Late Neogene chronology: New perspectives in high-resolution stratigraphy

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ABSTRACT

We present an integrated geochronology for late Neogene time (Pliocene, Pleistocene, and Holocene Epochs) based on an analysis of data from stable isotopes, magnetostratigraphy, radiochronology, and calcareous plankton biostratigraphy. Discrepancies between recently formulated astronomical chronologies and magnetochronologies for the past 6 m.y. have been resolved on the basis of new, high-precision Ar/Ar ages in the younger part of this interval, the so-called Brunhes, Matuyama, and Gauss Epochs (= Chrons C1n–C2An; 0–3.58 Ma), and revised analysis of sea floor anomalies in the Pacific Ocean in the older part, the so-called Gilbert Epoch (= Chron C2Ar–C3r; 3.58–5.89 Ma). The magneto- and astrochronologies are now concordant back to the Chron C3r/C3An boundary at 5.89 Ma.

The Neogene (Miocene, Pliocene, Pleistocene, and Holocene) and Paleogene are treated here as period/system subdivisions of the Cenozoic Era/Erathem, replacements for the antiquated terms Tertiary and Quaternary. The boundary between the Miocene and Pliocene Series (Messinian/Zanclean Stages), whose global stratotype section and point (GSSP) is currently proposed to be in Sicily, is located within the reversed interval just below the Thvera (C3n.4n) Magnetic Polarity Subchronozone with an estimated age of 5.32 Ma. The Pliocene/Pleistocene boundary, whose GSSP is located at Vrica (Calabria, Italy), is located near the top of the Olduvai (C2n) Magnetic Polarity Subchronozone with an estimated age of 1.81 Ma. The 13 calcareous nannoplankton and 48 planktonic foraminiferal datum events for the Pliocene, and 12 calcareous nannoplankton and 10 planktonic foraminiferal datum events for the Pleistocene, are calibrated to the newly revised late Neogene astronomical/geomagnetic polarity time scale.

INTRODUCTION

The past decade has witnessed a significant improvement in the precision and accuracy of late Neogene geochronology. This has been achieved through the merging of improved radiochronologic dating techniques and astronomical tuning methods with the already familiar and classical K-Ar method and geomagnetic polarity time scale (GPTS). The application of new techniques to old problems in geochronology, however, has resulted in a paradox; namely, discrepancies between radiometrically based and astronomically based time scales for the later part of the Neogene. It is a consideration of these problems (and their resolution) that forms the focus of this paper.

This study resulted from a request in 1991 by M. B. Cita (University of Milan and chair of the International Union of Geological Sciences Subcommission on Neogene Stratigraphy) to one of us (Berggren) to chair a working group with the task of investigating and resolving the age disagreements in the then-nascent late Neogene chronologic schemes being developed by means of astronomical/climatic proxies (Hilgen, 1987; Hilgen and Langereis, 1988, 1989; Shackleton et al., 1990) and the classical radiometric age calibration of the GPTS (Mankinen and Dalrymple, 1979). Accordingly, a working group was established (refer to authorship of this paper), assignments made, and the work initiated. The extended period required to complete this work reflects the intense activity in the field of Neogene geochronology during the past 5 yr and, in particular, resolution of the apparent discrepancies between the astronomical time scale and GPTS in 1993. This resolution was achieved primarily by means of new, high-precision Ar/Ar ages on the younger part of the stratigraphic interval treated here and by means of revised interpretations of sea floor anomaly patterns in the Pacific Ocean in the older part of the Pliocene–Pleistocene record.

In this paper we review the historical framework and recent evolution of studies in magnetostratigraphy, isotopic radiochronology, and astrochronology on late Neogene geochronology. We then discuss the chronostratigraphic framework of the late Neogene and conclude with a tabulation of regional and global biostratigraphic datum events that have been calibrated to the newly proposed astro-magneto-radiochronology.

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ASTROCHRONOLOGY

When the first deep-sea sediment $\delta^{18}O$ records were presented by Emiliani (1955), he proposed that the quasi-cyclic variations in $\delta^{18}O$ were due to astronomical forcing at the 41 000 yr tilt frequency. Although this has proven to be the case for the interval prior to 1 m.y. B.P. (Pisias and Moore, 1981; Ruddiman et al., 1986), better time control (Shackleton and Opdyke, 1973; Hays et al., 1976) demonstrated that climate records since then have been dominated by a 100 000 yr cycle, although the shorter orbital cycles of tilt and precession are also preserved. The first step toward astrochronology was taken by Hays et al. (1976), who proposed that a modest adjustment to their initial time scale, well within geological constraints, would bring the peak variance of the obliquity cycles to exactly the same frequency calculated for obliquity by astrophysicists (i.e., 41 000 yr). Moreover, after making this adjustment they found improved agreement between the shorter cycles and the orbital precessional frequencies at 19 000 and 23 000 yr. Thus, the method of "tuning" time scales was established.

By the middle 1980s, the large number of oxygen isotope records that had been generated were being used to further refine the "orbital" time scale (Prell et al., 1986; Pisias et al., 1984; Imbrie et al. 1984; Martinson et al., 1987), which, for the past 300 000 yr, was proposed to be accurate to within 5000 yr (Martinson et al., 1987). The astronomically calibrated time scale was extended to ca. 800 000 yr B.P., still without generating any conflict with published geological constraints (Imbrie et al., 1984). These time scales proved extremely popular because they proposed ages for many characteristic features of the late Pleistocene $\delta^{18}O$ record that are easily recognizable in deep-sea records (e.g., isotopic stages; Fig. 1).

Initial attempts at extending the orbital chronology back to the Pliocene (Pisias and Moore, 1981; Williams et al., 1988) were hampered by the relatively poor quality of the available core material. The development of the hydraulic piston corer by the Deep Sea Drilling Project, which allowed the recovery of continuous, high-sedimentation-rate sections, led to the generation of continuous, high-resolution isotopic records back to the middle Pliocene (and by now back to the Miocene). Using a continuous and orbitally tuned $\delta^{18}O$ record from Deep Sea Drilling Project Site 607, Ruddiman et al. (1989) and Raymo et al. (1989) recalculated the ages of magnetic reversals back to the top of the Gauss Chron. Once again, the study posed no serious conflicts with prior radioisotopic estimates other than a significantly shorter Olduvai event (170 000 yr as compared with the published duration of 220 000 yr).

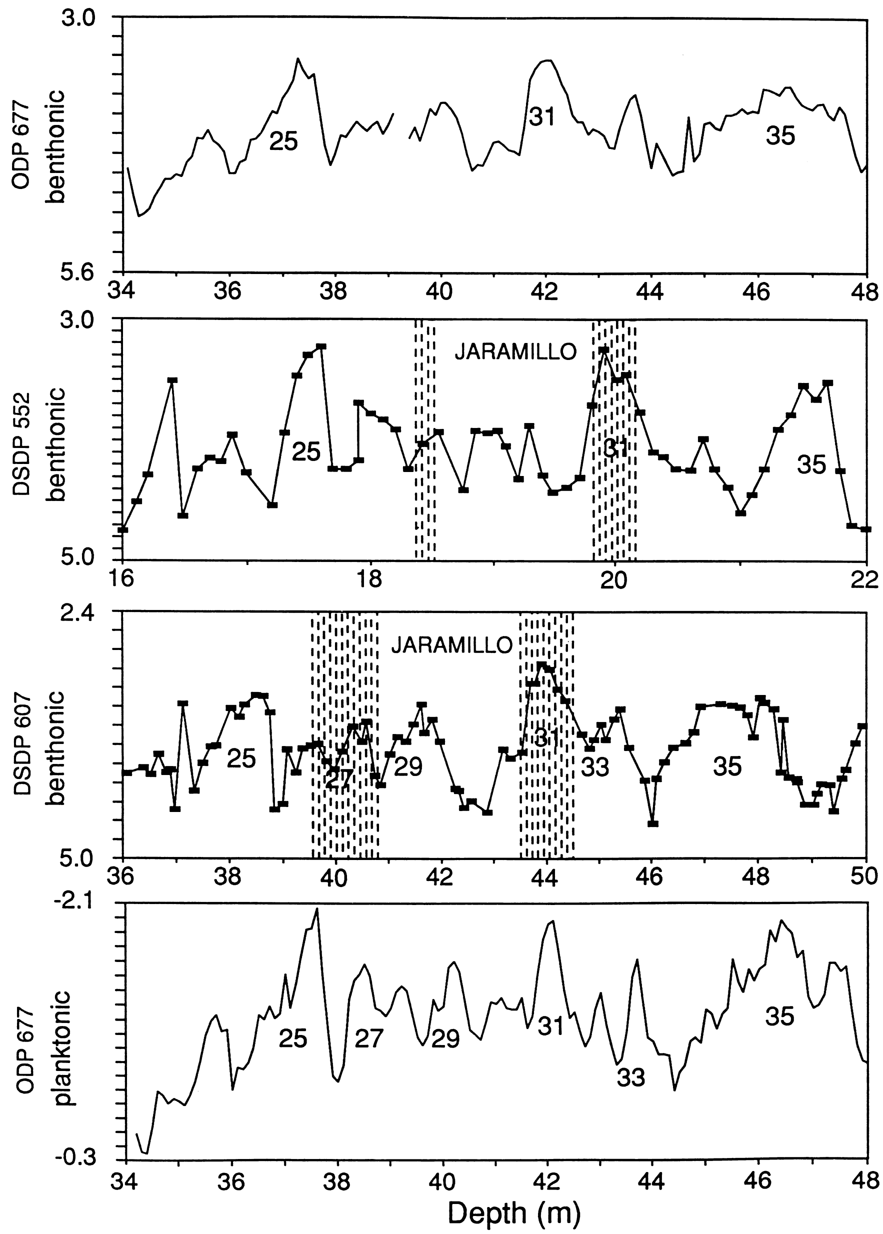

Figure 1. Oxygen isotope records from Deep Sea Drilling Project Sites 552A and 607, and Ocean Drilling Program Site 677, for the interval spanning isotope stages 24–35. Shaded intervals indicate depth range within which the Jaramillo magnetic reversals (top and bottom) fall in Sites 552 and 607 (after Shackleton et al., 1990).

Shortly after this, a "radical" orbital time scale was proposed, which required a significant (≈6%) adjustment to the published ages of the Pliocene–Pleistocene magnetic reversals (Shackleton et al., 1990; hereafter SBP90). In favor of the SBP90 time scale was the significant improvement in the coherency of the planktonic isotopic records with the highly modulated record of orbital precession. Hilgen (1991a), simultaneously examining Pliocene sedimentary sections in southern Italy, also proposed an astronomical chronology with the age of the Matuyama/Gauss boundary 6% older than the accepted value.

The orbitally tuned time scale of SBP90 was based on the premise that one can correlate specific cycles in $\delta^{18}O$ records to specific well-dated orbital variations. The greatest potential for error in such an exercise arises from the possibility that one might not recognize all the orbitally forced "cycles" in the sedimentary record, either because of natural sedimentation rate perturbations or because of a hiatus due to natural causes or the coring process. The possibility of overlooking a cycle increases in the older sections where fewer high-resolution records exist for intercomparison. By concentrating on two records which had been double-cored, SBP90 were able to avoid problems due to drilling. Detailed comparisons between the two isotope records, one from the Pacific and one from the Atlantic, suggested that neither of the cores had any obvious hiatuses above 2.6 Ma.

To correlate the isotopic climatic cycles to orbital cycles, SBP90 first established that the 0.9 m cycle and the 1.6 m cycles apparent in the Ocean Drilling Program (ODP) Site 677 isotopic record reflected variations in climate at the precessional and obliquity frequencies. The 0.9 m cycles were most obvious in the planktonic $\delta^{18}O$ record (Fig. 1D), presumably because they reflect sea-surface temperature or precipitation variations at the equator, a region where precession dominates incoming solar radiation. As the precessional signal is strongly modulated by eccentricity, SBP90 focused on extracting and correlating this component of the planktonic isotopic record to the inferred orbital forcing. The correlation was made by matching the modulation of the isotopic record at this frequency, determined by statistically filtering the data, to that observed in the precessional signal (see SBP90 for detailed discussion of the methodology). This strategy clearly was better than working with the benthic isotope records (Figs. 1A–1C), which are dominated by the poorly modulated obliquity signal. While it is fairly straightforward to identify obliquity cycles, it is difficult, without the benefit of a strongly modulated signal, to be certain of the exact match into the orbital obliquity record.

By using the coherency of the precessional signal as their guide, SBP90 were able to show that the previous orbital chronology of Ruddiman et al. (1989) missed identifying two obliquity cycles in the early Brunhes and early Pleistocene sections of the record, in part because they accepted the literature age for the Matuyama/Brunhes boundary as being essentially correct. The SBP90 time scale results in a downward (older) revision of ages (by ≈5%–7%) for the magnetic reversals, biostratigraphic datums, and oxygen isotopic stages contained within the stratigraphic record of the past 3.5 m.y. In particular, one can use oxygen isotope stratigraphy to identify the stratigraphic level of magnetic reversals in cores without magnetic data (e.g. the Jaramillo Subchron in Fig. 1; Table 1). Clearly, the assignment of accurate ages to the characteristic features of the global $\delta^{18}O$ record provides an alternative, extremely high-resolution correlation and dating tool for researchers working with Pliocene–Pleistocene marine sediments.

TABLE 1. OXYGEN ISOTOPE STAGES CORRELATIVE WITH PLIOCENE–PLEISTOCENE MAGNETIC REVERSALS

Stage*	Magnetic reversal
Brunhes/Matuyama	19
Jaramillo T	27
Jaramillo B	31
Olduvai T	63–64
Olduvai B	71–72
Gauss/Matuyama	103–104
Kaena T	G21–G22
Kaena B	K3–KM2
Mammoth T	KM6
Mammoth B	MG1
Gauss/Gilbert	MG6–Gi1
Cochiti T	Co1–Co2
Cochiti B	CN1
Nunivak T	N1–N2
Nunivak B	N7–N8
Sidufjall T	Si1–Si2
Sidufjall B	Si6
Thvera T	T1
Thvera B	TG2

*Oxygen isotope stage designations after Ruddiman et al. (1989), Raymo et al. (1989), and Shackleton et al. (1995).

CYCLOSTRATIGRAPHY OF THE MEDITERRANEAN

In the Mediterranean region similar work was being carried out by Hilgen (1991a). In that basin, sedimentary cycles occur frequently and are widespread in the marine sequences of the Neogene Period. They include a wide variety of rock types such as marl-limestone alternations, sapropels, diatomites, and even turbidites and evaporites (e.g. Ryan, 1972; Vai and Ricci Lucchi, 1977; McKenzie et al., 1979; Meulenkamp et al., 1979; Hilgen, 1987; Rio et al., 1989; Postma et al., 1993; Weltje and de Boer, 1993). These sedimentary cycles, in particular the sapropel record of the upper Pleistocene, are also linked to astronomically induced variations in climate (e.g., Ryan, 1972; Cita et al., 1978; Vergnaud-Grazzini et al., 1977; Calvert, 1983; Rossignol-Strick, 1983; Thunell et al., 1984; Rohling and Gieskes, 1989).

In southern Italy, sapropel-bearing sequences are found in the upper Pliocene–lower Pleistocene Narbone Formation (Fig. 2). As in upper Pleistocene piston cores, sapropels in the Narbone Formation are not distributed evenly in the stratigraphic record but occur in clusters on various scales: large-scale clusters commonly comprise several (two to three) small-scale clusters that in turn contain one to four individual sapropels. The Narbone Formation is underlain by the rhythmically bedded marls of the lower Pliocene Trubi Formation. The conspicuous rhythmic bedding in the Trubi consists mainly of quadripartite depositional sequences displaying a distinct gray-white-beige-white color layering, with the gray and beige beds being less indurated and $CaCO_3$-poor. In addition to these small-scale color cycles, larger-scale cycles are easily distinguished by the regular recurrence of relatively thick, indurated and $CaCO_3$-rich intervals.

Although it had been suggested that these bedding cycles were connected with the Earth's orbital cycles (Meulenkamp et al., 1979), the inferred astronomical forcing and its potential geochronological application had never been studied systematically. Two main factors contributed to this disregard: the lack of a sufficiently long and continuous rhythmically bedded sequence, and the lack of an accurate (first-order) time-stratigraphic framework. The first of these problems was solved by field-stratigraphic surveys in the Capo Rossello area and in Calabria, which identified a continuous composite sequence of the Trubi and the lowermost part of the Narbone Formation (Hilgen, 1987, 1990).

The second problem was resolved by a magnetostratigraphic study of the new Rossello composite section, which showed that the Trubi and lower part of the Narbone Formation range in age from the oldest reversed subchron of the Gilbert Chron to the oldest reversed subchron of the Matuyama Chron (Langereis and Hilgen, 1991). The upper part of the Narbone Formation as exposed in the Singa and Vrica sections (Calabria) ranges in age from the youngest normal subchron of the Gauss Chron to younger than the Olduvai Subchron (Tauxe et al., 1983; Zijderveld et al., 1991).

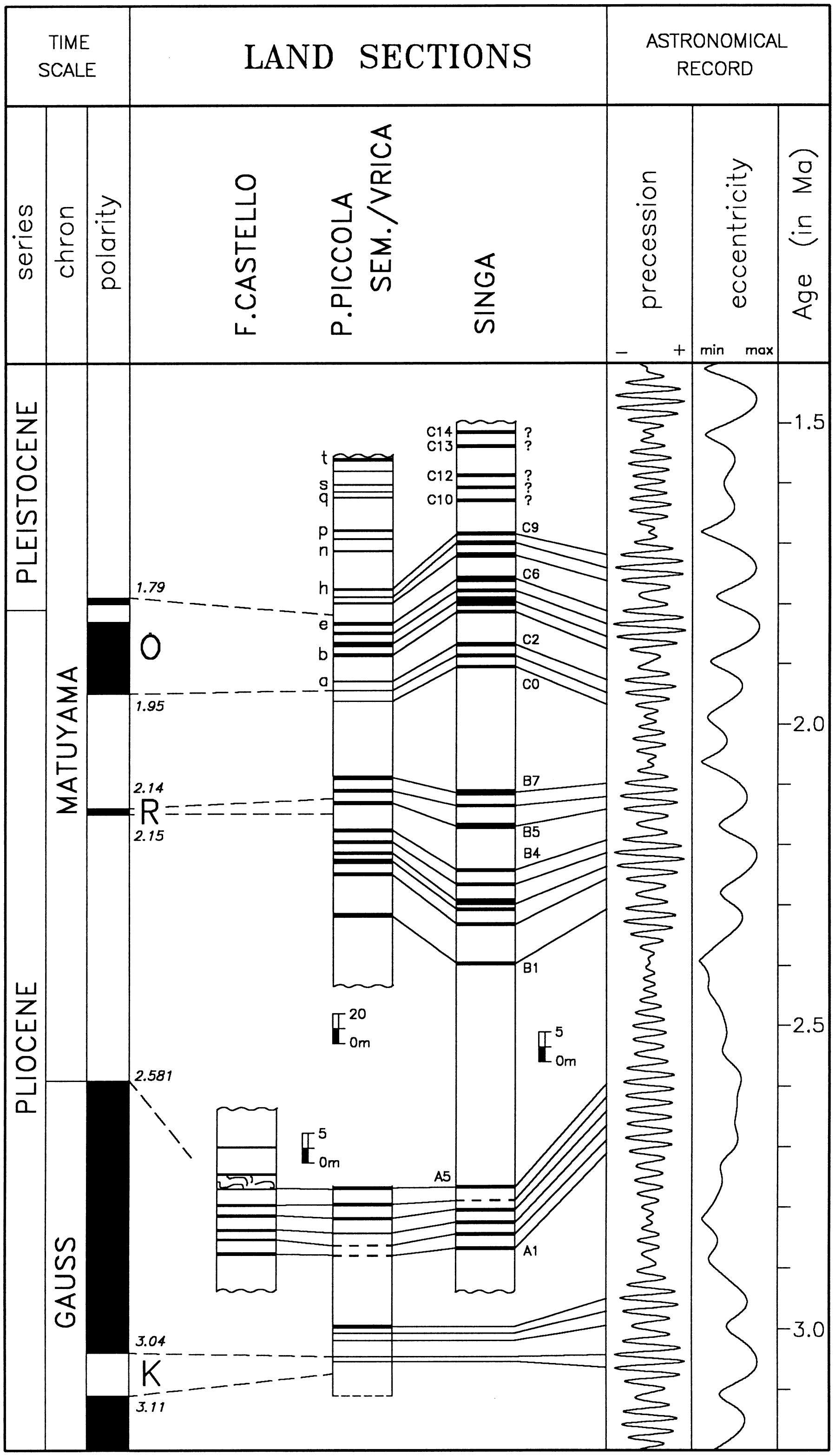

Figure 2. Astronomical calibration of sapropels in the Narbone Formation (modified after Hilgen, 1991a). Sapropel codifications after Selli et al. (1977) and Verhallen (1987) for the Vrica and Singa sections, respectively. Astronomical time series after solution BER90 (Berger and Loutre, 1991). Note that a detailed study of the Gauss/Matuyama boundary has led to a slight revision of the initial astronomical age for this reversal (Langereis et al., 1994), which is adopted here.

The continuous and magnetostratigraphically controlled composite sequence of cyclically bedded marine sediments of the Trubi and Narbone Formations served as a starting point for constructing an astronomically tuned time scale. The observation that periodicities of $CaCO_3$ cycles in the Trubi differ consistently from periodicities of the astronomical cycles (Berger and Loutre, 1991) if the geomagnetic polarity time scale of Berggren et al. (1985a, 1985b) was used (Hilgen and Langereis, 1989) gave further impetus to this attempt.

The astronomically calibrated time scale was constructed in two steps. First, the upper Pliocene–lower Pleistocene sapropel sequences were correlated to the orbital time series. Phase relations between sapropel and astronomical cycles were established by tying the youngest sapropels of late Pleistocene age to the astronomical record. This procedure revealed that individual sapropels correlate with minima of the precession index and that small-scale and large-scale sapropel clusters correspond to eccentricity maxima at 100 and 400 ka, respectively. These phase relations were used to calibrate the older sapropels in our land sections to the astronomical record (Hilgen, 1991a; Hilgen et al., 1993; see also Fig. 2). Because these sapropels remain restricted to the upper Pliocene–lower Pleistocene, the $CaCO_3$ cycles of the Trubi (lower Pliocene) were employed to extend the time scale back to the Miocene/Pliocene boundary (see Hilgen, 1991b, for discussion of methodology). Accepting the accuracy of the astronomical solution of Berger and Loutre (1991), this calibration of sedimentary cycles dates all sapropels and $CaCO_3$ cycles with an accuracy of ≈1–2 k.y., assuming a time lag of 3 k.y. between orbital forcing and maximum climate response (e.g., sapropel formation).

The presence of high-resolution magnetostratigraphic and biostratigraphic records in all sections meant that all bioevents and polarity reversals, from the top of the Olduvai Submagnetozone down to the bottom of the Thvera Submagnetozone, could be dated with a similar accuracy (of 5–10 k.y.; see Table 2). Astronomical ages were further obtained for the Pliocene/Pleistocene and Miocene/Pliocene boundaries. All ages were significantly older than ages given by earlier time scales based on radiometric dating; however, they were in essential agreement with the astronomically derived ages of SBP90. (Note that high-resolution studies of the polarity transitions in these sediments revealed a complication in establishing the

TABLE 2. AGE ESTIMATES MADE SINCE 1979 FOR CHRON AND SUBCHRON BOUNDARIES BASED ON VARIOUS METHODOLOGIES

Polarity chron and subchron*			Isotopic		Geomagnetic BKV (3)	Astronomical		Isotopic				Spline fit C&K (10)	Isotopic O&I (11)	Astroisomagnetic SHCK (12)
			M&D (1)†	McD (2)		S (4)	H (5)	B (6)	W (7)	McD (8)	T (9)			
Brunhes														
			0.73	0.72	0.73	0.78		0.783	>0.75	0.780	0.75	0.780	0.78	**0.78**
Matuyama	Jaramillo (N)	T	0.90	0.89	0.91	0.99			>0.99	0.915	0.992	0.984	>0.955	**0.99**
		B	0.97	0.94	0.98	1.07				1.010		1.049	>1.00	**1.07**
	Cobb Mtn (N)	T	1.10 (mid)									1.201	1.21	**1.21**
		B				1.19						1.212	1.24	**1.24**
	Olduvai (N)	T	1.67	1.76	1.66	1.77	1.79			1.78		1.757		**1.77**
		B	1.87	1.91	1.88	1.95	1.95		1.98–2.01	1.96		1.983	>1.96	**1.95**
	Reunion II (N)	T	2.01	2.07			2.14	2.11–2.13		2.11			2.11	**2.14**
		B	2.04				2.15	2.11–2.15		2.15				**2.15**
	Reunion I (N)	T	2.12							2.19		2.197		
		B	2.14							2.27		2.229	>2.18	
			2.48	2.47	2.47	2.60	2.60			2.60		2.600		**2.58**
Gauss	Kaena (R)	T	2.92	2.91	2.92		3.04		3.04	3.02		3.054		**3.04**
		B	3.01	3.00	2.99		3.11		3.10	3.09		3.127		**3.11**
	Mammoth (R)	T	3.05	3.07	3.08		3.22		3.21	3.21		3.221		**3.22**
		B	3.15	3.17	3.18		3.33		3.30	3.29		3.325		**3.33**
			3.40	3.41	3.40		3.58			3.57		3.553		**3.58**
Gilbert	Cochiti (N)	T	3.80	3.82	3.88		4.18					4.033		**4.18**
		B	3.90	3.92	3.97		4.29					4.124		**4.29**
	Nunivak (N)	T	4.05	4.07	4.10		4.48					4.265		**4.48**
		B	4.20	4.25	4.24		4.62					4.432		**4.62**
	Sidufjall (N)	T	4.32	4.44	4.40		4.80					4.611		**4.80**
		B	4.47	4.57	4.47		4.89					4.694		**4.89**
	Thvera (N)	T	4.85	4.72	4.57		4.98					4.812		**4.98**
		B	5.00	4.94	4.77		5.23					5.046		**5.23**
				5.44	5.35									**5.89**
C3An														

Note: Comparison of age estimates for chron and subchron boundaries based on isotopic data, astronomical calibration, and cubic spline fit to sea-floor spreading magnetic anomalies. The specific (post-1979) studies cited are listed. Note that the number of Reunion events remains uncertain. The younger event, termed Reunion II, is now well dated and, in deep-sea records, is consistently found in oxygen isotope stage 81. It probably corresponds to the W-event in seafloor anomaly profiles. Cande and Kent (1992) obtained an age of 2.197–2.229 Ma for W, labeled Reunion I in the present table because of its discrepant age as compared with Reunion II; however, recalculation using linear interpolation and the astronomical ages for base Olduvai and G/M boundary yields an age of 2.169–2.202 Ma, making it virtually indistinguishable from the Reunion II. At present no consistent data set exists for an older Reunion event.

*N = normal, R = reversed, T = top, B = bottom.

†References: M&D (1), Mankinen and Dalrymple (1979) and Mankinen and Gromme (1982); McD (2), McDougall (1979); BKV (3), Berggren et al. (1985a); S (4), Shackleton et al. (1990); H (5), Hilgen (1991a, 1991b); B (6), Baksi et al. (1992, 1993); W (7), Walter et al. (1991, 1992) and Renne et al. (1992); McD (8), Spell and McDougall (1992) and McDougall et al. (1992); T (9), Tauxe et al. (1992); C&K (10), Cande and Kent (1992), O&I (11), Izett and Obradovich (1991, 1992), Obradovich and Izett (1992), Obradovich et al. (in press); SHCK (12), this work.

exact position, and thus age, of reversal boundaries [Linssen, 1988; Van Hoof and Langereis, 1991, 1992; Van Hoof et al., 1993; Langereis et al., 1994]. It appears that the polarity of the Earth's magnetic field [e.g., following a reversal] still can be imprinted at a considerable depth, 1–1.5 m below the sediment-water interface. This artifact, presumably related to recurrent and cyclic redox conditions [early diagenesis], implies that astronomical ages for polarity reversals in the Trubi may be too old by 5–10 k.y. This would slightly reduce but still not explain the age discrepancies between astronomically calibrated and conventional [K/Ar] datings of the polarity reversals.)

Isotopic Radiochronology

The GPTS of Mankinen and Dalrymple (1979) represents the latest major synthesis of isotopic data related to rocks having specific polarity signatures. Subsequent geomagnetic polarity time scales (Berggren et al., 1985a, 1985b; Harland et al., 1982, 1990) have used the Mankinen and Dalrymple GPTS as an integral part of their late Cenozoic studies. Johnson (1982) was the first to suggest that the radiometrically determined age of the Brunhes-Matuyama boundary was incorrect. By correlating major glacial extremes ($\delta^{18}O$ minima) to major insolation minima, he arrived at an estimated age of 0.79 Ma, in contrast to the age of 0.73 Ma reported for this boundary (Mankinen and Dalrymple, 1979). This article was largely ignored by the geochronological community; however, Shackleton et al. (1990) again raised the issue when they proposed a new astronomical calibration of the early Pleistocene time scale. Not only did this time scale challenge the age of the Brunhes/Matuyama boundary (0.78 versus 0.73 Ma) but, in combination with the study of Hilgen and Langeris (1989), questioned the correctness of all ages for chron and subchron boundaries during the past 5.0 m.y.

At the time, it was difficult for the geochronological community to perceive of any mechanism that could consistently bias the isotopically derived ages by 5%–7% over a period of >5 m.y. (Shackleton et al., 1990; Hilgen, 1991a and 1991b). Explanations for the consistent age discrepancies ranged from imprecision of individual age determinations, to diffusive loss of radiogenic argon, to inadequately known decay constants. The last was readily dismissed, because the current estimate of the error for the decay constant of ^{40}K is <0.5% (Gale, 1982). As >90% of the data utilized by Mankinen and Dalrymple are from whole-rock basalts and andesites, a more likely possibility was systematic loss of argon from the glassy matrix (the site of the bulk of the potassium) or a loss due to unrecognizable alteration. In this case, all of the data would have to be questioned despite great efforts made to provide criteria by which suitable whole-rock samples could be selected (Dalrymple and Lanphere, 1969; McDougall, 1971). Likewise, diffusive loss of argon from sanidine as well as biotite and plagioclase minerals also seemed an untenable explanation as diffusion studies on sanidine yielded high activation energies (Fetchig and Kalbitzer, 1966; Musset, 1969). It was McDowell (1983) who first suggested, but did not prove conclusively, that it was difficult to totally degas sanidine, leading to an underestimated age. Deino et al. (1989) later demonstrated that this resulted in young ages for nine sanidine samples employed by Mankinen and Dalrymple (1979). When ^{40}Ar-^{39}Ar ages were compared to conventional K-Ar ages on sanidine, the ^{40}Ar-^{39}Ar ages were invariably older, some times by as much as 20%, although 5%–10% would be more in keeping with the experience of one of us (Obradovich). Further, as the constraining ages on the Brunhes/Matuyama boundary and the Jaramillo Subchron were based on sanidine, this implied that the time scale of Mankinen and Dalrymple was indeed biased toward young ages.

The first ^{40}Ar-^{39}Ar studies (Izett and Obradovich, 1991) on sanidine from the normally magnetized Bishop Tuff (middle Pleistocene) increased the age determination from 0.74 Ma (Izett et al., 1988) to 0.77–0.78 Ma (the age determination depends on the choice of age for the irradiation monitor). At the same time, Baksi et al. (1992) demonstrated an ^{40}Ar-^{39}Ar plateau age of 0.78 Ma in whole-rock basalt samples with transitional polarity from Maui, Hawaii. These ^{40}Ar-^{39}Ar age estimates for the Brunhes/Matuyama boundary were in much closer agreement with those proposed by Johnson (1982) and Shackleton et al. (1990).

These initial results spurred a flurry of activity in which the Jaramillo, Cobb Mountain, Olduvai, Reunion, Kaena, and Mammoth Subchrons and the Matuyama-Gauss and Gauss-Gilbert Chron boundaries were investigated. For sake of brevity these recent studies are summarized in Table 2 along with the pertinent citations. Also included for comparison are the GPTS of Berggren et al. (1985a, 1985b), the relevant portion of the magnetic anomaly time scale of Cande and Kent (1992), and the astronomically based chronology adopted in this paper (column 12 in Table 2). What is evident from this compilation is that all reversal boundaries for the past 5 m.y. given in Mankinen and Dalrymple (1979) and McDougall (1979) are too young and that the new ^{40}Ar-^{39}Ar ages are in general agreement with the astronomically calibrated scale.

This brings into question the overall accuracy of conventional K-Ar ages on whole-rock basalt and andesite samples. Despite earlier statements by one of us (Obradovich), loss of radiogenic ^{40}Ar has taken place in these whole-rock samples either as a result of diffusion or alteration. Until ^{40}Ar-^{39}Ar incremental heating studies are carried out on most of the whole-rock samples making up the database of Mankinen and Dalrymple, however, it will be difficult to decide the intrinsic value of any one data point, particularly in those intervals having a large number of reversal events. Finally, some differences between the new isotopically and astronomically calibrated time scales cannot be explained by a bias resulting from calibration errors in the isotopic data. For instance, the ages for the Cobb Mountain Subchron and the base of the Olduvai deviate far more than would be expected given the resolution attributed to the astronomical approach ($\approx$5 k.y.).

A New Geomagnetic Polarity Time Scale

Age estimates of geomagnetic polarity reversals during Pliocene and Pleistocene time have been based traditionally on paired K-Ar age and magnetization polarity determinations on lavas distributed more or less at random in time and space. The initial impetus for this approach was to test the hypothesis of self-reversal as an explanation for reversed magnetizations in rocks, a localized phenomenon that would not be expected to produce a globally consistent temporal distribution of magnetic polarities (e.g., Cox et al., 1963, 1964). This methodology for the construction of a GPTS is limited by the nonuniform age distribution of the K-Ar and polarity data as well as by the accuracy and precision of the K-Ar radiometric ages themselves. To minimize these problems, Cox and Dalrymple (1967) introduced the chronogram method in which the most likely age of a reversal is that which minimizes the apparent inconsistency in a population of dated lavas of known polarity.

In the most recent comprehensive analysis of the K-Ar/polarity data set for the past 5 m.y., Mankinen and Dalrymple (1979) used the chronogram method to provide age estimates for the reversal boundaries between the Brunhes/Matuyama (0.73 Ma), Matuyama/Gauss (2.47 Ma), and Gauss/Gilbert (3.40 Ma) Chrons. These estimates have been used widely for Pliocene–Pleistocene chronostratigraphy, and the age of the Gauss/Gilbert boundary has been, until very recently, a critical calibration point for virtually all geomagnetic polarity time scales based on marine magnetic anomalies since the classic study of Heirtzler et al. (1968).

In contrast to lavas on land, continuous high-resolution records of geomagnetic polarity can be obtained from the magnetostratigraphy of deep-sea sediments and from sea-floor spreading magnetic anomalies. Magnetostratigraphy allows direct correlation of biostratigraphy with the reversal sequence necessary for the construction of integrated geologic time scales. The validity of the marine magnetic anomaly record is established by the high redundancy of data available from the various ridge systems in the global ocean. Indeed, the geomagnetic polarity sequence inferred from sea-floor spreading magnetic anomalies has effectively become the standard reference for geomagnetic reversals extending back to the Jurassic Period. However, the high-resolution geomagnetic reversal sequence obtained by these methods must be calibrated in time using some independent source of age information.

In comparison with the frequency of geomagnetic reversals, there have been relatively few reliable, analytically precise, and stratigraphically well-controlled radiometric dates available to establish a chronology. Oceanic basalts, the source of the magnetic anomalies, are notoriously difficult to date radiometrically, and datable horizons in magnetostratigraphic sections have been reported infrequently—hence the need for interpolation between selected calibration datums that have been correlated to the characteristic reversal pattern. For example, Cande and Kent (1992) used nine calibration tie points plus the origin for fitting a cubic spline to a revised magnetic anomaly sequence in the South Atlantic, to derive a GPTS (CK92) consisting of 181 reversals during the past 83 m.y.. The calibration points that served to constrain the Pliocene–Pleistocene portion of CK92 are the zero-age ridge axis, 0.78 Ma for the older end of the central anomaly (Chron C1n; Brunhes/Matuyama boundary), 2.60 Ma for the younger end of anomaly 2A (Chron C2An; Matuyama/Gauss boundary), and 14.8 Ma

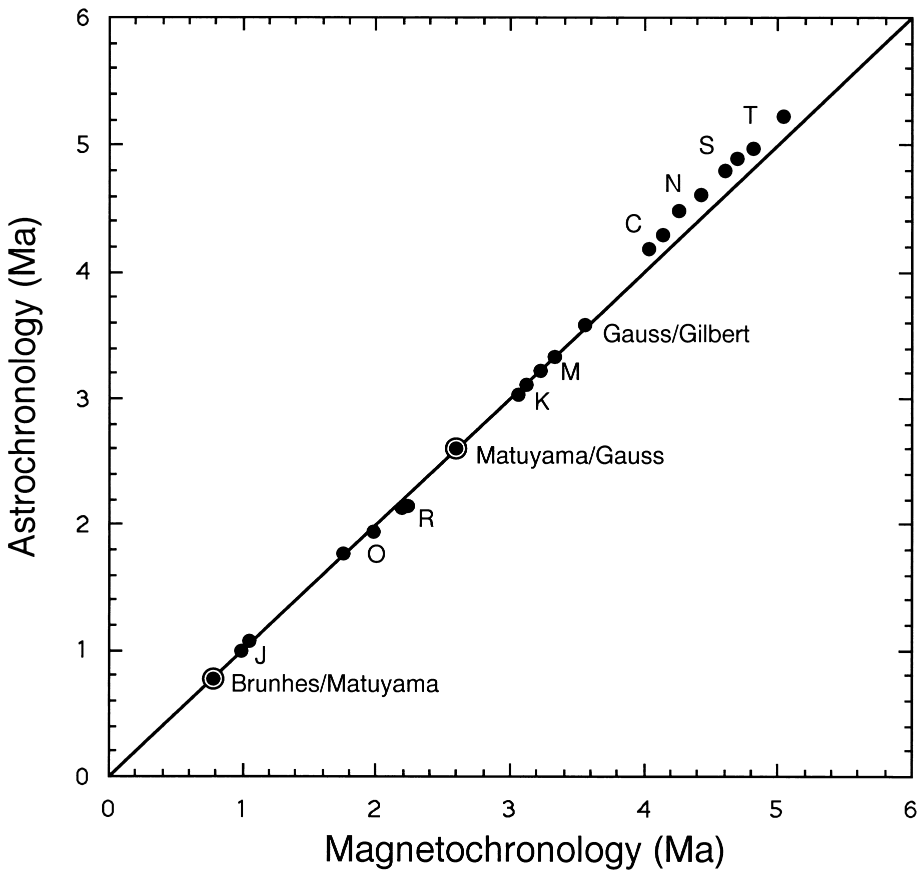

Figure 3. Ages of geomagnetic reversals determined from astrochronology (Shackleton et al., 1990; Hilgen, 1991b) compared to magnetochronology (Cande and Kent, 1992). Doubled circles are calibration points from astrochronology that were used in the magnetochronology of CK92. Chron boundaries indicated; subchron boundaries are J, Jaramillo; O, Olduvai; R, Reunion; M, Mammoth; K, Kaena; C, Cochiti; N, Nunivak; S, Sidufjall; and T, Thvera.

for the younger end of anomaly 5B (Chron C5Bn).

As described earlier, a new approach to geochronology has been to date high-resolution climate records by assuming that their characteristic variability is related to well-known variations in Earth's orbital parameters. Unlike discrete radiometric dates, which require interpolation using sedimentation or spreading rates, the astrochronologic method can effectively provide a continuum of independent age control; the precision is effectively a function of the shortest Milankovitch orbital variation ($\approx$20 k.y. precession cycle), and the accuracy depends on how reliably one can calculate orbital parameters back in time. As described above, Shackleton et al. (1990) developed an astrochronologic time scale to the level of the Gauss/Matuyama boundary. Hilgen (1991b) extended the astrochronology through the Thvera (C3n.4n) Subchron in the early part of the Gilbert Epoch (Gilbert Chron herein). Over the interval of overlap, there is excellent agreement in the ages derived from these independent sets of data, ages which have recently been confirmed by high-precision ^{40}Ar-^{39}Ar dating (e.g., Baksi et al., 1992; Tauxe et al., 1992; Walter et al., 1991; Renne et al., 1993); however, ^{40}Ar-^{39}Ar dates generally provide younger age estimates than conventional K-Ar dates for the earlier Cenozoic (e.g., Berggren et al., 1992; Cande and Kent, 1992), opposite of the trend in the Pliocene and Pleistocene.

A comparison of the astrochronology of Shackleton et al. (1990) and Hilgen (1991b) with the GPTS of Cande and Kent (1992) shows good agreement back to the base of the Gauss (Fig. 3). This is, of course, not unexpected because the calibration points at the Brunhes/Matuyama (0.78 Ma) and Matuyama/Gauss (2.60 Ma) boundaries in CK92 were based on these astrochronologic estimates. A significant difference, however, occurs in the Gilbert Chron, where astrochronology suggests ages for the subdivisions of Chron C3n (Cochiti, Nunivak, Sidufjall, and Thvera Subchrons) that are systematically 150–180 k.y. older than the ages inferred from cubic spline interpolation in CK92 (Fig. 3). The problem evidently lies in the interval between the base of the Gauss (Chron C2An), for which there is good agreement (within 30 k.y.) between the chronologies, and the top of the Cochiti (C3n.1n) Subchron, where the discrepancy suddenly becomes 147 k.y. It is now apparent that the magnetic anomaly sequence for this interval used by Cande and Kent (1992) is the likely source of the problem. This was demonstrated by Wilson (1993), who showed that the astrochronology gives a more consistent spreading history when applied to a set of revised spacings of anomalies on the Cocos-Nazca and other Pacific spreading ridges.

The success of the continuous chronology provided by the astronomical time scale means that it is not necessary to interpolate between a few discrete dated levels to construct a geomagnetic polarity time scale for this interval. Instead, the ages of reversals in the interval where astrochronology has been developed and confirmed simply become equivalent to the astrochronological values, thereby avoiding the promulgation of separate time scales. Astrochronological estimates for geomagnetic reversals back to the base of the Thvera Subchron are presented in Table 2 (Shackleton et al.; 1990 Hilgen, 1991b). Older than the Thvera Subchron, interpolation using marine magnetic anomalies is still necessary (Cande and Kent, 1995). This procedure gives a predicted age of 5.894 Ma for the base of the Gilbert (Chron C3r/C3An boundary). This interpolated age compares favorably with a tentative astronomical estimate of 5.882 Ma for this level based on results from ODP Leg 138 Site 852 (Shackleton et al., in press).

A NEW GLOBAL HIGH-RESOLUTION BIOSTRATIGRAPHY

In this section we delineate the main biostratigraphic datum events of the late Neogene succession in a series of tables. All datum events have been recalibrated to the new GPTS of Cande and Kent (1995), which is identical to the astronomical time scale of Shackleton et al. (1990) and Hilgen (1991a, 1991b). The ages of these datums differ slightly, but significantly in some instances, from the chronology in Berggren et al. (1985a and 1985b).

In Tables 3 and 4 age calibrations of Pliocene and Pleistocene calcareous nannoplankton global datum events are reported with their position versus the GPTS and stable isotope stratigraphy (references from Thierstein et al., 1977; Backman and Shackleton, 1983; Haq and Takayama, 1984; Berggren et al., 1985a; Backman and Pestiaux, 1987; Rio et al., 1990; Raffi et al., in press). These events are evolutionary appearances (first appearance datums, or FADs) and extinctions (last appearance datums, or LADs), and events caused by environmental changes such as fluctuations in presence and abundance of taxa (increase, acme, dominance, absence) and migrations (first occurrences, or FOs, and last occurrences, or LOs) in different biogeographic provinces. These events are reliable (reproducible in different sequences and among different authors), distinct, and easily recognized (appearance datums of taxa with unambiguous taxonomic definition). Bioevents excluded from this list are those with regional value (recorded and calibrated in restricted areas), events calibrated with no spatial accuracy (i.e., in a single locality), or events with uncertain biochronologic evaluations. For some of the taxa listed in Tables 3 and 4 a nonisochronous appearance datum (FAD or LAD) has been proved. Examples are: (1) the *Ceratolithus rugosus* FAD, which in oceanic areas occurs within the Thvera Subchron (Haq and Takayama, 1984), while in the Mediterranean is placed at the top of the Nunivak Subchron (Rio et al., 1990); (2) the *Helicosphaera sellii* LAD, diachronous with latitude in open-ocean sediments (Backman and Shackleton, 1983); (3) reentry of medium-size *Gepyrocapsa* spp., time transgressive between the low-latitude sections and the middle- to high-latitude sections (Raffi et al., 1993); and (4) the increase in abundance of *E. huxleyi* (reversal of dominance between *G. carribeanica* and *E. huxleyi*), time transgressive between tropical subtropical waters and transitional waters (Thierstein et al., 1977).

The numerous nannofossil datums result from rapid evolutionary turnover during the climatically unstable Pliocene and Pleistocene and provide a detailed biostratigraphic framework for this time interval. This set of calibrated bioevents, transformed to biochronologic datums, provides a high degree of stratigraphic resolution in the marine

TABLE 3. BIOCHRONOLOGY OF PLIOCENE CALCAREOUS NANNOFOSSILS

Datum event	Magnetic polarity	Age (Ma)*
Triquetrorhabdulus rugosus LAD[†]	In the uppermost part of the lowest reversed interval of Gilbert	5.34 (a, b)
Ceratolithus acutus FAD[§]	In the uppermost part of the lowest reversed interval of Gilbert	5.34 (a, b)
Ceratolithus rugosus FAD	Within Thvera (oceanic areas) Near top of Nunivak (Mediterranean)	5.0–5.23 (c) 4.5 (d)
Amaurolithus primus LAD	Near top of Sidufjall (oceanic areas) Within Nunivak (Mediterranean)	4.8 (a) 4.55 (d)
Discoaster asymmetricus FCO[#]	Near top of Cochiti (oceanic areas) In upper Gilbert (Mediterranean)	4.2 (d) 4.15 (c)
Reticulofenestra pseudoumbilicus LAD	In the uppermost part of the upper reversed interval of Gilbert	3.75 (e) 3.65–3.7** (d)
Sphenolithus spp. LAD	Near base of Gauss	3.6 (f) 3.55–3.6** (d)
Discoaster tamalis LAD	Near top of Gauss	2.78 (g) 2.73** (d)
Discoaster surculus LAD	Close to the Matuyama/Gauss boundary	2.55–2.59 (g) 2.55** (d)
Discoaster pentaradiatus LAD	Close to the Matuyama/Gauss boundary	2.46–2.56 (g) 2.55** (d)
Discoaster triradiatus	Onset acme in the early Matuyama	2.18 (g) 2.25** (d)
Discoaster brouweri LAD plus *Discoaster triradiatus* LAD	In the lowermost Olduvai (oceanic areas and Mediterranean)	1.95 (d, g)

Note: The position of the bioevents versus the magnetostratigraphy is indicated. Magnetobiochronology calibrated to the time scale shown in Table 2, column 12 (this work).

*References: a, Berggren et al. (1985a); b, Raffi and Flores (in press); c, Haq and Takayama (1984); d, Rio et al. (1990); e, Mazzei et al. (1979); f, Backman and Shackleton (1983); g, Backman and Pestiaux (1987).

[†]LAD = last appearance datum.

[§]FAD = first appearance datum.

[#]FCO = first common occurrence.

**Mediterranean.

TABLE 4. BIOCHRONOLOGY OF PLEISTOCENE CALCAREOUS NANNOFOSSILS

Datum event	Isotope stratigraphy (magnetic polarity)*	Age (Ma)*
Medium *Gephyrocapsa* spp. FAD[†] (bmG event)	Transition stages 59/60 (e) (above top of Olduvai)	1.67–1.70 (a)
Calcidiscus macintyrei LAD[§]	Upper stages 55 (e)	1.59 (c)
Large *Gephyrocapsa* spp. FAD (blG event)	Upper stage 48 (e)	1.46–1.48 (a)
Helicosphaera sellii LAD	Upper stage 49 Stage 37	(equatorial zone)1.47 (c) (mid-latitudes) 1.22 (b)
Large *Gephyrocapsa* spp. LAD (tlG event) Small *Gephyrocapsa* spp. beginning dominance	Stage 37 (e)	1.22–1.24 (a, b)
Medium *Gephyrocapsa* spp. reentrance (reemG event)	Stage 29 (within Jaramillo)	1.03 (a)
Small *Gephyrocapsa* spp. end dominance (=S2 event)	Stage 25 (above Jaramillo in Atlantic and Mediterranean)	0.96 (b)
Pseudoemiliania lacunosa LAD	Stage 12	0.46 (b, d)
Emiliania huxleyi FAD	Stage 8	0.26 (b, d)
G. caribbeanica–E. huxleyi	Stage 4 (in transition waters)	0.075
Reversal of dominance (=*Emiliania huxleyi* increase)	Stages 5a–5b (in tropical-subtropical waters)	0.09 (b, d)

Note: The position of the bioevents versus the oxygen isotope stratigraphy and magnetostratigraphy is indicated. Magnetobiochronology calibrated to the time scale shown in Table 2, column 12 (this work).

*References: a, Rio et al. (in press); b, Rio et al. (1990); c, Backman and Shackleton (1983); d, Thierstein et al. (1977); e, Raffi et al. (1993).

[†]FAD = first appearance datum.

[§]LAD = last appearance datum.

record (≈0.3 m.y. in the Pliocene and 0.2 m.y. in the Pleistocene). Some of the Pliocene bioevents of Table 3 have been recognized in the Mediterranean only (Rio et al., 1990) and are included because the standard Pliocene–Pleistocene chronostratigraphic (stage) units are defined in this region. These datums, together with planktonic foraminifera datums, provide an integrated biochronologic framework allowing reevaluation of the biostratigraphy and age of stratotype sections in the Pliocene and Pleistocene Series (Driever, 1988; Rio et al., 1991, in press).

Within this biochronologic framework, calcareous nannofossils represent the basic tool for long-distance correlations. Many of the nannofossil events are reasonably isochronous within widely separated areas, including the Mediterranean. They occur in the same isotope stage, within the same short magnetozone, or have the same relative spacing with respect to magnetozones. In particular, all but one of the Pleistocene nannofossil events appear isochronous between the Mediterranean and low- and middle-latitude oceanic areas (assuming 50 k.y. as a limit of isochrony; Raffi et al., 1993).

The biochronology of planktonic foraminiferal taxa used for biostratigraphic subdivision and correlation in the Pliocene and Pleistocene are listed in Tables 5 and 6, respectively. We note datum events that are restricted to the Mediterranean. Data tabulated in Berggren et al. (1985a) and believed to be still valid are cited from that source. Data developed since then come from various sources such as Dowsett (1988) for global occurrences, Weaver and Clement (1987) and Raymo et al. (1989) for the North Atlantic, Joyce et al. (1990) for the Gulf of Mexico, Srinivasan and Sinha (1993) and Chaproniere et al. (1994) for the Indo-Pacific, and Rio et al. (1990), Langereis and Hilgen (1991), and Zijderveld et al. (1991) for the Mediterranean. Finally, a Pliocene–Pleistocene (sub)tropical planktonic foraminiferal biostratigraphic zonation and magnetobiochronology, based on this and a related study (Berggren et al., 1995), is presented (Figs. 4 and 5). The standard calcareous nannofossil zonal schemes are calibrated to this chronology, also.

TABLE 5. BIOCHRONOLOGY OF PLIOCENE PLANKTONIC FORAMINIFERA

Datum event	Magnetic polarity	Age (Ma)*
Globorotalia margaritae FAD†	Mid-C3An.2n Mid-C3An.2n (Mediterranean)	6.4 (a–c) 5.32 (d, e)
Globorotalia tumida FAD	Early Gilbert reversed	5.60 (a, f, g)
Globorotalia sphericomiozea FAD	Early Gilbert reversed	5.60 (f, g)
Globorotalia pliozea FAD	Early Gilbert reversed	5.60 (f)
Globoquadrina dehiscens LAD§	Early Gilbert reversed	5.80 (a)
Sphaeroidinella dehiscens s.l. FAD	Early Gilbert reversed	5.20 (a, f)
Top *Sphaeroidinellopsis* spp.#	Basal Thvera Subchron (Mediterranean)	5.20 (d, e, i)
Globorotalia margaritae FCO**	Mid-Thvera (Mediterranean)	5.07 (d, e)
Globigerinoides seiglei LAD	Just above Sidufjall Subchron	4.70 (a)
Globorotalia puncticulata FAD	Nunivak Subchron Nunivak (Mediterranean)	4.50 (b) 4.50 (d, e, h, i, j)
Globorotalia crassaformis FAD	Nunivak Subchron Gilbert/Gauss boundary (Mediterranean)	4.5 (a, c) 3.58 (d, e)
Pulleniatina spectabilis LAD	Top Cochiti Subchron	4.20 (a)
Globoturborotalita nepenthes LAD	Top Cochiti Subchron	4.20 (a)
Globorotalia margaritae LCO††	Just above Cochiti Subchron (Mediterranean)	3.96 (d, e)
Pulleniatina§§	Just above Cochiti Subchron	3.95 (a)
Pulleniatina primalis LAD	Late Gilbert reversed	3.65 (a)
Globorotalia margaritae LAD	Gauss/Gilbert boundary Late Gilbert (Mediterranean)	3.58 (a, f) 3.79 (d, e, i, j)
Globorotalia miocenica FAD	C2An.3n (Atlantic Ocean)	3.55 (a)
Globorotalia puncticulata (temporary disappearance)	Basal Gauss (Mediterranean)	3.55 (d, e)
Globorotalia pertenuls FAD	C2An.3n	3.45 (a)
Pulleniatina (disappearance)	C2An.3n (Atlantic)	3.45 (a)
Globorotalia crassaformis (reappearance)	Below Mammoth (Mediterranean)	3.35 (d, e)
Globorotalia tosaensis FAD	Just below Mammoth	3.35 (a)
Globigerinoides fistulosus FAD	C2An.2r (base Mammoth)	3.33 (a)
Globorotalia puncticulata (reappearance)	Mammoth (C2An.2r) (Mediterranean)	3.31 (d, e)
Globorotalia sp. cf. *G. crassula* LAD	Mammoth (C2An.2r) (North Atlantic)	3.30 (b)
Globorotalia crassaformis§§	Mid-Mammoth (Mediterranean)	3.26 (d, e)
Globorotalia inflata FAD	Early Matuyama (Mediterranean) Same (North Atlantic)	2.09 (i, j, k) 2.09 (b)
Sphaeroidinella dehiscens s.s. FAD	Top Mammoth C2r.2r (Mediterranean)	3.25 (a) 2.19 (k)
Sphaeroidinellopsis seminulina LAD	Base Kaena Just above Mammoth (Mediterranean)	3.12 (b) 3.21 (d, e, h, i)
Globorotalia multicamerata LAD	C2An.1r (Kaena)	3.09 (a)
Dentoglobigerina altispira LAD	C2An.1r (Kaena) C2An.2n (late Mammoth, early Kaena) (Mediterranean)	3.09 (a) 3.17 (d, e)
Globorotalia crassaformis##	C2An.2n (late Mammoth, early Kaena) (Mediterranean)	3.17 (d, e)
Globorotalia crassaformis§§	C2An.1n (late Gauss) (Mediterranean)	3.00 (d, e)
Globorotalia crassaformis##	C2An.1n (late Gauss) (Mediterranean)	2.92 (d, e)
Neogloboquadrina atlantica (Mediterranean Pliocene entry)	C2An.ln (Mediterranean) ($\delta^{18}O$ stage 110)	2.72 (k)
Globorotalia pertenuis LAD	Top Gauss Subchron	2.60 (a)
Globorotalia truncatulinoides FAD	Gauss/Matuyama boundary (southwest Pacific) Just below Olduvai Subchron Just below Olduvai Subchron (Mediterranean)	2.58 (l) 2.00 (a) 2.00 (i, j, k)

Pliocene–Pleistocene Chronostratigraphic Framework

Geochronologic history, no matter how plausible and internally self-consistent, is still subject to stratigraphic cross examina-

TABLE 5. *(continued).*

Datum event	Magnetic polarity	Age (Ma)*
Neogloboquadrina atlantica LAD	Early Matuyama	2.41 (b)
	Early Matuyama (Mediterranean) ($\delta^{18}O$ stage 96)	2.41 (k, m)
Globorotalia puncticulata LAD	Early Matuyama (Atlantic Ocean)	2.41 (n)
	Early	2.41 (j, k)
Pulleniatina (local reappearance, Atlantic Ocean)	Early Matuyama (below Reunion Subchron)	2.30 (a)
Globorotalia miocenica LAD	Early Matuyama (below Reunion Subchron) (Atlantic Ocean only)	2.30 (a)
Globorotalia exilis LAD	Reunion Subchron (Atlantic Ocean only)	2.15 (a)
Globorotalia inflata (reappearance)	C2r.1r (Mediterranean)	1.99 (k)
Sphaeroidinella dehiscens (local LAD)	C2r.1r (Mediterranean	1.98 (k)
Neogloboquadrina pachyderma (sinistral)[#]	Top Olduvai Subchron (North Atlantic)	1.80 (b)
	Olduvai Subchron (Mediterranean)	1.80 (i, j)
Globigerinoides extremus LAD	Top Olduvai Subchron	1.77 (a)
	Top Olduvai Subchron (Mediterranean)	1.77 (i, j)
Globigerina cariacoensis FAD	Top Olduvai Subchron (Mediterranean)	1.77 (i, j)

Note: The position of the bioevents versus the magnetostratigraphy is indicated. Magnetobiochronology calibrated to the time scale shown in Table 2, column 12 (this work).

*References: a, Berggren et al. (1985a); b, Weaver and Clement (1986, 1987); c, Chaproniere et al. (1994); d, Langereis and Hilgen (1991); e, Hilgen (1991b); f, Srinivasan and Sinha (1993); g, Hodell and Kennett (1986); h, Channell et al. (1988); i, Glaçon et al. (1990); j, Channell et al. (1990); k, Zijderveld et al. (1991); l, Dowsett (1988); m, Zachariasse et al. (1989); n, Raymo et al. (1989).

†FAD = first appearance datum.

§LAD = last appearance datum.

#Acme.

**FCO = first common occurrence.

††LCO = last common occurrence.

§§Sinistral to dextral coiling change.

##Dextral to sinistral coiling change.

tion. Thus the Neogene, as a biochron, is framed in the biostratigraphy of the Mediterranean region (Lyell, 1833, 1839, 1857) and it is to *this* biostratigraphy that information from stratigraphic sequences of Neogene age, whether in the deep sea or in land sections, must be conveyed if such information is to have more than provincial significance. The means of conveyance is geochronology in the form of a Neogene time scale (e.g., Berggren et al., 1985a, 1985b; see discussion below). In the section below we review the present status of late Neogene (Pliocene and Pleistocene) chronostratigraphy, with particular reference to the epoch/series boundaries.

The term *Neogene* (Hörnes, 1853, 1864), like its counterpart, *Paleogene* (Naumann, 1858), was originally used in a biostratigraphic, rather than chronostratigraphic sense. These terms were introduced to characterize the faunas and floras of the Cenozoic Era. The Neogene was originally created to characterize the faunal (particularly molluscan) and floral changes that denote the middle part of the Cenozoic (beginning, but not clearly delimited, at the Oligocene/Miocene boundary) and that continue to the present day. In the course of preparing a monograph on the molluscan fauna of the Vienna Basin, Hörnes noted that Miocene and Pliocene faunas exhibited a closer relationship to each other than with those of the Eocene, sensu Lyell (he did not recognize a separate Oligocene or Pleistocene at the time). In creating the term *Neogene*, Hörnes (1853, 1864) referred to the biostratigraphic subdivision of the Tertiary and Quaternary made by the Austrian paleontologist Bronn (1838). Bronn's fifth period (termed the *Molasse Gebirge*) was subdivided into three groups (termed the *Molasse Gruppen*). The first was equated with Lyell's Eocene Epoch. The second group, the Molasse Group proper, included the Miocene and Pliocene Epochs sensu Lyell. The Miocene and Pliocene were treated as subunits within the second group and referred to as the *untere Abtheilung* and *obige Abtheilung*, respectively. The third Molasse Group contained *Alluvial und Quartär-Gebilde zum Theile*, the Pleistocene proper (*Knochen-Höhlen und der Loss*) already having been incorporated into the upper part of his second Molasse Group. The original Neogene concept thus included a priori all post-Eocene molluscan faunas, enclosing rock strata of the Vienna Basin up to and including those in glacial Löss and Diluvian deposits as well as correlative Mediterranean faunas in Sicily, Rhodes, and Cyprus that would now be included in the Pleistocene. It will be recalled that Lyell (1833) created the terms Eocene, Miocene, and Pliocene, and subsequently (Lyell, 1839, 1857) subdivided the latter into an Older (= Pliocene) and Younger (= Pleistocene) Pliocene. He never recognized the term Oligocene.

The Neogene, conceptually, when transferred from a biostratigraphic/biochrono-

TABLE 6. BIOCHRONOLOGY OF PLEISTOCENE PLANKTONIC FORAMINIFERA

Datum event	Isotope stratigraphy (magnetic polarity)	Age (Ma)
Globigerinoides fistulosus LAD[†]	Matuyama (just above Olduvai Subchron)	1.6 (a)
Pulleniatina finalis LAD	Matuyama (midway between Jaramillo and Olduvai Subchrons)	1.40 (a, b)
Globorotalia truncatulinoides excelsa FAD[§]	Middle Jaramillo Subchron (Mediterranean)	1.00 (c, d)
Globorotalia crassaformis hessi FAD	Basal Brunhes	0.75 (b)
Globorotalia tosaensis LAD	Early Brunhes	0.65 (a, e)
	Latest Brunhes–Jaramillo	0.8–1.0 (b)
Globorotalia hirsuta FAD	Middle Brunhes ($\delta^{18}0$ stage 12)	0.45 (f)
Globorotalia flexuosa FAD	Middle Brunhes ($\delta^{18}0$ stage 11)	0.401 (g)
Bolliella calida FAD	Late Brunhes ($\delta^{18}0$ stage 7)	0.22 (b)
Globoquadrina pseudofoliata LAD	Late Brunhes ($\delta^{18}0$ stage 7)	0.22 (b)
Globorotalia flexuosa LAD	Late Brunhes ($\delta^{18}0$ stage 4)	0.068 (g)

Note: The position of bioevents versus magnetostratigraphy and/or isotopic stratigraphy is indicated. Magnetobiochronology calibrated to the time scale shown in Table 2, column 12 (this work).

*References: a, Berggren et al. (1985a); b, Chaproniere et al. (1994); c, Glaçon et al. (1990); d, Channel et al. (1990); e, Srinivasan and Sinha (1993); f, Pujol and Duprat (1983); g, Joyce et al. (1990).

†LAD = last appearance datum.

§FAD = first appearance datum.

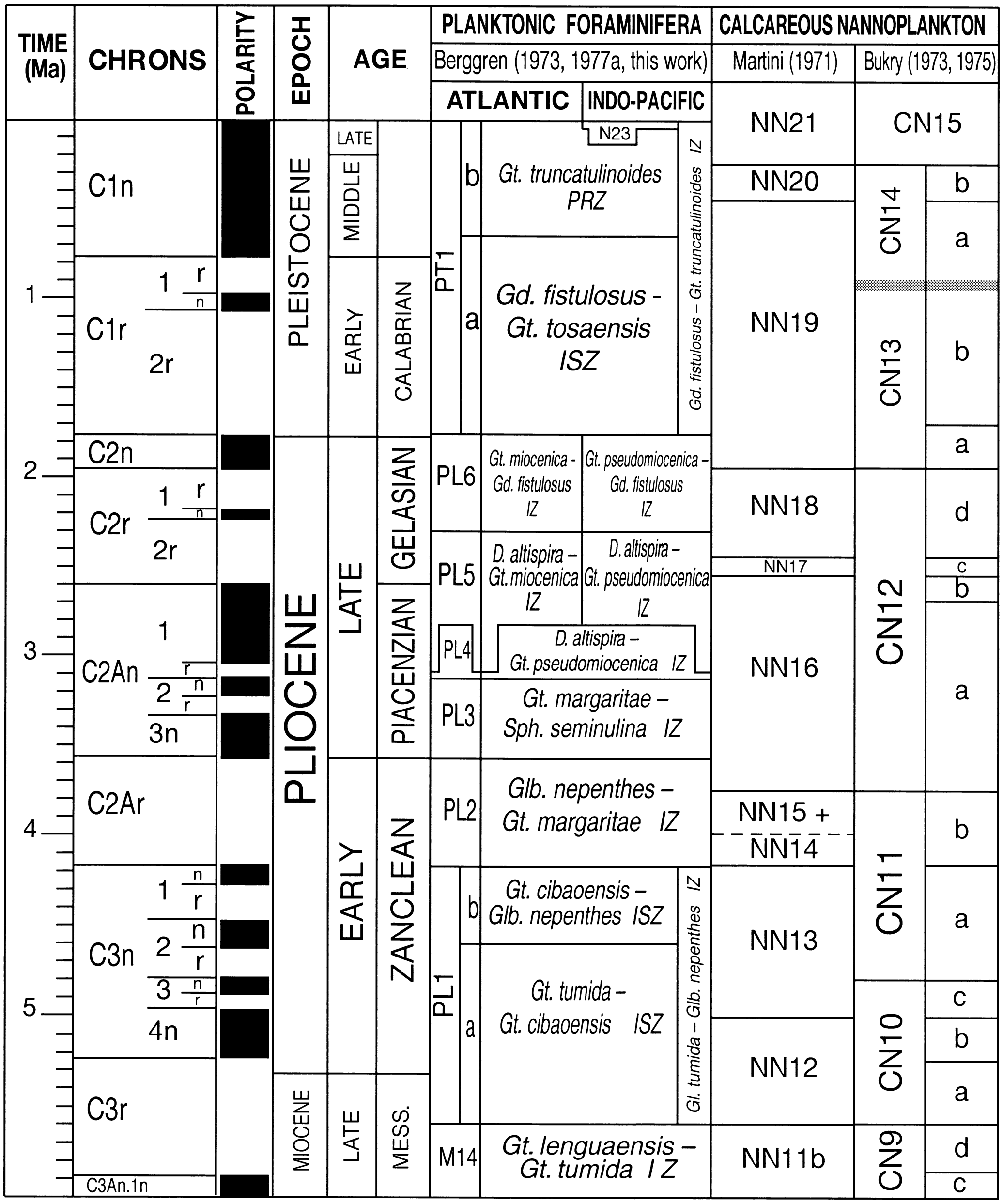

Figure 4. Late Neogene (Pliocene–Pleistocene) (sub)-tropical calcareous plankton biostratigraphy and magneto-biochronology.

logic (Lyellian) to a chronostratigraphic connotation, includes the stratigraphic record of the Miocene, Pliocene, and Pleistocene. The terms *Tertiary* and *Quaternary* are regarded here as inappropriate to the chronostratigraphic hagiography; we prefer a usage that subdivides the Cenozoic Era into two periods/systems, Paleogene and Neogene, with the latter consisting of the Miocene, Pliocene, and Pleistocene Epoch/Series. Harland et al. (1982, 1990) subdivided the Cenozoic Era into Tertiary ("useful and unambiguous") and Quaternary ("not a satisfactory name") Sub-eras, noting that the Tertiary has been replaced elsewhere by the terms Paleogene and Neogene (see Bassett, 1985, and Cowie and Bassett, 1989). The Neogene Period was limited to the Miocene and Pliocene Epochs and the Anthropogene or, alternatively, the Pleistogene Period was equated with the Pleistocene and Holocene Epochs. We have provided historical evidence above for including Pliocene and Pleistocene (sensu Lyell) in the Neogene.

Current practice in establishing global chronostratigraphic boundaries links the hierarchical elements in a lock-step equivalency. Thus, the series/epoch boundary coincides with the limits of its included stage/age boundary in a nested hierarchy, with the lower boundary of a series being drawn at the base of its lowest included stage (see Hedberg, ed., 1978; Schoch, 1989). There are problems associated with this approach, and the Miocene/Pliocene boundary (Messinian/Zanclean stage boundary) serves as a case in point.

The Pliocene Epoch

The Pliocene is traditionally considered to have begun with the restoration of normal marine conditions in the Mediterranean Basin following the late Miocene (Messinian)

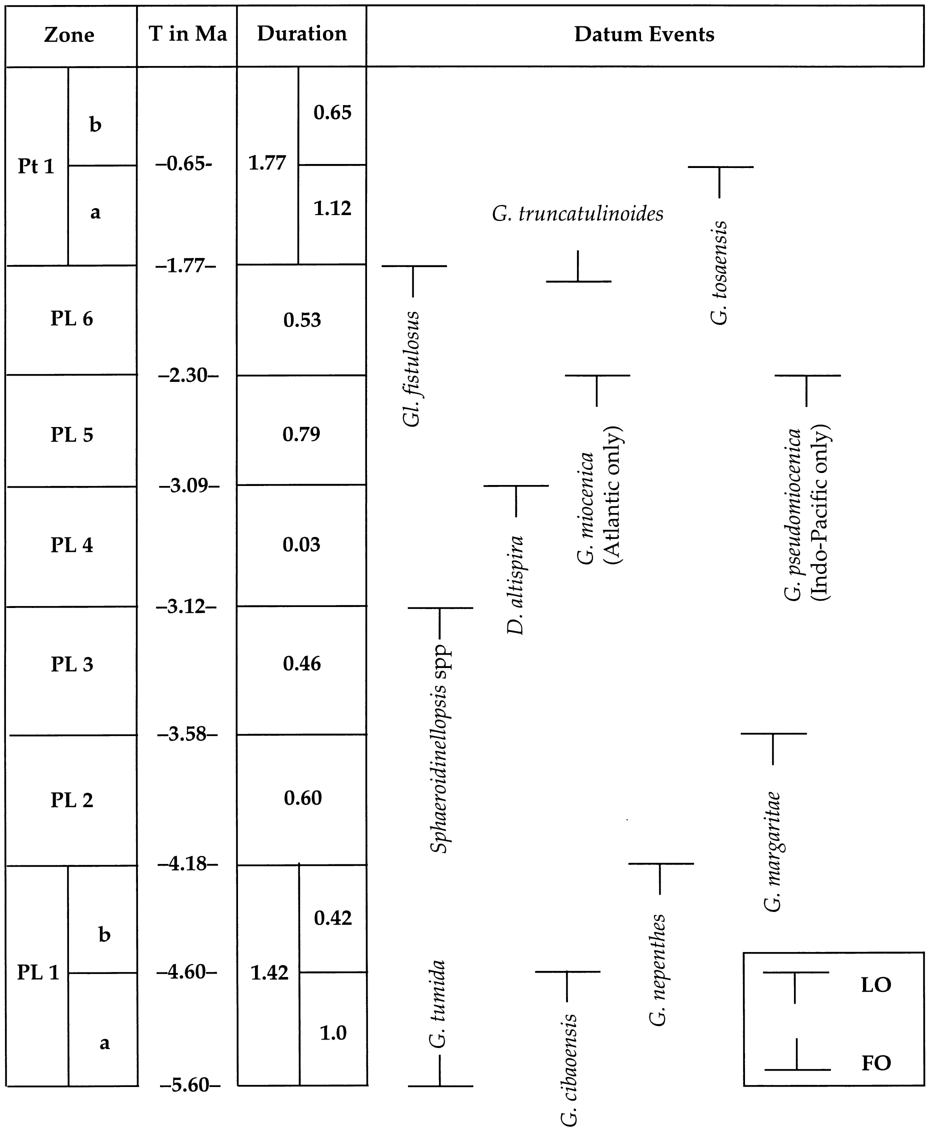

Figure 5. Late Neogene (Pliocene–Pleistocene) (sub)tropical planktonic formainiferal datum events used in developing a standard planktonic formainiferal zonation (Berggren et al., 1995). FO = first occurrence; LO = last occurrence.

desiccation, when the Mediterranean became essentially a basin of interior drainage with only sporadic connection with the global ocean. Cita (1975) proposed the establishment of a boundary stratotype at Capo Rosello (Sicily) at the base of the Trubi Formation (Zanclean Stage), which lies unconformably upon the terminal Messinian brackish water Arenazzola Formation. Recent integrated magnetobiostratigraphic studies (Zijderveld et al., 1986; Hilgen and Langereis, 1988; Hilgen, 1991a, 1991b; Channell et al., 1988) indicate that the base of the Trubi Formation in the Capo Rosello region and Eraclea Minoa sections (Sicily) and Singa and Capo Spartivento sections (Calabria) lies within the lower (reversed) part of the Gilbert Magnetozone. The estimated age ranges from 4.86 Ma (Hilgen and Langereis, 1988) to 4.93 Ma (Channell et al., 1988) based on downward extrapolation of sedimentation rates and precessional cycle counting within the sedimentary sequence between the base of the Thvera Subchron and above the regional unconformity below. Similar results have been obtained for western Mediterranean ODP Sites 652 and 654 (Channell et al., 1990). A more recent age estimate of this stratigraphic level, using an astronomically calibrated late Neogene time scale, is ca. 5.32 Ma (Hilgen, 1991b).

The currently accepted Miocene/Pliocene boundary definition, as the base of the Trubi Formation at Capo Rosello, is difficult to characterize biostratigraphically in the Mediterranean. An assessment of available data on magnetobiostratigraphic correlations of calcareous plankton to the Mediterranean region indicates the following: (1) None of the biostratigraphic events commonly used to estimate the lithostratigraphic position of the Miocene/Pliocene boundary in the global ocean are suitable for precise correlation within the Mediterranean Basin because of biogeographic or environmental exclusion or delayed entry (Rio et al., 1984; 1991). (2) Interpretation of meager biostratigraphic data as well as magnetostratigraphic data from ODP Sites 652 and 654 (Channell et al., 1990) and ODP Leg 138 (Raffi and Flores, in press) indicate that, globally, the base of the Zanclean Stage is probably bracketed by the highest occurrence of *Discoaster quinqueramus* in the middle part of the lower reversed interval of the Gilbert Chronozone (Chron C3r), and the lowest occurrence of *Ceratolithus acutus* and highest occurrence of *Triquetrorhabdulus rugosus* in the uppermost part of the lowest reversed interval of the Gilbert Magnetozone (Chron C3r). This latter pair of events occur ≈6 m above the base of the Trubi Formation at Capo Rosello (Cita and Gartner, 1973; Rio et al., 1984) and Capo Spartivento (Raffi, unpubl. data). Note that the *D. quinqueramus* extinction datum is not observed in the Mediterranean because it occurred during the salinity crisis. (3) In terms of planktonic foraminifera, the base of the Zanclean would appear to be bracketed by the lowest occurrence of *Globorotalia puncticulata* in the Nunivak Subchronozone, and the lowest occurrence of *Globorotalia tumida* in the lower part of the Gilbert Chronozone (this latter event is not recognized in the Mediterranean).

While the lower limit of the Pliocene (Zanclean Stage) can be recognized magnetostratigraphically, it has been argued (Rio et al., 1991) that the peri-Mediterranean lithologic break between nonmarine (below) and marine (above) sediments at the "boundary" renders this region in general, and the Capo Rosello composite section in particular, unsuitable for definition of a chronostratigraphic boundary as well as unamenable to precise biostratigraphic correlation. Indeed, these authors indicate (Rio et al., 1991) that "in order to properly define the Miocene/Pliocene boundary in a continuous marine sequence, a boundary stratotype should be selected from outside the Mediterranean region."

Two approaches to this dilemma have been suggested. One could determine the position of the base of the Zanclean Stage, as defined in the Mediterranean (Sicilian) section(s), in a continuous marine extra-Mediterranean section and place the Global Stratotype Section and Point (GSSP), the so-called "golden spike", at that level. This approach has been defended recently by Hilgen and Langereis (1993). A second approach would be to identify some recognizable biostratigraphic, magnetostratigraphic, or isotopic event in a continuous marine sequence spanning the interval represented by the unconformity separating the base of the Zanclean and top of the Messinian Stages and place the GSSP "golden spike" there. This alternative has been recommended by Benson and Hodell (1994) in a response to Hilgen and Langereis (1993).

Hilgen and Langereis (1993) propose that the base of the Pliocene can be recognized in the composite Italian stratotype(s) by counting down five precession cycles below the Thvera (C3n.4n) Subchron. Using this approach, they derive an age of 5.32 Ma for the oldest sediments of the Trubi Chalk Formation resting unconformably on the Messinian sediments.

Benson (1994, personal commun. to Berggren) and Benson and Hodell (1994) have discussed the problems of linking the Messinian/Zanclean boundary with the Miocene/Pliocene boundary and render the following observations: (1) The boundary, as currently accepted, is situated close to (if not virtually contiguous with) an unconformity and is based essentially on a local to regional event (i.e., establishment of the neo-Mediterranean Sea following the Messinian Salinity Crisis). Locating the boundary at this level does not allow the sequence of historical events directly prior to the boundary point to be recognized and thus precludes precise biostratigraphic recognition and correlation elsewhere. (2) The boundary in Sicily, as determined by Hilgen and Langereis (1993) by counting down five precession cycles below the base of the Thvera Submagnetozone, is not globally correlatable, and the determination of the boundary on the basis of cyclostratigraphy in the absence of a clearly determined magnetobiostratigraphic and/or isotopic framework renders correlation outside the Mediterranean region tenuous at best. (3) There is a difference between criteria chosen to designate a GSSP (a lithic horizon in a continuous section) and the means by which this horizon may be recognized and correlated away from the type section—usually a fossil datum. Defining the base of the Pliocene as the reestablishment of marine conditions in the Mediterranean establishes this as the criterion by which the base Pliocene is recognized elsewhere, obviously an impossibility. Rather, the Miocene/Pliocene boundary must be defined by a GSSP at a specific horizon, then correlated elsewhere by means of a denotative character (datum event, magnetostratigraphic signature, isotope feature, etc.). (4) Nineteenth-century subdivisions of the Cenozoic by Lyell, d'Orbigny, Mayer-Eymar, and others were imprecise and biostratigraphic and/or lithostratigraphic (Berggren, 1971) in concept, and boundaries were located frequently at unconformities. The boundary, as currently accepted, separates faunas (primarily molluscan) that historically exhibit a "Miocene" and "Pliocene" aspect, respectively; however, Pliocene molluscan faunas continue to exhibit a Miocene aspect well into the Pliocene. Calcareous plankton faunas and floras, characteristic of the early Pliocene, were already becoming established >1 m.y. earlier in the Atlantic and Indo-Pacific Oceans (i.e., outside the Mediterranean).

Accordingly, Benson and Hodell (1994) have argued for recognizing the Messinian and Zanclean Stages for what they are, namely the local regional expression of particular historical events in the late Neogene evolution of the Mediterranean, and decoupling them from the question of the Miocene/Pliocene boundary. They propose linking the Miocene/Pliocene boundary with the Chronozone C3r/C3An boundary in a newly proposed boundary stratotype in the Bou Regreg section near Rabat, northwestern Morocco, in a continuous marine sequence of Atlantic facies. (Note that this is essentially the same position of the Miocene/Pliocene boundary that was thought to correlate with the base of the Zanclean Stage [Cita, 1975; Berggren et al., 1985a, 1985b].)

Hilgen and Langereis (1994) responded as follows: (1) They have not proposed that the Miocene/Pliocene boundary be determined solely on a cyclostratigraphic approach but rather within an astronomically calibrated cyclostratigraphic framework that incorporates an integrated bio-magneto-isotopic stratigraphy. (2) The current boundary is already located very close (within ≈100 000 yr) to the next younger (Thvera = C3n.4n Subchron) geomagnetic polarity boundary, which is easily discernible. (3) Magnetostratigraphic data from Bou Regreg is inadequate. (4) It would be more appropriate to determine the precise position of the present boundary using an astronomically calibrated cyclostratigraphic framework in the open ocean and Moroccan sequences and then determine which biostratigraphic datums are most closely associated with this level. The option to define the actual boundary in the Mediterranean (Eraclea Minoa), Moroccan (Bou Regreg), or other location(s) could then be determined at some future date.

This debate has essentially come full circle and we wish to point out that two Miocene/Pliocene boundaries are currently vying for recognition, and ultimately acceptance, by the International Union of Geological Sciences (IUGS) commission on stratigraphy (ICS). They are (1) the base of the Zanclean Stage in the Capo Rosello composite section, which lies within the upper part of the lower (reversed) interval of the Gilbert (Chronozone C3r), and (2) the base of the Gilbert (Chronozone C3r/C3An boundary). The latter boundary has not been formally proposed to the ICS for consideration but is commonly used in paleoceanographic studies.

The Pliocene is generally subdivided into lower (Zanclean) and upper (Piacenzian) stage units, but Rio et al. (1991) have argued recently for a tripartite subdivision into Zanclean (lower) and Piacenzian (middle) Stages and an unnamed unit for the upper Pliocene (the name Gelasian has been proposed recently by Rio et al. (1994) for this stratigraphic interval). They believe that none of the stage names in current use adequately represent the upper Pliocene (ca. 2.6–1.8 Ma in the chronology adopted here). Both the top of the Zanclean and the base of the Piacenzian stratotypes are marked by unconformities (Rio et al., 1984, 1991). The base of the Piacenzian stratotype is now considered (Rio et al., 1991, 1994) to lie slightly above the highest occurrence of *Globorotalia margaritae*, between the highest occurrence of *Reticulofenestra pseudoumbilica* and the highest occurrence of *Globorotalia puncticulata* (i.e., within the younger part of the Gilbert [Chron C2r]). The top of the Piacenzian stratotype lies between the highest occurrence of *Discoaster tamalis* and the highest occurrence of *D. pentaradiatus*, within the confines of the relatively short *D. pentaradiatus* (NN17) biochron, which essentially straddles the Gauss/Matuyama boundary (Rio et al., 1991; Fig. 4). The base of the proposed Gelasian Stage is situated slightly below the highest occurrence of *Discoaster pentaradiatus* and is correlative with the major growth/expansion of northern hemisphere glaciation near the Gauss/Matuyama boundary.

The Pleistocene Epoch

Attempts at formulating a coherent and

unified chronostratigraphy of the Pleistocene have suffered from the climatic overprint that characterizes the late Neogene part of Earth history (see discussions in Hays and Berggren, 1971; Berggren and Van Couvering, 1982). While the formal terminology for the classic glacial/interglacial deposits has been based on northern European sequences, the standard chronostratigraphic terminology is based on Mediterranean (Italian) sections. These demonstrably incomplete stratigraphic sections have been correlated to the continuous oxygen isotope stratigraphy of the oceans for the late Pleistocene (Shackleton and Opdyke, 1973, 1976), which have, in turn, been extended into the early Pleistocene and Pliocene (Ruddiman et al., 1986; Raymo et al., 1989; Joyce et al., 1990).

Following a decade of study and discussions by the International Union for Quaternary Research (INQUA) Subcommission 1a (Pliocene/Pleistocene Boundary) on Stratigraphy and International Geological Correlation Programme Project 41 (Neogene/Quaternary Boundary), a draft proposal on the choice of a boundary stratotype for the Pliocene/Pleistocene boundary was submitted and approved by the INQUA Commission on Stratigraphy and the Subcommission on Quaternary Stratigraphy of the ICS at the 1982 Moscow INQUA Congress. A formal proposal was subsequently submitted to and accepted by the ICS in 1983 (see review by Pasini and Colalongo, 1982), and published 2 yr later (Aguirre and Pasini, 1985) together with the announcement (Bassett, 1985) that the content of the proposal had been formally ratified by the IUGS executive as the GSSP for the base of the Pleistocene.

The "golden spike" was placed at the base of a claystone unit conformably overlying the sapropelic marker bed *e* in the Vrica section of Calabria in southern Italy. This level has been shown to lie at or near the top of the Olduvai (C2n) Subchronozone. Current attempts to relocate the Pliocene/Pleistocene boundary to coincide with the late Pliocene climatic change at ca. 2.6 Ma are judged here to be inappropriate and to ignore stratigraphic first principles.

With respect to intra-Pleistocene chronostratigraphic subdivision, considerable controversy surrounds the use of lower Pleistocene stage names. As the Calabrian Stage, as stratotypified at Santa Maria di Catanzaro, is probably the approximate stratigraphic correlative of the supposedly suprajacent Sicilian Stage (Ruggieri and Sprovieri, 1979), it has been suggested that the term *Calabrian* be abandoned and supplanted by the Selinuntian Superstage (with the Santernian, Emilian, and Sicilian as sequential stages within). Ruggieri et al. (1984) subsequently reduced the Selinuntian to stage status and the other three terms to chronozone or substage rank. The Selinuntian had been defined (Ruggieri and Sprovieri, 1979) as the interval between the Pliocene/Pleistocene boundary stratotype and the boundary between the lower and middle Pleistocene (at the top of the Sicilian).

Van Couvering (in press) has argued elegantly and forcefully that this procedure is inappropriate and only exacerbates the problem of early Pleistocene chronostratigraphy. He correctly observes that the boundaries of lower chronostratigraphic units (e.g., Selinuntian) cannot be defined ab initio in terms of the established boundaries of higher units (epoch/series) as this is contrary to logic and the rules of stratigraphic procedure. Thus, the base of the Quaternary, a system of the Cenozoic in common use, is defined at and by the base of the included Pleistocene Series and the Pleistocene, in turn, by the base of its included stage(s). The location of the GSSP in the Calabrian section at Vrica has the serendipitous effect of making that level the Calabrian boundary stratotype.

In the interest of nomenclatorial stability, it is advisable to revise the definition of the Calabrian Stage and preserve a well-known and long-used name. This would appear to be the approach adopted at a recent meeting of the Italian Working Group for Quaternary Stratigraphy (Bari-Taranto, Italy, September 1994). The Santernian would then be the appropriate lowermost substage of the Calabrian Stage. A further step in resolving the chronostratigraphic subdivision of the lower Pleistocene is the recent proposal for the erection of a Santernian/Emilian boundary stratotype 6 m above marker bed "o" (71.1 m above the Pliocene/Pleistocene GSSP) in the Vrica section, approximately coincident with the local lowest occurrence of *Hyalinea baltica* and the lowest occurrence of "large" *Gephyrocapsa.*

SUMMARY AND CONCLUSIONS

(1) We recommend a twofold subdivision of the Cenozoic Era/Erathem into the Paleogene (Paleocene, Eocene, Oligocene) and Neogene System/Series based on a review of historical data and modern stratigraphic usage. The Neogene Period/System encompasses all post-Oligocene chronostratigraphic (epoch/series) units: Miocene, Pliocene, Pleistocene, and Holocene.

(2) Astrochronology (or astronomical calibration of climatic proxy records) offers a major step forward in the development of a precise and accurate time scale for the Neogene. The high resolution provided by astrochronology implies that it is no longer necessary to interpolate between a few discrete calibration points to construct a geological time scale. Likewise, it is now possible to unravel the complex phase relationships between various components in the ocean-climate system and between these components and radiation forcing.

(3) Apparent discrepancies between the recently developed late Neogene astronomical time scale and the GPTS have been resolved by means of significant improvements to isotopic dating ($^{39}Ar/^{40}Ar$) of the younger part of this interval (Chrons C1n–C2An) and revised analysis of sea-floor anomaly patterns in the older part (Chrons C2Ar–C3r). The two chronologies are now concordant back to the Chron C3/C3An boundary at ca. 5.89 Ma. The ages of all polarity boundaries are provided in Table 2 together with a historical review of earlier chronologies.

(4) Correlation of Pliocene and Pleistocene calcareous nannoplankton and planktonic foraminiferal biostratigraphic datum events to the GPTS provides a new high-resolution magnetobiochronology for regional and global correlations.

(5) Current research on the subdivision of the Pliocene and Pleistocene of the Mediterranean region is devoted to establishing chronostratigraphic stability in the form of stage definitions and stage boundary GSSPs. A formalized threefold subdivision of the Pliocene Series has been proposed recently with the recommendation of adding an upper Pliocene Gelasian Stage to complement the well-founded Zanclean and Piacenzian Stages (lower and middle Pliocene, respectively). The Calabrian Stage appears to be winning back friends as the appropriate name for the lower Pleistocene (with the Santernian, Emilian, and Sicilian as component substages). A GSSP has been proposed for the Santernian/Emilian boundary in the Vrica section, 71 m above the Pliocene/Pleistocene boundary GSSP.

POSTSCRIPT

At the recently convened 10th Regional Committee of Mediterranean Neogene Stratigraphy (RCMNS) in Bucharest, Romania, a meeting of the Subcommission on Neogene Stratigraphy (SNS) was held on 6 September 1995 concerning the placement of the Miocene/Pliocene boundary GSSP (Cita, 1995). An alternative, third proposal was introduced by J.-P. Suc to define the GSSP at a stratigraphic level in the Bou Regreg section correlative with the reestablishment of open marine conditions in the Mediterranean. The SNS agreed to a 2 yr moratorium during which the Capo Rosello and Bou Regreg sections will be examined in greater detail before a final decision is recommended.

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