Ultrafast oceanic spreading of the Marsili Basin, southern Tyrrhenian Sea: Evidence from magnetic anomaly analysis

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ABSTRACT

Spectral analysis of both shipborne and airborne magnetic maps of the southern Tyrrhenian Sea reveals seven subparallel positive-negative magnetic anomaly stripes over the flat-lying deep floor of the Marsili oceanic basin. This represents the first evidence of oceanic magnetic anomalies in the Tyrrhenian Sea. The central positive stripe is along the Marsili seamount, a superinflated spreading ridge located at the basin axis. The stratigraphy of Ocean Drilling Program Site 650 and K/Ar ages from the Marsili seamount suggest that the Marsili Basin opened at the remarkable full-spreading rate of ~19 cm/yr between ca. 1.6 and 2.1 Ma about the Olduvai subchron. This is the highest spreading rate ever documented, including that observed at the Cocos-Pacific plate boundary. Renewed but slow spreading during the Brunhes chron (after 0.78 Ma), coupled with huge magmatic inflation, gave rise to the Marsili volcano. Our new data and interpretation show that backarc spreading of the Tyrrhenian Sea was episodic, with sudden rapid pulses punctuating relatively long periods of tectonic quiescence.

Keywords: magnetic anomalies, Tyrrhenian Sea, ocean spreading, Marsili Basin.

INTRODUCTION

The analysis of the magnetic anomaly stripes of the oceans has greatly advanced knowledge of the crustal evolution of Earth. These anomalies provide fundamental constraints for determining the modes and timing of oceanic spreading.

In the Mediterranean domain (Fig. 1), the diachronous backarc basins of the Liguro-Provencal and Tyrrhenian Seas opened as a result of the generally eastward retreat of the Ionian-Adriatic slab, which has been passively sinking into the mantle since late Oligocene time (Malinverno and Ryan, 1986). Slab rollback also induced progressive eastward migration of an Alpine-Apenninic orogenic wedge, and other intervening microplates and terranes such as Corsica-Sardinia and Calabria (Patacca et al., 1990).

Drilling results from Ocean Drilling Program (ODP) Leg 107 (Kastens et al., 1988, 1990) show that the Tyrrhenian backarc basin has spread at an average rate of 6 cm/yr since ca. 10 Ma (late Miocene). Core data from the plains of the deepest Vavilov and Marsili Basins (3000–3500 m) (Fig. 1) suggested that ~15,000 km^2 of the Tyrrhenian Sea is floored by oceanic crust, emplaced between the lower-middle Pliocene and Pleistocene. However, both shipborne (Chiappini et al., 2000) and airborne (AGIP, 1981) magnetic surveys of the Tyrrhenian Sea revealed little evidence of the typical oceanic linear anomaly features (Speranza and Chiappini, 2002), and cast doubt on emplacement of oceanic crust in the Vavilov and Marsili Basins (Sartori, 2004). Here we show results from a reanalysis of the shipborne (Chiappini et al., 2000) and airborne (AGIP, 1981) magnetic maps, which provide clear evidence for oceanic-type magnetic anomalies on the floor of the Marsili Basin.

MARSILI BASIN AND SEAMOUNT: CHARACTERISTICS AND AGE

The Marsili Basin is a flat, deep (3500 m on average), roughly elliptical abyssal plain covering ~8000 km^2 (Fig. 1), with a NW-SE–trending major axis (~110 km). Its young oceanic lithosphere is thinner than 30 km (Marani and Trua, 2002) and crustal thickness is ~7 km (Scarascia et al., 1994).

The Marsili Basin is centered on the Marsili seamount, which rises more than 3000 m from the surrounding plain. This huge volcano is strikingly elongated NNE-SSW; it is ~50 km long and 16 km wide (Marani and Trua, 2002).

The formation period of both the Marsili plain and seamount is well constrained. Close to the western basin margin, Site 650 from ODP Leg 107 recovered 602 m of Pleistocene–uppermost Pliocene clastics, mudstones, and nannofossil oozes (Kastens et al., 1990). Although the underlying altered vesicular basalts were drilled for only 32 m, seismic reflection data and the velocity computed for the uppermost acoustic basement suggest that the cored basalts represent the basement, rather than a flow overlying older sediments (e.g., Kastens et al., 1988).

Kastens et al. (1988, 1990) assigned the basal sediments of Site 650 to the Olduvai subchron. However, the occurrence of the planktonic foraminifer *Globorotalia truncatulinoides truncatulinoides* in the lowermost sediment (Kastens et al., 1990) indicates an age 0.1–0.2 m.y. older than the bottom of the Olduvai subchron. Thus, the normal polarity may have arisen from the delayed acquisition of magnetization.

At the Marsili seamount, lava samples dredged on the summit of the volcano yielded K/Ar ages of 0.1–0.2 Ma (Selli et al., 1977). These dates and the normal magnetization retrieved for the volcanic edifice (Faggioni et al., 1995) suggest that the volcano was emplaced over the past 780 k.y., during the Brunhes polarity chron. Marani and Trua (2002) suggested that the Marsili volcano is a superinflated spreading ridge of the surrounding basin, emplaced during slow-spreading plate separation. Age constraints from the top of the volcano, and from the likely older basin floor sector, yield an average half-spreading rate estimate of 1.5–2.0 cm/yr (Kastens et al., 1990).

EVIDENCE OF OCEANIC MAGNETIC ANOMALIES IN THE MARSILI BASIN

Shipborne (Chiappini et al., 2000) and airborne (AGIP, 1981) magnetic surveys of the Marsili Basin mapped very similar anomaly

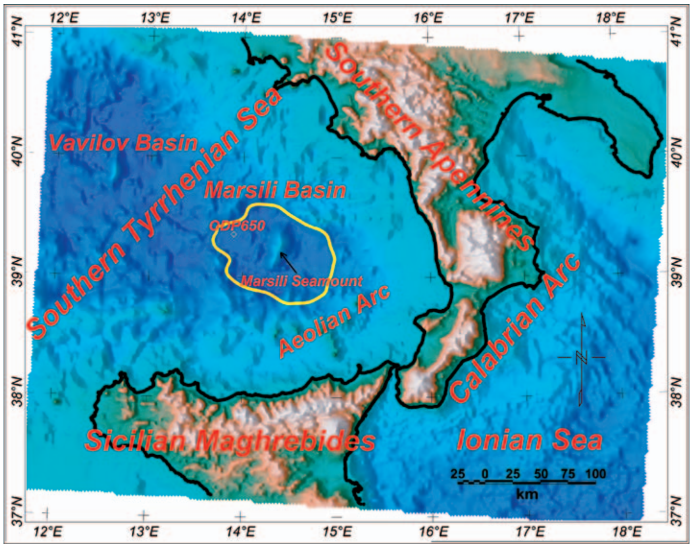

Figure 1. Digital elevation model (from National Geophysical Data Center ETOPO2 at http://www.ngdc.noaa.gov/mgg/fliers/01mgg04.html) of southern Tyrrhenian Sea and surrounding areas. Yellow line encircles flat-lying floor of Marsili Basin. Location of Ocean Drilling Program (ODP) Site 650 is from Kastens et al. (1988).

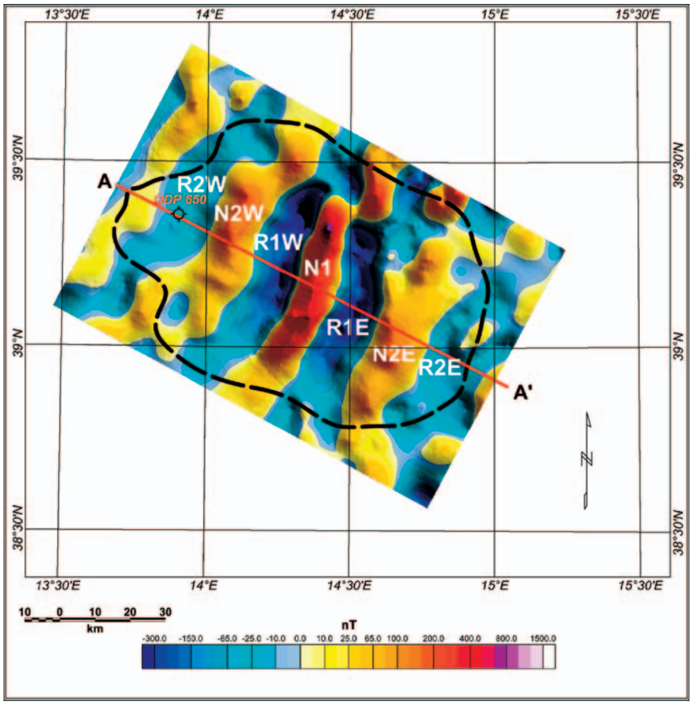

Figure 2. Shipborne magnetic anomaly map of Marsili Basin from Chiappini et al. (2000) filtered to emphasize anomaly components with wavelengths of ~40 km or less along WNW-ESE direction (see text). Dashed line is margin of deep basin floor shown in Figure 1. A–A′ magnetic profile is shown in Figure DR6 (see footnote 1).

patterns. The survey data define a NNE-SSW–elongated, high-intensity (>1000 nT) positive anomaly over the Marsili volcano (see GSA Data Repository Figs. DR1 and DR2[1]) with mostly negative magnetic anomalies in the adjacent flat oceanic floor of the Marsili Basin (e.g., Faggioni et al., 1995).

The magnetic anomaly pattern of the relatively small Marsili Basin is affected by the magnetic signature of the huge basaltic edifice of the Marsili volcano, and the magnetization of the seamounts located off the basin. Fourier analysis of the shipborne profiles across the NNE-SSW axis of the Marsili seamount found that the dominant wavelength component occurred at ~35 km. To enhance this trend in both the ship and airborne data, we designed a composite high-pass, strike-sensitive filter that passed essentially the 40 km and smaller anomaly components trending orthogonally to the NNE-SSW axis of the Marsili seamount. The applications of the composite filters to the ship and airborne magnetic data are given in Figure 2 and Figure DR3 (see footnote 1), respectively.

The filtering on both data sets reveals seven roughly parallel, alternating anomaly stripes centered on the axial positive N1 anomaly in Figures 2 and DR3 (see footnote 1). Superimposed over the Marsili seamount, the largest amplitude linear anomaly N1 essentially reflects the magnetic effects of the seamount. The three flanking pairs of anomaly stripes over the flat-lying floor of the oceanic basin are not associated with morphological features. Thus, we interpret the three paired anomaly stripes to be due to the highly magnetized extrusive basalts of the 2A seismic layer in the uppermost part of the oceanic crust, which is normally <1 km thick (Gee and Kent, 1994; Bazin et al., 2001). The more poorly defined outermost negative R2W and R2E anomaly stripes (Fig. 2) may also include the magnetic influence of the crust and seamounts located off the Marsili Basin and other boundary effects.

After removing the high-frequency component, the residual magnetic anomaly field is dominated by the axial linear positive anomaly centered over the Marsili seamount and broad negative anomalies over the flanking basin components (Figs. DR4 and DR5; see footnote 1). Positive residual anomalies are also observed at the western and eastern margins of the basin. We can relate the filtered and residual magnetic anomalies to a variety of crustal sources because of the apparently layered structure of the oceanic crust (e.g., Dyment and Arkani-Hamed, 1998).

Using the bathymetric data and the filtered magnetic anomalies of Figure 2, we obtained by inversion (Bear et al., 1995) a model of the magnetizations of the 2A basalt layer of the oceanic crust for the Marsili Basin given in Figure 3. We used an array of 2000 × 2000 × 1000 m volumetric cells to model the magnetization variations of the 2A basalt layer. We placed the top of the layer at a depth of ~500 m under the flat basin floor, assuming the uppermost 500 m of crust was made up of virtually nonmagnetic sediments (the log of the ODP Site 650 recovered 602 m of sediments before drilling the basalts; Kastens et al., 1988). The magnetizations in Figure 3 compare well with those actually measured for the oceanic crust (Smith and Banerjee, 1986; Gee and Kent, 1994) and closely mirror the filtered anomalies in Figure 2.

The broad negative residual anomalies overlying the ocean floor flanking the Marsili seamount (Figs. DR4 and DR5; see footnote 1) indicate that the deeper part of the oceanic crust in the Marsili Basin (likely corresponding with the gabbro layer) hosts a predominant reverse-polarity magnetization. The boundaries between the normally and reversely magnetized stripes of oceanic crust may be far from the vertical (Tivey, 1996), so that polarity layering of the oceanic crust at depth can result (Tivey, 1996; Hall and Muzzati, 1999). Moreover, the higher rate of oceanic spreading produces fewer vertical contacts between the oppositely magnetized crustal blocks due to conductive cooling of the crust (Blakely, 1976; Cande and Kent, 1976) and/or the delayed generation of the magnetic minerals over time arising from

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[1] GSA Data Repository item 2006150, Figures DR1–DR6, Tyrrhenian magnetic anomalies, is available online at www.geosociety.org/pubs/ft2006.htm, or on request from editing@geosociety.org or Documents Secretary, GSA, P.O. Box 9140, Boulder, CO 80301, USA.

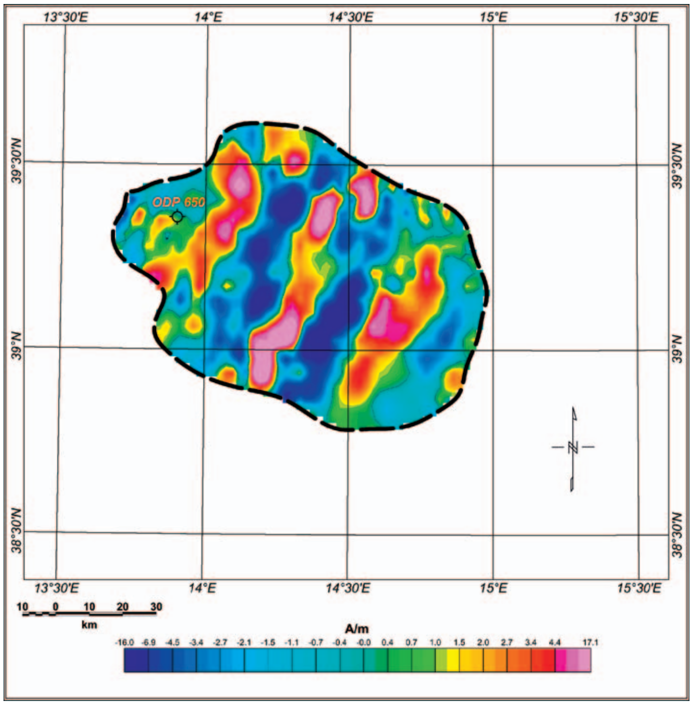

Figure 3. Magnetizations of 2A basalt layer estimated from inversion of filtered anomalies in Figure 2.

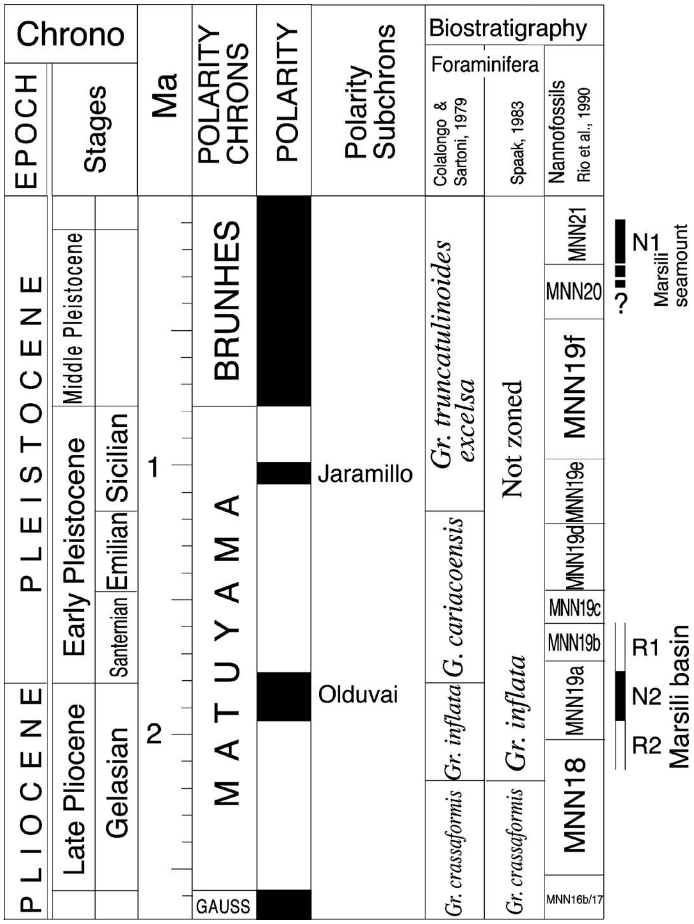

Figure 4. Dating N1, R1, N2, and R2 indicates magnetic anomaly stripes of Figure 2 in context of magnetic polarity (Cande and Kent, 1995) and biostratigraphic (Cita et al., 1999) scales for late Pliocene–Pleistocene development of Marsili Basin and seamount. For simplicity, Brunhes excursions and Cobb Mountain and Reunion events were omitted. (References are provided in the Data Repository, see footnote 1).

hydrothermal alteration processes (Hall and Muzzati, 1999). Thus, polarity layering of the crust in the Marsili Basin may reflect a high spreading rate during basin development.

CORRELATION OF MAGNETIC ANOMALY STRIPES WITH POLARITY CHRONS: IMPLICATIONS FOR THE SPREADING RATE

The K/Ar ages of 0.1–0.2 Ma from the top of the Marsili seamount (Selli et al., 1977) and the normal magnetic polarity of the seamount from our data inversion (Fig. 3) indicate that the axial positive N1 stripe in Figure 2 reflects magnetization acquired during the Brunhes chron, as Faggioni et al. (1995) proposed. Therefore, the adjacent R1W and R1E negative anomaly stripes are inferred to represent oceanic crust emplaced during the Matuyama reverse chron, which agrees with previous suggestions (Faggioni et al., 1995).

Our work, however, also resolves the paired positive N2W and N2E and negative R2W and R2E anomaly stripes (Fig. 2). ODP Site 650 is located at the easternmost edge of the reversely magnetized R2W crustal stripe (Fig. 3). The older sediments overlying the basalts drilled at Site 650 were deposited ~0.1–0.2 m.y. before the Olduvai subchron. Therefore, the normally magnetized oceanic crust of the N2W stripe should be synchronous and/or slightly younger than the lowermost sediments cored at Site 650, and older than the upper part of the Matuyama chron (C1r) of the adjacent R1W stripe. We conclude that the normal-polarity N2W stripe and its complementary symmetrical N2E stripe were produced by oceanic crust emplaced during the 1.77–1.95 Ma Olduvai subchron (Cande and Kent, 1995).

The crustal stripes producing the N2W and N2E anomalies have an average width of 17 ± 1 km (Fig. 3). Therefore, their correlation with the Olduvai subchron, which lasted 0.18 m.y., translates into a half-spreading rate of 9–10 cm/yr for the Marsili Basin. This result is consistent with the high spreading rate suggested by the in-depth polarity layering of the oceanic crust. This spreading rate is significantly greater than the normal geologically averaged 2–7 cm/yr spreading rates of backarc basins (Taylor, 1995), although the Lau Basin in the western Pacific is currently expanding at the rate of 16 cm/yr (Bevis et al., 1995). In general, the spreading rate of the Marsili Basin is comparable to the fastest known full spreading rate of 18–21 cm/yr calculated for the middle Miocene Cocos-Pacific plate boundary (Wilson, 1996).

Extrapolating the 9–10 cm/yr rate evaluated for the N2W and N2E stripes to the younger crust may suggest that 2 93-km-wide reversely magnetized crust stripes formed during chron C1r (0.78–1.77 Ma). The crustal components of the negative R1W and R1E anomalies, by contrast, are on average only 16 km wide (Fig. 3). Thus, we conclude that the spreading of the Marsili Basin occurred through two episodes with completely different spreading rates.

In the chronographic context of Figure 4, ultra-rapid spreading of ~19 cm/yr generated most of the basin during the uppermost Pliocene–lower Pleistocene, between ca. 2.1 and 1.6 Ma. A dramatic slowdown or perhaps total cessation of spreading followed for a period of ~1 m.y. In the mid-late Pleistocene (after 0.78 Ma), plate spreading resumed, but at a much reduced rate so that magmatic superinflation produced the Marsili volcano. The deeper gabbroic layer of the oceanic crust of the Marsili Basin was produced rapidly, partly during the normal polarity Olduvai subchron. However, it subsequently cooled and acquired magnetization, mostly during the reverse-polarity upper Matuyama chron (C1r), with minimal Olduvai subchron imprinting due to the extremely high spreading rate.

CONCLUSIONS

Seven alternating magnetic anomaly stripes in the Marsili Basin provide insight on the timing of oceanic spreading in this sector of the southern Tyrrhenian Sea. Age constraints from both the Marsili Basin and seamount were used to correlate the individual anomalies to the global magnetic polarity time scale. The two N2 positive anomaly stripes from the peripheral parts of the basin can be unambiguously related to oceanic crust emplaced during the Olduvai subchron. The 17 km mean width of the inferred normally magnetized crustal stripes and the 0.18 m.y. duration of the Olduvai subchron imply that most of the Marsili Basin opened over a few hundred thousand years during the Pliocene-Pleistocene transition, ca. 1.6–2.1 Ma, with a full spreading rate of ~19 cm/yr. This spreading rate is similar to the fastest known full spreading rate of 18–21 cm/yr calculated for the Cocos-Pacific plate boundary (e.g., Wilson, 1996). However, it is significantly greater than geologically averaged values from all other backarc basins. This ultrarapid spreading stopped or decreased dramatically after the early Pleistocene. After 0.78 Ma slow spreading resumed during the Brunhes chron, giving rise to the huge Marsili seamount, a superinflated spreading ridge.

These new data reveal a strikingly discontinuous geodynamic evolution of the Tyrrhenian Sea. In the Marsili Basin, a relatively long period (~1 m.y.) of tectonic quiescence separated the rapid spreading and magmatic inflation episodes. The accelerated development of the southern Tyrrhenian Sea may also be related to an ultra-rapid early Pleistocene paleomagnetic rotation of the southern Apennine belt (Mattei et al., 2004). However, the age uncertainty of the paleomagnetic rotation also allows the ultra-rapid events in the southern Apennines and Tyrrhenian Sea to be partly coeval and thus genetically related.

The development of the Marsili Basin between 1.6 and 2.1 Ma at the full spreading rate of ~19 cm/yr required a similar coeval drift velocity for the Calabrian block, which migrated southeastward ahead of the backarc spreading basin. However, the Tyrrhenian Sea spreading and the migration of the orogenic wedge since late Miocene time developed at the average rate of 6 cm/yr (Patacca et al., 1990). We conclude that the evolution of the Tyrrhenian arc-backarc system occurred through a succession of ultra-rapid and slow spreading events separated by periods of relative quiescence.

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REFERENCES CITED

AGIP spa, 1981, Carta Magnetica-Anomalie del Campo magnetico residuo: Milanese, Italy, S. Donato, scale 1/500,000.

Bazin, S., Harding, A.J., Kent, G.M., Orcutt, J.A., Tong, C.H., Pye, J.W., Singh, S.C., Barton, P.J., Sinha, M.C., White, R.S., Hobbs, R.W., and Van Avendonk, H.J.A., 2001, Three-dimensional shallow crustal emplacement at the 9°03′N overlapping spreading centre on the East Pacific Rise: Correlations between magnetization and tomographic images: Journal of Geophysical Research, v. 106, no. B8, p. 16,101–16,117, doi: 10.1029/2001JB000371.

Bear, G.W., Al-Shukri, H.J., and Rudman, A.J., 1995, Linear inversion of gravity data for 3-D density distribution: Geophysics, v. 60, p. 1354–1364, doi: 10.1190/1.1443871.

Bevis, M., Taylor, F.W., Schutz, B.E., Recy, J., Isacks, B.L., Helu, S., Singh, R., Kendrick, E., Stowell, J., Taylor, B., and Calmantli, S., 1995, Geodetic observations of very rapid convergence and back-arc extension at the Tonga arc: Nature, v. 374, p. 249, doi: 10.1038/374249a0.

Blakely, R.J., 1976, An age-dependent, two layer model for marine magnetic anomalies: Geophysics of the Pacific Ocean Basin and its Margins, v. 19, p. 227–234.

Cande, S.C., and Kent, D.V., 1976, Constraints imposed by the shape of marine magnetic anomalies on the magnetic source: Journal of Geophysical Research, v. 81, p. 4157–4162.

Cande, S.C., and Kent, D.V., 1995, Revised calibration of the geomagnetic polarity time scale for the Late Cretaceous and Cenozoic: Journal of Geophysical Research, v. 100, p. 6093–6095.

Chiappini, M., Meloni, A., Boschi, E., Faggioni, O., Beverini, N., Carmisciano, C., and Marson, I., 2000, Onshore-offshore integrated shaded relief magnetic anomaly map at sea level of Italy and surrounding areas: Annali di Geofisica, v. 43, no. 5, scale 1/1,500,000.

Cita, M.B., Rio, D., and Sprovieri, R., 1999, The Pliocene series: Chronology of the type Mediterranean record and standard chronostratigraphy, *in* Wrenn, J.H., and Suc, J.P., eds., Palynology, climate and sequence stratigraphy of the Pliocene: American Association of Stratigraphic Palynologists, Special Volume, p. 49–63.

Dyment, J., and Arkani-Hamed, J., 1998, Contribution of lithospheric remanent magnetization to satellite magnetic anomalies over the world's oceans: Journal of Geophysical Research, v. 103, no. B7, p. 15,423–15,441, doi: 10.1029/97JB03574.

Faggioni, O., Pinna, E., Savelli, C., and Schreider, A.A., 1995, Geomagnetism and age study of Tyrrhenian seamounts: Geophysical Journal International, v. 123, p. 915–930.

Gee, J., and Kent, D.V., 1994, Variations in layer 2A thickness and the origin of the central anomaly magnetic high: Geophysical Research Letters, v. 21, p. 297–300, doi: 10.1029/93GL03422.

Hall, J.M., and Muzzati, A., 1999, Delayed magnetization of the deeper kilometre of oceanic crust at Deep Sea Drilling Project Site 504: Journal of Geophysical Research, v. 104, p. 12,843–12,851, doi: 10.1029/1999JB900105.

Kastens, K.A., and 21 others, 1988, ODP Leg 107 in the Tyrrhenian Sea: Insights into passive margin and back arc basin evolution: Geological Society of America Bulletin, v. 100, p. 1140–1156.

Kastens, K.A., and 20 others, 1990, Proceedings of the Ocean Drilling Program, Scientific results, Volume 107: College Station, Texas, Ocean Drilling Program, 722 p.

Malinverno, A., and Ryan, W.B.F., 1986, Extension in the Tyrrhenian Sea and shortening in the Apennines as result of arc migration driven by sinking of the lithosphere: Tectonics, v. 5, p. 227–245.

Marani, M., and Trua, T., 2002, Thermal constriction and Tera faulting at the origin of a super-inflated spreading centre: The Marsili volcano, Tyrrhenian Sea: Journal of Geophysical Research, Solid Earth, v. 107, no. B9, p. EPM 3–1.

Mattei, M., Petrocelli, V., Lacava, D., and Schiattarella, M., 2004, Geodynamic implications of Pleistocene ultrarapid vertical-axis rotations in the Southern Apennines, Italy: Geology, v. 32, p. 789–792, doi: 10.1130/G20552.1.

Patacca, E., Sartori, R., and Scandone, P., 1990, Tyrrhenian basin and Apenninic arcs: Kinematic relations since late Tortonian times: Memorie della Società Geologica Italiana, v. 45, p. 425–451.

Sartori, R., 2004, The Tyrrhenian back-arc basin and subduction of the Ionian lithosphere: Episodes, v. 26, p. 217–221.

Scarascia, A., Lozej, A., and Cassinis, R., 1994, Crustal structures of the Ligurian, Tyrrhenian and Ionian seas and adjacent onshore areas interpreted from wide-angle seismic profiles: Bollettino di Geofisica Teorica Applicata, v. 36, no. 141–144, p. 5–21.

Selli, R., Lucchini, F., Rossi, P.L., Savelli, C., and Del Monte, M., 1977, Dati geologici, petrolchimici e radiometrici sui vulcani centro-tirrenici: Giornale di Geologia, v. 42, p. 221–246.

Smith, G.M., and Banerjee, S.K., 1986, Magnetic structure of the upper kilometer of the marine crust at Deep Sea Drilling Project Site 504B, eastern Pacific Ocean: Journal of Geophysical Research, v. 91, no. B10, p. 10,337–10,354.

Speranza, F., and Chiappini, M., 2002, Thick-skinned tectonics in the external Apennines, Italy: New evidence from magnetic anomaly analysis: Journal of Geophysical Research, Solid Earth, v. 107, no. B11, p. 2290, doi: 10.1029/2000JB000027.

Taylor, B., 1995, Backarc basins: Tectonics and magmatism: New York, Plenum, 524 p.

Tivey, M., 1996, The vertical magnetic structure of ocean crust determined from near bottom magnetic field measurements: Journal of Geophysical Research, v. 101, p. 20,275–20,296.

Wilson, D.S., 1996, Fastest known spreading on the Miocene Cocos-Pacific plate boundary: Geophysical Research Letters, v. 23, p. 3003–3006, doi: 10.1029/96GL02893.

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