Palaeostrait tectonosedimentary facies during late Cenozoic microplate rifting and dispersal in the western Mediterranean

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Abstract: Palaeostraits are common features formed during the extension and fragmentation of active continental margins, when back-arc basins are formed and microplates develop through upper-plate extension. In this contribution, we focus on narrow seaways and straits associated with successive pulses of late Cenozoic rifting along the southwestern Eurasian continental margin, and the ensuing formation and dispersal of microplates. The incomplete and largely non-cylindrical Africa–Eurasia continental collision in the Mediterranean region provides snapshots of the various phases of microplate fragmentation and dispersal, punctuated by the presence of narrow seaways/straits. These peculiar physiographical conditions induced current amplification, and led to the accumulation of characteristic large-scale, cross-stratified deposits and tidal facies in an otherwise generally microtidal setting. Although not univocally related to rifting, the identification of such distinctive sedimentary facies within deformed orogenic belts may ultimately help in terrane analysis and in the discrimination of individual microplates, which were then amalgamated. Recognition and dating of palaeostrait tectonosedimentary facies can thus constrain the presence and the onset of extension in the geological record.

Rift basins can originate along plate margins as well as in intraplate settings (e.g. Contreras *et al.* 1997; Gawthorpe *et al.* 1997; Gupta *et al.* 1998). Their structural architecture and orientation are fundamental to infer the regional palaeostress field, and an analysis of their basin-fill successions is useful in constraining phases of rift nucleation, propagation and linkage (e.g. Lambiase and Bosworth 1995). Many modern and ancient extensional depocentres have evolved through the interconnection of previously isolated, adjacent small endorheic basins, lined along the main extensional rift axis. Extensional fragmentation of continental margins is a recurrent phenomenon in Earth's history, with the generation of tectonically controlled extensional corridors, formed by coalescent graben and half-graben, shaped into block-faulted compartments with local topographical narrowing or funnelling (e.g. Morley 1999). Where flooded by marine waters, these settings may turn into straits, with characteristic kilometre-scale margin constrictions or shallowing (as in the modern Messina, Gibraltar and Cook straits). In other instances, tectonics may produce wider marine passages on the scale of tens to hundreds of kilometres known as seaways (e.g. Bering, Florida and Hudson). Common to both scales of setting is the oceanographic circulation pattern of marine or tidal currents flowing parallel to the axis of these basins. In straits, marine waters are usually subject to significant turbulence due to the convergence of water masses and their consequent amplification, resulting in current acceleration and a predictable distribution of erosional/depositional zones along the strait bottom (Longhitano 2013). In seaways, these phenomena are less pronounced or absent, and oceanographic currents are likely to be more persistent than in straits, running for great distances along the coastlines and influencing marginal-marine systems such as tidally deflected deltas and tidally modulated shorefaces (Dashtgard *et al.* 2009; Steel *et al.* 2012).

Straits and seaways are recognized as loci of accumulation of clastic sediments under current-dominated conditions (for a review see Longhitano and Steel 2016; Longhitano and Chiarella 2020; Dalrymple 2022; Rossi *et al.* 2023). These subaqueous deposits result from bedload parting exerted by oceanographic or tidal currents (Harris and Collins 1991; Harris *et al.* 1995) generating favourable conditions for the accumulation of extensive sand banks hosting a variety of bedforms, including dunes, bars (i.e. sand waves) and sand ridges (Desjardins *et al.* 2012). Such deposits have a good preservation potential as they can be subject to moderate lowstand

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incision or be rapidly buried by fine-grained sediment after major trangressions (Longhitano 2013). Strait deposits exhibit diagnostic sedimentological features such as: (i) thickness of the order of tens to hundreds of metres; (ii) large-scale cross stratification; (iii) wide erosional discontinuities; and, in case of tidal dominance, (iv) a variety of tidal indicators, including reactivation surfaces, tidal bundles and herringbone cross-strata (Longhitano *et al.* 2012*b*). Although these physical features can be common to other tide-dominated systems, including estuaries, deltas and tidal flats, strait-fill facies associations exhibit diagnostic axis-orientated facies tracts, reflecting the downcurrent energy depletion from the strait centre towards its wider exits (Longhitano 2013). Coarse-grained gravel pavements mark the zone of maximum current acceleration, and merge downcurrent into ribbons and then into large-scale cross-bedded complexes (Caterina *et al.* 2022). The latter are commonly tens to hundreds of metres thick because of the lateral constriction and are deposited where dominantly asymmetrical tidal currents decelerate, with characteristic 3D-to-2D stacking of cross-beds (Longhitano and Chiarella 2020). Large-scale cross-beds grade downcurrent into mud-rich or heterolithic distal facies containing sporadic cross-laminated lenticular intercalations reflecting the strait exit, where tidal current slacken and invert their direction, starting a new tidal cycle. These characteristics are commonly used as physical criteria to identify ancient straits and seaways in the rock record, and to envisage episodes of oceanographic connections between adjacent basins of the past (Longhitano and Chiarella 2020; Rossi *et al.* 2023).

Straits and seaways may also punctuate phases of rifting, particularly in back-arc regions, and their occurrence marks important stages of plate fragmentation, basin opening and continental mass drift during Earth's history. In this contribution, we review and compare some of the best-known palaeostrait successions that were deposited in an extensional regime in the western Mediterranean region during the Cenozoic (Fig. 1). The Mediterranean palaeostrait successions related to compressional tectonics and lithospheric flexure, such as the Betic and Rifian corridors of the Gibraltar region and the peri-Alpine corridor (Fig. 1), are also considered and compared with extensional palaeostrait successions based on their stratigraphic trends.

Starting from the Oligocene, following the maximum indentation along the Alpine collisional front, the western Mediterranean underwent repeated episodes of back-arc extension and plate fragmentation with a resulting oceanographic connection with the adjacent Atlantic Ocean, commonly corresponding with episodes of relative sea-level highstand (Blanc 2002; Meijer and Krijgsman 2005; Loget and Van Den Driessche 2006). Tidally modulated oceanic water penetrated into the western Mediterranean

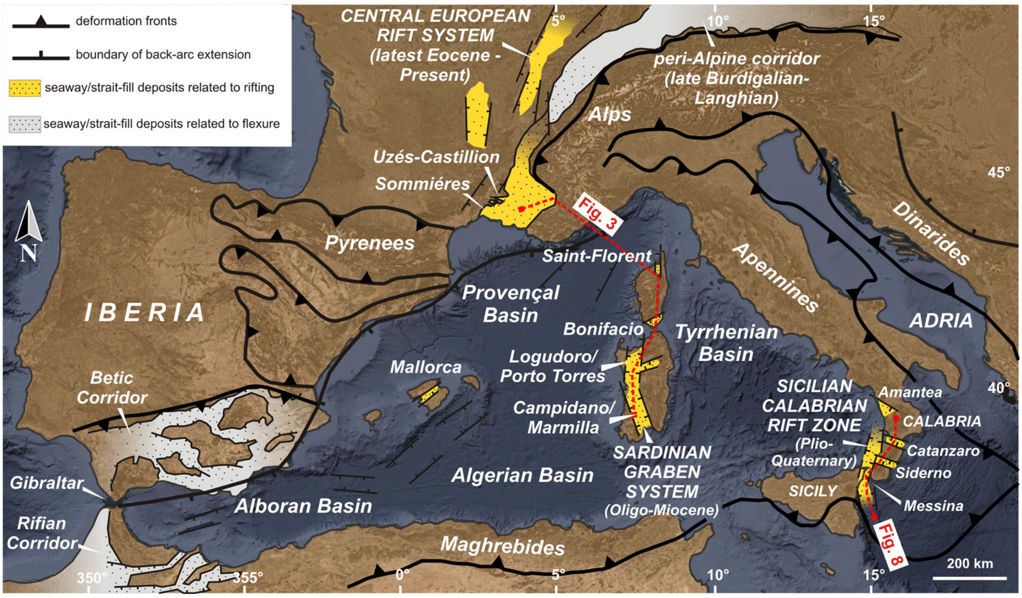

Fig. 1. Schematic structural map of the western Mediterranean region showing the main orogenic fronts and back-arc extensional areas (based on Carminati *et al.* 2012). Oligocene–Quaternary seaway/palaeostrait successions preserved onshore are differentiated based on their structural setting. Red dotted lines indicate stratigraphic correlation panels of major tidalite-bearing successions of southern France–Corsica–Sardinia (Fig. 3) and the CPT (Fig. 8).

through the Betic and Rifian corridors (Fig. 1) in the region of the present-day Gibraltar Strait. This oceanographic influence and the contemporaneous interconnection of previously isolated endorheic areas stimulated the onset of strait/seaway sedimentation in sectors subjected to tectonic extension during specific time intervals of the Neogene–Quaternary. In this paper, we: (i) discuss the close association between the peculiar sedimentological facies invariably associated with these straits and the late Cenozoic geodynamic context of the western Mediterranean region; and (ii) argue that such facies, although not necessarily connected with microplate fragmentation/dispersal, may be of great help in palaeogeographical/palaeotectonic reconstructions as a tool for dating episodes of rifting worldwide and identifying (micro)terranes in accretionary orogens.

The Neogene–Quaternary evolution of the western Mediterranean region

Following the Eocene collision between the African (Adriatic) and European plates along the Alpine front, the Mediterranean region has been affected by widespread extensional tectonics. Oligocene–Quaternary post-collisional domains of the western Mediterranean include, from west to east, the Alboran, Provençal and Tyrrhenian basins (Fig. 1). Such extensional domains have been interpreted as back-arc basins driven by the progressive roll-back of a subducted lithospheric slab in the general framework of: (i) a progressively slower convergence between the African and the European plates; and (ii) the lateral, non-cylindrical structural relationships between deeply indented promontories and vast embayments along the morphologically irregular European and African converging continental margins (e.g. Malinverno and Ryan 1986; Carminati *et al.* 1998; Cavazza *et al.* 2004; Cloetingh *et al.* 2010). Another major area of lithospheric extension occurs in southern France and continues to the north within the European continental interior to form the Central European Rift System (Cloetingh *et al.* 2010). Such a compounded rift system had already experienced an early phase of subsidence in the Late Eocene in response to northerly directed compressional stresses (Ziegler and Dèzes 2006), followed by a post-collisional extensional phase during the Oligocene, when rifting propagated southwards into the Gulf of Lions (Vially and Trémolières 1996) and the Valencia Trough (Roca 2001) (Fig. 2a). A vast amount of literature on this complex tectonic scenario is available; in the following, a very simplified literature synthesis is presented.

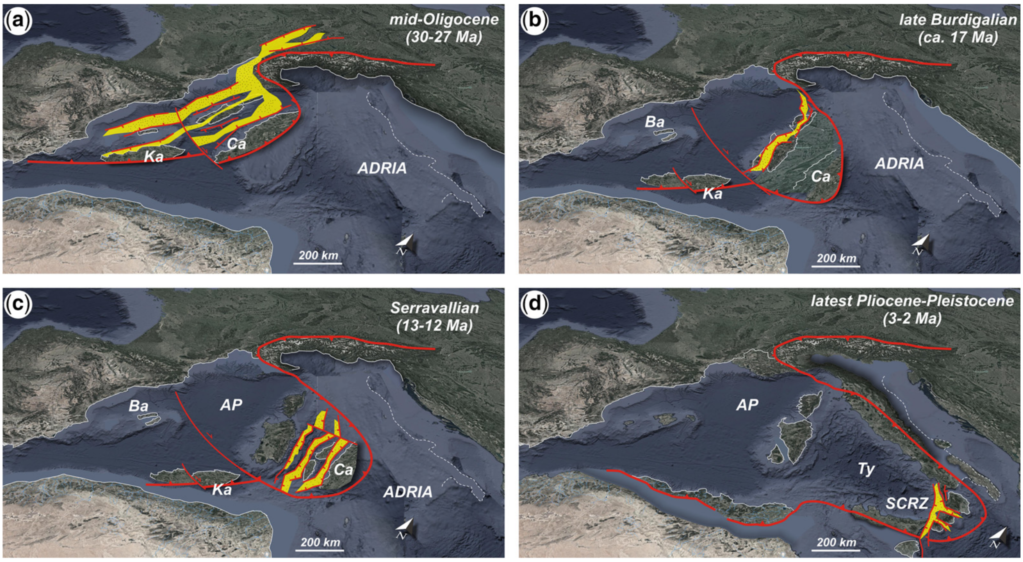

Fig. 2. Schematic palaeogeographical reconstructions of the western Mediterranean (after Séranne 1999, with integrations from Stampfli *et al.* 1998). Following Adriatic indentation and the Alpine collisional orogeny, progressive extension of the southwestern Eurasian continental margin was accommodated by sequential rifting and the ensuing opening of back-arc oceanic basins (in yellow). The early rifting phases of these basins are characterized by the deposition of large-scale, cross-bedded deposits (see the text for a discussion). AP, Algerian–Provençal Basin; Ba, Balearic block; Ca, Calabria–Peloritani Terrane; Ka, Kabylian terranes; SCRZ, Sicilian–Calabrian Rift Zone; Ty, Tyrrhenian Basin.

During Neogene time, in those areas not directly indented by Adria, a number of small fragments of continental lithosphere (Kabylides, Balearic Islands and Corsica–Sardinia–Calabria) rifted off the southern margin of Europe–Iberia and drifted towards the south and SE in response to roll-back of the NNW-dipping Alpine Tethys subducting slab and the opening of the Algerian–Provençal back-arc basin (Malinverno and Ryan 1986; Doglioni *et al.* 1997; Carminati and Doglioni 2004) (Fig. 2a, b). Such small blocks are now either thrust upon the northern margin of the African Plate (Petit and Grande Kabylie, Calabrian microplate) or left behind within areas of thinned continental crust or newly formed oceanic crust (Balearic Islands and Corsica–Sardinia) (Roca 2001; Cavazza *et al.* 2004). Western Mediterranean microplate rifting and dispersal occurred sequentially. The following is a concise description of the major pulses of extension, which will be then described in more detail in the following sections.

A first pulse of extension, microplate fragmentation/dispersal and back-arc basin formation occurred during the early Miocene (*c.* 20–16 Ma), when post-collisional extension in the northwestern Mediterranean induced rifting of a microplate (including what are now Corsica, Sardinia and Calabria) from the southern European–Iberian continental margin (Malinverno and Ryan 1986; Doglioni *et al.* 1997; Carminati *et al.* 1998). During drifting this microplate rotated counter-clockwise (Gattacceca *et al.* 2007), forming in its wake the oceanic Algerian–Provençal Basin, until in the late Burdigalian it collided with the African continental margins (Cavazza and Barone 2010) (Fig. 2b). To the west, a similar mechanism drove the rifting of the Balearic and Kabylian lithospheric blocks (Roca 2001). The latter drifted southwards until in the late Burdigalian they collided with the African margin (Bruguier *et al.* 2009), whereas the Balearic block was left behind.

A second pulse of extension occurred in middle–late Miocene time, when continued roll-back along the Ionian subduction zone induced further back-arc rifting in the upper plate, the opening of the Tyrrhenian Basin and the progressive separation of the Calabria–Peloritani terrane (CPT) from the Corsica–Sardinia block (Fig. 2c). Docking of the CPT in its current position between Adria to the north and the Sicilian–Maghrebian margin to the south is generally considered as Pliocene (Fig. 2d) (for a review see Bonardi *et al.* 2001).

Ancient connections between the Mediterranean Basin and the adjacent oceans

During late Mesozoic–early Cenozoic time the area which nowadays corresponds to the Mediterranean Basin connected the Atlantic Ocean to the west and the Neotethys to the east (Stampfli and Borel 2004; Barrier and Vrielynck 2008; Stampfli and Hochard 2009). This peculiar palaeogeographical/palaeoenvironmental setting induced water-mass exchanges (Bialik *et al.* 2019), affected the global climate and was an important palaeobiogeographical factor for the global distribution of palaeontological taxa. Starting from the Late Cretaceous, because of the northward subduction of the African Plate beneath the Eurasian Plate, the width of the Mediterranean connection decreased progressively.

The Mediterranean Basin was entirely disconnected from the Indian Ocean in the Middle Miocene (Harzhauser *et al.* 2007; Hamon *et al.* 2013; de la Vara and Meijer 2016; Bialik *et al.* 2019) and, episodically, from the Atlantic Ocean in the latest Miocene; the latter event resulting in the Messinian Salinity Crisis (Flecker *et al.* 2015; Meilijson *et al.* 2019). Data documenting the eastern connection of the Mediterranean with the Indian Ocean are mostly based on biogeographical variations following the closure of the so-called ‘Mesopotamian Seaway’ (Rögl 1998, 1999; Harzhauser *et al.* 2007; Reuter *et al.* 2009, 2013). A secondary connection across the Red Sea has also been hypothesized up until at least the Aquitanian (Cao *et al.* 2017; Segev *et al.* 2017), although with limited water-mass exchanges between the Mediterranean Sea and the Indo-Pacific oceans.

Unlike the eastern Mediterranean, where there is no evidence of oceanographic connections with the Indian Ocean after about 19 Ma (e.g. Harzhauser *et al.* 2007), the stratigraphic record of the western Mediterranean region provides abundant proof of intermittent connection with the Atlantic Ocean during the Neogene and the Quaternary. Linear belts of tide-dominated sedimentary strata crop out extensively along the so-called Betic and Rifian corridors to the north and south, respectively, of the present-day Gibraltar Strait (Fig. 1) (Flecker *et al.* 2015; Capella *et al.* 2017; de Weger *et al.* 2020). Episodic opening of these passageways may have played a fundamental role in providing the oceanographic tidal power in the semi-enclosed, micro-tidal Mediterranean, a prerequisite to activate the sedimentation of large-scale cross-bedded deposits along palaeostraits (Longhitano 2021), as discussed later. Considering the various ages of the seaway/strait-fill successions exposed in several Mediterranean areas (Fig. 1), one can conclude that a major influence on the oceanic Atlantic inflow occurred repeatedly during the late Cenozoic. Western Mediterranean Neogene–Quaternary cross-bedded deposits indicating bidirectional, tidally modulated, current patterns point to conditions of geomorphic constriction: that is, the second fundamental prerequisite for tidal amplification and the resulting accumulation of sand-rich, dune-bedded strata (Longhitano 2013, 2021). Clearly, the late Cenozoic back-arc rift basins discussed above provided ideal conditions for the generation of narrow seaways and the accumulation of thick successions of large-scale cross-bedded sedimentary deposits, particularly when the water exchange with the Atlantic Ocean was enhanced.

Latest Oligocene–Quaternary palaeostraits of the western Mediterranean

During late Oligocene–Quaternary time the western Mediterranean experienced crustal thinning, continental-margin fragmentation and microplate drifting as consequences of the progressive southeastward roll-back of the African Plate (Malinverno and Ryan 1986; Doglioni *et al.* 1997; Carminati *et al.* 1998) (Fig. 2a–d). Many basins generated during this important stage occupied a back-arc position and underwent intermittent periods of extension, transtension and block rotation (Ferrandini *et al.* 2003; Fellin *et al.* 2005). During major transgressions these areas were episodically connected with the Atlantic Ocean through the Gibraltar passageways (Fig. 1) and were thus affected by its tidal pulses, ultimately producing phenomena of current amplification in areas of geomorphic/bathymetric restrictions. The successions discussed in this paper were deposited in discrete time intervals of the Miocene (i.e. late Aquitanian–Burdigalian, Langhian–Serravallian and late Tortonian) and the Early Pleistocene (Gelasian). Depositional basins discussed below include the southern terminus of the Central European Rift System in southern France, various interconnected depocentres in the Corsica–Sardinia block and portions of the present-day CPT.

The Rhodanian–Provençal Basin

In the late Aquitanian–Burdigalian, the sea flooded a series of fault-bounded alluvial plains of southeastern France, accumulating sandy cross-stratified transgressive tidal deposits presently exposed in scattered outcrops. At that time this sector was at the intersection between the southwestern termination of the Alpine foreland basin (Upper Marine Molasse Seaway: Allen and Homewood 1984) and the southern end of the Central European Rift System (Dèzes *et al.* 2004) (Figs 1 & 2a). Atlantic tidal currents penetrating from the Rifian–Betic corridors affected these areas and were amplified when compressed through local straits. The present-day onshore Rhodanian–Provençal Basin (Fig. 1) (Sissingh 2001; Besson 2005) includes a number of subbasins where tidal deposits have been documented: Basse-Provence (Rubino 1988), Carpentras (Parize *et al.* 1997), Crest (Lesueur *et al.* 1990), Digne (Tessier and Gigot 1989), Forcalquier (Rubino *et al.* 1990), Saumane-Venasque (James *et al.* 2014), Sommières (Reynaud and James 2012), Uzès (Reynaud *et al.* 2006) and Valréas (Lesueur *et al.* 1990).

The tidalite-bearing successions of SE France (Fig. 3) have been attributed to semi-enclosed coastal settings, such as embayments or incised valleys, which were flooded and connected with the Mediterranean Sea during major episodes of transgression (Reynaud *et al.* 2006; Reynaud and James 2012). Cross-stratified, sand-size deposits comprise bioclasts deriving from barnacles, bryozoans, coralline algae (encrusting, branching, and rhodoliths), echinoids and foraminifers. This association reflects the presence of temperate carbonate factories adapted to high-energy, tide-winnowed environments associated with subaqueous dunes (Tessier and Gigot 1989; Lesueur *et al.* 1990; Parize *et al.* 1997). These deposits are stratigraphically organized into two depositional sequences marking stages of major transgression into the incised valley system in this region of France (Reynaud *et al.* 2006; Reynaud and James 2012). Subaqueous tidal dunes associated with storm deposits were deposited during the transgressive cycle of each depositional sequence. This feature is common to all areas of the Rhodanian–Provençal Basin, which comprises Burdigalian tidal deposits overlain by highstand fine-grained sediments that reflect the cessation of tidal amplification.

The Corsica–Sardinian Graben System

The Corsica–Sardinia–Calabria block is a lithospheric fragment that rifted off the southern continental margin of Iberia–Europe and drifted southeastwards with a counter-clockwise rotation of about 45° (Alvarez *et al.* 1974; Malinverno and Ryan 1986). Rifting occurred in the Oligocene (*c.* 30–22 Ma) (Vially and Trémolières 1996; Speranza *et al.* 2002), followed by drifting in the Early Miocene (*c.* 22–15 Ma) (Ferrandini *et al.* 2003) (Fig. 2a, b). Extension moved to the SE during the Late Miocene, when the opening of the Tyrrhenian Sea separated Calabria from Corsica–Sardinia (Doglioni *et al.* 1997; Carminati *et al.* 2012) (Fig. 2c). On the islands of Corsica and Sardinia, a series of graben and supradetachment basins bear the structural and stratigraphic record of prolonged extension. Scattered marine deposits featuring high-energy tractive bedforms suggest the existence of a seaway or a system of marine channels that became intermittently interconnected during periods of Miocene major transgressions (Longhitano *et al.* 2017; Telesca *et al.* 2020) (Fig. 4).

Strait-fill deposits of Corsica. During Burdigalian–Langhian time, Corsica experienced a combination

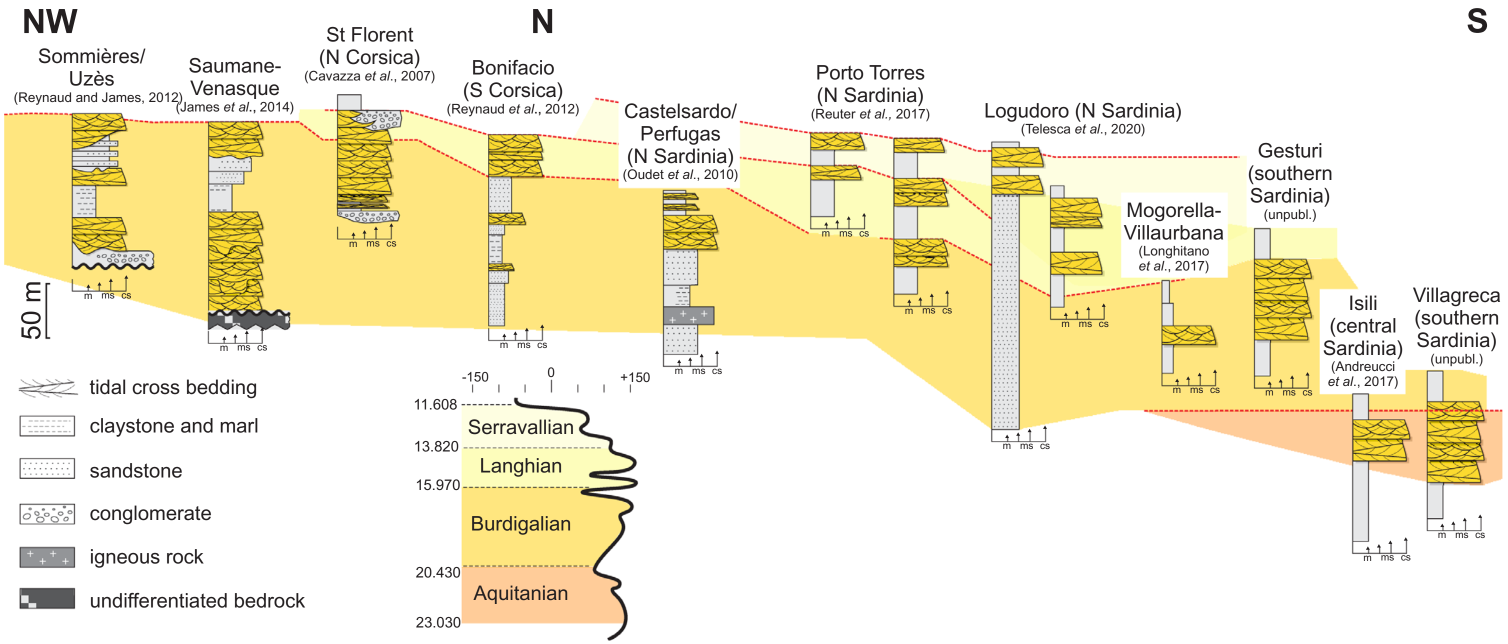

Fig. 3. Lithostratigraphic correlation between the lower–middle Miocene tidalite-bearing successions along the Provençal–Corsica–Sardinia transect (see Fig. 1 for the location). The eustatic sea-level curve is after Haq *et al.* (1987). m, mudrock; ms, medium-grained sandstone; cs, coarse-grained sandstone.

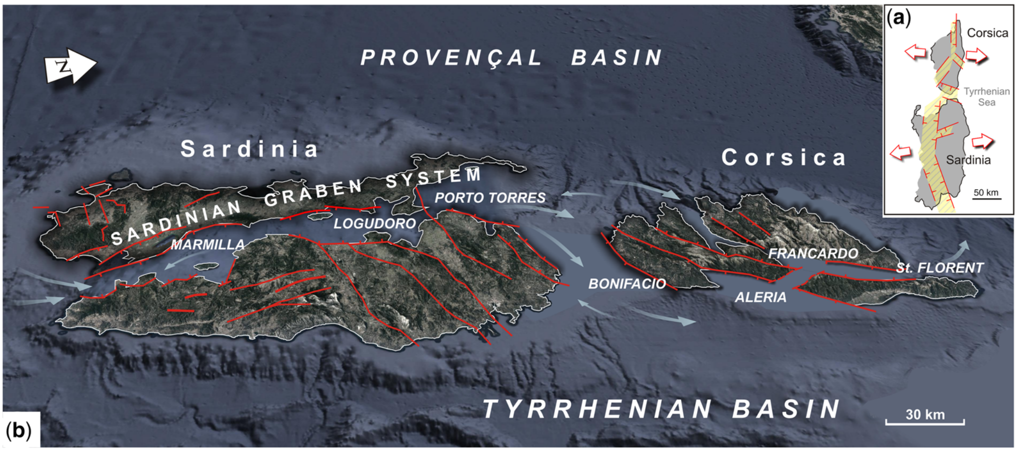

Fig. 4. Tentative palaeogeographical reconstruction of the Corsica–Sardinia block during the Early–Middle Miocene. The Sardinian Seaway was a marine conduit bordered by extensional faults through which tidal currents penetrated from the south and produced local effect of amplification and accumulation of sand-rich sand waves, which are preserved in discrete sectors of the various basins indicated in the figure. Present-day erosional remnants of sand-wave deposits in Corsica are too scattered to support the existence of a continuous seaway along the entire length of the island, although this possibility cannot be ruled out.

of: (i) eustatic sea-level rise; (ii) post-collisional structural inversion of the main boundary thrust system between the Alpine orogenic wedge and the foreland; and (iii) subsidence related to the development of the Ligurian–Provençal Basin (Faccenna *et al.* 2004). These concomitant processes created accommodation space for continental to shallow-marine sedimentary successions along narrow, elongated basins (Fig. 4a) where tidal currents were amplified, leading to the accumulation of distinctive large-scale cross-bedded sediments. Much of these peculiar deposits have been eroded and presently only a few scattered outcrop areas remain, most notably at Saint-Florent and Bonifacio. These deposits are correlative with strata exposed in Sardinia (Fig. 3), as discussed below.

At Saint-Florent in northern Corsica, the sedimentological and palaeontological characteristics of the palaeostrait succession point to a progressive deepening, from laterally discontinuous fluvial–alluvial deposits of undetermined age (Fium Albinu Formation) filling erosional depressions, through beach/nearshore deposits (Torra Formation: Burdigalian), to shallow-marine deposits (Monte Sant'Angelo Formation: late Burdigalian– early Langhian), which at a thickness of 250 m represent the bulk of the Saint-Florent sedimentary succession, and finally pelagic sediments (Farinole Formation: late Langhian–middle Serravallian) (Ferrandini *et al.* 1998; Fellin *et al.* 2005; Cavazza *et al.* 2007).

The Monte Sant'Angelo Formation (Vigliotti and Kent 1990; Ferrandini *et al.* 1998, 2003; Fellin *et al.* 2005; Cavazza *et al.* 2007) is dominated by intrabasinal bioclastic detritus (corallinacean red algae, bivalves, bryozoans and foraminifera), with subordinate siliciclastic extrabasinal detritus mostly derived from the erosion of the Alpine orogenic wedge. The formation is characterized by large-scale (up to 5 m thick) sets of cross-bedded, well-sorted, coarse biocalcarenite, with limited occurrences of megaripple cross-lamination in the bottom portion of the foresets. Co-sets are bounded by laterally extensive, planar erosional surfaces. Mud drapes are generally absent. Such facies association was produced by the migration of large subaqueous dunes, mainly two-dimensional (i.e. straight-crested). Most palaeocurrent indicators of the Monte Sant'Angelo Formation yield an average palaeoflow direction towards the NNE, with a limited number of indicators pointing in the opposite direction. Overall, these features show that the dunes formed in response to effectually unidirectional currents; subordinate tidal current are likely to have remained below the critical threshold for sediment transport, thus resulting in bedforms with unidirectional foresets (Cavazza *et al.* 2007).

The Burdigalian–Langhian Saint-Florent sedimentary succession is the remnant of an originally larger basin that developed after the post-collisional collapse of the Alpine orogenic wedge (Jolivet *et al.* 1991; Cavazza *et al.* 2001). In fact, 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 central

Corsica. This points to the possible existence in late Burdigalian–Langhian time of a roughly NNW–SSE-trending elongate and narrow seaway swept by intense currents and connecting the Ligurian–Provençal Basin with the Aleria Basin in the northern Tyrrhenian region (Fig. 4).

Time- and facies-correlative deposits of the Saint-Florent succession crop out on the exposed margins of the modern Bonifacio Strait connecting the Provençal Basin to the west and the northern Tyrrhenian Sea to the east (Fig. 1). Most of the Miocene deposits lie on the French side of the strait, where they fill structural lows formed between extensionally faulted, NE–SW-trending blocks of Hercynian plutono-metamorphic basement rocks (Orszag-Sperber and Pilot 1976; Ferrandini *et al.* 2002) (Fig. 4). One small equivalent Miocene outcrop occurs at Capo Testa along the Sardinian side of the strait. All of these outcrops collectively represent the onshore part of a much larger basin that occupies most of the Bonifacio Strait, where the Miocene succession reaches a maximum thickness of about 500 m (Berra *et al.* 2019). To the south, this basin connects to the Castelsardo Basin in the northern part of the Sardinian Graben System (Figs 2b & 4), which is discussed below.

The surface and subsurface geology of the Miocene Bonifacio Strait succession is relatively well known, both on land and in the offshore (for a review, see Orsini *et al.* 2010, 2011; Reynaud *et al.* 2013; Reynaud and Ferrandini 2021). The basin-fill succession, which crops out spectacularly along the coastal cliffs and is completely drilled inland, includes three formations with a stratigraphic/sedimentological trend similar to the one described above for the Saint-Florent Basin in terms of lithology, chronostratigraphy and palaeoenvironmental (from continental to shallow marine) evolution. The lowermost lithostratigraphic unit (Balistra Formation: late Aquitanian–early Burdigalian) is made up of continental sandstone and conglomerate. The two overlying units (Cala di Labra and Bonifacio formations) are fully marine, and define a progressive compositional and provenance change from siliciclastic extrabasinal at the base to carbonate intrabasinal (bioclastic) at the top. The stratigraphy, sedimentology and faunal content of the two upper lithostratigraphic units have been studied by a number of authors (e.g. Orszag-Sperber and Pilot 1976; Ferrandini *et al.* 2002, 2010; Jadoul *et al.* 2009; André *et al.* 2011; Orsini *et al.* 2011; Reynaud *et al.* 2013; Tomassetti and Brandano 2013). The Cala di Labra Formation (late Burdigalian) records the transgression in the Bonifacio Basin. Dominantly siliciclastic to mixed carbonate–siliciclastic, it comprises at least three orders of high-energy carbonate reefs that developed in a complex framework of varying sea levels. The overlying Bonifacio Formation (early Langhian), up to 200 m thick, consists mainly of large-scale cross-stratified bioclastic arenite (Fig. 5), which is interpreted as recording the migration of large submarine dunes under the influence of amplified tidal currents with a prevalent WNW direction. High-resolution seismic profiles show that the Bonifacio Formation extends into the central part of the Corsica–Sardinia Strait (Reynaud *et al.* 2013).

Increased carbonate production and the strong currents recorded by the Bonifacio Formation cannot be linked to a significantly higher eustatic sea level in the Langhian than in the late Burdigalian (Miller *et al.* 2020). The transition between the Cala di Labra and the Bonifacio formations is interpreted instead as being result of the structurally controlled opening and deepening of the Bonifacio Strait (Reynaud and Ferrandini 2021), as shown by the faulted Miocene depocentres and the abundance of synsedimentary deformation features. The opening of the Bonifacio Strait caused major regional palaeoceanographical changes as it linked the Provençal Basin to the west with the Aleria Basin (also known as Corsica Basin) to the east (Fig. 4). Being morphologically and bathymetrically different, these two basins are likely to be characterized by different amphidromic systems, as well as by tides of different phase and range (Reynaud and Ferrandini 2021). The resulting hydraulic gradient caused strong tidal currents to form across the Bonifacio Strait; a mechanism active even in microtidal settings such as the present-day Messina Strait (Colella 1990; Longhitano 2018*a*).

Strait-fill deposits of Sardinia. The Sardinian Graben System (*auct.*) dominates the geology of the western half of Sardinia (Figs 1 & 4). It consists of two segments. The main segment (Oligo-Miocene) occurs in north and central Sardinia with a north–south trend, it is *c.* 200 km long and 30–60 km wide, and it is intersected by subordinate and coeval NE–SW-trending transtensional sub-basins along its eastern margin. The second segment (Pliocene) occurs in southern Sardinia with a NW–SE trend, it is about 80 km long and 20 km wide, and cross-cuts and reactivates the southern end of the Oligo-Miocene rift (Sowerbutts 1997; Monaghan 2001). The entire system has considerable along-strike variability in the nature and timing of active structures. Available seismic lines and borehole data indicate a thickness of up to 2 km of sedimentary and volcanic rocks in half-graben structures (see Casula *et al.* 2001; Carmignani *et al.* 2012 for a review).

The Oligo-Miocene graben system of Sardinia developed within the context of the general extension of the southern European margin following the Alpine collision (Fig. 2a, b). Starting from the early Miocene, the Sardinian Graben was inundated

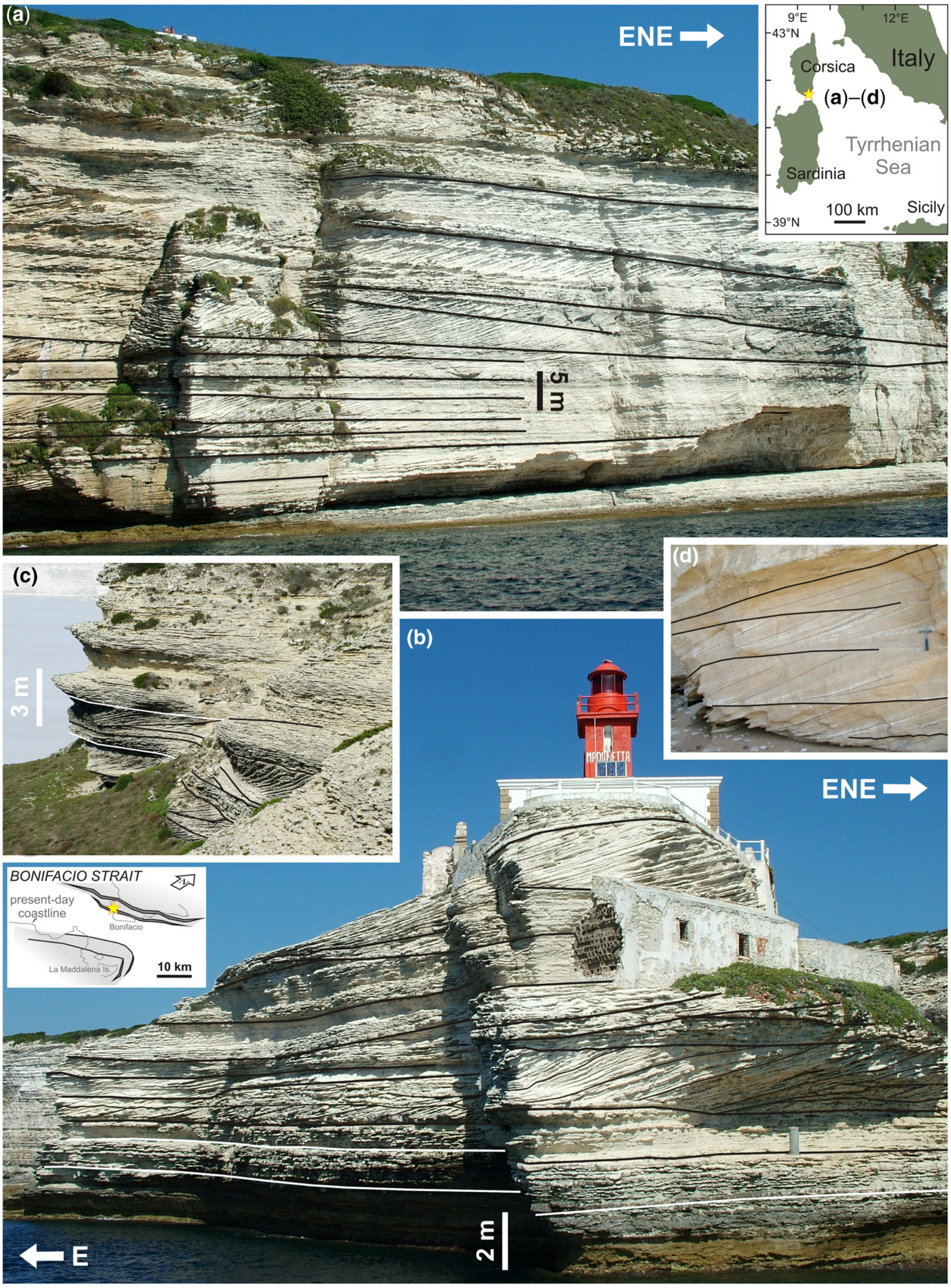

Fig. 5. Palaeostrait successions exposed near the town of Bonifacio (southern Corsica). (**a**) Tabular-based (planar) cross-bedding showing dominantly unidirectional foresets and indicating superimposition of straight-crested (2D) tidal dunes migrating towards the WSW (left of the photograph). (**b**) Large-scale tidal cross-bedding exposed east of the entrance to the Bonifacio harbour (‘Faro de la Madonetta’), comprising sigmoidal to tangential unidirectional foresets (direction of dune migration is towards the left of the photograph). The inset shows the Bonifacio Strait in Serravallian times and the location of the outcrops (star). (**c**) and (**d**) Close-up of tidal foresets showing the internal architectural complexities and tidal bundling indicated by lamina-set colour differences (photographs courtesy of V.M. Rossi).

by marine waters, turning progressively into a seaway characterized by a tidal circulation. The stratigraphy of the Castelsardo, Porto Torres and Logudoro basins of northern Sardinia (Fig. 3) shows strong correlative elements with the Bonifacio basin-fill succession described earlier in this chapter. The areas where all these basins lie correspond with the modern Corsica–Sardinian marine passageway and occupy a structurally complex transverse zone that partly accommodated the differential rotation between the Sardinian and Corsican margins (Muttoni 1998; Sissingh 2006; Telesca *et al.* 2020).

Based on integrated $^{40}Ar/^{39}Ar$, micropalaeontological and palaeomagnetic data, Oudet *et al.* (2010) showed that two major episodes of relative sea-level rise in the tectonically controlled Castelsardo Basin occurred in the early (19.7–18.8 Ma) and early–late (Chron C5Dr reversal–eustatic Bur 4 event) Burdigalian. The younger transgressive event correlates with the carbonate-rich, tidalite-bearing succession exposed in the Porto Torres Basin to the south (Figs 3 & 4). Sedimentary rocks cropping out at Porto Torres are dominantly carbonates and have been subdivided into three depositional sequences separated by basin-wide unconformities (Reuter *et al.* 2017). The lower sequence is made up of plain-parallel stratified, mixed siliciclastic–bioclastic arenite. Bioclasts consist of bivalves, coralline red algae, benthic and planktic foraminifers, echinoderms, bryozoans, balanids, and serpulids (Reuter *et al.* 2017). The uppermost part of this lower sequence is markedly progradational. The middle depositional sequence (Langhian) is also bioclast-rich, with large-scale cross-stratification at the base indicating migration of (tidal?) dunes towards the NE. The third depositional sequence (Serravallian) represents a carbonate ramp showing marked progradation and channelling (Vigorito *et al.* 2006; Murru *et al.* 2015; Reuter *et al.* 2017).

Burdigalian–Serravallian stratigraphic intervals correlative with those described above have been documented in the Logudoro Basin (Funedda *et al.* 2000), a half-graben filled by three main depositional sequences (middle Burdigalian–late Messinian) separated by regional unconformities (Telesca *et al.* 2020) (Figs 3 & 4). The middle sequence (Serravallian–Tortonian) records the onset of the marine connection between the Logudoro and the Porto Torres basins, which were previously isolated. The oceanographic opening of a seaway, at the beginning of a marine transgression, is suggested by facies indicating strong current reworking of sediment transiting along submerged deltaic platforms, where cross-bedded, vertically stacked, metre-thick sandstone intervals prevail (Fig. 3). These deposits are laterally equivalent to the large-scale cross-bedded sandbodies in the axial part of the seaway, where high-velocity currents promoted the migration of large-scale bars and subaqueous dunes. The overlying deposits, accumulated during a late stage of transgression, show higher textural maturity and exhibit herringbone cross-bedding, reactivation surfaces and lamina bundles with a clear tidal signature (Fig. 3). At this stage the seaway was swept by bidirectional and rotatory current patterns orientated roughly parallel to the basin axis. Carbonate-rich, large-scale bedforms (i.e. ridges) developed during this late stage of open-marine circulation, pointing to possible terrigenous sediment starvation in the basin (Telesca *et al.* 2020).

The late Burdigalian transgression is recorded in the Logudoro Basin by shallow-marine deposits that show no connection with the southern adjacent basins of the Sardinian Graben System. However, in central and southern Sardinia, scattered coastal successions may comprise tide-influenced stratigraphic intervals (e.g. the Mogorella–Villaurbana, Isili, Gesturi and Villa Greca successions in Fig. 3) (Vigorito *et al.* 2005; Andreucci *et al.* 2017; Longhitano *et al.* 2017). Although these deposits indicate no clear linkage between the small basins of Sardinia, the tidal influences recorded along marginal-marine settings and semi-enclosed embayments suggest that tides were amplified along local fault-controlled graben and half-graben.

Palaeostrait deposits related to the opening of the Tyrrhenian Sea

A family of graben filled by Serravallian–Quaternary sedimentary deposits trends parallel to the Tyrrhenian margin of Calabria, both on land and in the offshore, following the oroclinal bending of the CPT. A symmetrical set of coeval graben is present in the offshore of the conjugate Sardinian margin. Such marginal basins have recorded the progressive rifting/drifting of the CPT from the Corsica–Sardinia block and the resulting opening of the Tyrrhenian Basin (see Bonardi *et al.* 2001 for a review) (Fig. 2c). A good example is the middle–late Miocene Amantea Basin of NW Calabria (Fig. 1).

The Amantea Basin formed by the coalescence of small coastal graben (Argentieri *et al.* 1998) where a >600 m-thick, deepening-upward succession was deposited during Serravallian–Messinian time. The Amantea basin-fill succession is divided into three unconformity-bounded units (Colella 1995; Mattei *et al.* 2002): (i) The lower unit (middle–late Serravallian) covers non-conformably the tilted basement complex, is up to 280 m thick and consists of alluvial-fan/fan-delta deposits overlain by shallow-marine deposits. (ii) The middle unit (Tortonian–early Messinian) is up to 260 m thick and is made up of continental to shallow-marine conglomerate–sandstone overlain by large-scale, cross-stratified biocalcarenite showing a suite of tidal sedimentary structures, including herringbone cross-bedding, reactivation surfaces and simple-compound tidal foresets. Single cross-stratified beds can reach a thickness in excess of 20 m and result from the migration of 2D dunes forming a mounded longitudinal complex (Longhitano and Nemec 2005; Longhitano 2012) (Fig. 6). The deposition of large-scale, cross-bedded calcarenite dunes in the Amantea Basin occurred coevally with the initial phases of the opening of the Tyrrhenian Sea (Fig. 2c). The biocalcarenites are overlain by latest Tortonian–early Messinian mudstone interbedded with calcarenite. (iii) The upper unit (late Messinian) overlies the Messinian erosional surface formed during the eponymous sea-level drawdown (s). It is composed of a transgressive succession from alluvial-fan and fan-delta conglomerate to transitional and shallow-marine sandstone interbedded with mudstone.

The Tortonian biocalcarenitic succession of the Amantea basin-fill was deposited in a physiographically complex peri-Tyrrhenian shelf embayment (Figs 1 & 2c), whose northern part was linked with the southern part by a tectonic graben referred to as the Monte Pellegrino Strait (Colella 1995; Longhitano and Nemec 2005; Longhitano 2012). The graben was 1.5–2 km wide and 4 km long, had shallowly submerged margins, and a south-sloping sublittoral floor swept by strong, asymmetrical tidal currents. The water depth fluctuated due to episodes of rapid subsidence driven by extensional tectonics.

The Sicilian–Calabrian Rift Zone

The Sicilian–Calabrian Rift Zone (SCRZ) is a north–south-striking latest Pliocene–Quaternary extensional belt running along western Calabria and eastern Sicily for a total length of about 370 km (Fig. 7), and accommodating an ESE–WNW-orientated regional extension (Tortorici *et al.* 1995; Catalano *et al.* 2008) triggered by the progressive roll-back of the Ionian subduction zone. From north to south, the SCRZ comprises the Crati, Paola, Catanzaro, Mesima, Siderno and Messina basins, many of which formed during the Plio-Pleistocene as a series of coalescent straits (multiple strait system of Longhitano and Steel 2016). Starting from *c.* 2.5 Ma, these marine corridors underwent a tide-dominated oceanographic circulation, with amplified bidirectional currents regulated by semi-diurnal tidal inversions occurring between two deep-water bodies: the Tyrrhenian Sea to the NW and the Ionian Sea to the SE (Fig. 7) (Colella 1995; Longhitano *et al.* 2012*a*). As a result, mixed siliciclastic–bioclastic deposits, tens of metres thick, accumulated in the depositional portions of these straits, commonly separated by a central erosional sill where sediment bypass dominated because of the stronger tidal currents (Longhitano 2012, 2013; Longhitano and Chiarella 2020) (Fig. 8). Palaeostrait successions of the SCRZ display large-scale cross-stratification due to repeated superposition of straight- and sinuous-crested tidal dunes that formed extensive bedform fields or elongate sand ridges (e.g. Longhitano *et al.* 2012*a*, 2014, 2020, 2021*c*; Longhitano 2018*b*) (Fig. 9). Palaeocurrent directions indicate average sediment transport pathways that were orientated roughly parallel with respect to the axis of each strait. This feature is a crucial element that is associated with the orientation of the main bounding normal fault, and differentiates the main direction of structural extension at the local scale and defines the orientation of the rift basin at a broader (semi-regional) scale.

Some of the SCRZ basins are orientated orthogonally to the curvature of Calabria (Figs 1 & 7). They were caused by a combination of transtension deformation and oroclinal bending during the docking of the CPT at its northern and southern ends (Ghisetti 1979) (Fig. 2d), whereas passive subduction of the Ionian oceanic lithosphere has continued underneath its central portion. The stratigraphic architecture of their basin fills is characterized by marginal fanglomerates interfingering basinwards with shallow-marine facies (e.g. Colella and D'Alessandro 1988; Martini *et al.* 2001). A significant component of strike-slip movement is indicated by laterally continuous fan-delta fringes and by systematic offlap and younging trends along the basin-margin facies (Colella 1988). Late Piacenzian–Calabrian sand-wave deposits made up of carbonate bioclasts of shallow-marine benthic organisms are common, and their geographical distribution points to the existence of narrow seaways connecting the Ionian Basin with the developing Tyrrhenian Sea. Such large carbonate sand waves required strong currents and high biological productivity, which were probably induced by current amplification along the palaeostraits and the corresponding upwelling of nutrients, respectively. These peculiar palaeoceanographical conditions ended with an abrupt uplift of up to 1000 m around the middle Pleistocene (Tortorici *et al.* 1995).

The present-day Messina Strait (Figs 1 & 7) is an impressive example of a tide-dominated marine seaway (see Longhitano 2018*a* for a review). A peculiar hydrodynamic configuration – due to the combination of coastal narrowing and a semi-diurnal phase inversion between the Tyrrhenian Sea and the Ionian Sea – produces tidal currents that reach a velocity of 3 m s^{-1}. Currents follow bidirectional patterns along the strait, accelerating in the erosion-dominated strait central restriction (where the minimum water depth is *c.* 70 m) and decelerating towards the strait exits because of the progressive enlargement of the cross-sectional area (Longhitano 2018*a*; Martorelli *et al.*

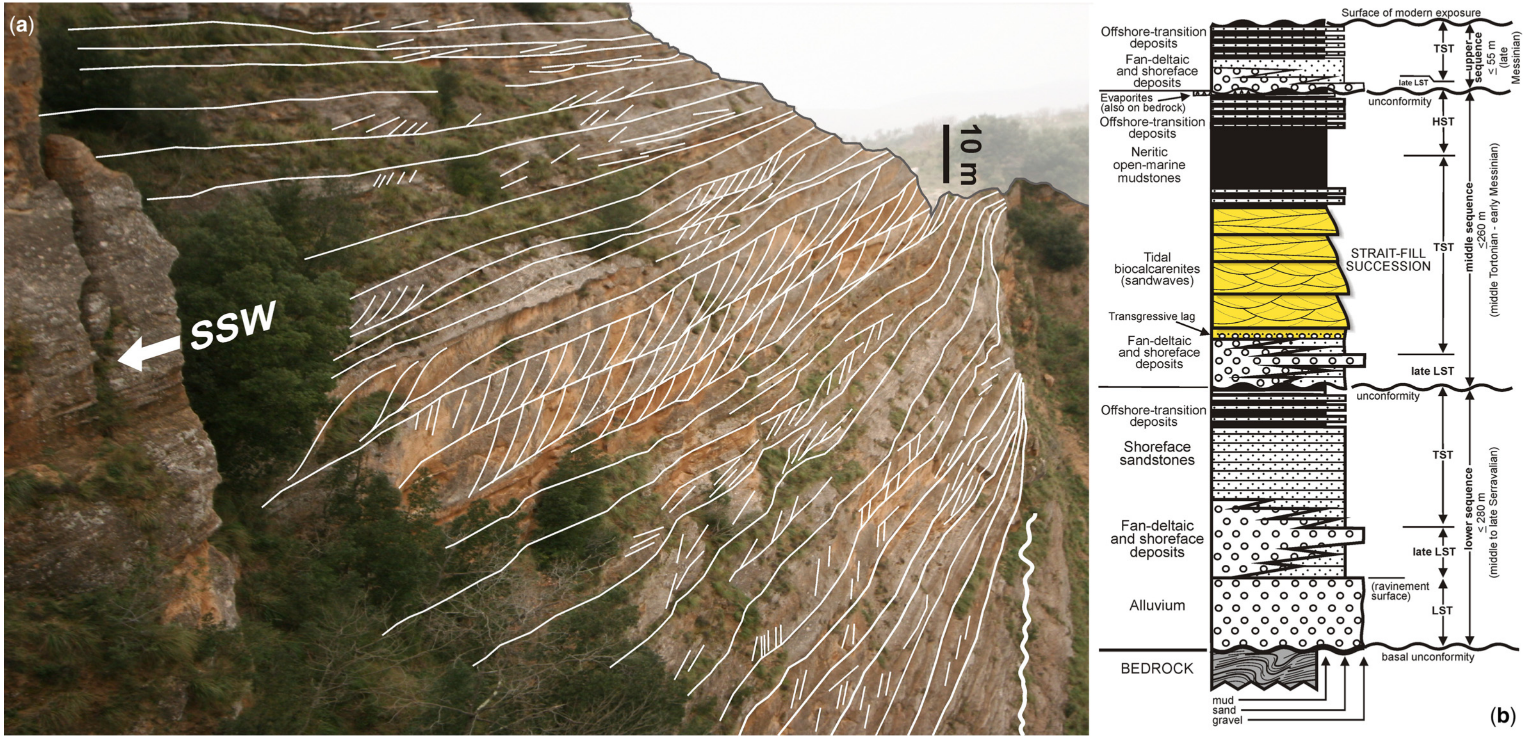

Fig. 6. Strait-fill, tidalite-bearing, succession in the upper Miocene of the Amantea Basin of Calabria, southern Italy (see Fig. 1 for the location). (**a**) Large-scale, tabular-based, cross-beds are exposed near Mt Pellegrino along the axis of a structurally controlled semi-graben. Mixed siliciclastic–bioclastic arenites are organized into unidirectional foresets and record the superimposition of 2D tidal dunes migrating prevalently towards the SSW. (**b**) Schematic stratigraphic column of the Amantea Basin showing the interval recording conditions of a tide-dominated strait (yellow) (modified from Longhitano and Nemec 2005). HST, highstand systems tract; LST, lowstand systems tract.

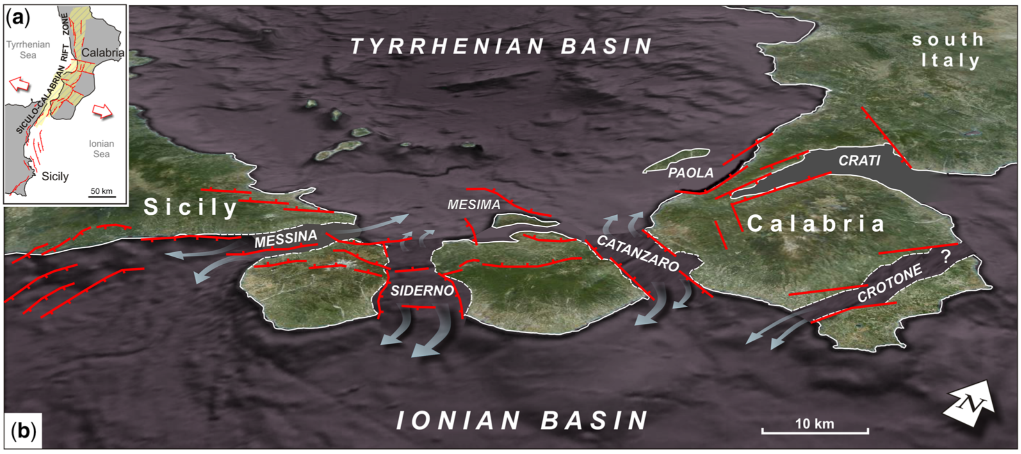

Fig. 7. (**a**) Structural sketch map of Calabria and eastern Sicily in southern Italy, showing the extent (in yellow) of the Sicilian–Calabrian Rift Zone and associated main normal faults (after Tortorici *et al.* 1995; Catalano *et al.* 2008). (**b**) Palaeoenvironmental reconstruction of Calabria and NE Sicily in the Early Pleistocene, showing the main marine basins of the Sicilian–Calabrian Rift Zone, the sedimentary infill of which is nowadays partly exposed in onshore areas (modified after Longhitano 2013).

2022). Consequently, sediment dynamics along the strait are quite variable, with individual sedimentary zones characterized by distinct bottom features and textural/dimensional attributes. Of particular relevance for this paper is the presence of large fields of sand waves at water depths ranging between *c.* 200 and 330 m on both sides of the strait. Such bedforms, of both 2D and 3D geometry, can be up to 15 m in height and reach a wavelength of 120 m. Other, smaller-scale bedforms with a variety of shapes and

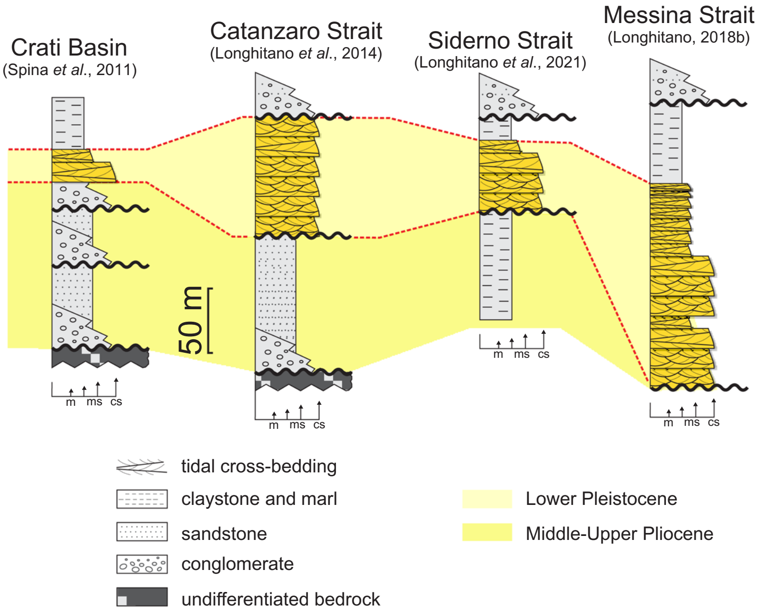

Fig. 8. Simplified stratigraphy of some of the main extensional basins associated with the Sicilian–Calabrian Rift Zone correlated with an interval bearing cross-stratified tidal deposits.

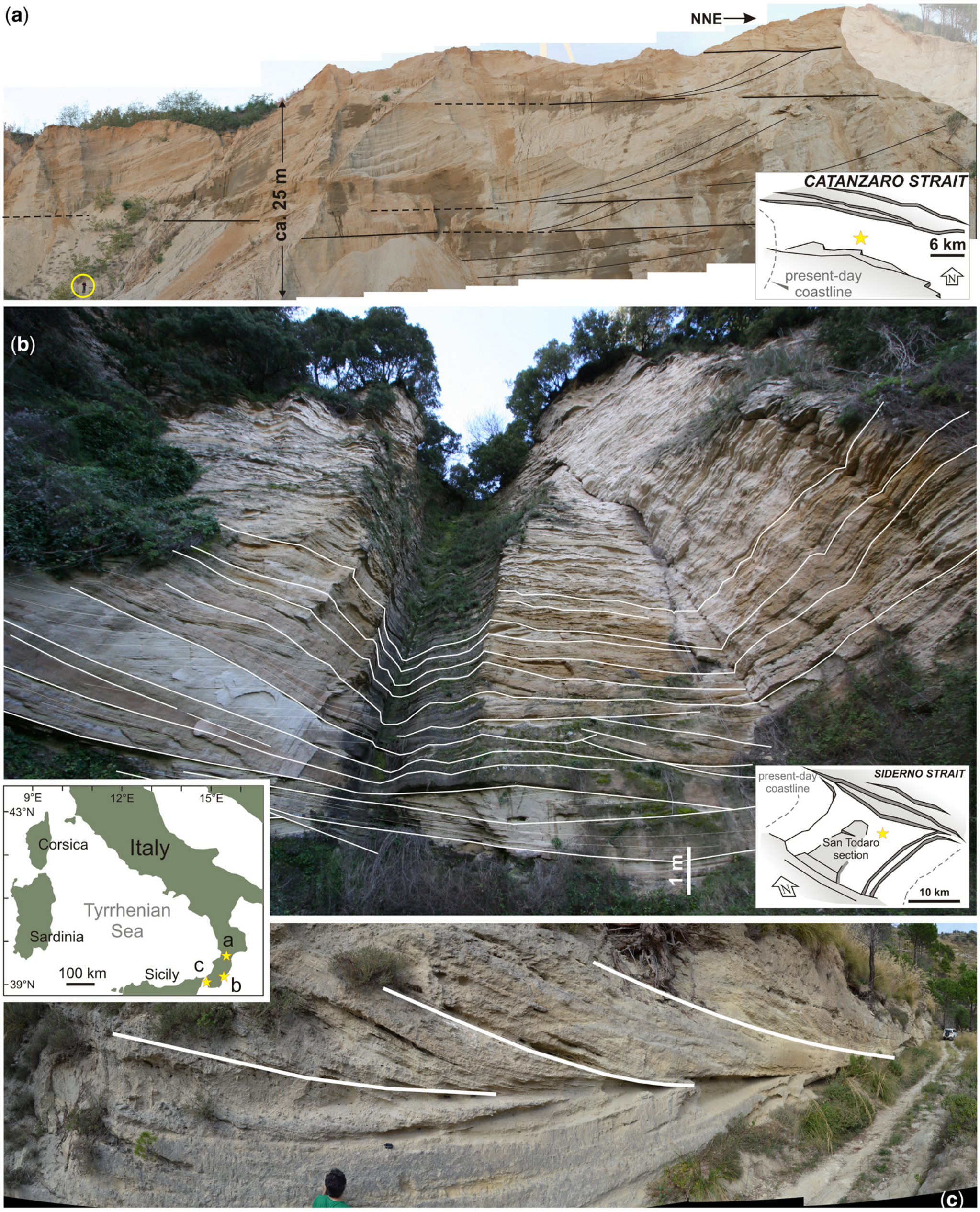

Fig. 9. Palaeostrait sedimentary facies of the Sicilian–Calabrian Rift Zone. (**a**) Very-large-scale cross-bedding due to the migration (towards the SSW, to the left in the photograph) of tidal dunes in the early Pleistocene section of the Catanzaro Strait. (**b**) Strongly aggradational trough (3D), cross-bedding exposed in the northeastern sector of the early Pleistocene Siderno Strait (for location see Fig. 7). (**c**) Tangential foresets rhythmically separated by reactivation surfaces in cross-bedded sandstone deposits of the Pleistocene Messina Strait (modified after Longhitano and Nemec 2005; Longhitano *et al.* 2014, 2021*c*; Rossi *et al.* 2017; Longhitano 2018*b*).

orientation are also present in lower-energy areas laterally adjacent to the larger bedform fields. Because of tidal asymmetry, the general direction of migration of the larger bedforms is towards the north. Sediment composition is mostly siliciclastic with a subordinate (<30%) carbonate fraction.

The Messina Strait is the remnant of a somewhat larger strait active from the latest Pliocene (e.g. Catalano *et al.* 2008; Longhitano 2018*b*). Large quantities of mixed siliciclastic–bioclastic sediment accumulated in the Early Pleistocene under a tidal hydrodynamic regime analogous to the modern one. The 250 m-thick palaeostrait-fill succession is beautifully exposed along the margins of the modern strait. It consists of cross-stratified sandstone, associated with conglomerate and structureless sandstone–siltstone (Figs 8 & 9c). Cross-stratified complexes are interpreted as the sedimentary record of the migration of several superimposed generations of tidal dunes of various size and geometry. The lower boundary of the tidal cross-stratified deposits is a widespread paraconformity that becomes an angular unconformity near the palaeostrait margins; this tectonically induced surface has been compared to similar features at the base of other correlative tidal cross-stratified successions filling adjacent tidal straits of the CPT (Colella and D'Alessandro 1988; Barrier *et al.* 1993; Chiarella *et al.* 2012; Longhitano 2012; Longhitano *et al.* 2012*a*, 2014; Zecchin *et al.* 2015; Rossi *et al.* 2017), as discussed earlier in this paper.

Discussion

Extensive fields of large-scale subaqueous dunes are a distinctive element of modern marine straits, as they represent the most spectacular and easily recognizable result of the increased competence of amplified tidal currents (Dalrymple 2010). Likewise, the interpretation of the global stratigraphic record points to the recurrent existence of ancient epicontinental seaways/straits (e.g. Allen and Homewood 1984; Colella and D'Alessandro 1988; Allen and Bass 1993; Martel *et al.* 1994; Mellere and Steel 1996; Anastas *et al.* 1997, 2006; Martín *et al.* 2001, 2009, 2014; Longhitano and Nemec 2005; Wells *et al.* 2005*a*, *b*; Betzler *et al.* 2006; Blackwood 2006; Longhitano 2011; Longhitano *et al.* 2011, 2012*a*, 2014; Olariu *et al.* 2012; Reynaud *et al.* 2013; Longhitano and Steel 2016; Rossi *et al.* 2017). Modern and fossil cross-stratified strait-fill deposits have common features that distinguish them from similar deposits such as tidal sand ridges and estuary-mouth sand bars: (i) stacked dune successions of unusually large average thickness, which can reach hundreds of metres because of subsidence and lateral confinement; (ii) a monotonous but well-ordered internal stacking of dunes (mud intercalations are rare because of the high-energy setting), occasionally punctuated by erosional master surfaces of regional significance; and (iii) recurrent bidirectionality of foreset beds, reflecting flood and ebb tidal reversals flowing axially to the strait. Other diagnostic characteristics of seaway/palaeostrait successions are listed in Table 1. Being highly recognizable and consistently associated with a discrete palaeoenvironmental setting, thick units of large-scale cross-bedded deposits can be easily identified and interpreted as evidence of the existence of palaeostraits even when they are deformed and structurally incorporated into an orogenic belt. The following is a discussion on the value of such deposits for highlighting the transitory existence of relatively small sedimentary basins in extensional settings such as the western-central Mediterranean during the Oligocene–Quaternary.

Large-scale, cross-bedded lithostratigraphic units as a tectonic facies

Lithospheric fragments of various size can be detached from an extending continental margin, inducing the development of back-arc rift basins. Such fragments can then drift out oceanwards before being accreted along subduction zones and eventually being incorporated into orogenic belts. Microplate rifting/dispersal along active non-collisional continental margins is promoted by decreasing plate convergence rates, inducing roll-back of the subducting plate and upper-plate extension. In collisional settings, upper-plate extension may be active in those embayments of the continental margins where collision did not take place. Either way, due to their geodynamic setting, structural configuration, thermal state and density distribution, back-arc basins are rheologically weak, transitory features sensitive to even minor stress-field variations, and are easily affected by shortening once the regional stress changes from extensional to compressional (e.g. Ziegler 1990; Cloetingh *et al.* 2005). Structural inversion of back-arc basins can result from increased convergence rates along subduction zones, as well from continental collisions, when horizontal compressional stress can travel far from suture zones and focus along pre-existing rheological discontinuities at distances of the order of several hundred kilometres (e.g. Ziegler *et al.* 1998, 2002; Cloetingh *et al.* 2005; Gusmeo *et al.* 2021). As discussed above, the Oligocene–Neogene geological evolution of the western Mediterranean provides a clear example of how rifting/drifting of small lithospheric blocks may temporarily create narrow seaways characterized by distinctive and relatively rare sedimentary facies. We propose here that thick large-scale, cross-bedded units, albeit locally developed, can be considered a 'tectonic facies', in that their peculiar sedimentological features are characteristic of palaeostraits and seaways commonly produced by rifting and microplate drifting (Fig. 10a). Palaeostraits and seaways may also occur in

Table 1. *Diagnostic characteristics of extensional seaway/palaeostrait successions*

Overall thickness	20–50 up to 220 m, pinching towards the strait centre and thickening towards the strait ends[1–3]			
Thickness of single cross-bed sets	2–12 m[4, 5]			
Predominant rock type	Siliciclastic–bioclastic sandstone and granule conglomerate[3–5] with subordinate mudrock, limestone and breccia (strait margin)[6]			
Sedimentary facies[6]	A: Strait centre zone[6, 7]	B: Dune bedded strait zone[1, 3–5, 8–13, 18]	C: Strait end zone[1, 2]	D: Strait margin zone[1, 2, 6, 15, 16]
	Patches of fossiliferous conglomerate, with local crude cross- beds erosionally lying on bare substrate	Metre-thick large-scale trough and planar cross-beds merging downcurrent to facies C and laterally to facies D	Highly bioturbated heterolithic mudstones with decimetre-thick cross-bedded sandstone intercalations	Channelized fan deltas, canyon-head deltas and scree-cone subaqueous aprons with intercalations of facies B
Bedding pattern	No preferential bedding pattern	Strongly aggradational to moderately progradational; monotonous stacking (no mud intercalations)[6]	Aggradational to retrogradational	Progradational to retrogradational[15, 16]
Internal architecture	Erosional discontinuities, metre-deep scours and chutes, locally filled by decimetre-thick coarse-grained cross-beds[1, 2, 12]	Large- to small-scale foresets, herringbone cross-bedding, tidal bundles[1, 3–5, 8–12, 18]	Isolated, dune- to ripple-scale cross-beds with bidirectional foresets [1]	Channel fills associated with supercritical-flow structures, such as backsets, cyclic steps and antidunes, large-scale substrate blocks and isolated foresets[1, 2, 15, 16]
Composition	Dominantly siliciclastic, mostly reflecting the underlying substrate	Variable ratio of siliciclastic (extrabasinal) v. carbonate (intrabasinal) detritus, mostly depending on climate[14]	Dominantly siliciclastic	Variable ratio of siliciclastic (extrabasinal) v. carbonate (intrabasinal)[14]
Structural context	Block-faulted horsts or narrow graben[17]	Depocentres associated with normal-fault hanging walls[1]		Block-faulted, marginal high-angle normal faults[1, 16]

References: [1]Longhitano (2018*b*); [2]Longhitano and Chiarella (2020); [3]Longhitano *et al.* (2014); [4]Anastas *et al.* (1997, 2006); [5]Longhitano *et al.* (2021*c*); [6]Longhitano (2013); [7]Shaw *et al.* (2012); [8] Martín *et al.* (2014); [9]Olariu *et al.* (2012); [10]Reynaud *et al.* (2013); [11]Archer *et al.* (2019); [12]Dalrymple (2010); [13]Longhitano *et al.* (2012*b*); [14]Longhitano *et al.* (2012*a*); [15]Rossi *et al.* (2017); [16]Longhitano *et al.* (2021*a*); [17]Longhitano and Nemec (2005); [18]Longhitano *et al.* (2021*b*).

compressional settings such as flexural foreland basins but their stratigraphy is different from that of their extensional counterparts (Fig. 10b), as discussed later in this section.

A tectonic facies combines the lithological and structural features that allow systematic recognition of past tectonic settings (Robertson 1994). The tectonic facies concept is a field-based approach that allows plate-scale processes to be recognized at the outcrop scale. From this perspective, a number of tectonic facies, representing individual tectonic settings, can be distinguished. Early attempts at categorizing the tectonic facies concept were made by Schaer and Rodgers (1987) and Hsu (1991). Robertson (1994) provided firm criteria to differentiate between some of the most common tectonic settings,

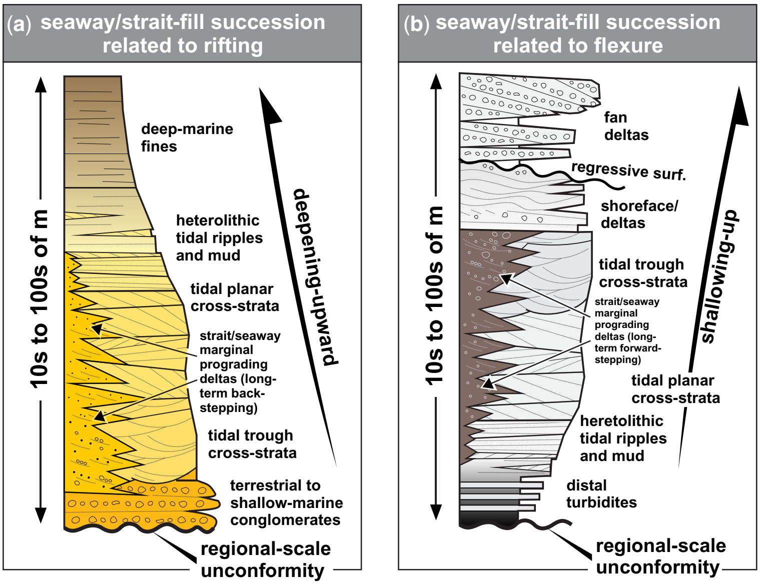

Fig. 10. Conceptual lithostratigraphic columns of straits and seaways in (**a**) rifting and (**b**) flexural tectonic settings. In rift-related straits and seaways (the focus of this study), initial transgression creates optimal conditions for tidal current amplification followed upsection by deeper-marine non-tidal conditions as the basinal cross-section widens. The strait/seaway floor would commonly be subsiding slowly but not necessarily resulting in continuous deepening. In seaway/straits related to lithospheric flexure, thick large-scale, cross-bedded units are encased along a regressive vertical trend, mostly governed by the reduction in accommodation rate. See Figure 1 for the geographical distribution of these two tectonic facies within the Mediterranean Basin. Note the inverted mutual relation between trough- and planar-cross tidal beds in each column, suggested as a key element for distinguishing transgressive v. regressive strait-fill trends (Longhitano and Chiarella 2020).

including, among others, rifts, passive continental margins, seamounts, volcanic arcs, forearcs and back-arcs. The advantage of the application of the tectonic facies concept to the analysis of orogenic belts is that it provides a means for standardization and comparison, as similar tectonic processes produced very similar tectonic facies in different orogens. In order to interpret and palinspastically restore an orogenic belt, one must recognize its main tectonostratigraphic components. A useful means for identifying such components is the tectonic facies concept. The tectonic facies concept is effective for the recognition of many ancient tectonic settings in an orogenic belt, and works particularly well in cases where modern settings are well documented and where facies models for individual tectonic settings are unique.

The western Mediterranean provides snapshots of the sequential phases of microplate creation, dispersal and accretion (Figs 1 & 2). Despite being formed over a time span of *c.* 20 myr, from the Burdigalian to the present, the examples of thick sedimentary successions dominated by large-scale cross-beds discussed above show remarkably consistent sedimentological and structural features pointing to a common origin (Table 1). Late Cenozoic tidal conditions in the western Mediterranean Sea were microtidal, although during much of the Miocene – up until the Messinian, when the Gibraltar region was subjected to shortening – a somewhat wider connection with the Atlantic Ocean made it easier for the Atlantic tide to propagate into the Mediterranean. Based on the descriptions above, the key controlling factor for the development of large dune fields seems to have been the opening of narrow graben during either (i) early rifting or (ii) the oroclinal bending and the ensuing transtensional deformation of microplates during docking (Figs 1 & 2). Even in microtidal settings, such structural configurations may amplify tidal currents through local coastal constrictions such as straits and embayments, leading to an out-of-phase set-up of tidal prisms between adjacent deep-water bodies. Structural/physiographical control on sedimentation style is evident in the present-day Messina Strait and its fossil deposits cropping out along the modern strait margins (Longhitano 2018*a*, *b*). The same conclusion is applicable retrospectively to virtually all significant accumulations of late Cenozoic western Mediterranean tidal deposits, including the Burdigalian–Langhian

Saint-Florent and Bonifacio successions of Corsica, the Aquitanian–Serravallian Sardinian Graben System, the Serravallian–Messinian Amantea Basin, and the Plio-Pleistocene Catanzaro, Siderno and Messina straits (Figs 1, 4 & 7).

The Miocene was characterized by major climatic and oceanographic changes, arguably the result of the final Arabia–Eurasia collision and the definitive closure of the marine connection between the Atlantic and the Indian Ocean. The late Burdigalian–Langhian time frame (*c.* 17–14 Ma) was characterized by a climatic optimum and by eustatic highstands (Miller *et al.* 2020) that could have facilitated the establishment of seaways along flooded continental margins. If some of the palaeostrait successions discussed in this paper were indeed deposited during the Miocene climatic optimum, most other examples are not coeval with such event (Figs 3 & 8) and do not fit the highstands of the global eustatic sea-level curve (Miller *et al.* 2020). The Monte Sant'Angelo Formation of the Saint-Florent Basin in northern Corsica fits precisely the duration of the climatic optimum; nonetheless, it is evident that even sedimentation of such unit was in essence structurally controlled, as shown by widespread field evidence of synsedimentary extensional tectonics (Cavazza *et al.* 2007). In summary, although extensional/transtensional tectonics was a key factor in all examples presented here, we cannot rule out the possibility that eustasy also played a role.

Activation/deactivation of the Mediterranean Sea straits might also have been influenced by short-term changes in water temperature and salinity triggered by variations in the volume of the Atlantic Ocean inflow penetrating into the Mediterranean as a result of variations in the hydraulic section of the Gibraltar corridor(s). The circulation of thermohaline oceanic currents represents a mighty climatic regulator for coastal and sublittoral areas (e.g. Flecker *et al.* 2015; Martin-Garcia 2019).

Geophysical and geological evidence from the western Mediterranean area shows that some segments of the Alboran, Algerian and south-Tyrrhenian back-arc basin margins are presently undergoing compressional tectonics (e.g. Serpelloni *et al.* 2007). Basin inversion will be likely to lead to subduction inception and to the final suture between Nubia and Eurasia (Billi *et al.* 2011), thus bringing to an end a convergence process that started in the Late Cretaceous and is still active, as is also shown by the GPS velocity field. From this viewpoint, the opening of the Provençal–Algerian and Tyrrhenian back-arc basins represents a temporary adjustment in an otherwise continued progression of plate convergence. However, what will be the geological record of the former existence of such back-arc basins after the final continental collision, when the basins are obliterated, and the various microplates amalgamated, deformed and affected by significant along-strike movements? Following subduction of the quantitatively limited back-arc oceanic lithosphere, the tectonostratigraphic assemblage of the resulting collisional orogen will be dominated by deep-marine sedimentary deposits indistinguishable from those of other tectonic settings. In such a context, the presence of thick successions of large-scale, cross-bedded rocks could indicate the transitory existence of tectonically controlled seaways/palaeostraits, thus pointing to episodes of decreasing plate convergence or providing clues as to the morphology of the encroaching continental margin (promontories v. embayments) before final collision.

Extensional palaeostrait successions in the geological record

In a preliminary search of the literature we identified a few potential examples of ancient palaeostrait/seaway successions now deformed to varying degrees and incorporated into orogenic belts (Table 2). In the Late Neoproterozoic Port Askaig Formation of the Dalradian Basin (Scotland), a *c.* 100 m-thick succession of giant cross-beds (individual sets up to 11 m thick) is interpreted to be the result of the migration of large subaqueous dunes in a tide-dominated submarine setting (Arnaud 2004). Deposition occurred during the rifting associated with the break-up of the Rodinia supercontinent in the initial phases of the opening of the Iapetus (proto-Atlantic) Ocean. Abrupt lateral facies variations are attributed to localized extensional faulting delimiting narrow and elongated sub-basins (e.g. Anderton 1985). The closure of the Iapetus Ocean and the resulting Caledonian Orogeny deformed the Dalradian basin-fill, which was then dismembered by the opening of the northern Atlantic.

Strike-slip faulting can create physiographical configurations conducive to the development of narrow seaways, particularly along highly deformed continental margins affected by transform plate boundaries. Different segments of a sinuous strike-slip fault system may experience subsidence (transtension) or uplift (transpression), thus creating bathymetric variations, changes in the hydraulic section and localized current amplifications (e.g. Gardner and Dorsey 2021). Tidal dunes of the Late Triassic succession of East Svalbard were deposited along small transtensional graben during the unsuccessful break-up of Pangaea in the northern Atlantic, and later deformed by Late Cretaceous–Paleocene transpressional deformation to create the Svalbard fold-and-thrust belt (Smyrak-Sikora *et al.* 2020). The entire cycle, from deposition to deformation, occurred along one of the major structural features of the Arctic region, the long-lasting dextral

Table 2. *Potential examples of palaeostrait/seaway successions incorporated into orogenic belts*

Lithostratigraphic unit	Location	Tectonic setting	Main features	References
Port Askaig Formation (Late Neoproterozoic)	Dalradian Basin (Scotland)	Incipient phases of Rodinia break-up and Iapetus Ocean rifting; later incorporated into Scottish Caledonides	Individual cross-beds up to 11 m thick; interbedded with and laterally equivalent of coarse-grained glacially influenced debris-flow basin marginal deposits	Arnaud (2004)
Kvalpynten succession (Late Triassic)	East Svalbard	Oblique rifting during unsuccessful Pangaea break-up; later incorporated into Svalbard fold-and-thrust belt during Late Cretaceous–Paleocene transpressional deformation	Metre-thick cross-beds (co-sets up to 7.5 m thick) interpreted as sand-rich tidal sand dunes and bars reworking marginal deltaic subaqueous environments	Smyrak-Sikora *et al.* (2020)
Waimai, Orahiri and Otorohanga units of Te Kuiti Group (Oligocene–Early Miocene)	North Island of New Zealand	Alternating episodes of transtension and transpression associated with the subduction of the Pacific Plate under the Australian Plate along the Hikurangi subduction zone	15–50 m-thick large-scale, unidirectional cross-stratified biocalcarenite interbedded with horizontally bedded terrigenous siliciclastic siltstone and sandstone	Anastas *et al.* (1997, 2006)
Large-scale, cross-bedded biocalcarenites (Tortonian)	Amantea Basin (southern Italy)	Late Miocene rifting followed by Plio-Quaternary thrusting during docking of the Calabria–Peloritani Terrane	Individual cross-beds up to 7 m thick; aggradational, large-scale, cross-bedding in biocalcarenites deposited under strong bidirectional tidal currents	Longhitano and Nemec (2005)

strike-slip fault system separating northern Greenland from the Arctic and Barents shelves (Ziegler 1989). Another example of thick tidal sediments that were deposited in strike-slip depocentres with alternating episodes of transtension and transpression is the Late Oligocene–Early Miocene succession of the North Island of New Zealand. The Te Kuiti Group contains 15–50 m-thick packages of large-scale, unidirectional cross-stratified calcarenites that were deposited in water depths of 40–60 m in a 50–100 km-wide seaway characterized by a complex internal physiography and bound laterally by palaeobathymetric highs (Anastas *et al.* 1997, 2006). From initial accommodation to basin inversion and uplift, the entire process was the result of the dynamics of the Hikurangi subduction zone and the ensuing development of the strike-slip Alpine Fault.

As described earlier in this chapter, the 120 m-thick Tortonian sand-wave complex of the Amantea Basin in southern Italy was deposited in a narrow palaeostrait formed during incipient Tyrrhenian Sea rifting (Figs 1 & 2) along the western margin of the Calabrian microplate. Large volumes of mixed siliciclastic–carbonate sediment filled the strait under the influence of amplified tidal currents (Longhitano and Nemec 2005; Longhitano and Steel 2016) (Fig. 6). The cumulative structural effects of the docking of the Calabria–Peloritani microplate (Fig. 2d) and isostatic uplift since mid-Pleistocene time have deformed and exposed the palaeostrait succession.

Extensional v. compressional palaeostrait successions

Within the Mediterranean region, palaeostraits formed not only during lithospheric thinning but also during foreland flexure due to tectonic loading caused by encroaching orogenic wedges. This process is best exemplified by the Burdigalian peri-Alpine corridor of the northern Alpine foreland basin (Kuhlemann and Kempf 2002; Bieg *et al.* 2008; Grunert *et al.* 2011) (Fig. 1). Between 20 and 18 Ma, strong mechanical coupling between the Alpine orogenic wedge and its hinterland induced subsidence in the foreland basin and the occasional establishment of a >1000 km-long marine connection between the western Mediterranean and the western Paratethys to the east (e.g. Homewood *et al.* 1986). Sediments of the Burdigalian seaway were mainly deposited under a meso- to macro-tidal regime, as shown by sand waves, bidirectional cross-stratification and tidal-bundle sequences recognized in outcrops in France, Switzerland, southern Germany and Austria (see Bieg *et al.* 2008 for a review). Overall, the stratigraphy of the Burdigalian seaway and the overlying Upper Freshwater Molasse defines a shallowing-upward trend from shallow-marine to continental palaeoenvironments (e.g. Schlunegger *et al.* 1996).

Other smaller-scale examples of palaeostrait environments in a foreland setting are those of the Late Miocene Betic and Rifian marine corridors, north and south, respectively, of the present-day Gibraltar Strait (Fig. 1). They developed during the latest stage of collision of the Betic–Rif arc (Martín *et al.* 2009; Capella *et al.* 2017; Beelen *et al.* 2022). The complex geometry and physiography of these corridors is due to the fact that they were mostly located on a partially submerged orogenic system (wedge-top basins) subjected to a significant strike-slip component of deformation (e.g. Lonergan and White 1997). An early single seaway (the North-Betic Strait: Martín *et al.* 2009) developed in the Serravallian and soon closed in the early Tortonian. In a following phase, the Atlantic–Mediterranean connection featured numerous coalescing seaways that were then progressively confined and reduced in number. Progressive confinement led to current amplification and the deposition of thick sections of giant cross-bedded units. The corridors finally disappeared in the Pliocene with the establishment of the present-day Gibraltar Strait. Despite local variations, the stratigraphy of the Betic and Rifian corridors evolves from upper bathyal depths to continental environments (Krijgsman *et al.* 1999; Martín *et al.* 2001, 2014; Betzler *et al.* 2006; Puga-Bernabéu *et al.* 2022).

Discrimination of ancient (and deformed) extension- and compression-related palaeostrait successions in the stratigraphic record must be based on the analysis of the overall sedimentological/stratigraphic context. ‘Extension-related’ palaeostrait deposits (Fig. 10a) are characteristically part of a relatively simple sedimentary succession commonly directly overlying the basin floor. The overall vertical palaeoenvironmental trend may be deepening upward as progressive rifting induces crustal thinning and ensuing subsidence (Fig. 10a). The siliciclastic component of the sediments has a typical basement-derived petrographical signature (e.g. Dickinson *et al.* 1983). ‘Compressional’ palaeostrait successions within a foreland basin (Fig. 10b), however, are part of a complex basin-fill stratigraphy that may include thick turbidite sections, as well as repeated megacycles of shallow-marine to continental sediments related to orogenic wedge dynamics and related flexure of the foreland lithosphere. The overall vertical palaeoenvironmental trend is shallowing upwards as the entire accommodation space in the foreland basin tends to be progressively filled by orogen-derived sediments with their characteristic petrographical signature. Extensional and contractional palaeostrait successions can be also differentiated through subsidence analysis as the subsidence curves of rift and foreland basins are markedly different (Allen and Allen 2013).

Conclusions

The presence of dunes or compound dunes is a broadly recurring element of modern tidal straits and seaways, despite their variability in terms of sedimentary dynamics and environments. Thick successions of large-scale, cross-bedded sandstone in the rock record may thus indicate the existence of palaeostraits or seaways. As shown in this chapter, in the western Mediterranean distinctive successions of large-scale, cross-bedded deposits are characteristically associated with the creation of back-arc basins during successive episodes of microplate rifting and dispersal. As back-arc basins are transitory features liable to structural inversion by compressional stresses brought about by accretion tectonics or even by relatively minor changes in subduction dynamics (e.g. Cloetingh *et al.* 1989; Horton *et al.* 2016; Gusmeo *et al.* 2021), the presence of strait/seaway successions may be the only clue to former episodes of extension punctuating an overall compressional tectonic context. As such, they may play an important role in terrane analysis of collisional belts and in the correct identification of variations of stress regime obscured by later phases of deformation.

A rigorous sedimentological analysis of the overall stratigraphic context, taking into account sand-body geometries, facies internal heterogeneities, motifs of vertical cross-bedding stacking and diagnostic sedimentary features, is necessary to discriminate between extension- and compression-related palaeostrait successions within deformed collisional orogens. Further study in the peculiar late-collisional tectonic setting of the Mediterranean region may provide insights into the process of tectonic inversion eventually leading to the closure of back-arc basins, final suturing between continents and the incorporation of seaway-related sedimentary successions within the resulting orogens. The recognition of ancient deformed palaeostrait successions as a tectonic facies indicating the former presence of extensional basins may have significant application in the retrodeformation of orogenic belts, the identification of accreted terranes and, more generally, in palaeotectonic and palaeogeographical reconstructions.

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