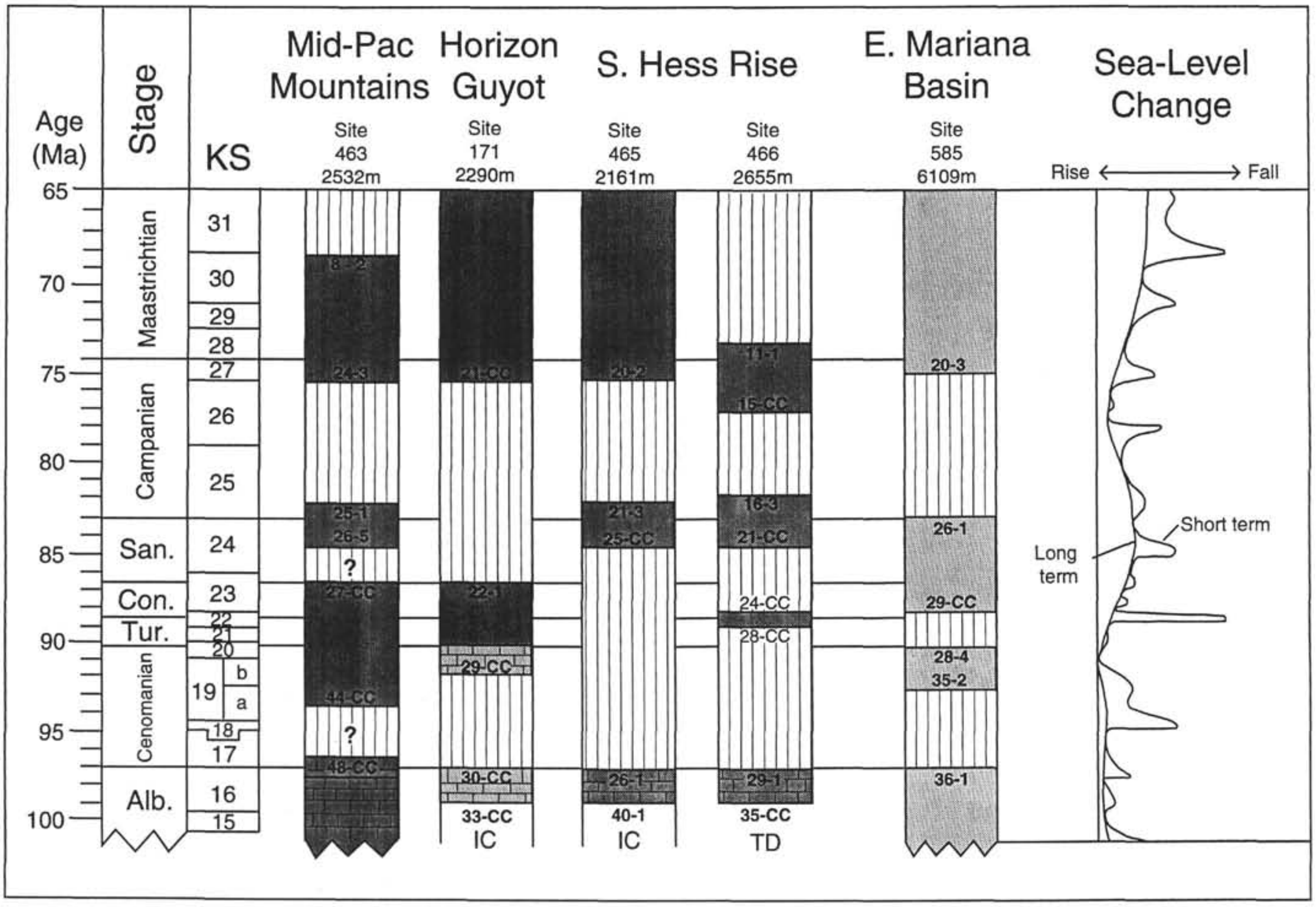

Figure 5. Age of Cretaceous sediments dated by planktonic foraminifers from selected DSDP sites in the central Pacific Ocean and East Mariana basins. Zone assignments as in Figure 2. Patterns and abbreviations as in Figure 3. Lithologic symbols shown for chalk and limestone. IC = igneous crust. Sea-level curve modified from Haq et al. (1987).

pyritic clay in Core 62-466-28 was placed at the limestone/ooze transition (Thiede, Vallier, et al., 1981).

Planktonic foraminifers, identified by Boersma (1981) from the limestone facies in Cores 62-466-35 to -29, are referred to the upper Albian *Rotalipora appenninica* Zone (KS 16). A major hiatus at the limestone/ooze transition was identified by the assemblage in Cores 62-466-28 to -24, assigned to the upper Turonian to lower Coniacian *Marginotruncana sigali* Zone (KS 22). This interval also includes a fragmented orbitolinid in Core 62-466-25 that identifies redeposition from a shallow-water source. The occurrence of *Dicarinella asymetrica*, among others, in Sample 62-466-21-CC yields a Santonian age (KS 24) that is probably equivalent to the upper Santonian assemblage at Site 465. This assemblage extends upsection to Core 62-466-16, where the presence of the nannofossil *Broinsonia parca* indicates an early Campanian age (Čepek, 1981). Cores 62-466-15 to -11 were placed in the upper Campanian *Radotruncana calcarata* Zone by Boersma (1981), based on the abundance of Campanian species, although the nominate species was not found. A more definitive age for this interval is derived from nannofossils that were assigned by Čepek (1981) to the *Tetralithus gothicus* Zone in Cores 62-466-15 to -12 and the *Tetralithus trifidus* Zone in Core 62-466-11. Accordingly, the interval ranges from the upper Campanian to the lower Maastrichtian (Fig. 5).

The Cretaceous sequence at Site 585 in the East Mariana Basin consists of volcaniclastic turbidites of Aptian to Albian age, followed by siliceous siltstone and zeolitic claystone of late Albian to Santonian age, and finally nannofossil claystone, zeolitic claystone, and chert of Campanian to Maastrichtian age. The entire sequence was deposited at abyssal water depths, well below the calcite compensation depth. Throughout the sequence, episodic pulses of calcareous shallow-water material were deposited on the basin floor and occasionally preserved. The ages of these events are based primarily on hydrodynamically sorted benthic foraminifers (Sliter, 1986), and secondarily on scattered, poorly preserved planktonic foraminifers (Premoli Silva and Sliter, 1986) and nannofossils (Bergen, 1986). Consequently, the age limits are broad and often poorly defined (Fig. 5).

Major resedimentation events occurred during the Aptian/Albian and Maastrichtian, with minor pulses during the Cenomanian/Turonian and Coniacian/Santonian (Sliter, 1986). The Aptian/Albian pulse extended into the late Albian, as evidenced by a meager assemblage of planktonic foraminifers in Core 89-585-36 that includes *Ticinella primula* and forms resembling *Biticinella breggiensis* and *Ticinella praeticinensis*, associated with ooids, orbitolinids, and echinoid fragments. This assemblage ranges in age from the middle Albian *Biticinella breggiensis* Zone to the late Albian *Rotalipora ticinensis* Zone (KS 15).

The Cenomanian/Turonian pulse of sedimentation extended from the upper Cenomanian to the lower Turonian, as determined by nannofossils (Bergen, 1986) and planktonic foraminifers (Premoli Silva and Sliter, 1986). The latter range extends from the *Rotalipora cushmani* Zone (KS 19b) to the *Whiteinella archaeocretacea* Zone (KS 20). Shallow-water material includes ooids, sponge spicules, and ostracodes. Organic carbon-rich layers found within this interval at both holes drilled at Site 585 in Cores 89-585-32 and 89-585A-8 represent the widespread Bonarelli layer, which occurs near the Cenomanian/Turonian boundary (Schlanger et al., 1987). The minor Coniacian/Santonian pulse was identified by benthic foraminifers in Core 89-585-29 and includes carbonaceous material and sponge spicules. The Maastrichtian pulse of sedimentation, with volcanogenic grains and redeposited neritic benthic foraminifers, was dated by nannofossils and rare planktonic and benthic foraminifers.

Thus, despite the effects of local volcanism during the Aptian/Albian and Maastrichtian sequences at Site 585, the pattern of sedimentation at the DSDP sites in the Central Pacific Basin is remarkably similar. Even the pace of episodic resedimentation at Site 585 showed lulls in the early Cenomanian, late Turonian, and early-to-middle Campanian that correspond to hiatuses on the topographic highs. Clearly, the pervasiveness of these widespread unconformities in both platform and basin settings requires an oceanographic or regional tectonic cause to produce interruptions in sedimentation on such a broad scale.

One measure of global environmental change is provided by variations in relative sea level. Comparing the overall trend in pelagic sedimentation with the modified sea-level curve of Haq et al. (1987) shows a correspondence between high sea level and low accumulation rates (Fig. 5). This correspondence, which has been documented in Cretaceous cratonic, continental margin, and pelagic deposits (e.g., Fischer and Arthur, 1977; Vail et al., 1977), represents a global response to changing paleoceanographic conditions.

DISCUSSION AND CONCLUSIONS

Cretaceous planktonic foraminifers from ODP Leg 143 provide a record of platform drowning and subsequent fluctuations in sea level in the western Central Pacific Ocean Basin. Comparing the pelagic record from Leg 143 with the record from DSDP sites in the Central Pacific area identifies both local and regional geologic events.

Late Albian Event

One well-documented event in the northwest Pacific is the demise of carbonate platform sedimentation near the Albian/Cenomanian boundary (review in Winterer et al., 1993). Timing of this event at Allison and Resolution guyots is poorly constrained owing to the limited recovery of the overlying Cretaceous pelagic sediments. The youngest platform age is latest Albian based on both the meager planktonic faunas (Fig. 3) and the associated, more abundant, benthic foraminifers (Arnaud-Vanneau and Sliter, this volume). The oldest surviving pelagic sediments are Turonian in age based on planktonic foraminifers from Allison Guyot. Within this time frame, the platform at Allison Guyot may have emerged, developed a karstic topography, and/or was drowned. Drowning presumably took place prior to the Turonian, but older sediments either did not accumulate or were eroded by physical or chemical processes perhaps similar to those for modern seamounts at intermediate water-depths that are swept clean by water currents (Hein et al., 1988).

Boundary relationships between Albian and younger sediments from DSDP sites in the Mid-Pacific Mountains, southern Hess Rise, and perhaps Horizon Guyot are important as they document similar lithologic and temporal changes in a coeval but pelagic environment. These records show, that even at bathyal water-depths, the abrupt transition from upper Albian pelagic limestone to younger nannofossil chalk typically is marked by a hiatus that involves much of the lower to middle Cenomanian (Fig. 5).

At Site 463 from the Mid-Pacific Mountains, however, the limestone/chalk transition takes place in the uppermost Albian, and the chalk apparently extends into the lower Cenomanian before being interrupted by a hiatus. This sequence is significant as it documents paleoceanographic events across the Albian/Cenomanian boundary at an apron site flanking Resolution Guyot. The limestone/chalk transition is attributed to a lowered rate of sedimentation and early cementation at the sediment water interface, perhaps augmented by current erosion and winnowing. The cause of the lithologic change remains ambiguous. The lowered sedimentation rate may indicate a drop in sea level or the waning contribution of carbonate from the adjacent drowning seamount. Following this event, the pelagic chalk was deposited during a subsequent rise in sea level in the early Cenomanian. If this interpretation is correct, the possible emergence and subsequent drowning of the Central Pacific guyots took place during the latest Albian *Rotalipora appenninica* Zone (KS 16; Figs. 5 and 6). A similar conclusion was reached by Grötsch and Flügel (1992) based on the examination of dredged material from numerous guyots in the western Pacific. In this case, they identified the first post-drowning sediment as early Cenomanian in age.

Late Albian to early Cenomanian events outside the Pacific Basin indicate that the late Albian change in sea level, accompanied by the demise of Pacific platform sedimentation, is indeed widespread and not a regional event. Examples include:

1. Presence of a widespread disconformity near the Albian/Cenomanian boundary (review in Schlanger, 1986).
2. Truncated and/or drowned platform carbonate deposits of late Albian age at DSDP Sites 536 and 538 on the Campeche Escarpment and Catoche Knoll, respectively (Buffler, Schlager, et al., 1984), at ODP Site 627 off Little Bahama Bank (Watkins and Verbeek, 1988), and between Florida, north Cuba, and the Bahamas (Denny et al., 1994). The latter broad platform may have been subaerially exposed or karstified during a late Albian (?) to middle Cenomanian drowning judging from irregularities on sequence boundaries.
3. An erosional phase in the Gulf of Mexico, involving displaced shallow-water material from adjacent carbonate platforms, that began in the late Albian and reached maximum intensity in the early and middle Cenomanian (Sliter and Premoli Silva, 1984).
4. A deep-water Cenomanian to Santonian hiatus found throughout the Southern Hemisphere that typically begins in the late Albian (Sliter, 1977; Krasheninnikov and Basov, 1983). These latter studies show the upper Albian sequence on the Falkland Plateau is truncated, increases in water depth at the top, and contains a change in bivalve fragments from *Aucellina* dominated throughout the lower to middle Albian section to *Inoceramus* dominated in the upper Albian as the intensity and paleotemperature of deep-water currents were modified.

The timing and correlation of late Albian events in shallow-water carbonate sequences in the Tethys and Atlantic Ocean and deep-water transitions in the Southern Hemisphere support the interpretation of a link between changing paleoceanographic conditions in the world ocean and the demise of carbonate platform sedimentation in the western Pacific.

Younger Cretaceous Events

A second, but not so well-documented, group of events represented by hiatuses or times of reduced sedimentation are recognized in the pelagic cover of Cretaceous Pacific seamounts (Sliter, 1992; Kroenke et al., 1993; Sliter and Leckie, 1993). The meager record of Cretaceous pelagic sedimentation from Allison Guyot shows only broad hiatuses in the Cenomanian to Maastrichtian sequence (Fig. 3). A more complete record emerges when the data are combined with the recovery from the deeper-water DSDP sites in the central Pacific (Fig. 5). This combined, regional, pattern shows hiatuses in the lower-to-middle Cenomanian (as noted above), upper Turonian, lower Santonian, lower-to-middle Campanian, and upper Maastrichtian sequence.

Several of these events have been recognized elsewhere, for example, such as reduced rates of sedimentation in the Cenomanian, Santonian, Campanian, and Maastrichtian sections near Gubbio, Italy (Premoli Silva and Sliter, in press), dissolution in Cenomanian to Santonian deep-water sediments from the southern hemisphere (Sliter, 1977), and the widespread lower Campanian disconformity in the

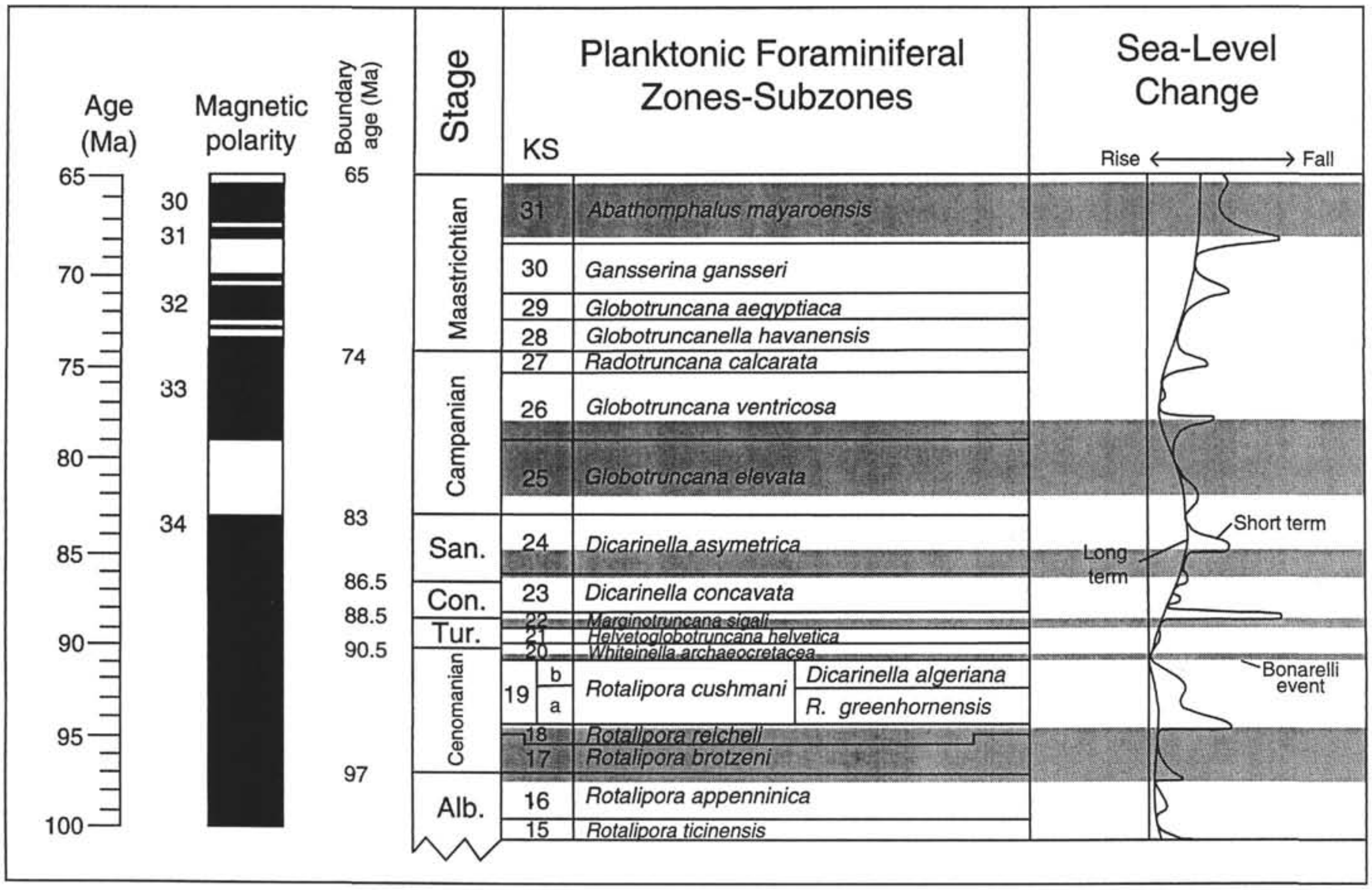

Figure 6. Stratigraphic distribution of hiatuses in Late Cretaceous pelagic cover at DSDP sites in the central Pacific Ocean Basin relative to changes in sea level. Hiatuses = shaded areas. Sea-level curve modified from Haq et al. (1987).

Southern Ocean (Huber and Watkins, 1992). The best documented of these events, is the short dissolution event at or near the Cenomanian/Turonian boundary, the Bonarelli event, that is associated with widespread deposition of organic carbon-rich sediments during a sea-level highstand (Schlanger et al., 1987).

Determining the correlation and duration of these events (represented by changes in sedimentation rate) presently is a problem owing to a combination of factors, chief of which is uncertainty in age assignment. This is especially a problem in Santonian and older Pacific Ocean sequences where chert inhibits core recovery in oceanic settings and diagenesis and lithification hamper biostratigraphic resolution. And yet, the brevity of several documented paleoceanographic events during the mid-Cretaceous (e.g., Weissert and Lini, 1991) underscores the need for temporal resolution. The Bonarelli event is a case in point as it falls within a single planktonic foraminiferal zone, the *Whiteinella archaeocretacea* Zone (KS 20), of 1-m.y. duration or less (Schlanger et al., 1987). Accordingly, although the hiatuses shown in Figures 5 and 6 are recognizable, their stratigraphic position and duration is, as yet, biostratigraphically coarse and may include one or more condensed sections.

Paleoceanographic Model

Regional hiatuses in the Cretaceous pelagic cap are ascribed to times of intensified bottom-water activity (Sliter and Brown, 1993). In the mid-Cretaceous, increased current flow was likely linked to fluctuations in sea level. Following this hypothesis, during highstands in sea level, intensified oceanic circulation at intermediate depths resulted from the increased production of warm, saline bottom water (WSBW) in flooded marginal seas (Brass et al., 1982). Whether from the horizontal spreading of the dense water at intermediate depths through dilution (e.g., Price et al., 1993) or from upwelling of less dense deeper-water, the result was increased density-driven oceanic circulation and localized upwelling, a landward shift in productive areas and carbonate deposition, a rise in the calcite compensation depth (CCD), and episodes of carbonate removal from topographic highs by dissolution and mechanical erosion.

Increased ocean current flow in the Campanian and Maastrichtian, although also associated with periods of volcanism and fluctuations in sea level, seem more closely related to long term climatic cooling and paleogeographic changes such as the opening of circum-Antarctic gateways (Huber and Watkins, 1992). These factors, together with biogeographic distribution patterns (e.g., Sliter, 1977), differentiate paleoceanographic conditions in the latest Cretaceous from those in the mid-Cretaceous. Accordingly, I here restrict my comments to the latter.

A paleoceanographic model linking sea-level highstand and the formation of WSBW with organic carbon deposition was presented by Arthur et al. (1987) and refined by Thurow et al. (1992). Support for this model and the corresponding increase in current strength at intermediate water depths comes from the recognition of condensed intervals and associated strong seismic reflectors that approximate times of organic carbon burial in the pelagic cover of Shatsky Rise. Sliter and Brown (1993) argued that the reflectors, with erosional truncation and channeling on their surface, correspond to times of enhanced upwelling and productivity during a period of increased oceanic circulation.

Changes in the diversity and morphologic development of pelagic microfossils likewise mirror the paleoceanographic events. Using planktonic foraminiferal morphotypes from the Upper Cretaceous sequence near Gubbio, Italy, Premoli Silva and Sliter (in press) interpreted fluctuations in the depth of the thermocline and stability in the upper water column. Of particular interest are times when the thermocline was shallow and the upper water column was seasonally unstable, weakly stratified, or both. As a result, oceanic conditions presumably were strongly influenced by upwelling and increased nutrient levels in near-surface waters. These times correspond to the *Rotalipora reicheli* Zone (KS 18) in the Cenomanian, the *Whiteinella archaeocretacea* Zone (KS 20) near the Cenomanian/Turonian boundary, the *Dicarinella asymetrica* Zone (KS 24) in the Santonian, and the late *Abathomphalus mayaroensis* Zone (KS 31) in the Maastrichtian. Each of these times approximates a hiatus or condensed interval in Figure 6.

Similar changes in plankton and ammonite evolution are reported in a recent study of the Cenomanian sedimentary sequence in Tunisia (Robaszynski et al., 1993) where a major faunal boundary is associated with a sea-level lowstand in the *Rotalipora reicheli* Zone (KS 18). These authors note that this lowstand appears to have the greatest amplitude of any other Cenomanian change in sea level. In addition, widespread fluctuations in the abundance of radiolarians at the Cenomanian/Turonian boundary and in the lower Campanian reported by Thurow et al. (1992) likewise are ascribed to increased productivity and variation in oceanic circulation.

Thus, the biotic changes associated with fluctuations in productivity suggest that the termination of at least some late Albian carbonate platforms may have been linked to excessive nutrient levels (Hallock and Schlager, 1986). Perhaps most vulnerable to nutrient fluctuations were oceanic platforms located in or near the paleoequatorial divergence during the late Albian and low-latitude continental margin platforms located in zones of upwelling such as in northern Venezuela (Barron, 1985). Shipboard magnetic measurements from Resolution Guyot, however, suggest paleolatitudes greater than 17°S and possibly as high as 29°S in the late Albian (Sager, Winterer, Firth, et al., 1993) and, thus, well outside the influence of the equatorial divergence.

A clue to the processes involved in the removal of carbonate during a hiatus (i.e., dissolution or mechanical erosion) is found in the record from deep-water basin Sites 869 and 585. Despite less-than-optimal age control and the local effects of volcanism (during the Cenomanian and Campanian at Site 869 and the Aptian/Albian, and perhaps Maastrichtian, at Site 585), the remaining pattern of sedimentation is remarkably similar to that from the topographic highs (Figs. 3, 5). The in-phase sedimentation events at the basin sites represents redeposition of the capping sediments preserved during the pelagic intervals. Redeposition perhaps resulted from mass flow processes during buildup of the cap and/or by periodic erosion from currents or a subsequent erosional phase.

The lulls in basin sedimentation correspond to the hiatuses on the highs. Thus, either the pelagic sediment was scoured off the highs by intensified currents and dissolved at depth or it was dissolved on top of the highs at intermediate water depths. Having noted the correlation between intensified currents and a rise in CCD, the hiatuses are ascribed primarily to variations in dissolution intensity together with increased current scouring. This implies, that during sea-level highstands, the CCD rose at least to intermediate water depths. If periods of resedimentation indeed were the sole or primary factor involved in the removal of the pelagic carbonate, one would expect the opposite effect or a lowering of the CCD to accompany the influx of carbonate.

The most direct evidence of eustatic control from the basin sites is the resedimentation event at the Cenomanian/Turonian boundary from Site 585 and, to a lesser degree, at Site 869. At Site 585, a reworked assemblage of Cenomanian and Turonian planktonic foraminifers associated with size- and shape-sorted radiolarians occur in organic carbon-rich turbidites (Premoli Silva and Sliter, 1986). This case documents mechanical erosion and redeposition associated with the Cenomanian/Turonian transgression. Perhaps not coincidently, this is one of the better dated intervals at Site 585.

If the hypothesis of increased current flow at intermediate water depths associated with the development of WSBW is accepted, several inferences follow. Sea level dynamics have long been correlated with climate change. Transgressions tend to correlate with warmer climates, regressions with cooler times. Certainly, transgressions tend to moderate climate, although the influence on heat distribution is less related to the rise and fall of sea level than to the mechanism of sea-level change as noted by Fischer and Arthur (1977). Recent studies focus on mantle outgassing of CO_2 as the major moderator of global climate (e.g., Berner et al., 1983; Arthur et al, 1985; Lasaga et al., 1985). Accordingly, increased seafloor spreading rates in the Late Cretaceous, largely a result of mid-plate volcanism in the Pacific, resulted in atmospheric warming, rise in sea level, increased precipitation rates, continental weathering, and basin-shelf fractionation of carbonate.

These relationships suggest that the variations in dissolution intensity and mechanical erosion recognized in the pelagic sequences are linked to climatic oscillations. Thus, the hiatuses in the mid-Cretaceous, correlated to highstands and the formation of WSBW, represent warmer periods, whereas carbonate accumulation correlates with cooler periods. Several intervals appear to reflect brief cooling events such as in the late Albian *Rotalipora appenninica* Zone (KS 16), as suggested by Grötsch and Flügel (1992), within or just above the middle Cenomanian *Rotalipora reicheli* Zone (KS 18), and in the late Turonian to early Coniacian *Marginotruncana sigali* Zone (KS 22). Others may reflect longer periods of perhaps less intense cooling, such as during the early Cenomanian, the early Santonian, the early Campanian, and the late Maastrichtian. These Late Cretaceous climatic oscillations resemble the Aptian fluctuations described by Weissert and Lini (1991) and, thus, apparently represent an extension of climatic cycles begun in the mid-Cretaceous "greenhouse" state.

The Cretaceous hiatuses thus appear analogous to those present in the Neogene, likewise ascribed by Barron and Keller (1982) to the intensification of bottom-water circulation. In the case of the mid-Cretaceous hiatuses, however, erosion is related to periodic warm climatic events rather than cool events as in the Neogene. The presence of hiatuses in the Cretaceous pelagic cover increases the likelihood that corresponding changes in physical properties may be associated with prominent acoustic reflectors.

These very general inferences require confirmation through future interdisciplinary study of Cretaceous sequences. Improved biostratigraphic resolution is needed as are high-resolution geophysical, geochemical and sampling surveys. Obvious targets for attack are the major seamounts and plateaus in the western Pacific with their thick pelagic cover as well as deeper-water apron and basin sites.

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[*] Abbreviations for names of organizations and publications in ODP reference lists follow the style given in *Chemical Abstracts Service Source Index* (published by American Chemical Society).

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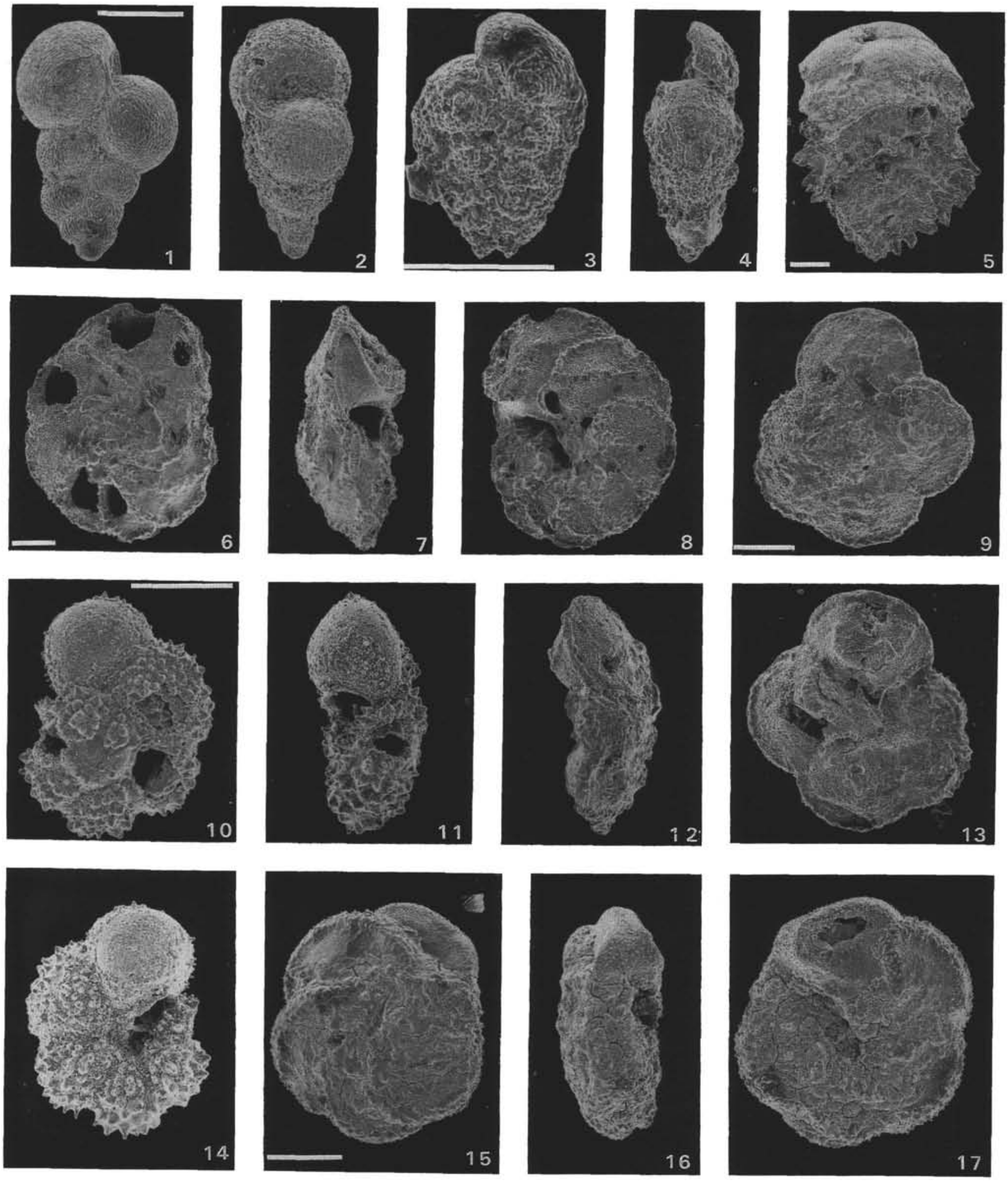

Plate 1. Scale bar = 100 µm. **1, 2.** *Heterohelix reussi* (Cushman), side and edge views, Sample 143-865B-17H-CC, 14–16 cm. **3, 4.** *Heterohelix carinata* (Cushman), side and edge views, Sample 143-869B-28R-CC. **5.** *Reussella szajnochae* (Grzybowski), side view, Sample 143-869B-15R-CC. **6–8.** *Rotalipora greenhornensis* (Morrow), spiral, side and umbilical views, Sample 143-869B-36R-13, 13–14 cm. **9, 12–13.** *Rotalipora cushmani* (Morrow), spiral, side and umbilical views, Sample 869B-31R-1, 72–74 cm. **10, 11, 14.** *Dicarinella algeriana* (Caron), spiral, side and umbilical views, Sample 143-869B-44R-1, 127–128 cm. **15–17.** *Dicarinella canaliculata* (Reuss), spiral, side and umbilical views, Sample 143-869B-31R-1, 72–74 cm.

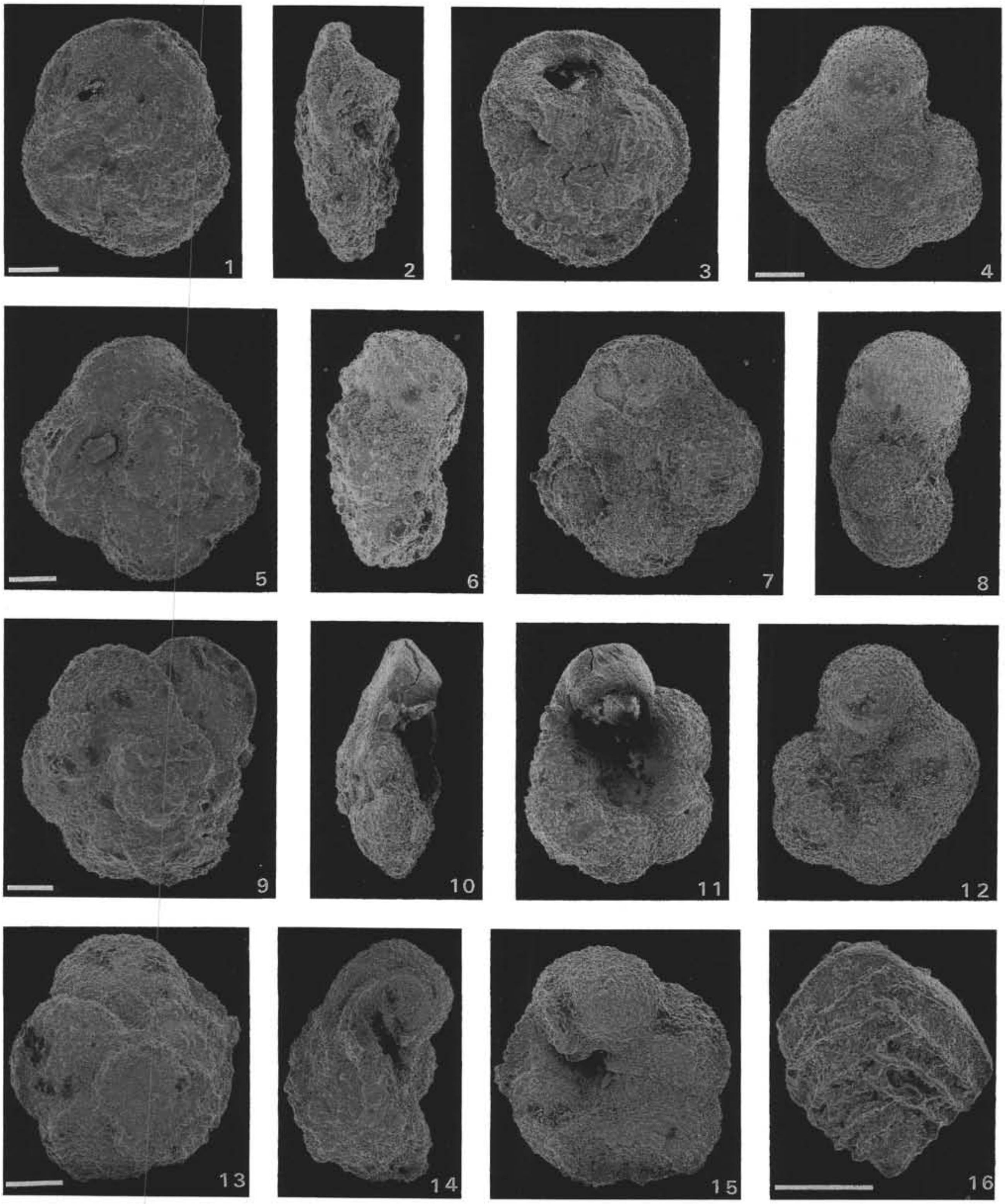

Plate 2. Scale bar = 100 µm. **1–3.** *Dicarinella primitiva* (Dalbiez), spiral side and umbilical views, Sample 143-869B-31R-1, 72–74 cm. **4, 8, 12.** *Whiteinella baltica* (Douglas and Rankin), spiral, side and umbilical views, Sample 143-869B-31R-1, 72–74 cm. **5–7.** *Helvetoglobotruncana helvetica* (Bolli), spiral, side and umbilical views, Sample 143-865B-17X-CC, 14–16 cm. **9–11.** *Marginotruncana sigali* (Reichel), spiral, side and umbilical views, Sample 143-869B-31R-1, 72–74 cm. **13–15.** *Praeglobotruncana gibba* (Klaus), spiral, side and umbilical views, Sample 143-865B-17X-CC, 14–16 cm. **16.** *Aragonia ouezzaensis* (Rey), side view, Sample 143-869B-11R-CC.

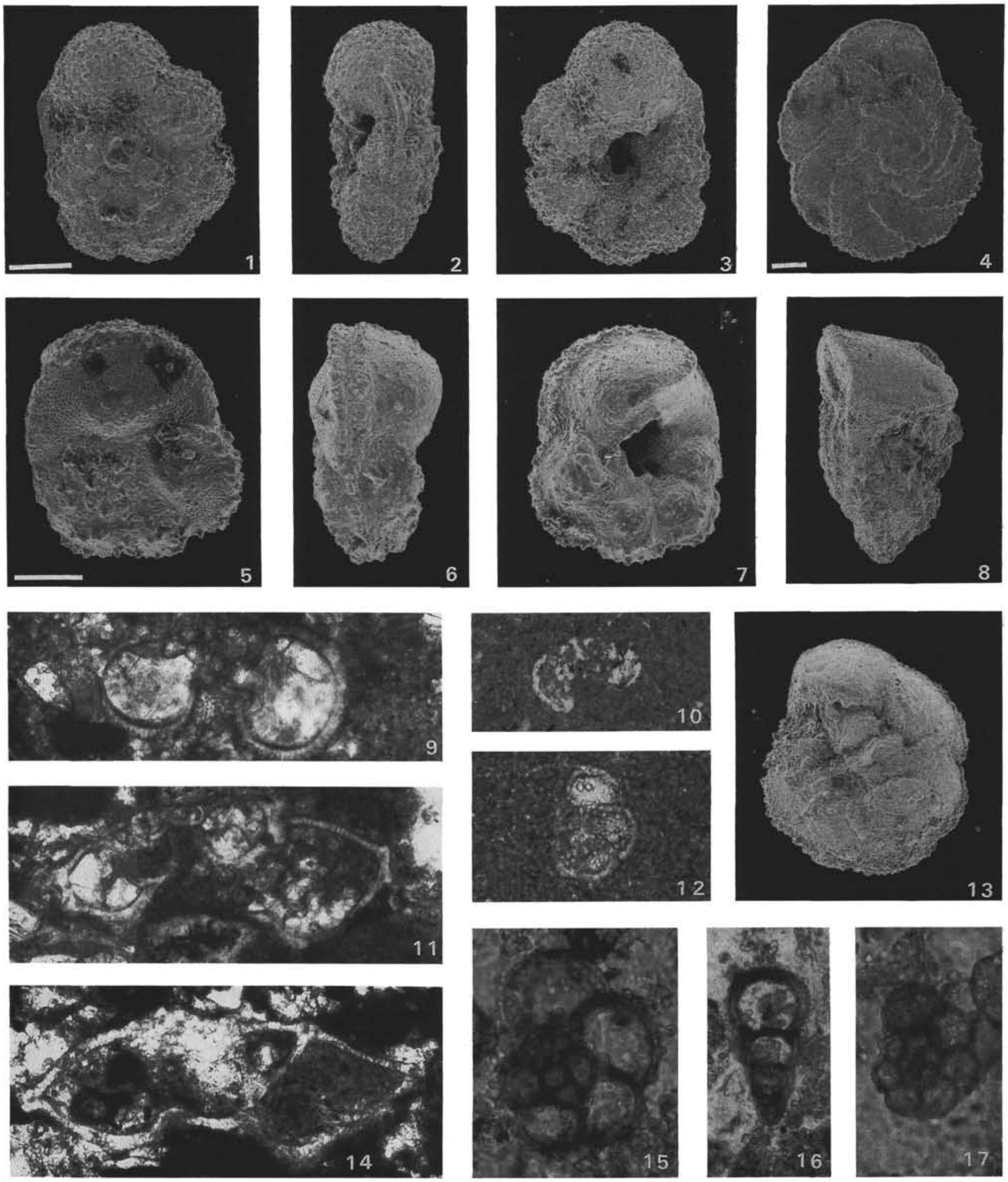

Plate 3. Scale bar =100 µm. **1–3.** *Whiteinella aprica* (Loeblich and Tappan), spiral, side and umbilical views, Sample 143-865B-17X-CC, 14–16 cm. **4, 8, 13.** *Globotruncanita elevata* (Brotzen), spiral, side and umbilical views, Sample 143-869B-12R-CC. **5–7.** *Globotruncana aegyptiaca* (Nakkady), spiral, side and umbilical views, Sample 143-869B-10R-CC. **9.** *Whiteinella baltica* (Douglas and Rankin), axial section, Sample 143-869B-34R-1, 142–144 cm, 150×. **10.** *Hedbergella* sp. cf. *H. trocoidea* (Gandolfi), transverse section, Sample 143-865A-79R-1, 32–33 cm, 300×. **11.** *Dicarinella imbricata* (Mornod), axial section, Sample 143-869B-35R-1, 44–46 cm, 150×. **12.** *Ticinella*? sp. cf. *T. primula* (Luterbacher), transverse section, Sample 143-866A-36R-1, 2–4 cm, 300×. **14.** *Rotalipora* sp. cf. *R. greenhornensis* (Morrow), oblique axial section, Sample 143-969B-52R-5, 12–15 cm, 150×. **15** *Hedbergella monmouthensis* (Olsson), transverse section, Sample 143-869B-30R-1, 24–26 cm, 300×. **16.** *Heterohelix reussi* (Cushman), edge section, Sample 143-869B-30R-1, 24–26 cm, 300×. **17.** *Hedbergella planispira* (Tappan), transverse section, Sample 143-869B-30R-1, 24–26 cm, 150×.