The Miocene Saint-Florent Basin in northern Corsica: stratigraphy, sedimentology, and tectonic implications

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ABSTRACT

Late early–early middle Miocene (Burdigalian–Langhian) time on the island of Corsica (western Mediterranean) was characterized by a combination of (i) postcollisional structural inversion of the main boundary thrust system between the Alpine orogenic wedge and the foreland, (ii) eustatic sealevel rise and (iii) subsidence related to the development of the Ligurian-Provençal basin. These processes created the accommodation for a distinctive continental to shallow-marine sedimentary succession along narrow and elongated basins. Much of these deposits have been eroded and presently only a few scattered outcrop areas remain, most notably at Saint-Florent and Francardo. The Burdigalian–Langhian sedimentary succession at Saint-Florent is composed of three distinguishing detrital components: (i) silicilastic detritus derived from erosion of the nearby Alpine orogenic wedge, (ii) carbonate intrabasinal detritus (bioclasts of shallow-marine and pelagic organisms), and (iii) silicilastic detritus derived from Hercynian-age foreland terraines. The basal deposits (Fium Albino Formation) are fluvial and composed of Alpine-derived detritus, with subordinate foreland-derived volcanic detritus. All three detrital components are present in the middle portion of the succession (Torra and Monte Sant'Angelo Formations), which is characterized by thin transitional deposits evolving vertically into fully marine deposits, although the carbonate intrabasinal component is predominant. The Monte Sant'Angelo Formation is characteristically dominated by the deposits of large gravel and sandwaves, possibly the result of current amplification in narrow seaways that developed between the foreland and the tectonically collapsing Alpine orogenic wedge. The laterally equivalent Saint-Florent conglomerate is composed of clasts derived from the late Permian Cinto volcanic district within the foreland. The uppermost unit (Farinole Formation) is dominated by bioclasts of pelagic organisms. The Saint-Florent succession was deposited during the last phase of the counterclockwise rotation of the Corsica–Sardinia–Calabria continental block and the resulting development of the Provençal oceanic basin. The succession sits at the paleogeographic boundary between the Alpine orogenic wedge (to the east), its foreland (to the west), and the Ligurian-Provençal basin (to the northwest). Abrupt compositional changes in the succession resulted from the complex, varying interplay of post-collisional extensional tectonism, eustacy and competing drainage systems.

INTRODUCTION

The islands of Corsica and Sardinia and the surrounding continental shelves are a fragment of European–Iberian lithosphere, which, together with Calabria, rifted off the southern continental margin of Europe during Oligocene–Miocene time and drifted southeastward by counterclockwise rotation (Alvarez, 1972, 1976; Alvarez et al., 1974) of about 45°. Malinverno & Ryan (1986) first proposed that rifting of Corsica–Sardinia–Calabria from Europe was the result of upper plate extension related to slab roll-back along the Ionian subduction zone. Although the exact timing is still far from precisely known (see Speranza, 1999, for a discussion), rifting started as early as the early Oligocene (Vially & Trèmolières, 1996) whereas drifting occurred between ca. 22 and 15 Ma (Ferrandini et al., 2003). Rifting reactivated Pyrenean compressional structures of the Late Cretaceous–Early Eocene Provençal foldbelt (Vially & Trèmolières, 1996). Subsequent rifting of Calabria from Corsica–Sardinia during the late Miocene isolated Corsica–Sardinia between the Ligurian-Provençal basin to the northwest and the Tyrrhenian Sea to the east (see Cavazza et al., 2004a, b, for a review) (Fig. 1).

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As the Corsica–Sardinia–Calabria block rifted away from Europe, the progressively widening Ligurian-Provençal basin was flooded by marine water. Situated between the extending Provençal margin of Europe and a fragment of the collapsing western Alpine orogenic wedge (Fig. 2), the basin was filled with a diverse assemblage of lithofacies that include both carbonate and siliciclastic components. Remnants of the older portions of this basin fill are preserved in Corsica and are referred to as the Saint-Florent Basin (Fig. 1). Most striking are stacked deposits of very large-scale gravelly and sandy bedforms and intercalated sediment-gravity flows. The tectonic setting of the Saint-Florent Basin between the extending orogenic wedge and the foreland produced a narrow paleostrait.
where oceanic (probably tidal) currents were strongly amplified. The purpose of this paper is to document the Saint-Florent Basin fill and to assess its potential significance for understanding the history of rifting and orogenic collapse in northern Corsica, as well as its implications for the Oligocene–Neogene tectonic evolution of the western Mediterranean. Our results should be useful for evaluating other Neogene basins in the Mediterranean region, where upper plate (or back-arc) rifting events in response to subduction zone retreat (Royden, 1993) are widespread (e.g. Jolivet & Faccenna, 2000; Cavazza et al., 2004a).

**GEOLOGIC SETTING**

Corsica is divided into two distinct geologic terranes (Fig. 1). The western part of the island is characterized by Carboniferous–Permian granitoid rocks and acid volcanic rocks related to the Hercynian orogeny, with Precambrian–middle Palaeozoic metamorphic host rocks and scattered outcrops of Palaeozoic sedimentary rocks. The northeastern portion of the island (‘Alpine Corsica’) is a nappe stack dominated by Jurassic oceanic crust and its sedimentary cover (Durand-Delga, 1978, 1984). These rocks were metamorphosed at high pressures and low temperatures during the Alpine orogeny (see Gibbons et al., 1986, and Caron, 1994, for reviews). Although data pointing to a metamorphic event older than Late Eocene are scarce and questionable (see Brunet et al., 2000, for a discussion), the compressional development of the Alpine orogenic wedge traditionally has been considered to span Late Cretaceous to Eocene time. Major thrusting in Corsica occurred during the Eocene, producing much of the documented nappe stacking and metamorphism (Caron, 1994). Local remnants of an Eocene foreland basin (e.g. the autochthonous clastic units of the Balagne region of Fig. 1) are present between the Hercynian crystalline basement complex and Alpine Corsica (Nardi et al., 1978; Jourdan, 1988).

From north to south, Miocene rocks are present in Corsica only in four relatively small areas: Saint-Florent, Francardo, the Aleria plain and Bonifacio (Fig. 1) (e.g. Orszag-Sperber & Pilot, 1976; Alesandri et al., 1977; Groupe Francais d’Etude du Neogene, 1989; Cabells et al., 1994; Ferrandini et al., 1998, 2002, 2003; Loye-Pilot et al., 2004; Fellin et al., 2005). The successions at Saint-Florent, Francardo and Bonifacio are thin and cover almost precisely the same time span (Burdigalian–Langhian). The sedimentary succession of the Aleria Plain spans Burdigalian to Pliocene time and represents the onland portion of the > 8 km–thick offshore Corsica basin to the east (Maufrett et al., 1999), which developed during rifting of Calabria from the Corsica–Sardinia block. This paper describes and interprets the Miocene succession at Saint-Florent within the framework of the post-collisional evolution of Corsica and the western Mediterranean.

**STRUCTURE OF THE BALAGNE–BASTIA TRAVERSE**

The Saint-Florent Basin lies along the Balagne–Bastia traverse (Fig. 2), which arguably constitutes the best-studied part of Corsica. From the west to the east along the traverse, the major tectonostatigraphic domains include the following: (i) The autochthonous, relatively undeformed basement complex of the European–Iberian plate (‘Hercynian Corsica’), made of Carboniferous to early Permian granitoid rocks, their pre-Hercynian metamorphic host rocks and Permian volcanic rocks (Mt. Cinto volcanic district); (ii) remnants of the Balagne foreland basin and thrust belt with middle Eocene turbidites, overthrust by W-verging nappes of Jurassic ophiolites and their Cretaceous–Eocene, unmetamorphosed sedimentary cover; (iii) the highly deformed parts of the European continental margin, including the dome-shaped Tenda massif of Hercynian granitoid and Palaeozoic volcanic rocks, and slices of crystalline basement of the Olette–Serra di Pigno–Farinole crystalline unit interspersed in the ‘Schistes Lustrés’ nappe; (iv) the Nebbio nappe, composed of rock units similar to those of the Balagne nappe and similarly unmetamorphosed (Dallan & Puccinelli, 1986, 1995). The Nebbio
nappe overlies the previously deformed and metamorphosed ‘Schistes Lustrés’ nappe and is unconformably overlain by the Miocene Saint-Florent succession. The Nebbio, Balagne and Macinaggio (NE Corsica; not shown in Figs 1 and 2) nappes are the uppermost, unmetamorphosed tectonic units of the Alpine orogenic wedge (‘nappe superieure’). The basal thrust faults of these nappes, emplaced during the Eocene (e.g. Durand-Delga, 1984), were tectonically inverted as low-angle detachment faults during the late Oligocene (Daniel et al., 1996); (v) the ‘Schistes Lustrés’ nappe is composed of ophiolitic sequences covered by metamorphosed marine sedimentary rocks. It contains eclogitic associations (Caron et al., 1981; Lahondère, 1988; Fournier et al., 1991) of Late Cretaceous age (Lahondère and Guerrot, 1997) that were retrogressed to blueschist facies (Gibbons et al., 1986; Lahondère, 1988; Waters, 1989) during Eocene time (Brunet et al., 2000). Westward emplacement of the ‘Schistes Lustrés’ nappe onto the Hercynian basement and its Eocene sedimentary cover ended in late Eocene time (Durand-Delga, 1978).

**STRUCTURAL SETTING OF THE SAINT-FLORENT BASIN**

From the structural viewpoint, the Saint-Florent region is dominated by the East Tènda shear zone (Figs 1–3), a major Alpine top-to-the-west thrust fault that was reactivated as a detachment fault since the Oligocene (Waters, 1990; Fournier et al., 1991; Brunet et al., 2000) and then cut by high-angle normal faults during late-early to early-middle Miocene time (Fellin et al., 2005). The dome-shaped
Tenda massif (Fig. 2) is the footwall of the detachment fault and is mostly composed of Hercynian granitoid rocks and Palaeozoic volcanics. These rocks were recrystallized in blueschist conditions during west-vergent Alpine thrusting at peak pressures/temperatures of 0.5 GPa/300 °C to 1.1 GPa/500 °C (minimum burial depth of 27 km) and later overprinted by ductile extension in greenschist facies along its eastern border during Oligocene time (Waters, 1990; Fournier et al., 1991; Brunet et al., 2000; Tribuzio & Giacomini, 2002).

In its presently outcropping extent, the Saint-Florent Basin is bounded on the east by an N-striking, W-dipping normal fault system (Fig. 3) active during Burdigalian–Langhian time (i.e. during deposition of the Saint-Florent succession), based on the integrated results of structural analysis and apatite–zircon fission-track analyses (Fellin et al., 2005). The western boundary of the basin is a N-to–NW-striking fault partially obscured by Quaternary alluvium of the Aliso River. According to Fellin et al. (2005), this high-angle normal fault displaces the East Tenda shear zone and juxtaposes terrains exhumed at different ages, thus substantiating the notion that high-angle normal faulting postdated and was unrelated to detachment faulting along the East Tenda shear. Syndepositional tectonism along minor associated normal faults is documented also by widespread soft-sediment deformation and growth faults affecting the basin fill, both particularly common along the northern end of the basin.

Julivet et al. (1990) and Fournier et al. (1991) interpreted the Saint-Florent Basin as the result of extension along the East Tenda shear zone, in spite of the rather large time gap between the youngest ages in the exhumed metamorphic rocks (33–32 Ma), and the age of the base of the basin fill at about 19–18 Ma. Brunet et al. (2000) reported Ar/Ar ages from the Tenda massif as young as 25 Ma, thus reducing the time gap to 6–7 Ma.

Structural and fission-track analyses indicate that during the early late Miocene (Tortonian) the Saint-Florent Basin was uplifted, acquiring its open synclinal shape. Kinematic indicators suggest that uplift occurred through right-lateral transpressional reactivation of the boundary faults under a NNE-oriented maximum compressional axis (Fellin et al., 2005). Partial annealing of apatites from the basement beneath the basin indicates that significant erosion occurred and that the basin fill was originally much thicker.

**STRATIGRAPHY AND SEDIMENTOLOGY OF THE SAINT-FLORENT SUCCESSION**

The region of Saint-Florent is characterized by a thick succession of Neogene sedimentary rocks unconformably overlying the Nebbio tectonic unit (Fig. 3) (Dallan & Puccinelli, 1986, 1995; Dallan et al., 1988; Rossi et al., 1994). The Saint-Florent sedimentary succession delineates a large, open, north-plunging syncline. The lower contact of the Saint-Florent succession is exposed locally along the northern boundaries of the outcrop area (Fig. 3). Ferrandini et al. (1998, Fig. 2) and Fellin et al. (2005, Fig. 6) reported total thicknesses of 550 and ca. 600 m, respectively. Our measured stratigraphic sections, integrated from observations throughout the Saint-Florent Basin, point to a maximum total thickness of about 400 m for the preserved basin fill, in agreement with Durand-Delga (1978), Rossi et al. (1994) and Dallan & Puccinelli (1995).

From the bottom to the top, the succession begins with a thin (maximum 60 m), discontinuous unit named the Fium Albino Formation by Ferrandini et al. (1998) (Fig. 4). This unit crops out locally in erosional depressions cut into the underlying Nebbio nappe (Fellin et al., 2005) and is composed of pebble conglomerate and very coarse- to coarse–grained sandstone that were interpreted by Ferrandini et al. (1998) as the deposits of an alluvial/fluvial environment. Conglomerate clasts were mostly derived from the Alpine orogenic wedge, whereas sandstone lithic fragments have a more varied provenance (see following section). No direct age determination exists for this unit.

Poor outcrops hinder unambiguous interpretation of the contact between the Fium Albino Formation and the overlying Torra Formation (Fig. 4); Ferrandini et al. (1998) inferred it to be an angular unconformity, whereas Fellin et al. (2005) interpreted the contact as conformable. Most likely, local minor unconformities are the result of syndepositional deformation, the effects of which are also visible in the overlying calcarenites of the Monte Sant‘Angelo Formation. In most places the Torra Formation directly

![Fig. 4. Generalized stratigraphic section for the Saint-Florent area. The entire succession is here considered as conformable, with only minor unconformities locally present between the fluvial deposits of the Fium Albino Formation and the transitional ones of the Torra Formation Data sources: Orszag-Sperber & Pilot, 1976, Loyo-Pilot & Magné, 1987, Ferrandini et al. (1998), Fellin et al. (2005), and unpublished data. Time scale after Berggren et al. (1995).](image-url)
overlies the rock units of the Nebbio nappe. The Torra Formation is about 50–60 m thick (Fig. 5c) and it is made of massive medium- to coarse-grained sandstone and pebble conglomerate with abundant oyster and pectinid shells, echinoids, bryozoans and Skolithos burrows. The upper portion of the unit comprises thick carbonate-dominated beds rich in corallinaceous algae and is transitional to the overlying Monte Sant’Angelo Formation.

Reaching ca. 250 m in thickness, the Monte Sant’Angelo Formation represents the bulk of the Saint-Florent sedimentary succession (Figs 4 and 5). This unit conformably overlies the Torra Formation and is overlain with a sharp contact by the Farinole Formation. The Saint-Florent conglomerate is a local coarse-grained facies that cuts erosively into and is partly a lateral equivalent of the Monte Sant’Angelo Formation. The age of the Monte Sant’Angelo Formation is late Burdigalian–early Langhian, based on foraminifera and nanoplankton biostratigraphy (Ferrandini et al., 1998; Fellin et al., 2005) and on magnetostratigraphy (Vigliotti & Kent, 1990; Ferrandini et al., 2003).

The Monte Sant’Angelo Formation is dominated by bioclastic detritus (corallinean red algae, bivalves, bryozoans and foraminifera) although varying amounts of siliciclastic detritus are present throughout (on average 15% of sand-sized grains consist of noncarbonate extra-basinal detritus). The extra-basinal clasts are dominated by
material derived from the Alpine orogenic wedge. Rhodoliths (i.e. nodules of red corallinacean algae concentrically encrusted) are very common (Orszag-Sperber et al., 1977), often concentrated in 1.0–2.5-m-thick beds intercalated with large-scale cross-bedded calcarenites virtually devoid of rhodoliths (Figs 6, 7 and 8a). Amalgamated beds are common in both the rhodolitic and calcarenitic facies. Rhodoliths reach a maximum diameter of 12 cm (average 4–5 cm) and have either bioclastic or siliciclastic nuclei (Fig. 8b). The rhodolitic beds are tabular, laterally continuous and their bases are flat and not obviously erosional (Fig. 8c). Although some of these beds contain weakly developed horizontal stratification, most are not internally stratified (Fig. 8c, d). These beds exhibit distinctive grain-size and grain-compositional variations. In beds with abundant rhodoliths, the denser siliciclastic clasts, either free or as cores within rhodoliths, are concentrated in the lower part of the bed, whereas densely packed rhodoliths are concentrated in the upper part (Fig. 8e). Other beds are composed primarily of siliciclastic clasts. In either case, the siliciclastic clasts are usually normally graded (Fig. 8f); in a few cases we observed inverse grading (Fig. 8g) and inverse to normal grading (Fig. 8g) among the siliciclastic clasts. Out-sized clasts are present in many of these beds, floating in an unsorted and unstratified matrix.

Large-scale sets of cross-bedded calcarenite are up to 5 m thick (Figs 5, 6 and 8a, c) and are made of well–sorted, very coarse- to coarse-grained intrabasinal bioclastic sandstone (biocarparite and biomicroparite) with a subordinate siliciclastic extrabasinal detrital component. Rhodoliths are scarce in cross-bedded calcarenites and are always concentrated at the base of the foresets. At the outcrop scale the bounding surfaces of the cross-sets are mostly planar; only in a few cases were curved bounding surfaces visible (Fig. 8a). Reactivation surfaces are common. Foreset dips are commonly greater than 20°. The internal geometry of the large-scale sets of cross-bedded calcarenite is
forests. Foresets are commonly burrowed, particularly near the bottom, with *Thalassinoides*, *Ophiomorpha* and *Macaronichnus* as the most common ichnogenera (Fig. 8h). *Spreiten* structures are also present. Cosets commonly are bounded by laterally extensive, sheet-like erosional surfaces. No mud drapes were detected.

Well-rounded, well-organized pebble-to-cobble conglomerate made of rhyolitic ignimbrite clasts is interbedded with the Monte Sant’Angelo Formation along the coast north of the town of Saint-Florent (Fig. 3) and also crops out just east of the town in the core of the syncline. These conglomerates have been referred to as *poudingues de Fortino* (Hollande, 1917), *poudingues de Saint-Florent* (Maury, 1909), *Formation de Saint-Florent* (Ferrandini et al., 1998) and *St Florent conglomerate* (Fellin et al., 2005). The rhyolite (ash-flow tuff) clasts are distinctive in colour, texture and composition, and were derived from the Permian volcanic complex of Monte Cinto, located about 30 km to the southwest of the study area. This rock type is so distinctive that derivation of the conglomerate clasts from the Monte Cinto complex was first proposed by Reynaud in 1833. Possibly because of its siliciclastic nature – markedly different from the carbonate-dominated majority of the Saint-Florent succession – this conglomerate has long been considered to postdate the deposition of the carbonate succession, in spite of several outcrops along the coast north of Saint-Florent where the conglomerate is interbedded with the carbonates. In addition, in the coastal outcrops just east of the town of Saint-Florent, these conglomerates have the same structural orientation as the Monte Sant’Angelo Formation (Durand-Delga, 1978, p. 158; Dallan et al., 1988; Dallan & Puccinelli, 1995, p. 51), suggesting that the two units are laterally equivalent to each other. Therefore, based on our observations and those of Fellin et al. (2005), we propose to discontinue the use of the terms *poudingues de Fortino*, *poudingues de Saint-Florent*, and *Formation de Saint-Florent* for these characteristically well-rounded conglomerates. Because they are intimately associated with and laterally equivalent to the thick carbonate layers of the uppermost Monte Sant’Angelo Formation, we consider this distinctive unit as an informal member of the Monte Sant’Angelo Formation with the name ‘Saint-Florent conglomerate’. In spite of the lack of direct age control, we agree with Fellin et al. (2005) that the interfingering with the upper Monte Sant’Angelo Formation points to an early Langhian age for this conglomerate.

The Saint-Florent succession is capped by foraminifera-rich marls and calcarenites of the Farinole Formation (Fig. 4). This unit is about 80 m thick and crops out only in the northeastern part of the study area (Fig. 3). Its age is late Langhian–middle Serravallian, based on nannoplankton and foraminifera biostratigraphy (LoyÈ-Pilot & MagnÈ, 1987; Ferrandini et al., 1998; Fellin et al., 2005) and magnetostratigraphy (Ferrandini et al., 2003). Compared with the underlying Monte Sant’Angelo Formation, the Farinole Formation was deposited in a deeper marine environment, as indicated by the presence of a pelagic
Fig. 8. Field photographs of sedimentary facies within the Monte Sant’Angelo Formation of the Saint-Florent sedimentary succession. (a) Large-scale cross-bedding in sandwave deposits (base of Campu Maggiore stratigraphic section). (b) Detail of a massive rhodolith bed with ophiolitic clasts at the core of several rhodoliths (Strutta Rau stratigraphic section). (c) Alternating massive rhodolith beds (on) and cross-bedded sandwave deposits (sw); cross-beds dip away from observer. Note the nonerosive contacts between depositional units (Strutta Rau stratigraphic section). (d) Structureless, 3-m-thick bed of rhodoliths. (Campu Maggiore stratigraphic section). (e) Bed containing both siliciclastic and rhodolithic grains, with siliciclastic grains concentrated in lower part and showing subtle inverse grading (Punta di Saeta). Scale bar is 10 cm long. (f) Bed of mixed rhodoliths and siliciclastic grains, with inverse to normal grading (Campu Maggiore). Scale bar is 10 cm long. (g) Bed containing a mixture of rhodoliths and siliciclastic clasts. Note the absence of stratification, and the upward fining size of the siliciclastic components (Strutta Rau). (h) Ichnofacies (Spreiten structures, Thalassinoideas and Macaronichnus) of the Monte Sant’Angelo Formation (Campu Maggiore).
microflora and microfauna. A few channelized conglomerate bodies composed of rhyolite clasts are present.

One hundred eighty-six paleocurrent indicators were measured throughout the stratigraphic thickness of the Saint-Florent Basin fill over its entire outcropping area. In order of decreasing abundance, measured paleocurrent indicators include (i) sandwave foresets, (ii) imbricated clasts and (iii) axes of trough cross-beds. Outcrops of the Fium Albino Formation are too poor for measuring unequivocal paleocurrent indicators. Despite some scatter, paleocurrent indicators of the Monte Sant’Angelo Formation yield an average paleoflow direction towards the NNE (Fig. 3). Foresets dipping in opposite directions at any one location are rare. The Saint-Florent conglomerate also shows well-defined paleocurrent directions toward the NNE. Paleocurrent indicators are absent in the Farinole Formation.

**PALAEOENVIRONMENTAL INTERPRETATION OF THE SAINT-FLORENT SUCCESSION**

Overall, the sedimentological and palaeontological characteristics of the Saint-Florent Basin fill point to a progressive deepening, from the fluvial–alluvial deposits of the Fium Albino Formation, through the transitional deposits of the Torra Formation, to the shallow marine deposits of the Monte Sant’Angelo Formation and the pelagic sediments of the Farinole Formation.

The macropalaeontological association of the Monte Sant’Angelo Formation, which constitutes the bulk of the Saint-Florent sedimentary succession, comprises coralline red algae, bivalves, bryozoans, foraminifera and echinoderms, and suggests a shallow-water warm-temperate association (Flügel, 2004), in agreement with the approximate palaeolatitude of Corsica of 37–38°N during Burdigalian–Langhian time (e.g. Dercourt et al., 1993). The micropalaeontological association (flora and fauna) points to an outer shelf depth range; yet the coarser-grained bioclastic detritus was derived from a littoral environment (Ferrandini et al., 1998). Along with the broken and abraded nature of the larger bioclasts, this indicates that reworking and redeposition of shallow marine biogenic detritus into a deeper environment were common in the Saint-Florent Basin.

The Monte Sant’Angelo Formation has been interpreted as beach deposits (e.g. Durand-Delga, 1978; Orszag-Sperber, 1980), with the large-scale cross-beded deposits representing beach ridges. However, more recent sedimentological analyses led to the interpretation of the cross-beded facies as the deposits of large sandwaves (Cavazza & DeCelles, 1995, 1996; Ferrandini et al., 1998). The common interbedding of sandwave deposits and rhodolith beds is interpreted here as the result of sediment-gravity flows reworking the rhodoliths and other bioclastic detritus from a shallow nearshore environment into a deeper marine realm swept by intense, mostly unidirectional currents. Periodic deposition of thick rhodolith beds can be viewed within the framework of the intense syndepositional tectonism documented by growth faults and soft-sediment deformation (Fig. 9).

A key component of our environmental reconstruction is the sediment-gravity flow interpretation of the rhodolith and mixed rhodolith–siliciclastic beds. We interpret these beds as the deposits of concentrated and hyperconcentrated density flows (genus Mulder & Alexander, 2001). Several aspects of these beds require this interpretation: (1) The general absence of internal traction–current stratification (Fig. 8d) indicates deposition from turbulent high-density/high-concentration flows in which typical traction-current bedforms were suppressed, probably because of a high rate of suspended-load fall-out (Lowe, 1982, 1988; Arnott & Hand, 1989; DeCelles & Cavazza, 1992) and extreme unsteadiness of the flows (Allen, 1991). (2) The common presence of normal grading, particularly among the denser siliciclastic clasts (Fig. 8e), is consistent with rapid suspension fall-out. Similarly, the density stratification within these beds (Fig 8e), with the siliciclastic clasts concentrated in the lower parts of beds (dense-grain tail normal grading, Mulder & Alexander, 2001), indicates that rapid suspension fall-out, rather than traction flow, was the dominant depositional process. (3) Inverse and inverse-to-normal grading (Fig. 8e, f) are common attributes
Table 1. Framework composition of sandstone samples

<table>
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<th>Formation Samples</th>
<th>Fium Albino Fm</th>
<th>Torra and Monte Sant’Angelo formations</th>
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<th>Farinole Fm</th>
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<td>60.9</td>
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<td>25.5</td>
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</table>

Notes: Lv, volcanic lithic fragments; Ab, albite; Ep, epidote; CE, extrabasinal carbonate clasts; CI, intrabas. carb. clasts; NCE, noncarb. extrabasinal clasts; Q, total quartz grains; F, feldspars; L, aphanitic lithic grains.
of hyperconcentrated density flows, and result from inertial
grain-to-grain collisions (Lowe, 1982; Mulder & Alexander,
2001). Analogous high-concentration turbidites are widely
documented in the literature (Surlyk, 1978, 1984; Postma,
1986; Ghibaudo, 1992; Cavazza & DeCelles, 1993; Mulder &
Alexander, 2001; Naruse & Masuda, 2006; Eyles & Januszcz-
ak, 2007; Patros et al., 2007). Although the water depth in
which these deposits accumulated is not known, we can con-
fidently state that they must have been below storm wave
base because of the general absence of typical storm-related
coarse-grained lithofacies such as hummocky and swaley
cross-stratification, storm- and poststorm lags and strongly
erosional basal surfaces (e.g. Clifton, 1981; DeCelles, 1987;
Duke, 1987; Dumas & Arnott, 2006).

Using the terminology proposed by Ashley (1990) the
Saint-Florent sandwaves can be classified as large subaque-
sous dunes, mostly two-dimensional (i.e. straight-crested).
Altogether, the sedimentological features described above
indicate that the sandwave deposits of the Saint-Florent
succession formed in response to effectively unidirectional
currents, perhaps because the reversing tidal flow
remained below the critical threshold for sediment trans-
port, resulting in bedforms with unidirectional, angle-
of-repose foresets. The simple internal structure indicates
that temporal and spatial variability of processes produ-
cing the bedforms was low.

The predominance of corallineaceous red algal detritus
at Saint-Florent fits well with global chronostratigraphic
patterns. In fact, late-early to early-late Miocene
eritic carbonate settings are characterized by a worldwide
bloom of coralline red-algal (rhodalgal; Carrannante et al.,
1988) facies. Coralline red algae are commonly known as
encrusters in shallow coral reefs and on rocky substrates
or, in the absence of hard substrates, can also occur as
rhodolith nodules inhabiting sandy substrates. Rhodalgal
facies are known to develop extensively in areas of nutri-
ent-rich upwelling and under reduced-light conditions.
Halfar & Mutti (2005) attributed the global shift from
coral- to rhodolith-dominated carbonate communities
during the Burdigalian to early Tortonian to Miocene
global changes. According to these authors, a combination
of a global enhancement of trophic resources and declin-
ing temperatures led to oceanographic conditions that fa-
voured the expansion of rhodalgal communities.

**DETRITAL COMPOSITION**

In order to determine sandstone detrital modes and prove-
nance of the Saint-Florent Basin fill 25 samples were col-
lected, thin-sectioned and examined petrographically. Of
these, fifteen samples were point-counted following the
Gazzi-Dickinson method (Gazzi, 1966; Dickinson, 1970),
a procedure that minimizes the variation of composition
with grain size. Three hundred points per thin section
were counted and assigned to one of seventeen categories.
Sandstone point-count data were then recalculated to
produce the grain parameters indicated in Table 1. Prove-
nance was also qualitatively assessed on 14 conglomerate
outcrops covering the entire Saint-Florent sedimentary
succession. The following is a synthesis of all quantitative
and qualitative observations.

The detrital modes of the Fium Albino Formation are
characterized by abundant siliciclastic lithic fragments:
phyllite, felsitic volcanics, albite– epidote semischist, basic
volcanics, serpentinite and serpentine schists (Table 1). A
few aplite grains are also present. The relative proportions
of these lithic grains vary locally, suggesting control by the
local drainage network of the small fluvial systems repre-
ented by this lithostratigraphic unit. These rock types
match those presently exposed in the Alpine orogenic
wedge nearby to the cast, except for the felsitic volcanics,
which do not have a counterpart within the orogenic wedge
and were derived from the Permian Mt. Cinto volcanic dis-
trict within the foreland. Sandstones of the Torra and Monte
Sant’Angelo Formations have a predominant carbonato-
clastic intrabasinal component (Table 1, Figs 11 and 12).
Most common bioclasts are calcareous algae, bryozoans,
benthic foraminifera and mollusks. Siliciclastic rock frag-
ments are identical to those in the Fium Albino Formation.
Averaging 2.5%, felsitic volcanic grains are less abundant in
these units than in the underlying Fium Albino Formation.
The Saint-Florent conglomerate is composed almost exclu-
sively (> 98%) of felsitic volcanic clasts (ash-flow tuffs).
Such volcanic clasts are identical to those present in the un-
derlying lithostratigraphic units. The fine-grained sand-
stones of the Farinole Formation are dominated by carbonates
intraclasts (planktic foraminifera) with subordi-
nate quartz and feldspars grains (Fig. 10); minor amounts
of phyllite and felsitic volcanic clasts also are present.

On a carbonate lithic grains plot (Fig. 12), all sandstone
samples of the Saint-Florent succession lack extrabasinal
carbonate grains. The litharenites of the Fium Albino and
Saint-Florent Formations are devoid of intrabasinal car-
bronate detritus; the detrital modes of all other samples are
dominated (52–94% of the total framework grains) by
intrabasinal carbonate grains.

On a standard QFL plot (Fig. 12), most sandstone sam-
ple of the Saint-Florent succession fall within the ‘rec-
cycled orogen’ field of Dickinson et al. (1983), i.e.
compositions resulting from the orogenic uplift and ero-
sion of fold belts and thrust sheets. The only exceptions
are the litharenites of the basal Fium Albino Formation
(Alpine metamorphic rock fragments) and Saint-Florent
conglomerate (late Hercynian volcanic rock fragments).

The abundance of serpentinitic siliciclastic grains – typi-
cal of sandstones derived from collisional orogenic ter-
ranes (Sueck & Ingersoll, 1985) – indicates that ophiolitic
units of the Alpine orogenic wedge were available as sedi-
ment source rocks.

Based on the qualitative observation of conglomerate
composition, Fellin et al. (2005) concluded that the extra-
basinal provenance of the Saint-Florent succession is
dominated by the Alpine units, which still form the east-
ern margin of the basin, with a subordinate detrital input
from the Tenda gneissic dome to the west. In their inter-
deposits and high-density turbidity current deposits composed of reworked shallow-marine carbonate bioclasts, raises the prospect that this assemblage may have paleotectonic/paleoenvironmental significance. In particular, we propose that this blend of lithofacies should be characteristic of relatively deep marine environments that are swept by both powerful bottom currents and sporadic turbidity currents.

Offshore marine gravel and sandwave deposits have been documented in a number of turbidite fan systems (e.g. McHugh & Ryan, 2000), but these deposits typically exhibit very low-angle cross-stratification. High-angle gravel and sandwave deposits are known from a range of water depths (100–1000 m) in the Messina Strait and its approaches (e.g. Ryan & Heezen, 1965; Santoro et al., 2002, 2004), and in shallower (30–106 m) shelf environments near the Golden Gate of San Francisco, USA (Barnard et al., 2005, 2006). Both these settings are characterized by powerful bottom currents (1.2 m/s at ~300 m in the Messina Strait) produced mostly by bathymetric focussing of tidal currents. However, only deeper environments such as the Messina Strait are also sites of deposition by means of gravity flows. At this location, at depths in excess of 400 m, Selli et al. (1975) documented the presence of turbidite layers (average recurrence time ca. 60 years) interbedded with sandwave deposits and derived from the rugged and tectonically unstable basin margins. The Monte Torre paleostrait of southern Italy – which linked the Tyrrhenian and Ionian seas during Plio–Pleistocene time (Colella & D’Alessandro, 1988) – provides a well-documented fossil analogue of the Messina Strait. Sedimentation along the axis of the paleostrait was characterized by gravel or very coarse sandwave deposits composed mostly of carbonate bioclasts. Such large carbonate sandwaves required strong currents and high biological productivity, which were induced by current amplification along the paleostrait and the corresponding upwelling of nutrients, respectively. The Monte Torre sandwave deposits are interbedded with thick turbidites – the result of intermittent high-density turbidity currents mobilizing mixed bioclastic–siliciclastic detritus from the tectonically active basin margins. Based on ichnofacies, stratigraphic relationships and sedimentary facies analyses, Colella & D’Alessandro (1988) concluded that deposition in the Monte Torre paleostrait occurred at water depths in excess of 350 m.

Combined with the structural–tectonic setting, the Saint–Florent Basin lithofacies assemblage may be interpreted as a type of textfacies, characteristic of narrow, relatively deep-water (upper bathyal?) basins subject to bathymetric focussing of (tidal) currents and derivation of coarse-grained turbidites from nearby rugged topographic highlands, a physiographic situation characteristic of the early stages of microplate fragmentation and drift. In support of this hypothesis, similar bioclastic sandwave deposits of Tortonian age occur along the Tyrrhenian coast of southern Italy

DISCUSSION

Paleoenvironmental significance of the Saint–Florent Basin

The peculiar assemblage of lithofacies in the Saint–Florent Basin, characterized by intercalation of sandwave
(Longhitano & Nemec, 2005) and mark the early stages of rifting of the Tyrrenian back-arc basin (Bonardi et al., 2001), when the Calabria–Peloritani terrane drifted off the Corsica–Sardinia microplate (see following section).

**Tectonic significance of the Saint-Florent Basin**

The Saint-Florent sedimentary succession was deposited in a tectonic context of regional extension that character-
Rifting in the Ligurian–Provençal basin area occurred at least from the Oligocene (34 Ma) to the middle Aquitanian (21 Ma), according to the age of the sediments drilled in the Gulf of Lions beneath the break-up unconformity (Gorini et al., 1993). A number of grabens have been seismically imaged and drilled in Provence, both on land and offshore, in order to test their hydrocarbon potential (Vially & Trémolières, 1996). The development of the Provençal rifts traditionally has been interpreted as the result of the collapse of the Languedoc–Provence thrust belt, an eastward extension of the Pyrenees. From a wider perspective, the Provençal rifts are part of a regional extensional area cross-cutting the Pyrenees–Languedoc–Provençal orogen and ultimately driven by incipient roll-back of the Ionian subduction zone.

Drifting and the creation of the central, oceanic portion of the Provençal basin took place during the Burdigalian, as indicated by paleomagnetic data (Vigliotti & Langenheim, 1995) and by the transition from syn-rift to postrift subsidence of its margins (Vially & Trémolières, 1996; Roca, 2001). Although still somewhat controversial, paleomagnetic studies (Alvarez et al., 1974; Vigliotti & Langenheim, 1995; Speranza et al., 2002) indicate counterclockwise rotation of the Corsica–Sardinia block between ca. 19 and 16 Ma (Burdigalian), synchronous with the formation of oceanic crust in the Provençal basin (Fig. 13). The rifted margins of Provence and western Corsica display the same Oligocene–Aquitanian synrift sequences, being the conjugate margins of the same Neogene Provençal oceanic domain. Postrift thermal subsidence commenced during the middle Miocene in the adjacent Provençal and Algerian basins (Roca, 2001; Carminati et al., 2004; Roca et al., 2004).

Despite its small size, the Saint-Florent Basin has played a significant role in tectonic interpretations of the western Mediterranean region. Traditionally, the accretionary prisms of northern Corsica and the northern Apennines have been considered as separate entities. The west-vergent orogenic wedge of Corsica is considered the continuation of the western Alps (see Dal Piaz et al., 2003, and Lacombe & Jolivet, 2005, for reviews). Similarly, the crystalline rocks of Calabria in southernmost peninsular Italy have been long considered a fragment of the Alpine orogen separated from the Corsica–Sardinia block by the development of the Tyrrhenian basin since the Tortonian (see Bonardi et al., 2001, for a review). The correlation between Corsica and the western Alps is based on a wealth of structural and stratigraphic similarities, both in the orogenic wedges and in the foreland basins. Such similarities point to the existence of an east-dipping subduction zone (present-day coordinates) beneath the Corsican Alpine orogenic wedge. The northeast-vergent northern Apennines are considered to have been produced by limited, possibly passive, subduction of the Adriatic sub-plate towards the west (see Vai & Martini, 2001, and Elter et al., 2003, for reviews) in a different tectonic regime that mostly postdated Alpine compression and thus was not related to the genesis of the Corsican orogenic wedge.
Brunet et al. (2000) proposed an alternative interpretation envisioning a continuous time–space migration of deformation from Corsica to the Apenninic frontal thrust from at least 35 Ma to the present in a context of continued W–dipping subduction. From this viewpoint, ductile structural inversion of Alpine thrusts in Corsica (Jolivet et al., 1990; Fournier et al., 1991) and the development of the Saint-Florent extensional basin (Burdigalian–Langhian) would represent a bridge between the opening of the Provençal basin (early Miocene) and of the Tyrrenian Sea (latest Miocene–Pliocene), thus supporting the notion of an extensional wave moving towards the east.

In contrast with this hypothesis, available isotopic ages from Alpine Corsica point to a substantial time gap between the youngest metamorphic ages and the age of the oldest sedimentary unit in the Saint-Florent Basin fill. According to Brunet et al. (2000), two generations of white micas coexist in the basement rocks of the Tenda massif. A first generation is the relict of blueschist–facies contractional metamorphism around 45–35 Ma (i.e. the effect of late Eocene Alpine thrusting), whereas the second formed around 25 Ma during lower grade metamorphism associated with the structural inversion of Alpine top-to-the-west thrust faults. The mid-Burdigalian (ca. 19–18 Ma) age of the base of the Saint-Florent Basin fill would seem to preclude a direct tectonic relationship between the basin and low-angle normal faulting, although it is not excluded that significant increments of displacement along the East
1. LATE OLIGOCENE: beginning of orogenic collapse, inversion of Alpine boundary fault

Tenda shear zone might have occurred later in the brittle field.

The exact crosscutting relationships between the E Tenda shear zone and the high-angle normal faults are far from being completely understood. Fellin et al. (2005) have interpreted the high-angle normal faults as cutting the detachment, an interpretation we have incorporated in Fig. 14. On the other hand, the supradetachment location of the basin suggests that the high-angle faults might sole out into the main detachment, and that isostatic uplift owing to tectonic unroofing of the Tenda core complex could have delayed subsidence and at least temporarily prevented the development of large amounts of sediment accommodation.

The apatite and/or zircon fission-track studies of Cavazza et al. (2001), Jakni et al. (2000), Zarki-Jakni et al. (2004) and Fellin et al. (2005) indicate that Corsica underwent a widespread episode of rapid exhumation in early-middle Miocene time with a cluster of ages at 22–15 Ma. The data show that before exhumation most of Corsica was covered by at least 2 km of Alpine thrust sheets and foreland deposits. There are no systematic age differences across the main boundary fault system separating the Alpine orogenic wedge and the foreland, indicating that large-scale inversion of this structure was no longer active during the early Miocene.

Following this line of reasoning, the late Oligocene–early Miocene structural evolution of northern Corsica can be summarized as shown in Fig. 14. During the late Oligocene, following maximum thickening of the Alpine orogenic wedge, the boundary fault system between the Alpine wedge and its foreland was tensionally reactivated as a low-angle detachment fault. This first phase corresponds to the development of the marginal grabens along the Provençal conjugate margin. During early Miocene time, Corsica was uplifted wholesale although high-angle
brittle normal faults locally created accommodation space for basin development both on land (Saint-Florent) and offshore (Corsica basin). This interpretation is also supported by geomorphological data pointing to the existence of a piedmont-type planation surface, which formed during a phase of tectonic quiescence in the middle Miocene following early Miocene uplift (Kuhlemann et al., 2005).

CONCLUSIONS

The Saint-Florent sedimentary succession in northern Corsica represents the remnant of a basin that developed in Burdigalian–Langhian time after the postcollisional orogenic collapse of the Alpine tectonic wedge. In spite of its supradetachment setting, basin development was controlled by high-angle normal faults structurally unrelated to the underlying detachment fault system.

Owing to postdepositional erosion, the maximum areal extent of the Saint-Florent Basin cannot be reconstructed. Nevertheless, scattered outcrops of time-equivalent continental, transitional and shallow-marine deposits occur elsewhere along the boundary between the Alpine orogenic wedge and its foreland, at least from Saint-Florent to Corte in central Corsica. This points to the possible existence in late Burdigalian–Langhian time of roughly N-S-trending elongate and narrow seaway(s) possibly linking the developing Ligurian–Provençal basin and the shallow epicontinental sea to the east and southeast of Corsica (Fig. 13; 16 Ma).

The bulk of the Saint-Florent Basin fill is composed of large-scale sandwave deposits made of biogenic carbonate detritus whose paleocurrents point consistently to the NNE. Thick, graded beds composed of rhodoliths are locally interbedded with the sandwave deposits and represent the result of sediment-gravity flows from the basin margins, possibly triggered by seismic shocks. The middle Miocene palaeogeography of northern Corsica was thus characterized by long, narrow furrow(s) that were swept by intense currents, as demonstrated by the large sandwave deposits. The high biological productivity can be explained as the result of upwelling induced by favourable oceanographic conditions, i.e. narrow and shallow straits connecting large, deeper bodies of water.

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