5.10 Geomagnetic Excursions

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5.10.1 Introduction

5.10.1.1 History of the Polarity Timescale and Excursions

David (1904) and Brunhes (1906) were the first to measure magnetization directions in rocks that were approximately antiparallel to the present Earth's field. Brunhes (1906) recorded magnetizations in baked sedimentary rocks that were aligned with reverse magnetization directions in overlying Miocene lavas from central France (Puy de Dome). In so doing, Brunhes (1906) made first use of a field test for primary thermal remanent magnetization (TRM) that is now referred to as the 'baked contact' test (see Laj *et al.* (2002) for an account of Brunhes' work). Matuyama (1929) was the first to attribute reverse magnetizations in (volcanic) rocks from Japan and China to reversal of geomagnetic polarity, and to differentiate mainly Pleistocene lavas from mainly Pliocene lavas based on the polarity of the magnetization. In this respect, Matuyama (1929) was the first person to use magnetic stratigraphy as a means of ordering rock sequences.

The modern era of paleomagnetic studies began with the studies of Hospers (1951, 1953–54) in Iceland, and Roche (1950, 1951, 1956) in the Massif Central of France. The work of Hospers on Icelandic lavas was augmented by Rutten and Wensink (1960) and Wensink (1964) who subdivided Pliocene–Pleistocene lavas on Iceland into three polarity zones from young to old: N–R–N. Magnetic remanence measurements on basaltic lavas combined with K/Ar dating, pioneered by Cox *et al.* (1963) and McDougall and Tarling (1963a, 1963b, 1964), resulted in the beginning of development of the modern geomagnetic polarity timescale (GPTS). These studies, and those that followed in the mid-1960s, established that rocks of the same age carry the same magnetization polarity, at least for the last few million years. The basalt sampling sites were scattered over the globe. Polarity zones were linked by their K/Ar ages, and were usually not in stratigraphic superposition. Doell and Dalrymple (1966) designated the long intervals of geomagnetic polarity of the last 5 My as magnetic epochs, and named them after pioneers of geomagnetism (Brunhes, Matuyama, Gauss, and Gilbert). The shorter polarity intervals (events) were named after localities: e.g., Jaramillo (Doell and Dalrymple, 1966), Olduvai (Grommé and Hay, 1963, 1971), Kaena and Reunion (McDougall and Chamalaun, 1966), Mammoth (Cox *et al.*, 1963), and Nunivak (Hoare *et al.*, 1968). The nomenclature for excursions has continued this trend by naming excursions after the localities from where the earliest records were derived (e.g., Laschamps in France).

The fit of the land-derived polarity timescale, from paleomagnetic and K/Ar studies of exposed basalts, with the polarity record emerging from marine magnetic anomalies (MMAs) (Vine and Matthews, 1963; Vine, 1966; Pitman and Heirtzler, 1966; Heirtzler *et al.*, 1968) resulted in a convincing argument for synchronous global geomagnetic polarity reversals, thereby attributing them to the main axial dipole. The results relegated self-reversal, detected in the 1950s in the Haruna dacite from the Kwa district of Japan (Nagata, 1952; Nagata *et al.*, 1957; Uyeda, 1958), to a rock-magnetic curiosity rather than to a process generally applicable to volcanic rocks.

The first magnetic stratigraphies in sedimentary rocks may be attributed to Creer *et al.* (1954) and Irving and Runcorn (1957) who documented normal and reverse polarities at 13 locations of the Proterozoic Torridonian Sandstones in Scotland, and in rocks of Devonian and Triassic age. For the Torridonian Sandstones, normal and reverse magnetizations were observed from multiple outcrops, and an attempt was made to arrange the observed polarity zones in stratigraphic sequence. Meanwhile, Khramov (1958) published magnetic polarity stratigraphies in Pliocene–Pleistocene sediments from western Turkmenia (central Asia), and made chronostratigraphic interpretations based on equal duration of polarity intervals. Early magnetostratigraphic studies were carried out on Triassic red sandstones of the Chugwater Formation (Picard, 1964), on the European Triassic Bundsandstein (Burek, 1967, 1970), and on the Lower Triassic Moenkopi Formation (Helsley, 1969). The above studies were conducted on poorly fossiliferous mainly continental (red) sandstones and siltstones; therefore, the correlations of polarity zones did not have support from biostratigraphic correlations. It was the early magnetostratigraphic studies of Plio-Pleistocene marine sediments, recovered by piston coring in high southern latitudes (Opdyke *et al.*, 1966; Ninkovitch *et al.*, 1966; Hays and Opdyke, 1967) and in the equatorial Pacific Ocean (Hays *et al.*, 1969), that mark the beginning of modern magnetic stratigraphy. These studies combined magnetic polarity stratigraphy and biostratigraphy and in so doing refined and extended the GPTS derived from paleomagnetic studies of basaltic outcrops (e.g., Doell *et al.*, 1966).

Heirtzler *et al.* (1968) produced a polarity timescale for the last 80 My using South Atlantic MMA record (V-20) as the polarity template. They assumed constant spreading rate, and extrapolated ages of polarity chrons using an age of 3.35 Ma for the Gilbert/Gauss polarity chron boundary. This was a dramatic step forward that extended the GPTS to $\sim$ 80 Ma, from $\sim$5 Ma based on magnetostratigraphic records available at the time (e.g., Hays and Opdyke, 1967). Heirtzler *et al.* (1968) made the inspired choice of a particular South Atlantic MMA record (V-20). Several iterations of the polarity timescale (e.g., Labrecque *et al.*, 1977) culminated with the work of Cande and Kent (1992a) that re-evaluated the MMA record, and used the South Atlantic as the fundamental template with inserts from faster spreading centers in the Indian and Pacific Oceans. They then interpolated between nine numerical age estimates that could be linked to polarity chrons over the last 84 My. Eight of these ages were based on 'high-temperature' radiometric ages (no glauconite ages were included), and the youngest age (2.60 Ma) was the astrochronological age estimate for the Gauss/

Matuyama boundary from Shackleton *et al.* (1990). In the subsequent version of their timescale, hereafter referred to as CK95, Cande and Kent (1995) adopted astrochronological age estimates for all Pliocene–Pleistocene polarity reversals (Shackleton *et al.*, 1990; Hilgen, 1991a, 1991b) and modified the age tie-point at the Cretaceous–Tertiary boundary from 66 to 65 Ma.

The astrochronological estimates for Pliocene–Pleistocene polarity chrons (Shackleton *et al.*, 1990; Hilgen, 1991a, 1991b), incorporated in CK95, have not undergone major modification in the last 15 years, and indicate the way forward for timescale calibration further back in time. Subsequent to CK95, astrochronological estimates of polarity chron ages have been extended into the Miocene, Paleogene, and Cretaceous (Krijgsman *et al.*, 1994, 1995; Hilgen *et al.*, 1995, 2000; Abdul-Aziz, 2000, 2003, 2004). Many of these advances are included in the timescale of Lourens *et al.* (2004). At present, more or less continuous astrochronologies tied to polarity chrons are available back to the Oligocene (e.g., Billups *et al.*, 2004).

Recognition of brief polarity excursions as an integral part of the Earth's paleomagnetic field behavior has developed alongside astrochronological calibration of the polarity timescale in the last 20 years. Although the first recognized polarity excursions (the Laschamp and Blake excursions) were documented in the late 1960s (Bonhommet and Babkine, 1967; Smith and Foster, 1969), excursions were widely considered to represent either spurious recording artifacts or, at best, local anomalies of the geomagnetic field and of doubtful utility in stratigraphy. It was not until the late 1980s that the tide began to turn. As high-resolution sedimentary records from the deep sea became available, it became accepted that excursions are frequent (with perhaps more than seven in the Brunhes Chron) and that they appear to be globally recorded and not local anomalies of the geomagnetic field. Geomagnetic excursions were not recorded in the early days of magnetic stratigraphy because the high sedimentation rate sequences, required to record brief (few kiloyears duration) magnetic excursions, were generally not targeted during conventional piston coring expeditions. Indeed, at the time, low sedimentation rate sequences were often preferred as they allowed the record to be pushed further back in time. The development of hydraulic piston coring (HPC) techniques, first used in early 1979 during the Deep Sea Drilling Project (DSDP) Leg 64, brought high sedimentation rate sequences within reach by increasing penetration by a factor of $\sim$15, from $\sim$20 m for conventional piston coring to $\sim$300 m for the HPC.

5.10.1.2 Nomenclature for Excursions and Polarity Intervals

The International Stratigraphic Commission (ISC) has guided the use of magnetostratigraphic units (polarity zones), their time equivalents (polarity chrons), and chronostratigraphic units (polarity chronozones) (see Anonymous, 1977; Opdyke and Channell, 1996). The revelation in the last 20 years of numerous short-lived excursions within the Brunhes and Matuyama chrons, and probably throughout the history of the Earth's magnetic field, requires an extension of terminology to include excursions and brief polarity microchrons (**Table 1**). Terminology based on duration of polarity intervals is obviously problematic at the low end of the duration spectrum. Records are always compromised by limitations of the recording medium and, in addition, estimates of duration are limited by the availability of chronological tools of adequate precision. In the realm of MMAs, La Brecque *et al.* (1977) applied the term 'tiny wiggles' to minor, but lineated, MMAs. The time equivalent of the 'tiny wiggle' was labeled 'cryptochron' (Cande and Kent, 1992a, 1992b), expressing the uncertain origin of 'tiny

Table 1 Nomenclature for polarity intervals and excursions

Magneto-stratigraphic polarity zone	Geochronologic (time) equivalent	Chronostratigraphic equivalent	Duration (yr)
Polarity megazone	Megachron	Megachronozone	10^8–10^9
Polarity superzone	Superchron	Superchronozone	10^7–10^8
Polarity zone	Chron	Chronozone	10^6–10^7
Polarity subzone	Subchron	Subchronozone	10^5–10^6
Polarity microzone	Microchron	Microchronozone	$<10^5$
Excursion zone	Excursion		Brief departure from normal secular variation
Polarity cryptochron	Cryptochron	Cryptochronozone	Uncertain existance

wiggles' as either short-lived polarity intervals or paleointensity fluctuations. These authors placed the duration separating 'polarity chrons' and 'cryptochrons' at 30 ky, representing an estimate of the minimum duration of polarity intervals that can be resolved in MMA records. For magnetostratigraphy, Krijgsman and Kent (2004) advocated a duration cut-off for separating 'subchrons' and 'excursions', at 9–15 ky . The drawback of such a scheme is that, in most cases, the chronostratigraphic precision will be inadequate to establish the distinction. An alternative is that an excursion be defined in terms of a 'brief' ($<10^4$ years) deviation of virtual geomagnetic poles (VGPs) from the geocentric axial dipole (GAD) that lies outside the range of secular variation for a particular population of VGPs. Some authors have adopted an arbitrary VGP cut-off (say a co-latitude of 45°) to define an excursion, although Vandamme (1994) has advocated a method of calculating the cut-off for a specific VGP population based on VGPs lying outside 'normal' secular variation. As higher fidelity records of excursions have become available, it appears that the majority of 'excursions' is manifest as directional changes through ~180°, followed by a return to the pre-excursional directions within a few thousand years. For example, the Laschamp and Iceland Basin excursions, although short-lived with durations of a few kiloyears, represent paired reversals of the geomagnetic field as VGPs reach high southerly latitudes for both excursions (Channell, 1999; Laj *et al.*, 2006). 'Excursions' displaying low or mid-latitude VGPs, rather than high latitude reverse VGPs, probably very often reflect inadequacy of the recording medium (Roberts and Winklhofer, 2004, Channell and Guyodo, 2004), inadequate rates of sediment accumulation, and/or inadequate sampling methods, rather than geomagnetic characteristics. We therefore advocate the use of the term Microchron for brief polarity chrons with established duration less than 10^5 years (**Table 1**). The term 'excursion' would then be used only for features that represent departures from normal secular variation, for which full polarity reversal has not been established. As these features become better documented, they could then be elevated to the status of Microchron, a term that denotes a brief polarity chron. Under this nomenclature, the Laschamp and Iceland Basin 'excursions' would be elevated to Polarity Microchron status as it has been established that these represent paired full polarity reversals defining a distinct polarity interval.

5.10.2 Geomagnetic Excursions in the Brunhes Chron

5.10.2.1 Introduction

In the last few decades numerous geomagnetic excursions have been discovered in the previously believed stable Brunhes Chron (**Figure 1** and **Table 2**).

The Laschamp excursion, which is now known to have an age of ~40 ka, was the first geomagnetic excursion to be recognized, in lavas from the French Massif Central (Bonhommet and Babkine, 1967). It is the most thoroughly studied excursion, and its existence is proved beyond doubt. This is not the case for other reported excursions in the late Brunhes Chron, and some of these can be attributed to sedimentological and/or sampling artifacts: for example, the Starno event (Noel and Tarling, 1975), the Gothenburg 'flip' (Mörner and Lanser, 1974), and the Lake Mungo excursion (Barbetti and McElhinny, 1972). Other early papers that documented excursions have been ratified by later work: for example, the reported excursion (Blake Event) in marine isotope stage (MIS) 5 from the Blake Outer Ridge (Smith and Foster, 1969). Wollin *et al.* (1971) documented three short intervals of reverse paleomagnetic inclination in cores from the Caribbean and the Eastern Mediterranean. An approximate age model suggested that the most recent of these was coeval with the Blake Event, and the ages of the other two were estimated to be around 180–210 and 270 ka. Kawai *et al.* (1972) and Yaskawa *et al.* (1973) reported a paleomagnetic study of a 197 m core from Lake Biwa (Japan) that documented five short episodes of reverse polarity in the Brunhes Chron. On the basis of a tentative correlation with the results of Wollin *et al.* (1971), the youngest of these episodes was correlated to the Blake Event and the other two were labeled Biwa I at about 176–186 ka and Biwa II at about 292–298 ka. Later, Kawai (1984), on the basis of fission track ages on zircons from tephra layers interbedded in the sediments, obtained ages of 100, 160, and 310 ka for the first three episodes, and ~380 ka for a fourth episode which he called Biwa III. As with much of the evidence for directional excursions published in the 1960s and 1970s, it was based on magnetization directions that were poorly defined (by modern standards) and on age control that provided plenty of latitude in their correlation to other supposed excursions. Apart from the Blake and the Laschamp excursions that have stood the test of time, we advocate abandoning the labels such as Biwa, that refer to excursions that are poorly defined with ages that are poorly constrained.

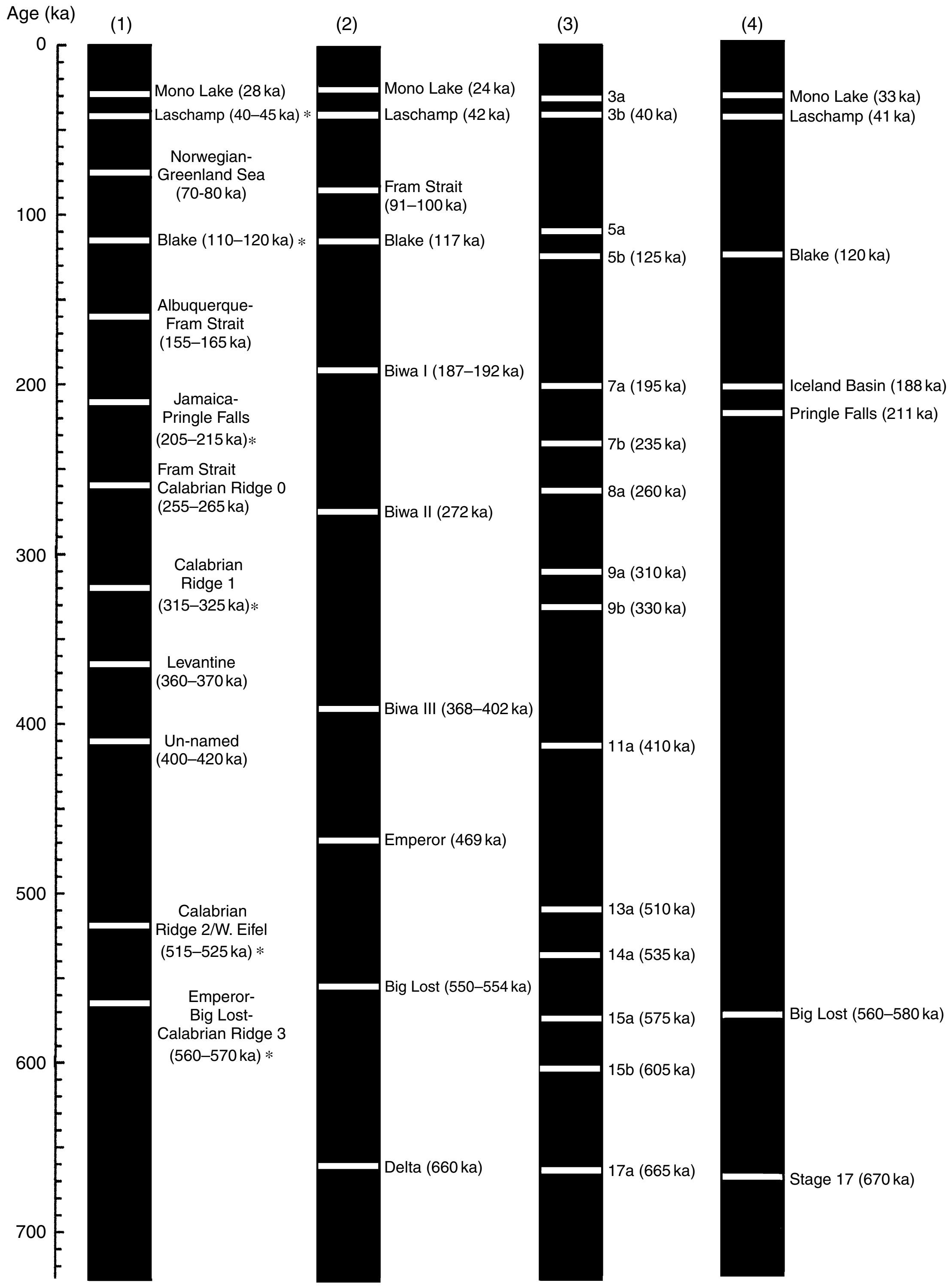

Figure 1 Geomagnetic excursions in the Brunhes Chron according to various authors. Column 1: Langereis *et al.* (1997): asterisks mark 'well-dated, global' excursions, others were deemed 'restricted or not as certain'; column 2: Worm (1997); column 3: Lund *et al*. (2001a); column 4: well-documented excursions with acceptable age control (this study).

Table 2 Excursions within the Brunhes Chron

Excursion name	MIS	Estimated age (ka)	Estimated duration (kyr)	Location or ODP Leg	Principal references
Mono Lake	3	33	1	Mono Lake/Arctic/152	(1)
Laschamp	3	41	1.5	Atlantic realm/Arctic/177/172/152	(2)
Norwegian-Greenland Sea		*60*		*Arctic*	(*3*)
Blake	5d/5e	120	5	North Atlantic/Med/172	(4)
Iceland Basin	6/7	188	3	Atlantic/Pacific/Baikal/152/162/172/177/184	(5)
Pringle Falls	7	211		Western US/New Zealand/172/152	(6)
Calabrian Ridge 0	*8*	*260*		*Med/172*	(*7*)
Calabrian Ridge I	*9*	*325*		*Med/172*	(*8*)
Un-named	*11*	*400*		*172*	(*9*)
Calabrian Ridge II	*13*	*525*		*Med/172*	(*10*)
Big Lost	14/15	560–580		Yellowstone/172/162	(11)
Stage 17	17	670		Osaka Bay/172/162	(12)

Italics indicate poorly documented excursions, MIS: marine isotope stage.
Reference: (1) Denham and Cox (1971); Liddicoat and Coe (1979); Negrini *et al*. (1984, 2000); Lund *et al*. (2001a); Nowaczyk and Antonow (1997); Nowaczyk and Knies (2000); Channell (2006). (2) Bonhommet and Babkine, (1967); Bonhommet and Zahringer (1969); Levi *et al*. (1990); Laj *et al*. (2000); Lund *et al*. (2001a); Mazaud *et al*. (2002); Laj *et al*. (2006); Channell (2006). (3) Nowaczyk and Frederichs (1994). (4) Smith and Foster (1969); Tric *et al*. (1991); Zhu *et al*. (1994); Thouveny *et al*. (2004). (5) Channell *et al*. (1997); Weeks *et al*. (1995); Roberts *et al*. (1997); Channell (1999); Oda *et al*. (2002); Stoner *et al*. (2003); Laj *et al*. (2006); Channell (2006). (6) Herrero-Bervera *et al*. (1989, 1994); McWilliams (2001); Singer *et al*. (2005); Channell (2006). (7) Kawai (1984); Langereis *et al*. (1997); Lund *et al*. (2001b). (8) Lund *et al*. (2001b). (9) Lund *et al*. (2001b). (10) Langereis *et al*. (1997), Lund *et al*. (2001b). (11) Champion *et al*. (1988); Lund *et al*. (2001b); Quidelleur *et al*. (1999); Singer *et al*. (2002); Channell *et al*. (2004). (12) Biswas *et al*. (1999); Lund *et al*. (2001b); Channell and Raymo (2003); Channell *et al*. (2004).

During a search for the Laschamp excursion in exposed sediments at Mono Lake (California), an excursion apparently younger than the Laschamp excursion was documented by Denham and Cox (1971) and subsequently by Liddicoat and Coe (1979). This excursion was named the Mono Lake excursion after its type locality. Although the excursion has apparently been observed elsewhere, the age of the excursion at the type locality remains controversial.

Creer *et al.* (1980) studied a drill-core section at Gioia Tauro in southern Italy, and suggested that an episode of reverse polarity found in the upper part of the section could be the Blake Event, at about 105–114 ka. Four additional episodes of low inclination were named α, β, γ, and δ. Kawai (1984) suggested that the youngest of the Biwa episodes could correspond to the Blake Event, the α episode to Biwa I, the β excursion to Biwa II, and the γ episode to Biwa III. Poor age control and poor definition of magnetization directions at Gioia Tauro (and Lake Biwa) make these correlations equivocal.

Champion *et al.* (1981) found evidence for a brief reversal in a sequence of basalt flows in the Snake River Plain (Idaho) to which they assigned an age of 465 ± 50 ka, using K–Ar ages of bracketing normally magnetized flows. They correlated this event with the Emperor Event of Ryan (1972). Later, however, Champion *et al.* (1988) revised the age of these reverse polarity lavas to 565 ± 10 ka. The new age implies a new reverse episode, which they named the Big Lost excursion. Based on literature available at the time, Champion *et al.* (1988) proposed the existence of eight reverse polarity microchrons in the Brunhes Chron (**Figure 1**).

Langereis *et al.* (1997) reviewed evidence for excursions in the Brunhes Chron, made the case for seven well-dated 'global' excursions and five 'restricted' less-certain Brunhes excursions, for a total of 12 excursions in the Brunhes Chron (**Figure 1**). From a central Mediterranean core, these authors added a series of excursions (Calabrian Ridge 0, 1, 2, 3) to the excursion vocabulary, the oldest (CR3) being correlative to the Big Lost excursion. Worm (1997) suggested a link between geomagnetic reversals/excursions and glaciations, and listed 10 geomagnetic excursions in the Brunhes Chron that he considered to be adequately documented (**Figure 1**). In more recent papers attempting a synthesis, Lund *et al.* (2001a, 2001b, 2006) proposed 17 excursions during the Brunhes Chron. The database for geomagnetic excursions has clearly evolved rapidly in recent years, but with little consensus on the existence or age of Brunhes-aged excursions (**Figure 1**).

In this section, we first describe evidence for the five best-documented excursions in the Brunhes

Chron: the Laschamp, Mono Lake, Blake, Iceland Basin, and Pringle Falls excursions. We then describe other less well-documented excursions recorded in the early Brunhes Chron (**Table 2**).

5.10.2.2 The Laschamp Excursion

The Laschamp excursion was the first reported geomagnetic excursion, and is certainly the best known excursion in the Brunhes Chron. Bonhommet and Babkine (1967) discovered this excursion and named it after the Puy de Laschamp, in the French Chaîne des Puys (Massif Central, France). The flow carries an anomalous characteristic paleomagnetic direction, up to 160° away from the expected dipole field direction. The excursion was initially discovered in the flow and scoria in the Puy de Laschamp, but the same anomalous direction has also been found in the nearby Olby flow (**Figure 2**).

Considerable effort has been dedicated to determination of the age of the Laschamp excursion. The first determination, using K/Ar methods on whole rock by Bonhommet and Zahringer (1969), yielded an age between 8 and 20 ka. Subsequent determinations did not confirm such a young age: Hall and York (1978) obtained whole rock K–Ar and $^{40}Ar/^{39}Ar$ dates of 47.4 ± 1.9 and 45.4 ± 2.5 ka. Condomines (1978), using the $^{230}Th/^{238}U$ radioactive disequilibrium method, obtained an age of 39 ± 6 ka, which is more than double the age of Bonhommet and Zahringer (1969). Gillot *et al.* (1979), using the unspiked (Cassignol) K–Ar method, obtained an age of 43 ± 5 ka for the Laschamp flow and 50 ± 7.5 ka for the Olby flow. An early attempt to date the Laschamp excursion using thermoluminescence (Wintle, 1973) failed because of anomalous fading of the signal. Huxtable *et al.* (1978) mitigated the problem by using sediments baked by the Olby flow and obtained an age of 25.8 ± 1.7 ka, which is somewhat younger than the 35 ± 5 ka age determined by Gillot *et al.* (1979) from quartz contained in a granitic inclusion found in the Laschamp flow. Gillot *et al.* (1979) also obtained an age of 38 ± 6 ka from five quartz pebbles found in a baked paleosol beneath the Olby flow.

In the most recent determination of the age of the excursion, Guillou *et al.* (2004) combined K–Ar and $^{40}Ar/^{39}Ar$ results from the two basaltic flows at Laschamp and Olby to better resolve the age of the Laschamp excursion. This was possible in part due to recent advances in $^{40}Ar/^{39}Ar$ incremental heating methods using a resistance furnace which can yield ages for basaltic lava flows between 100 and 20 ka with a precision better than 5% at the 95% confidence level (Heizler *et al.*, 1999; Singer *et al.*, 2000). Similarly,

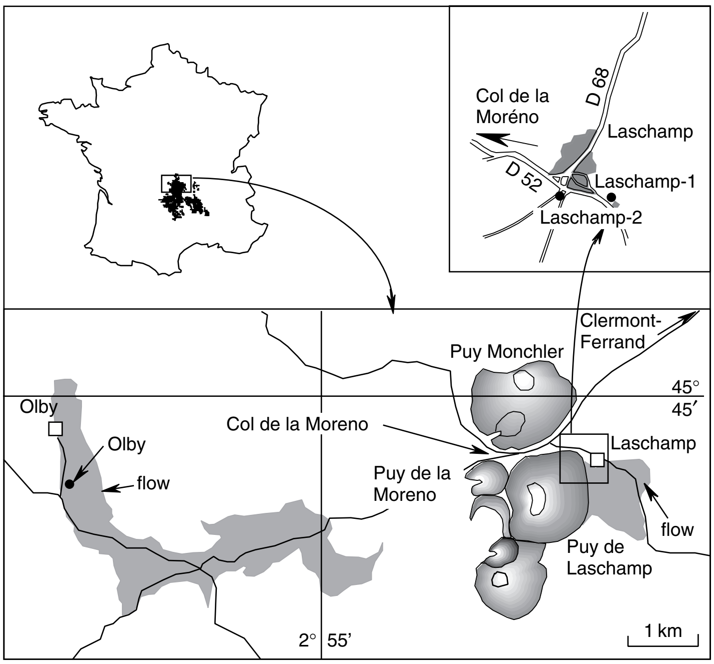

Figure 2 Location map of the sites at Laschamp and Olby in the Chaîne des Puys, Central France. From Guillou H, Singer BS, Laj C, Kissel C, Scaillet S, and Jicha B (2004) On the age of the Laschamp geomagnetic event. *Earth and Planetary Science Letters* 227: 331–343.

the unspiked K–Ar dating method, with the new calibration procedure of Charbit *et al.* (1998), allows accurate and precise measurements of minute quantities of radiogenic argon (Guillou *et al.* (1998). The results of six new unspiked K–Ar and 13 $^{40}Ar/^{39}Ar$ incremental heating experiments on subsamples from three sites in the Laschamp and Olby flows are concordant and yield a weighted mean age of 40.4 ± 1.1 ka. Consideration of the uncertainties in the $^{40}K \rightarrow {}^{40}Ar$ decay constant led to 40.4 ± 2.0 ka (2σ analytical error plus decay constant uncertainties) as the most probable age for the Laschamp excursion. The incremental progress in determining a precise age for the Laschamp excursion is depicted in **Figure 3**.

Over 10 years after the Laschamp excursion was first reported in the literature, doubts were expressed about its geomagnetic origin. Heller (1980) and Heller and Petersen (1982) noted partial or complete self-reversal of the NRM of many Olby samples, and to a lesser extent of Laschamp samples, during thermal laboratory experiments. They suggested that the almost reverse magnetization in the two flows could be attributed to self-reversal properties, rather than to the geomagnetic field. This suggestion was at odds with earlier work of Whitney *et al.* (1971) who concluded that the excursional magnetization was carried by single-domain magnetite grains, a magnetic mineralogy inconsistent with suggested mechanisms for self-reversal. Roperch *et al.* (1988) pointed out that, when more than one magnetic phase is present, the species with the lowest blocking temperature usually experiences self-reversal by magnetostatic interactions, while the species with higher (>200 °C) blocking temperature invariably carries the primary reverse direction. Furthermore, Roperch *et al.* (1988) found a reverse direction in a sample of clay which had been baked by the overlying Olby flow. Although limited to one sample (due to the difficulty of finding clays baked by the flow), this result provided evidence for a geomagnetic origin of the Laschamp excursion. Roperch *et al.* (1988) also used the Thellier–Thellier paleointensity method to obtain a value of 7.7 μT (i.e., less than one-sixth of the present field) for the reverse polarity flows at Laschamp and Olby. They argued that this low value is more characteristic of transitional geomagnetic field behavior, and, therefore, that the paleomagnetic directions of the Laschamp and Olby flows were not acquired during a stable period of reverse polarity. In the Chaîne des Puys, Barbetti and Flude (1979) obtained low paleointensity values from sediments baked by the lava flow at Royat. Here, paleointensity data imply a field strength about 30% of its present value, and an $^{40}Ar/^{39}Ar$ age of around 40 ka (Hall *et al.*, 1978), which is close to that of the Laschamp and Olby flows. In the Chaîne des Puys, Chauvin *et al.* (1989) obtained paleointensity results from different flows on the Louchadière volcano, where K–Ar ages indicated approximate synchroneity with the Laschamp flow. Chauvin *et al.* (1989) documented anomalous paleomagnetic directions (although different from the Laschamp directions) and determined a paleointensity of 12.9 ± 3.3 μT, which is about one-third the present field value, which confirms the transitional character of the Louchadière flow.

In Iceland, Kristjansson and Gudmundsson (1980) found evidence for an anomalous paleomagnetic direction from three different localities in the Reykjanes Peninsula, which they called the Skalamaelifell excursion. Levi *et al.* (1990) subsequently identified the same excursion at four additional localities in the same region, where the K–Ar age is 42.9 ± 7.8 ka, thereby associating the Skalamaelifell excursion with the Laschamp excursion. Levi *et al.* (1990) also obtained a paleointensity determination of 4.2 ± 0.2 μT, which is consistent with an earlier result by Marshall *et al.* (1988) and similar to

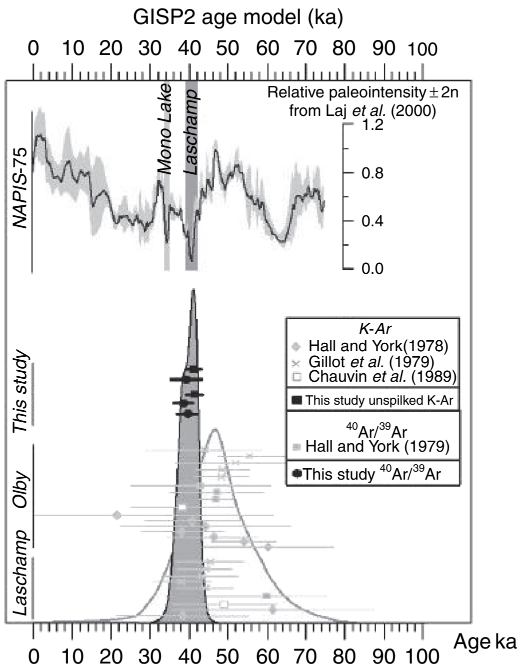

Figure 3 Radioisotopic ages of the Laschamp and Olby flows. *N* is the number of individual age determinations. The broad probability–density curve with a maximum at 46 ka represents previously published K–Ar and $^{40}Ar/^{39}Ar$ ages, whereas the narrow envelope peaking at 40.5 ka is from Guillou *et al*. (2004).

that determined by Roperch *et al.* (1988) on the Laschamp and Olby flows, and almost an order of magnitude less than the present field strength in Iceland.

In central France, the search for the Laschamp excursion in sediments from Lac du Bouchet, a maar lake only about 100 km from Laschamps (Creer *et al.*, 1990), did not yield evidence of a departure of the geomagnetic field vector from the normal direction, but the interval corresponding to the excursion was marked by low paleomagnetic field intensity. In a later paper, Thouveny and Creer (1992) discussed in more detail possible reasons for the absence of directional changes in the Lac du Bouchet record and concluded that, given the sediment accumulation rate, the duration of the Laschamp excursion could not have exceeded some 200 years, a duration that has subsequently proved to be an underestimate (see Laj *et al.*, 2000). This negative result is probably due to the inability of the Lac du Bouchet sediments to record the brief directional changes associated with the Laschamp excursion (see Roberts and Winklhofer, 2004).

Although sedimentary records often document the characteristic paleointensity minimum associated with the Laschamp excursion, the majority of records fail to document excursional directions associated with the minimum. This is likely due to low sedimentation rates (low resolution of the sedimentary records) combined with the fact that the directional anomaly associated with the Laschamp excursion may be brief (less than a few kiloyears) compared to the duration of the associated paleointensity minimum. Any delay in remanence acquisition due to bioturbation and a progressive magnetization lock-in may lead to smearing of the NRM directions, thereby inhibiting the recording of brief directional events (Roberts and Winklhofer, 2004; Channell and Guyodo, 2004).

Nowaczyk and Antonow (1997) found evidence for the Laschamp excursion in four sediment cores from the Greenland Sea. Subsequently, Nowaczyk and Knies (2000) reported directional and relative paleointensity records of the Laschamp and Mono Lake excursions from the Arctic Ocean. One of the cores documented in this study (PS 2212-3 KAL) illustrates an anomaly common to these high latitude cores. The cumulative percentage thickness of reverse polarity zones in sediment accumulated during the last few 100 ky is far greater than expected ($\sim$50% for the top 4 m of the section representing the last 120 ky in PS2212-3KAL). Fortuitous fluctuations in sedimentation rates, that are not resolvable in the age models, appear to have 'amplified' these excursional records.

In the North Atlantic Paleointensity Stack (NAPIS-75) of Laj *et al.* (2000), all but one of the six cores display abrupt directional changes at around 41 ka. In three of the cores, the component inclinations reach negative values in excess of $-30°$. The chronology of NAPIS-75 is based on correlation of the planktic $\delta^{18}O$ record from core PS2644-5 (Voelker *et al.*, 1998) to the $\delta^{18}O$ record from the GISP2 ice core (Grootes and Stuiver, 1997). Correlation among the NAPIS-75 cores was determined using isotopic stage boundaries with a second step matching cycles observed in the down-core profiles of the anhysteretic remanent magnetization (ARM). The cores were then aligned to the Greenland ice core chronology by correlation of each ARM record to that of PS2644-5 (Kissel *et al.*, 1999). The Laschamp excursion occurred at the end of interstadial 10 (Baumgartner *et al.*, 1997; Laj *et al.*, 2000; Wagner *et al.*, 2000), at about 41 ky, consistent with the most recent radiometric dating of the excursional lava flow at the Laschamp locality (Guillou *et al.*, 2004).

The Laschamp excursion has also been recognized in several piston cores from the sub-Antarctic South Atlantic Ocean (Channell *et al.*, 2000) where the age models indicate an excursional age of $\sim$40 ka and an excursion duration of <2 ky. In the cores that best display the directional excursion, mean sedimentation rates in the excursional interval exceeded 20 cm ky^{-1}. Age models were determined by $\delta^{18}O$ data from the same cores, and by utilization of accelerator mass spectrometry (AMS) ^{14}C ages from a nearby core (RC11-83) (Charles *et al.*, 1996) that could be correlated to the magnetically studied cores using both $\delta^{18}O$ and $\delta^{13}C$ data.

Mazaud *et al.* (2002) resolved the Laschamp excursion in core MD94-103 from the southern Indian Ocean (east of Kerguelen Plateau). Sedimentation rates approached 20 cm ky^{-1} in the excursional interval, the paleointensity record can be unambigously correlated to NAPIS-75, and the age and duration of the excursion can be estimated as $\sim$41 ka and approximately 2 ky, respectively.

Lund *et al.* (2005) presented two records of the Laschamp excursion from sediment cores from the Bermuda Rise and the Blake Outer Ridge in the Western North Atlantic Ocean. The high sediment accumulation rate (estimated to be 20–28 cm ky^{-1} in the interval recording the Laschamp excursion) and the uncomplicated magnetic properties resulted in well-defined characteristic remanent magnetizations. The excursional directions are present during times of apparently low paleomagnetic field intensity. The declinations display almost a full reversal (120° change in direction), while the inclinations change from 49° to −49° with intermediate values reaching −80°. The directions have been interpreted to display oscillatory behavior, growing in amplitude during the decrease in paleointensity, with an estimated time constant of 1200 years based on a constant sedimentation rate of 28 cm ky^{-1} during the excursion. The duration of the excursional interval was estimated to be ~2 ky, in agreement with previous estimates (Laj *et al.*, 2000; Channell *et al.*, 2000), but much shorter than the duration estimate from the Arctic (Nowaczyk and Knies, 2000). Lund *et al.* (2005) calculated the rates of intensity and directional changes during the excursion, based on the oxygen isotope stratigraphy. In paleointensity, these rates are typically less than 50 nT yr^{-1} and never more than 150 nT yr^{-1}. Although these values are averaged over some 100–130 years, and are strongly dependent on the age model, they are not significantly different from the mean annual rate of change of the historic field (Peddie and Zunde, 1988). Similarly, the rates of directional change are of the order of 30–70 arc min yr^{-1}, which compares with the maximum rate of 40 arcmin yr^{-1} observed historically.

Laj *et al.* (2006) reported five records of the Laschamp excursion obtained from rapidly deposited sediments at widely separated sites (the Bermuda Rise (MD95-2034), the Greenland Sea (PS2644-5), the Orca Basin of the Gulf of Mexico (MD02-2551 and MD02-2552), and the southern Indian Ocean (MD94-103). The records of the Laschamp excursion from cores JPC-14 (Blake Outer Ridge) and 89-9 (Bermuda Rise) (Lund *et al.*, 2005), as well as from cores 4-PC03 and 5-PC01 from the sub-Antarctic South Atlantic (Channell *et al.*, 2000), provide evidence for global manifestation of the Laschamp excursion (**Figure 4**).

A coherent picture of excursional field behavior during Laschamp excursion has emerged (Laj *et al.*, 2006): the excursional VGPs trace a clockwise loop (**Figure 5**), moving southward over east Asian/western Pacific longitudes, reaching high southern latitudes, followed by a northward-directed VGP path over Africa and western Europe. The turning point where the VGPs change from being southward- to northward-directed coincides with the minimum in relative paleointensity. A recently published record of the Laschamp Event at ODP Site 919 in the Irminger Basin (off east Greenland) (Channell, 2006) yields a VGP path that also describes a large clockwise loop, albeit not exactly on the same longitudes as those described above. Here, mean sedimentation rates in the vicinity of the excursion exceed 20 cm ky^{-1} and, according to the oxygen isotope age model, the age and duration of the excursion is 40 ka and <2 ky, respectively.

It is now firmly established that the strength of the geomagnetic field is the most important factor controlling cosmogenic radionuclide production (Lal and Peters, 1967; Masarik and Beer, 1999): the smaller the field intensity, the larger the production rate. As shielding of cosmic rays by the geomagnetic field occurs at distances of several Earth radii, only changes in the dipole field (i.e., global changes) are relevant to the regulation

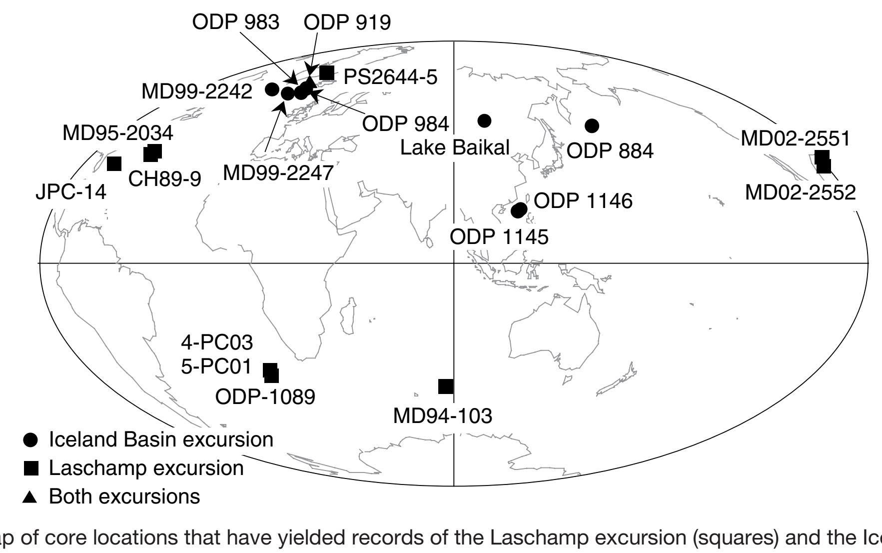

Figure 4 Map of core locations that have yielded records of the Laschamp excursion (squares) and the Iceland Basin excursion (circles).

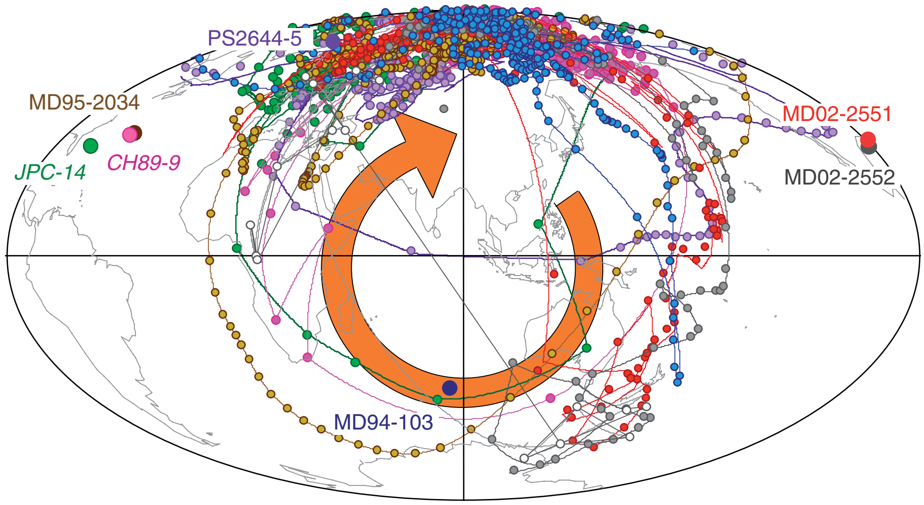

Figure 5 Transitional magnetization directions for the Laschamp excursion represented as virtual geomagnetic polar (VGP) paths. The large arrow illustrates the sense of looping which is consistently clockwise.

of cosmogenic radionuclide production. Cosmogenic nuclide flux therefore provides an independent way of establishing the global character of the intensity minimum associated with the Laschamp excursion. Initial work along these lines demonstrated that ^{10}Be flux and geomagnetic field intensity anticorrelate in a North Atlantic core (Robinson *et al.*, 1995). Subsequently, Frank *et al.* (1997) correlated a reconstruction of paleofield intensity, obtained from ^{10}Be flux records in marine sediments, to long-term trends in the SINT-200 paleointensity stack (Guyodo and Valet, 1996). The 96–25 ka record of ^{36}Cl flux from the GRIP ice core (Baumgartner *et al.*, 1998) agrees reasonably well with a production rate calculation from a paleointensity stack from the Somali basin (Meynadier *et al.*, 1992). More recently, Wagner *et al.* (2000) compared the NAPIS-75 paleointensity stack (Laj *et al.* 2000) to the field intensity record estimated from an improved, higher-resolution ^{36}Cl record, assuming that the variations in the ^{36}Cl flux are entirely due to modulation by the geomagnetic field. The ^{36}Cl-derived profile has been smoothed out using a 3000-year window in order to filter the solar modulation. The coincidence of a prominent maximum in ^{36}Cl flux with the paleointensity profile through the Laschamp excursion (**Figure 6**) leaves little doubt concerning the geomagnetic nature of the increase in cosmogenic production at around 40 ka, and provides strong evidence for the global nature of the intensity minimum associated with the Laschamp excursion.

5.10.2.3 The Mono Lake Excursion

About 35 years ago, the age of the Laschamp excursion was thought to be ~20 ka from studies in central France (Bonhommet and Zahringer, 1969). Denham and Cox (1971) sampled the Wilson Creek Formation at Mono Lake (California) in search of the same excursion. Based on radiocarbon ages available at the time, this section was thought to span the 13-30 ka time interval. Denham and Cox (1971) detected an excursion at Mono Lake, at a horizon with an estimated age of 24 ka. This excursion has since been named the Mono Lake excursion. Denham (1974) interpreted the excursion as arising from eastward drift of the nondipole field, and showed that the pattern of observed paleomagnetic directions could be reproduced assuming an eccentric radial dipole source first increasing and then decreasing during its eastward movement.

Liddicoat and Coe (1979) confirmed the apparent existence of a geomagnetic excursion at Mono Lake, documented the directional record in more detail, and extended the record further down section. There was initially widespread skepticism about the validity of this excursion. Verosub (1977) did not find evidence for it in sediments from Clear Lake (California) and Palmer *et al.* (1979) failed to find it in gravity cores from Lake Tahoe (California). Turner *et al.* (1982) obtained a distinctive pattern of 'normal' paleosecular variation, but no excursion, from an 18 m sedimentary sequence at Bessette Creek (British Columbia), that was radiocarbon

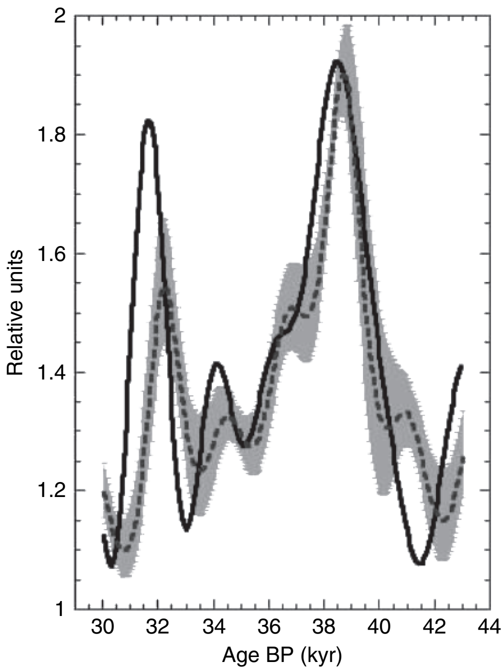

Figure 6 Comparison of the ^{36}Cl production rate (dashed line) calculated from paleomagnetic field data (modified NAPIS-75 stack) and the decay-corrected ^{36}Cl flux (solid line) from the GRIP ice core. The shaded area corresponds to the $\pm 2\sigma$ uncertainties in the ^{36}Cl production rate due to uncertainties in the relative paleointensity data. Both records are low-pass filtered and normalized to the present field values. From Wagner G, Beer J, Laj C, (2000) Chlorine-36 evidence for the Mono lake event in the Summit GRIP ice core. *Earth and Planetary Science Letters* 181: 1–6.

dated to 31.2–19.5 ka. Verosub *et al.* (1980) did not find this excursion in sediments exposed on the shores of Pyramid Lake (Nevada), which covers the time interval 25–36 ka and is only about 230 km from Mono Lake. Absence of the excursion at both Pyramid Lake and Clear Lake led Verosub *et al.* (1980) to conclude that the estimated age of the Mono Lake excursion was in error.

The interpretation of the excursion at Mono Lake has progressively evolved. Liddicoat *et al.* (1982) discovered the excursion in sediments from Lake Lahontan exposed in Carson Sink (Nevada), some 60 km east of Mono Lake. Negrini *et al.* (1984, 2000) discovered the excursion in sediments from Summer Lake, Oregon, 550 km north of Mono Lake. The excursion was apparently detected in sediments recovered by hydraulic piston core from the DSDP Site 480 in the Gulf of California (Levi and Karlin, 1989). Liddicoat (1992) resampled at Carson Sink and Pyramid Lake (Nevada), at the same locality studied by Verosub *et al.* (1980), and showed that, while the older half of the excursion is not recorded, the younger half can be recognized. Liddicoat (1996) also reported the excursion from the Lahontal Basin, Nevada.

As was the case for the Laschamp excursion, proof of the global character of the Mono Lake excursion has come from the study of marine cores. Nowaczyk and Knies (2000) obtained a record from cores in the Arctic Ocean, in which they identified directional and relative paleointensity changes that they associated to the Laschamp and to the Mono Lake excursions (dated at 35 and 27–25 ka, respectively). In the Irminger Basin (ODP Site 919), off east Greenland, two excursions with duration <2 ky are centered at 34 and 40.5 ka according to the oxygen isotope age model (Channell, 2006). These examples represent the rare occurrence of both the Mono Lake and Laschamp directional excursions in the same section, and there is now accumulating evidence for the existence of two excursions (Mono Lake and Laschamp) separated by 7–8 ky.

As for the Laschamp excursion, although the Mono Lake directional excursion is rarely recorded due to its brevity, a relative paleointensity minimum at about 32–34 ka is often recorded in relative paleointensity records. This Mono Lake paleointensity minimum is manifest as an increased cosmogenic isotope flux in ice core records. For example, in the ^{36}Cl record obtained from the GRIP ice core (Wagner *et al.*, 2000), in addition to the peak in production associated with the Laschamp excursion described above, a second distinct peak is present between Dansgaard–Oeschger (D–O) events 6 and 7, at approximately 33 ka in the GISP2 age model (**Figure 6**). This production peak can therefore be attributed to the geomagnetic field intensity minimum associated with the Mono Lake excursion, providing evidence for the global character of this paleointensity minimum.

Kent *et al.* (2002) have obtained new radiocarbon data from lacustrine carbonates (ostracodes and tufa nodules) from 11 stratigraphic horizons from the lower part of the Wilson Creek (Mono Lake) section. These authors argued that previous age estimates for the excursion recorded in the Mono Lake section, based on a series of 27 published radiocarbon measurements on tufa or ostracodes (Benson *et al.*, 1990, 1998; Lund *et al.*, 1988), do not take into account radiocarbon reservoir effects, modern carbon contamination effects, or radiocarbon production variations. As it proved virtually impossible to find clean shells, Kent *et al.* (2002) made measurements on pairs of uncoated and variably tufa-encrusted ostracodes and tufa nodules. Invariably, the more encrusted ostracode or tufa nodule samples yielded ages that were younger than the uncoated samples (by 700–2100 years). This could be attributed to modern carbon contamination, as supported by the younging ages from sample aliquots subjected to progressive acid leaching. Because (maximum) age plateaus were not always recorded during the leaching process, Kent *et al.* (2002) concluded that the radiocarbon ages should be viewed as minimum age constraints. The difference between these new determinations and those of Benson *et al.* (1990) increases with age, and reaches $\sim$4 ka in the interval of the Mono Lake excursion.

Kent *et al.* (2002) also obtained $^{40}Ar/^{39}Ar$ age estimates from sanidine crystals imbedded in the different ash layers in the Wilson Creek section. These ages are at the younger limit of the method, and are further complicated by residence time in the magma chamber, and other sources of inherited (old) argon. As a consequence, the ash layers are usually characterized by a wide range of sanidine ages. In the middle of the Mono Lake excursion, 34 individual sanidine samples from Ash #15 yielded ages between 49 and 108 ka, a range that far exceeds analytical precision. In such cases, Chen *et al.* (1996) suggested that the youngest $^{40}Ar/^{39}Ar$ ages provide realistic estimates of eruption age, as they are generally closer to accompanying radiocarbon ages. Using similar logic, Kent *et al.* (2002) considered 49 ka as the maximum depositional age for Ash #15.

Using radiocarbon dates as minima and $^{40}Ar/^{39}Ar$ dates as maxima, Kent *et al.* (2002) derived a new age model for the Wilson Creek section, and estimated the age of the lower part of the Wilson Creek section as >46 ka. On this basis, Kent *et al.* (2002) proposed that the excursion recorded at Mono Lake is, in reality, the Laschamp excursion. Indirect evidence for this interpretation is provided by the absence of a second excursion that might otherwise be identified with the Laschamp excursion in the lower part of the section.

Benson *et al.* (2003) argued against the conclusions of Kent *et al.* (2002). First, they noted that Kent *et al.* (2002) assumed a reservoir age of 1000 years, while previous work by Benson *et al.* (1990) had concluded that the carbonates deposited in the Mono Basin during the past 650 calendar years exhibit a reservoir effect ranging from 1100 to 5300 years. The radiocarbon dates obtained by Kent *et al.* (2002) may therefore have provided overestimates of the age of deposition. In addition, Benson *et al.* (2003) took advantage of the unique chemical composition of Ash #15, which allows it to be distinguished from other tephra layers outside the Mono Basin, and dated this layer in the Pyramid Lake Basin. The advantage of this strategy is that the total organic carbon (TOC) fraction in the Pyramid Lake Basin is mostly composed of algae that obtain their carbon from dissolved CO_3^{2-} and HCO_3^{-}. Therefore, while the radiocarbon ages from the Pyramid Lake Basin still incorporate reservoir effects, they are not likely to be seriously contaminated by modern carbon. Based on a previous estimate of the reservoir effect of around 600 years in the last 3000 years, Benson *et al.* (2003) concluded that the age of Ash #15 is 28 620 $\pm$ 300 years (not corrected for the reservoir age). Benson *et al.* (2003) therefore concluded that the Mono Lake excursion recorded at Wilson Creek occurred in the 31.5–33.3 ka interval, which is consistent with the estimated age of the Mono Lake excursion in the Irminger Basin (Channell, 2006), with the paleointensity minimum of this age in NAPIS-75 (Laj *et al.*, 2000), and with the peak in ^{36}Cl flux observed in the GRIP ice core record (Wagner *et al.*, 2000). The age of the excursion reported at the Mono Lake type locality remains controversial. The uncertainty in age makes this excursion difficult to distinguish from the Laschamp excursion ($\sim$41 ka).

5.10.2.4 The Blake Excursion

Smith and Foster (1969) first defined the Blake Event from a paleomagnetic study of four deep-sea cores recovered from the Blake Outer Ridge. Their study established the existence of a short interval of reverse polarity in the later part of the Brunhes Chron. The paleomagnetic declinations could not be reliably determined in these piston cores, but the event was clearly revealed by anomalous inclinations, reaching $-10°$ to $-70°$, compared to an expected inclination at this location of $+40°$. On the basis of the position of the episode within the biostratigraphy of Ericson *et al.* (1961), and the age estimate by Broecker *et al.* (1968), the boundaries of the Blake Event were placed at 108 and 114 ka. Subsequently, the oxygen isotope data of Broecker and Van Donk (1970) indicated that the Blake excursion occurred within marine isotope stage (MIS) 5.

Denham and Cox (1971) and Denham (1976) provided corroborating evidence for the Blake Event from the Greater Antilles Outer Ridge and the Blake-Bahama Outer Ridge. These studies also provided the first evidence that the Blake Event might consist of two short intervals of almost reverse polarity separated by a short interval of almost normal polarity. The evidence for the Blake Event in these early studies came from a limited geographic area and the reality of the Blake excursion as a global phenomenon remained questionable. Denham (1976) considered the possibility of a 'local' reversal. He designed a model involving a small dipole source antiparallel to the main dipole and situated near the core–mantle boundary, directly beneath the 'Blake region'. This model accounted for a reversal in the vicinity of the zone where the Blake excursion had been documented, but affecting only a small portion of the globe.

Subsequently, Creer *et al.* (1980) reported evidence for the Blake Event in clays from Gioia Tauro in southern Italy, where the double structure of the episode was also apparent. Verosub (1982) pointed out that the record of the Blake excursion is characterized by two reverse intervals separated by a short normal interval in records of the Blake excursion from Italy, Japan (Lake Biwa record), and the North Atlantic Ocean, thereby providing evidence for the global character of this complex directional characteristic.

Tucholka *et al.* (1987) reported a paleomagnetic and oxygen isotope study of five cores from the eastern and central Mediterranean that could be intercorrelated using sapropel occurrences. The quality of the results from cores from the central Mediterranean (sampled several years after coring) was not high, but the eastern Mediterranean cores yielded unambiguous results. In all three eastern Mediterranean cores, a change to negative inclinations was observed, accompanied by reverse declinations in two cores and a smaller declination change in the third. The excursion was recorded in sediment directly overlying sapropel S5. In two of the cores, the record is probably incomplete, as no underlying normal polarity directions were found above the sapropel layer. One record, on the other hand, shows the bipartite structure that is characteristic of the Blake Event in the North Atlantic, Italy, and Japan. Detailed oxygen isotope analyses, made on four of the cores, established the stratigraphic position of the Blake Event between MIS 5e and MIS 5d. Estimates of the duration of the event, based on constant sediment accumulation rate between tie points, varied in the 2.8–8.6 ky range, with a mean of 5.3 ± 2.7 ky.

Tric *et al.* (1991) reported a detailed record of the Blake Event from two Mediterranean cores, one from the Tyrrhenian Sea and the other being one of the cores already studied by Tucholka *et al.* (1987) that was resampled at much higher resolution. In the Tyrrhenian Sea core, only five excursional (intermediate) directions were recorded possibly due to a lower sediment accumulation rate synchronous with the Blake excursion. On the other hand, the eastern Mediterranean core yielded much more detailed results, with 70 intermediate and reverse polarity directions, making this record of the Blake Event the most detailed obtained so far, and the only one for which intermediate VGPs were obtained (**Figure 7**). Although no record of field intensity was obtained, the directional variations provide insight into the complex dynamics of the Blake Event. First, a sudden jump of inclination from positive to negative values was observed, accompanied by intermediate declinations and ending with a recovery of normal polarity. The VGPs in this first phase, which lasted about 1100 years according to the age model, lie over North and South America followed by an abrupt jump to Australia. The excursion in this core is characterized by two intervals of reverse polarity directions with an intervening normal polarity period. The total duration of the excursion, based on the isotope stratigraphy, appears to be close to 5 ky.

Thouveny *et al.* (2004) observed two intervals of negative magnetization inclination in the 115–122 ka

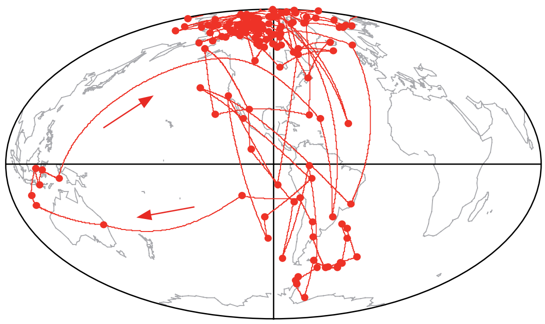

Figure 7 Transitional magnetization directions represented as virtual geomagnetic polar (VGP) paths for the Blake excursion recorded in Core MD84-627 from the Eastern Mediterranean. Data from Tric E, Laj C, Valet J-P, Tucholka P, Paterne M, and Guichard F (1991) The Blake geomagnetic event: Transition geometry, dynamical characteristics and geomagnetic significance. *Earth and Planetary Science* Letters 102: 1–13.

interval of core MD95-2042 from the Portuguese Margin. Oxygen isotope stratigraphy in this core places this record of the Blake excursion in MIS 5c according to table 2 of Thouveny *et al.* (2004), although it appears to span the MIS 5d/5e boundary in their figure 12. The latter designation would be consistent with the findings of Tucholka *et al.* (1987).

Zhu *et al.* (1994) obtained a record of the Blake excursion from a loess section at Xining (western China). The sequence of directional changes is contained in paleosol S1, which corresponds to MIS 5 (An *et al.*, 1991; Li *et al.*, 1992). Zhu *et al.* (1994) estimated the top and base of the studied section to be at 72 and 130 ka, and estimated ages of 117 and 111 ka for the onset and demise of the Blake excursion, respectively, yielding a duration estimate of 5.6 ky. The directional record of the excursion is characterized by stop-and-go behavior, reminiscent of the (Miocene) volcanic record at Steen's Mountain (Mankinen *et al.*, 1985). The Xining record contains three clearly defined periods of reverse polarity separated by two short intervals of normal polarity. The structure of this excursional record is more complex than has been previously reported for this excursion. The documented nonuniform timing of acquisition of magnetization for Chinese loess/paleosol sequences (e.g., Heslop *et al.*, 2000; Spassov *et al.*, 2003) may contribute to this apparent complexity, and to the fact that searches for the Blake Event in Chinese loess have not always been successful (Parès *et al.*, 2004).

Although the Blake excursion has also been documented in the Junzhoutai loess of the excursion section at Lanzhou (Fang *et al.* 1997), the structure is not the same as at Xining. At Lanzhou, it comprises two short reverse polarity intervals separated by a short normal polarity interval, which is similar to some records of the excursion in marine sediments. Based on thermoluminescence and astronomically tuned cycles of magnetic susceptibility, the age of the Blake Event was bracketed between Paleosol S1-c (equivalent to MIS 5e) and Loess 2-2 (MIS 5d) (Fang *et al.*, 1997). This corresponds to an age range of 120–115 ka, and a total duration of 5.5 ky, which is consistent with the estimate of Tucholka *et al.* (1987) from the Mediterranean Sea.

Observations from the highest resolution marine cores and from loess sections indicate that the Blake excursion is a global geomagnetic feature, with a characteristic structure comprising two short periods of almost reverse polarity separated by a short period of almost normal polarity.

5.10.2.5 The Iceland Basin Excursion

Over the last 10–15 years, many studies have provided evidence for geomagnetic excursions in the 180–220 ka interval. Labels Jamaica, Pringle Falls, and Biwa I have been used somewhat arbitrarily by different authors in referring to excursions with estimated ages in the 180–220 ka interval. As the fidelity and age control of available records has improved, evidence has accumulated for two excursions in the 180–220 ka interval. The younger one is labeled the Iceland Basin excursion (~185–190 ka), and the older one takes its name from Pringle Falls (~211 ka).

As mentioned above, paleomagnetic analyses of cores from Lake Biwa provided early evidence of several excursions during the late Brunhes Chron (Kawai *et al.*, 1972; Yaskawa *et al.*, 1973). The authors suggested that the youngest could be the Blake event, and two other episodes of reverse inclination were tentatively correlated to intervals of anomalous inclination in the record of Wollin *et al.* (1971) from the North Pacific Ocean. From this correlation, an age between 176 and 186 ka was assigned to an event that they called the Biwa I event. These early records have remained problematic due to inadequate paleomagnetic analysis by modern standards, and poor age control.

In the marine realm, records corresponding to this time interval obtained from Norwegian–Greenland Sea revealed several intervals of negative inclination (Bleil and Gard, 1989). These intervals have also been observed in a series of piston cores from further North in Fram Strait (Nowaczyk and Baumann, 1992) and on the Yarmak Plateau (Nowaczyk *et al.*, 1994). These early studies, however, lacked a well-defined $\delta^{18}O$ record, leading to uncertainties in assessing the precise age of the intervals of negative inclination.

Weeks *et al.* (1995) reported a paleomagnetic study of four piston cores from the North Atlantic Ocean. The age model was based on oxygen isotope stratigraphy from planktic foraminifera. The paleomagnetic record indicates that a large swing in inclination to negative values and a marked decrease in relative paleointensity occurred around 180–200 ka. Lehman *et al.* (1996) reported a paleomagnetic study of three marine cores in the Açores area in the North Atlantic Ocean. For one of the cores, a detailed $\delta^{18}O$ record was obtained, and the two others were correlated to it using sediment grayscale reflectance data. The paleomagnetic record contains fluctuations in declination and inclination and a marked drop in paleointensity at ~190 ka, but no clear shift in the inclination. Nowaczyk and Antonow (1997) produced $\delta^{18}O$ and magnetostratigraphic records from the Greenland Basin that indicate the presence of negative inclinations with ages corresponding to the Mono Lake (27–28 ka) and Laschamp (~40 ka) excursions, with an additional excursion at around 188 ka. Nowaczyk and Antonow (1997) refer to this latter excursion as the 'Biwa I/Jamaica' excursion, thereby associating it with the excursions documented in early papers by Wollin *et al.* (1971), Ryan (1972), and Kawai *et al.* (1972). Roberts *et al.* (1997) documented an excursion in the North Pacific Ocean at ODP Site 884 (**Figure 4**). Preservation of foraminifera was poor at this site, so the age model was obtained by transferring the oxygen isotope stratigraphy from ODP Site 883 (Keigwin, 1995), by correlation of magnetic susceptibility records. However, the $\delta^{18}O$ record was not well defined at Site 883 in the vicinity of the excursion (MIS 6/7 boundary), so the age of the excursion is relatively poorly constrained.

Channell *et al.* (1997) reported $\delta^{18}O$ and paleomagnetic (directions and paleointensity) data from rapidly deposited ($\geq$10 cm ky^{-1}) sediments recovered at ODP Site 983 on the Gardar Drift in the Iceland Basin (**Figure 4**). Paleomagnetic analyses revealed, in the 186–189 ka interval, a short-lived excursion in which the VGPs move to high southern latitudes. The age of this excursion is constrained close to the MIS 6/7 boundary (~188 ka) and appears to be distinct from the Pringle Falls excursion (see below). A paleointensity minimum with an onset age of 218 ka at ODP Site 983 may be coeval with the Pringle Falls excursion, although no directional excursion is recognized in association with this paleointensity minimum at this site. Channell *et al.* (1997) named this excursion the Iceland Basin Event (now Iceland Basin excursion) and this labeling is adopted here when referring to the geomagnetic excursion at the MIS 6/7 boundary. The Iceland Basin excursion appears to be coeval with the geomagnetic excursions recognized in the central North Atlantic Ocean by Weeks *et al.* (1995), as well as in the western equatorial Pacific Ocean (Yamazaki and Yoka, 1994).

The record of the Iceland Basin excursion from ODP Sites 984 (Channell, 1999) and 980 (Channell and Raymo, 2003), where mean sedimentation rates in the Brunhes Chron exceeded 11 cm ky^{-1}, is similar to that obtained at Site 983. At all three sites, inclinations are negative in the 180–195 ka interval, and the estimated duration of the excursion is ~3 ky. The apparent discrepancy in age at Sites 980, 983, and 984 can be attributed to uncertainties in the isotopic age model, particularly at Site 984 where the age model is not well defined. The VGP paths for the Iceland Basin excursion from Sites 983 and 984 are similar (**Figure 8**) and feature a large-scale counterclockwise loop that is located over Europe and Africa in the first N → S part of the loop, and returns to northern latitudes over the western Pacific Ocean. The directional changes coincide with a prominent paleointensity minimum that is a characteristic feature of sedimentary relative paleointensity records.

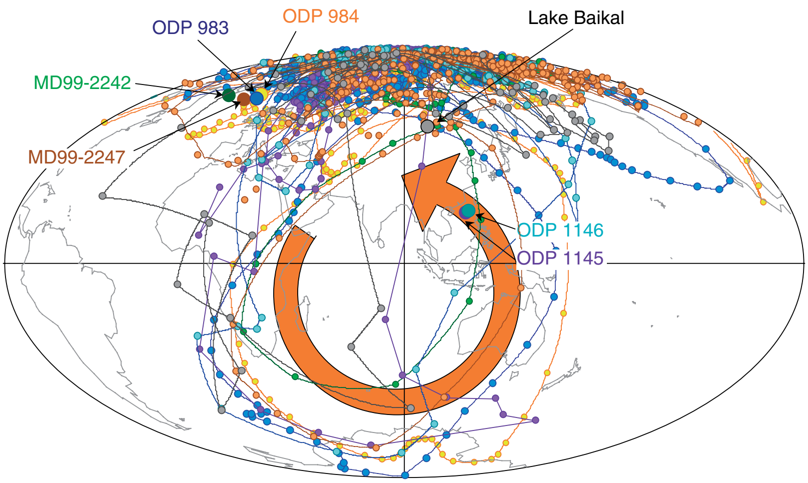

Figure 8 Transitional magnetization directions represented as virtual geomagnetic polar (VGP) paths for the Iceland Basin excursion. The large arrow illustrates the sense of looping of the VGPs which is consistently counterclockwise.

Oda *et al.* (2002) obtained a paleomagnetic record, including a paleointensity record, from core Ver 98-1 from Academician Ridge, Lake Baikal. The age model was derived using sediment bulk density as a proxy for biogenic silica content. Interglacial intervals were then correlated to maxima in silica (density). An age-depth plot was then produced based on SPECMAP ages (Martinson *et al.*, 1987) for isotope substages back to MIS 7.3. The average sedimentation rate was estimated to be around 4.5 cm ky^{-1}. A directional excursion in the paleomagnetic record between 6.70 and 6.96 m below lake floor was estimated to have an age of 177–183 ka. The excursion coincides with a marked minimum in the relative paleointensity proxy that can be correlated with the record from ODP Site 983. Oda *et al.* (2002) also redated an excursion from another core from Academician Ridge previously inferred to record the Blake event (Sakai *et al.*, 1997). The new age is 223 ka, which provides evidence for an excursion (possibly Pringle Falls) preceeding the Iceland Basin excursion in sediments from Lake Baikal.

The general pattern of the VGP paths for the Lake Baikal record is similar, albeit much more scattered than those observed at ODP Sites 983 and 984. After an initial swing over the western Atlantic Ocean (which does not appear in the ODP records), the VGPs move southward over Africa reaching high southern latitudes. The S $\rightarrow$ N return path is defined by only three points, all of which are in the Southern Hemisphere, which consequently makes the return path ill-defined. The overall counterclockwise loop is, however, similar to that observed from the two North Atlantic ODP cores.

Stoner *et al.* (2003) recorded the Iceland Basin excursion at ODP Site 1089 in the sub-Antarctic South Altantic. Mean sedimentation rates are in the 15–20 cm ky^{-1} range for a record that extends back to ~580 ka. The relative paleointensity record from this site can be correlated to those from the North Atlantic and Lake Baikal that record the same excursion; the excursion occurs within the same prominent paleointensity low. At this site, the oxygen isotope age model yielded an age of 189–191 ka and a duration of ~2 ky for this excursion.

Laj *et al.* (2006) reported four records (directions and relative paleointensities) of the Iceland Basin excursion. Two of the records are from the North Atlantic, core MD99-2242 was taken on the Eirik Drift (south of Greenland), and core MD99-2247 was from the western flank of the Reykyanes Ridge (**Figure 4**). The two other records were obtained from the South China Sea, at ODP Sites 1145 and 1146 (ODP Leg 184). Average sedimentation rates ranged from 7 to 10 cm ky^{-1} for cores in the Atlantic Ocean, and ~15 cm ky^{-1} for cores from the South China Sea. The excursion can be correlated to the MIS 6/7 boundary. The VGP paths are uniform for each of the dispersed site locations, and are consistent with the paths obtained from ODP Sites 983 and 984 (**Figure 8**). During the first part of the excursion, the VGPs move southward over Africa along a narrow band of longitudes before crossing the equator and reaching high southern latitudes (fully reverse polarity directions). The VGP path for core MD99-2242 is less well resolved, reaching mid-latitudes in the Indian Ocean. For all four cores, the VGP return path to the Northern Hemisphere lies within a longitudinal band over east Asia.

Channell (2006) recorded the Iceland Basin excursion at ODP Site 919 in the Irminger Basin (off east Greenland). According to the oxygen isotope age model, the excursion occurs in the 180–188 ka interval. Its manifestation at this site is considerably more complex than other records of the same excursion. Although the sediments have all the attributes for high fidelity recording, and the excursion is recorded by both discrete samples and u-channel samples, the excursion is characterized by a first VGP loop to high southern latitudes characterized by an outward and return path over Africa, followed by a second VGP loop with an outward path over east Asia and a return path over Africa. These two VGP loops are followed by a complex group of VGPs at low latitudes over western Africa. In view of the uniformity in magnetic hysteresis properties through the excursion, the complexity of the VGP path cannot be readily attributed to lithological variability. There are two feasible explanations for the discrepancy between this record and those compiled by Laj *et al.* (2006): either unrecognized sediment deformation has affected parts of this record or excursional records are more complex when fully recorded at some (high latitude) locations.

5.10.2.6 The Pringle Falls Excursion

A detailed record of a geomagnetic excursion obtained from a sedimentary lacustrine sequence near Pringle Falls (Oregon) was initially thought to represent the Blake Event (Herrero-Bervera *et al.*, 1989). Subsequently, however, the age of this episode was revised on the basis of $^{40}Ar/^{39}Ar$ dating of plagioclase feldspars from an ash layer (Ash D) located close to the base of the excursion (Herrero-Bervera *et al.*, 1994). The isochron age and the plateau age obtained from step heating were not significantly different, but the isochron age had a larger relative uncertainty, resulting from low radiogenic yield. The authors, therefore, took the plateau age of 218 ± 10 ka to be the best estimate of the age of the Ash D layer, which corresponds to the onset of the excursion. Herrero-Bervera *et al.* (1994) associated this excursion with the Jamaica excursion of Wollin *et al.* (1971) and Ryan (1972). By modern standards, the Jamaica excursion was not adequately resolved either paleomagnetically or stratigraphically in these early publications. Following the practice of naming geomagnetic excursions after the location where they have been unambiguosly recorded, we advocate use of Pringle Falls as the name for this excursion.

Herrero-Bervera *et al.* (1994) used the chemical and petrographic characteristics, stratigraphic position, and available age data to correlate the tephra from Pringle Falls to tephra layers present in other sequences in western North America, including Summer Lake, Mono Lake, and Long Valley. Coeval paleomagnetic records of excursions from each of these localities had been previously published (Negrini *et al.*, 1998, 1994; Liddicoat and Bailey, 1989; Liddicoat, 1990; Herrero-Bervera and Helsley, 1993), but only after this correlation of tephras were they recognized as records of the same excursion. The paleomagnetic records from two sites at Pringle Falls (the second one being at a distance of 1.5 km from the original one) and from Long Valley, some 700 km away, are strikingly similar (**Figure 9**). The VGP paths lie over the Americas in the first N → S part of the excursion and then move to the northwest Pacific for the S → N return path (Herrero-Bervera *et al.*, 1994).

Subsequently, a series of transitional paleomagnetic directions were obtained from the Mamaku ignimbrite in the North Island of New Zealand (Tanaka *et al.*, 1996). Three new $^{40}Ar/^{39}Ar$ ages from plagioclase from the Mamaku ignimbrite (Houghton *et al.*, 1995) yielded a weighted mean of 223 ± 3 ka, which is statistically indistinguishable from the age of Ash D at Pringle Falls according to Herrero-Bervera *et al.* (1994). McWilliams (2001) plotted 29 VGPs for the Mamaku ingimbrite together with those from the two sites at Pringle Falls and at Long Valley. The agreement in time and space is remarkable (**Figure 9**). The VGPs from volcanic rocks in New Zealand record only a fraction of the total excursion and are not nearly as complete as the records from Pringle Falls or Long Valley, as expected due to the episodic nature of volcanic eruption. Nonetheless, this remarkable agreement suggests that the excursion was manifest globally, and that the transitional field was dominated by a dipolar field component (McWilliams, 2001).

Singer *et al.* (2005) recently reported 16 laser incremental heating ages from plagioclase crystals derived

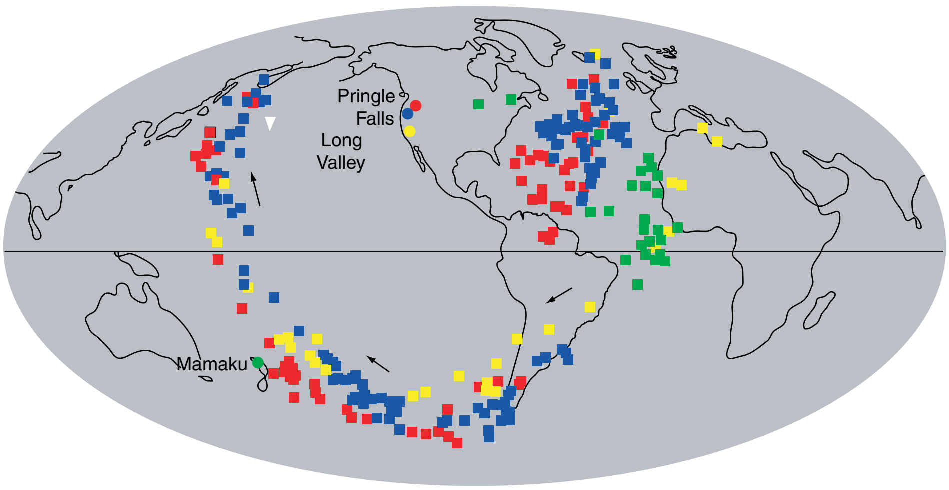

Figure 9 Transitional magnetization directions represented as virtual geomagnetic polar (VGP) paths for the Pingle Falls excursion. Squares are individual VGPs for the event recorded at Pringle Falls (two records), Long Valley, and New Zealand. Circles are sampling sites, color-coded to match respective points on the VGP paths. Arrows indicate the progression of the VGP paths from older to younger. From McWilliams M (2001). Global correlation of the 223 ka Pringle falls event. *International Geological Reviews* 43: 191–195.

from Ash D (deposited during the onset of the excursion) at Pringle Falls, that define an $^{40}Ar/^{39}Ar$ isochron of 211 ka ±13 ka. At ODP Site 919, off east Greenland, Channell (2006) recorded an excursion that predates the Iceland Basin excursion and, according to the stable isotope age model, correlates to the 206–224 ka age interval. The VGP path has a complex series of loops to equatorial latitudes over the Pacific and South America, followed by a clockwise loop to high southern latitudes. This last loop is in the same sense (clockwise) and follows a similar path as those recorded at Pringle Falls, Long Valley, and New Zealand for supposedly the same excursion. On the other hand, there is no trace in the Pringle Falls and Long Valley records of the initial complex structure seen at ODP Site 919, despite the high sediment accumulation rates at these two localities (the excursion is recorded over stratigraphic intervals of 100 cm at Long Valley and over 700 cm at Pringle Falls).

Well-grouped excursional site mean magnetization directions (mean declination = 101°, mean inclination = −36°) were obtained from 65 sites collected from the six major flows that represent the Albuquerque Volcanoes Field (Geissman *et al.*, 1990). Age determinations using the $^{238}U-^{230}Th$ method yielded an isochron age of 156 ka ± 29 ka (Peate *et al.*, 1996). More recent $^{40}Ar/^{39}Ar$ analyses in the Albuquerque volcanics yield an isochron age of 211 ± 22 ka (Singer *et al.*, 2005) implying that the excursional magnetization directions in the Albuquerque Volcanoes Field are coeval with the Pringle Falls excursion in Oregon.

5.10.2.7 Excursions in the Early Brunhes Chron

Although fewer studies have been conducted for the Brunhes Chron prior to 250 ka, there is evidence for geomagnetic excursions in the early part of the Brunhes Chron. The use of different names, combined with imprecise age control, has led to confusion in labeling and correlation.

Ryan (1972) defined the so-called Emperor Event as a short reverse polarity zone at the base of Caribbean core V12-22. Based on the work of Ericsson *et al.* (1961) and Broecker and Van Donk (1969), the age of this interval was estimated to be about 460-480 ka. The excursion was, however, defined by only one sample that was only demagnetized in peak alternating fields of 5–15 mT. Nevertheless, some support for the excursion was found in the study of axial MMA records from the Galapagos spreading center that suggested a brief reverse polarity event at 490 ± 50 ka (Wilson and Hey, 1981). This age depends on the assumption of linear spreading rate of the Galapagos Ridge and is therefore not robust. Interestingly, the apparent occurrence of the Emperor Event on the Galapagos Ridge MMA data provides the basis for the one/only 'cryptochron' within the Brunhes Chron in the timescale of Cande and Kent (1992a).

As mentioned above, initial dating of a reverse polarity flow in Idaho at 490 ± 50 ka (Champion *et al.*, 1981) provided land-based support for the existence of the Emperor Event. Later, however, Champion *et al.* (1988) revised the age of these reverse polarity lavas to 565 ± 10 ka which therefore corresponds to a different (older) reverse episode, which they named the Big Lost excursion. In their seminal paper, Champion *et al.* (1988) reviewed evidence for excursions from both volcanic and sedimentary sequences and concluded that at least eight excursions existed during the Brunhes Chron. In addition to the newly discovered Big Lost excursion and the well-known excursions discussed above, they presented evidence for: (1) the δ excursion, first documented by Creer *et al.* (1980) at Gioia Tauro in Southern Italy, dated at 640 ka; (2) the Emperor excursion (~470 ka) (Ryan, 1972); (3) the Biwa III excursion (~390 ka); and the (4) Levantine excursion (or Biwa II) (~290 ka) (Ryan, 1972; Yaskawa *et al.*, 1973; Kawai *et al.*, 1972; Kawai, 1984).

Langereis *et al.* (1997) documented four short excursions in piston core KC-01B collected from the Calabrian Ridge in the Ionian Sea (Mediterranean Sea). Astronomical calibration of the age sapropels, recognized on the basis of rock-magnetic and geochemical properties, allowed development of an age model, according to which the ages of the four excursions are 255–265 ka, 318 ± 3 ka, 515 ± 3 ka, and 560–570 ka. Langereis *et al.* (1997) labeled these excursions Calabrian Ridge 0 (CR0), CR1, CR2, and CR3. The authors proposed that CR0 could correspond to the Fram Strait excursion (Nowaczyk *et al.*, 1994), while the oldest one could be the Big Lost excursion (which they associated with the Emperor excursion). The complex redox conditions in the multicolored sediments of core KC-01B (see Langereis *et al.*, 1997), and the fact that the CR excursions are defined by single samples, indicates that they require independent corroboration prior to their incorporation into the library of geomagnetic excursions.

Biswas *et al.* (1999) reported a magnetostratigraphy of a 1700 m core from the Osaka Bay, southwestern Japan, spanning the last 3.2 My. Although the mean sedimentation rate in the Brunhes Chron (50 cm ky^{-1}) is far higher in these coastal sediments than in deep marine sequences, only one excursion is recorded in the Brunhes Chron. This Brunhes Chron excursion is recorded in lacustrine silty clays that are immediately overlain by marine clays, and is characterized by intensity fluctuations with steep inclinations (up to $+72°$ and $-82°$). It occurred during marine isotopic stage (MIS) 17, probably during substages 17.4–17.3, based on a sea-level interpretation of the marine/terrestrial sequence. This would correspond to an excursion age of 690 ka. The same age is obtained from the assumption of uniform sedimentation rate between the Matuyama/Brunhes (M/B) boundary and a characteristic tuff layer (Aira Tuff dated at 24.5 ka). Biswas *et al.* (1999) named this excursion the 'Stage 17 Event'. They correlated it to the δ event reported by Creer *et al.* (1980) from Gioia Tauro. The age of the δ event can be revised to 680 ka, using the modern value of 780 ka for the M/B reversal. Interestingly, the Stage 17 excursion is manifest in Osaka Bay as a double excursion with steep negative inclinations and an intervening interval with positive inclinations. An excursion within MIS 17 has also been recognized at ODP Site 980 collected from the Feni drift, North Atlantic (Channell and Raymo, 2003). The oxygen isotope age model yields an excursional age of 687–696 ka within MIS 17, and an excursional duration of 9 ky. Paleomagnetic inclinations reach low negative values accompanied by a $\sim 180°$ swing in declination.

Lund *et al.* (2001a, 2001b) documented many apparent excursions in the Brunhes Chron in marine cores taken from sediment drifts in the Western North Atlantic Ocean (Blake Outer Ridge, Bahama Outer Ridge and Bermuda Rise) during ODP Leg 172. From initial shipboard measurements, made on half-cores with a low-resolution pass-through cryogenic magnetometer, 14 geomagnetic episodes were identified as 'plausible' Brunhes Chron magnetic field excursions (Lund *et al.*, 1998). Laboratory measurements conducted on u-channels and discrete samples documented narrow intervals of relatively low or high inclination or westerly and easterly declinations, which could be traced among the records from independent holes separated by distances <1 km and were correlated using variations in magnetic susceptibility. Shore-based measurements led Lund *et al.* (2001a, 2001b) to confirm 12 of the 14 originally defined excursions, one of which appears to be synchronous with the Stage 17 excursion. The large number of geomagnetic excursions in the Brunhes Chron at ODP Sites 1060–1063 (ODP Leg 172) indicates that they are not rare, episodic disturbances of an otherwise stable geomagnetic field, but an integral component of the field. In addition, Lund *et al.* (2001b) proposed that most of the excursions tend to occur in bundles of two or three close together separated by intervals of 'regular' secular variation. This observation, in addition to highlighting an important characteristic of the geomagnetic field, may also explain why there has been considerable difficulty in identifying and distinguishing among near-coeval excursions (i.e., Iceland Basin – Pringle Falls) prior to the availability of high-resolution stratigraphy.

Lund *et al.* (2001b) stated, "for almost any previously identified excursion anywhere in the world, we can find an excursion record in Sites 1060–1063 that is not significantly different in age". Here lies the problem: the ODP Leg 172 sites lack continuous oxygen isotope data that would aid precise age model construction. The most recent age models for ODP Leg 172 sediments were determined by tuning filtered records of carbonate percentage, derived from gray-scale reflectance records calibrated with shipboard and postcruise carbonate measurements, to the astronomical solution for precession and obliquity (Grützner *et al.*, 2002). The age models are more robust for the shallow water sites (Sites 1055–1059) than for the deeper water sites (Sites 1060–1063). Precession-related cycles are weak, particularly in the MIS 6-7 interval and prior to 0.5 Ma for the deeper water sites (Grützner *et al.*, 2002). The tuning was performed on one 'reference' site from each group (Sites 1058 and 1062), and the other sites in the group were correlated to these reference sites using the filtered and unfiltered carbonate records. The correlation and labeling of ODP Leg 172 excursions would be substantially aided by age models based on oxygen isotope data.

Some claims for new excursions in the early Brunhes Chron remain controversial, particularly those from lava flows where precise radiometric dating is often lacking and critical to adequate correlation among excursions. For instance, Quidelleur and Valet (1996) conducted a paleomagnetic study in the Barranco de los Tilos, La Palma (Canary Islands) at a location previously studied by Abdel-Monem *et al.* (1972). Initially, they proposed that transitional magnetization directions from the southern side of the Barranco represented post-transitional rebound associated with the M/B boundary reversal. Later, on the basis of three transitionally magnetized lavas that gave a mean unspiked K/Ar age of 602 ± 24 ka, Quidelleur *et al.* (1999) recognized a new, significantly younger episode and proposed the name La Palma excursion. In support of this claim, Quidelleur *et al.* (1999) noted that the proposed excursion coincides with a marked minimum in the SINT-800 paleointensity stack of Guyodo and Valet (1999). Subsequently, Singer *et al.* (2002) obtained $^{40}Ar/^{39}Ar$ ages of 580.7 ± 7.8 ka for the same transitional flows studied by Quidelleur *et al.* (1999). This age is indistinguishable at the 95% confidence level from the $^{40}Ar/^{39}Ar$ age (558 ± 20 ka) obtained by Lanphere *et al.* (2000) for the Big Lost excursion in the Snake River Plain.

Table 2 provides a summary of excursions within the Brunhes Chron. Those entries in normal type are considered (by us) as adequately defined, whereas the entries in italics await further ratification either in definition of the magnetic data or in refinement of the age models.

5.10.3 Geomagnetic Excursions in the Matuyama Chron

5.10.3.1 Background

The age and structure of the predominantly reverse polarity Matuyama Chron has been progressively refined since the coupled K/Ar and paleomagnetic studies on basaltic lavas that began with the work of Cox *et al.* (1963) and McDougall and Tarling (1963a, 1963b, 1964). The Olduvai Subchron takes its name from normal polarity lavas dated at 1.72 Ma from the Olduvai Gorge, Tanzania (Grommé and Hay, 1963). The Jaramillo Subchron was first recognized by Doell and Dalrymple (1966) and takes its name from Jaramillo Creek (New Mexico). The Réunion Subchron originates from the work of Chamalaun and McDougall (1966) who found both normal and reverse magnetizations in basaltic rocks yielding K/Ar ages close to 2.0 Ma from the island of La Réunion. These normal polarity directions were, at that time, considered to be coeval with those from Olduvai Gorge as documented by Grommé and Hay (1963). McDougall and Watkins (1973) sampled two basaltic sections on La Réunion and documented a normal polarity zone dated by K/Ar methods to the 1.95–2.04 Ma interval, which corresponds to ~2.07 Ma using more modern decay constants (Steiger and Jager, 1977). By the early 1970s, it was realized that the Réunion Event is significantly older than the Olduvai Subchron. Grommé and Hay (1971) considered that a bimodal distribution of K/Ar ages for normally magnetized lavas with ages of ~2.00–2.14 Ma from a variety of locations indicated the existence of two Réunion Events, although there was, and continues to be, no evidence for two events within any single stratigraphic section (see review in Channell *et al.*, 2003a). In their compilation of the GPTS for the last 5 My, Mankinen and Dalrymple (1979) adopted the two Réunion Events proposed by Grommé and Hay (1971) and estimated their ages as 2.01–2.04 and 2.12–2.14 Ma, respectively.

Mankinen *et al.* (1978) detected normal polarities in 1.1 Ma volcanics from Cobb Mountain (Coso Range, California); however, Mankinen and Dalrymple (1979) were not sufficiently confident in a normal polarity subchron of this age to include it in their timescale. Evidence for this normal polarity subchron was strengthened by further study of volcanics in the Coso Range by Mankinen and Grommé (1982). Subsequent observation in high sedimentation rate sediment cores from the Carribbean (Kent and Spariosu, 1983) and North Atlantic (Clement and Kent, 1987; Clement and Martinson, 1992) cemented the Cobb Mountain Subchron as a feature of the Matuyama Chron.

The GPTS of Mankinen and Dalrymple (1979), based on coupled K/Ar and paleomagnetic studies of basaltic lavas (**Figure 10** and **Table 3**), remained the reference for the 0–5 Ma GPTS for over 10 years. Beginning in the 1980s, astrochronologies from sedimentary sequences demonstrated that the K/Ar ages for polarity reversals in volcanic rocks compiled by Mankinen and Dalrymple (1979) were young by an average of about 7% (due to argon loss). The first astrochonological evidence that the generally accepted K–Ar age for the M/B boundary (0.73 Ma) was too young can be attributed to Johnson (1982) who gave an age of 0.79 Ma for the M/B reversal based on matching the orbital insolation curve to oxygen isotope records from two cores (V28-238/9) that record the M/B reversal. It was, however, the study of ODP Site 677 in the equatorial Pacific Ocean (Shackleton *et al.*, 1990) that opened the door to astrochronological revision of the polarity timescale. Benthic oxygen isotope data at ODP Site 677 are dominated by orbital obliquity, and the planktic record is controlled by orbital precession. Because the precession signal is modulated by eccentricity, the planktic oxygen isotope data at

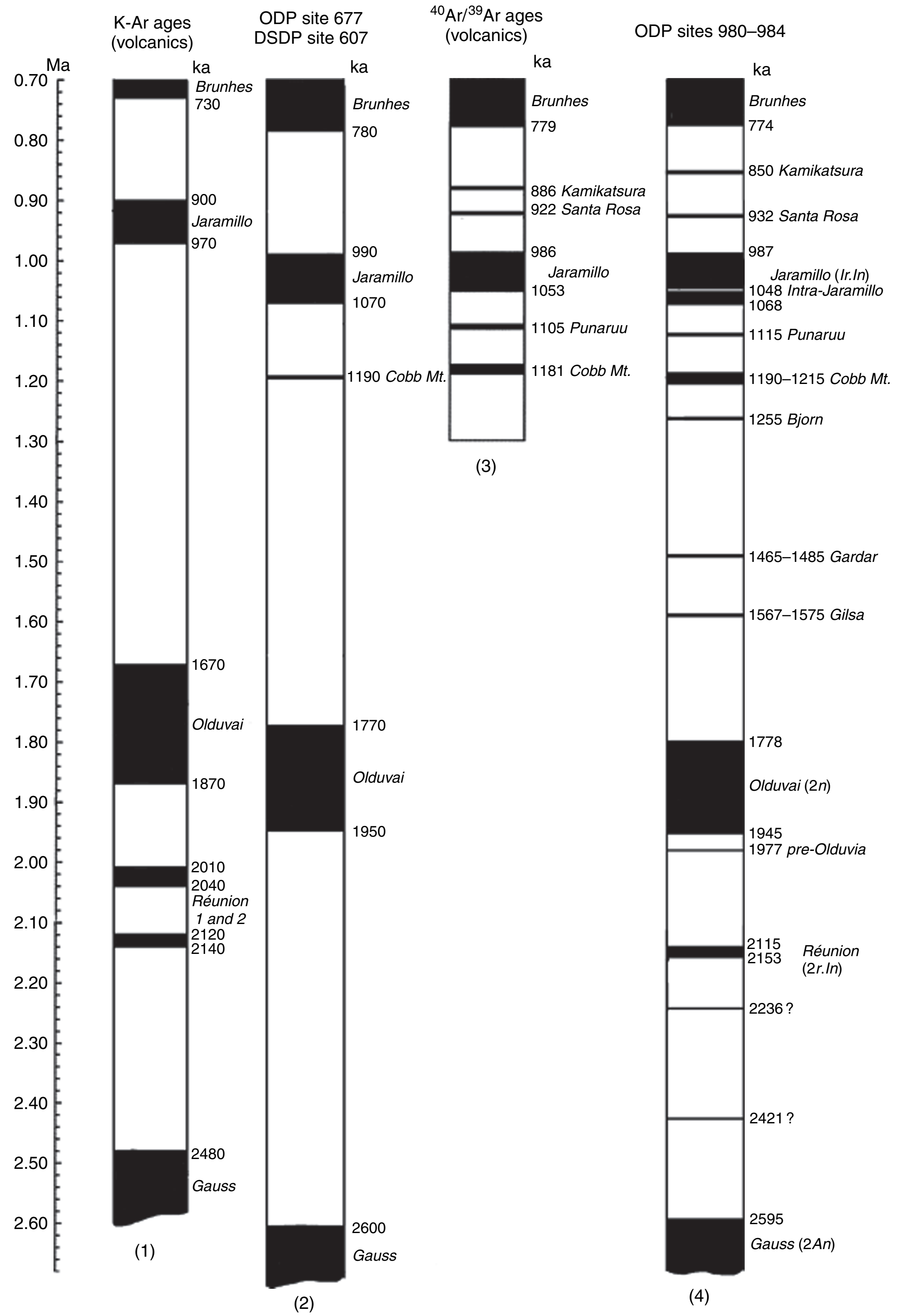

Figure 10 The geomagnetic polarity timescale for the Matuyama Chron. From left: (1) coupled K–Ar and paleomagnetic studies summarized by Mankinen and Dalrymple (1979); (2) ODP Site 677 and DSDP Site 607 (Shackleton *et al.*, 1990) as adopted by Cande and Kent (1995); (3) ^{40}Ar/^{39}Ar and paleomagnetic studies of volcanics (summarized by Singer *et al.*, 2004); (4) ODP Sites 980–984 (Channell and Kleiven, 2000; Channell *et al.*, 2002, Channell and Guyodo, 2004).

Table 3 Excursions within the Matuyama Chron

Excursion or subchron	Label	MIS Sites 607, 609, 677 Ref. 1	MIS Site 659 Ref. 2	MIS Italy Ref. 3	MIS (Sites 980–984) Ref. 4	Age (ka) (Sites 980–984) Ref. 4	Duration (kyr) (Sites 980–984) Ref. 4	Age (ka) $^{40}Ar/^{39}A$ Ref. 5
base Brunhes	base 1n	base 19			19	774		791
Kamakitsura					21	850	?	899
Santa Rosa					top 25	932	3	936
top Jaramillo	top 1r.1n	mid 27	27		base 27	985	8	1001
	1r.1n.1r				base 30	1050	3	
base Jaramillo	base 1r.1n	mid 31	31		base 31	1070	5	1069
Punaruu					mid 34	1115	5	1122
top Cobb. Mt.		base 35			base 35	1190		
Cobb Mt.							35	1194
base Cobb Mt.		base 35			top 37	1125		
Bjorn					top 38	1255	3	
Gardar					49	1465–1485	20	
Gilsa		53			54	1567–1575	8	
top Olduvai	top 2n	base 63	64	64	63	1778	4	1775
base Olduvai	base 2n	base 71	72	71	base 71	1945	5	1922
pre-Olduvai					top 73	1977	3	
Huckleberry Ridge					75	2040		2086
top Reunion	top 2r.1n	79		81	79	2115		
Reunion							38	2137
base Reunion	base 2r.1n			81	81	2153		
pre-Reunion 1					85/86	2236	?	
pre-Reunion 2					95	2421	?	
top Gauss		104			base 103	2595		

MIS: marine isotope stage
Ref. 1: Ruddiman et al. (1989); Raymo *et al*. (1989); Shackleton *et al*. (1990). Ref. 2: Tiedemann *et al*. (1994). Ref. 3: Lourens *et al*. (1996). Ref. 4: Channell and Kleiven (2000); Channell *et al*. (2002); Channell *et al*. (2003a); Channell and Guyodo (2004). Ref. 5: summarized in Singer *et al*. (2004).

ODP Site 677 can be matched to astronomical data with more confidence than the obliquity signal at this site or at DSDP Site 607 (Ruddiman *et al*., 1989) where the number of obliquity cycles within the Brunhes Chron was underestimated by a single cycle, thereby inadvertently supporting the K–Ar M/B boundary age (0.73 Ma). Although ODP Site 677 played a pivotal role in revamping the GPTS for the Matuyama Chron, the site itself did not yield a polarity stratigraphy. The Cobb Mt. Subchron, the M/B and Gauss–Matuyama (G/M) boundaries, and the boundaries of the Jaramillo and Olduvai subchronozones, were recorded at DSDP Site 607 (Clement and Robinson, 1987), and their ages were determined by transferring astrochronological ages from ODP Site 677 to DSDP Site 607 using oxygen isotope correlations between the two sites (Shackleton *et al*., 1990) (**Figure 10** and **Table 3**).

The sequence of reversals of the Matuyama Chron in the CK95 differs from that in the timescale of Mankinen and Dalrymple (1979). In CK95, one (as opposed to two) normal polarity subchron comprises the Réunion Subchron. CK95 adopted the astrochonological age estimates for Pliocene–Pleistocene polarity reversals (Shackleton *et al*., 1990; Hilgen, 1991a, 1991b). These original astrochronological age estimates have generally stood the test of further astrochronological dating of sediments during the last 15 years (see **Table 3** and **Figure 10**). The exception is the age of the Réunion Subchron in CK95 (2.14–2.15 Ma with 10 ky duration, derived from Hilgen (1991a, 1991b)), which should now be amended to 2.115–2.153 Ma (38 ky duration) based on data from North Atlantic high-sedimentation-rate drift sites (Channell *et al*., 2003a).

Soon after the K–Ar ages for Pliocene–Pleistocene polarity reversals (Mankinen and Dalrymple, 1979) were superceeded by astrochronological determinations, a large number of $^{40}Ar/^{39}Ar$ ages confirmed the astrochronological ages of polarity chrons. For example, $^{40}Ar/^{39}Ar$ age determinations for the M/B boundary and boundaries of the Jaramillo Subchron (Izett and Obradovich, 1991; Spell and McDougall, 1992; McDougall *et al.*, 1992; Tauxe *et al.*, 1992; Baksi *et al.*, 1993), for the Cobb Mountain Subchron (Turrin *et al.*, 1994), for the Réúnion Subchron (Baksi *et al.*, 1993) and for the boundaries of the Olduvai Subchron (Walter *et al.*, 1991; Baksi, 1994) were all close to the astrochronological estimates. The exercise becomes somewhat academic in view of the suggestion of Renne *et al.* (1994) that the $^{40}Ar/^{39}Ar$ standard (Fish Canyon sanadine, Mmhb-1) should be calibrated using the astrochronological ages of polarity reversals.

In CK92/95, 'tiny wiggles' in MMA data, interpreted either as brief polarity chrons (with duration <30 ky) or paleointensity fluctuations, were labeled 'cryptochrons'. Fifty-four 'cryptochrons' were recognized over the last 83 My, since the middle of the Late Cretaceous. In CK92/95, two 'cryptochons' are listed within the Matuyama Chron at 1.201–1.212 Ma (Cobb Mt. Subchron) and 2.420–2.441 Ma (Anomaly X in Heirtzler *et al.* (1968)). As outlined above, the Cobb Mt. Subchron is now a well-established feature within the Matayama Chron, whereas Anomaly X has yet to be well established in magnetostratigraphic records.

The Jaramillo, Réunion, and Olduvai normal polarity subchrons were identified almost 40 years ago. Since then, starting with the unequivocal identification of the Cobb Mountain Subchron in marine sediments (e.g., Clement and Kent, 1987), up to nine additional normal polarity excursions, and one reverse polarity excursion within the Jaramillo Subchron, have now been identified within the Matuyama Chron (**Figure 10** and **Table 3**).

5.10.3.2 Excursions between the Gauss–Matuyama Boundary and the Réunion Subchron

At ODP Site 982, two intervals with anomalous magnetization directions are observed in MIS 85/86 and MIS 95 (**Figure 10** and **Table 3**) (Channell and Guyodo, 2004). The older of the two coincides in age (2420 ka) with Anomaly X from the MMA data of Heirtzler *et al.* (1968) and hence with the 'cryptochron' of this age featured in CK92/95.

5.10.3.3 Huckleberry Ridge

Reynolds (1977) recorded anomalously shallow inclinations and southwestward-directed declinations in paleomagnetic data from 57 sites representing 23 separate localities in the Huckleberry Ridge Tuff (Yellowstone group). The anomalous directions were considered by Reynolds (1977) to record a polarity transition or an excursion, possibly associated with the Réunion Subchron. The tuff has recently been dated at 2.06 Ma using $^{40}Ar/^{39}Ar$ methods (Lanphere *et al.*, 2002), indicating that the transitional directions are younger than the Réunion Subchron for which the mean $^{40}Ar/^{39}Ar$ age is 2.14 Ma (Baksi and Hoffman, 2000; Baksi *et al.*, 1993; Roger *et al.*, 2000; Singer *et al.*, 2004). At ODP Site 981, low inclinations are recorded in MIS 75/76, stratigraphically above the Réunion Subchronozone which occurs in MIS 79–81 (Channell *et al.*, 2003a). It remains to be confirmed whether the anomalous directions within MIS75/76 correlate to the Huckleberry Ridge excursion as recorded in the Huckleberry Ridge Tuff.

5.10.3.4 Gilsa

The name Gilsa, in a geomagnetic context, owes its origin to McDougall and Wensink (1966) who detected two normal polarity flows separated by a reverse polarity flow within the lower part of the Matuyama Chronozone at Jokuldalur (Iceland). They assigned the older normal polarity flow to the Olduvai Subchronozone and the younger normal polarity interval, dated at 1.60 Ma, was named the Gilsa Event. Watkins *et al.* (1975) subsequently sampled the same sections and measured normal polarity directions in superposed lava flows that yielded ages of 1.58 and 1.67 Ma, but with no intervening reverse polarity interval. More recently, Udagawa *et al.* (1999), from work on five separate sections in the Jokuldalur region, confirmed the existence of a reverse polarity flow below the 1.60 Ma normal polarity flow and above the normal polarity flows associated with the Olduvai Chronozone. This work supports the conclusion of McDougall and Wensink (1966) that the Gilsa excursion (at

~1.60 Ma) is an interval of normal polarity distinct from the underlying Olduvai Chronozone.

In view of the Watkins *et al.* (1975) study, Mankinen and Dalrymple (1979) were not sufficiently confident to include the Gilsa Event in their timescale, nor was it included in CK95 as a 'cryptochron' as it is not evident in MMA records. The presence of a normal polarity interval at ~1.55 Ma in two holes at DSDP Site 609, however, confirmed its existence (Clement and Kent, 1987). This was corroborated at ODP Sites 983 and 984, where the correlative normal polarity excursion correlates to MIS 54 at 1567–1575 ka (**Table 3**), implying duration of 8 ky (Channell *et al.*, 2002).

5.10.3.5 Gardar and Bjorn

At ODP Sites 983 and 984, two normal polarity excursions are present between the Gilsa excursion zone and the Cobb Mountain subchronozone (Channell *et al.*, 2002). These two normal polarity intervals which occur in MIS 49 (Gardar excursion) and MIS 38 (Bjorn excursion), with estimated durations of 8 ky (Gardar) and 3 ky (Bjorn), have not been detected elsewhere, presumably due to their brief duration and paucity of deep-sea cores of this age with sufficiently high sedimentation rates. The names for these excursions are derived from the locations of Sites 984 (Bjorn Drift) and 983 (Gardar Drift); however, both excursions are present at both sites.

5.10.3.6 Cobb Mountain

Mankinen *et al.* (1978) documented a normal polarity site in the Alder Creek rhyolite at Cobb Mountain (California) which has yielded an $^{40}Ar/^{39}Ar$ age of 1186 ka (Turrin *et al.*, 1994). Further work by Mankinen and Grommé (1982) in the Cosa Range (California) supported the existence of the Cobb Mountain Event. DSDP Site 609 in the central North Atlantic Ocean provided the first unequivocal documentation of a normal polarity zone of similar age in deep-sea sediments (Clement and Kent, 1987). At Site 609, the so-called Cobb Mountain Subchron can be correlated to MIS 35/36 (Ruddiman *et al.*, 1989) (**Table 3**). This subchron has also been recognized at ODP Hole 647B in the southern Labrador Sea (Clement and Martinson, 1992), in the Celebes and Sulu Seas (Hsu *et al.*, 1990; Clement, 1992), in the Lau Basin (Abrahamsen and Sager, 1994), in New Zealand (Pillans *et al.*, 1994), off the California Margin (Guyodo *et al.*, 1999; Hayashida *et al.*, 1999), in the western Philippine Sea (Horng *et al.*, 2002, 2003), on the Bermuda Rise (ODP Leg 172) (Yang *et al.*, 2001), and at ODP Sites 980 and 983/984 where it correlates to MIS 35 at ~1.2 Ma and has an estimated duration of 35 ky (Channell *et al.*, 2002; Channell and Raymo, 2003).

5.10.3.7 Punaruu

The Punaruu excursion originates from the Punaruu Valley (Tahiti) where normal polarity magnetizations were recorded in basaltic lava flows stratigraphically below, and distinct from, the Jaramillo Subchronozone (Chauvin *et al.*, 1990). The Punaruu excursion in the type section on Tahiti has yielded an age of 1105 ka using $^{40}Ar/^{39}Ar$ methods (Singer *et al.*, 1999). This excursion appears in the sediment record from ODP Site 1021 (California Margin) where an age of 1.1 Ma is based on assumed uniform sedimentation rate between the M/B boundary and the top of the Jaramillo Subchronozone (Guyodo *et al.*, 1999). At both sites 983 and 984, the same excursion lies within MIS 34, which corresponds to an age of 1115 ka (Channell *et al.*, 2002).

5.10.3.8 Intra-Jaramillo Excursion (1r.1n.1r)

An intra-Jaramillo excursion has been detected in the Jingbian loess sequence from Northern China within loess 10 (Guo *et al.*, 2002), in marine cores from New Zealand (Pillans *et al.*, 1994), and at ODP sites 983 and 984 where it is correlated to MIS 30 at 1048 ka (Channell and Kleiven, 2000; Channell *et al.*, 2002).

5.10.3.9 Santa Rosa

Transitional magnetization directions from volcanic rocks at Cerro Santa Rosa I dome in New Mexico (Doell and Dalrymple, 1966; Doell *et al.*, 1968) were originally interpreted as recording the polarity transition at the end of the Jaramillo Subchron. $^{40}Ar/^{39}Ar$ ages (Spell and McDougall, 1992; Izett and Obradovich, 1994) from this dome are significantly younger than the end of the Jaramillo Subchron. More recently, anomalous (transitional) magnetization directions (mean: dec/inc: $103°/-63°$, $\alpha_{95} = 9.5°$) and a mean $^{40}Ar/^{39}Ar$ age of 936 ka (Singer and Brown, 2002) support the existence of a Santa Rosa excursion. An excursion of similar age (932 ka) has been observed at ODP Sites 983 and 984 at the top of MIS 25

(Channell *et al.*, 2002) and at the same position in the western Philippine Sea (Horng *et al.*, 2002, 2003).

5.10.3.10 Kamikatsura

The name 'Kamikatsura excursion' originates from the work of Maenaka (1983) who documented excursional magnetization directions in the Kamikatsura Tuff of the Osaka group (SW Japan). The existence of this excursion is supported by a 0.83 Ma normal-polarity flow from near Clear Lake (Mankinen *et al.*, 1981), and was promoted by Champion *et al.* (1988) in their review of Matuyama–Brunhes excursions. More recently, Takatsugi and Hyodo (1995) documented another excursion in marine clays about 10 m above the Kamikatsura Tuff, and 1 m above another tuff (Azuki Tuff) that has yielded a K–Ar age of 0.85 Ma. The Kamikatsura and Azuki Tuffs are separated by about 10 m of pebbly sand in the Osaka group (see Takasugi and Hyodo, 1995) and rapid deposition of this facies may mean that the two excursional intervals record a single geomagnetic excursion. On the other hand, two excursions with age estimates of 0.89 and 0.92 Ma have been detected in loess 9 (L9), which corresponds to MIS 22 according to the loess chronology of Heslop *et al.* (2000), in the Baoji loess section in southern China (Yang *et al.*, 2004). This observation follows earlier work in the Lishi and Luochuan regions (China) where an apparent excursion, also within L9, has been recorded (Wang *et al.*, 1980; Liu *et al.*, 1985). Although excursional directions in loess deposits have often been dismissed as remagnetizations in view of the probable delay of remanence acquisition of Chinese loess (see Spassov *et al.*, 2003), an interval of anomalously low VGP latitudes also occurs at ODP Site 983 in MIS 21 at about 850 ka (Channell and Kleiven, 2000). An interval of negative (reverse) inclination in equatorial Pacific core KK78O30 between the M/B boundary and the Jaramillo Subchronozone (Laj *et al.*, 1996) was assigned to the Kamikatsura excursion. Its age (based on uniform sedimentation rate between the M/B boundary and the top Jaramillo Subchronozone) is closer to that of the Santa Rosa excursion, which was not recognized at the time. Low VGP latitudes from volcanics on Maui and Tahiti associated with the Kamikatsura excursion yield mean ages of 866 ka (Singer *et al.*, 1999). Coe *et al.* (2004) associated a 25 m thick interval of anomalous declinations on Maui with the Kamikatsura excursion, and gave a weighted mean $^{40}Ar/^{39}Ar$ age of 900.3 ± 4.7 ka for this interval. To add to the confusion in this interval, a transitionally magnetized flow from La Palma (Canary Islands) has yielded a K–Ar age of 821 ka (Quidelleur *et al.*, 2002). There is clearly much to be done to resolve the behavior of the geomagnetic field in this interval immediately prior to the M/B boundary. The relationship of the Kamikatsura excursion and paleointensity minimum immediately prior to the M/B boundary (Kent and Schneider, 1995; Hartl and Tauxe, 1996) remains to be determined.

5.10.4 Geomagnetic Excursions in Pre-Matuyama Time

In the Gauss and Gilbert chrons, there are no well-documented excursions in magnetostratigraphic records, nor are there 'tiny wiggles' in MMA data denoting 'cryptochrons' of this age. Based on 'tiny wiggles' in MMA data, CK92/95 included four 'cryptochrons' in the Late Miocene, 18 in the Oligocene, 3 in the Late Eocene, and 23 in the Paleocene and Early Eocene. These 'tiny wiggles' can often be correlated between ship's tracks, and they have been thought to represent fluctuations in geomagnetic field intensity (Cande and LaBrecque, 1974; Cande and Kent, 1992b) and/or short polarity intervals (e.g., Blakely and Cox, 1972; Blakely, 1974).

From magnetostratigraphic studies of sedimentary sequences, there is evidence for ~14 polarity excursions in the Miocene (**Figure 11**), four of which correlate with cryptochrons in CK92/95. The Miocene magnetostratigraphic record, like that in the Brunhes Chron, implies that there are far more polarity excursions than are represented in the cryptochron record from MMA data. On the other hand, whereas 44 Paleocene to Oligocene (Paleogene) cryptochrons are given in CK92/95, only three excursions have been documented in the Paleogene, and only one of these appears to correlate with a cryptochron in CK92/95.

In the Middle Cretaceous, at the base of the Creteceous Long Normal interval, the existence of reverse polarity subchrons younger than CM0 has been advocated (e.g., Tarduno *et al.*, 1992) although their existence remains controversial. Here, the uncertainty is due to the lack of continuous sedimentary sections recording both CM0 (at the base of the

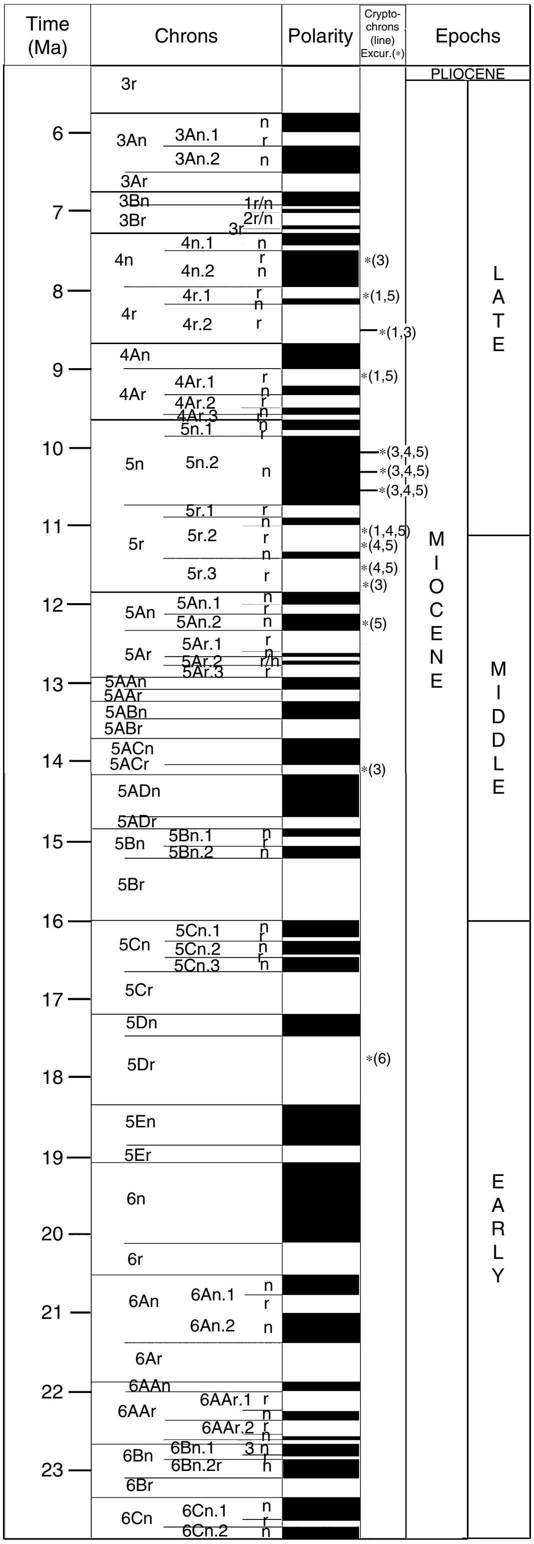

Figure 11 Miocene geomagnetic polarity timescale of Cande and Kent (1995) with cryptochrons from Cande and Kent (1992a) and additional polarity subchrons/excursions according to various authors: 1 = Schneider (1995), 2 = Acton *et al*. (2006), 3 = Evans and Channell (2003), Evans *et al*. (2004), 4 = Abdul Aziz and Langereis (2004), 5 = Krijgsman and Kent (2004), 6 = Channell *et al*. (2003b).

Cretaceous Long Normal) and the younger proposed subchrons, and the inability of $^{40}Ar/^{39}Ar$ age dating to clearly distinguish these subchrons in discontinuous volcanic sequences.

5.10.4.1 C5n.2n (Late Miocene)

Late Miocene polarity subchron C5n.2n in the CK92/95 GPTS includes three 'cryptochrons' first recorded as 'tiny wiggles' in MMA profiles from the North Pacific Ocean (Blakely, 1974). Over the last 25 years, 'tiny wiggles' in MMA records of C5n.2n have been represented as polarity subchrons in several versions of the GPTS notably those of Ness *et al.* (1980) and Harland *et al.* (1982, 1990). They were not included in the GPTS by Lowrie and Alvarez (1981) or Berggren *et al.* (1985) because of the lack of confirmation for polarity subchrons of this age from magnetostratigraphy.

A single track of deep-tow magnetic anomaly data, covering polarity subchron C5n.2n, was collected in 1998 at 19°S on the flanks of the East Pacific Rise (Bowers *et al.*, 2001). The half-spreading rate at this site was estimated to have been 9 cm yr^{-1} for the Late Miocene. This profile revealed more than twice as many 'tiny wiggles' as seen in profiles from the North Pacific region, where Blakely (1974) first identified the four short wavelength anomalies within C5. Bowers *et al.* (2001) identified a number of short wavelength magnetic anomalies within C5n.2n, three of which were interpreted as correlative to the three cryptochrons in CK92/95. Bowles *et al.* (2003) addressed the question as to whether these tiny wiggles within C5n.2n represent polarity subchrons/excursions or paleointensity fluctuations. These authors compared the directional and paleointensity record from C5n.2n in the sedimentary sequence at ODP Site 887 with the deep-tow record of Bowers *et al.* (2001). The absence of directional excursions in the magnetostratigraphic record of C5n.2n led Bowles *et al.* (2003) to conclude that the tiny wiggles in MMA records of C5n.2n were produced by paleointensity fluctuations rather than directional excursions or brief polarity subchrons.

The magnetic polarity stratigraphy at ODP Site 884, on the slopes of the Detroit Seamount (NW Pacific), was initially interpreted from the shipboard pass-through magnetometer measurements of split half-cores (Weeks *et al.*, 1995). The C5n part of the section was reinterpreted by Roberts and Lewin-Harris (2000) who recognized two short polarity subchrons and one excursion within a polarity zone correlative to C5n.2n (see figure 3 in Roberts and Lewin-Harris, 2000). The excursion and subchrons were estimated to have durations of 6, 23, and 28 ky, respectively, assuming constant sedimentation rates in C5n.2n.

At ODP Site 1092, in the sub-Antarctic South Atlantic, the stratigraphic interval correlative to C5n.2n is 34 m thick with a mean sedimentation rate of 3.3 cm ky^{-1}. At this site, three polarity excursions are recognized within C5n.2n, all of which have estimated durations <10 ky (Evans and Channell, 2003; Evans *et al.*, 2004). The presence of directional excursions within C5n.2n has recently been supported by measurements of new discrete samples collected at DSDP Site 608 (Krijgsman and Kent, 2004), although at this site the excursions correlated to C5n.2n are represented by single samples with shallow negative inclinations.

Continental records of short polarity events in C5n.2n have been reported from NE Spain (Garces *et al.*, 1996), Bolivia (Roperch *et al.*, 1999), and western China (Li *et al.*, 1997). The section from the Pyrenees in NE Spain consists of a composite section from two terrestrial sequences that cover the interval from 8.7 to 11.1 Ma (Garces *et al.*, 1996). The magnetic stratigraphy can be interpreted on the basis of polarity zone pattern fit to the GPTS aided by the first occurrence datum of *Hipparion*. The polarity zone correlative to C5n.2n has a thickness of 175 m and includes a ~5 m polarity zone in its upper part that was correlated with cryptochron C5n.2n-1 in CK92/95. A 4.5 km thick composite section of red beds from the Bolivian Altiplano produced a magnetic stratigraphy for the 9–14 Ma interval (Roperch *et al.*, 1999). The polarity zone correlative to C5n.2n is 1 km thick in this section and contains a single 5 m thick polarity zone that was correlated with cryptochron C5n.2n-1. The interpretation of the polarity stratigraphy is supported by $^{40}Ar/^{39}Ar$ age determinations on two tuff layers within the section, one of which falls within C5n.2n. An upper Cenozoic section in western China recorded three short reverse polarity zones within C5n.2n (Li *et al.*, 1997). The polarity interpretation at the site is fairly unambiguous, and is corroborated by the presence of vertebrate fossils. The events are, however, only represented by single discrete samples. The authors correlate these events with the three cryptochrons in CK92/95.

5.10.4.2 Other Miocene Excursions/ Subchrons

In sediments from the equatorial Pacific Ocean recovered during ODP Leg 138, a normal polarity subchron was recognized within C4r.2r (Schneider, 1995) that correlates to polarity subchrons recorded in sediments from ODP Site 1092 in the South Atlantic (Evans and Channell, 2003; Evans *et al.*, 2004) and from ODP Site 1095 off the Antarctic Peninsula (Acton *et al.*, 2006), as well as to a cryptochron in CK92/95 (**Figure 1**). Independent verification from multiple sites argues strongly for an excursion or brief subchron within C4r.2r (**Figure 11**).

Three other subchrons or excursions within C4r.1r, C4Ar.1r, and C5r.2r (**Figure 11**) were apparent from the study of ODP Leg 138 sediments (Schneider, 1995). All three of these subchrons (or excursions) are apparently supported by recent data from a resampling of DSDP Site 608 (Krijgsman and Kent, 2004). One of these subchrons/excursions, that within C5r.2r, is also supported by the studies of Abdul-Aziz and Langereis (2004) in continental sediments from NE Spain. Abdul-Aziz and Langereis (2004) documented a total of three subchrons within C5r (**Figure 11**). The oldest of the C5r excursions (that within C5r.3r) may correlate with a subchron found within the same polarity chron at ODP Site 1092 (Evans and Channell, 2003). At ODP Site 1092, yet another newly recognized subchron has been observed within chron C5ACr (reinterpreted from the initial placement within C5AAr by Evans and Channell, 2003).

The plethora of Late and Middle Miocene subchrons recognized in stratigraphic sections clearly out-numbers the cryptochrons identified in CK92/95 derived from MMA data (**Figure 11**). At least 13 subchrons that were not included in the CK92/95 polarity chron template have been documented in stratigraphic section since the publication of CK95. Four of 13 newly recognized subchrons correlate with cryptochrons in CK92/95.

5.10.4.3 Oligocene and Eocene

Three 'cryptochrons' appear in C13r in the CK92/95 GPTS (**Figure 12**) based on 'tiny wiggles' in MMA data. Bice and Montanari (1988) found one normal polarity sample in the top third of C13r at Massignano, Italy, in the Eocene–Oligocene boundary stratotype section. Lowrie and Lanci (1994) resampled the Massignano section and concluded that there were no normal polarity intervals within C13r. This conclusion appears to be corroborated by the study of cores recovered from a borehole drilled at Massignano (Lanci *et al.*, 1996). At ODP Site 1090, in the sub-Antarctic South Atlantic Ocean, the high sedimentation rates in C13r appear to have facilitated the recording of a short subchron represented by a 3 m thick normal polarity zone, implying a duration for the C13r.1n subchron of 79 ky, which is

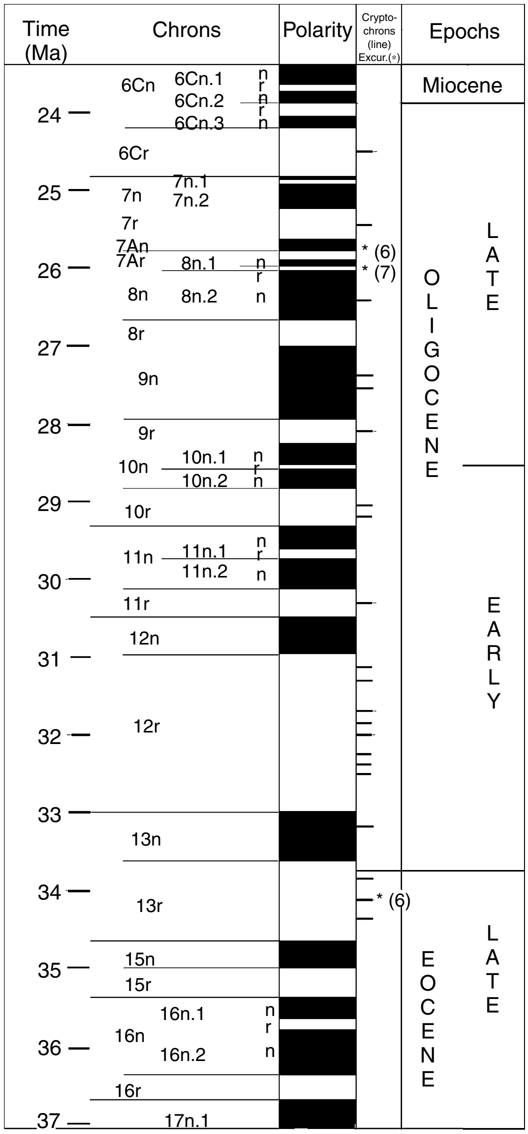

Figure 12 Late Eocene–Oligocene geomagnetic polarity timescale of Cande and Kent (1995) with cryptochrons from Cande and Kent (1992a) and additional polarity subchrons/excursions according to: 6 = Channell *et al*. (2003b) and 7 = Lanci *et al*. (2005).

comparable to the duration of the Jaramillo Subchron (Channell *et al.*, 2003b).

Apart from the apparent subchron within C13r, two additional polarity subchrons not included in the CK92/95 GPTS are observed at ODP Site 1090 (Channell *et al.*, 2003b). One of these (within C5Dr) is listed as a cryptochron in table 7 of CK92, and was included as a subchron in versions of the GPTS prior to CK92/95. This normal polarity subchron was originally identified in the North Pacific MMA stack (Blakely, 1974). Two other normal polarity subchrons have been documented in the Late Oligocene (**Figure 12**). One of these occurs within C7Ar and is recorded at ODP Site 1090 (Channell *et al.*, 2003b) and another occurs within C8n.1n at ODP Site 1218 in the equatorial Pacific Ocean (Lanci *et al.*, 2005). Neither of these subchrons has been recognized in other magnetostratigraphic records, or in MMA records.

5.10.4.4 Middle Cretaceous

The record of polarity subchrons in the Middle Cretaceous begins with the so-called ISEA reversal that was originally recognized in a bed of reddish Aptian limestone along a road outcrop in the Umbrian Apennines (Italy) by VandenBerg *et al.* (1978). Lowrie *et al.* (1980) resampled this anomalous limestone bed, and confirmed the reverse polarity magnetization but were unable to substantiate the existence of this reverse polarity zone in other coeval Umbrian sections. Some 10 years later, Tarduno (1990) documented reverse polarity magnetizations in two samples within the *G. algerianus* foraminiferal zone of the middle Aptian at DSDP Site 463. Documentation of this short polarity subchron is strengthened by two samples with reverse magnetization spanning a 43 cm interval at ODP Site 765 (Ogg *et al.*, 1992). Reverse magnetizations in basalts from the Tarim Basin (China) yielding $^{40}Ar/^{39}Ar$ whole rock and plagioclase fraction ages of 113 and 119 Ma, respectively (Sobel and Arnaud, 2000), have been associated with ISEA (Gilder *et al.*, 2003). According to some timescales (e.g., Channell *et al.*, 1995a), an age of 119 Ma is less than 2 My younger than CM0 (the youngest reverse polarity chron of the M-sequence) at the Aptian/Barremian boundary. On the other hand, in the timescale of Gradstein *et al.* (2004), an age span of 113–119 Ma would correspond to the Late Aptian, with 119 Ma corresponding closely to the supposed age of ISEA. An andesitic lava sequence from Liaoning Province (China) also yields reverse magnetization directions and $^{40}Ar/^{39}Ar\pm$ ages of 116.8 Ma $\pm$ 3.0 Ma (Zhu *et al.*, 2004) and these authors associate the reverse magnetizations either to ISEA or to CM0.

Tarduno *et al.* (1992) documented seven reverse polarity zones in the middle Albian of the Contessa section (Umbria, Italy). The cyclostratigraphy of Herbert *et al.* (1995) applied to the Contessa section implies a duration of $\sim$800 ky for the thickest of these polarity zones. As this duration is greater than the estimated duration of CM0, the polarity chron at the base of the Aptian that is easily recognized in MMA records, it is unlikely the Albian polarity zones documented in the Contessa section represent reverse polarity chrons. As the reversely magnetized limestones in the Albian at Contessa are reddish, and the NRM is partly carried by hematite, the hematite magnetizations may be recording a Late Cretaceous or younger magnetization rather than an Albian one. On the other hand, reverse polarity magnetizations have been retrieved from sediments of Albian age recovered during ODP Leg 171 although these varicolored sediments indicate extensive iron mobilization and probably indicate a (Late Cretaceous) remagnetization (Ogg and Bardot, 2001).

The template for the Late Jurassic and Early Cretaceous polarity sequence (CM0–CM25) stems from the MMA records published by Larson and Hilde (1975). Correlation of this sequence of polarity chrons to biozonations/stage boundaries and numerical ages has been reviewed by Channell *et al.* (1995a). The only modification of this template in the intervening 30 years since publication of Larson and Hilde (1975) has been the identification of a second reverse polarity subchron between CM11 and CM12. The two reverse polarity subchrons within CM12n have been recognized in MMA records (Tamaki and Larson, 1988) and in magnetostratigraphic sections in Italy (Channell *et al.*, 1987; Channell *et al.*, 1995b).

5.10.5 Duration of Geomagnetic Excursions

Precise evaluation of the duration of geomagnetic excursions has become a point of interest since the proposal of Gubbins (1999) that, during excursions, the geomagnetic field reverses polarity in the Earth's liquid outer core but that this outer-core field does not persist for long enough in a reversed polarity state for diffusion of the field into the solid inner core. Diffusion times of $\sim$3 ky for the inner core, therefore, provide a prediction for excursion duration. It is not trivial to determine excursion durations of a few kiloyears in the geologic record. Only in cases where astronomical tuning is possible to the level of orbital precession can we expect durations of the order of thousands of years to be adequately resolved. Interpolation, assuming constant sedimentation rates, among tie points matching an oxygen isotope record to an (astronomically tuned) isotope target curve is unlikely to provide duration estimates of sufficient precision. Radiometric ($^{40}Ar/^{39}Ar$ or K/Ar) ages are also unlikely to have realistic uncertainties within a few kiloyears, limiting their utility for estimating the duration of excursions.

Nonetheless, for the Laschamp excursion, Laj *et al.* (2000) presented a duration estimate of 1500 years based on correlation of marine cores to the GISP2 (Greenland) layer-counted ice-core record. Here, correlation among the NAPIS-75 cores was achieved through correlation of the susceptibility records (Kissel *et al.*, 1999), and the oxygen isotope record from one of the cores provided a correlation to GISP2 (Voelker *et al.*, 1998). Approximately, the same duration estimate was obtained from each of the five NAPIS-75 cores that record the excursion. As these cores are spread over a distance of about 5000 km, it is unlikely that the duration estimate is affected by local changes in the sediment deposition rate at the time of the excursion. Further evidence for the duration of the Laschamp excursion is obtained from the record of flux of ^{36}Cl in the GRIP ice core. Wagner *et al.* (2000) showed that variations in ^{36}Cl flux (assumed to be entirely due to modulation by the geomagnetic field) are similar (in inverse sense) to changes in geomagnetic field intensity (e.g., **Figure 6**). A duration estimate of 1500 years for the ^{36}Cl anomaly, corresponding to the paleointensity minimum associated with the Laschamp excursion, supports the excursion duration determined from paleomagnetic records. The ^{36}Cl flux ice-core record also indicates that a second peak, associated with the Mono Lake excursion, has approximately the same duration, about 1500–2000 years (on the GRIP–GISP age model).

Although none of the sedimentary records containing the Laschamp excursion are accompanied by primary astrochronologic tuning to orbital solutions, there is good consistency in estimates of excursion duration. For example, Lund *et al.* (2001a, 2005) estimated a duration of 2 ky for the Laschamp excursion using age models based on AMS radiocarbon ages from Keigwin and Jones (1994). A similar duration is apparent for records of the Laschamp excursion in the South Atlantic Ocean (Channell *et al.*, 2000) and the southern Indian Ocean (Mazaud *et al.*, 2002). In the Irminger Basin, both the Mono Lake and Laschamp excursions have apparent durations of $\sim$1 ky (Channell, 2006). Similarly, duration estimates for the Iceland Basin excursion are 2–3 ky for the records that have oxygen isotope age control (e.g., Channell *et al.*, 1997; Channell, 1999; Stoner *et al.*, 2003; Laj *et al.*, 2006). A longer duration ($\sim$8 ky) for the Iceland Basin excursion recorded at ODP Site 919 (Channell, 2006) may be attributable to inadequacy of the oxygen isotope age model. A somewhat longer duration has been documented for the Blake Event, with values of 5–8 ky (Tric *et al.*, 1991). These duration estimates are about 2–3 times larger than most of those observed for the Laschamp, Mono Lake, and Iceland Basin excursions, although the duration estimates are heavily dependent on the quality of the individual age models.

In general, longer duration estimates for geomagnetic excursions have come from high-latitudes cores from the Atlantic and Arctic Oceans. In the work of Nowaczyk and Antonow (1997) and Nowaczyk and Knies (2000), excursional durations of >10 ky are implied by the age models. The structure of the excursions is also different, with rapid transitions, and distinct periods of time with fully reversed polarity VGPs. As mentioned above, the results from the Arctic cores are anomalous in that, for the last 100 ky, the reverse polarity intervals occupy almost 50% of the recovered sedimentary sequence, which is obviously much greater than expected. It has been suggested that these anomalously long duration estimates for excursions in this region is due to large increases in sedimentation rate during the excursions, which would be particularly noticeable at high latitudes, since, as noted by Worm (1997), many excursions seem to occur preferentially during cold or cooling climatic stages. According to Clement's (2004) review, reversal duration at the Matuyama–Brunhes boundary, and at the boundaries of the Jaramillo and Olduvai subchrons, varies with latitude and falls in the 2–10 ky range. This distribution of reversal durations with latitude is broadly consistent with a model in which nondipole fields are allowed to persist while the axial dipole decays to zero and then builds in the opposite direction. The apparent increase in excursion duration in high latitude cores (Nowaczyk and Antonow, 1997; Nowaczyk and Knies, 2000) appears to be greater than can be accommodated by such a model. King *et al.* (2005) reported preliminary results from Pleistocene sediments from the Lomonosov Ridge, Central Arctic Ocean, collected during IODP Leg 302. Excursional magnetization directions are apparently present during the early Brunhes Chron. However, these authors observe a strong correlation between rock magnetic variations, color changes and physical properties, and the observed excursions. They therefore attribute the observed Arctic paleomagnetic behavior to environmental controls rather than to anomalous geomagnetic behavior.

In summary, the duration estimates for the Laschamp, Mono Lake, and Iceland Basin excursions (**Table 2**) and for most Matuyama Chron excursions (**Table 3**) provide support for the excursion mechanism of Gubbins (1999) in that the excursion durations are estimated to be no more than a few kiloyears, which is comparable with the magnetic diffusion time of the inner core. This result is consistent with the concept that excursions can lead to a full polarity reversal and a subsequent prolonged polarity interval only if the excursion (with directions approximately antiparallel to the pre-excursional field) is maintained for times exceeding the magnetic diffusion time of the inner core.

5.10.6 Excursional Field Geometry

Determination of the field geometry during an excursion requires availability of multiple records from different and widely separated geographical locations. Only an incomplete picture can be obtained, because only two excursions, the Laschamp and Iceland Basin excursions, have been studied at distant localities from rapidly deposited sediments that yield detailed VGP paths (e.g., Laj *et al.*, 2006). For the Iceland Basin excursion (**Figure 8**), a consistent picture of VGP paths is obtained for widely distributed site locations. VGPs move first southward over Africa along a rather narrow band of longitudes before crossing the equator and reaching high southern latitudes. The return paths to the Northern Hemisphere are contained in a longitudinal band over eastern Asia. The overall picture is that of a large counterclockwise loop of the VGPs. A coherent picture is also observed for VGP paths for the Laschamp excursion (**Figure 5**). The first southward part of the paths passes over east Asian–western Pacific longitudes and then reaches high southern latitudes. The northward-directed part of the VGP paths proceeds over Africa and Western Europe. For the Laschamp as well as for the Iceland Basin excursion, the turning point, where the VGPs change from being southward to northward, coincides with the minimum in relative paleointensity. In other words, we see no clear evidence for recovery of paleointensity within the excursion interval, although a recovery in paleointensity may be filtered by the DRM acquisition process.

There are similarities among the records obtained for the Laschamp and Iceland Basin excursions. For instance, although the sense of movement of the VGP path is opposite for the Iceland Basin (counterclockwise) and Laschamp (clockwise) excursions (**Figures 5** and **8**), the two sets of VGP paths pass over similar longitudinal bands. This suggests that a similar core–mantle boundary structure may have prevailed during the two excursions. In addition, the repetitive structure of the VGP paths for dispersed site locations can be taken as an indication of a simple, possibly dipolar, geometry for the transitional geomagnetic field for both excursions. The VGP paths and the intensity records for the two excursions are consistent with a decrease in strength of the axial dipole, a substantial transitional equatorial dipole, and a reduced nondipole field relative to the axial dipole. During the first, $N \rightarrow S$ part of the Iceland Basin VGP path, over Europe and Africa, the g_1^1 term of the equatorial dipole appears to be preponderant, while h_1^1 appears to be dominant during the $S \rightarrow N$ part of the excursion. For the Laschamp excursion, the opposite is true, with the first part dominated by h_1^1 and the second part dominated by g_1^1 (Laj *et al.*, 2006).

These results do not support the view that the geomagnetic intensity minima that coincide with directional excursions reflect the emergence of nondipole geomagnetic components (e.g., Merrill and McFadden, 1994; Guyodo and Valet, 1999). If nondipole fields were dominant, one should observe widely different VGP paths at the different sites, which is not the case here, nor was it the case in Clement's (1992) analysis of the Cobb Mountain subchron or in McWilliams' (2001) analysis of the Pringle Falls excursion. As originally pointed out by Valet and Meynadier (1993), the intensity of the geomagnetic field is substantially reduced during excursions. As an important corollary, the dominance of nonaxial dipolar fields during excursions implies that the amplitudes of the nondipole components may have been relatively reduced during excursions. Remarkably, the transitional VGPs appear to follow the same paths for the Laschamp and Icelandic Basin excursions which are separated in time by $\sim$140 ky. This is obviously longer than the time constant associated with fluid motions in the outer core, and therefore might suggest a deep Earth (lower mantle) control on the excursional field geometry (Laj *et al.*, 2006). The Blake excursion, however, situated in time between the Laschamp and Iceland Basin excursions, does not appear to have the same transitional field geometry. For the Blake excursion, there is only a single record (Tric *et al.*, 1991) that incorporates enough transitional magnetization directions for the transitional VGP path to be mapped. The VGP path for this record lies over the Americas during the N $\rightarrow$ S part reaching high southern latitudes (**Figure 7**). The return path, although much less detailed, is situated over Australia and South Eastern Asia. The VGP path, albeit based on a single record, is thus different from that of the Laschamp and Iceland Basin excursions. For the Pringle Falls excursion, on the other hand, several records from different sites yield consistent results. As noted by McWilliams (2001), this is an indication that the transitional field had a large dipolar component during the Pringle Falls excursion. Curiously, the path for the Pringle Falls excursion is similar to that observed for the Blake excursion from the Mediterranean record of Tric *et al.* (1991).

5.10.7 Concluding Remarks

The paleomagnetic records for the two best-documented excursions (Laschamp and Iceland Basin excursions) imply that the field during excursions was characterized by a simple transitional geometry. The axial dipole underwent a substantial decrease in strength, while equatorial dipoles were apparently relatively enhanced during the excursions. Contrary to a commonly held view, the nondipole field may have undergone a decrease during excursions. Although the sense of looping of the VGP paths is opposite for the Iceland Basin and Laschamp excursions, the VGPs for the two excursions (separated in time by $\sim$140 ky) followed the same path, which suggests a repetitive lower mantle control on excursional field geometry. Some excursions, notably the Blake excursion (Tric *et al.*, 1991; Zhu *et al.*, 1994) and some records of the Iceland Basin and Pringle Falls excursions (Channell, 2006), appear to be characterized by multiple VGP swings to high latitudes. Such field instability may be explained by the observation that the critical Reynolds number for the onset of core convection is very sensitive to the poloidal field, and the strength of core convection varies wildly in response changes in magnetic field strength particularly during intensity minima (Zhang and Gubbins, 2000).

The duration of excursions in the Brunhes Chron (**Table 2**), as well as excursions with the Matuyama Chron (**Table 3**), based mainly on constraints from oxygen isotope stratigraphy, is usually estimated to be $<$5 ky. This duration is comparable with the $\sim$3 ky timescale for diffusive field changes in the Earth's solid inner core, which must reverse polarity in order for a full geomagnetic reversal to be sustained. The fact that our estimate of excursion duration is comparable with the $\sim$3 ky time constant for inner core magnetic diffusion provides support for the suggestion by Gubbins (1999) that there is a mechanistic distinction between polarity reversals, that define polarity chrons, and excursions. Nevertheless, the similarity of some VGP paths of excursions and reversal transitions bounding polarity chrons suggests an inherent link between the mechanisms that give rise to geomagnetic excursions and reversals. For instance, the East Asian longitudinal bands identified in the excursional VGP paths coincides with one of the preferred longitudinal bands for transitional VGP paths during reversals (Laj *et al.*, 1991; Clement, 1991), and are featured in VGP clusters that appear in volcanic and high-resolution sedimentary records of reversal transitions (e.g., Hoffman, 1992; Channell and Lehman, 1997; Channell *et al.*, 2004). The longitudinal band over western Europe and Africa, on the other hand, is less evident in compilations of sedimentary VGP reversal paths; however, this region does feature in a recent compilation of transitional VGP paths from volcanic rocks (**Figure 13**; Valet and Herrero-Bervera, 2003).

An increasing number of relative paleointensity records, based on normalized remanence data from marine sediments, are now available for the Matuyama Chron (Valet and Meynadier, 1993; Meynadier *et al.*, 1995; Kok and Tauxe, 1999; Hayashida *et al.*, 1999; Guyodo *et al.*, 1999, 2001; Channell and Kleiven, 2000; Dinares-Turell *et al.*, 2002; Channell *et al.*, 2002; Horng *et al.*, 2003). The sedimentary sequences that have yielded high-quality relative paleointensity records only rarely capture the directional changes associated with excursions. Presumably, paleointensity minima are captured more readily because the paleointensity features are longer lasting than the accompanying directional

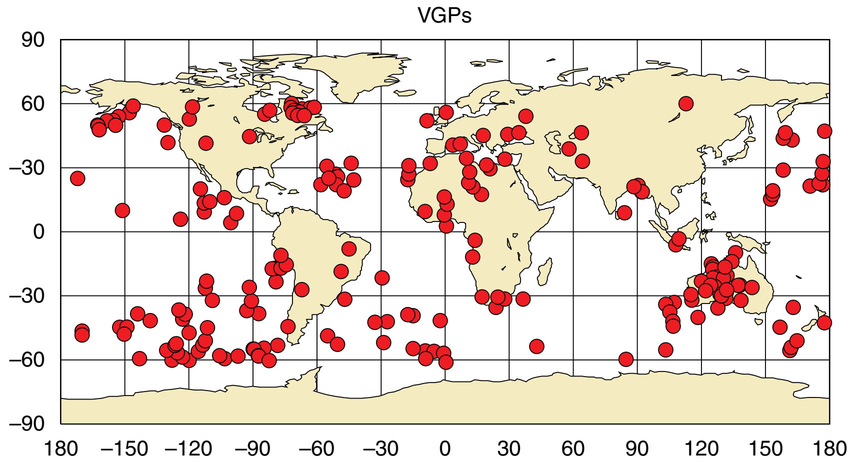

Figure 13 Polarity transition VGPs with latitudes lower than 60° obtained from volcanic sites with large global distribution. Adapted from Valet J-P and Herero-Bervera E (2003) Some characteristics of geomagnetic reversals inferred from detailed volcanic records. *Comptes Rendus Geoscience* 335: 79–90.

excursions, and are perhaps less susceptible to magnetic overprinting and the filtering affect of a finite magnetization lock-in zone during DRM acquisition. The short duration of directional excursions (**Tables 2 and 3**) result in them being rarely recorded in sediments with mean sedimentation rates of less than several cm ky^{-1}. It has often been noted that the ages of paleointensity minima in paleointensity records correspond to the excursions found elsewhere (e.g., Valet and Meynadier, 1993). As an example of this correspondence, we show the ODP Site 983 paleointensity record for the 750–1800 ka interval (**Figure 14**) with the position of polarity reversals and excursions as recorded in the same

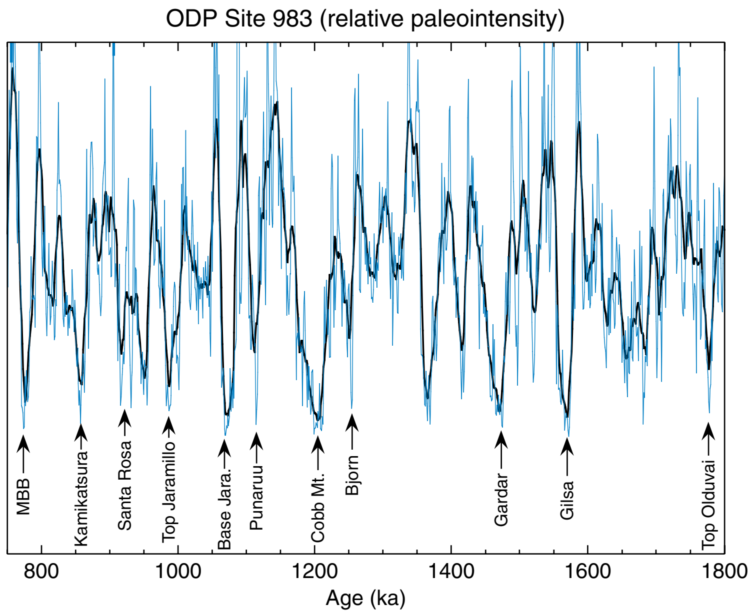

Figure 14 ODP Site 983 relative paleointensity record (blue) with a 10 kyr running mean (black) of the same record. Ages of principal polarity reversals and excursions are indicated, and correspond to paleointensity minima. Data from Channell JET and Kleiven HF (2000) and Channell *et al*. (2002).

sediment sequence (Channell and Kleiven, 2000; Channell *et al.*, 2002).

In the next few years, paleomagnetic studies of rapidly deposited sediments will be combined with astrochronological age models, and coupled with $^{40}Ar/^{39}Ar$ dating and paleomagnetic studies from volcanic rocks. These will converge toward a concensus on the short-term (excursional) behavior of the geomagnetic field. At present, the majority of documented excursions is in the Brunhes Chron (**Figure 1**), the Matuyama Chron (**Figure 10**), and in the Late and Middle Miocene (7–14 Ma) interval (**Figure 11**). It remains to be seen whether these time intervals are representative of excursion frequency in general, or whether they remain exceptional, as the catalog of excursions expands. It is interesting to note that the high frequency (and short duration) of excursions in these intervals, if applied to the entire Cenozoic GPTS, would lead to a distribution of chron/excursion durations inconsistent with a Poisson distribution (see Lowrie and Kent, 2004), implying that polarity excursions are mechanistically distinct from polarity chrons.

The study of polarity excursions, today, is analogous to the study of polarity chrons 40 years ago when the principal polarity chrons of the last 5 My were in the process of being resolved by coupled K–Ar and paleomagnetic studies in volcanic rocks, and by studies of conventional piston cores from the oceans. The challenge of resolving brief excursions, that have remained obscure and poorly documented until very recently, has only just begun.

Acknowledgments

The authors thank M. McWilliams and J.-P. Valet for providing **Figures 9** and **13** respectively, and N. Nowaczyk and H. Oda for providing numerical data, respectively.

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